From: Guinotte, John
To: Grizzle, Betty
Subject: Re: Home range mapping
Date: Wednesday, January 31, 2018 8:10:40 AM
Attachments: romo_range_female2.jpg

Hi Betty, here it is w primary. If it was only RMNP, it looks like it would be 2 females and one male.

John Guinotte
Spatial Ecologist
Ecological Services
U.S. Fish and Wildlife Service
Mountain Prairie Region 6
134 Union Blvd., Lakewood, CO 80228
303-236-4264
john_guinotte@fws.gov

On Wed, Jan 31, 2018 at 7:27 AM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:

> See message below from Bob Inman, recommending adding his PH.
>
> ---------- Forwarded message ----------
> From: **Inman, Bob** <bobinman@mt.gov>
> Date: Tue, Jan 30, 2018 at 3:46 PM
> Subject: RE: Home range mapping
> To: "Grizzle, Betty" <betty_grizzle@fws.gov>
>
> Hey Betty,
>
> I would like to see primary habitat on there too. Might be able to get a better feel of how they might spread themselves out that way.
>
> Five females and one male is a ratio that I have not heard of before. 2-3 females per male is typical, so if 5 females are in the area, I would think 2 males would be appropriate. In Grand Teton Park, we essentially had 2 pairs, a male and female in the south end and the same in the north end.
>
> Hope you are well.
>
> -Bob Inman

Robert M. Inman, PhD

Carnivore-Furbearer Coordinator

Montana Fish, Wildlife and Parks

1420 East 6th Ave., PO Box 200701,

Helena, MT 59620-0701

406-444-0042 (o)

406-570-5326 (c)

bobinman@mt.gov

From: Grizzle, Betty [mailto:betty_grizzle@fws.gov]
Sent: Tuesday, January 30, 2018 4:19 PM
To: Inman, Bob <bobinman@mt.gov>
Subject: Home range mapping

Hi Bob - We did a quick spatial analysis to estimate the potential number of female wolverines that could occupy the Rocky Mountain National Park area. This is one area included in our climate change/snow persistence modeling analysis. We used home range estimates from your Greater Yellowstone study (approximate) and your maternal habitat layer. I am attaching a Word version of the map my colleague in Region 6 prepared (original jpg is too large to send).

What are your thoughts on this? It seems to show that you might be able to have (at most) 5 females in this study area, but maybe only 1 male. Does that make sense?

I am teleworking today, but back in the office tomorrow, if you have questions.

Thanks for your time,

Betty

--

Betty J. Grizzle, D.Env.

Fish and Wildlife Biologist

U.S. Fish and Wildlife Service

Carlsbad Fish and Wildlife Office

2177 Salk Ave, Suite 250

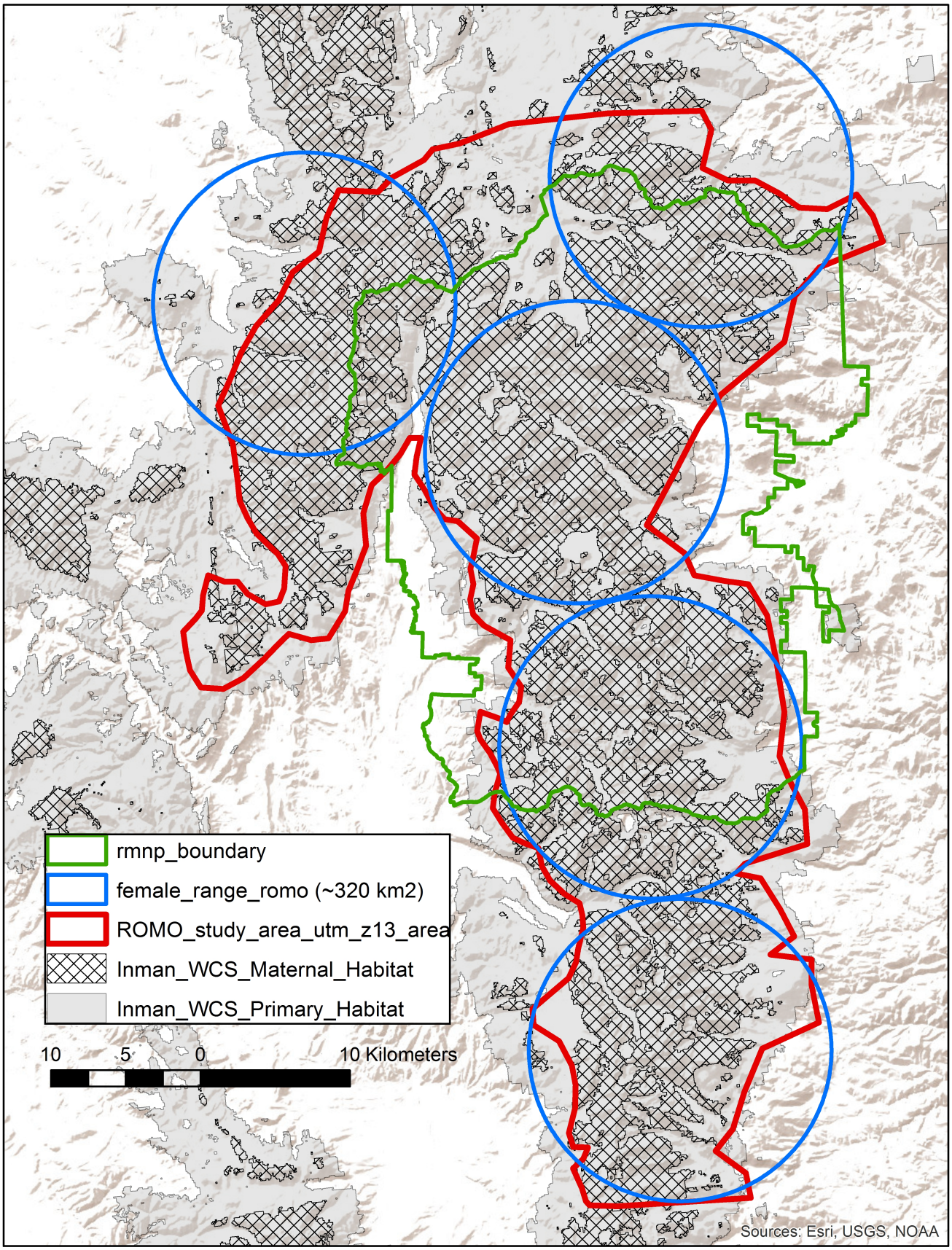
Carlsbad, CA 92008

760-431-9440, ext. 215

760-431-5901 fax

--

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760-431-5901 fax


rmnp_boundary
female_range_romo (~320 km2)
ROMO_study_area_utm_z13_area
Inman_WCS_Maternal_Habitat
Inman_WCS_Primary_Habitat
10 5 0 10 Kilometers
Sources: Esri, USGS, NOAA

From: Grizzle, Betty
To: Inman, Bob
Subject: Re: Home range mapping
Date: Wednesday, January 31, 2018 8:23:42 AM
Attachments: Rocky Mtn NP_Wolv home range.docx

Thanks for your review. Here is another version with the primary habitat added.
For Rocky Mtn National Park only (the area outlined in **green**) would 2 females and 1 male seem about right?

On Tue, Jan 30, 2018 at 3:46 PM, Inman, Bob <bobinman@mt.gov> wrote:

> Hey Betty,
>
> I would like to see primary habitat on there too. Might be able to get a better feel of how they might spread themselves out that way.
>
> Five females and one male is a ratio that I have not heard of before. 2-3 females per male is typical, so if 5 females are in the area, I would think 2 males would be appropriate. In Grand Teton Park, we essentially had 2 pairs, a male and female in the south end and the same in the north end.
>
> Hope you are well.
>
> -Bob Inman
>
> Robert M. Inman, PhD
>
> Carnivore-Furbearer Coordinator
>
> Montana Fish, Wildlife and Parks
>
> 1420 East 6th Ave., PO Box 200701,
>
> Helena, MT 59620-0701
>
> 406-444-0042 (o)
>
> 406-570-5326 (c)
>
> bobinman@mt.gov

From: Grizzle, Betty [mailto:betty_grizzle@fws.gov]
Sent: Tuesday, January 30, 2018 4:19 PM
To: Inman, Bob <bobinman@mt.gov>
Subject: Home range mapping

Hi Bob - We did a quick spatial analysis to estimate the potential number of female wolverines that could occupy the Rocky Mountain National Park area. This is one area included in our climate change/snow persistence modeling analysis. We used home range estimates from your Greater Yellowstone study (approximate) and your maternal habitat layer. I am attaching a Word version of the map my colleague in Region 6 prepared (original jpg is too large to send).

What are your thoughts on this? It seems to show that you might be able to have (at most) 5 females in this study area, but maybe only 1 male. Does that make sense?

I am teleworking today, but back in the office tomorrow, if you have questions.

Thanks for your time,

Betty

--

Betty J. Grizzle, D.Env.

Fish and Wildlife Biologist

U.S. Fish and Wildlife Service

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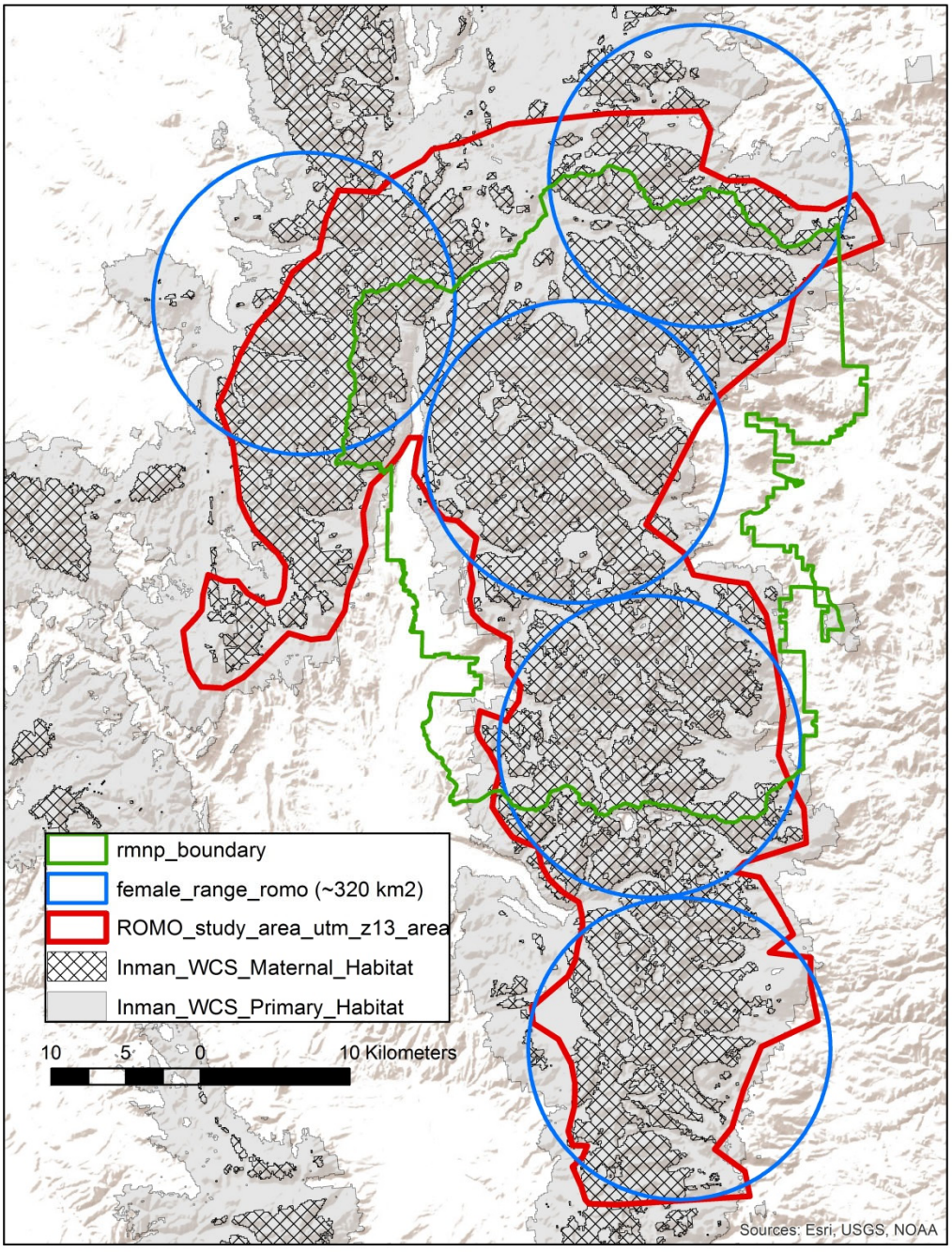
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--

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rmnp_boundary
female_range_romo (~320 km2)
ROMO_study_area_utm_z13_area
Inman_WCS_Maternal_Habitat
Inman_WCS_Primary_Habitat
10
5
0
10 Kilometers
Sources: Esri, USGS, NOAA

From: Grizzle, Betty
To: Bush, Jodi
Cc: Guinotte, John; Shoemaker, Justin; Stephen Torbit
Subject: Re: Request to use map (from your 2016 paper)
Date: Wednesday, January 31, 2018 9:16:26 AM

I have asked Shevenell if she can add the 8 den locations to the map she sent. Not sure if they are able to release this information just yet, but I will cite this new (and important) den location information as pers. comm.

On Wed, Jan 31, 2018 at 8:12 AM, Bush, Jodi <jodi_bush@fws.gov> wrote:

> fine with me. JB
>
> Jodi L. Bush
> Office Supervisor
> Montana State Ecological Services Office
> 585 Shepard Way, Suite 1
> Helena, MT 59601
> (406) 449-5225, ext.205
>
> On Tue, Jan 30, 2018 at 10:11 AM, Guinotte, John <john_guinotte@fws.gov> wrote:
>
> > The color version makes a big difference. I think it should go in the SSA not appendix.
> >
> > John Guinotte
> > Spatial Ecologist
> > Ecological Services
> > U.S. Fish and Wildlife Service
> > Mountain Prairie Region 6
> > 134 Union Blvd., Lakewood, CO 80228
> > 303-236-4264
> > john_guinotte@fws.gov
> >
> > On Tue, Jan 30, 2018 at 10:01 AM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:
> >
> > > FYI - Looks like we can use the map Shevenell provided in her email (with accompanying credit). I recommend placing this directly in the SSA Report rather than in another appendix. Thoughts?
> > >
> > > ---------- Forwarded message ----------
> > > From: **Shevenell Webb** <Shevenell.Webb@ab-conservation.com>
> > > Date: Tue, Jan 30, 2018 at 8:55 AM
> > > Subject: Re: Request to use map (from your 2016 paper)
> > > To: "Grizzle, Betty" <betty_grizzle@fws.gov>
> > >
> > > Hi Betty,
> > >
> > > I have attached a color map version showing spring snow cover and wolverine locations in Alberta. We are working on a manuscript right now on Characteristics of wolverine dens in the lowland boreal forest of north-central Alberta. We visited 8 unique

wolverine dens outside Copeland et al. spring snow cover layer. I don't know when USFWS will be making final recommendations, but we should have a decent draft available in ~1 month if that is at all useful?

cheers, Shevenell

_____

Shevenell Webb, Wildlife Biologist

Alberta Conservation Association

phone: 207-242-2875

email: shevenell.webb@ab-conservation.com

**I am in the office Tuesday, Thursday, and Friday.

From: Grizzle, Betty <betty_grizzle@fws.gov>
Sent: Friday, January 26, 2018 10:54 AM
To: Shevenell Webb
Subject: Request to use map (from your 2016 paper)

Hi Shevenell - We are addressing a few final comments on our wolverine status review. Regarding your 2016 paper in The Journal of Wildlife Management, would it be possible for our agency to use Figure 1 from that paper in our status review document? If so, do you have a better quality image than what is presented in the published paper?

Thanks so much and let me know if you have questions about this request,
Betty

--

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Carlsbad Fish and Wildlife Office
2177 Salk Ave, Suite 250
Carlsbad, CA 92008

> > 760-431-9440, ext. 215
> > 760-431-5901 fax
>
> --
>
> Betty J. Grizzle, D.Env.
> Fish and Wildlife Biologist
> U.S. Fish and Wildlife Service
> Carlsbad Fish and Wildlife Office
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--

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760-431-5901 fax

From: Inman, Bob
To: Grizzle, Betty
Subject: RE: Home range mapping
Date: Wednesday, January 31, 2018 9:27:43 AM

That looks very reasonable to me. Thanks Betty

-Bob Inman

Robert M. Inman, PhD
Carnivore-Furbearer Coordinator
Montana Fish, Wildlife and Parks
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Helena, MT 59620-0701
406-444-0042 (o)
406-570-5326 (c)
bobinman@mt.gov

From: Grizzle, Betty [mailto:betty_grizzle@fws.gov]
Sent: Wednesday, January 31, 2018 8:24 AM
To: Inman, Bob <bobinman@mt.gov>
Subject: Re: Home range mapping

Thanks for your review. Here is another version with the primary habitat added.
For Rocky Mtn National Park only (the area outlined in **green**) would 2 females and 1 male seem about right?

On Tue, Jan 30, 2018 at 3:46 PM, Inman, Bob <bobinman@mt.gov> wrote:

> Hey Betty,
>
> I would like to see primary habitat on there too. Might be able to get a better feel of how they might spread themselves out that way.
>
> Five females and one male is a ratio that I have not heard of before. 2-3 females per male is typical, so if 5 females are in the area, I would think 2 males would be appropriate. In Grand Teton Park, we essentially had 2 pairs, a male and female in the south end and the same in the north end.
>
> Hope you are well.
>
> -Bob Inman
>
> Robert M. Inman, PhD
> Carnivore-Furbearer Coordinator

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406-570-5326 (c)
bobinman@mt.gov

From: Grizzle, Betty [mailto:betty_grizzle@fws.gov]
Sent: Tuesday, January 30, 2018 4:19 PM
To: Inman, Bob <bobinman@mt.gov>
Subject: Home range mapping

Hi Bob - We did a quick spatial analysis to estimate the potential number of female wolverines that could occupy the Rocky Mountain National Park area. This is one area included in our climate change/snow persistence modeling analysis. We used home range estimates from your Greater Yellowstone study (approximate) and your maternal habitat layer. I am attaching a Word version of the map my colleague in Region 6 prepared (original jpg is too large to send).

What are your thoughts on this? It seems to show that you might be able to have (at most) 5 females in this study area, but maybe only 1 male. Does that make sense?

I am teleworking today, but back in the office tomorrow, if you have questions.
Thanks for your time,
Betty

--

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--

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760-431-5901 fax

From: Grizzle, Betty
To: A J
Subject: Re: new paper
Date: Wednesday, January 31, 2018 11:18:05 AM

I am finishing up a few edits based on comments received. The effective population size topic is a bit of a challenge.
Feel free to give me a call and I can fill you in on Shevenell's info.

On Wed, Jan 31, 2018 at 10:13 AM, A J <222wsheridan@gmail.com> wrote:

> Hi Betty
> I don't know what else Malin is preparing. Have you contacted her? I know she has been in the field but may be in contact now.
>
> Yes, I know Shevenell. I didn't know about her draft paper.
>
> I have been studying the Heinemeyer report. I've been in touch with Jenny Mattisson about use distribution in wolverine home ranges when determined by GPS over part of the year. She sent me several home ranges showing partial use of home ranges even without human use like recreation. It is quite common for wolverines to limit movements to various portions of there home ranges seasonally whether they are denning or not. So I don't think the analysis in the report can be used to conclude that the recreation footprint explains the use distribution in their study areas. The areas "avoided" may have been "not used" even without a human footprint.
> How is it going?
> Audrey
>
> Sent from my iPhone
>
> On Jan 31, 2018, at 7:58 AM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:
>
> > Thanks for sending. Do you know if the other manuscripts in Malin's thesis are due to publish soon? (e.g., "The link between local territorial dynamics and dispersal patterns for female wolverines").
> >
> > Also, do you know Shevenell Webb in Alberta (Alberta Conservation Association)? She has been helpful in providing data to inform our report and they are preparing a (still draft) report documenting wolverine den locations.
> >
> > On Tue, Jan 30, 2018 at 1:10 AM, A J <222wsheridan@gmail.com> wrote:
> >
> > > new paper from Sweden
> >
> > --
> >
> > Betty J. Grizzle, D.Env., MPH
> > Fish and Wildlife Biologist
> > U.S. Fish and Wildlife Service
> > Carlsbad Fish and Wildlife Office

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Carlsbad, CA 92008
760-431-9440, ext. 215
760-431-5901 fax

From: Grizzle, Betty
To: Bryon Holt
Subject: Help with locating references
Date: Wednesday, January 31, 2018 11:26:43 AM
Attachments: Sospedra-Alfonso et al 2015_highlighted.pdf

Hi Bryon - You offered to help with finalizing the SSA Report, so am writing to ask if you can locate and download the references highlighted in yellow on the first page of the attached publication. There are 7 total and full citations are at the end.

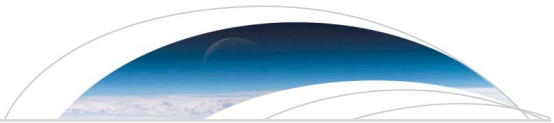
Please let me know whether or not you have time to help. If not, I can ask others.
Thanks.

--

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AGU PUBLICATIONS

Geophysical Research Letters





RESEARCH LETTER
10.1002/2015GL063898

Key Points:
- An effective elevation divides P- and T-driven alpine snowpack variability
- This threshold elevation is defined for a large, thermally heterogeneous region
- T control of the snowpack increases linearly with mean winter temperatures

Supporting Information:
- Figure S1
- Text S1, Figure S1, and Tables S1 and S2

Correspondence to:
R. Sospedra-Alfonso,
sospedra@uvic.ca

Citation:
Sospedra-Alfonso, R., J. R. Melton, and W. J. Merryfield (2015), Effects of temperature and precipitation on snowpack variability in the Central Rocky Mountains as a function of elevation, *Geophys. Res. Lett.*, *42*, 4429–4438, doi:10.1002/2015GL063898.

Received 18 MAR 2015
Accepted 2 MAY 2015
Accepted article online 11 MAY 2015
Published online 2 JUN 2015

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Effects of temperature and precipitation on snowpack variability in the Central Rocky Mountains as a function of elevation

Reinel Sospedra-Alfonso[1], Joe R. Melton[1], and William J. Merryfield[1]

[1]Canadian Centre for Climate Modelling and Analysis, Environment Canada, University of Victoria, Victoria, British Columbia, Canada

Abstract We employ a regression-based methodology to study the impact of temperature and precipitation on snowpack variability as a function of elevation in the Central Rocky Mountains. Because of the broad horizontal coverage and thermal heterogeneity of the measurement sites employed, we introduce an elevation correction based on the sites' climatological temperature. For the elevation range investigated (1295–2256 m), and assuming an average atmospheric lapse rate of −6.5°C/km, we find a mostly linear relationship between effective elevation and correlation of temperature or precipitation with snow water equivalent and snowpack duration. We estimate a threshold elevation, 1560 ± 120 m, below (above) which temperature (precipitation) is the main driver of the snowpack. This threshold elevation is robust under a range of assumed atmospheric lapse rates. Locations below this elevation are likely to be affected by projected rising temperatures, with important effects on ecosystems and economic activities dependent on snow.

1. Introduction

Seasonal snow cover in mountainous regions has a direct effect on local climate and hydrology as well as the functioning of alpine and subalpine ecosystems [*Pelto*, 2008; *Keller et al.*, 2005]. By acting as frozen storage in the water balance, the snowpack provides an important source of freshwater recharge in the spring, influencing runoff, soil moisture, and groundwater [*Barnett et al.*, 2005]. Alpine snow also affects human activities through its effects on water supply and hydroelectric power generation, which is the dominant source of electricity in western Canada and the northwestern United States [*NWPCC*, 2010; *Utzig and Schmidt*, 2011], as well as the economic activities based on winter sports and tourism [*Burakowski and Magnusson*, 2012].

Because of the importance of the alpine snowpack, the study of its spatial and temporal behavior as well as the processes which control the snowmelt is crucial. The properties of the snowpack can be largely inferred from temperature, solar radiation, and precipitation. The primary conditions for the accumulation of snow are sufficiently cold air temperature to favor precipitation in the form of snow, and the persistence of temperature below freezing to preserve the snowpack. Snowmelt on the other hand is largely driven by downwelling shortwave and longwave radiations [*Sicart et al.*, 2006]. In many studies, air temperature and precipitation have been used as the sole explanatory variables of the snowpack [*Cayan*, 1996; *Hamlet et al.*, 2005; *Mote*, 2006; *Brown and Mote*, 2008; *Beniston*, 2012; *Morán-Tejeda et al.*, 2013a]. This is in part because air temperature is known to serve as a good proxy indicator of snowmelt [*Anderson*, 2002] and because there is generally a lack of radiation observations, whereas air temperature and precipitation are readily available from climatological and operational hydrometeorological networks. Several studies have shown that for most conditions, temperature-based snowmelt models (e.g., SNOW-17) [*Anderson*, 1973] perform comparably to energy balance models [*Ohmura*, 2001; *Zappa et al.*, 2003; *Franz et al.*, 2008].

Among previous investigations that considered temperature and precipitation as snowpack drivers, *López-Moreno* [2005] studied the impact of temperature and precipitation in the interannual variability of snow depth in the Spanish Pyrenees. *Morán-Tejeda et al.* [2013b] carried out similar studies for several Iberian mountainous regions, aiming to quantify the sensitivity of the snowpack to the joint influence of temperature and precipitation. For western North America, the impacts of rising temperatures on the reduction of the mountain snowpack and the retreat of glaciers were studied by *Cayan* [1996], *Knowles et al.* [2006], *Hamlet et al.* [2005], *Barnett et al.* [2008], and *Pelto* [2008] among others. *Mote* [2003, 2006] examined the roles of

temperature and precipitation on April snow water equivalent (SWE) trends and sensitivity to warming temperatures. *Pierce et al.* [2008] showed that these warming temperatures cannot be explained by natural variability alone and concluded that anthropogenic warming is taking place in the western U.S. *Klos et al.* [2014] estimated the rain-snow transition zone in the western U.S. for historic and projected climate and discussed implications for the timing of the SWE peak. *Schnorbus et al.* [2014] investigated the effect of increasing temperatures on historical and projected SWE in three British Columbia watersheds.

Recent work has extended such approaches by examining the relative influences of temperature and precipitation as a function of elevation. *Morán-Tejeda et al.* [2013a] considered three mountainous regions of Switzerland and found that the influences of temperature and precipitation on snowpack variability vary approximately linearly with elevation; the impact of temperature tends to decrease with altitude, whereas that of precipitation tends to increase. The elevation at which these influences are equal defines a threshold below (above) which temperature (precipitation) is the main driver of the snowpack. For the Swiss mountains, the authors estimated a threshold elevation equal to 1400 ± 200 m. By analyzing long-term records of daily snow depth, air temperature and precipitation, the authors concluded that there has been a slight increase of this threshold elevation (approximately 140 m to 280 m since the late 1960s, depending on the snow index considered), which may be related to the atmospheric warming in the Swiss alpine region over the past few decades.

The current study applies this type of analysis to the U.S. Rocky Mountains of northern Idaho and northwestern Montana and generalizes its methodology to enable larger, thermally heterogeneous regions to be considered. In assessing the elevation dependence of temperature and precipitation effects on snowpack variability in this region, we rely on snow and meteorological data from measurement sites that are scattered over a relatively wide latitude and longitude range, covering an area approximately twice as large as the Swiss alpine region considered by *Morán-Tejeda et al.* [2013a]. This poses a challenge since temperatures at a given altitude tend to decrease as latitude increases. Also, the eastern side of the central Rockies has a continental climate with relatively cold winters, whereas the western side has some maritime influence with more moderate winter temperatures. Therefore, we must take into account the effect of temperature on snowpack variability not only as a function of elevation but also as a function of horizontal position, and as a consequence the methodology of *Morán-Tejeda et al.* [2013a] cannot be applied straightforwardly. Instead, this methodology is extended to assign effective elevations to the measurement sites based on their climatological temperature, in a manner consistent with an average atmospheric lapse rate. Specifically, we assume a linear relationship between the sites' effective elevations and their climatological temperatures given by an average adiabatic lapse rate, taken as -6.5°C/km [*Wallace and Hobbs*, 2006; *Roe and O'Neal*, 2005; *Arnold et al.*, 2006]. Since the actual lapse rate may differ from this value [*Minder et al.*, 2010], we further show that the threshold elevation separating precipitation-controlled snow regions from those mainly controlled by temperature is nearly insensitive to the precise value chosen.

The region of study and climate data are discussed in section 2.1. In section 2.2, we discuss the methodology and describe the snow and meteorological indices introduced by *Morán-Tejeda et al.* [2013a]. The sites' elevation corrections based on climatological temperature is described in section 2.3. We describe and discuss the results in section 3, and conclusions are provided in section 4.

2. Region of Study, Data, and Methods

2.1. Region of Study and Climate Data

We examine the eastern central region of the Columbia river basin in the Central Rocky Mountains, encompassing northern Idaho and northwestern Montana, U.S. Winter (December–March) temperatures in this region are typically below -3°C. Air masses which generally arrive from the west deposit moisture over the mountains, whose orography can trigger abundant precipitation, mostly in the form of snow at high elevations. November has more precipitation than any other month, SWE peaks in April, and the SWE/P ratio, characterizing the fraction of solid precipitation, peaks in January [*Serreze et al.*, 1999].

Automated daily observations of temperature (T), precipitation (P), and SWE are from the Snow Telemetry (SNOTEL) system operated by the Natural Resources Conservation Service (NRCS) of the United States Department of Agriculture (www.wcc.nrcs.usda.gov/snow/). To maximize the number of sites and time records while maintaining limited spatial horizontal coverage, we employ data that span 1996–2014, from 38 sites distributed over $\sim 300 \times 300$ km^2 with elevations ranging from 1295 m to 2256 m (Figure S1 and Table S1

in the supporting information). This period is sufficiently long to enable the snow climatology analysis performed here, although it is too short for long-term trends to be analyzed as in *Morán-Tejeda et al.* [2013a]. The climate records are mostly complete, except as noted in Table S1 and are used as provided by the U.S. NRCS without any quality assurance procedure [e.g., *Durre et al.*, 2010]. All other sites in this region were discarded due to short time span, longer gaps in their data records, or other reasons described in the supporting information.

Oyler et al. [2015] has recently pointed out systematic errors in temperature measurements at some SNOTEL sites which affect inferred warming trends across the western U.S. alpine region. Since we do not examine trends but only the influences of temperature and precipitation on snowpack variability, we do not expect these systematic errors to have a significant impact on our results.

2.2. Snow and Meteorological Indices

We employ the snow and meteorological indices introduced by *Morán-Tejeda et al.* [2013a] to characterize the interannual variability of the snowpack amount and duration, except that here we consider SWE instead of snow depth, measure the "abundant" snowpack with the climatological SWE 70th rather than 90th percentile (see below), and employ more meteorological indices to account for the various lengths of the sites' snow seasons. In addition, we apply a climatological temperature-based correction to the measurement sites' elevations, discussed in the next section, to compensate for the broad difference in latitude and continentality which make temperature changes in the region dependent not only on elevation.

The snow indices considered as in *Morán-Tejeda et al.* [2013a] are the monthly mean (obtained from daily values) of SWE for January (SWE_{Jan}), February (SWE_{Feb}), March (SWE_{Mar}), and April (SWE_{Apr}). Also, as a measure of the snowpack duration (SD), we use the number of days in the snow year between October and June with more SWE than the climatological SWE 50th and 70th percentiles (SD_{50} and SD_{70}, respectively, denoting a measure of the "normal" and "abundant" snowpack), with the climatological values obtained as daily mean SWE from October to June averaged over 1996–2014.

The meteorological indices, considered predictors of the snow indices, are defined as follows. For every monthly snow index, i.e., SWE_{Jan}, SWE_{Feb}, SWE_{Mar}, and SWE_{Apr}, we define several climatic predictors characterizing the aggregation of precipitation and temperature of previous months, starting in October. For instance, SWE_{Jan} has four predictor pairs of temperature and precipitation given by ($T_{Oct–Jan}$, $P_{Oct–Jan}$), ($T_{Nov–Jan}$, $P_{Nov–Jan}$), ($T_{Dec–Jan}$, $P_{Dec–Jan}$), and (T_{Jan}, P_{Jan}) denoting the average (accumulated) temperature (precipitation) from October to January, November to January, December to January, and in January, respectively. Similarly, SWE_{Feb} has five predictor pairs, and so forth. We employ eight predictor pairs for the snowpack duration indices, SD_{50} and SD_{70}, defined by the temperature (precipitation) averages (accumulation) over October–March, November–March, October–April, November–April, October–May, November–May, October–June, and November–June. These aggregated months account for the lengths of the sites' snow seasons.

For each SNOTEL site, we produce annual values (1996–2014) of each snow index and compute the Pearson correlation with the time series of the corresponding meteorological indices. To illustrate the predictive skill of the meteorological indices, we show in Figure 1 observed and estimated monthly mean SWE using multiple linear regression of temperature and precipitation predictors ($T+P$) for three different elevations. The coefficient of determination (R^2) between observed and estimated SWE is used as a diagnostic of the linear model performance. For comparison, we also show performance when only the T or P index is used as predictor. These regression relations could in theory be applied to forecasting of SWE based on seasonal predictions of T and P. This potential application, however, lies outside the scope of the present study.

At the highest elevation site (Hawkins Lake, 1966 m), SWE_{Apr} estimated with predictors $T_{Fe–Apr} + P_{Oct–Apr}$ provides a very good agreement with observations ($R^2 = 0.93$). At this site, the agreement is almost as good if we only use $P_{Oct–Apr}$ as predictor ($R^2 = 0.89$), but much reduced if we use only $T_{Feb–Apr}$ ($R^2 = 0.19$). This indicates that precipitation is the main driver of the snowpack at this elevation. For this high-elevation site, the best performance is obtained with the P predictor having the highest number of aggregated months implying that maximizing the memory of previous snowfalls yields a better estimate of a precipitation-driven snowpack. For the intermediate elevation site (Mosquito Ridge, 1585 m) observed SWE_{Apr} has good agreement with estimated values obtained with predictors $T_{Jan–Apr} + P_{Dec–Apr}$ ($R^2 = 0.81$). The performance is almost as good if we use $P_{Dec–Apr}$ alone ($R^2 = 0.80$) but is also significant if we use $T_{Jan–Apr}$ ($R^2 = 0.57$) as predictor. This indicates that temperature has an increasing role in controlling the snowpack at decreasing elevation, although

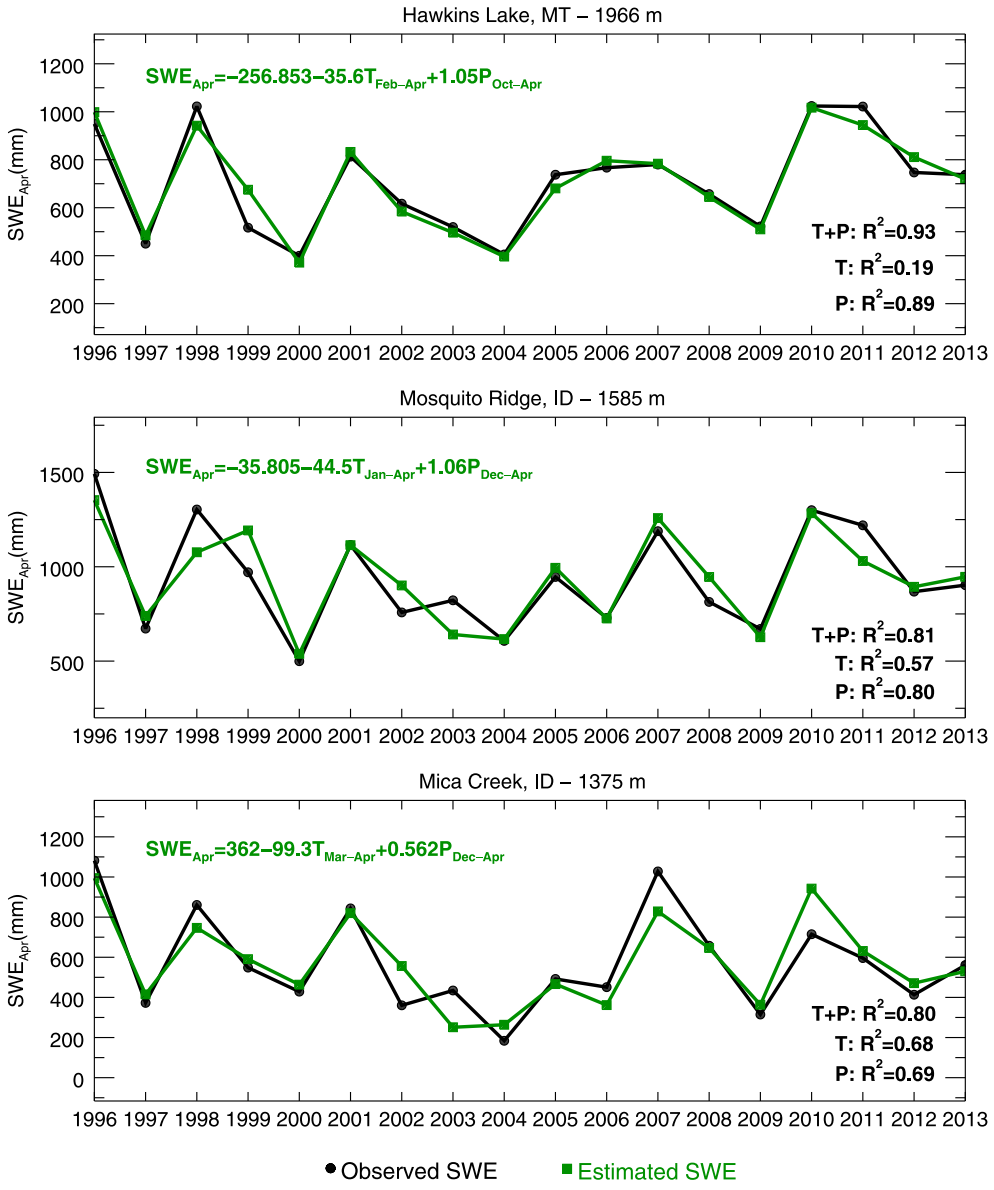


Figure 1. Observed (black) and modeled (green) monthly mean SWE at three different elevations. Modeled SWE is obtained by multiple linear regression of predictors *T* and *P*, as indicated by the linear model shown in the figure. Agreement between observed and modeled SWE is quantified with the coefficient of determination (R^2). Also, R^2 is shown for SWE modeled with *T* or *P* predictors alone.

precipitation is still the main driver of the snowpack here. For the site with the lowest elevation among the three (Mica Creek, 1375 m), estimated SWE_{Apr} using both $T_{Mar–Apr}+P_{Dec–Apr}$ is in good agreement with observations ($R^2 = 0.80$). However, the performance of SWE estimated with $T_{Jan–Feb}$ ($R^2 = 0.68$) or $P_{Jan–Feb}$ ($R^2 = 0.69$) as predictor is about the same. This indicates that both temperature and precipitation drive the snowpack at this location approximately equally. The examples depicted in Figure 1 show the increasing (decreasing) role of temperature (precipitation) in controlling the snowpack with decreasing elevation.

2.3. Elevation Correction Based on Temperature Climatology

Because of the wide horizontal distribution (~3° latitude ×4° longitude) and thermal heterogeneity of the measurement stations, the effect of temperature on snowpack variability depends not only on elevation but also on latitude, longitude, and slope orientation. As a result, we cannot apply the analysis of *Morán-Tejeda et al.* [2013a] straightforwardly, but first introduce an elevation correction to the observation sites based

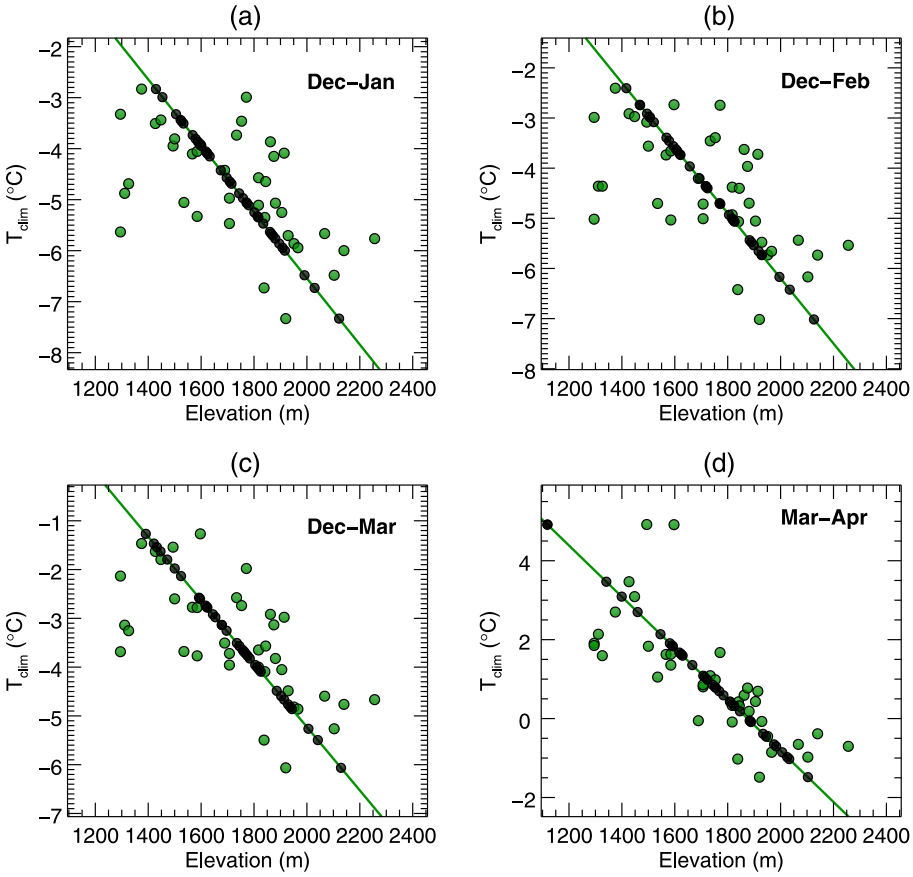


Figure 2. Climatological temperature (T_{clim}) versus sites' elevations (green dots); T_{clim} is the mean temperature over the aggregated months specified in the panels, averaged over 1996-2014; aggregated months correspond to the temperature predictor having the best linear relationship with T_{clim}; straight lines are centred at the sites' mean T_{clim} and elevation, with slope −6.5°C/km (a representative average atmospheric lapse rate); black dots indicate the effective elevations obtained by inverse mapping the sites' T_{clim} according to these lines.

on their climatological temperature (T_{clim}), in a way consistent with a representative average atmospheric lapse rate.

Figures 2a–2d show scatterplots of the site's elevations versus T_{clim}, computed as the mean temperature over the aggregated months specified in the figures and averaged for the climatological period 1996–2014. The aggregated months correspond to the temperature predictors of SWE_{Jan} (Figure 2a), SWE_{Feb} (Figure 2b), SWE_{Mar} (Figure 2c), and SWE_{Apr} (Figure 2d) having the best linear relationship with T_{clim}, as determined in the next section. For example, Figure 2a corresponds to T_{clim} averaged over December–January, since among the four temperature predictors of SWE_{Jan}, the correlation of $T_{Dec–Jan}$ versus SWE_{Jan} is found to have the best linear dependence with T_{clim} (i.e., the highest R^2, shown in the uppermost left panel of Figure 4).

For a given latitude and longitude, we expect temperatures to vary approximately linearly with altitude according to a representative average atmospheric lapse rate, indicated by the straight lines in Figures 2a–2d passing through the sites' mean T_{clim} and elevation (centers of mass of the green dots), and having slopes equal to −6.5°C/km. For a sufficiently small study area the site locations would be such that their climatological temperatures would lie relatively close to these lines, and we could determine the role of temperature on the snowpack solely as a function of elevation. For the larger area considered here, however, the sites' T_{clim} are scattered as a result of their spatial distribution. We thus adjust the sites' elevations by inverse mapping their climatological temperature according to these lines. Note that these elevation corrections have a robust physical basis for the temperature indices, but not necessarily for the precipitation indices, since precipitation is more a local phenomenon that is more strongly dependent on local topography than latitude or longitude. The extent to which the correlation of precipitation versus SWE is independent of latitude and longitude is

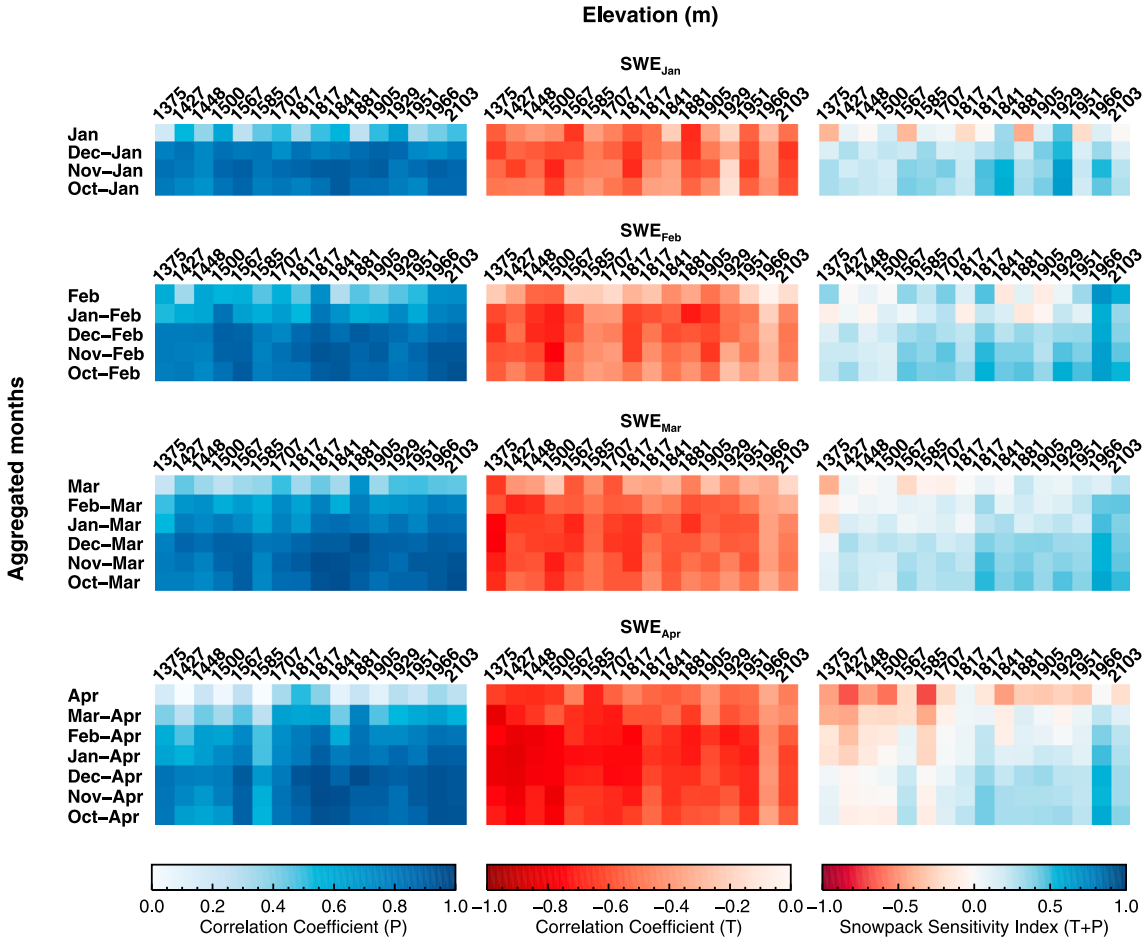


Figure 3. Pearson correlations of SWE indices versus (left column) precipitation and (middle column) temperature predictors having aggregated months as indicated; (right column) sum of left and middle values; sites' elevations are given without correction.

unclear, as this correlation may depend on temperature as well (simply put, the covariance of temperature and snowfall precipitation is not zero). However, the strong linear dependence of the correlation between snow and precipitation indices with elevation, shown in the next section, suggests that such corrections should be minimal.

3. Results and Discussion

Figure 3 shows correlations for snow versus meteorological indices at 16 selected SNOTEL sites ordered by increasing elevation. These representative stations have climatological temperatures lying within 120 m of the atmospheric lapse rate lines of Figure 2, and as a result they need minimal or no elevation correction. Figure 3 (left column) provides, for each of the snow indices SWE_{Jan}, SWE_{Feb}, SWE_{Mar}, and SWE_{Apr}, the correlation between that index and the precipitation (*P*) indices representing different numbers of aggregated months, as a function of elevation and aggregated months. Analogously, the middle column provides correlations between the snow indices and aggregated temperature (*T*) indices. Figure 3 (right column) shows the sum of Figures 3 (left column) and 3 (middle column), which lies between −1 and 1 in view of the opposite signs of the correlation coefficients for *P* and *T*. A near-zero value (showing as white) indicates that the snow index has a similar degree of correlation with the corresponding *T* and *P* predictors, whereas a negative (positive) value indicated as red (blue) indicates that the snow index has a stronger correlation with the temperature (precipitation) index. These matrices of correlation values show the effect of increasing the number of aggregation months on the response of the snowpack to temperature and precipitation, and its dependence on elevation. The precipitation-based correlations reveal that the snow indices tend to correlate better with the precipitation indices having more aggregated months, and at higher elevations. This is indicative of the

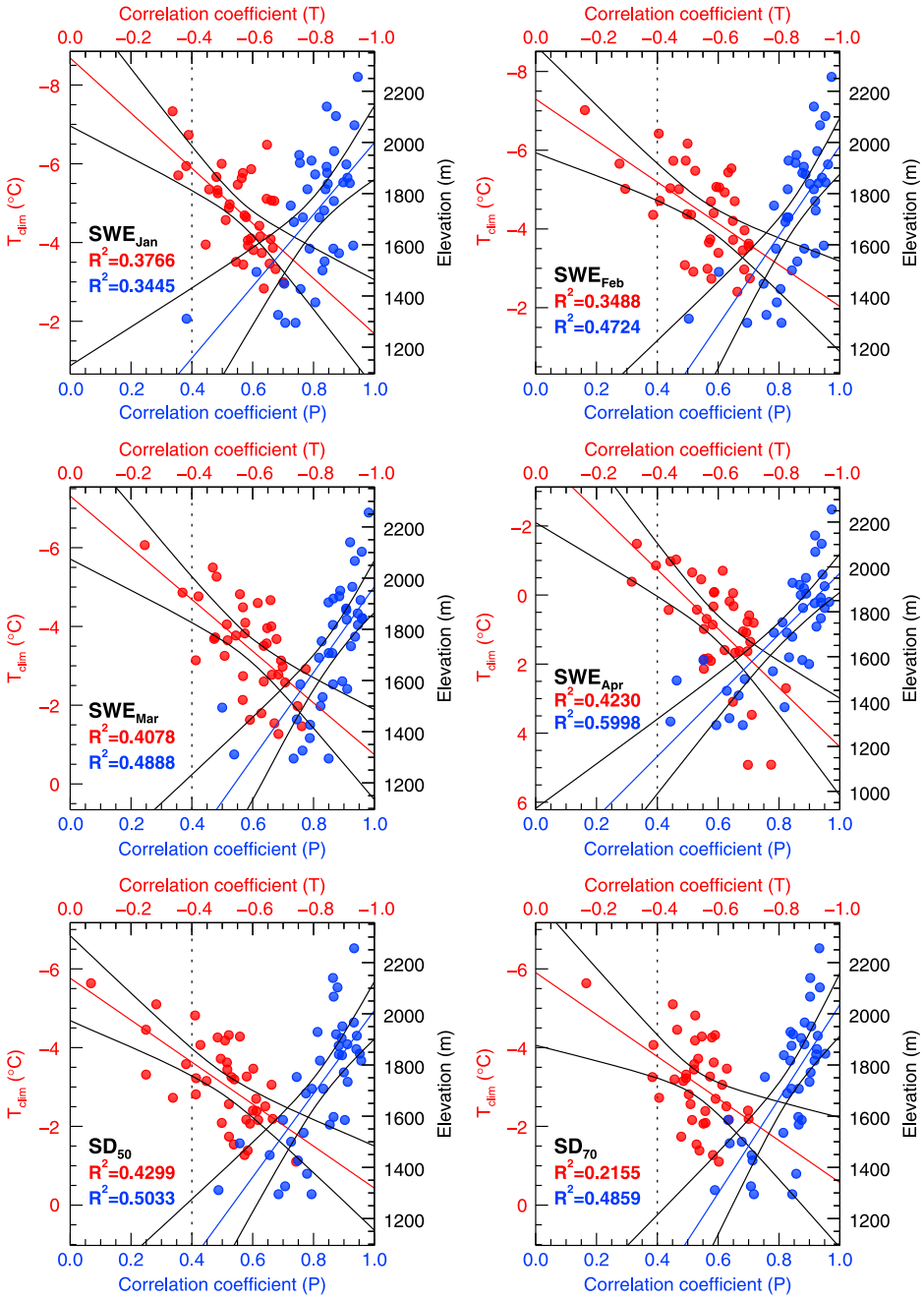


Figure 4. Pearson correlations between snow indices and (red) *T* or (blue) *P* predictors as a function of elevation. Dependence of the correlation coefficients between snow indices and *T* predictors on climatological temperature (T_{clim}) is also indicated (axes colored red). This dependence is valid only for correlations involving *T* predictors. For every snow index, we use the *P* predictor giving the best linear dependence with elevation and the *T* predictor giving the best linear dependence with T_{clim}. Elevations and T_{clim} are related by the linear mapping of Figure 2. Linear fits and 95% confidence intervals are indicated by the colored lines and black curves, respectively, and the vertical dotted line indicates the threshold above which correlation absolute values are statistically significant ($p < 0.05$).

snowpack's sensitivity to previous snowfall events, and the predominance of precipitation driving the snowpack at higher elevations. The dependence of the temperature index correlations on aggregation length and elevation is less distinct. However, correlations do tend to be stronger at lower elevations (particularly for March and April), where temperatures are warmer and thus have more potential to control the snowpack. The comparative effect of temperature and precipitation on snowpack variability shown in Figures 3 (right column) indicates that the influence of temperature tends to be strongest for short aggregations and

at low altitudes (red color), whereas that of precipitation tends to dominate for long aggregation and at high altitudes (blue color).

Figure 4 displays the sites' correlation coefficients of snow index versus both T and P predictors to infer a threshold elevation dividing where temperature or precipitation primarily drive the snowpack. For each snow index, we employ the P predictor that provides the best linear dependence between the sites' correlation coefficients and elevation, measured in terms of R^2. For SWE_{Jan}, SWE_{Feb}, SWE_{Mar}, and SWE_{Apr}, these predictors are found to be the ones with maximum number of aggregated months (i.e., from October to the month corresponding to the snow index), whereas $P_{Oct–Mar}$ and $P_{Oct–Apr}$ are the best P predictors of SD_{50} and SD_{70} respectively, indicating that for most sites snowmelt occurs well before June. The T predictors employed are those having the best linear dependence between the sites' correlation coefficients and climatological temperature T_{clim} (rather than elevation), as quantified by R^2. For each observation site, T_{clim} equals the mean temperature over the aggregated months corresponding to the temperature predictor and averaged over 1996–2014. We employ T_{clim} as a proxy for elevation because of the issues discussed in section 2.3. The correspondence between correlation coefficients and effective elevation is then inferred through the linear relationship between elevation and T_{clim}, assuming a lapse rate of −6.5°C/km as discussed in section 2.3. For SWE_{Jan}, SWE_{Feb}, SWE_{Mar}, and SWE_{Apr}, the aggregated months of the T predictors are those specified in Figures 2a–2d. For both SD_{50} and SD_{70}, the T predictor is $T_{Nov–Mar}$.

The least squares regression lines for the correlation coefficients versus elevation corresponding to T or P predictors are also shown in Figure 4, together with their 95% confidence bands. For each regression line $R^2 > 0.3$, except for SD_{70}, which has $R^2 = 0.22$ for the T predictor. In general, the correlations between snow indices and T predictors tend to decrease with elevation and are significant ($p < 0.05$) in most cases, corresponding via a t test to absolute values >0.4, indicated by the vertical line. Conversely, the correlations between snow indices and P predictors increase with elevation and are >0.4 in nearly all cases. The threshold effective elevation, above which precipitation is the main driver of the snow pack and below which temperature is the main driver, is determined by the intersection of the two regression lines.

The threshold elevations corresponding to SWE_{Jan}, SWE_{Feb}, SWE_{Mar}, and SWE_{Apr} are 1577 ± 87, 1558 ± 89, 1557 ± 84, and 1578 ± 76 m, respectively. Despite the altitude increase of the 0°C isotherm from January to April (Figure 4), these threshold elevations are relatively close (within the combined confidence intervals for all snow indices), showing little variation during the snow season. The threshold elevations corresponding to the snowpack duration indices are 1552 ± 81 (SD_{50}) and 1555 ± 114 (SD_{70}) m, which are approximately the same as those for the other snow indices. Table S2 shows that the threshold elevations and confidence intervals are robust with respect to the precise choice of atmospheric lapse rate, with relatively small changes, <25 m per 1°C/km difference in the lapse rate. In addition, the accuracy (R^2) of the linear dependence of the correlation coefficients for snow indices versus temperature predictors on elevation was found to be similar regardless the assumed lapse rate.

4. Conclusions

We have shown that in the elevation range investigated, the effect of precipitation on snowpack variability in the Central Rocky Mountains can be characterized by a linear dependence with elevation. Also, the potential of temperature to control the snowpack increases proportionally to the climatological seasonal mean temperature (T_{clim}). Thus, by assuming that T_{clim} decreases with altitude as per an average atmospheric lapse rate, we showed that the influence of temperature on snowpack variability decreases linearly with elevation. The opposite behaviors of temperature and precipitation in controlling snowpack variability enables us to obtain an effective threshold elevation (1560 ± 120 m depending on the various snow indices employed) below (above) which temperature (precipitation) is the main driver of the snowpack. This threshold elevation is robust under the range of average atmospheric lapse rates considered (−7.5 to −5.5°C/km).

In the context of a warming climate, our result has a direct implication for the hydrology and ecosystems of the studied region as well as the economic activities dependent on snow, since rising temperatures are likely to directly affect the amount of snow and runoff timing below the threshold elevation. Previous studies have emphasized the reduction of observed SWE over the western U.S. for elevations near the freezing level. *Mote* [2003] showed that April SWE in the Pacific Northwest has decreased significantly from 1950 to 2000 for elevations below ~1800 m as a result of temperature increase rather than decrease in precipitation. *Pierce et al.* [2008] showed that in the same period, negative trends of SWE/P have been observed for most of the western

U.S. at elevations below ~1900 m, thus confirming the dominant influence of temperature on the snowpack at those elevations. In addition, the approximately 0.8°C temperature increase observed in the U.S. Northwest during the past century [*Mote et al.*, 2014] has likely driven the shifts in the timing of the snowmelt runoff pulse in the Central Rocky Mountains, which has occurred 5 to 15 days earlier in recent decades at basins below ~2000 m [*Stewart et al.*, 2004; *Fritze et al.*, 2011] and are expected to occur at least 15 to 25 days earlier as temperature is projected to rise another 1.8°C to 5°C during this century [*Mote et al.*, 2014]. With this work, we have extended the methodology of *Morán-Tejeda et al.* [2013a] to estimate a threshold elevation in the Central Rocky Mountains below which rising temperatures are likely to alter the snowpack duration and hydrological regime of this large, thermally heterogeneous region. It remains to be seen whether this threshold elevation will increase as a result of a changing climate.

Acknowledgments

This research is partly supported by the Natural Sciences and Engineering Research Council of Canada's (NSERC) Climate Change and Atmospheric Research (CCAR) program through the Canadian Sea Ice and Snow Evolution (CanSISE) Network (www.cansise.ca). J.R.M. is supported by a NSERC Visiting Postdoctoral Fellowship. We acknowledge the Natural Resources Conservation Service Snow Telemetry (SNOTEL) network for making their snow cover and meteorological data openly accessible (data available online at http://www.wcc.nrcs.usda.gov/snow/). We are also grateful to Vivek Arora and Chris Derksen for helpful comments and for pointing out relevant bibliography.

The Editor thanks Ignacio Lopez Moreno and an anonymous reviewer for their assistance in evaluating this paper.

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From: Guinotte, John
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Date: Wednesday, January 31, 2018 4:47:14 PM
Attachments: Fig2-4_resolution_comparison.png

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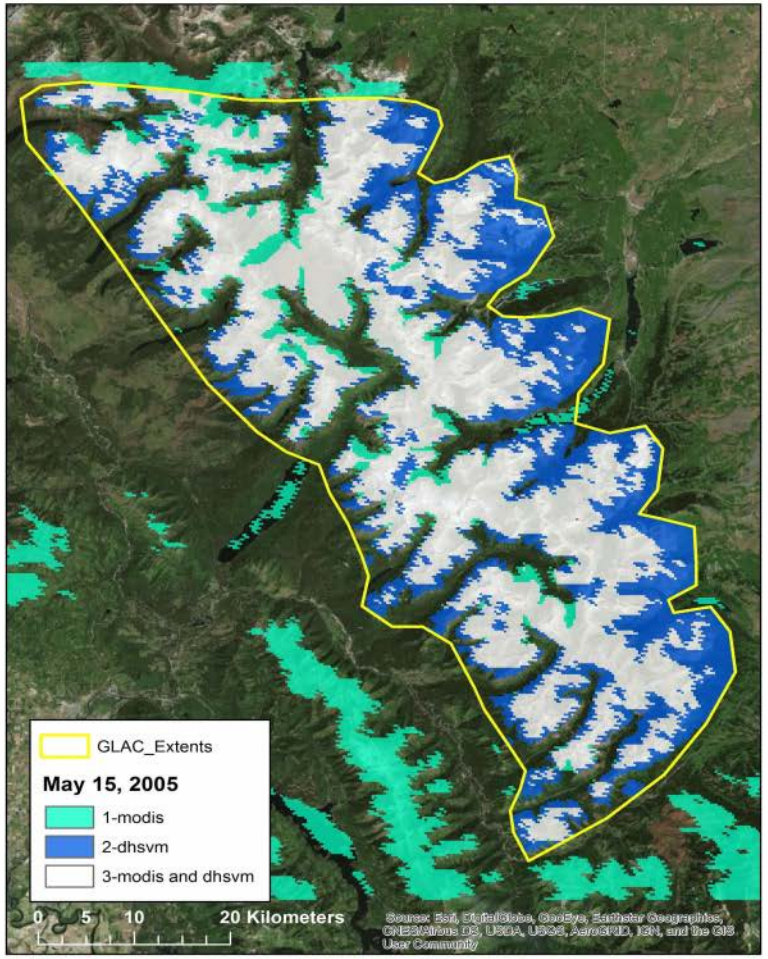
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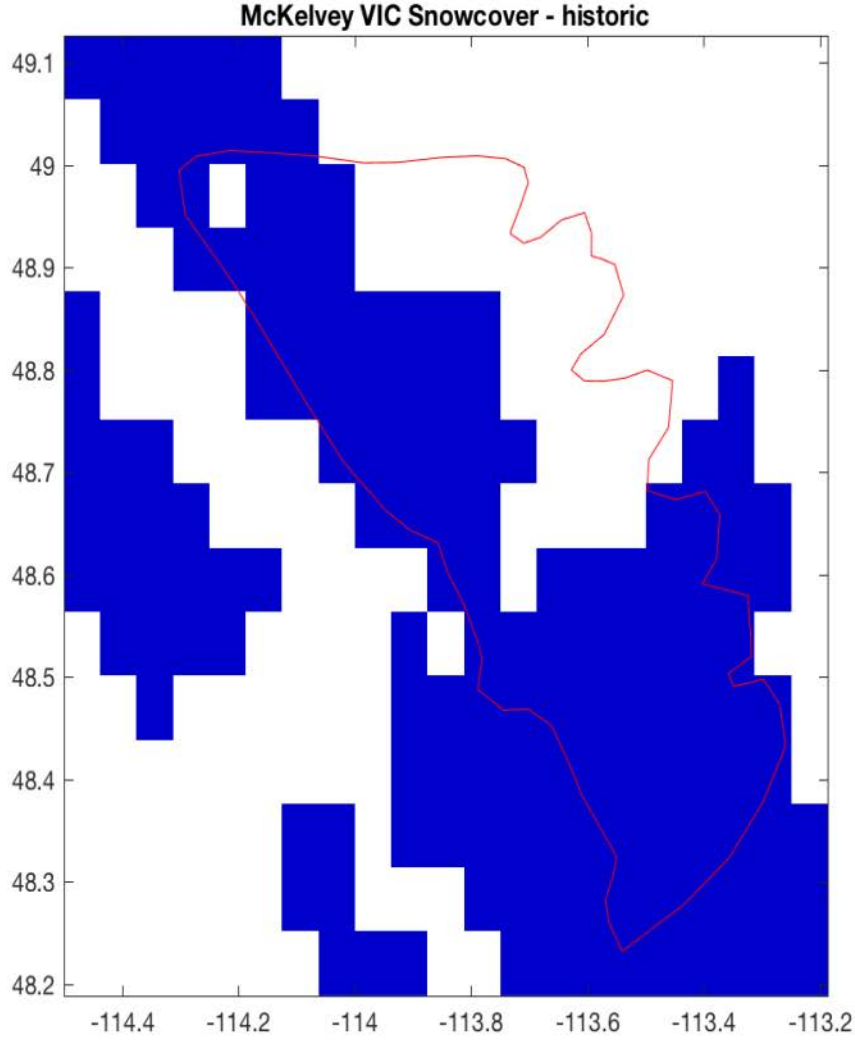
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-114
-113.8
-113.6
-113.4
-113.2

From: Nelson, Marjorie
To: Bush, Jodi
Cc: Justin Shoemaker/R6/FWS/DOI; Betty Grizzle
Subject: Re: PDD briefing for wolverine.
Date: Thursday, February 1, 2018 11:10:40 AM

Thanks Jodi, I'll try to add you and Betty. Duh - I was thinking lynx.
Hopefully you can lead the briefing, Betty!
Marj

Marjorie Nelson
Chief, Division of Ecological Services
Mountain-Prairie Region
U.S. Fish and Wildlife Service
303-236-4258 direct
720-582-3524 cell

On Thu, Feb 1, 2018 at 10:59 AM, Bush, Jodi <jodi_bush@fws.gov> wrote:

> Marj. I haven't seen the calendar invite but will add it if you can tell me time.
>
> Alas Jim wants no part of the wolverine show but I have cc'd Betty here. JB
>
> Jodi L. Bush
> Office Supervisor
> Montana State Ecological Services Office
> 585 Shepard Way, Suite 1
> Helena, MT 59601
> (406) 449-5225, ext.205
>
> On Thu, Feb 1, 2018 at 10:40 AM, Nelson, Marjorie <marjorie_nelson@fws.gov> wrote:
>
> > I see that got scheduled for Feb 20. I'll follow up with Stephanie on scheduling a dry run. I would think Noreen would want one for this species. Like stoneflies, field SME can present. If Noreen doesn't want a dry run, we should have one without her (ideally with Mike).
> >
> > thanks!
> > Marj
> >
> > Marjorie Nelson
> > Chief, Division of Ecological Services
> > Mountain-Prairie Region
> > U.S. Fish and Wildlife Service
> > 303-236-4258 direct
> > 720-582-3524 cell

From: Grizzle, Betty
To: Ed Turner
Subject: Map
Date: Thursday, February 1, 2018 11:05:09 AM
Attachments: Whitlock et al 2017_MT Climate Assessment_Executive Summary.pdf

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Here is the pdf. Map is on page 2. Thank you!

--

Betty J. Grizzle, D.Env., MPH
Fish and Wildlife Biologist
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Carlsbad Fish and Wildlife Office
2177 Salk Ave, Suite 250
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760-431-9440, ext. 215
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Marias River. Photograph courtesy of Scott Bischke.

EXECUTIVE SUMMARY

WHAT IS THE MONTANA CLIMATE ASSESSMENT?

The Montana Climate Assessment (MCA) is an effort to synthesize, evaluate, and share credible and relevant scientific information about climate change in Montana with the citizens of the State. The motivation for the MCA arose from citizens and organizations in Montana who have expressed interest in receiving timely and pertinent information about climate change, including information about historical variability, past trends, and projections of future impacts as they relate to topics of economic concern. This first assessment reports on climate trends and their consequences for three of Montana's vital sectors: water, forests, and agriculture. We consider the MCA to be a sustained effort. We plan to regularly incorporate new scientific information, cover other topics important to the people of Montana, and address the needs of the state.

The process to develop the first MCA was driven by stakeholder input and informed by the best-available science. Insights regarding topics to cover were developed from questionnaires, conversations, and listening sessions held across the state. A team of researchers, educators, and stakeholders used that feedback to select the topics covered.

The Montana Institute on Ecosystems, a statewide center based at both Montana State University and University of Montana, has taken on the responsibility of organizing the MCA. The 2017 MCA is the result of two years of work by university faculty and students, state and federal agency researchers, non-profit organizations, resource managers, and citizens from across Montana.

The assessment begins with an analysis of Montana's recent climate trends and how climate is projected to change in the future (Chapter 2). This information is used throughout the assessment to explain the key impacts of climate change observed in recent decades and projected in the future. Discussion of climate change impacts on Montana's water (Chapter 3), forests (Chapter 4), and agriculture (Chapter 5) are presented next. The assessment concludes with an analysis of major knowledge gaps—and thus areas for future research—related to climate change and its impacts on the three sectors covered herein (Chapter 6).

MONTANA'S CLIMATE

Understanding current climate change and projecting future climate trends is of vital importance, both for our economy and our well-being. The Climate chapter serves as a foundation for the MCA, providing information on present-day climate as well as climate terminology, past climate trends, and future climate projections. The chapter is an introduction to climate science and the important processes that determine whether climate remains constant or changes.

Climate basics

Climate is driven largely by energy from the sun, and the manner in which this incoming solar radiation is reflected, absorbed, transformed (as in photosynthesis), or re-radiated (as heat). Each of these processes influences climate through changes to temperature, the hydrologic cycle, vegetation, and atmospheric and ocean circulation patterns. Climate change, as defined by the US Global Change Research Program (USGCRP undated), includes:

> *Changes in average weather conditions that persist over multiple decades or longer. Climate change encompasses both increases and decreases in temperature, as well as shifts in precipitation, changing risk of certain types of severe weather events, and changes to other features of the climate system.*

Such changes are driven in large part by the greenhouse effect, the trapping of greenhouse gases in Earth's atmosphere and consequent warming of the planet. The rapid rate of climate change since the Industrial Revolution has resulted from changes in atmospheric chemistry, specifically increases in greenhouse gases due to increased combustion of fossil fuels, land-use change (e.g., deforestation), and fertilizer production (Forster et al. 2007). The primary greenhouse gases in the Earth's atmosphere are carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), water vapor (H_2O), and ozone (O_3).

Montana's unique features

To understand climate change in Montana, we must first understand Montana's unique geography. Montana is the fourth largest state in the nation and its location within North America exposes the state to a mix of diverse weather systems that originate from the Pacific Ocean, the Arctic, and sometimes subtropical regions. The Continental Divide, which has a predominantly north-south alignment in Montana, effectively splits the state into climatically distinct western wet and eastern dry regions with respect to moisture from eastward-flowing Pacific Maritime air. The state also includes the beginnings of three major river basins—the Missouri, Snake/Columbia, and Saskatchewan—two of which encompass almost one-third of the landmass of the conterminous United States. Consequently, Montana's climate influences the water supply of a large portion of the country, and its water supports communities, ecosystems, and economies far beyond its borders.

Our analysis

Montana's unique geography means that climate varies across the state, as it does across the nation. Throughout the MCA, we aggregate past climate trends and future climate projections into seven Montana climate divisions, as shown in Figure I. These seven climate divisions are a subset of the 344 divisions defined by the National Oceanic and Atmospheric Administration (NOAA) based on a combination of climatic, political, agricultural, and watershed boundaries (NOAA undated).

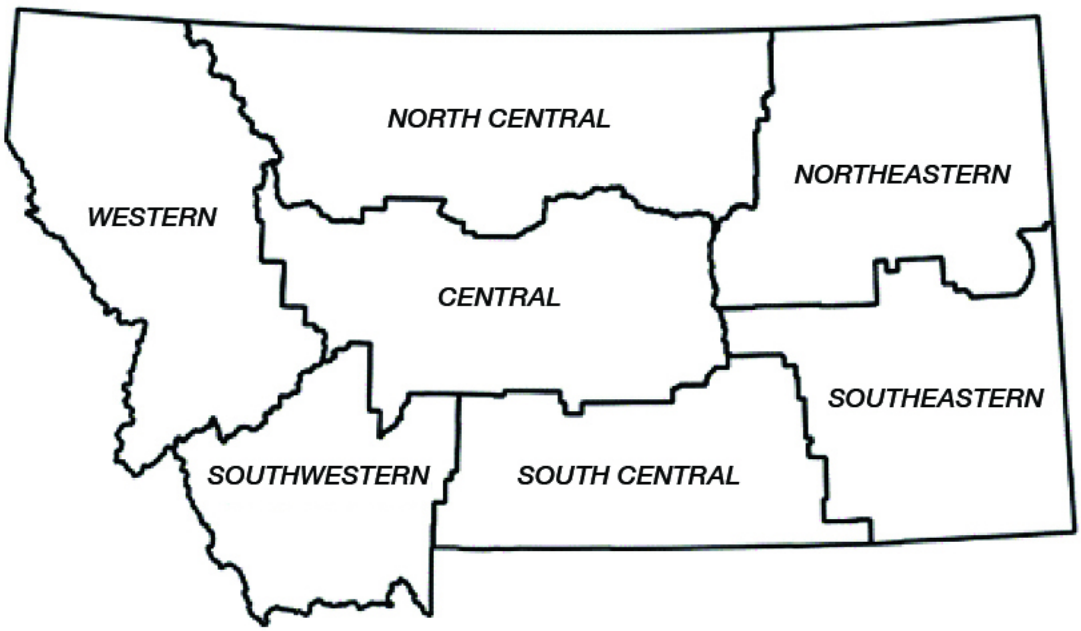


Figure I. Montana's seven climate divisions.

To assess Montana's historical climate, we evaluated temperature and precipitation trends since the mid-20th century by using standard statistical methods to analyze records of temperature and precipitation. To assess future projected changes to Montana's climate, we employed an ensemble of climate models from the fifth iteration of the Coupled Model Intercomparison Project (CMIP5), and utilized a statistically downscaled dataset.

Major findings

The results of this analysis produced several key messages, some of which are shown below, about Montana's historical and future climate (for a complete list of key messages, see the Climate chapter):

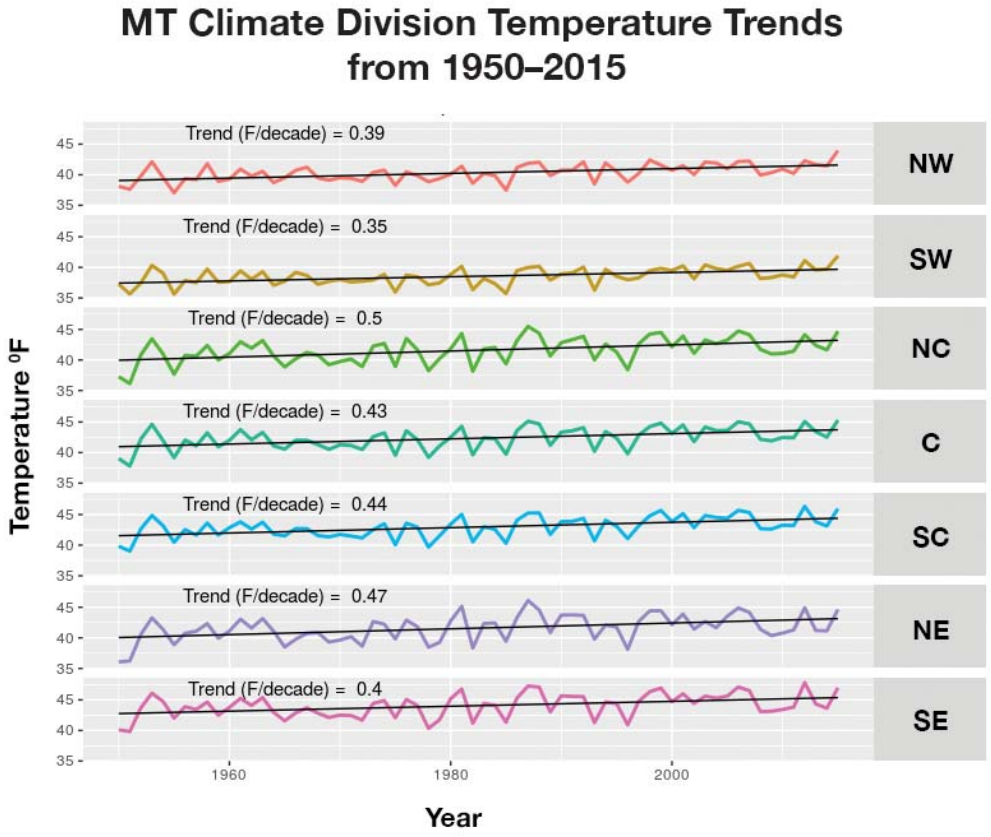


Figure II. Trends in annual average temperature across each climate division (Figure I) in Montana. The divisions are northwestern (NW), southwestern (SW), north central (NC), central (C), south central (SC), northeastern (NE), and southeastern (SE).

Mid-century and End-of-century

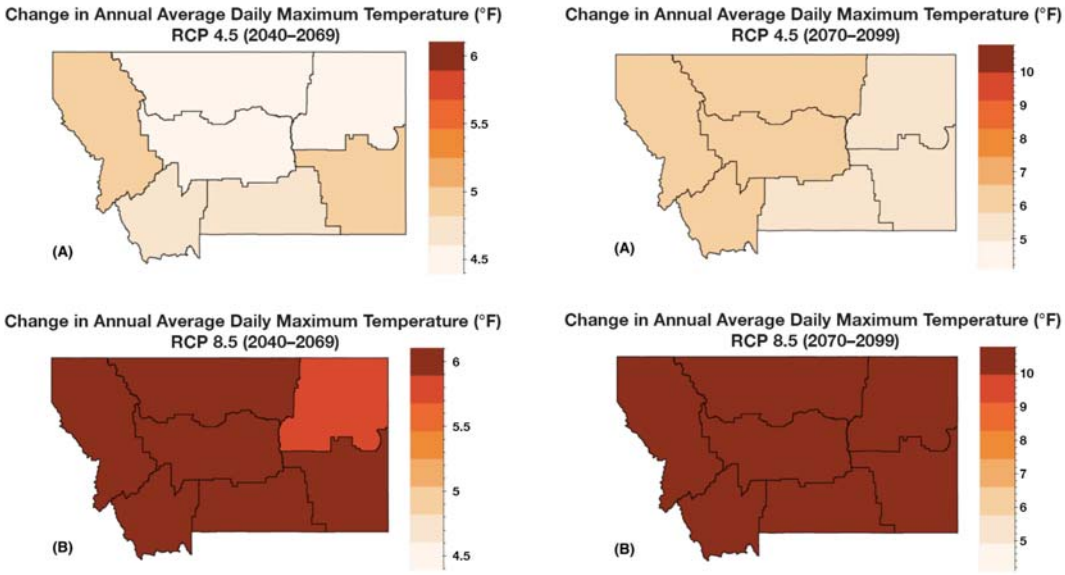


Figure III. The projected increase in annual average daily maximum temperature (°F) for each climate division in Montana for the periods 2049-2069 and 2070-2099 for (A) stabilization (RCP4.5) and (B) business-as-usual (RCP8.5) emission scenarios.

- Annual average temperatures, including daily minimums, maximums, and averages, have risen across the state between 1950 and 2015. The increases range between 2.0-3.0°F (1.1-1.7°C) during this period (see Figure II). *[high agreement, robust evidence]*[1]
- Despite no historical changes in average annual precipitation between 1950 and 2015, there have been changes in average seasonal precipitation over the same period. Average winter precipitation decreased by 0.9 inches (2.3 cm), which can largely be attributed to natural variability and an increase in El Niño events, especially in the western and central parts of the state. A significant increase in spring precipitation (1.3-2.0 inches [3.3-5.1 cm]) also occurred during this period for the eastern part of the state. *[moderate agreement, robust evidence]*
- Montana is projected to continue to warm in all geographic locations, seasons, and under all emission scenarios throughout the 21st century. By mid century, Montana temperatures are projected to increase by approximately 4.5-6.0°F (2.5-3.3°C) depending on the emission scenario. By the end-of-century, Montana temperatures are projected to increase 5.6-9.8°F (3.1-5.4°C) depending on the emission scenario. These state-level changes are larger than the average changes projected globally and nationally (Figure III). *[high agreement, robust evidence]*

- Across the state, precipitation is projected to increase in winter, spring, and fall; precipitation is projected to decrease in summer. The largest increases are expected to occur during spring in the southern part of the state. The largest decreases are expected to occur during summer in the central and southern parts of the state. *[moderate agreement, moderate evidence]*

Table I provides a summary of climate metrics developed under the MCA.

Table I. Summary of climate metrics.

Climate Metric—	Trend and future scenario
Atmospheric CO2 concentrations	Global atmospheric carbon dioxide concentrations have increased over 100 ppm since Montana statehood and are projected to increase under both future scenarios considered here.
Average temperature	Since 1950, average statewide temperatures have increased by 0.5°F/decade (0.3°C/decade), with greatest warming in spring; projected to increase by 3-7°F (1.7-3.9°C) by mid century, with greatest warming in summer and winter and in the southeast.
Maximum temperatures	Maximum temperatures have increased most in spring and are projected to increase 3-8°F (1.7-4.4°C) by mid century, with greatest increases in August and in the southeast.
Days above 90°F (32°C)	Extreme heat days are projected to increase by 5-35 additional days by mid century, with greatest increases in the northeast and south.
Minimum temperatures	Minimum temperatures have increased most in winter and spring and are projected to increase 3-7°F (1.7-3.9°C) by mid century, with greatest increases in January and in the southeast.
Frost-free days	Frost-free days are projected to increase by 24-44 days by mid century, particularly in the west.
Average precipitation	Statewide precipitation has decreased in winter (0.14 inches/decade [-0.36 cm/decade]) since 1950, but no significant change has occurred in annual mean precipitation, probably because of very slight increases in spring and fall precipitation. Precipitation is projected to increase, primarily in spring (0.2-0.7 inches [0.5-1.8 cm]) in the northwest; a slight statewide decrease in summer precipitation and increased year-to-year variability of precipitation are projected, as well.
Number of consecutive dry days	Little projected change, with a maximum increase of 3 days to -3 days under the most severe scenario by end of the century. However, increased variability in precipitation suggests potential for more severe droughts, particularly in connection with climate oscillations.
Number of consecutive wet days	No substantial change projected.

IMPACTS TO MONTANA'S WATER

Water in Montana

Montana depends on an adequate supply of clean water for nearly every aspect of our economy, including food production, hydroelectric power, domestic and industrial uses, and sustaining our natural ecosystems. The vast majority of water that enters Montana comes as rain or snow at higher elevations (MT DNRC 2014a, b, c, d; MT DNRC 2015). Although some of Montana's water originates in Wyoming or adjacent Canadian provinces, over 80% is derived from within state boundaries, hence Montana's designation as a "headwaters state." The major rivers of Montana export more than 40 million acre-feet of water/yr (4.9x1010 m^3/yr)[2] —more than twice the capacity of Flathead Lake—with the majority, approximately 60%, generated in the Clark Fork and Kootenai river basins west of the Continental Divide.

Groundwater is another large and important component of the water cycle in Montana, with most groundwater coming from shallow sand or gravel aquifers in river floodplains. Groundwater resources are critical for water users, but also contribute significantly to natural streamflow throughout the year. In Montana, much of the winter snowfall that accumulates in the mountains melts in spring to produce streamflow and recharge groundwater aquifers. Projected changes in temperature will have large effects on how water enters Montana (e.g., as rain or snow), how it is distributed among major storage pools, and how it moves or changes from one component of the water cycle to another.

Our analysis

To best represent the influence of climate variations on water resources, the Water chapter focuses on eight rivers and their watersheds (Figure IV; note that some watersheds—for example, that of Poplar River—extend beyond the state boundaries). These focal rivers and watersheds, chosen across the state's seven NOAA climate divisions,[3] include:

- Climate division 1
  —Clark Fork River at Saint Regis
  —Middle Fork of the Flathead River at West Glacier
- Climate division 2—Missouri River at Toston
- Climate division 3—Marias River near Shelby
- Climate division 4—Musselshell River at Mosby
- Climate division 5—Yellowstone River at Billings
- Climate division 6—Poplar River near Poplar
- Climate division 7—Powder River near Locate

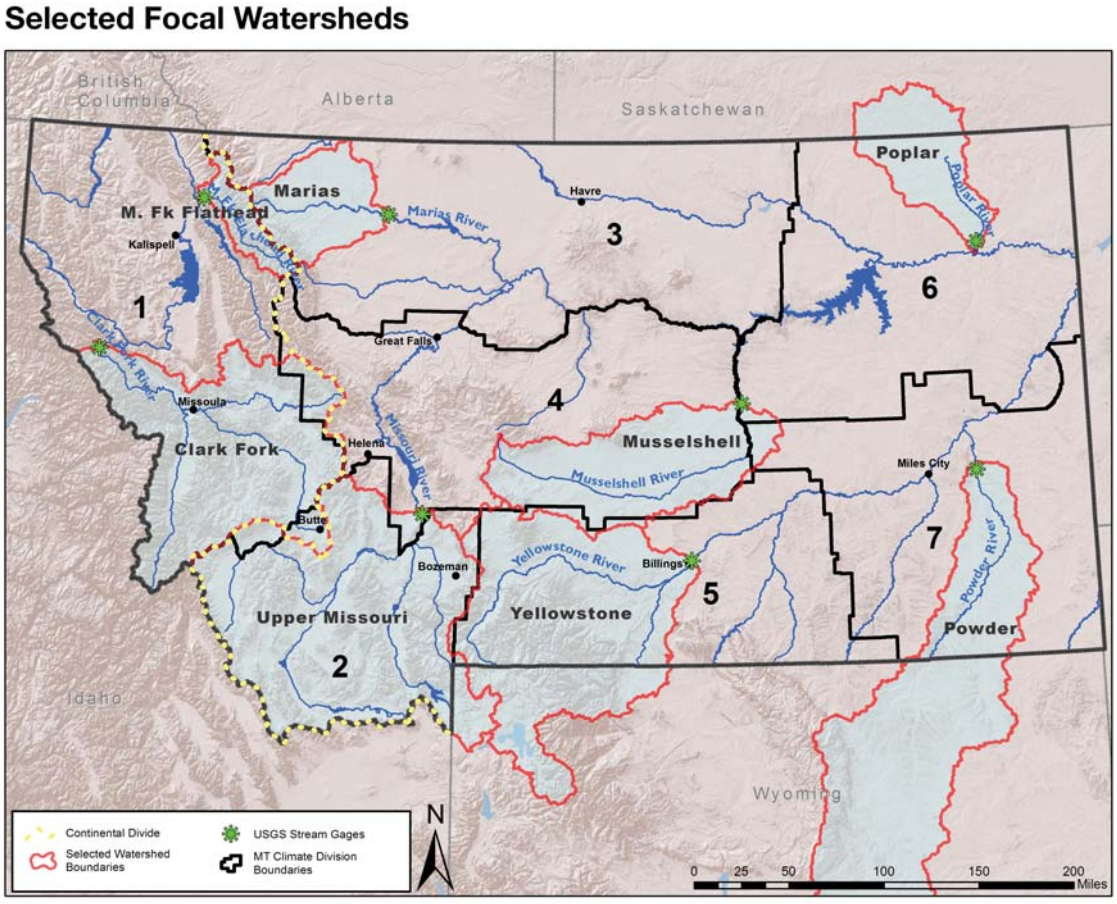


Figure IV. The focal rivers for this assessment, including black outlines of the seven climate divisions (see Water chapter), contributing watersheds (red), river gage locations (green), and the Continental Divide (dotted).

Complex computer models (see Climate chapter) provide a method for projecting future climate scenarios in Montana. By linking climate models to water-cycle models, we generate projections about how climate change is likely to influence water resources. For the projections in the Water chapter, we present results from as many as 31 climate models that are linked to a water-cycle model. We utilize these projections to discuss how climate change may affect key components of the water cycle, including:

- Snowpack
- Snowmelt runoff and timing
- Annual streamflow
- Groundwater resources
- Drought

Major findings

The results of this analysis produced several key messages, some of which are shown below, about how climate change will affect Montana's water resources (for a complete list of key messages, see the Water chapter).

Rising temperatures will reduce snowpack, shift historical patterns of streamflow in Montana, and likely result in additional stress on Montana's water supply, particularly during summer and early fall. Key messages associated with these findings follow:

- Montana's snowpack has declined over the observational record (i.e., since the 1930s) in mountains west and east of the Continental Divide; this decline has been most pronounced since the 1980s. *[high agreement, medium evidence]*
- Warming temperatures over the next century, especially during spring, are likely to reduce snowpack at mid and low elevations. *[high agreement, robust evidence]*
- Historical observations show a shift toward earlier snowmelt and an earlier peak in spring runoff in the Mountain West (including Montana). Projections suggest that these patterns are very likely to continue into the future as temperatures increase. *[high agreement, robust evidence]*
- Earlier onset of snowmelt and spring runoff will reduce late-summer water availability in snowmelt-dominated watersheds. *[high agreement, robust evidence]*
- Groundwater demand will likely increase as elevated temperatures and changing seasonal availability of traditional surface-water sources (e.g., dry stock water ponds or inability of canal systems to deliver water in a timely manner) force water users to seek alternatives. *[high agreement, medium evidence]*

Rising temperatures will exacerbate persistent drought periods that are a natural part of Montana's climate. Key messages associated with these findings follow:

- Multi-year and decadal-scale droughts have been, and will continue to be, a natural feature of Montana's climate *[high agreement, robust evidence]*; rising temperatures will likely exacerbate drought when and where it occurs. *[high agreement, medium evidence]*
- Changes in snowpack and runoff timing will likely increase the frequency and duration of drought during late summer and early fall. *[high agreement, medium evidence]*

IMPACTS TO MONTANA'S FORESTS

Forests in Montana

In the Forest chapter, we interpret how past and projected changes in climate—as described in the Climate chapter—may influence Montana forests. There are approximately 23 million acres (9.3 million ha) of forested land in Montana, and most are publicly owned, in the western part of the state and dominated by Douglas-fir, lodgepole pine, and ponderosa pine (Figure V). Forest conditions in Montana are varied, and potential impacts from climate change will overlay on existing stresses to forests. Ultimately, forest managers will need to consider specific adaptation actions in response to current and potential climate changes.

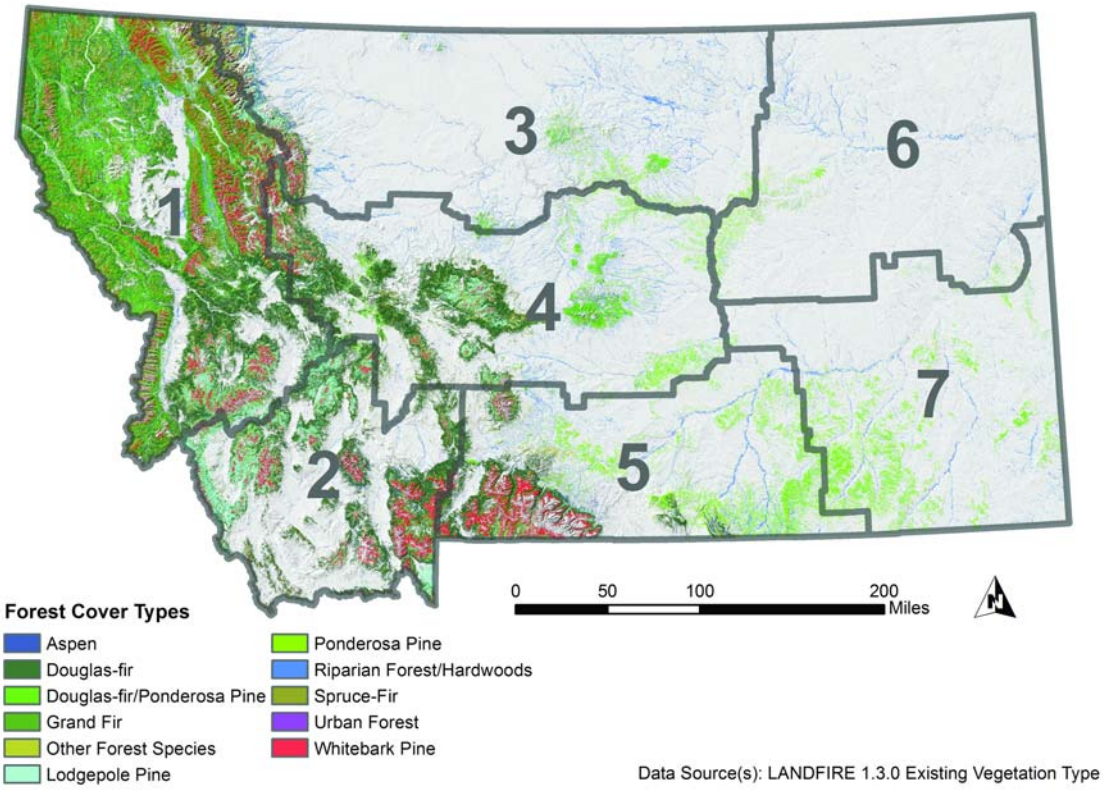


Figure V. Existing forest cover type in Montana (Landfire 2012). Gray boundaries delineate climate divisions (see Figure I).

Our analysis

In the face of changing climate, forest managers can best maintain forest health and stable product yield by understanding past trends and planning for a range of climate scenarios. The analysis in the Forest chapter is based on the climate trends for which we had sufficient data and climate projections that represent plausible future scenarios, as described in the Climate chapter. It is important to note that current forest conditions will largely determine the potential impacts from current and future climate change. Forest conditions vary across land ownership types, and many Montana forests are under stress due to past forest management practices. In addition, we do not detail potential responses of individual tree species to climate shifts in this assessment; instead, we focus on the direct and indirect effects of climate change on forests. We point the reader to Chapter 6 in the Northern Region Assessment Program report (Keane et al. forthcoming) for species-level information.

Major findings

The direct effects of climate change on forests include increased temperatures and shifts in precipitation that together alter humidity, soil moisture, and water stress. Direct effects can be beneficial or detrimental to forest growth and survival. The results of this analysis on the direct effects of climate change on Montana's forests produced several key messages, some of which are shown below (for a complete list of key messages, see the Forests chapter):

- Increased temperatures will have positive or negative effects on individual trees and forest-wide processes, depending on local site and stand conditions, but impacts from increased extreme heat will be negative. *[high agreement, moderate evidence]*
- Direct effects of climate change on individual trees will be driven by temperature in energy-limited forests and moisture in water-limited forests. *[high agreement, moderate evidence]*
- The speed and magnitude of climate change may mean that increased forest mortality and contractions in forest distribution will outpace any gains in forest growth and productivity over the long run, leading to a net loss of forested area in Montana. *[medium agreement, limited evidence]*

Table II. Summary of potential climate-related direct effects to forests.

Direct effect	Possible impacts	Projected net effect
Establishment and regeneration	**Positive:** Higher CO_2 concentrations and temperatures may lead to increased tree fecundity **Negative:** Higher temperatures and reduced water availability could reduce seedling survival	Possible positive or negative effects are superimposed on climate oscillations, such as the Pacific Decadal Oscillation, which can produce decades of cooler and wetter conditions that may be more favorable for establishment and regeneration
Growth and productivity	**Positive:** Increased vegetation water use and increased growth and productivity as a result of longer growing season **Negative:** Reduced growth and productivity in water limited areas	Possible increased growth and productivity concurrent with climate oscillations that increase water availability, particularly at higher elevations and where stand density is low; extreme high temperatures would have net negative impact, regardless of water availability

Mortality	**Positive:** Few opportunities for reduced direct climate effects on mortality but possibility for reduced mortality from indirect effects **Negative:** Increased acute and background mortality from increased temperatures and indirectly from increased disturbance	Increased mortality, although may be driven by indirect effects; patterns of mortality will be dependent on initial stand and local site conditions, but more arid regions more susceptible
Range shifts and forest distribution	**Positive:** Potential range expansion with warmer temperatures and sufficient moisture **Negative:** Potential range contraction where temperature is too high or in water-limited locations	Possible faster range contraction compared to expansion, with net range reduction particularly in drier areas; no clear direction of elevational shifts; responses will be highly species and location dependent

Indirect effects of climate change on forests include disturbance—a key component of forest ecology—and may be more important, immediate, and longer lasting than direct effects. As with direct effects, indirect effects can compound existing forest conditions and impacts from past and future human land-use activities (Moritz and Agudo 2013).

The results of this analysis on the indirect effects of climate change on Montana's forests produced several key messages, some of which are shown below (for a complete list of key messages, see the Forests chapter):

An increase in fire risk (i.e., probability of occurrence)—including an increase in size and possible frequency and/or severity (i.e., tree mortality)—is expected in the coming century as a result of a) prolonged fire seasons due to increased temperatures, and b) increased fuel loads from past fire suppression. *[high agreement, robust evidence]*

Rising temperatures are likely to increase bark beetle survival [high agreement, strong evidence], but climate-induced changes to other insects and forest pathogens are more varied and less certain *[medium agreement, moderate evidence]*

There may be a reduction in the amount of carbon stored in forests. *[low agreement, limited evidence]*

Table III provides a summary of potential climate-related indirect effects to forests.

Table III. Summary of potential climate-related indirect effects to forests.

Indirect effect	Possible impacts	Projected net effect
Disturbance: fire	Positive: Increased forest heterogeneity (long-term, post-burn) Negative: Decreased forest diversity and heterogeneity (immediately post-burn); increased social and economic impacts from fire; increased release of forest carbon	Increased fire severity resulting primarily from warmer weather and past fire suppression; increased release of forest carbon from fire
Disturbance: pathogens	Positive: Some pathogen species may decline and result in decreased forest mortality Negative: Some pathogens species may increase and result in increased forest mortality and increased susceptibility to beetle attack	Uncertain climate effects on pathogens, dependent on moisture regimes, pathogen species, and host species
Disturbance: insects	Negative: Increased forest mortality; reduced forest diversity with shift towards non-host tree species	Increased temperatures likely to result in increased insect disturbance, but dependent on elevation, insect species and host availability
Soil responses and carbon storage	Positive: Increased organic matter if increased productivity; increased nitrogen availability Negative: Decreased organic matter (with increased decomposition rates); decreased mycorrhizal support; increased soil acidity; increased release, or decreased removal, of atmospheric CO2	Uncertain climate effects on soil responses, but projected reductions in soil and forest carbon storage

IMPACTS TO MONTANA'S AGRICULTURE

Agriculture in Montana

Agriculture is a key industry in Montana, generating over $5.2 billion in 2014 through the sale of agricultural commodities (USDA-NASS 2015). Montana's large agricultural industry consists of both crops and livestock. Montana's farm and ranchland support a mosaic of dryland and irrigated agriculture, commodity and specialty cropland, and native and planted rangeland. Although more Montanans live in cities than on farms and ranches, we think of Montana as an agricultural state, where the non-forested landscape is dominated by livestock and crop production.

Our analysis

Montana agriculture has always faced volatility, extreme events, and variability across the state and these conditions will continue to be the case with projected climate changes in Montana. Climate model projections show a warmer Montana in the future, with mixed changes in precipitation, more extreme events, and mixed certainty about upcoming drought. The Agriculture chapter examines potential impacts of projected climate change on commodity crops, livestock, pollinators, disease, pests, and weeds. However, any effort at assessing climate impacts on agriculture faces multiple levels of uncertainty, including uncertainty that a) accompanies all climate projections, b) is specific to agricultural projections, and c) is created by adaptive actions (human interventions) that can mask a direct climate signal. Climate impacts on agriculture in other regions of the world can also create uncertainty and have a major impact on Montana agriculture by changing commodity prices and input costs. Increasing uncertainty due to complex interactions, whether through volatility or new and hard-to-predict temperature and moisture trends, can disrupt agricultural decision-making and will probably become an even more important direct agriculture decision-driver in the years ahead (See Figure VI).

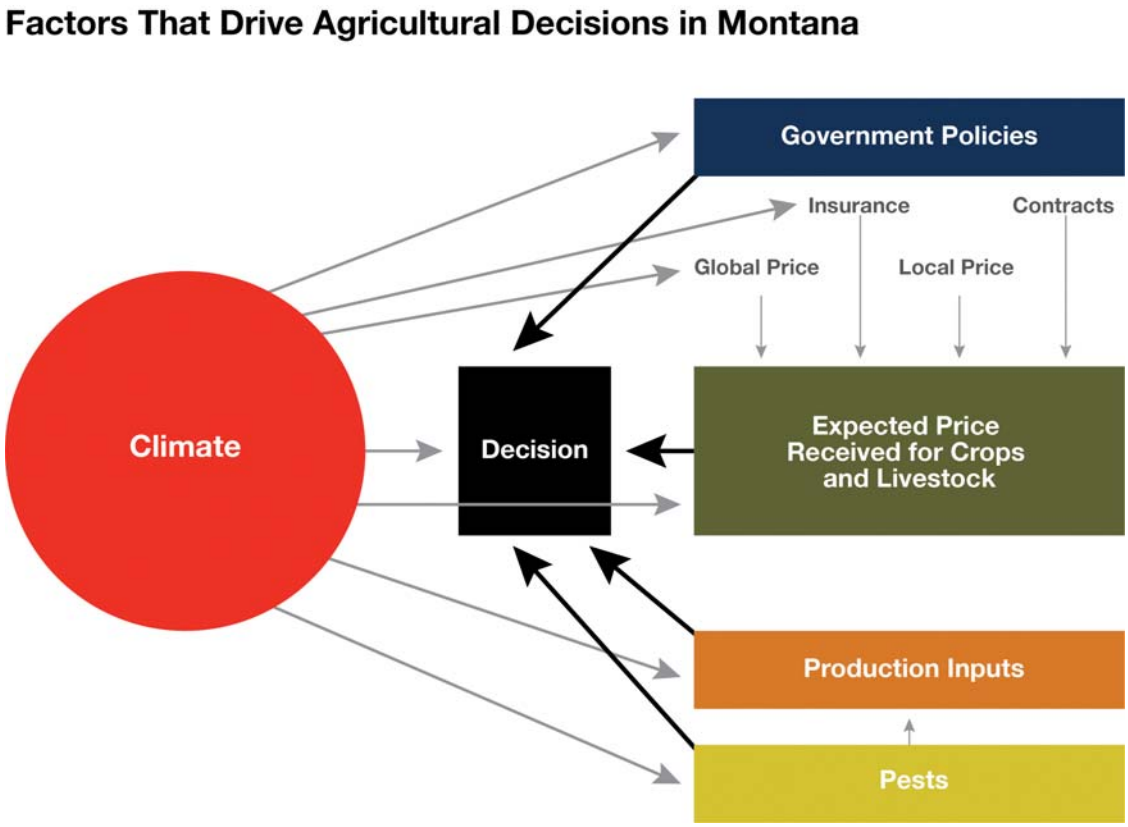


Figure VI. Factors that drive agricultural decisions in Montana. The size of bubble and arrows qualitatively represents the relative importance of each factor's influence on agricultural production decisions.

Major findings

The results of this analysis produced several key messages, some of which are shown below, about how climate change will affect Montana agriculture (for a complete list of key messages, see the Agriculture chapter):

- Every component of agriculture—from prices to plant pollinators and crop pests—exhibits complex relationships to climate, depending on the location, weather variability, and agricultural and economic practices and policies (Figure VI). Social and economic resilience to withstand and adapt to variable conditions has always been a hallmark of Montana farmers' and livestock producers' strategies for coping with climate variability. *[high agreement, robust evidence]*
- Decreasing mountain snowpack will continue to lead to decreased streamflow and less reliable irrigation capacity during the late growing season. Reduced irrigation capacity will have the greatest impact on hay, sugar beet, malt barley, market garden, and potato production across the state. *[high agreement, robust evidence]*
- Increases in temperature will allow winter annual weeds, such as cheatgrass, to increase in distribution and frequency in winter wheat cropland and rangeland. Their spread will result in decreased crop yields and forage productivity as well as increased rangeland wildfire frequency. *[high agreement, medium evidence]*
- Climate change affects global-price-determined commodity agriculture differently than it affects non-commodity agriculture. Commodity crops, such as small grains, are more directly driven by global markets and agricultural subsidies, whereas non-commodity crops tend to be more directly tied to local or specialized non-local markets and local micro-climates. *[high agreement, medium evidence]*
- Diversified cropping systems, including rotation with pulse crops and innovations in tillage and cover-cropping, along with other measures to improve soil health, will continue to allow adaptation to climate change. *[medium agreement, low evidence]*

CONCLUSIONS

The 2017 Montana Climate Assessment focused on three sectors that Montana stakeholders identified as important to their lives: water, forests, and agriculture. The MCA found that all three of these sectors have experienced impacts from climate change over the last half century. In addition to exploring how the past climate has changed and its effects on Montana, the MCA explored how future projected climate change would also affect water, forests, and agriculture across the state. The overall objective of the MCA is to inform Montanans about the state's changing climate so that they can better plan for the future.

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NEXT CHAPTER (/chapter/introduction)

(/)

UNDERSTAND
Climate Assesment
(/Understand)

EXPLORE
Multimedia, Education
(/Explore)

FOLLOW
News, Social Media, Events
(/Follow)

CONNECT
Partners, Stakeholders
(/Connect)

From: Shoemaker, Justin
To: Sarah Fierce; Marjorie Nelson; Kurt Broderdorp
Subject: Wolverine schedule
Date: Friday, February 2, 2018 7:52:57 AM

Marj,

We have two entries for wolverine on our R6 workload chart. Now that we have direction from the recommendation meeting, I say we could delete the earlier one and keep the one w/ the 9/30/18 due date as the Final withdrawal. But the date in pack track is 7/15/18. If your talking w/ HQ today about workload, we should sort the date we should be aiming for.

Justin Shoemaker
Classification and Recovery Biologist
U.S. Fish and Wildlife Service, Region 6
Phone: 309-757-5800 x214
Email: justin_shoemaker@fws.gov

From: Stephen Torbit
To: Betty Grizzle; John Guinotte
Subject: Document1
Date: Friday, February 2, 2018 1:04:37 PM
Attachments: Document1.docx

My write-up on fractional snow-cover by cardinal direction produced from Ray et. al. I retained the figure numbers used by Ray, because I did not know how Betty wanted to handle this info. Do we want it in the body of the SSA or in an appendix.

Let me know if this helps.

Steve

To illustrate the importance of aspect on snow retention, Ray et. al. (2017) presented MODIS derived summary statistics for snow-cover for March 1 to July 1 for all years 2000-2017. Figure (4.5 in Ray) presents a graphic of the number of years with snow-cover by aspect for Glacier National Park and Fig. (4.12 in Ray) presents the same snow-cover information for Rocky Mountain National Park. Ray et. al. (2017) calculated the fraction of total land area within each octant that was snow-covered for each of the 17 years in the historical record. The asymmetry in Fig 4.5 shows that for Glacier, the northeast directions retain a much larger fractional snow covered area than other aspects. Even in dry years, more than 60 percent of the NE facing slopes are snow-covered on May 15th. Similar results are revealed in Figure 4.12 for Rocky Mountain, except that northwest facing slopes retain a larger amount of fractional snow-cover than other aspects. These results clearly illustrate the differential retention of snow by aspect; therefore, topography must be evaluated for meaningful biological interpretation of future snow conditions.

From: A J
To: Grizzle, Betty; Inman, Bob
Subject: good summary article
Date: Sunday, February 4, 2018 1:29:38 PM
Attachments: jarnevich et al. 2015.pdf

Good summary paper of caveats to consider for modeling approaches

Ecological Informatics 29 (2015) 6–15

Contents lists available at ScienceDirect

Ecological Informatics

journal homepage: www.elsevier.com/locate/ecolinf

Caveats for correlative species distribution modeling

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ARTICLE INFO

Article history:
Received 5 March 2015
Received in revised form 18 June 2015
Accepted 21 June 2015
Available online 24 June 2015

Keywords:
Correlative models
Species distribution modeling
Niche modeling
Presence–absence methods
Mapping
Species-environmental matching models

ABSTRACT

Correlative species distribution models are becoming commonplace in the scientific literature and public outreach products, displaying locations, abundance, or suitable environmental conditions for harmful invasive species, threatened and endangered species, or species of special concern. Accurate species distribution models are useful for efficient and adaptive management and conservation, research, and ecological forecasting. Yet, these models are often presented without fully examining or explaining the caveats for their proper use and interpretation and are often implemented without understanding the limitations and assumptions of the model being used. We describe common pitfalls, assumptions, and caveats of correlative species distribution models to help novice users and end users better interpret these models. Four primary caveats corresponding to different phases of the modeling process, each with supporting documentation and examples, include: (1) all sampling data are incomplete and potentially biased; (2) predictor variables must capture distribution constraints; (3) no single model works best for all species, in all areas, at all spatial scales, and over time; and (4) the results of species distribution models should be treated like a hypothesis to be tested and validated with additional sampling and modeling in an iterative process.

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1. Introduction

Examples of species distribution models abound in the current literature. They are being used to explore potential impacts of climate change (Kearney et al., 2010), establish conservation priority areas and select

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http://dx.doi.org/10.1016/j.ecoinf.2015.06.007
1574-9541/Published by Elsevier B.V.

reserve designs (Fuller et al., 2008; Kremen et al., 2008; Pawar et al., 2007), guide restoration efforts (Fei et al., 2012), identify new populations of rare and endangered species (de Siqueira et al., 2009; Guisan et al., 2006), predict distributions of known and unknown species (e.g., Raxworthy et al. visited high diversity locations from combined distribution models and found previously undescribed species; Pearson et al., 2007; Raxworthy et al., 2003), aid invasive plant management (Crall et al., 2013), as a component of risk analysis (Ficetola et al., 2007), and in phylogeographic studies (Miller and Knouft, 2006; Moritz et al., 2009). To focus the scope of this paper, our discussion is limited to correlative species distribution models, rather than process-based models. Correlative models develop mathematical relationships between observed locations of species occurrence (presence or absence) and environmental variables. Correlative species–environment relationships often are applied to *develop*, as opposed to *test*, hypotheses regarding drivers of species distributions or are applied to unsampled locations to predict occurrence of the species or its suitable environmental conditions. These same models are frequently used with climate change scenarios in ecological forecasting. Our review was precipitated by the fact that several important caveats are overlooked or unstated by many investigators. Or, even when they are explicitly stated, may not be clearly respected in applications. Thus, the purpose of this paper is to provide a summary document that could be referred to by the modeling community and resource managers and other users of model output.

Easy access to species distribution modeling software and greater computing power has increased their use. Many techniques can be implemented in freely available programs (e.g., see examples in Elith et al. (2006); EcoMod by Guo and Liu (2010); SAHM by Morisette et al. (2013); Maxent by Phillips et al. (2006); and BIOMOD by Thuiller et al. (2009)), R packages (e.g., biomod2 and dismo) and R code provided in many papers (e.g., Broennimann et al., 2012). Some implementations are very user friendly, making it easy to obtain output. This does not, however, mean that models are being implemented with an understanding of the assumptions and limitations of a given statistical technique or of species distribution modeling in general (see Joppa et al., 2013). There are important considerations when following the model development steps of preparing data, selecting and implementing a model, applying a model to a novel location (e.g., native range to invaded range) or time (e.g., climate change), and evaluating and interpreting results from a model. In recent years the literature has been flooded with many studies examining several aspects related to modeling in great depth. Given the plethora of papers and the great need for and interest in applications of these techniques, a review of caveats and modeling considerations is warranted.

There are several review papers relating to some caveats of species distribution modeling that provide a good foundation for improved species distribution modeling practice and should be read. Our review builds on these older papers that do not cover topics debated in more recent literature (e.g., Guisan and Zimmermann, 2000), cover more specific topics (e.g., Heikkinen et al. (2006) and Thuiller et al. (2007) only examine climate change applications), or deal with the more theoretical rather than applied aspects of model development (e.g., Araujo and Guisan, 2006; Austin, 2007). We build on recent reviews examining appropriate use of bioclimatic models (e.g., Araújo and Peterson, 2012) by briefly reviewing caveats and considerations for cautious implementation of correlative species distribution modeling techniques.

We focus on techniques that characterize habitat suitability. The techniques for habitat suitability identification use presence-only or presence/absence data, whereas abundance models require a continuous value such as count data for animals or foliar cover for plant species. Though we do mention some aspects of species abundance, we primarily focus on species occurrences (i.e., presence locations). The locations of occurrence or abundance are related to environmental characteristics, and the patterns found between occurrence/abundance and the environment aid in predicting suitability/ abundance at other unsampled locations in a confined area (i.e., the extent). Such efforts could be fraught with uncertainty. Understanding four primary caveats to the habitat suitability modeling process might help communicate the utility of species distribution models, despite areas of high uncertainty. Fig. 1 gives an outline of the species distribution modeling workflow. The caveats offered in this paper will follow the same outline: (1) all sampling data are incomplete and potentially biased; (2) predictor variables must capture distribution constraints; (3) no single model works best for all species in all areas at all spatial scales and over time; and (4) the results of species distribution models should be treated like a hypothesis to be tested and validated with additional sampling and modeling in an iterative process.

2. Primary caveat #1: all sampling data are incomplete and potentially biased

Biological field sampling typically represents a biased snapshot of species distribution and abundance (Stohlgren, 2007). In general, field sampling is restricted in time and space to a small subset of all potential species occurrences, abundances, persistence, fecundity, and dispersal potential. A sampling design that poorly captures the entire range of the species or population of concern (e.g., Fig. 2a) may lead to sample biases which in turn will affect model results. Locations should not only capture the geographical range of a species but also the environmental gradients over which a species is found. A given "presence location" may occur in moderately suitable habitat, or may represent temporary use of unsuitable habitat, yet species distribution models generally treat all locations as being equally suitable with the assumption that locations represent viable populations. An occurrence "point" may represent a microsite and different environmental conditions than indicated by the environmental resolution of the model. For example, a plant species may be restricted to tiny springs or seeps (mesic microsites) within a large grid cell characterized as a xeric environment. Thus, to create an accurate and useful model, the resolution of the model needs to match the spatial context of the field sample and, relatedly, the biological phenomenon being modeled rather than the resolution of readily available environmental predictors.

2.1. Location data errors

True "presence data" must consist of observations that are concurrent with the environmental predictors and must be taxonomically accurate. The issue of data quality was highlighted by Lozier et al. (2009), when they generated a fictitious distribution of Sasquatch, when the data represented misidentifications of the black bear, *Ursus americanus*. While this is a humorous example, it does highlight the importance of using high quality data in model generation. Dormann et al. (2008) identified uncertainty in species breeding status at locations as a primary source of uncertainty in model results (27.4% of identified uncertainty) when compared to uncertainty from model type, variable selection within model method, and collinearity treatment. Other errors relate to the quality of georeferenced information associated with the data. Some modeling methods may be less sensitive than others to moderate location errors, and useful predictions may still be generated when errors are present (Graham et al., 2008). If a model is well specified and only a small proportion of samples have errors (e.g., misidentification or inaccurate coordinates), it should not greatly affect model results (see Rodda et al., 2011). Spatial resolution of the environmental predictors can be adjusted to accommodate locational uncertainties in species occurrences. For example, if occurrence data have locational uncertainties of 1–2 km distance then using 4–5 km resolution data layers may result in more accurate model predictions (e.g., Kumar et al., 2014). Location of the samples is also important as incomplete sampling across important environmental gradients affects interpolation and extrapolation.

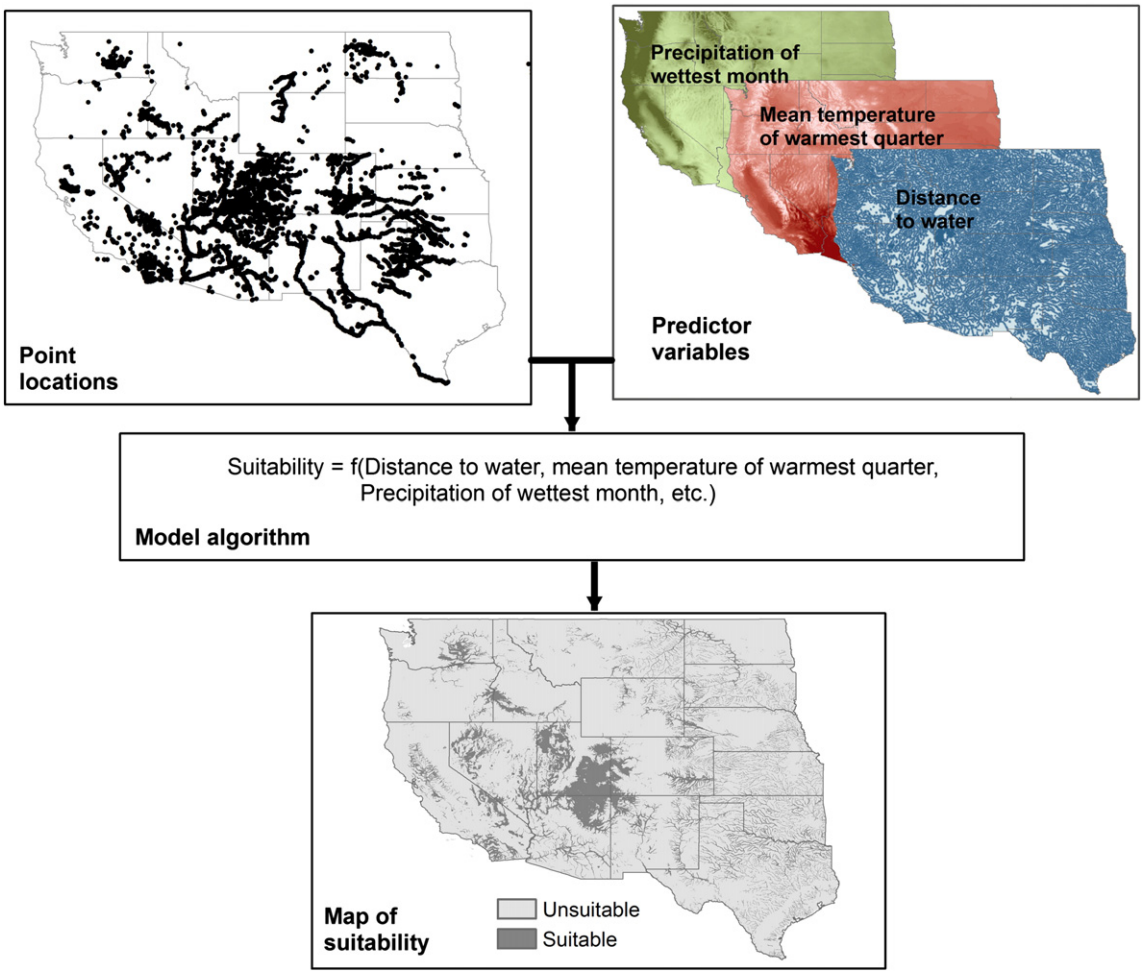


Fig. 1. Phases of the modeling process. These phases correspond to the primary caveats and include (1) sampling data; (2) predictor variables; (3) model algorithm; and (4) the results of species distribution models.

True "absence" may be impossible to model over large areas through time. The observed "absence" of species during field sampling cannot differentiate between undetected presence, suitable environmental conditions not yet established by a species, and unsuitable environmental conditions (e.g., Fig. 2b). For example, seeds may be buried in soil, going unreported. Small organisms, or certain life stages of organisms, may be cryptic, or microscopic (e.g., insect eggs buried in the soil or insect larvae developing inside a tree or fruit before emerging as adults). Imperfect detection of a species because of the type of survey method used, species characteristics and abundance across the landscape (cryptic species have low detectability), and heterogeneity of environmental conditions (harder to sample all environmental conditions), can also result in false absence (Boulinier et al., 1998; MacKenzie, 2005). True absence data at appropriate spatial scales also may not be stable over time, as climate and other conditions change, particularly for invasive species (likely to be expanding) or threatened and endangered species (concerned to be shrinking; Table 1). Still, where absence data are available they should be used as they generally produce better models (Brotons et al., 2004), and research using artificial species indicates that using randomly generated pseudo- absence locations may be better than no absence data (Wisz and Guisan, 2009). The best we can do with species distribution modeling tools is to provide a snapshot of a species distribution or abundance with estimated levels of uncertainty over a defined period of time.

It is important to remember that all species distribution models are affected in some way by the quality, completeness, and potential biases of the data (Stohlgren, 2007). The best models cannot make up for poor quality data. Ideal location data would consist of presence locations covering suitable environmental conditions accessible to a species and absence data covering unsuitable environmental conditions accessible to a species. In any study the description of the field sampling method used should be explicit, and should include information regarding the spatial consistency between observations and the model grid resolution, temporal consistency between samples and predictor layers, and how absence observations or presence-only considerations are handled.

2.2. Sampling prevalence and sample size

The influence of prevalence, defined as the ratio of presence to absence locations, in species presence/absence datasets has received a lot of attention. A 50% sampling prevalence is often preferred (Fig. 2c), but anything ranging from below 20% to above 75% prevalence of presence locations may provide decent results (McPherson et al., 2004). It is important to note that McPherson et al. (2004) only tested linear regression and non-linear discriminant analysis, and other techniques may perform differently. In contrast, Jimenez-Valverde and Lobo (2006) suggest that prevalence is not a factor if: (1) models are developed with good predictors (representative), and accurate training data (few false absences and adequate sample size); and (2) the data are not spatially or environmentally biased (e.g., Fig. 2a where one branch of a riparian area was missed or Fig. 2c where the north part of the area of interest was not surveyed). Franklin et al. (2009) manipulated prevalence for each of a suite of species ranging from common to rare across the landscape. They determined that prevalence did not have

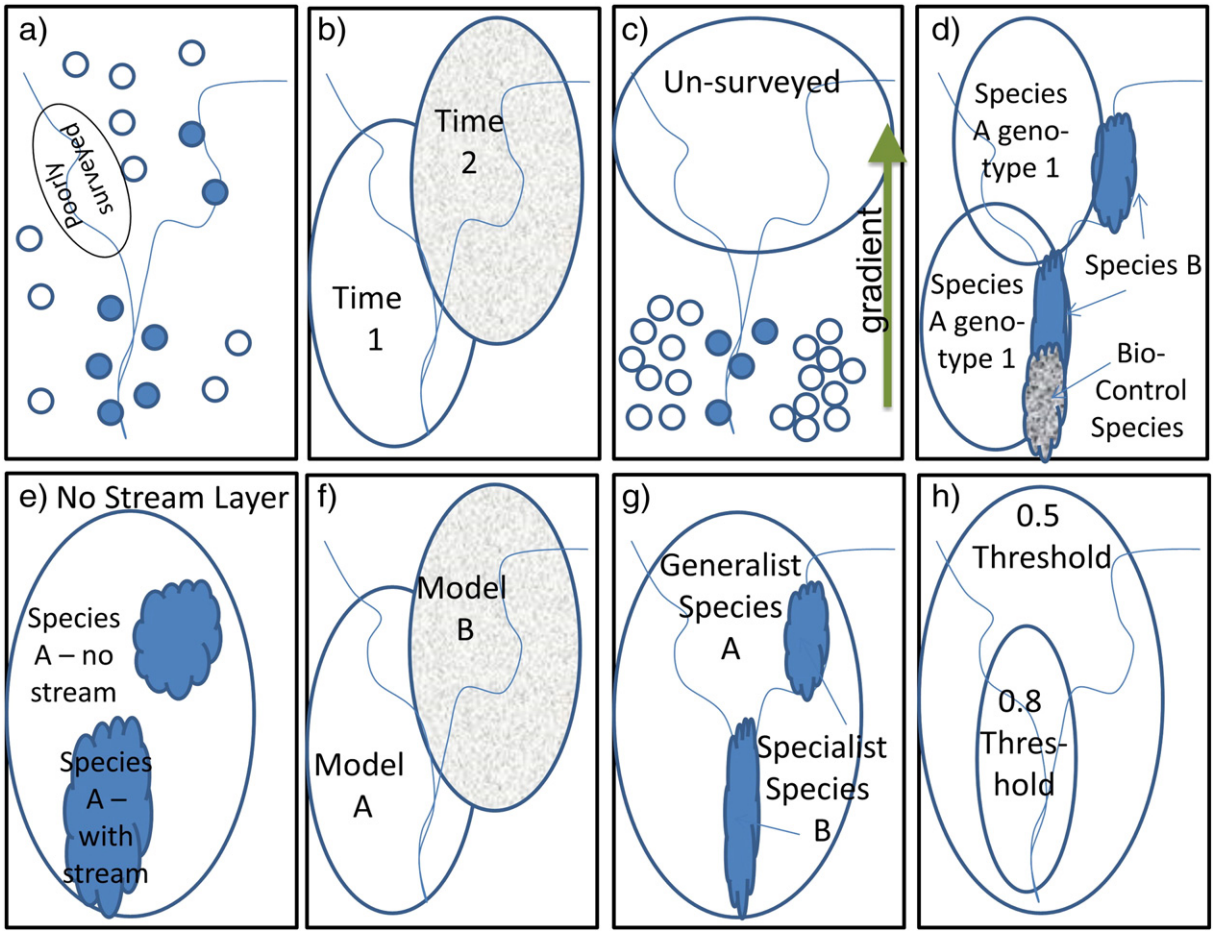


Fig. 2. Issues from species distribution modeling. Illustrated examples of different issues that may arise when developing species distribution models, including (a) spatial sampling bias — riparian zones may be important and here an entire branch of a river was missed, although there were several absence locations (open circles (absence) versus closed circles (presence)) in the vicinity; (b) temporal sampling, which could bias predictions if the species had an unstable distribution — if this is where an invader newly arrived at Time 1, there may be several untrue absence locations in the area overlapped by the future Time 2; (c) skewed prevalence and distribution of samples — ideally samples should cover the entire area being predicted, unlike this picture where there are many more absence locations and the northern part of the environmental gradient is missed; (d) non-environmental factors affecting a species' distribution, such as biotic interactions (here, biological control for a species) or genetic diversity; (e) missing important predictors, such as a stream layer here for a riparian species, may decrease model performance (species A with the stream layer results in a more restricted distribution than species A without the stream layer); (f) models for the same species using the same predictors but trained on different geographic areas will produce different results; (g) generalist species (distribution encompassed by circle) may be harder to predict than specialist species (limited to the riparian zone here); (h) the impact of threshold selection to define locations as suitable or unsuitable will affect the extent predicted as suitable.

Table 1
Major underlying assumptions of presence-only models, presence–absence models, and abundance models.

Model type	Assumptions or idiosyncrasies
Presence-background (e.g., Maxent and GARP)	1. Background locations adequately sample areas accessible to the species (see Fig. 2a) 2. Samples cover important environmental gradients 3. Produces a model of relative habitat suitability (e.g., one location is more suitable than another) but values do not reflect occupancy
Presence–absence models (e.g., logistic regression)	1. Absence recorded at a specific location is assumed to denote absence at the resolution of the cell in the model 2. Absence locations are assumed to be unsuitable environmental conditions 3. Samples cover important environmental gradients (see Fig. 2c) 4. Produces a model of occupancy
Abundance models (e.g., multiple linear regression)	1. Abundance measurements at a point are assumed to denote abundance at the resolution of the cell in the model. 2. Abundance is not affected by time since colonization (see Fig. 2b). 3. Samples cover important environmental gradients (see Fig. 2c).

an effect and rather that more accurate models can be developed for habitat-specialist species than habitat-generalist species (also see Evangelista et al., 2008; Fig. 2g); this may, however, be an artifact of the evaluation metric (Lobo et al., 2008; see species characteristics matter section below). The ideal conditions proposed by Jimenez-Valverde and Lobo (2006) are difficult to meet, so checking prevalence in your data may still be a good idea.

The minimum sample size required to create a useful model is another issue. In one study, no algorithm from those tested predicted consistently well at small sample sizes ($n < 30$; Wisz et al., 2008). However, these small samples may still provide useful information to guide further survey efforts and enable more efficient location of unknown populations. For example, using jackknife techniques with Maxent with as few as five presence locations, Pearson et al. (2007) successfully identified unknown populations, highlighting locations with similar environmental conditions rather than range boundaries. The sample size also affects the number of predictors that should be included in a model, with a recommendation of at least 10 data points per variable (Hosmer and Lemeshow, 2000). Categorical variables imply further consideration for sample size, in that there should be sufficient points for each class in a categorical predictor variable (e.g., at least 10 data points per class).

2.3. Spatial extent and background selection

Ideally, the spatial extent of the modeling exercise (and thus the location data) should cover the area that has been accessible to the species of interest over the relevant time period, and background or absence data should be limited to this same extent (Barve et al., 2011). However, if presence data do not cover the entire range accessible to a species, geographic extent of these data should still define the extent for model development. If background or absence data are drawn from areas outside the presence location sampling range, results will be altered and assessment metrics inflated. VanDerWal et al. (2009) varied the spatial extent from which they drew background points (10 distance bands around presence points; 10 km to 500 km) and found background points randomly extracted from within a distance band of 200 km from presence locations produced the best Area Under the Curve results (AUC; discussed in the 'Evaluation and calibration' section in more detail) for 12 species in northwestern Australia. Smaller areas developed misleading models while larger areas inflated test statistics, over-predicted distributions, and altered response variable selection. The number of background points also affects model predictions and a large number of points, usually 10,000 or more, are recommended to represent background environment (Barbet-Massin et al., 2012; Northrup et al., 2013). However, this number would vary based on the extent of the study area and heterogeneity of the environment as the background data need to adequately capture environmental gradients available to the species.

Projecting predictions to large extents from a model created for a small area may also lead to problems. If capturing potential distribution is desired, location data to develop the model will ideally be drawn from the broadest geographic and environmental range possible. These modeling techniques utilize correlative relationships between a species and the environment. Species relationships to environmental parameters and relationships among environmental parameters may not be constant in space, and factors important in controlling a species distribution in one region may be different from those in a different region (Fig. 2f). For example, Duncan et al. (2009) examined the transferability of climate envelope models built for dung beetles native to Africa and introduced to Australia. For three of five species, models created using data from the native range (Africa) did not accurately predict distribution in the introduced range (Australia). This may potentially be due to species climatic niche shift during the invasion process (Broennimann et al., 2007), or because factors other than climate may be controlling the distribution.

3. Primary caveat #2: predictor variables must capture distribution constraints

Predictor variables considered in the modeling should be biologically meaningful, selected based on species' eco-physiological tolerances or habitat requirements (Austin, 2007; Guisan and Zimmermann, 2000). The choice of the set of independent or predictor variables can affect the model predictions and performance (Johnson and Gillingham, 2005). For example, one study used three different sets of climate predictors to model the potential distribution of a vulnerable butterfly species. Depending on the set of predictors used, the species was either projected to experience little alteration in suitable environmental conditions, or to have a great range expansion with climate change (Harris et al., 2013).

Given these potential effects, the combination of variables retained during screening processes, such as removal of correlated variables, is very important (Braunisch et al., 2013). The removal of highly correlated variables is important as they should be "independent" (among themselves)—not highly collinear (correlation coefficient $|r| > 0.70$, assessed using Pearson, Spearman Rank, or Kendall correlation metrics; Dormann et al., 2013). Collinearity is especially problematic when there is interest in applying a model to a novel geographic location or time, in particular if the correlation structure among variables is not constant. This constitutes extrapolation rather than interpolation. High collinearity among variables can also make the interpretation of their relative importance very difficult.

Variables should also not be greatly affected by spatial autocorrelation. Spatial autocorrelation in response and predictor variables might result in a lack of independence among model residuals, thus violating the spatial independence assumption of some techniques (e.g., ordinary least squares regression). It can also result in parameter coefficient shift, underestimation of standard errors, and overestimation of statistical significance (Hawkins, 2012; Mauricio Bini et al., 2009). A lack of independence among residuals could also be due to a missing important environmental predictor (e.g., Fig. 2e) or clustered or geographically and environmentally biased occurrence data. Therefore, spatial patterns in model residuals should be examined for spatial autocorrelation (Legendre and Legendre, 1998) using metrics such as Moran's *I* coefficient of spatial autocorrelation, variograms and spatial correlograms (see Dormann et al., 2007 for details). Spatial autocorrelation in residuals can be addressed by either including spatial structure in the model error term (e.g., Lichstein et al., 2002) or using additional predictor variable(s) that are ecologically meaningful and also match the spatial pattern of the residuals. The spatial leave-one-out procedure suggested by Le Rest et al. (2014), which is a special case of spatial cross-validation where each point left out is spatially independent from the others, can also be used to account for residual spatial autocorrelation. Higher spatial autocorrelation in species occurrence data can result in highly inflated test evaluation statistics and erroneous model projections (Boria et al., 2014). A spatial filtering approach for dealing with spatial autocorrelation in occurrence data could be used (Boria et al., 2014) and a spatial cross-validation could be used to reduce inflation in test evaluation statistics.

Variables that have a direct influence on a species' distribution are more transferable than indirect predictors (Guisan and Zimmermann, 2000). If you want to project a model to a new area or to a future time, such as using an invasive species' native range to predict an invaded range, use direct predictors (Fig. 2f). Something like elevation, which is a surrogate for other predictors like temperature, may not mean the same thing in different locations. An elevation of 2000 m at 20° latitude will probably be warmer on average than an elevation of 2000 m at 40° latitude.

Models generally only include abiotic factors, and these are often limited to climatic information. Many factors other than abiotic conditions can affect species distributions including genetic diversity, dispersal ability (including anthropogenic spread pathways which are often ignored), the presence of competing or predatory species, and natural succession, adaptations, and evolution (Fig. 2d). These factors could be included in a model. For example, models could be created for different genetic sub-populations (e.g., Jarnevich et al. (2014) created models for Africanized honey bees which are a hybrid of several subspecies of *Apis mellifera*). For an invasive species one model could be created for dispersal, where predictors would include anthropogenic factors associated with spread such as distance to transportation corridors or movements of human populations, and another for establishment, where predictors would include climatic factors that may limit habitat suitability. Many species-environmental matching models inadequately represent complex interactions in the real world. Missing a key environmental layer can drastically affect model results (Fig. 2e), and poor model performance can alert a modeler to this issue. For example, missing a stream layer (for distance from stream calculations) for a phreatophyte species may weaken the models substantially, resulting in poor evaluation metrics. If a model performs poorly, re-evaluating the predictor layers incorporated into the model may be helpful.

4. Primary caveat #3: no single model works best for all species in all areas at all spatial scales and over time

As evidenced by the results from different model comparison studies and multiple model runs within the studies themselves (Elith et al., 2006; Tsoar et al., 2007), no single model works best for all species in all areas at all spatial scales and over time. Likewise, all models contain sources of uncertainty and share some common assumptions. Below, we describe several related caveats and considerations when selecting, applying or interpreting species distribution models.

All species-environmental matching models have underlying assumptions that are rarely fully met. For example, the models generally assume equilibrium or semi-equilibrium conditions, whereby species distributions are largely unaffected by the time since establishment (i.e., the time needed to fill most available niches), dispersal limitations, or other major ecosystem processes such as competition or disturbance (Austin, 2002; Guisan and Thuiller, 2005; Guisan and Zimmermann, 2000; Soberon and Peterson, 2005). As highlighted in caveats one and two, all models assume adequate sampling across important environmental gradients and the area of interest (model extent) with predictor variables capturing the constraints of the species' distribution (Table 1). Within the statistical model the model generated functional forms of relationships between occurrence data and predictors are assumed to be correct. Modelers may find it difficult to meet these assumptions, especially when dealing with changing conditions (climate, land use) or a new or highly invasive species, but can still produce useful models.

There are additional assumptions and idiosyncrasies of most presence-only models, presence–absence models, and species-abundance models (Table 1). True presence-only models develop relationships between a species' known locations and the environment at those locations, developing some sort of box around the conditions (see Hirzel et al., 2002). Other methods often referred to as presence-only actually require background locations that reflect the available environmental conditions in the region to compare to environmental conditions at known presence locations. Presence–absence techniques require both presence data and absence data, although the absence data could be pseudo-absences (generated locations to represent absences) rather than measured absences. Models using either background locations or pseudo-absence locations can suffer from sampling bias (geographic and/or environmental), if the background or pseudo-absence locations do not mimic sampling bias in the presence data. If possible, bias in presence locations should be mimicked in selection of background or pseudo-absence locations. Many different methods have been developed to try and mimic the sampling bias found in the presence data. See Phillips et al. (2009) for an example using target-group occurrence data as background points in Maxent to reduce the bias and improve models. Background and pseudo-absence should also only be drawn from regions accessible to the species (Barve et al., 2011). Any of the parametric techniques, such as logistic regression, assume locations are randomly collected, which is often not the case.

4.1. No two models are the same

No two models will produce identical results, whether holding inputs constant and switching modeling methods or using the same method with different inputs (Fig. 2f). Sensitivity of the model to variable choice was discussed above in caveat two. The choice of modeling method depends on the type of dataset (e.g., presence-only, presence–absence, or abundance) and the specific study objectives. You can either choose the best performing modeling method(s) based on previous model comparison studies or test multiple methods on your own. There are several comparison studies in the literature (e.g., Elith et al., 2006). Some compare different methods for specific aspects or applications such as presence-only modeling (Ortega-Huerta and Peterson, 2008), presence–absence versus presence-only methods (Brotons et al., 2004), habitat use (i.e., habitat specialist versus generalist; Evangelista et al., 2008), grain size (Guisan et al., 2007a), sample size (Wisz et al., 2008), species range sizes (McPherson et al., 2004), species' ecology (Guisan et al., 2007b; McPherson and Jetz, 2007), prevalence, latitudinal range and clumping (spatial autocorrelation; Marmion et al., 2009) and many others. However, most papers that have tried to quantify different contributions to uncertainty have identified model type as the top contributor (e.g., Diniz et al., 2009; Dormann et al., 2008).

Ensemble modeling may help with these differences among models (Araujo and New, 2007; Marmion et al., 2009; Roura-Pascual et al., 2009), and can at least help identify uncertainty resulting from different choices made during the modeling process. Ensemble modeling can be used to highlight where different models agree and where they disagree (Stohlgren et al., 2010). The ensemble model will only be as good as the models used, and will rely on the accuracy of individual models along with sampling and predictive layers mentioned earlier (e.g., sample size and distribution relative to major environmental gradients and the resolution and accuracy of the predictive layers). And all models contain other uncertainties associated with species characteristics, sampling design, and the completeness, accuracy, and resolution of predictive layers (Stohlgren et al., 2010).

4.2. Species characteristics matter

Many researchers have suggested that generalist species are harder to model than habitat specialist species (Brotons et al., 2004; Evangelista et al., 2008; Tsoar et al., 2007) in species–environment matching models (Fig. 2g). Species also may be generalists over some geographic extents, and specialists over other portions of their distributions. In general, habitat-generalist and habitat-specialist species differ in the ratio of occurrence extent occupied to the extent of the area of interest. The ratio will be much smaller for habitat-specialist species (small patches scattered across the landscape of interest compared to large swaths in the landscape of interest), and it will be more likely that absence locations drawn from the entire area of interest will differ from the presence locations for specialists when quantifying species–environment relationships with the models. This ability to discriminate presence from absence locations is what most evaluation metrics measure. It should be kept in mind that AUC values are not directly comparable between models unless the exact same location data (presence and absence/ pseudo-absence/ background data) are used to compute them.

4.3. Evaluation and calibration

As a first step, model results should be examined to determine if they match what is known about the biology of a species and its distribution. If the results contradict what is known, model implementation and input data should be reviewed. An independent data set is ideal for statistical model evaluation and calibration, but often is unavailable. Alternatives include cross-validation, where the data set is divided into n equal parts (or folds) and n models are run withholding a different subset for testing each time (spatial or not; see caveat two section on autocorrelation). This option has advantages over the single split sample approach because it is less sensitive to spurious results due to a single random selection.

It is wise to calculate multiple evaluation statistics to assess model performance rather than relying on a single statistic. For example, Cohen's Kappa may be sensitive to the prevalence, while AUC (Fielding and Bell, 1997), or True Skill Statistic (TSS; Allouche et al., 2006) may be less affected. Comparing models using multiple evaluation statistics will allow for better overall evaluation (Lobo et al., 2008), and there are several packages, including most of those mentioned early in this article, that make this easy to do.

AUC is perhaps the most commonly used metric for evaluation and is a discrimination statistic representing the likelihood of a presence location having a higher predicted value than an absence location (Pearce and Ferrier, 2000). It is important to note that an AUC value of 0.5 represents a model that is no better than random. So, while AUC can range from 0 to 1, the practical range is from 0.5 to 1.0. Generally, AUC values less than 0.7 imply a questionable model fit (Swets, 1988). However, AUC can be sensitive to the spatial extent from which absences or background points are drawn (Lobo et al., 2008). For presence-background or presence and pseudo-absence data, AUC values should be interpreted cautiously because the values reflect equal weights to the errors of commission and omission (see Jiménez-Valverde, 2012 for more discussion).

Often the purpose of creating a distribution model is to predict suitable environmental conditions in new locations (e.g., model a native range and project to an invaded range) or time (e.g., model current climate conditions and project to future climate conditions). However, if your model performs poorly, it would be questionable to extrapolate that model to other spatial or temporal situations.

Model calibration, which assesses the ability of a model to correctly predict conditional probability of presence of a species (i.e., agreement between predicted probabilities of occurrence and the observed proportion of locations occupied by the species), is generally not calculated for models. Discrimination is only useful to evaluate how well a model performs for a given species in a specific geographic location and point in time, while model calibration is more broadly applicable (through space and time; Jiménez-Valverde et al., 2013). Calibration is important if model results are intended to be used as an absolute measure of probability of occurrence; if relative suitability of locations is the goal then discrimination measures may be sufficient.

5. Primary caveat #4: The results of species distribution models should be treated like a hypothesis to be tested and validated with additional sampling and modeling in an iterative process

5.1. Model interpretation

Given all that is reviewed above, it seems clear that there is an art and a science to model interpretation. Interpretation of results, whether the model aimed to explain patterns of distribution or to predict actual or potential geographic distributions of the target species, heavily depends on the type of modeling algorithm used (Jimenez-Valverde et al., 2008). In general, the simpler methods (e.g., BIOCLIM, and environmental envelopes) may better estimate potential distribution of the species (where the species could survive) whereas more complex techniques that use presence–absence data (in this case real or true absences) may better predict actual realized geographic distributions of the species (where a species actually survives; e.g., Neural Networks; Jimenez-Valverde et al., 2008). Complex presence-only methods like Maxent (Phillips et al., 2006), and GARP (Stockwell and Nobel, 1992) will predict all or parts of realized and potential geographic distributions of a species (Jimenez-Valverde et al., 2008).

It seems prudent for users to remain open to alternative explanations of the same results. Given the caveats described here, models should be considered to *create* (rather than *test*) hypotheses regarding species realized and potential distributions and the factors that may be driving them. When more than one explanation of model results seems plausible, this creates an opportunity for the modelers to consult iteratively with taxon experts in the process of exploring the existing data in more detail, collecting more data, or otherwise obtaining more information. It is important to keep in mind that the models represent (more or less complex) statistical relationships between predictors and field observations. Simply put, these models represent correlation more than causation. The key point is that when interpreting the models, results should not be viewed as a definitive explanation of the way things are, but rather hypotheses about species distributions and their relationship with environmental conditions that can be further tested. Therein is the "art" of modeling.

Accurately modeling species distributions in space is difficult, and ecological forecasting to new geographic areas or in time is even more challenging. The caveats above are compounded when extrapolating to new areas or times. For example, if the predictors used are not what is actually controlling the species distribution, you end up with a spurious prediction. Sax et al. (2013) illustrate the difference between the realized niche for a species and the 'tolerance' niche (climatic space under which the species can survive) by comparing the climate space encompassed by the species' native range to the expanded space when including naturalized or non-native populations. In their example the native range obviously is not limited by climatic constraints but is controlled by other factors that could be explored (e.g., other abiotic conditions such as soil type, biotic factors such as competition). In this example a distribution model based on native range localities and climate would perform fine to capture the current realized range but would yield poor results for the potential range globally or under future climate conditions.

These challenges have not stopped investigators from integrating future climate change scenarios into correlative species distribution models (Franklin et al., 2013; Jarnevich and Stohlgren, 2009). It is not possible to provide a full disclosure of caveats related to climate modeling and projections in association with species distribution models here. Fortunately, the primary caveats for the uncertainties associated with climate modeling have been covered by Tebaldi et al. (2005) and Pielke and Wilby (2012). Often cited caveats include non-analogous climate, a sparse and short climate record, novel processes and interactions as species shift distributions independently of one another, and extremely limited means of evaluating uncertainty for projections in space and time, which can lead to questionable species distribution predictions (Franklin et al., 2013). While such models begin to elucidate a few fundamental species–environment relationships, reliable projections of range changes likely will only result from monitoring species abundance, persistence, dispersal, and migration over time. Other variables including soils; genetics; species interactions; and changes in disturbance regimes, land use, and trade and transportation which continually bring in new invaders are all likely to be important and are unaccounted for in most extant distribution models (Stohlgren et al., 2013). Still, when combined with a careful long-term monitoring program, preliminary species distribution models are essential for identifying potential leading and trailing edges of distributions (invasion or change due to changing climate or land use), and the dominant environmental factors associated with species presence, abundance, and persistence. If the diagnostics indicate a strong model fit and the assumptions of the modeling technique are met by the given application, projecting future distributions is reasonable to consider but should be done with the understanding that the uncertainty inherent in these types of models is compounded by the unknowns of projecting the predictor layers and the predictor–species relationships into the future (Wilby and Dessai, 2010).

In short, we treat all species distribution model outputs as hypotheses to be tested in an iterative process of gathering new data, validating, and improving models over time (Crall et al., 2013; Stohlgren and Schnase, 2006). Thus, we recognize that our "predictions" should be viewed as "possible outcomes", which are impossible to fully assess in terms of accuracy, precision, and uncertainty, except by waiting to see if the predictions hold true with further data collection and assessment.

5.2. Creating "yes/no" maps with thresholds

Models created using presence-only or presence–absence data generally produce results with predicted probability values ranging between 0 and 1. For many model applications, diagnostics, and further analyses, the model predictions are converted into binary maps

representing suitable and unsuitable locations for the species of interest. Selection of a threshold above which to convert locations to suitable and below which to convert locations to unsuitable is important as different thresholds can produce drastically different estimates of suitable range (e.g., Fig. 2h).

Thresholds are generally based on criteria related to weighting false presence or false absence. Weighting one error over another may be more desirable for certain applications such as minimizing errors of commission to assist in reserve design to ensure that the species of interest has suitable environmental conditions within the reserve. Jimenez-Valverde and Lobo (2007) found that the sensitivity–specificity difference minimizer and sensitivity– specificity sum maximizer threshold criteria performed best across a wide-range of species, prevalences, and sample sizes. Liu et al. (2005) also found these two to be good along with three others of twelve they tested. These two methods were also highly correlated with each other and with prevalence (Jimenez-Valverde and Lobo, 2006). A recent examination of thresholds for presence-only models found that some thresholding techniques did not appear to be sensitive to not having absence data (Liu et al., 2013). Ultimately, the application of model output should guide the selection of a decision threshold. For example, if the model is to be used to send crews to control an invasive species, it is more important to have the species be there (higher threshold) whereas if the purpose is for early detection it is more important to not miss any locations (lower threshold).

Table 2
List of guidelines we follow when developing a species distribution model.

Issue	Resolution
Sample size	Minimum of 30 points, but still warn about instability with small sample sizes; smaller samples only used to guide further sampling; always remove duplicates found in the same pixel; fit simpler models with fewer predictors (e.g., roughly 10 locations for each predictor)
Background/pseudo-absence and absence data	If absences are available, use them (except potentially for species with very unstable distributions such as a new invader); draw points from areas accessible to the species; mimic sampling bias in presence data using a bias surface or target group sampling for background/pseudo-absence
Spatial extent	Ensure that the model is trained on area sampled; evaluate bias in samples; background or absence data limited to area accessible to species
Variable selection	Predictors should match direct effects on species; collinearity should be handled (e.g., remove one from each pair of variables with correlations $\|r\| > 0.7$)
Thresholds	Continuous surfaces best but if needed select threshold with care; keep objectives in mind; generally thresholds related to sensitivity and specificity perform well
Evaluation statistics	Examine discrimination using multiple statistics such as AUC, TSS, sensitivity, specificity, kappa, correct classification rate; question models with low performance (e.g., AUC < 0.7); if occupancy is important examine calibration; independent data are best but if unavailable use a cross-validation approach with large sample size or a jack knife approach with small sample size
Model transferal in space or time	Ensure that the model performs well on training area before bothering with transferal; examine stability of correlation structure among predictors (models assume similar structure); highlight areas with novel conditions as having high uncertainty; utilize Multivariate Environmental Similarity Surface (MESS) maps or some other mechanism to explore interpolation versus extrapolation
Response curves	Examine shapes of species response curves: do they make biological sense? and are they overly complex?; very complex curves can often result in nonsensical projections and indicate overfitting; always apply an eco-plausibility filter (e.g., do response curves and distribution maps make biological sense) rather than solely relying on evaluation metrics

6. Conclusions

George E.P. Box said, "Essentially, all models are wrong, but some are useful" (Box and Draper, 1987). We highlighted several caveats related to species distribution modeling, but do not mean to imply that these models are not useful. Rather, we hope to highlight the importance of considering the limitations and sources of uncertainty in models. We strongly recommend including a caveat section in papers based on these models. Pay attention to most or all of the above considerations for better modeling and mapping of species' geographic distributions. Develop a set of rules or guidelines (e.g., Table 2), to consult and follow when you develop a model. Identifying and keeping the purpose of the model in mind may help guide the decisions made in model development. If the principles outlined here are considered and followed, species distribution models can produce results useful to resource managers.

Four take-home messages bear repeating: (1) all sampling data are incomplete and potentially biased; (2) it is important to understand the assumptions and limitations of the model being used; (3) no single model works best for all species in all areas at all spatial scales; and (4) the results of species distribution models should be treated like a hypothesis to be tested and validated with additional sampling and modeling in an iterative process.

Acknowledgments

We would like to thank the U.S. Geological Survey Invasive Species Program for financial support for this work. We thank Colorado State University and the U.S. Geological Survey Fort Collins Science Center for logistical support. We thank Tim Kern and Alan Swanson for providing comments on an early version of our manuscript. This research was partially supported by USDA CSREES/NRI 2008-35615-04666. Any use of trade, product, or firm names is for descriptive purposes only and does not imply endorsement by the U.S. Government.

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From: Holt, Bryon
To: Grizzle, Betty
Subject: Re: Help with locating references
Date: Monday, February 5, 2018 12:34:29 PM
Attachments: Anderson 2002.pdf
Beniston 2012.pdf
Brown and Mote 2009.pdf

Hi Betty,

I've got all but one document. The one document I don't yet have is by Moran-Tejeda, but I have requested it from our Conservation Library and should have it soon to send to you. I will send the ones I have now in a couple of emails to minimize size issues. Note: Brown and Mote is 2009 not 2008 - could not find 2008. May be a typo?

Bryon

On Wed, Jan 31, 2018 at 10:47 AM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:

> Thank you! Will save me time pulling this final draft together.
>
> On Wed, Jan 31, 2018 at 10:44 AM, Holt, Bryon <bryon_holt@fws.gov> wrote:
>
> > I can help you with this.
> >
> > On Wed, Jan 31, 2018 at 10:26 AM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:
> >
> > > Hi Bryon - You offered to help with finalizing the SSA Report, so am writing to ask if you can locate and download the references highlighted in yellow on the first page of the attached publication. There are 7 total and full citations are at the end.
> > >
> > > Please let me know whether or not you have time to help. If not, I can ask others. Thanks.
> > >
> > > --
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Calibration of Conceptual Hydrologic Models for Use in River Forecasting

By
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August 2002

Acknowledgments

The following people help greatly with the preparation of this manual by reviewing the entire contents: Tom Econopouly, Northeast RFC, Reggina Garza, Southeast RFC, Don Laurine, Northwest RFC, Scott Lindsey, Alaskan RFC, and Mike Smith, Hydrology Laboratory. A number of other people provided assistance by reviewing various portions of the manual: Tom Adams, Ohio RFC, Norm Bingham and Tim Buckelew, Northeast RFC, Bob Cox and others, Missouri Basin RFC, and Victor Koren, Hydrology Laboratory. Mike Smith also provided the coordination with the Hydrology Laboratory for the preparation of many of the figures and other material. Many thanks are extended to all these people.

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List of Acronyms

ALERT	Automated Local Evaluation in Real Time (system and network)
API	Antecedent Precipitation Index
ARS	Agricultural Research Service
ASOS	Automated Surface Observation System
BPA	Bonneville Power Authority
CAP	Calibration Assistance Program
CSSL	Central Sierra Snow Laboratory
ESP	Ensemble Streamflow Prediction
ET	Evapotranspiration
FSL	Forecast Systems Laboratory
FWS	Free Water Surface (evaporation)
GIS	Geographical Information System
HRAP	Hydrologic Rainfall Analysis Project
HRL	Hydrologic Research Laboratory (currently Hydrology Laboratory)
ICP	Interactive Calibration Program
IDMA	Interactive Double Mass Analysis
IFLOWS	Integrated Flood Observing and Warning System
IFP	Interactive Forecast Program
MAP	Mean Areal Precipitation
MAPE	Mean Areal Potential Evaporation
MAT	Mean Areal Temperature
MPE	Multisensor Precipitation Estimator
NCDC	National Climatic Data Center
NCEP	National Centers for Environmental Prediction
NDVI	Normalized Difference Vegetation Index
NOAA	National Oceanic and Atmospheric Administration
NHDS	NOAA Hydrologic Data System
NOHRSC	National Operational Hydrologic Remote Sensing Center
NRCS	Natural Resources Conservation Service (was SCS - Soil Conservation Service)
NWS	National Weather Service
NWSRFS	National Weather Service River Forecast System
OFS	Operational Forecast System
OPT	Automatic Optimization Program
PE	Potential Evaporation
PRISM	Precipitation Regressions on Independent Slopes Model
PXPP	Precipitation Preliminary Processing (program)
QPF	Quantitative Precipitation Forecast
RFC	River Forecast Center
SNODAS	Snow Data Assimilation System
SNOTEL	Snow Telemetry (network)
STATSGO	State Soil Geographic Data Base

List of Acronyms (continued)

TAPLOT	Temperature Plotting (program)
TVA	Tennessee Valley Authority
UCSL	Upper Columbia Snow Laboratory
USGS	United States Geological Survey
WFO	Weather Forecast Office

Chapter 1 – Introduction and Background

Introduction

Continuous, conceptual hydrologic models were first developed in the early 1960's with the advent of digital computer technology. Such models have been used for many applications over the succeeding years. One of the main applications is for river forecasting. In order to get the maximum benefits from the use of conceptual models, especially when used for river forecasting applications, the models must be properly calibrated at each point on the river system where a forecast is to be generated. This manual describes the steps and procedures needed to calibrate conceptual models for use in river forecasting and issues involved in the implementation of the calibration results into operational use. Many aspects of this manual are applicable to any conceptual model or any river forecasting system, however, the manual is specifically focused on the models in the National Weather Service River Forecast System (NWSRFS) and the use of these models for river forecasting in the United States.

NWSRFS was initially developed in the 1970's. In the mid 1980's the system was completely redesigned. NWSRFS contains programs for the processing of historical data and the calibration of models using these data, as well as programs for operational forecasting. In recent years graphical, interactive interfaces have been developed for many of the programs. The portion of NWSRFS where the hydrologic models reside is referred to as the forecast component. This component is very modular. Models and other techniques needed to produce simulations and forecasts are coded as separate modules called operations. The operations can be strung together in whatever sequence is appropriate for a given application. This sequential group of operations is referred to as an operations table. Information is currently passed from one operation to another in the form of time series. In a future redesign, operations will be able to process and communicate with each other via gridded fields.

There are several types of river forecasts. There are specific forecasts of river stage or discharge for time periods of hours or days into the future. There are extended forecasts, generally of a probabilistic nature, for weeks or months into the future. There are also area wide forecasts, generally involving very fast responding, i.e. flash flood, events, that warn of the possibility of flooding over a specific city, county, or other entity. In the National Weather Service (NWS), short term (hours or days into the future) forecasts of river conditions at many specific locations are produced at 13 River Forecast Centers (RFCs), each with areas of responsibility encompassing major drainage basins. The RFCs are also increasingly more involved with producing extended probabilistic predictions of river conditions using ensemble techniques. Ensemble techniques are also being applied to short term forecasts in order to generate probability statements for this type of prediction. The public issuance of river forecasts within the NWS is performed by the local Weather Forecast Office (WFO). The WFOs also issue site specific and area wide flash flood watches and warnings using information supplied by the RFCs. NWSRFS is the river forecast system for use by the RFCs, thus it needs to be able to generate short term river and flood forecasts, ensemble streamflow predictions (ESP), and initial state and

guidance information needed by the WFOs to produce flash flood forecasts. In NWSRFS different programs are provided for calibration and historical simulation, short term deterministic forecasts and model adjustments, and ESP, however, all the programs use the exact same models and techniques.

Even though NWSRFS contains a number of rainfall-runoff models, this manual will concentrate on the calibration of the Sacramento Soil Moisture Model (SAC-SMA) [*Burnash et al.*, 1973]. This is the rainfall-runoff model used by almost all of the RFCs. The Sacramento Model and the NWSRFS snow accumulation and ablation model, referred to as SNOW-17 [*Anderson*, 1973], are the primary models referred to in this manual though various channel response and routing models, as well as other NWSRFS operations, are needed to generate streamflow forecasts and thus will be mentioned.

The proper calibration and correct operational implementation and use of a model is critical in obtaining the maximum benefits of river forecasting. The primary benefits of a good calibration and proper operational application of a model are:

- *Short term river forecasts should more closely track observations, thus improving forecast accuracy and lead time and requiring fewer adjustments by the forecaster.*

- *Models can be used to generate reliable extended probabilistic predictions.*

When models are not well calibrated and/or not defined or used properly in an operational mode, the simulated results will quickly deviate from observed values under most circumstances. This reduces the accuracy and lead time of short term forecasts and requires that the forecasts be updated much more frequently. In addition, an improperly calibrated model cannot be used to generate predictions further out into the future, i.e. weeks and months, if it can't reproduce what has happened in the past. For many hydrologic models, values of model parameters or coefficients can be assigned *a priori* based on soils information or other physiographic factors or based on values for a similar nearby watershed. These parameter sets can produce adequate results for some applications, but for river forecasting they will not provide the level of accuracy and reliability that a well calibrated and properly applied model can produce. These *a priori* methods of determining parameter values are helpful in getting initial parameter estimates, but significant improvement is usually possible by continuing the calibration process.

For use in river forecasting conceptual models must be properly calibrated and applied over large areas. While there are many references in the literature to calibration techniques, in almost all cases the emphasis is on the calibration of models to a single headwater drainage area. When applying models to all the river basins in an RFC area of responsibility, the procedures used must not only produce a quality simulation at each point that is modeled, but must be efficient and generate consistent data input and model parameters over the entire region. Efficient procedures are required so that the initial calibration and subsequent updates and recalibrations can be accomplished within a reasonable time period and without an unrealistic amount of resources.

Consistent results are necessary so the spatial variability of the input data and model parameters make physical sense and to make it easier for forecasters to make realistic operational modifications to data and model state variables as they proceed down a river system. In order to produce high quality simulation results that are spatially consistent in an efficient manner requires a regional approach to calibration. That is the type of approach described in this manual. Data are processed for entire river basins involving many forecast points at one time utilizing regional analyses of spatial and temporal variations of the variables. The determination of model parameters is accomplished by first calibrating the headwater area with the best data and least complications and then using these parameters, plus information on the variability of hydrologic conditions over the river basin, as a basis for determining appropriate parameters for the other drainages.

This manual will discuss the steps necessary to properly calibrate and operationally implement and use conceptual models for river forecasting. It will describe the reasoning and procedures behind each step in the process. The manual will focus on the data needed for calibration, methods to analyze the data, techniques for calibrating models in an efficient and consistent manner, and suggestions for the operational implementation of calibration results. References will be made to some of the main NWSRFS historical data processing and model calibration programs, but the manual doesn't contain program input summaries. The manual will not describe the algorithms of individual models and operations, but will concentrate on how to use the models. The manual will also not cover the details of software or web based tools to access historical data and physiographic information. The user should refer to the NWSRFS User's Manual to get information on specific program input and descriptions of models and operations and to individual user guides for other relevant software (the NWSRFS User's Manual can be accessed via www.nws.noaa.gov/oh/hrl/nwsrfs/users_manual/htm/formats.htm - some of the other user guides can be found at www.nws.noaa.gov/oh/hrl/general/indexdoc.htm). This manual is concerned with the concepts behind the calibration process rather than details concerning the software and data sets. The manual attempts to provide the user with the steps and procedures to follow to produce an efficient, consistent, and reliable application of the models to a large area with many forecast points. Opinions expressed in the manual are those of the author.

Conceptual Models and Methods of Application

A conceptual hydrologic model is a model that represents the major physical processes of the portion of the hydrologic cycle it is intended to mimic, however, these processes are described in a simplified form. Most conceptual models, including all those in this manual, are applied continuously and not on an event basis. Conceptual models can be applied on a lumped or distributed basis. A distributed application generally involves breaking a watershed down into many small subareas either based on a grid or physical characteristics. This manual covers the lumped application of such models which includes cases where a watershed is divided into a few subareas based on elevation, travel time, or significant difference in other physiographic factors.

When models are applied on a lumped basis, the input variables are areal averages. If conditions vary considerably over the watershed, the drainage area may be subdivided into a few zones to account for significant variations in precipitation, temperature, snow cover, melt rates, etc. or significant differences in watershed properties. In that case areal average values of the input variables are used for each zone. This is a common approach in mountainous regions. For the lumped application of a conceptual model to produce reasonably reliable results for river forecasting, several factors are important:

- Significant events must have runoff contributions from most of the area. Subdivision of the watershed can improve the results if there are sufficient data to define the spatial variability of the input.

- The amount of runoff must be reasonably large (as a rough guide annual runoff should exceed about 5 inches for marginal results and 10 inches for generally satisfactory results).

- Significant runoff from snow, though adding to the complexity of forecasting, will improve the chances of acceptable results in regions with low runoff amounts since the accumulated snow cover smooths out the spatial and temporal variability of individual precipitation events and makes it more likely that most of the watershed contributes runoff.

These factors are based on experience with applying conceptual models to a wide variety of watersheds under different climatic conditions throughout the United States. Based on these factors, Figure 1-1 gives an indication of how the applicability of lumped, conceptual models varies across the lower 48 states. This figure indicates that for most of the eastern portion of the country and along the west coast there are frequent, large scale storms that produce enough runoff so that simulation results using a lumped application of a conceptual model are generally satisfactory. Also in the high mountain areas of the west satisfactory results are generally attainable due to the significant snow cover that occurs. In the plains region and the lower elevation portions of the intermountain west storms are generally infrequent and localized, resulting in unsatisfactory results when trying to apply a model on a lumped basis. In-between these areas a transition occurs where generally the results from using a lumped application of a model are only marginally satisfactory. This assessment is based on having areal precipitation estimates based on multiple gages.

In recent years gridded estimates of precipitation have become available allowing for the potential to apply hydrologic models on a distributed or at least pseudo distributed basis. Prior to this, distributed applications of models were only possible in areas with very dense gage networks which are very rare under operational forecasting situations. In the NWS, gridded estimates of precipitation are generated on a hourly basis by combining radar derived precipitation estimates with precipitation gage data and other variables [*Seo et al.*, 2000]. The merged radar-gage estimates of precipitation are generated on a 4 km grid. The quality of these estimates varies from generally good results during summer type storms to less than adequate estimates under winter storm conditions, especially snowfall events, and in mountainous areas.

Early studies also showed that a significant bias (under estimation) generally occurred when comparing radar estimates to gage catch over an extended period. New processing methods have been developed and are being implemented to minimize this bias.

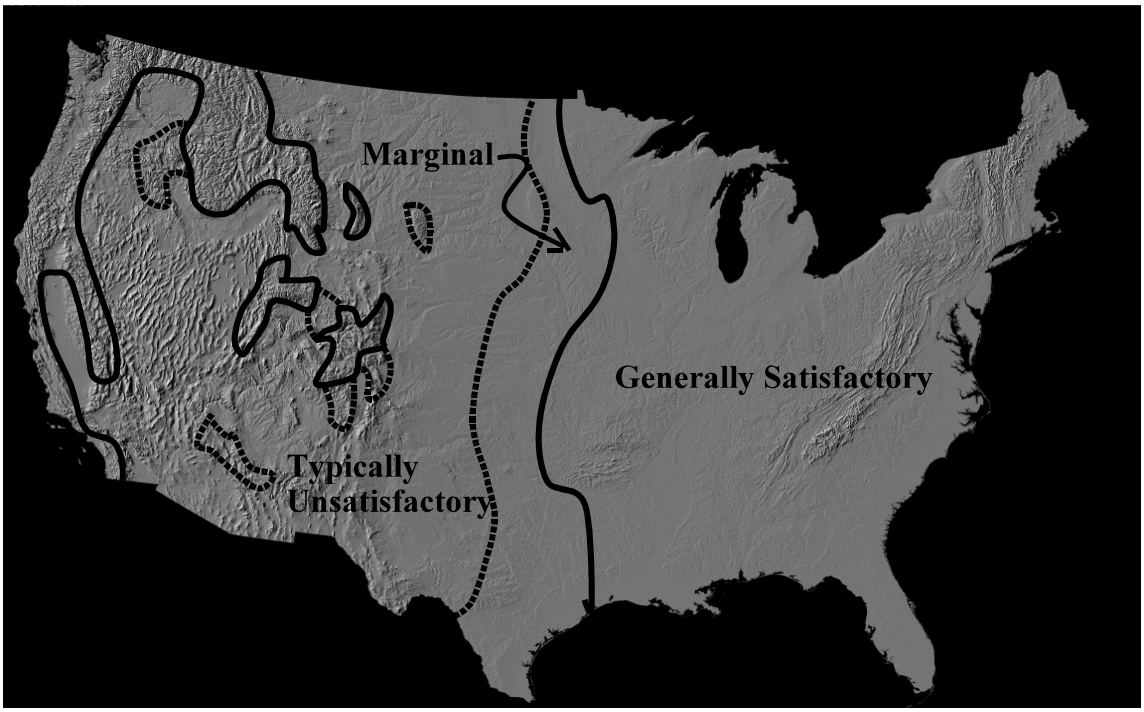


Figure 1-1. Applicability of lumped, conceptual rainfall-runoff models

It is very important to minimize any bias between the input data being used to calibrate a model and the estimates of the input variables being used operationally. Most hydrologic models are very sensitive to data bias. If the data used operationally are biased compared to the calibration input, the benefits of calibration can be negated (discussed in detail in chapter 8). Because of the relatively short period that gridded precipitation estimates have been available and because of periodic changes to the methods used to process the radar data and generate the precipitation estimates, there is generally not a sufficient length of record of these data to fully calibrate a model. Thus, historical gage data must be used to generate input data for calibration. The spatial and temporal density of the climatological networks are such that only lumped applications of the models are possible with the historical data. Also it is normally not possible to obtain reasonably reliable estimates of input variables for time periods less than about 6 hours due to the sparseness of hourly historical data. Thus, besides possible bias in input data values between calibration and gridded operational estimates, there are also issues involving how the models are affected by the different spatial and temporal scales.

The NWS has had a major research effort for a number of years focused on how to best apply models in a distributed mode for operational river forecasting [*Smith et al.*, 1999]. This research has focused on quantifying problems with the radar-gage estimates of precipitation and trying to

determine under what types of hydrologic conditions distributed applications of the models will produce significant improvements in forecast accuracy. The research is also trying to determine how the results are affected by the method and degree of subdividing the watershed, how to determine distributed estimates of model parameters, how to operationally adjust state variables across multiple subareas when simulated and observed conditions differ, and looking at alternative modeling approaches. Early results have indicated that in some areas there is little difference in simulation accuracy between those obtained with lumped and distributed applications of a conceptual model while in other areas there can be significant improvements. These results seem to indicate that the more variation in soils and other physiographic factors and the greater the spatial variability in the precipitation patterns, the better the chance of obtaining improved results by applying the models in a distributed mode. Also it is logical to expect more improvement in areas where the streams rapidly respond to precipitation than in regions with a more damped response.

Another factor affecting the application of hydrologic models in a distributed mode for river forecasting is the case of extended streamflow predictions. It is generally only feasible for an RFC to maintain a single operational model across their area of responsibility. The amount of effort to monitor model simulations and make operational adjustments to input variables and model states as needed, as well as all the other operational and development tasks that must be done, make it impossible to maintain separate models for short term and extended forecasts. Presently extended predictions are generated by taking the current states of the short term forecast models and applying many possible future input scenarios, typically involving 20 to 50 years of data, to generate an ensemble of likely streamflow traces. The input scenarios are generated by attempting to modifying historical data sequences for short term and extended meteorological predictions. Historical data are used because of the difficultly of statistically generating input traces that maintain the proper relationship between different data types, e.g. precipitation and temperature, and the correct spatial correlations over a large region. The input traces currently used for extended predictions are the same as are generated and used for model calibration from climatological networks. Thus, the problem to be resolved for those rivers where extended predictions are needed and a distributed application of the models would be beneficial, is how to generate short term forecasts and maintain model states in a distributed mode and yet make extended predictions efficiently using lumped historical data.

Taking all of these factors into account seems to indicate that the greatest potential for distributed applications of models to improve river forecasts is in the typically unsatisfactory and marginally satisfactory regions from Fig. 1. There is also the potential for improvements in the portions of the generally satisfactory regions where the basins respond rapidly, especially in areas with significant spatial variability in rainfall patterns. However, it is also clear, given all of the data and operational problems, that the lumped application of models and their calibration using historical data will continue to be important for many years into the future in order for the NWS to meet its river forecasting responsibilities in a reliable and timely manner. Thus, the material in this manual should continue to be relevant.

Chapter 2 – Calibration Overview

Introduction

The calibration of a conceptual model to a large region, such as the entire area of responsibility of a RFC, is a major undertaking. This effort must be carefully planned and monitored in order complete the process within a reasonable time frame and end up with consistent, quality results. In order to keep the overall task manageable, it is recommended that the region be broken down into individual river basins or portions of basins involving in the order of 75-200 precipitation stations, 25-100 temperature stations, and 20-50 river locations. The steps and procedures in this manual can then be applied to each river basin until the complete area is calibrated.

General Calibration Requirements

There are a number of factors that are important in order to be able to complete a large scale calibration effort in a reasonably efficient manner and end up with quality results. Some of the most important requirements are knowledge, experience, teamwork, and leadership for the people involved, computerized tools to aid in doing the work, and following proven procedures and strategies.

- Knowledge and Experience – Calibration of a conceptual model is not a skill that one becomes proficient at by merely taking a course or studying a report. True expertise in the calibration process can take years. Even after being involved with model calibration for many years, one still learns something new from almost every river basin that they work on. Thus, it is important when starting a calibration effort or when a person first begins to work on model calibration, that sufficient time be allocated to learn about the process and gain experience. Those that have gained a reasonable level of experience should be assigned to mentor those that are beginning their first calibration. Usually by the time that someone finishes their first calibration they know enough to realize that they should go back and redo what they have just done. If a person receives the proper training, the right guidance, and a reasonable period to gain some experience (in the order of 6 months), they should be able to become a productive member of a calibration team. It should also be noted that the knowledge and experience gained during calibration should pay large dividends in how well the person is able to use the models and procedures to produce operational forecasts.

- Teamwork – There are many aspects to the total model calibration and implementation process. The overall process involves having people with skills in areas such as data analysis, calibration of snow and soil moisture models, development of parameters for routing models, reservoir operations, operational procedures, and the application of geographical information systems (GIS). It is nearly impossible for one person to have the necessary expertise in all these areas, thus calibration becomes a team effort. In building a calibration team it is important to assess the skills and aptitudes of each member. Not everyone has the background and propensity to become proficient at calibrating headwater

models, plus those that are good at calibrating snow and soil moisture models may not have the ability to do the best job at analyzing data or calibrating routing models. Thus, it is important to the successful completion of a major calibration effort to have people with a variety of skills involved and to make maximum use of the abilities of each team member to accomplish the final goal.

• Leadership – Even though a major calibration project requires a team effort, it is also very important that the team has the proper leadership and that the results of the project are carefully monitored and evaluated. Leadership is generally best provided by someone with experience in and a good knowledge of all aspects of the calibration process. This person must also have the necessary leadership skills to work with the other team members and be able to clearly communicate goals, progress, and resource needs to those in charge of the office and others interested in the results. The leader, possibly with the help of others in the office, should carefully review the results of each step of the process to make sure that the results meet performance standards adopted by the office concerning such factors as model selection and basin splitting and that parameter and data values are consistent with nearby watersheds and river basins.

• Computerized Tools – In order for the overall calibration process to be done in an efficient manner, computer tools must be available and used properly to perform many of the necessary tasks. These tools range from number crunching programs to process data and perform data analysis or model computations to GIS applications to display information and generate new data fields to interactive, graphic interface programs that allow the user to better understand the inner workings of models and to interact with the models and data . Without a complete set of tools, portions of the calibration process will become very labor intensive and severely affect the time required to complete the process. In addition such tools assist the user in gaining the knowledge and experience to better calibrate and operationally apply the models and procedures. Tools that exist in NWSRFS or are needed in order to efficiently do calibration in all types of situations include:

- Programs to access historical data bases, inventory the data, and convert the needed data to the proper form for subsequent processing,
- Data processing programs to analyze gage data and compute areal time series of precipitation, temperature, and evaporation,
- Programs that perform the computations for all of the operations necessary to simulate streamflow and other hydrologic variables,
- Calibration Assistance Program (CAP) to perform computations necessary to prepare the input to the data analysis and model calibration programs and to use GIS features to display and generate various data fields, and
- Interactive, graphical interface programs that allow the user to easily view results, make changes, and rerun computations for portions of the process such as model calibration and data consistency analysis.

Many of these tools exist. Others are in the process of being developed. Throughout this manual further details will be given regarding the tools needed.

• Follow Proven Procedures and Strategies – Innovation is periodically required during the calibration process, but there is much that has already been tried and learned. Many procedures and strategies have been developed, tested, and proven that can make the process efficient and more likely to end up with consistent, quality results. It is much better to follow these procedures and benefit from this experience than to try to invent new methods for doing things. If some of the procedures and strategies clearly don't work properly in a given situation, then alternative techniques can be tried. This manual tries to cover the recommended calibration procedures and strategies that have been proven effective.

Basic Steps in the Calibration Process

When calibrating hydrologic models to a river basin there are six steps that are recommended to be followed.

• Gather Information and Data -- The first step is to gather all the information and data necessary for the calibration of the river basin. This includes all available historical data, plus a determination of what real time data are available. It includes gathering maps and data sets that describe physiographic features such as topography, vegetation, and soils, as well as past analyses of the variability of quantities such as precipitation, temperature, evaporation, and snow cover. It also includes information about control structures and their effect on streamflow, plus data on diversions into or out of the basin or between watersheds within the basin and any irrigation effects. Information is also needed on current and possible future forecast requirements. After all this information and data are gathered, the pertinent values need to be accessed and put into the form needed for further processing.

• Assess Spatial Variability of Hydrologic Factors – Before proceeding further with the calibration process it is necessary to take the gathered information and data and get an idea of how various hydrologic factors vary over the river basin. This includes such variables as precipitation, temperature, evaporation, and snow cover, as well as features such as topography, vegetation, soils, and geology. Hydrographs from headwater drainages and local areas can be used to get an idea of how different portions of the river basin respond to the integrated effects of these factors. A knowledge of the spatial variability of hydrologic factors is important in determining how the basin should be divided for modeling purposes and which procedures should be used to analyze the data.

• Select Flow-points and Period of Record for Calibration – The determination of where streamflow is to be simulated during the calibration process is dependent on a number of factors. These include availability of historical streamflow and reservoir data, location of current and future forecast points including those needed to meet all user requirements, location of diversions, irrigation, and control structures, and the variability of hydrologic

factors over the basin. The period of record to be used for calibration is dependent on the period of record of the historical data, especially streamflow, and information on changes that have occurred within the basin over time such as the building of control structures or diversions, increase or decrease in irrigated acreage, and vegetation and land use changes.

• Analyze Historical Data and Put in Form Needed by Models – This step involves the generation of areal average values of precipitation, temperature, and evaporation for the local drainage above each simulation point. In mountainous areas it also involves first determining which drainages must be subdivided into elevation zones and what elevation bands are appropriate. Other areas may also be subdivided if the hydrograph shape varies significantly depending on where the runoff occurs and if there is sufficient gage data to adequately define the input for each subarea. This step also includes the checking of discharge data and the adjustment of these data, if needed, to account for diversions and other factors. Regional water balance computations are an important part of this step especially in mountainous areas to insure that the precipitation, evaporation, and runoff estimates are physically reasonable and consistent.

• Calibrate Hydrologic Models (Snow, Soil Moisture, River, and Reservoir) – Finally after completing all the previous steps it is time to calibrate the hydrologic models to the individual headwater drainages and local areas. The recommended procedure involves the calibration of the headwater with the best data and least complications first and then proceeding to other drainages. Parameters from previously calibrated points are used as initial values for the subsequent areas and only those parameters that clearly need to be changed are altered. Areas that can't be calibrated due to the effect of control structures or because the local contribution is small compared to the total flow are assigned parameter values from the most similar calibrated watershed.

• Implement Calibration Results for Operational Use – The final step in the process takes the results of the data analysis and the model calibration and implements them into the operational systems. This includes the NWSRFS Operational Forecast System (OFS) used for maintaining state variables and producing short term river forecasts and the ESP system used to make extended predictions. The most important factor in the operational implementation of the results is to not produce any bias between the operational application and the historical simulations produced during calibration while still trying to reduce random variations to a minimum through the use of new data sources, dynamic data analysis methods, and real time model adjustment techniques. Bias can occur due to differences in data networks, data types and processing methods, and operational modifications made to state variables.

The subsequent chapters of this manual will describe in detail the recommended procedures to follow for each of these steps in the calibration process.

Extension of Historical Records and Recalibration

Besides the initial calibration of the models to each river basin in an RFC area, it will generally be necessary at times to extend the historical record and to recalibrate at least portions of the area. Reasons for extending the historical record include:

- To have more years of data to use for ensemble streamflow predictions, especially extended predictions. Extending the length of record should improve the confidence of the probabilistic statements that are generated by providing a greater assortment of climatological conditions that might occur in the future. Extensions might be done on a regular 5 or 10 year interval when extended streamflow predictions are being generated for the basin.

- The occurrence of events that were not included in the previous model calibration, such as a new record flood, prolonged dry spell, record snow cover, or surface runoff occurring for the first time. The additional data record can then be used to check the extrapolation capabilities of the models and to make any necessary parameter adjustments (doesn't typically require redoing the calibration).

- Changes occurring within portions of the river basin, such as new control structures, changes in how reservoirs or diversions are operated, changes in agricultural practices, large forest fires, or land use changes. When such changes occur, some of the models being used and/or model parameters for the part of the basin affected typically will need to be modified which may require a recalibration of those watersheds.

Reasons for recalibrating include:

- The people doing the calibrations have gained considerable experience and realize that they can now do a much better job at improving all of the calibration objectives.

- Operational use has uncovered problems and situations that were overlooked or modeled improperly during the original calibration.

- Significant changes have occurred in one or more watersheds within the basin as a result of new agricultural practices, large forest fires, or modifications to land use.

- There is a need to establish some new forecast points at locations with historical streamflow data that were not included in the original calibration. This typically doesn't require a recalibration of all models. Generally it involves subdividing the area between original calibration locations so that simulations are produced at the new forecast points. This requires producing new data input and parameters for channel models, but usually only slight modifications to soil moisture and snow model parameters.

- New methods of determining input data, such as precipitation computed from a

combination of gage, radar, and other data, become available which cannot be made totally consistent with the data used for the original calibration.

• It is decided to change the way that the historical data were analyzed and processed, such as switching to mountainous area procedures when non-mountainous techniques were originally used.

• New models become available which will replace existing models or simulate situations that were not modeled in the original calibration.

• Climatic changes have occurred which could alter climatological average estimates of ET-Demand or other model parameters.

It is very important when extending the existing historical data record to make sure that the new data are consistent and unbiased compared to the data used to determine the current model parameters. This requires that station and areal means used in the computation of the areal estimates of precipitation, temperature, and evaporation for the new period must be the same as those used to generate the data on which the current calibration is based. For example, if the current calibration is based on data from the period 1949 to 1990 and the record is now being extended for the period 1991 to 2000, the station and areal means established for the 1949 to 1990 period must be used in the processing programs when generating data input for the 1991 to 2000 period. Likewise, if the data are further extended in the future, e.g. for the period 2001 to 2010, the station and areal means for the 1949 to 1990 period must continue to be used so that the additional data remain consistent and unbiased with the period on which the calibration is based. If new stations are included when generating the extensions, the means for these stations should be determined by using ratios (precipitation) or differences (temperature) with long established stations that have well defined mean values for the period used for the calibrations.

Recalibrations can be based on the initially processed period of record plus any consistent and unbiased extensions to it or the entire data period can be regenerated prior to a recalibration. If the entire historical record is regenerated involving new types of data or different processing methods, then the models used for the entire area must be recalibrated unless somehow one can guarantee that the new data values are consistent and unbiased compared to the data previously used for calibration. As examples we can related these statements to the reasons for recalibration listed previously in this section.

• If new data types, such as radar estimates of precipitation, or new processing methods, such as switching from non mountainous to mountainous area procedures, are being used to generate the input data, then the data for the entire data period (may not be the same as the original data period due to the length of record for the new data types) needs to be regenerated and the entire area recalibrated.

• When new models are being implemented, calibrations are being redone based on increased

experience, or operational problems indicate a need to reexamine the calibrations, it may or may not be necessary to regenerate the input data. As long as the new models can use the existing data (though additional data types may be needed depending on model requirements) or the simulation problems are not related to the data record, there is no need to regenerate the input for the models. If the new models are being implemented or the existing models recalibrated over just a portion of the river basin, then the possibility of regenerating the input data only needs to be considered for this part of the area.

- Land use changes, new agricultural practices, large forest fires, and climatic changes typically do not require regenerating the historical data record. Generally this situation involves extending the data to cover the period after the action took place and then using this period to modify the model parameters.

- The need to establish new forecast points involves subdividing existing drainages and thus the creation of new areal data estimates. This typically requires that a complete new data record be generated for each new subarea. This only needs to be done for the affected drainage areas. Whenever generating new data input for only a portion of the area it must be remembered that many of the same stations will be used for these areas as are used for surrounding drainages and thus, the station means used to compute the new data input records should be those used when the data were processed for the latest full calibration of the river basin.

When extending the period of historical record or recalibrating all or part of a river basin, the same general principles and recommendations apply as when doing an initial calibration except that certain of the basic steps may not be necessary or require less attention. The gathering of information and data definitely will be a part of an extension of the data record or a regeneration of the model input involving new data types and likely needed when changing processing methods, but, for the most part, will not needed when the existing historical record is sufficient for any modifications to the model parameters. Analyzing the spatial variability of hydrologic factors typically would not be necessary unless completely new data types are now involved though a review of this step could be beneficial in many cases. The considerations regarding the selection of flow-points and the period to be used for calibration should definitely be examined whenever a recalibration, especially a complete recalibration of the entire river basin, is being done. The recommendations regarding the analysis of historical data and the generation of model input applies whenever a new historical data record is being produced. Special considerations apply to the extension of existing records and these will be discussed further in the chapter covering this step. Likewise, the recommendations regarding the calibration of the hydrologic models apply whenever model parameters are being calibrated or modified and the chapter covering this step will include a discussion of items that need emphasis when doing a recalibration. The chapter on the operational implementation of calibration results will also include a discussion of operational changes that may be needed after completing a historical record extension or a recalibration.

Chapter 3

Step 1 - Gather Information and Data

Introduction

The first step in the process of calibrating a conceptual hydrologic model to an entire river basin is to gather all the information and data that will be needed. This includes basic information on the climatology of the area, physical features, control structures and other modifications to natural flow, and operational forecast data and requirements. It also includes all available sources of historical data. This step also involves accessing the relevant historical data and converting it to the form needed for further processing in subsequent steps.

When doing an extension of the data record the exact same historical data requirements pertain to the extension period as are described in this chapter for the initial calibration period. In addition to data for the extension period, data for about 10 years prior to that period should also be gathered so that the consistency of the records over time can be evaluated. When doing an extension of the historical record the basic information that should be gathered is that which indicates changes that occurred during this period such as new reservoirs or other control structures, new forecast requirements, and changes involving agricultural practices, vegetation cover, and land use.

This chapter lists the basic information and historical data that may be needed for the calibration of a river basin, possible sources of the information and data, and data selection criteria. The exact way that the information and data are used in the calibration process are described in subsequent chapters.

Basic Information Requirements

A variety of items should be gathered at the beginning of the calibration process. This includes information about the climate of the basin, physical features such as topography, soils, and vegetation, the location and operating procedures of control structures, and real time data and forecast locations. A number of these should be able to be accessed via the Calibration Assistance Program (CAP). The current features, capabilities, and user instructions for CAP are described in sources outside of this manual. For some basins reports from agencies such as the USGS and Corps of Engineers contain a variety of these types of information that can be helpful in understanding the hydrology of the area.

Climatological Information

- Isohyetal maps showing annual, seasonal, and/or monthly average values of precipitation over the area. This is most important in mountainous areas, but there may be other areas where the variation of average precipitation over the area cannot be ignored when analysing

precipitation data and generating areal average time series for model input. Such areas include those with lake or ocean influences (e.g. areas that include zones affected by lake effect snow patterns). The most common source of digital isohyetal maps is the PRISM maps produced by the NRCS and Oregon State University [*Daly et al.,1994*]. There may be other isohyetal maps for parts of the United States that also should be examined.

• Evaporation analyses that show the magnitude and variation of evaporation over the river basin. One source is NOAA Technical Reports NWS 33 and 34, "Evaporation Atlas for the Contiguous 48 United States" [*Farnsworth and Peck, 1982*] and "Mean Monthly, Seasonal, and Annual Pan Evaporation for the United States" [*Farnsworth and Thompson, 1982*]. These reports contain maps of May-October pan evaporation, May-October Free Water Surface (FWS) evaporation, annual FWS evaporation, and pan coefficients over the lower 48 states. The reports also contain tables of monthly, seasonal, and annual pan evaporation for various sites in each state derived from pan data and computed from meteorological factors. There is also some information on the variation of pan evaporation with elevation. There may be other reports, typically by government agencies or universities, on evaporation analyses for a given river basin or state which would be helpful to have available.

• Other climatological analyses involving variables such as temperature, runoff, snow cover, frozen ground and/or glaciers. The PRISM project has developed maps of average temperature values. In some river basins maps showing how runoff varies over the area have been produced. These maps are generally done as part of a water balance study. Maps of average snow depth or water equivalent may be available for some basins, as well as reports on the variation in frost depth. In Alaska and in the Pacific northwest it is important to know whether the basin is affected by glaciers and if so, the location and coverage of the glaciated area. Reports on how the glaciers are changing over time can also be very helpful.

Physical Information

• A basic map of the river basin. The map should show location of rivers and streams, major lakes and reservoirs, and contain topographic information. The maps may also contain general information about vegetation cover and show locations of glaciers. The maps can not only be used to get a general understanding of the physical layout of the river basin, but are useful to visualize drainage areas and gage locations.

• Information on vegetation cover and land use. It is important to know how much of the basin is covered by vegetation, and the type of vegetation, as well as how the vegetation or crop patterns may have changed over time (e.g. changes to the amount of forest cover due to reforestation, suburbanization, or large forest fires). It is also helpful know which portions of the basin contain agricultural areas and which are urbanized. In addition, NDVI Greenness fraction data can assist in determining the seasonal variation in vegetation activity.

• Soils information. Can be used to subjectively assess how various soil moisture parameters

may vary relatively over the basin. Soils data can also be used to compute initial estimates of parameters for the Sacramento Model. In some northern areas permafrost may exist over portions of the basin and can have an effect on streamflow.

• Geological information. Can be helpful in determining how parameters may vary over the basin, especially groundwater parameters. Such information can also help to determine if parts of the basin significantly contribute to deep groundwater recharge or if karst regions may exist.

Information on Controls and Transfers

• Reservoirs and large lakes. It is important to know where these features are located, how they control the flow, and the magnitude of their effect. Ideally these features should be modeled. In order to do this information such as area-elevation and elevation-discharge curves are needed. For reservoirs, details on how the dams are operated is important.

• Other types of structures. These include small dams (including farm detention reservoirs) and power plants. These structures primarily affect low flows, small storms, and flows after dry periods. It is important to know the general type of such structures and where they exist in the river basin.

• Diversions of water across watershed boundaries for purposes such as water supply, irrigation, and power generation. Diversions can also occur at high water levels when rivers or impoundments may overflow across drainage divides. It is critical to know where such diversions occur so that data can be obtained to model the diversion or adjust the streamflow records to reflect natural flow conditions if the magnitude is significant.

• Irrigation that occurs within a given watershed. Besides removing water from the river, the irrigated land is kept wetter than would naturally occur thus affecting the amount of runoff from subsequent storms, and some of the removed water may reenter the stream later as return flow. It is important to know the amount of irrigated acreage, the rules governing the removal of water, and seasonal changes in irrigation demands.

• Drainage systems. In some agricultural regions drain tiles have been installed under the ground to control the drainage of excess water from the soil. In other areas wetlands have been drained or farming practices may affect runoff generation. It is important to know about these items and when they occurred and how they may have changed over time.

• Large springs. May represent a large portion of the flow at low levels, or they may contain areas where a significant portion of the percolated water can go to deep aquifers. Both of these occurrences can have a important effect on the modeling of baseflow and need to be known and, if possible, quantified.

Operational Information

•Operationally available data. Knowledge of available real time data, especially river and reservoir observations, is needed to help determine how the river basin should be divided for modeling purposes.

• Location of current forecast points and future user requirements. Needed to help determine how the basin is divided for historical data processing and model calibration.

Historical Data Requirements

A variety of historical data needs to be gathered for direct use in calibration. Hopefully the data are already in digital form, however, in some cases critical data may only exist in tabular form and would have to be converted in order to be used. Besides finding the data and determining which records are needed for calibration, the actual data values need to be put into the format required for use by the calibration programs (currently in NWSRFS this is the DATACARD format).

The period of record needed for historical time series depends on how the calibration results are to be used for operational applications. If there is any possible application of the results for extended ensemble predictions, then the longest possible period of precipitation and temperature data should be used in order to generate the most statistically reliable results. Since only a limited number of stations are available in digital form in the National Climatic Data Center (NCDC) climatological records prior to 1948, the longest reasonable data period to process is that from water year 1949 to the time of the most recent available data. If significant changes occurred in the station network over the years, the data for this entire period might not produce time series that consistently represent a reasonable estimate of what occurred over the river basin. For example, in some western areas there were few if any high elevation stations until the advent of the SNOTEL network maintained by the Natural Resources Conservation Service (NRCS). In this case time series produced from only low elevation sites prior to the installation of the SNOTEL stations likely would not produce reasonable estimates of conditions, especially the timing and amount of precipitation, at the higher elevations. Only the period of record after the beginning of the SNOTEL network would likely produce reasonable estimates of model input values and thus would be the period selected for accessing historical data. Chapter 5 contains more information concerning this situation and how to determine if periods of record with very different networks can be mixed.

If the calibration results are only going to be used for short term river forecasting, then possibly a shorter period of record will be sufficient. In this case only the period of record needed to calibrate the models is required. Chapter 5 contains recommendations for selecting a calibration period for a given location, however, since the data are being processed for an entire river basin, the data period must cover the calibration periods of all flow points within the basin. This will allow for the calibration of each point and the routing of flows downstream.

If there are any questions concerning whether a given station should be included at this stage, it is best to include it. When the data are later analyzed, as described in Chapter 6, stations with problems or those that don't add information content can easily be removed, but it is inefficient to go back and retrieve additional data.

The exact programs or procedures to view and access the data are generally not included in this manual. Much of the historical data can be obtained via the NOAA Hydrologic Data System (NHDS). Some of the data are obtained from the agency responsible for collecting the information and then stored and possibly reformatted for use in the NHDS. Various NHDS programs are then available to view and access the data. Other data are obtained directly from agency web sites and then converted into the format needed for NWSRFS by NHDS provided utilities. In other cases individual RFCs have developed programs and procedures to obtain various data from other agencies. Since many of the procedures for obtaining data are changing periodically and may vary from one office to another, this information will not be covered. This manual concentrates on the data needed, selection criteria, and how to use the data.

Precipitation Data

Daily and hourly precipitation data are required in order to generate Mean Areal Precipitation (MAP) time series from historical data.

- Daily Precipitation from the National Climatic Data Center (NCDC). The recommended guidelines for selecting daily precipitation stations for further analysis are:

  - stations within the river basin boundaries and stations slightly outside the boundaries that could be assigned Thiessen weights when computing MAP or would be likely estimators for stations within the boundary that have significant periods of missing data,

  - in mountainous areas, stations that are outside the basin boundaries but represent elevation zones and orographic precipitation patterns within the basin, and

  - stations with at least about 5 years of generally complete data (about 5 years of data are needed to determine a reasonable estimate of the mean precipitation at a given site). Even such short periods of precipitation records are generally included because of its significant spatial variability.

  The NCDC climatological network is comprised primarily of volunteers who receive minimal payments for making observations, thus there are many cases when new observers must be found and stations are moved from one location to another. NCDC station data records are defined by a station number. In some cases when a station is relocated the number remains the same, while in other cases a new number is assigned and thus a new data record is established. Sometimes a station can be moved 10 miles and the number remains the same,

whereas other times one station is discontinued and a new station immediately established only a few miles away. In at least one case a new station was established when there was no site or equipment change, only the name was changed. One option when a station is discontinued and a new station established a short distance away, is to merge the data from the two stations into a single data record. This can be done even when the new location has less than 5 years of data and is especially important to do if the new station is part of the operational network.

• Daily precipitation data from other available networks or sources. One such source is the SNOTEL network that is maintained by the NRCS. This data source is very important in most of the west and Alaska. Daily precipitation data may also be available from other networks such as IFLOWS, statewide or local ALERT installations, regional climate centers, historical data bases maintained by some RFCs, and from Canadian or Mexican sources. The recommendations for which stations to include are the same as for the NCDC network. The situation of station relocations normally doesn't exist for automated networks.

• Hourly precipitation from NCDC. The recommended guidelines for selecting hourly precipitation stations for further analysis are:

- stations within the river basin boundaries, plus stations surrounding the basin that will be used to time distribute records of previously selected daily stations,

- in mountainous areas, stations that are outside the basin boundaries but represent elevation zones and orographic precipitation patterns within the basin, and

- stations with at least about 5-10 years of generally complete data.

By having hourly stations surrounding the daily stations, there is a tendency to dampen the precipitation intensity as a storm moves across the basin, however, this is probably better than having an hourly station in only one direction used to time distribute a daily amount and thus offsetting when the event occurred. Where daily and hourly records both exist at the same location, it is generally best to include both, since there is a tendency for more missing data and less accurate estimates of missing amounts at hourly stations using the NWSRFS programs (the estimation of missing hourly data is only based on other hourly stations, whereas missing daily data are filled in using all stations). Hourly stations that may be assigned weights when computing MAP need only about 5 years of data, whereas one may decide to use only those with a slightly longer data period to time distribute daily records. For those stations used only for time distribution of daily data, their period of record should overlap that of the stations they will be used to distribute. Comments regarding NCDC network relocations and merging of records given for daily stations also apply to hourly stations.

• Hourly precipitation data from other available automated networks or sources. The

recommendations for which stations to include are the same as for the NCDC network. The situation of station relocations normally doesn't exist for automated networks.

Temperature Data

Air temperature data are needed for snow model computations, frozen ground effects in conjunction with the Sacramento Model, and ET computations in the consumptive use operation (alternative to using a PE time series). The air temperature data are used to compute Mean Areal Temperature (MAT) time series. Currently the calibration program used to compute MAT only uses daily maximum and minimum air temperatures, as opposed to the operational MAT program which uses both instantaneous and max/min temperature data.

- Daily maximum and minimum temperatures from the NCDC climatological network. The recommended guidelines for selecting max/min temperature stations for further analysis are:

  - stations within the river basin boundaries and stations slightly outside the boundaries that could be assigned weight when computing MAT or would be likely estimators for stations within the boundary that have significant periods of missing data,

  - in mountainous areas, stations that are outside the basin boundaries but represent elevation zones within the basin that have limited temperature observations, and

  - if possible, stations with generally complete data for the entire or at least a majority of the period of record, though stations with shorter records should be included when there are elevation zones and portions of the basin that are not represented by stations with more complete data. Stations with shorter records are most commonly used in mountainous areas that don't have sufficient high elevation stations with extended records.

Air temperature doesn't have nearly the amount of spatial variability as precipitation, thus one can be more selective when choosing which temperature stations to use. It is not necessary to include stations with short records that represent portions of the basin or elevation zones that have sufficient stations with longer records.

When analyzing temperature data and computing mean areal time series, all stations must have mean monthly max and min values. Unlike for precipitation, there is currently no preliminary processing program that will estimate mean values for the entire period of record in a consistent manner from stations with varying lengths of data observations. Mean monthly values are computed from the existing data for each station. Thus in order to get a good estimate of the long term mean, the station should have data for as much of the period of record as possible (ideally for 70% or more of the historical data period being used). When stations with short data periods are included, the mean monthly values will need to be estimated manually using the procedure outlined in Section 6-4.

• Daily maximum and minimum temperature data from SNOTEL stations and any other available sources including regional climate centers and historical data bases maintained by some RFCs. The selection guidelines are the same as for NCDC stations.

Evaporation Data

Evaporation data are used by several of the NWSRFS rainfall-runoff models (SAC-SMA, API-CONT as an option, API-HAR, API-HAR2, and XIN-SMA), reservoir models to compute surface evaporation losses (RES-J and RES-SNGL as options), and some other operations (CHANLOSS as an option to compute stream evaporation loss and CONS_USE as an option to compute evaporation from irrigated areas). Several of these operations will accept evaporation input in two forms, daily time series or average mean monthly values. Also, the quantities needed vary from Potential Evaporation (PE), to mean values adjusted for vegetation activity, to lake evaporation adjusted for heat storage. In order to obtain the proper input for a given operation, the starting point is either daily or average monthly estimates of PE.

• Daily estimates of PE are obtained in one of two ways:

- Computed from meteorological factors at selected synoptic sites using NHDS programs. These programs currently estimate solar radiation from manual sky cover observations using the method of Thompson [1976]. Since the advent of the ASOS, these sky cover observations are no longer available. Other methods exist for determining solar radiation including direct observations, computing from percent sunshine data, and satellite derived values. None of these options are currently included in the NHDS programs. Based on a study by Lindsey and Farnsworth [1992], it was shown that using sky cover to estimate solar radiation causes the resulting PE estimates to be biased. The bias varies spatially over the lower 48 states. The NHDS procedure includes an adjustment for each available station to remove this bias so that the average annual PE that is produced matches the average annual lake evaporation given by NOAA Technical Report NWS 33.

- Generated using the Mean Areal Potential Evaporation (MAPE) program. This program can use daily PE time series computed from meteorological variables and daily pan evaporation time series to generate daily MAPE values. Daily pan data are obtained from NCDC. Missing pan data are estimated in the MAPE program. Pan coefficients, used to adjust pan observations to become lake evaporation estimates, must be included in the consistency corrections applied to the stations.

• Average monthly estimates of PE can be obtained in the following ways:

- Use the average monthly values computed from meteorological factors generated by the NHDS programs or tabulated in NOAA Technical Report NWS 34.

- Apply the proper pan coefficient to the average monthly pan evaporation values

obtained for NCDC stations with pan observations or tabulated in Technical Report 34.

- Use annual and seasonal average lake evaporation values shown in NOAA Tech Report NWS 33 and prorate these to monthly values using distributions based on stations given in NOAA Tech Report NWS 34 or obtained from NCDC stations with pan measurements. These lake evaporation values have been digitized and seasonally distributed using formulas based on Report 34 data so that monthly estimates of PE can be obtained using CAP.

Streamflow Data

Streamflow data are needed at all river locations where comparisons are going to be made between simulated and observed flows during the calibration of the river basin. Streamflow data may also be needed at locations just downstream from reservoirs so that reservoir inflow can be computed and thus compared to simulated inflow even if the operation of the reservoir itself is not going to be modeled (pool elevation or storage data are also required to compute inflow). Streamflow at locations below a reservoir or at the lower end of an area that is not yet or cannot be reasonably modeled, are also used to generate flows to route downstream. In addition, streamflow data can be used in calibration to remove the contribution from a portion of a watershed significantly affected by a control structure, diversions, irrigation, or another influence which is going to be accounted for operationally, but not during calibration (i.e. information is available in real time to estimate the contribution, but not historically – during calibration only the uncontrolled portion of the watershed is to be modeled).

Since at this point in the process a determination has not yet been made as to which river locations are to be included in the calibration process (to be determined during step 3 and described in chapter 5), it is best to just determine exactly what streamflow data are in a readily available form and wait to access the data and put it in the proper form until the river basin breakdown is completed.

Streamflow data can be in the form of mean daily flows or instantaneous discharges. Mean daily flow values are used to determine the volume of flow and its general timing. Instantaneous discharges are needed for faster responding streams and for locations with diurnal variations in flow in order to refine the timing of the simulated streamflows. Mean daily flows are needed on a continuous basis over many years to verify model performance, whereas it is generally sufficient to have instantaneous discharges available only during major events.

- Mean daily flow time series are primarily available at locations maintained by the U.S. Geological Survery (USGS) and available via the USGS web site. These data have been checked and quality controlled. There may also be some other sources of daily flow data in some parts of the country and in Canada and Mexico.

- Instantaneous discharge data is generally more difficult to obtain. Such data may be

available, at least for a portion of the historical record, from the local USGS district office. Typically the district offices have such data in digital form for more recent years. These may be raw values that have not had the same quality control checks and adjustments as the mean daily flow data. Data for earlier years are for the most part only available in a hard copy form, such as printouts of periodic stage readings or strip charts. In this case the values must be extracted and converted to discharge using the appropriate rating curves.

Instantaneous discharge data can also be obtained from the RFC's archived operational files. Programs are available within the NHDS to assist in extracting data from the operational files and generating time series. Instantaneous discharges obtained from operational files are based on real time stage reports and were converted to discharges using the rating curve defined at that time in the RFC files. There are no quality control checks or adjustments applied, plus the time when rating curve changes are applied may not be the same as used by the USGS when processing the mean daily flow data.

Instantaneous discharges may also be able to be obtained from other agencies or from Canadian and Mexican sources for rivers that cross these international boundaries.

In addition to continuous flow data, peak flow values can also be useful in model calibration. The USGS maintains an archive of peak flow data on their web.. Unfortunately the USGS only archives the peak flow, peak stage, and day of occurrence of the peak. The time of the peak during the day, which is available in the USGS Water Resources Data reports, is not maintained in the archive files. The peak flow data can be downloaded and put in a form that can be used by the NWSRFS PEAKFLOW operation by following the directions given in the NWSRFS User's Manual.

Reservoir and Lake Data

Reservoir data are needed during calibration to either compute reservoir inflow time series or to verify simulated reservoir operations. This also applies to lakes which have a significant influence on the timing and magnitude of stream flows. The main data that are frequently available historically for reservoirs and large lakes are pool or lake elevation and/or storage (storage can be computed from elevation or vice versa using an elevation-storage relationship). To compute inflows either pool elevation or storage values are needed, as well as outflow data. Some reservoirs may have design features and/or operating rules that would require obtaining additional historical time series in order to compute inflows or verify operations.

- The USGS maintains records of daily pool elevation or storage for a number of reservoirs around the country. These records are published in the USGS Water Resources Data reports for each state. At this point in time these reservoir and lake data are generally not available via the USGS web site and there is only limited reservoir data available via the NHDS.

- Reservoir data may also be available from the agency or private company that operates the

dam. This includes the Corp of Engineers, Bureau of Reclamation, government authorities such as BPA and TVA, local river basin commissions, and power companies.

Snow Data

Snow data can be used during calibration to verify computations of the snow model. This includes checking and adjusting, if necessary, the form of precipitation and comparisons of observed and computed water equivalent, areal snow cover, and snow depth. Thus, the snow data that are most useful during calibration are water equivalent, snowfall, depth of snow on ground, and areal extent of snow cover.

- Daily values of new snowfall and depth of snow on the ground are available at many climatological stations from NCDC. Water equivalent data are also available at a few of these stations.

- Daily water equivalent at SNOTEL sites as measured by sensors located under the snow cover.

- Periodic snow course data (water equivalent and depth) are collected at many sites throughout the U.S.. In most of the western states and Alaska the data are collected by the NRCS. In California, west of the Sierra Nevada mountains, snow course data are collected by the State of California. In the upper mid-west and northeast parts of the country, these data are collected by various government agencies and private power companies. Access or information on these data can be obtained from NOHRSC.

- The NOHRSC archives two types of snow data that they are responsible for collecting and processing. The first is snow water equivalent determined from aerial gamma measurements from airplanes. These data represent the average water equivalent over a flight line. Measurements are typically made when requested during periods of significant snow accumulation. The second is areal extent of snow cover determinations derived from satellite data. These data are generated periodically on a routine basis over most of the western states. Data are available for specific drainages and for specified elevation bands within many of the watersheds. Data are also available in a form for visual display and processing using GIS applications. Areal snow cover data are also available for some periods for other parts of the U.S. from the NOHRSC.

Other Data Types

There are other data types that are needed or can be helpful when calibrating a river basin. Which of these data types are available or are needed depends on the hydrology and climatology of the basin and the types of transfers across watershed divides.

- When water is transferred across watershed divides it is necessary to have data to quantify

the magnitude and timing of the diversions. In some cases it is possible to get only a general idea of the magnitude of the diversion. If the magnitude is small relative to the total amount of flow and actual data values are not readily available, the diversion may only implicitly be included in the modeling process. If the amount of the diversion is significant, it should be directly accounted for in one way or another. There are two general ways of handling significant diversions. One way is to adjust the mean daily flow hydrograph to reflect natural flow conditions by adding or subtracting daily diversion flows (may be routed first if travel times are significant). The other way is to explicitly model the diversion and verify by comparing simulated and observed diversion amounts.

Daily data on some diversions are available from the USGS via their web site. Monthly values of some diversions are also published in the Water Resources Data reports of the USGS. Monthly values in many cases can reasonably be distributed into daily amounts for use in adjusting mean daily flow data. It may also be possible to obtain diversion data from the agency, group, or private concern that operates the diversion.

• In some watersheds large springs exist which may contribute a significant portion of the overall flow. The flow from such springs should be explicitly accounted for if possible rather than lumping their effect into the rainfall-runoff computations. The USGS Water Resources Data reports contain periodic flow measurements just below such springs in some cases. These measurements can be used to adjust the daily flow records to remove the contribution from the spring.

• In some river basins water flows across watershed divides during times of high flows. This typically occurs in regions or parts of the basin with little topographic relief. The USGS Water Resources Data reports will normally note where such conditions exist and may at least indicate when such occurrences took place. Further inquiries may be needed in order to quantify the magnitude of such transfers.

• Data on frost depths may be available in some northern areas to help in verifying model simulations of frozen ground effects. These may be in the form of soil temperature measurements at specific sites or direct estimates of frost depth. For example, the Wisconsin State climatologist, gets measurements of frost depth from grave diggers and produces periodic maps of estimated frost depth over the state.

Meta Data

Besides gathering actual data values for use in calibration, it is also very important to obtain information on the methods of making the measurements and how these may have changed over time. This includes the type of equipment used, the observation time, and station relocations. For precipitation especially, it is important to know the type of gage that is used and whether the gage had a wind shield and whether either of these changed over time. Studies have shown significant differences in precipitation catch based on the type of gage and kind of wind shield.

The observation time of daily precipitation and temperature data is important in getting the data values assigned to the proper time. Information on equipment changes and site relocations are very important in trying to assess the consistency of the data and when determining if adjustments should be applied to correct for inconsistencies. General knowledge of measurement methods can be very helpful in assessing the magnitude of possible errors in all the data types and thus the overall reliability of the values. This will be of great benefit in evaluating possible problems in observed data and thus determining when simulated values should possibly not match observations.

- Information on station relocations and observation times of daily reports can be obtained for the NCDC climatological network via the NHDS. Observation times and changes to these times are available only for stations with long records that were used in computing 30 year normals. Similar information for climatological stations are also available from the B44 forms that are filed by the network managers periodically and when changes occur and from information published by NCDC in the Daily Climatological Bulletins. These sources may also contain information as to if and when precipitation gages were equipped with a wind shield. However as a word of caution, the meta data available from NCDC from these sources is not always accurate or complete. This is partly due to the minimal amount of money allocated to the climatological data program. Comparisons between these various sources of meta data for specific stations have shown that the information is not consistent and certainly no one source is complete. One cannot say with certainty that a relocation or equipment change did not occur just because it is not listed in one of these sources. Information on changes at synoptic stations used in the computation of PE from meteorological factors are documented in the NCDC Local Climatological Data Bulletins. Changes in anemometer height should be included in a file used by the NHDS program that computes PE from meteorological factors.

- Meta data for other measurement sites should be obtained from the agency that maintains the data. Information on changes over time to USGS stations is included in the Water Resources Data reports. Location information and observation times for SNOTEL sites is available from the NRCS.

Chapter 4

Step 2 - Assess Spatial Variability

Introduction

After gathering all the information and data that may be needed for the calibration of a river basin, the next step is to carefully analyze the spatial variability of physiographic factors and hydrologic conditions over the basin. This analysis will be very beneficial in all of the succeeding steps. The spatial variability of conditions over the basin is very important in determining which streamflow points should be included in the calibration. Climatic variability will determine the methods used to analyze and process the historical data. It will also be used to determine if the watersheds need to be subdivided into elevation bands or split based on significant variability of soils and/or vegetation. This analysis of spatial variability will also be of great benefit in predicting how model parameter values might vary over the river basin. It will also form the basis for selecting initial parameter values and for assigning values to model parameters for portions of the basin that cannot be calibrated.

There are two general ways to assess the spatial variability over the basin. The first is to look at the variability of various physiographic factors such as topography, precipitation, temperature, evaporation, snow cover, soils, vegetation, and geology. The second is to look at variations in hydrograph response which is an integration of all the hydrologic factors.

Analysis of Physiographic Factors

Introduction

There are a number of physiographic factors, physical and climatic, that should be examined to understand how conditions vary over the river basin. Much of the physiographic data and certain analyses of these data are available in a digital, gridded form that can be processed using GIS applications. Some of these data may still only be in hard copy form. Most of these data should be able to be viewed and analyzed via CAP. Since the data available in the form needed for GIS applications are ever expanding and the features in CAP are periodically changing, this manual will not describe the details of how to obtain, view, and analyze these data.

Physical Factors

The physical factors that are important to examine include topography, soils information, vegetation cover and types, and geologic information. Displays of this information are based on information gathered and analyzed at some point in time. Generally the topography and geology of an area remain constant over the time frames involved in calibration. Soil characteristics also should remain basically constant over the period of record. Vegetation types and patterns can

change in areas subject to land use changes, changes in agricultural practices, and forest variations due to harvesting or large fires.

• Topography – The main question when viewing topographic information is how much does elevation vary over the river basin and within possible subbasins. Significant changes in elevation (more than about 1000 feet) generally indicate that there could be differences in precipitation, temperature, and evaporation that will need to be taken into account when analyzing and processing the data. The greater the range in elevation, the more chance that these variations will be important. One can also look at area-elevation curves for the basin and possible subbasins to determine if the vast majority of the area is at similar elevations and only a small fraction of the area is significantly different or whether there are major portions of the area at different elevations. If there is a significant difference in elevation affecting major portions of the area, it is very likely that mountainous area procedures need to be used when processing the data. As the elevation range increases there is also the likelihood that the individual drainages may need to be divided into multiple elevation zones for data processing and modeling purposes. Significant elevation differences also indicate that, if possible, it is important to include data that represent what is occurring at a variety of elevations.

Topographic maps also generally show rivers and thus one can get a general idea of the shape of the individual watersheds with the basin. If there are watersheds that are long and narrow, this would indicate that these drainages may have a different shaped hydrograph depending on where the runoff occurs and may need to be subdivided if the necessary data are available. Also included on most topographic maps are lakes and reservoirs. It is important to note whether there are large lakes and/or reservoirs which may have a significant effect on hydrograph response and thus will need to be modeled separately. In extremely flat terrain, especially ones with many small lakes and ponds, it may be difficult to determine drainage boundaries and contributing areas. This will result in the drainage area, which is normally known, to be treated somewhat like a parameter in the calibration process. Topographic maps also typically show the location of large population centers and the general extent of forest cover. All of this information should be useful in the subsequent steps in the calibration process.

• Soils – The main thing to examine at this point regarding the soils data is whether there are significant differences in soil properties from one part of the river basin to another. In mountainous areas it could be helpful to see how the soil properties vary with elevation and whether a large change occurs within a particular elevation range. Differences in soil properties from one watershed to another within the river basin may indicate how certain model parameters should vary. Significant differences within watersheds may suggest that these watersheds possibly could be subdivided to improve simulation results. Since this author is not that knowledgeable regarding the details of the information contained in soils reports and gridded data sets of soil properties, only some general insights on what to look for will be offered. First, any information that indicates how the permeability of the soils may

vary over the basin is important. This could be measured properties or general classification of the type of soils (e.g. clays versus silts versus sandy soils). This information will be helpful later in judging how the parameters involved in the percolation equation might vary from one watershed to another. Second, information on how the depth of the soil layers varies over the basin could be helpful in determining variations in tension water capacities from one watershed to another.

• Vegetation – Several things are important to get an understanding regarding the variation in vegetation cover and types over the river basin. First is the general distribution of forest, open areas, and agricultural lands. Second is the variation in the types of forest cover, primarily between conifers and deciduous trees, and the type of crops in agricultural lands. In mountainous areas changes in vegetation cover and types with elevation, especially from semi-arid types of cover to forest types that require a significant amount of water every year, can be helpful in determining if and at what elevation watersheds in these regions should be subdivided. The vegetation information should be helpful in determining variations in certain snow model parameters from one watershed to another and how vegetation types and seasonal activity patterns effect changes in evaporation demand rates over the basin. Variations in vegetation over the basin can also serve as an indicator of how the LZTWM parameter in the Sacramento model may vary as this parameter is partly dependent on the depth of the root zone.

With vegetation it is also important to try to determine if significant changes have occurred in the amount and type of cover over time. These changes could be due to factors such as changing agricultural practices, large scale harvesting of trees, forest fires, or rural areas being suburbanized. In these cases it will be important to use a period of record that reflects current conditions to calibrate the models.

• Geology – Again this author is not that knowledgeable concerning the details of the information contained in geologic reports. Some people within the NWS have indicated that there are relationships between certain model parameters, especially lower zone free water storages in the Sacramento Model, and variations in the geology of the area. Since soil types are probably related in many cases to the underlying geology, soils data and geologic information may show similar general patterns. If geologic information is available, one mainly wants to be aware of significant changes in properties across the river basin. The more knowledge of geology that one has, the better this information can be put to use during the calibration process.

Climatic Factors

Climatic factors to examine at this point in time are precipitation, temperature, evaporation, snow cover, and frozen ground. One is primarily interested in analyses that show the average value of these quantities on a annual, seasonal, or in the case of snow cover and frozen ground, the average value at various times during the year. The values of these factors during extreme years

are also of interest. These climatic factors will have a significant influence on how the river basin is subdivided and how the data are analyzed and processed.

• Precipitation – Isohyetal analyses are needed to assess the variability of precipitation over the river basin. Ideally both annual, plus monthly or seasonal, analyses are available to examine. The primary source of these analyses are the maps produced for the NRCS using the PRISM technique developed at Oregon State University. In some areas there may also be other isohyetal maps available, some which may even involve the use of other data, such as runoff, to assist in the analysis. What is important at this stage is the variation of annual precipitation over the area. This will determine if mountainous area procedures should be used when analyzing the data. If the variation in annual precipitation is small, in the order of ±5% or less, or if the annual average changes very slowly across the basin, then non-mountainous area data processing procedures should be adequate. If there are significant variations in mean annual precipitation over the basin, then mountainous area procedures should be used when generating areal precipitation time series. Variations may be significant even if the terrain is relatively flat due to variations in precipitation amounts based on, for example, distance from a large water body. If mountainous area procedures are indicated, they should typically be used over the entire river basin even though the variation in precipitation may be slight in some subbasins. This is because the ratio of monthly averages between stations is used to estimate missing data in the mountainous area procedure. In order to ensure one station is not estimated from another with a significantly different average value, generally monthly average values need to be defined for all stations. Operationally, over an RFC area a transition can be made from defining monthly averages for each station to not providing these values, as long as monthly averages are defined for all stations that could possibly ever be used to estimate a mountainous station.

• Temperature – The variation in average temperature over the basin determines whether mountainous area procedures should be used when generating Mean Areal Temperature (MAT) time series. Variations in average temperatures can generally be determined using either annual or monthly analyses. These analyses normally use temperature and elevation data. Such analyses are available from NRCS as part of the PRISM project. Significant variations in temperature over portions of the basin indicate that the mountainous area procedure using a synthetic station for each subarea should be used for these regions. The use of the mountainous area procedure for temperature can vary from one watershed or subbasin to the next, depending on the variation of temperature over the area. In general a significant range in elevation indicates that there will be a significant range in temperature. In some cases the variation in temperature could be significant even with flat terrain (e.g. based on the distance from the ocean).

• Evaporation – NOAA Technical Report NWS-33 contains maps showing the variation in Lake or Potential Evaporation (PE) over the 48 contiguous states. In regions of the country where PE varies slowly, the non-mountainous area procedure can be used to determine Evapotranspiration (ET) Demand for use with the Sacramento Model. In regions of the

country where there is a considerable variation in PE over a river basin, these maps typically don't define the variation in evaporation in sufficient detail for modeling. In these areas, the mountainous area procedure involving the development of a ET Demand versus elevation relationship is recommended (see Section 6-5). This also true for Alaska.

• Snow Cover – Snow cover data are used first to determine if the snow model should be included when simulating the hydrograph. In areas where snow occurs infrequently or only in small amounts, snow depth data from climatic stations can be used to determine when the most significant snow events occurred. By examining the hydrograph response to snowmelt after such events, one can judge whether snow computations need to be included when simulating the hydrograph. If the response is small compared to even moderate rain events, then the snow model is probably not needed.

Snow Cover data are also used, along with precipitation and temperature data, to help determine if headwater and local drainages will need to be divided into multiple zones. When significantly different amounts of snow typically accumulate in different parts of a drainage, then the area probably needs to be divided into zones in order to get reasonable simulation results. Averages from snow course sites during the time of maximum accumulation can be used to make this judgment. Also, variations in snow depth from climatic sites during large accumulation periods could be used. Satellite snow cover data may also be helpful. In locations where multiple zones are needed, these comparisons will typically not only show different amounts of snow in different parts of the area, but also depletion of the snow cover will occur at different times.

• Frozen Ground -- In regions where data on frozen ground are available, these data can be used to determine if frozen ground needs to be considered when simulating the hydrograph. These data could consist of soil temperatures at individual sites or frost depth analyses. What is most important from a streamflow modeling point of view is whether significant frozen ground can occur and whether there is significant variation in the amount of frozen ground from one year to another. In region, where soils are frozen every year, e.g. permafrost areas, the soil moisture model parameters can probably absorb the effect the frozen soil has on runoff. In such regions frozen ground doesn't need to be included in model computations. Also in regions with significant amounts of snow every year, very little frost will develop in the soil, thus there is no need to model the effect of frozen ground. The primary areas where frozen ground needs to be explicitly accounted for is where cold temperatures occur frequently in the winter and the amount of snow cover varies considerably during these cold periods. It is also more likely that frozen ground is important in non-forested areas since a dense forest acts like a snow cover to insulate the ground. In addition, frozen ground will have a greater effect in regions where the soils are reasonably permeable than in areas with low soil permeability. This is because the existence of frost will have a greater effect on the percolation rates.

Integrated Analysis of Hydrologic Conditions

Introduction

The hydrograph at a point on a river integrates all of the factors affecting runoff generation above that point. By comparing hydrographs at different points within a river basin it is possible to determine if the factors controlling runoff are similar from one area to the next or whether differences exist. These differences may be partly the result of differences in climatic input, i.e. precipitation amounts, temperature, etc. and will hopefully be accounted for by the input time series used for modeling. Other differences may be the result of differences in soil properties, vegetation cover, etc., and thus indicate variations in model parameters from one part of the basin to another. Still other differences can be caused by the effects of lakes, reservoirs, channels, diversions, etc., and thus indicate the need for additional models or runoff adjustments. Generally two types of hydrograph comparisons are useful to examine. The first involves comparing watersheds with few man-made controls in order to get an indication of how parameter values might vary across the river basin. This includes headwater areas and those local areas where a good definition of the hydrograph can be generated by merely subtracting upstream flows from downstream flows. The second involves comparing hydrographs at locations with man-made controls and downstream river locations to determine the effects of such items as reservoirs, irrigation, and channel routing.

Headwater Comparisons

These comparisons involve headwater areas with few complications (i.e. no significant reservoirs, diversions, or regulation of flow). In some cases local areas without complications can be included if a good definition of the local hydrograph can be obtained by merely subtracting upstream from downstream daily flows or by using a simple routing of the upstream flows before computing the local contribution. Also headwater inflows to reservoirs can be included if a good definition of the inflow hydrograph can be computed from pool elevation or storage and outflows from the reservoir. The purpose of the comparisons at these sites is to determine if model parameters should be similar over the river basin or whether they are likely to vary. Variations in the hydrograph response from different parts of the basin can be used to predict which parameters are likely to change and the pattern of these changes. This information may be able to be associated with physical attributes such as soils and vegetation to help in determining which headwater area's parameters are best to use as the starting point when calibrating other headwaters, including those with complications, and local areas.

The comparisons are done by plotting the results using the WY-PLOT operation graphical display in ICP. Typically in order to not have too many lines on the screen so that one can properly visualize the differences between individual locations, no more than about 7 hydrographs should be included on a single WY-PLOT. In river basins with more than 7 gages to plot, several plots can be created by grouping the streamgage locations in various ways (e.g. by general location within the river basin or by separating those with no complications from those

with minor disturbances). Before plotting the hydrographs, some scaling needs to be done so that a proper assessment can be made of the displays. First, the hydrographs should all be scaled to a common drainage area. The drainage area of one of the locations is typically selected as the base and then the other hydrographs adjusted to that drainage area by using a WEIGH-TS operation for each. The WEIGH-TS operation multiples all the daily flows by the ratio of the base area to the area of the drainage being adjusted. Second, in river basins where there is a significant difference in runoff from one part of the basin to another, it may be helpful, in addition, to scale the flows to a common mean annual runoff amount. This again can be done by using WEIGH-TS operations and the ratio of a selected base annual runoff to the annual runoff at each location. A common period of record should be used to compute these ratios. Once the hydrographs are scaled, then the responses can be evaluated.

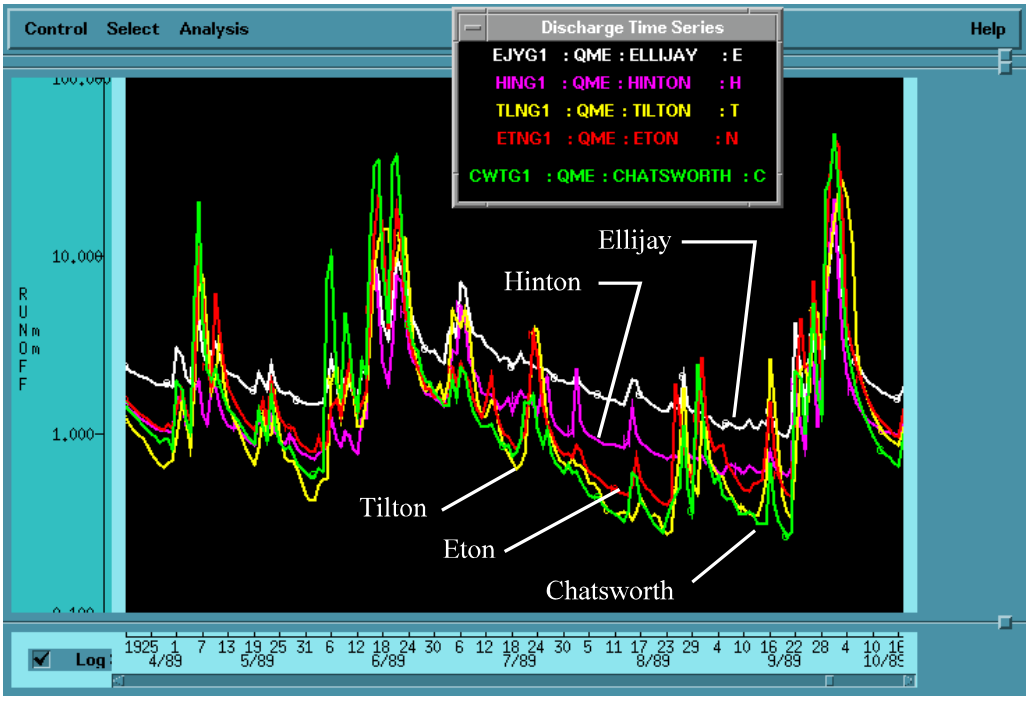


Figure 4-1 - Hydrograph comparisons for Oostanaula River basin.

Figure 4-1 shows a semi-log comparison of 5 locations within the Oostanaula River basin above Rome, Georgia. Eton and Chatsworth are small headwater areas above the Tilton streamgage. The Tilton plot is for the entire drainage above that location. First, it can be seen that the response for the Eton and Chatsworth drainages are very similar to the response for the entire area above Tilton. Since Eton and Chatsworth are not points that require a forecast, they were not modeled separately based on this comparison. If the response for Eton or Chatsworth were different from the entire area above Tilton, then improved simulation results would be likely by modeling the small headwaters separately from the rest of the area above Tilton. Second, it can be seen that there are significant differences between the responses from the Tilton, Ellijay, and Hinton drainages. The amount of baseflow is greatest above Ellijay and least above Tilton. The amount

of storm runoff is greatest for Tilton and least for Ellijay. This pattern clearly indicates that the percolation rates for these areas are quite different, since percolation separates fast response storm runoff from slow response baseflow. Most of this difference should be able to be modeled by altering the lower zone free water capacities (LZFSM and LZFPM) from one area to the next though there is also some evidence that the slope of the supplemental baseflow (model parameter LZSK) probably also varies between the drainages (greatest for Tilton, least for Ellijay). Since there are apparent differences in the immediate storm runoff, the UZFWM parameter may also vary between the drainages in addition to the percolation rates. All of this information will be very helpful in deciding how to divide the river basin to get the best simulation results, in making parameter adjustments, and in assigning parameters to other portions of the river basin.

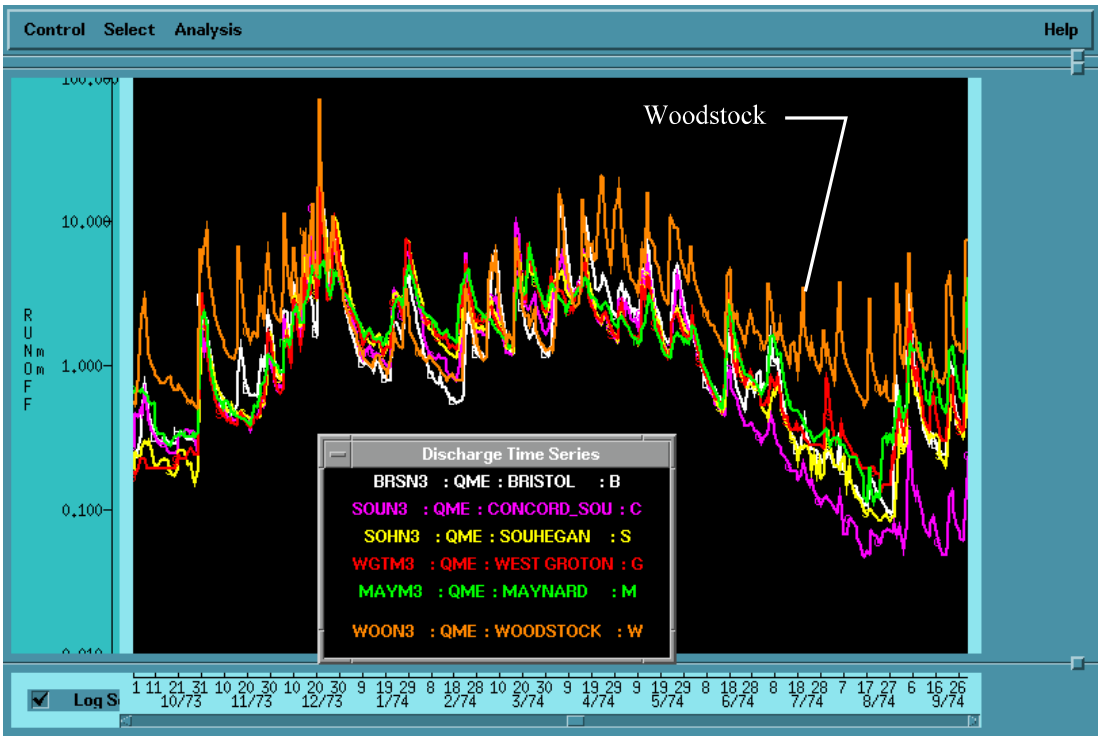


Figure 4-2 - Hydrograph comparison for the Merrimack River basin.

Figure 4-2 shows a comparison of a number of headwater drainages scattered across the Merrimack River basin in New Hampshire and Massachusetts. Even though there is some

difference in annual runoff amounts (especially more runoff above Woodstock), the plots are scaled only by drainage area. In this case the shape of the hydrograph response is, in general, very similar from one headwater to the next. This would indicate that there should only be small variations in model parameters from one area to the next. The area above Woodstock (in the White Mountains at the northern end of the river basin) shows some differences in response from the other areas. There is more and later snowmelt runoff in the spring which is expected because of the mountain location. Low flow levels in the summer are higher than for the other areas suggesting that the lower zone primary storage (LZFPM) may be greater for the White Mountain drainages. There is also more quick storm response which is partly due to more rain in the mountains, but may also indicate more frequent surface runoff in this region and thus a lower value of UZFWM.

Figure 4-3 is from a number of headwater areas draining from the east into Puget Sound in Washington. Since there is a very large difference in annual runoff from one drainage to another in this region, the hydrographs are scaled by annual runoff amounts in addition to drainage area. During the fall and winter period the hydrograph response is very similar for all the drainages. During the spring and summer, there are differences in the magnitude and timing of the response. In the spring most of the variation is due to differences in the amount and timing of snowmelt. Some of the areas, such as Arlington, get very little snow accumulation, while others, such as Goldbar, get a lot of snow. During the summer most of the variation is due to differences in glacial contributions, with National, which drains off Mount Ranier, having the greatest glacial effect. Thus, in order to determine whether the variations in hydrograph response between areas can likely be accounted for by varying model parameters or whether there are other causes, one needs to have a reasonable understanding of how factors such as topography, snow cover, and glaciers vary over the region. In this case, it appears that the model parameters should be very similar from one drainage to another and that the differences in the response will be accounted for by the magnitude and timing of snowmelt and the size of the glaciated area.

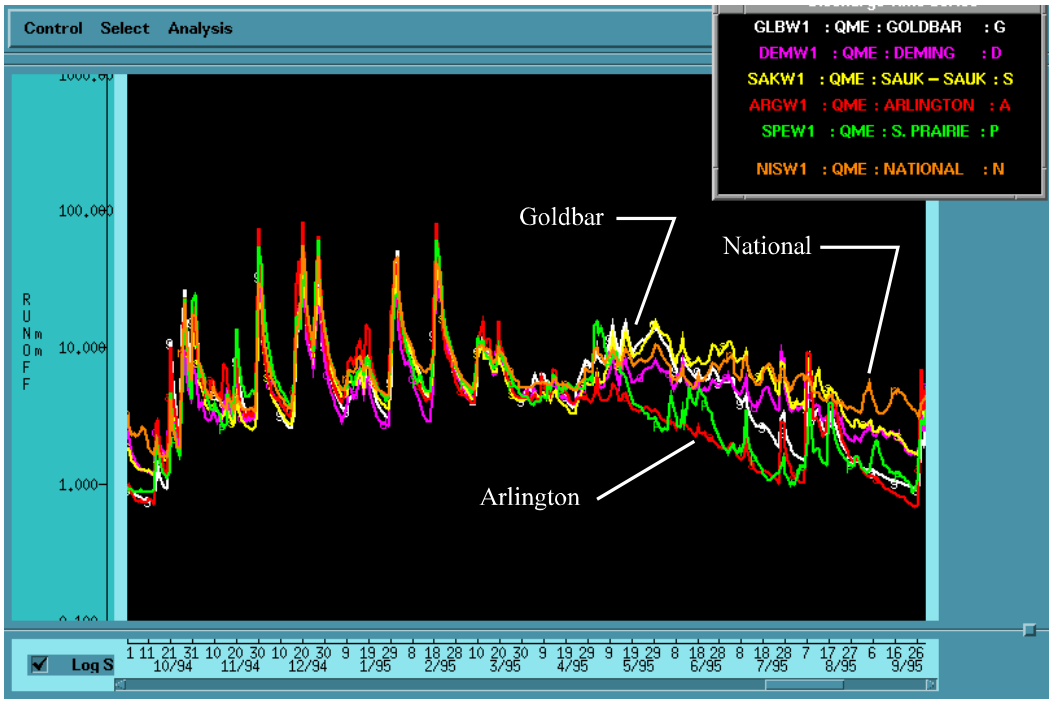


Figure 4-3 - Hydrograph comparison for eastern Puget Sound drainages.

Hydrograph Comparisons at Other Locations

Hydrograph comparisons can be made at other stream gage locations to assist in understanding the effect of features such as reservoirs, lakes, diversions, power plants, etc. and to get a general idea of the lag and attenuation of the flow as it passes downstream through the river channels. In general, these plots will not help with determining possible variations in model parameters, but will assist in determining what modeling procedures and adjustments will be needed to generate reasonable flow simulations and how to divide the river basin for modeling purposes. For these comparison plots, it may or may not be appropriate to

scale the hydrographs before plotting the values. If a semi-log plot is going to be used to view the streamflow traces, then the discharge time series should be scaled to a common area since the WY-PLOT operation uses a single area to convert daily flow into depth of runoff. This may be the case when one is comparing the response from headwater areas, affected by such features as lakes or reservoirs, diversions, or power plants. If an arithmetic plot is going to be used, which is normally the case with many of these comparisons, then no scaling is needed. An arithmetic plot is generally used when comparing the flows at a series of downstream river locations to assess the effects of the channel, reservoirs, or other features that affect the magnitude and timing of the flow.

Figure 4-4 shows a semi-log comparison of upstream locations in the Merrimack River basin that exhibit man-made effects. The hydrograph for Bristol is included to show the response for a location with minimal complications. The flows at Ashland, Lakeport, and Tilton are significantly affected by large lakes. Ashland is downstream from Squam Lake. Lakeport is at the outlet from Lake Winnipesaukee and Tilton is further downstream below the outlet from Lake Winnisquam. The effect of the lakes definitely needs to be explicitly modeled in order to produce reasonable simulations at these locations. The low flows at Goffstown during this period are affected by a power plant, which appears to release only a minimal flow on weekends. At this location it will be difficult to adjust the model parameters that control low flow with any degree of confidence. These parameters will most likely be obtained from a similar nearby watershed.

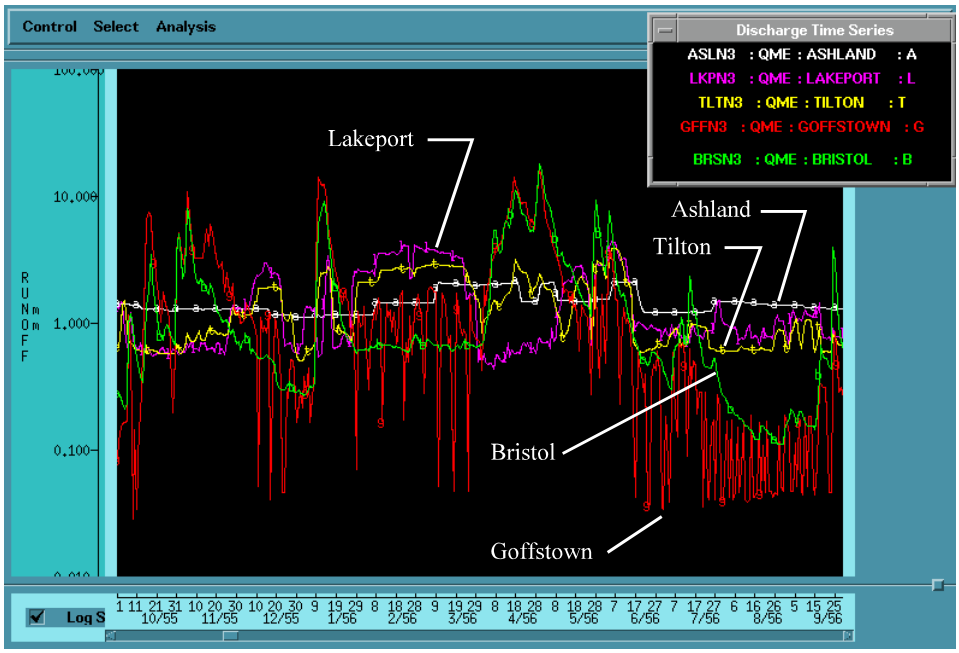


Figure 4-4 - Merrimack River basin locations made with man-made effects.

Figure 4-5 shows an arithmetic plot of the hydrographs at several downstream locations on the Merrimack River. This figure indicates that the amount of lag and attenuation is not large as the water moves through the channel system or does it vary significantly with the flow magnitude. It also shows that there are some minor low flow controls, at times, above the further downstream locations. This comparison indicates that channel routing should be fairly straight forward in this basin and that local area hydrographs, determined by subtracting upstream flows from downstream, will exhibit a lot of noise at lower discharges, though a reasonable definition of the local area contribution might be possible at high flows. The computation of the local area flow will also require subtracting gaged tributaries between these river locations and this will affect whether a reasonable local hydrograph can be obtained. Whenever the local area is small compared to the upstream areas, it is difficult to obtain a reasonable definition of the local hydrograph.

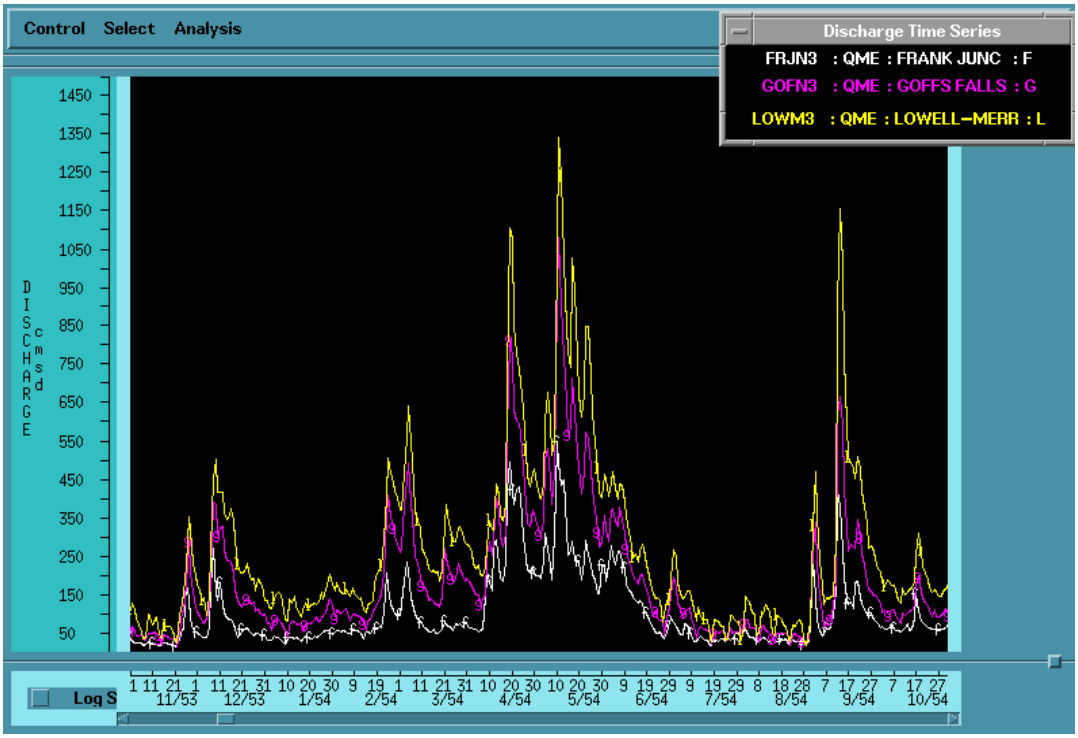


Figure 4-5 - Comparison of Merrimack River basin downstream locations

Chapter 5

Step 3 – Select Flow Points and Period of Record

Introduction

The next step in the calibration of a river basin is to select the locations at which streamflow will be modeled and the period of record to use. The locations to model streamflow are primarily selected based on the location of historical daily flow data, the location of forecast points, the real time operational data network, the spatial variability of hydrologic conditions, and the location of control structures. During calibration streamflow may need to be modeled at more locations than just forecast points in order to properly handle the variation in hydrologic response and the effect of reservoirs, diversions, etc..

There are two periods of record involved in the calibration process. The first is the period to use when processing the historical input data. This period is dependant on the length of the available historical precipitation, temperature, and evaporation records, the consistency of the historical network over time, and the type of forecast products to be generated. The second is the period of record to use when calibrating the models. This period is dependant on the length of the historical streamflow records, the variability and noise in the streamflow data, and changes in vegetation and land use.

It is important to carefully assess all these factors and make a deliberate decision regarding which locations to model and the periods of record to use before processing the data and beginning to calibrate the models. Making these decisions at this point will save a considerable amount of time and effort in the long run.

Selection of Points to be Modeled

The following steps can be followed to determine which river locations should be modeled during the calibration of a river basin.

1. All forecast points with historical streamflow data should be modeled. This includes both existing forecast points and new or potential forecast points. In the case of reservoirs and lakes where inflow or water level is to be forecast, historical pool elevations and possibly outflows will be needed. This is an excellent time to review forecast needs in the river basin and, with input from WFOs and users, decide if all existing forecast points are still needed, whether there are new locations for which forecasts should be provided, and try to anticipate future needs. This includes the consideration of all types of forecasts, including floods, low flow, water supply, recreation, navigation, and reservoir operations. It also includes both short term and extended forecasts. If there are forecast locations that don't have any historical data, then typically a procedure is later developed for operational use that takes the results from a location with historical data and extrapolates those results to the forecast point.

2. River locations that have real time data available and a historical record should in many cases be modeled even if they are not forecast points. Such locations can be valuable in that operationally computations can be verified and adjustments made to improve the lead time and accuracy of forecasts at downstream points. However, if the drainage area reflected at such a location is small and especially when data are sparse, the combination of streamflow measurement and data input errors may make it very difficult to make realistic adjustments. Errors need to be averaged over a sufficiently large area in order to make real time adjustments that will improve the forecast results at a downstream location.

3. Locations at the outlet of reservoirs or lakes that have a significant effect on movement of water through the river system and locations of diversions into or out of the stream network need to be considered for modeling during calibration. Typically there are two options at such locations. First, the control structure can be modeled. In this case historical data such as pool elevation, storage, outflows, or diverted quantities are needed so that a comparison can be made between simulated and observed values. Second, the outflow from a reservoir or lake or the amount of diverted water can be obtained from the agency that operates the structure. Operationally this must include actual, as well as forecast values, so that the operation of the structure will not need to be modeled. During calibration the historical streamflow record will need to be modified to reflect natural flow conditions (described in Section 6-6).

4. Other river locations with historical streamflow data also need to be considered for modeling during calibration based on the variability of hydrologic conditions over the river basin and the available data networks. These gages may be upstream from a headwater forecast or real time data point or they maybe at intermediate locations within the river system. If these gages are at locations where there is likely a significant variation in hydrologic response based on hydrograph comparisons or other information such as soils or vegetation data, then these gages should be considered as possible sites for calibration. Also if the hydrograph response for the area above or between previously determined calibration locations can vary considerably based on the spatial variability of rainfall, then these additional sites may need to be included during calibration. Another factor to consider is whether there is sufficient historical and real time input data, primarily precipitation, to adequately define the areal input if the drainages are subdivided in order to model these additional locations.

After considering all these factors the flow points to be modeled as part of the calibration process can be selected. In many cases, especially in areas with significant variability in hydrologic conditions or rainfall amounts, calibration simulations and comparisons with observed values will be produced at more locations than just forecast points. The inclusion of these extra locations should improve the simulation results at downstream locations and result in more accurate and timely operational forecasts.

Selection of Periods of Record

As mentioned in the introduction for this chapter there are two periods of record that need to be determined. First, is the period to generate input time series such as precipitation, temperature, and evaporation. This is referred to as the historical data period. Second, is the period used to determine the parameters for the hydrologic models. This period is referred to as the calibration period and is typically a subset of the historical data period.

1. Historical Data Period of Record – This period depends on the length of available historical data, the consistency of the network over time, and the type of forecast products to be generated. An appropriate period is based on the influence of each of these factors.

a. Generally in the United States digital records of climatic variables are only available starting in 1948. This is when NCDC started to put the data for all climatic stations in digital form. Records prior to that time have been added to the digital archives when additional funding was available, however, in most parts of the country these earlier records are only for selected locations and do not result in a network that is dense enough for hydrologic modeling.

b. While in most parts of the country the density of the historical network and the representativeness of gage locations are fairly similar over time there are some regions where significant changes occur in network coverage. A prime example is in the western states with the advent of the SNOTEL network maintained by the NRCS. The SNOTEL network consists primarily of high elevation sites where a sizeable snow cover can exist. In some portions of the west there were a reasonable number of high elevation sites prior to the advent of the SNOTEL network, but in other parts of the region there were few, if any, high elevation locations during the earlier years. Most of the runoff in much of the west is generated at the higher elevations. It is generally impossible to produce realistic time series of precipitation at high elevations based only on data from low elevation sites. One can adjust for the annual or seasonal difference based on an isohyetal analysis, but the time distribution and frequency can't be accounted for properly. Much of the precipitation at high elevation locations occurs when no precipitation is falling at lower elevations.

When there are significant changes in the network during the period of available historical data, it may be that only a portion of the data can be used to generate input time series for use in calibration. This is certainly the case in the west for basins where there were little or no high elevation data sites prior to the SNOTEL network. In these basins only the period of record after the advent of the SNOTEL network should be used to generate input time series. In other areas of the west there are a reasonable number of high elevation sites prior to SNOTEL. In some of these areas the entire period of available data can be used to generate time series that have similar properties. In other areas there may be statistical differences in the time series produced with data before and

after SNOTEL such that parameter determination and simulation results will differ significantly from one period to the other. In such cases it may be necessary to do some comparisons at selected locations to determine if the different periods of record are compatible. This can be done by generating input time series for the entire period of available data and then calibrating the models using the period with the best data coverage, typically the period with SNOTEL data. After this is done, the parameters determined from this period would be used to simulate the earlier period and a comparison made of the simulation results. If the results are similar for both periods then the entire length of available data can be used to generate input time series, whereas if there are significant differences in results then the two periods will not produce compatible time series and only the later period should be used during calibration.

c. If extended probabilistic forecasts are to be generated anywhere in the river basin or at locations further downstream using flows routed from the current basin, then the longest possible period of record that will produce consistent values should be used to generate input time series. In this case the time series will not only be used for determining model parameters, but will be used as inputs to generate possible streamflow traces with the ESP procedure. The more traces available to statistically analyze after an ESP run, the higher the confidence in the results. Extended probabilistic predictions produced using ESP are primarily of value in river basins where the current state of the system can have a significant effect on what happens in the future.

On the other hand, if extended applications of the ESP procedure will never be used in the river basin or at downstream locations, then only a period sufficient to determine model parameters needs to be used to generate input time series. The considerations in determining this period are discussed in the next section. Due to differences in periods of streamflow data at different locations and the effects of vegetation and land use changes on the period used to calibrate the models, the calibration period for one location within the river basin may not be the same as for others. The period used to generate input time series should overlap the calibration periods to be used for all locations.

2. Calibration Period of Record – This period is based on the length of historical streamflow records, the variability and noise in the streamflow data, and vegetation and land use changes. An appropriate period is selected for each point to be modeled after considering the influence of each of these factors.

a. At a given location within the regions where lumped, conceptual models produce satisfactory results (see Figure 1-1), typically about 10 years of streamflow data are adequate for determining model parameters. This length of record is needed so that one can have reasonable confidence that the noise in the data signals is random. The shorter the period of record the greater the chance that the data noise will not be random, thus resulting in biased parameter values. Parameter sets based on only a few years of data will typically vary from one time period to another. For these regions a calibration period

of about 10 years should result in fairly stable parameter values. In areas where the model results are generally marginal, a longer period of record likely is needed in order to have enough events and to minimize the noise caused by the typical large spatial variability of rainfall that occurs in these regions. This longer period is needed so that at least a reasonable degree of confidence can be assigned to the model parameter values. In regions where model results are typically unsatisfactory, it is unlikely that any length of record will be sufficient to determine parameter values with any significant degree of confidence. For the initial headwater basin to be calibrated (see chapter 7 for more details), it is recommended that besides the period used to calibrate the model parameters, that another period be available to verify the calibration results. This period should ideally be at least as long as the calibration period.

b. The period of record selected for model calibration should contain a variety of hydrologic conditions. There needs to be wet periods with high flows, as well as dry periods with low flows. If snow is a factor, there should be years with below normal snow cover and years with above normal cover, as well as near normal years. In regions where climatic indices are proven to have a significant effect, the calibration period should include a full range of values of the index (e.g. if the Pacific Oscillation is important, years with a strong El Nino effect, as well as years when La Nina dominates should be included). At least for the initial headwater calibration, it may be a good idea to exclude the flood of record and the lowest flow period from the calibration and use these events to verify the extrapolation capabilities of the calibration.

c. In areas where physiographic factors are essentially constant over the period used to generate the input time series, any portion of the period that contains sufficient variability in hydrologic conditions can be used to calibrate the models. However, in areas where there are significant changes in vegetation or land use, the portion of the period of record that should be used to calibrate the models is that which most closely reflects current conditions. This is typically the most recent period though, for example, in the case where a watershed has just recovered from the effects of a large forest fire, the period prior to the fire would better reflect the current state of the drainage. In agricultural areas that have seen large changes in crop type or farming practices or for watersheds near metropolitan areas that have undergone considerable suburbanization, the most recent period of record should be used to calibrate the models.

In many river basins the periods of historical streamflow data vary from one location to the next. Also vegetation or land use changes may only affect certain portions of the basin. The aim is to try to find an appropriate period to calibrate the models that is as common to as many flow points as possible. There are a couple of reasons for this. First, when calibrating downstream locations it is best to have observed data at the immediate upstream sites so that errors in the upstream calibrations are not propagated downstream. Second, having observed flows at each point within the basin during the period used will allow for a comparison with total simulated flows to mimic forecast and extended ESP conditions.

Chapter 6

Step 4 - Historical Data Analysis and Processing

Introduction

This step involves analyzing and processing those data that are not yet in the form needed for model calibration. These primarily are precipitation, temperature, and evaporation data. Lumped hydrologic models require mean areal estimates of these variables, thus this step involves the computation of mean areal values from the station data retrieved in step 1. Even if point model computations are to be done, the station data need to be checked for consistency and any missing data gaps filled in. In addition to the areal estimates of model input variables, mean daily discharge data may need to be checked and modified for the effects of diversions or other control structures. The other data needed for model calibration mentioned in chapter 3 should already be in the form needed for model calibration and do not require further processing. These additional data are typically used to verify or modify simulation results.

This is a very important step in the calibration process. There is a tendency in many cases for people to rush through this step in a mechanical, cookbook fashion without carefully evaluating data consistency, understanding the variability in the data fields, and determining how to properly use the available data to generate physically realistic estimates of these input variables. This can easily result in time series that are inconsistent and biased. This is especially true in areas with significant spatial variability in data values on a seasonal or annual basis such as mountainous regions. The snow and rainfall-runoff models are designed to represent those processes and are not designed to deal with the interpolation and extrapolation of data fields. The models do contain some very simple factors for making minor data adjustments (e.g. the SAC-SMA operation contains a precipitation and evaporation multiplier and the SNOW-17 operation includes a precipitation multiplier and a lapse rate adjustment), but these factors cannot make seasonal modifications or deal with inconsistencies. In order for the models to work properly, the input data should represent as closely as possible the scale and seasonal variations that actually occur in nature.

There is also a tendency for people to immediately start to process precipitation and temperature data after downloading the station information. It is very important to take the time for steps 2 and 3, i.e. assessing spatial variability and selecting flow points and period of records. An understanding of the variability of hydrologic and climatic conditions is essential to choosing the proper data analysis procedure. A careful determination of the flow points to be modeled and the period of record to use will avoid having to regenerate time series for different subdivisions of the river basin and different data periods during the calibration process. By carefully doing steps 2 and 3, a considerable amount of time and effort can be saved in the long run and the entire process made more successful and efficient.

The techniques used for data consistency checks and estimating missing data and the procedures

recommended for determining mean areal values and assigning station weights referred to in this chapter are not highly sophisticated. This chapter uses well established techniques and procedures based on engineering judgement. Most of the methods referred to in this chapter were devised by people primarily interested in hydrologic modeling. In order to use the models it is obviously necessary to get the data into the form needed for calibration and operational applications. The procedures described in this chapter for data analysis and processing are an engineering solution to the problem, not a scientific hydro-meteorological solution. The underlying objectives for coming up with these methods for preparing the data were to have estimates that reflect the proper scale of what actually occurs in nature and to be able to reasonably guarantee that there would be minimal bias between historical and operational estimates of data quantities. This last objective is critical to successfully using calibration results for operational applications. There is more discussion of potential sources of bias between historical and real time analyses in chapter 8.

This chapter provides a discussion of general topics and tasks that are important in the analyzing and processing of all types of historical data. Details for performing these tasks and recommendations on the steps to follow and procedures to use when working with individual data types are contained in the subsequent sections.

Precipitation, Temperature, and Evaporation

Effect of Data Bias on Model Results

The Sacramento and snow models are very sensitive to biased data. Thus it is very important to remove as much bias as possible when generating areal estimates of input variables. This not only includes overall bias as reflected in mean annual values, but also seasonal biases that can occur due to such things as variations in the precipitation versus elevation patterns and lapse rates. Again remember that the models are not designed to represent how variables like precipitation and temperature vary over the area. These variations need to be handled by the data analysis procedures.

Table 6-1. Change in runoff based on a change in precipitation

Watershed	Change in Precipitation	Change in Runoff
Leaf R. nr. Collins, MS	+10 %	+24.9 %
Bird Ck. nr. Sperry, OK	+10 %	+28.8 %
Smith R. nr. Bristol, NH	+10 %	+17.8 %
Animas R. at Durango, CO	+5 %	+9.7 %

Table 6-1 shows how a change in the overall amount of precipitation will effect the amount of runoff generated. As can be seen the percentage change in runoff is greater than the relative

change in the amount of precipitation. Typically the drier the region the greater the difference. It can be seen that a bias in the amount of precipitation will have a significant effect on the amount of runoff produced by the models. In order to compensate for data bias, either evaporation amounts will have to warped or model parameters, especially ones affecting the water balance such as tension water capacities, will have to be changed. Neither of these adjustments will produce the same results as if the precipitation data were unbiased. Also, evaporation demand curves and tension water capacities that should vary in an understandable pattern over the river basin will vary unrealistically if different amounts of precipitation bias exists from one watershed to another within the basin.

Figure 6-1shows the effect of changing the temperature by only 3°F on the timing of snowmelt for the Animas River at Durango, Colorado. This response is typical of a similar magnitude temperature change on other watersheds where snowmelt is significant. In this case the temperature for both the upper and lower elevation zones were changed by the same amount. As can be seen this relatively small change in temperature causes a fairly large shift in the timing of the snowmelt. If biased temperature data are used in a calibration, the non-rain melt factor parameters and parameters that affect the heat deficit and liquid water storage prior to melt could be changed to partly compensate for the biased temperature input, however, the results would not be nearly as good as if the data were properly corrected. In extreme cases one would be tempted to use a melt base (MBASE parameter) of other than 0°C in order to correct the timing of the snowmelt. The use of a non 0°C MBASE value almost always is an indicator of a bias problem with the temperature time series. Also, as with precipitation, one would find that the pattern of melt factors across the river basin would be unrealistic if temperature inputs were biased by different amounts from watershed to watershed.

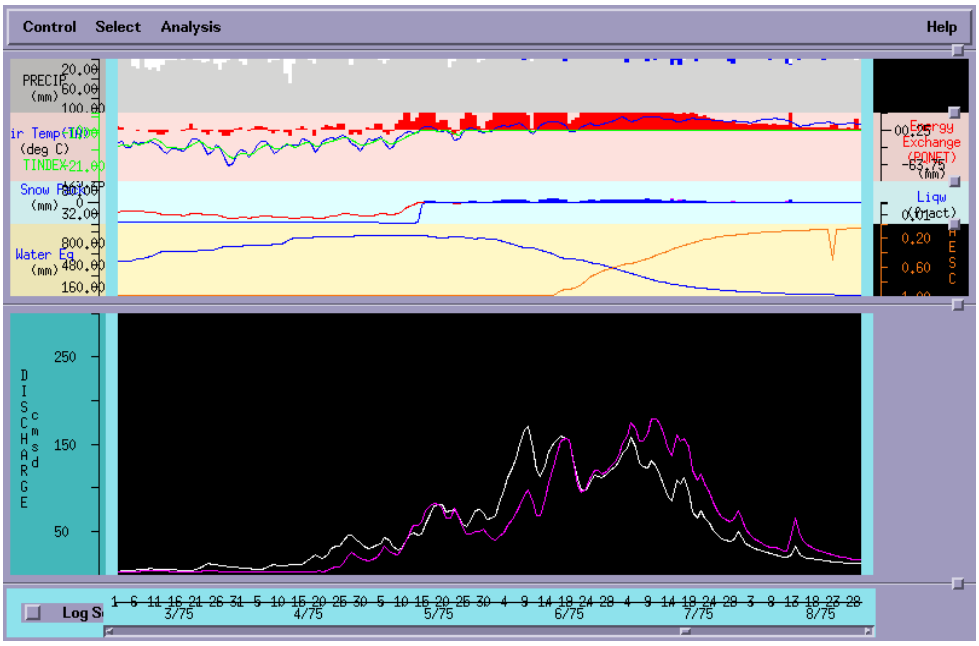


Figure 6-

1. Effect of a 3°F change in temperature - Animas River at Durango, CO

Biased evaporation estimates will also change the amount of runoff computed though the effect is not as great as with precipitation. Model runs on several basins indicated that the change in runoff due to modifying evaporation was about 40% of the change in runoff due to varying precipitation by a given percentage. Again, if such bias exists, it will cause model parameters to be distorted and the variations from watershed to watershed within the river basin will likely not follow expected patterns.

It is very important to carefully analyze the data during this step in the calibration process to minimize any data bias. This is most important in areas where the data values vary from location to location, typically mountainous areas. In such areas it is relatively easy to create biased input time series especially when extrapolating data to ungaged portions, such as high elevation zones.

Mountainous versus Non-Mountainous Area Procedures

Different techniques and procedures are recommended for use in analyzing and processing data based on whether the area is considered to be mountainous or non-mountainous. A non-mountainous area is defined as:

> *An area where the long term average values of the variable being analyzed are essentially the same at all locations within the area. This refers to both annual and seasonal averages.*

Typically for a non-mountainous area the mean values of the variable being considered should be within a range of about ±5% over the entire basin. There can be a general trend of increasing or decreasing average values as one moves across the area, but there should not be any rapid transitions. For non-mountainous areas the recommended techniques and procedures assume that any station can be used to estimate missing data at another station without making any adjustments for differences in magnitude and that areal averages can be computed by weighing stations merely as a function of their x,y plane location (sum of weights will always equal one).

A mountainous area is defined as:

> *An area where there are significant differences in the long term average values of the variable being analyzed over the area. These can be differences in annual or seasonal averages.*

Generally an area that needs to use mountainous area procedures will have terrain differences as these are the most common reason that long term means of a variable will differ from one location to another, however, areas with flat terrain can in some cases require the use of mountainous area methods. For example, long term averages of precipitation, temperature, or evaporation could vary significantly based on the distance from the coast or a large water body even though the terrain is generally level. In the mountainous area procedures, techniques for

estimating missing data account for the long term difference in the means between stations. Also, factors other than merely location are used to determine the station weights used to calculate mean areal values (sum of weights do not have to equal one).

It is very important to determine whether mountainous or non-mountainous area procedures should be used prior to analyzing and processing the data for model computations. As far as estimating missing data, if any portion of the river basin conforms to the mountainous area definition, then the entire river basin should use mountainous area techniques. When determining station weights, the use of mountainous versus non-mountainous area methods can vary from one area to another within the basin. To use non-mountainous area methods to determine station weights, all the stations that will be assigned weight for the watershed or subarea should have the same long term average value and all points within the boundaries should have the same mean.

Subdivision of Watersheds

When the flow points to be modeled are chosen, this defines the headwater and local areas for which data estimates need to be provided. However, in some cases these drainages need to be divided into subareas in order to properly model the variations that occur in such quantities as runoff, snow cover, and soil moisture. While areas could be subdivided in many different ways based on such factors as elevation, distance from gage, vegetation cover, soil type, and land use, there are two ways that are commonly used. The first is to divide the drainage into elevation zones and the second is to divide the area based on hydrograph response. Since the conceptual models are being applied in a lumped rather than distributed manner in this manual, the headwaters and local areas are divided into the minimum number of subareas needed to handle the effects of the spatial variability. Typically when subdivision is necessary only 2 zones are required, though in a few cases with extreme variations in conditions, 3 or even 4 zones may be required. From an operational standpoint, it has been found that the fewer the zones, the easier it is to manage real time adjustments to model states and computations.

Elevation zones are used in mountainous areas when the distribution of climatic variables and possibly physiographic factors create significant differences in the amount and timing of runoff. In the United States these differences are almost always related to varying amounts of snow cover and differences in the timing of snowmelt over the drainage. While these differences are not completely based on elevation, elevation is the dominant reason for the variations and elevation zones can easily be delineated. In most of the country when elevation zones are needed, two zones are sufficient with the models used in NWSRFS due to the inclusion of the areal depletion curve concept within the snow model. There are other snow models that divide a mountainous area into many elevation zones. These models don't use an areal depletion curve, but instead model areal cover based on which zones have snow and which are bare. However, even with the use of the depletion curve there are some areas that will require more than two elevation zones in order to get satisfactory results. These are typically drainages with large elevation variations or watersheds that are partly covered by glaciers. Also, if significant rainfall

events are common late in the melt season, additional zones may be needed. Typically there is more snowmelt than ET during much of this period so that the tension water storages in the soil model remain full, even though only a small portion of the area is covered by snow. In reality, when rain occurs late in the melt season, areas that have been bare of snow for some time have dried out and thus, the amount of runoff tends to be over computed. Adding additional zones can reduce this problem. Section 6-1 describes some criteria for determining when elevation zones are needed and guidelines for selecting the elevations between zones.

The second main reason for subdividing a headwater or local area into subareas is based on variations in hydrograph response due primarily to differences in the spatial distribution of rainfall. If the time to peak and/or the shape of the hydrograph can vary significantly based on where most of the rainfall occurs, then the drainage should possibly be subdivided so that these differences can be modeled. Besides differences in response, a precipitation gage network that can adequately define the spatial variation in the rainfall is necessary for the subdivided drainage to produce improved model results. One of the common cases for subdividing based on hydrograph response is when the shape of the drainage is long and narrow. In such a watershed, the response is normally quite different when most of the runoff comes from the upper end of the area as compared to a storm centered just above the gage. In this case the drainage would be subdivided by travel time. In other cases where the shape of the response varies, the subdivision might be based on the configuration of the channel network, e.g. one main branch of the stream network might be separated from another branch.

After making decisions as to whether any of the headwater and local areas need to be subdivided and, if so, delineating subareas based on elevation or drainage boundaries, the areas for which precipitation, temperature, and evaporation estimates must be computed are now defined. Also, at this point it should have been decided whether mountainous or non-mountainous area methods will be used to produce the areal estimates. The next step is to check the consistency of the data prior to computing the mean areal values.

Data Consistency Checks

The station data used to compute areal estimates of precipitation, temperature, and evaporation should first be checked for consistency and adjusted if necessary. By consistency we mean whether the data record is consistent over time, i.e. whether the record at a station maintains the same relationship over time to the other stations in the basin. Station consistency is checked by using a double mass analysis. In this technique the data for each station is plotted against the average of the data for a group of other stations. If the group is reasonably large, the average for the group is not significantly affected by inconsistencies in individual stations, but changes in the record for the station plotted against the group will be revealed. The main problem associated with consistency checks is separating real changes in the stations data record from natural variations that occur. When making consistency checks and adjustments it is very important not to remove the natural variability of the data. The underlying rule when making adjustments should be that if there is any doubt that the correction should be made, then don't make the

adjustment.

Inconsistencies in data records occur primarily for two reasons. First, the station is moved to a new location with a somewhat different climatic regime and a different exposure. This is the most common cause of an inconsistency. Second, there is a change that occurs at the site, such as an equipment change or a change in the exposure of the gage. This can include a new type of gage being installed, a wind shield installed or removed from a precipitation gage, a new building built near the site, or vegetation growing or removed from near the instrumentation. For many of the newer automated networks, moves and site changes are infrequent, thus inconsistencies in the data record are rare. Within the NCDC climatic network there are also stations that have remained in one place with the same equipment for many years, however, there are also stations that are moved periodically due to changes in observers. These are the stations that most frequently require adjustments for inconsistencies. Unfortunately, the records of station moves and equipment changes for the climatological network are not totally reliable. Comparisons made between different sources of station changes such as the Daily Climatological Bulletins, NCDC meta data files, B-44 forms filed by network managers, and records from state climatologists are not always consistent. Thus, changes may have occurred at a station even if the records you have available at the time do not indicate a move or equipment change. However, going along with the underlying rule mentioned earlier, if there is no documentation of a station change, don't make an adjustment unless it is very clear that a correction is needed.

The reason for making adjustments and removing inconsistencies prior to using the data to compute areal estimates of the input variables is to avoid having differences or bias between one portion of the historical record and another. If a reasonably sizeable inconsistency exists and is not removed and that station has a large weight in the computation of an areal estimate, then a significant bias will result from one portion of the areal estimate to another. As we have seen, the hydrologic models are quite sensitive to data bias. Model parameters determined using the record prior to the inconsistency will not be the same as parameters based on the period after. Results obtained using parameters determined from one period will be biased when those parameters are run on the other portion of the period of record. If the model parameters are based on a period that doesn't reflect the current status of the station, then operational results will be biased compared to those obtained during calibration.

Details on the procedures used to make consistency checks and adjustments and guidelines for using these procedures are contained in Section 6-2.

Computation of Mean Areal Values

Once the data have been checked for consistency and any corrections made, it is time to generate mean areal estimates of precipitation, temperature, and evaporation for use in model calibration. The analysis methods and techniques used vary as to whether an area is classified as mountainous or non-mountainous. The objective is the same in all cases, i.e. to obtain an estimate that is as unbiased as possible compared to what occurred in nature and to minimize the

random errors in those estimates. For precipitation and temperature the methods used in NWSRFS involve estimating all missing station data and then determining appropriate weights to assign to the stations to compute a mean areal estimate of the variable. This is also true for evaporation in some cases, but more frequently average climatic estimates of evaporation are used when calibrating the models. The recommended procedure for generating mean areal estimates of precipitation are in Section 6-3, temperature is in Section 6-4, and evaporation is discussed in Section 6-5.

Mean Daily Discharge Data

Possible Adjustments

Mean daily discharge data are used primarily to evaluate simulation results by comparing the model computed values to the observed data. For streams that do not have any man-made control structures, the model simulations can be compared directly against the observed natural flow in the river. When control structures exist, there are two options. The first is to model the operations of all of the controls so that a direct comparison can be made between the computed discharge and the observed controlled flow. The second is to adjust the observed daily discharge to remove the effects of the control structures so that these time series now reflect "observed" natural flow conditions. Prior to using the streamflow data, it is probably a good idea to check its consistency. Such a check could uncover variations in measurement techniques, but more likely will provide insight to alterations in runoff produced by such things as forest fires, changes in land use, different agricultural practices, variations in irrigated acreage, and changes to diversions which might otherwise not be noticed. Instantaneous streamflow data, when available, could also be adjusted for control structures, but this is more difficult and is usually not done. Section 6-6 describes possible checks and adjustments to mean daily discharge data.

Extension of Historical Data Record

As time goes by there likely will be a need to extend the historical data record. Reasons for extending the record were listed in chapter 2. When doing a data extension, it is very important to insure that the additional record is unbiased and consistent with the previous record. Since model parameters are based on the data used in the original historical analysis, the time series extensions must be unbiased as compared with the initial period used for calibration. When generating extensions, it is critical to do the following:

- New data for stations that were included previously should be checked for consistency with the prior data for these stations and adjusted if necessary (this typically requires that station data for some period prior to the extension, typically about 10 years, be available to use when generating consistency check plots - any consistency adjustments that were used for these stations for this period should be applied before producing the plots). Also, data for any new stations that are being added to the analysis should obviously be check for consistency.

- Station and areal means for precipitation, temperature, and evaporation that were determined for the original complete historical data analysis should continue to be used until another complete data analysis is performed (e.g. if 30 years of data were used when the river basin was first calibrated, and later the record was extended by 10 years, and now it is being extended by another 10 years, the means for the original 30 years of data should be used when processing this latest extension), and

- Mean values for new stations need to be determine based on their relationship to stations with a long historical record that were used in the original analysis using the techniques included in PXPP for precipitation (uses the average ratio of precipitation at each new station to the amount recorded at a well established station) or as outlined in Section 6-4 for temperature data (uses the average difference between each new station and a station with a long record). Typically, based on experience, about 5 years of data are needed at a new station in order to compute a stable estimate of the mean values. Less data can be used, but the estimate of the mean values will not be as reliable.

Historical data extensions are possible when the types of measurements, processing methods, and networks are the same as used in the original analysis. When new types of measurements (e.g. radar based estimates of precipitation), new processing methods (e.g. switching from non-mountainous area to mountainous area procedures), or significant changes in the network (e.g. the addition of a high elevation gage network in an area where previously there were low elevation stations) occur, the historical data analysis should generally be completely redone. In some of these situations it may be possible to produce values that are consistent and unbiased compared to the previous analysis and thus can be used to extend the existing record, but this is difficult to accomplish. In addition, for many of these cases the period of the historical data record that can be used will change since the new measurements or network are only available for a limited time. Unless the new data can be proven to be unbiased and consistent with the previous record, the models must be recalibrated so that the parameters are compatible with the new data. This new historical data period then becomes the base period to use for any future extensions of the record.

Section 6-1

Determination of Elevation Zones

Introduction

Elevation zones are used in mountainous areas primarily to model differences in the amount of snow that occurs and the timing of the resulting snowmelt. A watershed could be divided into elevation zones solely based on variations in the amount of rainfall and resulting runoff, but this seldom is the reason for using elevation zones in the United States. As far as what each elevation zone represents and methods for selecting the separating elevations, the United States can be divided into 2 general regions for discussion purposes. First is the intermountain west, i.e. the area east of the Pacific crest over to the plains, in other words, the Rocky Mountains and the east side of the Sierra Nevada and Cascade ranges. Basins in this region are typically dominated by snowmelt runoff and elevation zones are primarily used to separate the portion of the watershed that contributes significant runoff every year from those that normally don't produce much runoff except during an exceptionally large snow year. Second are mountainous regions where runoff is typically generated at all elevation zones and rainfall is a significant factor in runoff production. This includes a few areas in the eastern U.S., primarily in the northeast, the Pacific coastal drainages, and much of Alaska.

Intermountain West

In much of this region the annual hydrograph, at least from those watersheds that produce enough runoff to generate satisfactory or at least marginal results with a lumped, conceptual model, generally shows a large rise beginning in the spring that is the result of snowmelt and only a few responses at other times of the year caused by rainfall. Most basins in this region produce significant amounts of baseflow, thus there is typically a long recession following the snowmelt period. Rainfall in this region infrequently produces significant amounts of runoff over large areas (as opposed to small scale thunderstorms). Such events occur primarily in the northern part of the region due to moisture coming over the mountains from the Pacific northwest and in the southern part of the region due to moisture moving up from the south into Arizona, New Mexico, and southwestern Colorado. The primary reason for using elevation zones in this region is to separate the portion of each watershed that produces significant runoff every year from the part that generates runoff only during years with large snow accumulations. Throughout this region two elevation zones should be sufficient.

A method for selecting the elevation at which to divide the watersheds in this region is based on areal snow cover information and hydrograph response. The following data are needed:

1. Areal snow cover images at various dates just prior to and throughout the snowmelt period. These images are produced by the NOHRSC from analysis of satellite data. The images should be from a number of years with varying amounts of snow cover, though years with

below normal snow amounts are of most value.

2. Mean daily flow hydrographs for the watershed for the same years as the areal snow cover data.

The steps to follow are:

1. For each date with an areal snow cover image, determine the average snowline elevation over the watershed. Typically this elevation is similar for all watersheds within the river basin, but differences can occur due to factors that affect the accumulation and melt processes, such as slope, aspect, vegetative cover, and prevailing storm direction. The CAP should contain a method to access these images and determine the snowline elevation.

2. Tabulate the snowline elevations on the hydrograph plots for the watershed for each date. The snowline will typically be lowest prior to the onset of melt, but will also move down the mountains after periods of snowfall during the melt season

3. By examining all the years, select the elevation above which snow exists just prior to the beginning of significant runoff during almost all years. This is the elevation to use to separate the zones for modeling. During years with considerable snow and above normal runoff, the snowline at the onset of significant runoff will be lower than that during below normal runoff years. What one is looking for is the highest snowline elevation just prior to the beginning of significant runoff during all the years or, in other words, the elevation above which snowmelt always produces a rise in the hydrograph. This elevation is usually determined by looking at some of the below normal runoff years and possibly averaging the snowline elevations prior to significant runoff for these years. If there is a very low snow year, i.e. runoff far below normal, it is probably best to ignore that year. When later modeling such a low snow year hopefully the snowmelt runoff volume will be reasonable, though the timing of melt will be too early since the snow during that year is really only in the upper part of the upper zone.

Figure 6-1-1 shows an example of a below normal runoff year for the Animas River above Durango, Colorado which could be used to determine the elevation separating the upper and lower zones. In this case, the snowline was at about 10,000 feet just prior to the onset of significant runoff. The use of this zonal split elevation resulted in good simulation results. Figure 6-1-2 shows the snowline elevations for an above average snow year for the same basin. It can be seen that significant runoff was generated due to snowmelt at elevations below 10,000 feet. By splitting the watershed at 10,000 enabled the models to properly simulate both of these years. In water year 1981 very little runoff was generated from snowmelt in the lower zone while in 1985 this zone produced significant runoff. The upper zone produced runoff during both years. Figure 6-1-3 shows simulation results for the lowest snow year on record, 1977. The snowline elevation prior to melt during this year was over 11,000 feet and was significantly higher than for other low snow years and thus not used when deciding where to

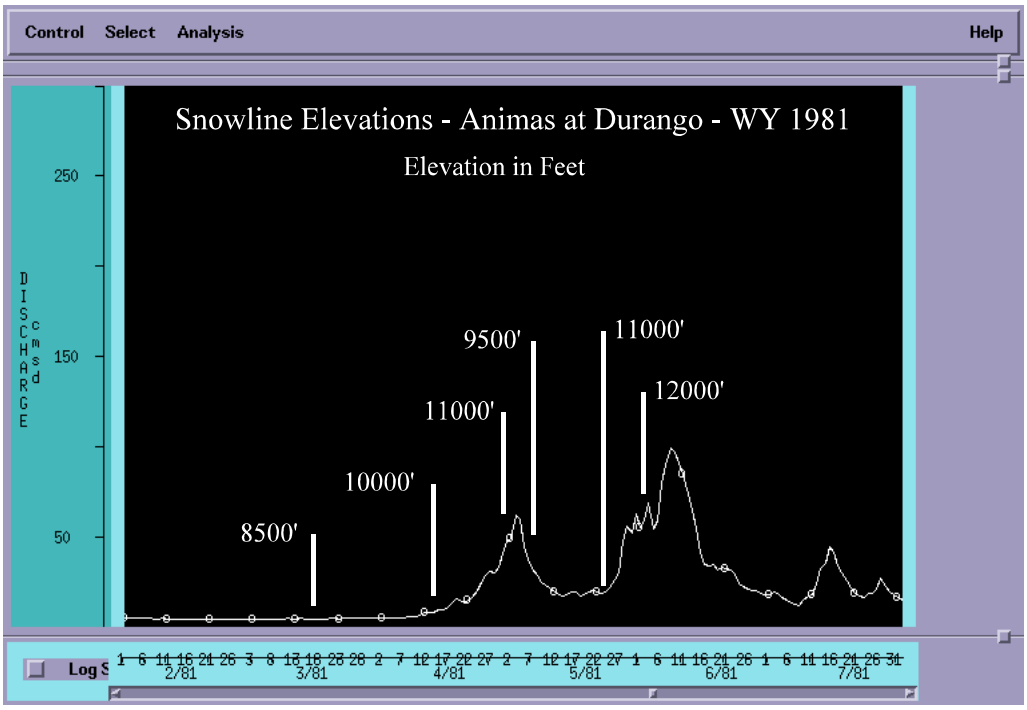


Figure 6-1-1. Snowline elevations during a below normal runoff year.

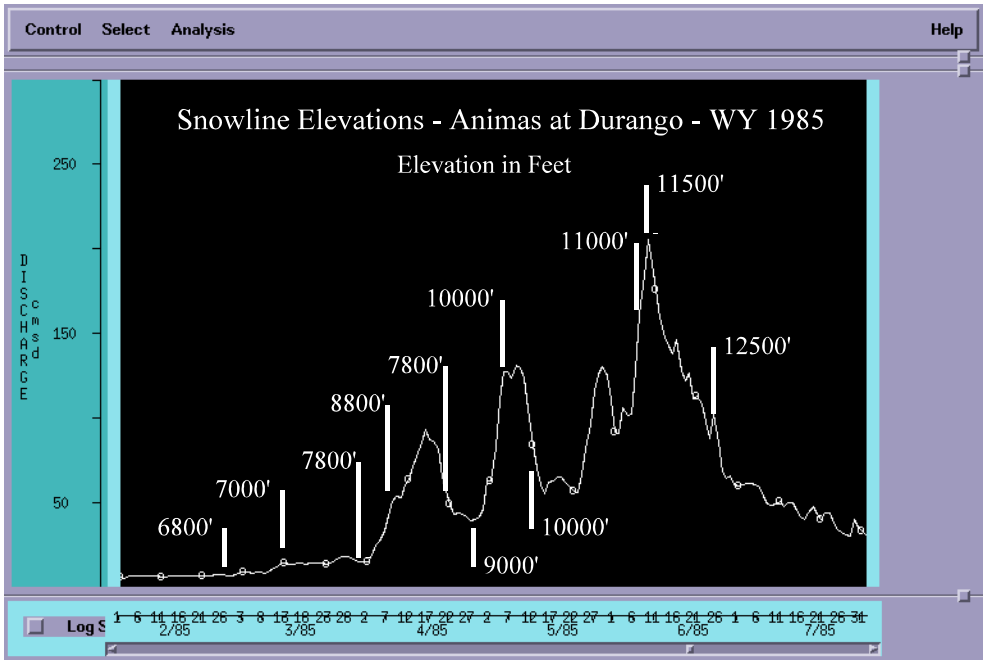


Figure 6-1-2. Snowline elevations during an above normal runoff year.

divide the two zones. The snowmelt runoff volume for this year was simulated quite reasonably,

but as expected the timing of the melt is early. Three elevation zones could be used to better simulate such a small year, but generally the added benefits don't justify the added work of operationally maintaining the state variables for an extra zone.

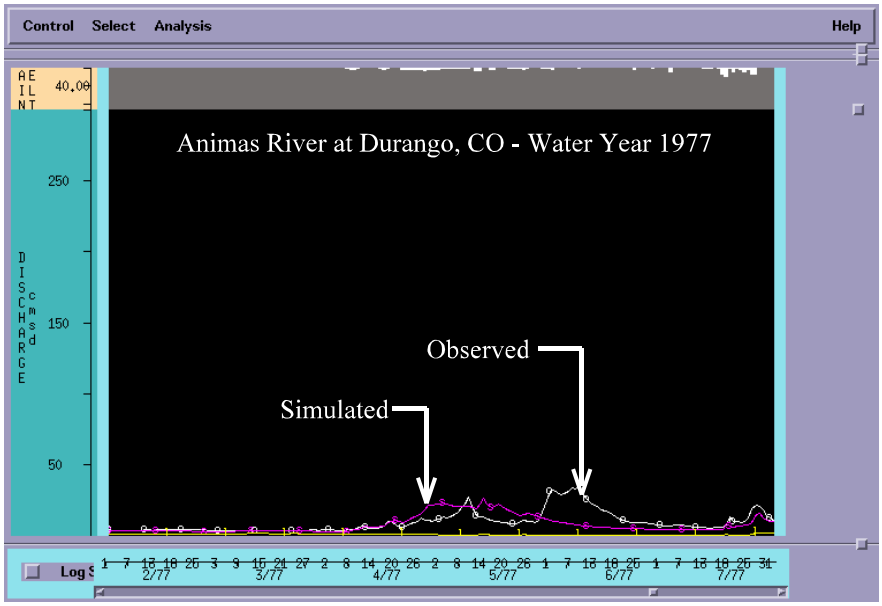


Figure 6-1-3. Typical response during the simulation of a very low snow year.

In addition to areal snow cover data and the response of the hydrograph, vegetation information can be very helpful in selecting the elevation separating the 2 elevation zones in the intermountain west. The type of vegetation that grows at a particular location is highly dependent on the prevailing moisture supply. Thus, in many cases there will be a transition from one type of vegetation to another at the elevation separating where runoff occurs almost every year and where it only occurs during very wet years. In much of the intermountain west this is the elevation where the transition from pines to a predominately fir-spruce forest occurs. In the southern part of the region, e.g. the Mogollon Rim area of Arizona, there is only a small portion of the area that produces significant snow runoff every year and has predominately fir-spruce forests, thus it may be better to split the watersheds at a slightly lower elevation where the transition from pinon pine-juniper to ponderosa pine forests occur. CAP should contain information on the types of forest cover and the distribution of vegetative types with elevation.

Northeast, Pacific Coast, and Alaska

In the northeastern part of the United States, the Pacific Coast drainages which primarily involve the west side of the Sierra Nevada and Cascade ranges, and Alaska, significant runoff is generated at all elevations, but the amount and timing of snowmelt runoff varies considerably with elevation. During most storm events precipitation occurs at all elevations, but in many

cases there is a transition from rain at the lower elevations to snow at the higher locations. Thus, in these regions elevation zones are needed to properly model the accumulation and ablation of the snow cover throughout the watersheds. The general breakdown in zones that you are trying to achieve in these regions are:

- zone where there is a significant snow cover during all years,
- zone where a significant snow cover exists only during substantial parts of some years,
- zone where snow seldom occurs and any snow cover exists for only short periods,
- zone that is permanently covered by snow and ice, i.e. glaciated areas which occur in some Pacific northwest basins and frequently in Alaska.

In the northeastern United States there are only a few areas with a significant range in elevation to justify using multiple zones. This primarily occurs in the White Mountain region of New Hampshire and possibly in a few watersheds in Maine, the Andirondack region in New York, and the Allegany highlands of West Virginia. In the southern Appalachian Mountains there are significant elevation differences, but snow is typically hydrologically unimportant. In the northeast 2 zones are sufficient when they are needed at all. Along the Pacific Coast and over portions of Alaska only 2 of these elevation zones comprise the vast majority of the drainage area and thus are sufficient for modeling. In glaciated basins in Alaska and fairly frequently in other parts of these regions, however, 3 or even all 4 zones exist and need to be included. In some of the watersheds a snow cover exists over the entire drainage every winter, but there is still a significant difference in the amount and timing of snowmelt with elevation. This is especially true in Alaska. In these cases, 2 elevation zones are still needed to model the seasonal snow cover and its contribution to runoff. If glaciers are present in these watersheds, then a third zone would be required.

The elevations used to separate the elevation zones in this region are primarily determined by examining areal snow cover information from NOHRSC and snow depth and water equivalent data at various locations. These data can be examined for a number of years and a subjective determination made as to what elevations to use to separate the zones. The elevations selected should be able to be used over a fairly wide area. Along the Pacific Coast the elevations separating the zones should likely decrease as one goes further north as average temperatures generally decrease and thus snow is more likely to accumulate at lower elevations. In Alaskan basins where there is substantial snow cover over the entire drainage every year, the tree line elevation may be a reasonable place to separate the zones.

Section 6-2

Consistency Checks for Precipitation, Temperature, and Evaporation

Basic Method

Consistency of precipitation, temperature, and evaporation data from individual stations is checked using a double mass analysis. NWSRFS contains consistency check options within each of the programs that are available for processing these data types, i.e. the PXPP and MAP programs for precipitation, the MAT program for temperature, and the MAPE program for evaporation. In addition a graphical user interface program, IDMA (Interactive Double Mass Analysis), is available to display the consistency plots generated by the processing programs and to interactively make consistency corrections. When using these processing programs the first step is to check the consistency of the data and make any adjustments needed prior to computing mean areal values for use in model calibration.

When using a double mass analysis technique it is essential to include enough stations so that inconsistencies in individual stations can't have a significant effect on the consistency of the computed group average. Typically when performing the data analysis for an entire river basin or on a regional basis, there are more than enough stations, however, in case the technique is being applied to a smaller network it is recommended that at least a minimum of 10 stations be included.

For precipitation and evaporation the consistency checks involve plotting the data for one station against the average of the data values from a large group of stations. This is because the relationship between precipitation and evaporation stations is normally expressed as a ratio, i.e. it is said that one station typically has a certain fraction of the value observed at another station, the average of a group of stations, or an areal averaged value. The accumulated value for the station is plotted against the accumulated value of the group average. In NWSRFS in order to more clearly see changes in the relationship between one station and the group, the deviation of the accumulated station average from the accumulated group average is plotted against the accumulation of the group average. The group average is computed based on all the stations in the group minus the station being plotted against the group. Thus, the accumulated group average generally changes slightly from one station to another and therefore there is not a direct relationship between the accumulated group average and a specific date in time.

For temperature the consistency check involves plotting the deviation between the accumulated average temperature for a single station and the accumulated average temperature for a large group of stations against time. This is because the relationship between temperature stations is normally expressed as a difference, i.e. it is said that one location is a certain amount colder or warmer than other locations on the average. Separate plots are generated for maximum and minimum temperatures.

For all these data types the record for the individual station is consistent over time if the relationship between it and the group plots generally as a straight line. However, there is a lot of variability in nature such that the relationship between an single station and a group of stations will not be perfectly straight. These random variations in the relationship will cause the plotted lines for each station to wobble. As long as the general trend of line is straight and there are not significant deviations from the general trend line, the data are consistent. Inconsistencies are indicated by changes in the slope of the line that are of a significant magnitude and last for a reasonably long period of time to be caused by something other than the natural variability in the data fields. The trick is to be able to recognize inconsistencies and adjust the appropriate data periods without removing the natural variations that exist in the data. Precipitation data typically are much more spatially variable than temperature and evaporation, thus consistency plots for precipitation generally exhibit more wobble than plots for the other variables.

NWSRFS Program Options

The NWSRFS programs with consistency checks contain options for specifying the number of groups, the stations assigned to each group, the stations to be included on each plot, and whether to do precipitation checks on a seasonal basis.

Generally all the available stations within the river basin are included in a single group. In this case each station is then plotted against the average for all the other stations. NWSRFS does contain the option to have multiple groups. It was initially thought that it would be better in mountainous areas not to put all the stations in one group, but instead group the data by some factor such as elevation or mean annual value. This feature still exists, but experience has shown that typically nothing is really gained by trying to use multiple groups when making consistency checks. Also, generally all the stations in the group are used to compute the group average even though the option exists to remove some stations from this calculation (option not in PXPP). This option is seldom needed since normally there are enough stations involved in the group that the effect of any one station is minimal. If the data for a station are so questionable that it should not be part of the group average, the station probably shouldn't be included in the analysis in the first place.

The consistency plots generated by the NWSRFS data processing programs display a maximum of 5 stations per plot. When a large number of stations are being included for a river basin, it is very important as to how the stations are grouped for plotting, especially in the case of precipitation. As indicated earlier, inconsistencies typically show as a change in the slope of the double mass plot that persists for a reasonably long period of time. For precipitation data such changes in slope can also occur due to shifts in storm tracks or storm types over an area like a river basin and can persist for a number of years. For example, changes in storm tracks can cause one side of the basin to catch more precipitation relative to the other side than normal for a period of several years or changes in prevailing storm types can change the relationship between high and low elevation stations for a period of time. Thus, the selection of which stations are included on each plot is important. Selection should be made so that stations in the same

geographical part of the basin or with similar elevations are displayed together. When this is done, then real shifts that occur over the basin due to climatic changes can be recognized and not inferred to be inconsistencies in the data. When many of the stations in the same portion of the basin or at similar elevations exhibit a change in the slope of their consistency plots at the same point in time, this is an indication that this is a real shift that occurred in nature and should not be adjusted. When only a single station exhibits a change in slope, then it may indicate an inconsistency. Temperature and evaporation stations should also be grouped by location or elevation for plotting purposes though such pattern shifts are less likely to occur with these data.

With precipitation data the option exists to perform consistency checks on a seasonal rather than an annual basis, i.e. separate plots are generated for the winter and summer seasons. The decision to do consistency checks on a seasonal basis is independent of whether station weights are specified on a seasonal basis though the NWSRFS programs do force one to use the same definition of the seasons for both options. Consistency checks for precipitation should be done on a seasonal basis when snowfall dominates some months during the winter. This is because the effect of gage exposure or equipment changes is generally much larger when the precipitation is in the form of snow rather than rain. Thus, if a station is moved or a wind shield added, the existence and magnitude of an inconsistency are highly dependent on the form of the precipitation. The NWSRFS programs allow only for the separation of precipitation into different consistency plots based on month of the year, not on an event or time interval basis. Thus, the winter season should include those months when snowfall predominates and the summer season when the majority of the precipitation is rain. Seasonal precipitation consistency plots should be used whenever snowfall is significant over the river basin.

Guidelines for Making Consistency Corrections

There are a number of factors to consider when analyzing the consistency plots and making corrections for apparent inconsistencies in the data.

- Consistency corrections can be applied only to periods with observed data, not to periods of estimated values. Thus, it is first important to delineate those periods that contain missing values and are thus estimated from those periods with observed data. The PXPP program works only with monthly precipitation totals. A month is treated as missing and thus estimated whenever any data are missing during the month. The other programs estimate only those days or hours that are missing, thus a month can contain a mix of observed and estimated data. The IDMA program tries to designate when data are observed and when the values are estimated by changes in the color of the plotted line. Estimated data periods may not be consistent with the observed data periods for some stations, i.e. the slopes of the plots during these periods may vary. However, once the inconsistencies have been corrected for all the stations, the estimated data periods for all stations should plot at the same slope as the observed data periods, i.e. the trend of the entire plot should be straight.

- Inconsistencies should occur only when a station is moved or there is an equipment change

that can affect the measurement. Thus, it is important to have the meta data that indicates the type and time of modifications to a station. Actual moves and equipment changes, such as adding a wind shield or converting from a weighing gage to a heated tipping bucket gage at a precipitation observation site, should be marked on the plots as possible times when inconsistencies can occur. Inconsistencies seldom occur in any network that is stable, i.e. there are no site relocations or equipment changes. The network that is most susceptible to inconsistencies is the national climatological network maintained by NCDC. This network relies on volunteer observers for the most part and station relocations occur quite frequently. In some cases when a station is relocated, a new station number and name are assigned and thus you have a new record. In other cases when a move occurs, the station number remains the same, though the name will be modified by changing the distance and direction from the nearest town, e.g. ANYTOWN 1N may be changed to ANYTOWN 5SW. When a move occurs there are not clear rules as to when a new station is established and when the number remains the same and just the name is modified. Sometimes a site can be moved only a couple miles and a new number is assigned, while in other cases a station can be moved 15-20 miles and the number remains the same. In one case a station was not moved at all, but the number was changed when the two word name was reversed thus changing is alphabetical position. Frequently when a station is discontinued and a new station is established just a short distance away, the records are merged prior to being used in the data analysis programs. Obviously the time of this move should be examined for a possible inconsistency in the record.

• Data should generally be made consistent with the current location of the instrumentation. This is especially true for a real time reporting site. However, if the data for the current location seems unreasonable when compared to nearby gages, then either the data should be made consistent with another portion of the period of record or the data for the current location should be ignored by setting it to missing. When using climatic stations that have been discontinued and are no longer in use, the data can be made consistent with any portion of the period of record that you choose. If a correction is applied to the current location of a real time reporting station, this correction should also be applied in OFS.

• For temperature data the mean monthly max and min temperatures at each station are used to estimate missing data. These mean monthly values are computed from the observed data for the period of record being used. When corrections are applied to portions of the observed record, the mean monthly values for that station are altered. For precipitation the computation of monthly means is done automatically within the PXPP program, however, no preliminary processing program currently exists for temperature, thus the mean monthly temperatures must be manually adjusted whenever a consistency correction is applied to a station. The MAT program contains the option to compute and output the new monthly means, but the user must manually take this output and use it to edit the input file. This should be done after every run that involves new consistency corrections. Since the revised monthly means are not computed until after the corrections are input, the program is typically using the monthly means from the previous run which are not correct whenever new

adjustments have been entered. Thus, it is necessary at the end to make one final run with no new adjustments

• Consistency plots will show random wobbles due to the natural variability of the data. Precipitation data will generally show more of this variability than temperature or evaporation. Consistency plots can also exhibit a seasonal wobble caused by variations in the relationships between stations, especially due to elevation. Especially in the intermountain west the precipitation relationship between high and low elevation sites varies considerably from one time of the year to another. There is also typically a seasonal variation in lapse rates for temperature. The extreme case is in portions of Alaska where temperature decreases with elevation generally in the summer, but in the winter inversion conditions can persist even for maximum temperatures at interior locations during the months with the least sunlight. These effects can cause natural seasonal wobbles in the consistency plots.

• Sudden jumps in a consistency plot for a station indicate that there may be some bad data in the record. In the case of precipitation, the jump could occur because that station received much more precipitation during a given month than any of the other stations, typically from an intense thunderstorm directly over the gage. However, in most cases, discontinuities in the plots are due to improper values being entered into the data record for a station. The bad data at one station can cause bad estimates at nearby stations therefore causing jumps in their plots also. A little detective work should find the culprit. There have been some problems, at least in the past, with hourly precipitation data not being able to be decoded properly, resulting in a number of months of data that should be missing being set to all zeros. If the problem period is long enough, this should cause a shift in the slope of the consistency plot. Such a period will not respond to a correction factor since zero times any factor is still zero.

• There is a greater chance of inconsistencies occurring in mountainous areas than in flat terrain. This is because in mountainous areas the amount of normal precipitation can vary by a considerable amount over a fairly small area due to orographic effects. Thus, when a station is relocated there is good chance that the average amount of precipitation being caught will change. There is also a greater chance for inconsistencies to occur in regions with substantial snowfall than in regions where most or all of the precipitation is rain. This is because snowfall catch is greatly affected by the exposure of the gage to wind. Station relocations will frequently change the exposure of the gage. In general, one will find a greater need to make consistency corrections in mountainous areas with substantial snowfall where there are periodic site relocations. In a region of flat terrain with only rainfall, inconsistencies should seldom occur.

• The aim is not to see how many “inconsistencies” one can find, but to preserve the variability that occurs in nature and correct only those periods that clearly need to be corrected. It is also not rational to go to the other extreme and assume that consistency adjustments will just distort the observed data and therefore avoid making consistency checks. There are clearly times when the measurements at a station are modified by a

relocation, equipment change, or alteration of the site surroundings. These changes to the observed data need to be adjusted so that model parameters can be reliably determined during calibration and that these parameters will be applicable for forecast applications. One should adjust those periods that clearly need to be corrected, but if there is any doubt as to whether an adjustment is needed, it is best to not make that correction.

By taking all of these factors into account, one should be able to adjust the data record for any significant inconsistencies without removing the variability that occurs in nature.

Mechanics of Computing Consistency Corrections

Even though the IDMA program will compute consistency corrections for you, it is a good idea to understand how the corrections are calculated. The calculations can be done in several ways. Following is one method.

1. For each station determine the line segment that the data should be made consistent with. As mentioned earlier this generally corresponds to the most recent location of the gage.

2. Draw a line that is parallel to the line segment from step1 that originates at the beginning of the period of observed data for the station. This is referred to as the "Corrected Line". Note that this line will start after the beginning of the period being analyzed when measurements aren't initiated for a given station until a later date, thus the data for the first part of the analysis period are estimated.

3. Compute corrections as:

Precipitation or evaporation:

$$A_{t,t'} = \frac{\left(S_t - S_{t'}\right) + \left(C_t - C_{t'}\right)}{\left(S_t - S_{t'}\right)} \qquad (6\text{-}2\text{-}1)$$

Temperature:

$$A_{t,t'} = \frac{\left(C_t - C_{t'}\right)}{\Delta N_{t,t'}} \qquad (6\text{-}2\text{-}2)$$

where: A = Adjustment, precipitation or evaporation (multiplying factor), temperature (degrees/month)
S = Accumulated precipitation or evaporation
C = Magnitude of correction (difference between current location and Corrected Line; positive if increase, negative if decrease)
ΔN = number of months

t’ = start time of adjustment
t = end time of adjustment

Figures 6-2-1 and 6-2-2 illustrate the computation of the consistency factors. These figures show how all the corrections for the stations can be computed at one time, however, when first learning to adjust data for inconsistencies, it is probably easiest to work with one station at a time. It is probably also best in this situation to make and verify one correction at a time, beginning with the earliest.

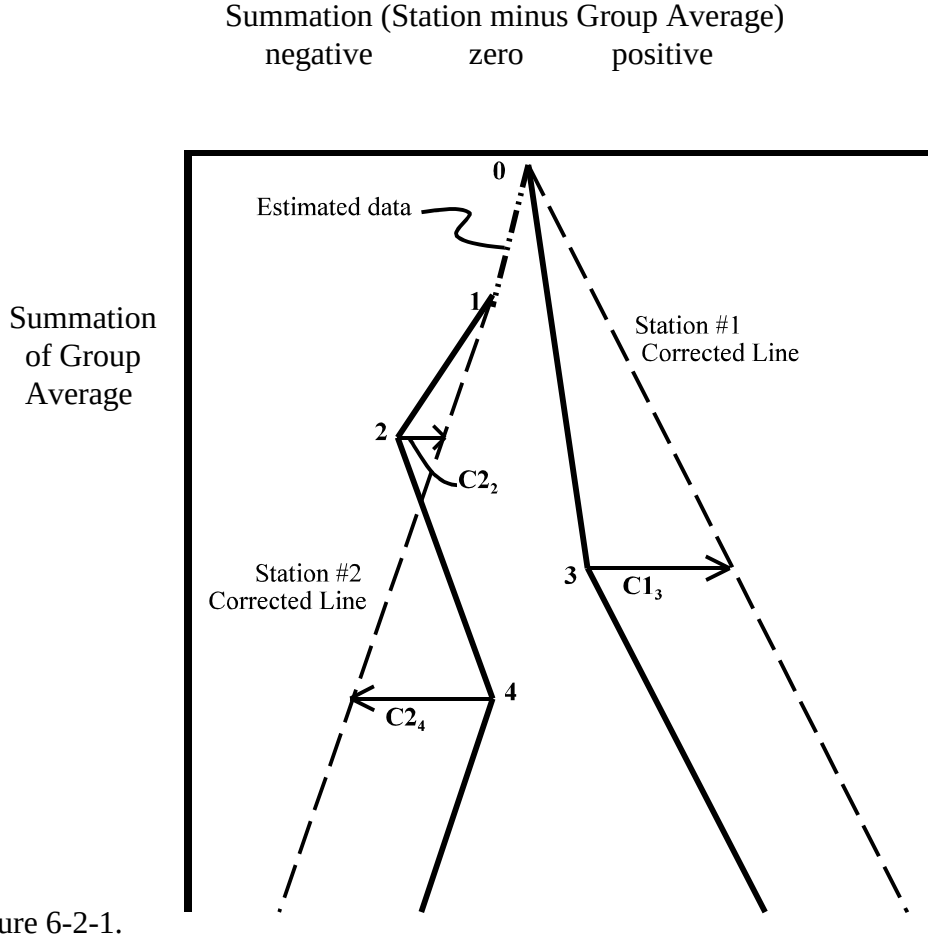


Figure 6-2-1.
Illustration of precipitation or evaporation consistency adjustments

For figure 6-2-1 the adjustments would be computed as:

For station 1:

$$A1_{3,0} = \frac{S1_3 + C1_3}{S1_3}$$ (greater than 1.0 since $C1_3$ is positive)

If the data record begins with October 1948 and the inconsistency for station 1 occurs in May 1975 and if $S1_3 = 860$ inches (i.e. accumulated precipitation for station 1, not the group average -- when using the PXPP, MAP, or MAPE programs this value is obtained from the tables that accompany the consistency plots using the date corresponding to when the inconsistency occurred) and $C1_3 = 260$ inches (computed from the plots using the station minus group average summation scale), then the adjustment is 1.30. This adjustment is applied starting in October 1948 and remains in effect through April 1975. In May 1975 the adjustment factor needs to be set back to 1.0 since the remaining data plot parallel to the Corrected Line.

For station 2:

$$A2_{2.1} = \frac{\left(S2_2 - S2_1\right) + C2_2}{S2_2 - S2_1}$$ (greater than 1.0 since $C2_2$ is positive)

$$A2_{4,2} = \frac{\left(S2_4 - S2_2\right) + \left(C2_4 - C2_2\right)}{S2_4 - S2_2}$$ (less than 1.0 since $C2_4$ and $-C2_2$ are both negative)

For figure 6-2-2 the adjustments would be computed as:

For station 1:

$$A1_{2,0} = \frac{C1_2}{N_2 - N_0}$$ (adjustment is negative since $C1_2$ is negative)

If the data record begins in October 1977 and the inconsistency for station 1 occurs in June 1984 (i.e. 80 months after the start of the observed record) and if $C1_2 = -120$ °F (computed from the plot using the deviation of accumulated means scale), then $A1_{2,0} = -1.5$ °F/month. This adjustment is applied beginning in October 1977. The adjustment is then set back to 0.0 in June 1984.

For station 2:

$$A2_{3,1} = \frac{C2_3}{N_3 - N_1}$$ (adjustment is positive since $C2_3$ is positive)

$$A2_{4,3} = \frac{C2_4 - C2_3}{N_4 - N_3}$$ (adjustment is negative since $C2_4$ and $-C2_3$ are both negative)

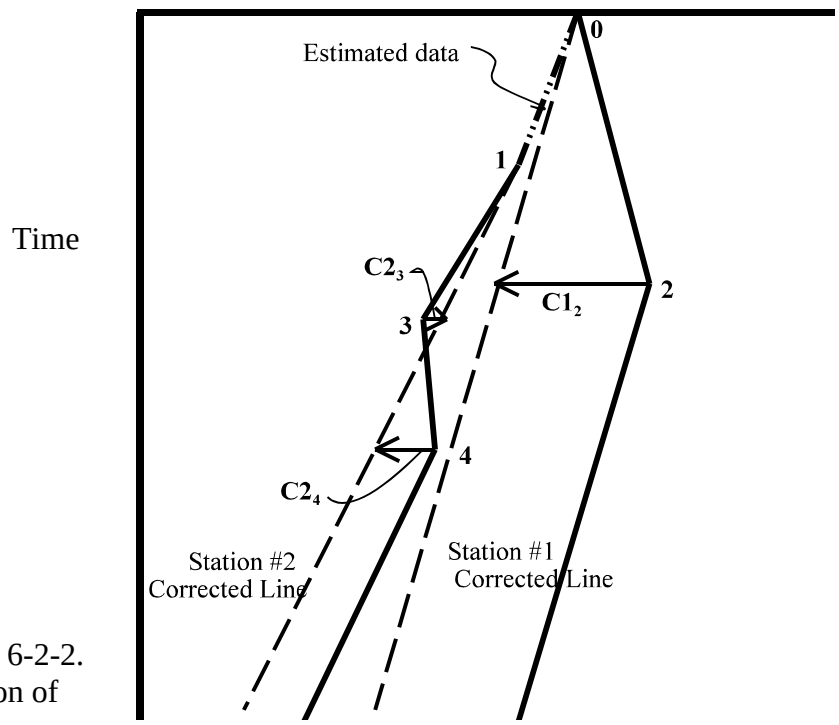


Figure 6-2-2. Illustration of consistency adjustments for temperature

Section 6-3

Computation of Mean Areal Precipitation

Introduction

Good estimates of precipitation are critical to the successful use of conceptual models. The precipitation estimates should first of all conform to the physical scale of what actually occurs in nature, i.e. the estimates should be unbiased compared to what actually occurs. Second, other statistical properties of the estimates, such as frequency of occurrence of various amounts and seasonal variability, should also be as realistic as possible. Third, the amount of random error in the estimates should be as small as possible. The random error is controlled to a large degree by the data network and spatial variability of precipitation over the region. In sections of the country where there is a high spatial correlation between precipitation amounts the typical climatological network can produce a fairly reliable areal estimates, whereas in areas with little spatial correlation the estimates will contain a large amount of random error. A large amount of noise in the precipitation input makes it very difficult to determine model parameters and obtain good simulation results. The quality of the areal precipitation estimates largely determines where lumped, conceptual models can be applied with satisfactory results (see Figure 1-1 in chapter 1).

Prior to doing the steps discussed in this section it is assumed that all the precipitation data are available in the proper form and that the data have been checked for consistency and adjusted if necessary. It is also assumed that the areas where MAP estimates are to be generated have been determined based on the definition of the flow points needed for calibration and the possible subdivision of the drainage areas into elevation zones or subareas specified by other means. It is also assumed that it has been determined as to whether mountainous or non-mountainous area procedures should be used to compute areal estimates.

Ideally precipitation data should be corrected for gage catch deficiencies prior to computing areal estimates for use in conceptual models. Catch deficiencies are a function of the type of precipitation, the type of gage, and the amount of wind at the orifice of the gage. The catch deficiencies for snow are much greater than for rain. The wind speed at the gage orifice is controlled by the current atmospheric wind speed and direction, the gage exposure, and the type of wind shield, if any. The catch deficiency varies throughout an event as meteorological conditions change. Some techniques have been proposed for applying monthly adjustments to precipitation data at a given station based on wind and temperature data from nearby synoptic sites and a general classification of the gage exposure including whether a wind shield is installed. Whether the application of such techniques would improve model simulation results has not be shown. In non-mountainous areas gage catch deficiencies are typically ignored when computing MAP. An average catch deficiency for snow events is later incorporated in the snow model computations (e.g. the SCF parameter in the SNOW-17 model). In mountainous areas some adjustment for the average catch deficiency can be implicitly included when water balance computations are used to obtain average areal precipitation estimates. Further adjustment is

likely when the models are calibrated (see discussions under the SCF parameter in sections 7-4 and 7-7).

Time Interval for Precipitation Computations

The Mean Areal Precipitation (MAP) processing program in NWSRFS used for calibration will allow for the generation of time series at a 1,3, and 6 hour interval. In most cases a 6 hour interval should be used. There are several reasons for using a 6 hour time interval when generating MAP time series from historical raingage data. First, the program used operationally in NWSRFS to generate MAP time series from raingage reports currently will only generate time series at a 6 hour time interval. Second, in most regions there is not a dense enough historical hourly raingage network to produce reliable areal estimates at less than a 6 hour interval. Third, since some of the models, especially the Sacramento Model, will produce different results depending on the time scale that they are applied, the calibration of the model and the operational application should use the same time interval.

Historically for river forecasting at an RFC a 6 hour time interval has been used for model computations. This has typically been the case for several reasons. First, real time data, especially precipitation, was only available on a 6 hour basis at a reasonable network density. Second, the time needed operationally to obtain and process data, perform model computations and make manual adjustments, and get the forecasts to the users limited the time between forecast updates to around 6 hours. Third, RFCs were not opened on a 24 hour basis except during times of significant flooding potential. Thus, when the current OFS MAP program was coded in the 1980's a 6 hour time interval was specified. The use of a 6 hour computational interval has also tended to limit the size of the basins that are modeled by the RFCs to those that can be modeled successfully at a 6 hour interval, i.e. basins that typically peak in 12 or more hours. Drainages with faster response times have been handled using site specific and regional flash flood procedures by the WFO or via local flood warning systems such as ALERT and IFLOWS. The WFO and local flood warning procedures seldom involve continuous modeling and rely on previously prepared guidance material.

With the automation of more data networks, there now are considerable data available operationally at intervals of 1 hour or less in some parts of the United States, however, the availability of hourly data from the climatological network is still limited and the OFS MAP program still produces time series at only a 6 hour interval. Techniques are now available in OFS to use gridded hourly precipitation estimates generated by merging radar, raingage, and other data, however, studies have shown that these estimates in most cases are biased as compared to MAP values generated from historical raingage data. There is of course also the scale effects of the models which must be accounted for prior to using hourly precipitation estimates operationally with models that are calibrated at a 6 hour time interval. A further discussion of historical versus operational bias and scale effects is included in chapter 8 of this manual.

If there are sufficient historical hourly stations, a 1 or 3 hour MAP time series could be generated and used for calibration. If snow is also a factor, the SNOW-17 model allows for the generation of the rain plus melt time series at the same interval as the precipitation time series even if temperature data are only at a multiple of the precipitation data time interval. Thus, for example, SNOW-17 will produce hourly rain plus melt from hourly MAP data and 6 hour temperature estimates (the NWSRFS historical temperature processing program only produces 6 hour mean areal estimates). In order to then use such calibration results for operational forecasting would require a procedure that would produce real time hourly estimates of precipitation from raingage data or the assurance that there was no bias between the historical MAP values and hourly radar based estimates available operationally.

Non-Mountainous Area Precipitation Estimates

Introduction

Even though it has been previously decided by looking at the station data and other factors that the non-mountainous area procedure can be used to generate MAP estimates, it is probably a good idea to reexamine the station data after the consistency checks and adjustments have been completed to make sure that the variation in average station precipitation over the river basin conforms to the definition of a non-mountainous area. If so, the computation of the MAP time series is quite straight forward.

Background

The MAP program of NWSRFS uses the following sequence when completing the data record by estimating all missing data values and time distributing accumulated amounts.

1. Estimate all missing hourly station values using only other hourly stations since these are the only data that always correspond exactly to the time period of the missing record. Also time distribute any accumulated amounts at hourly stations based on the other stations in the hourly data network. If there are no estimator stations available for certain periods, i.e. all hourly stations have missing data, then the estimates are set to zero for all the hourly stations for these periods. The program lists the number of hours that this occurs. If the number of hours that precipitation is set to zero is significant (something like more than 0.1% of the period of record), more hourly stations should be added to the analysis if possible. This is most important when some of the hourly stations are assigned weight for computing MAP values.

2. Time distribute all observed values at daily reporting stations into hourly estimates based on the observation time of the daily report and the stations in the hourly network. If no hourly data are available to time distribute a daily value, i.e. all hourly stations were missing or recorded no precipitation, the daily total is currently assigned to the hour of the observation time and the other hours in the day are set to zero. The program tabulates the

number of days for each station when the daily total can't be distributed for amounts greater and less than 0.5 inches. The maximum daily amount that couldn't be distributed is also tabulated for each station. If the number of cases for some stations are much larger than for the majority, this may indicate a problem with the observation times specified for those stations. Such discrepancies can also occur for daily stations that are not in the vicinity of any hourly stations or for high elevation daily reporting stations when all the closest hourly stations are at much lower elevations. Correcting the observation time histories and adding additional hourly stations, if available, are the only ways to reduce the number of days when the distribution of daily amounts is not possible. However, even in the best of cases there will always be many days when isolated storms effect only daily stations.

3. Estimate the missing values at daily reporting stations based on the observation time of the report and all available hourly estimates from both the hourly network and daily stations that were time distributed in step 2. The missing daily totals are then time distributed into hourly amounts just like the observed daily data.

After this process is complete, all of the stations have a complete record, i.e. no missing or accumulated data, and the data for each station are defined at an hourly time interval. Station weights can then be multiplied times the data values and the results summed up over the specified time interval to generate the MAP time series for each subarea.

The equation used for estimating missing precipitation data in non-mountainous areas is:

$$P_x = \frac{\sum_{i=1}^{i=n} \left\{ P_i \cdot \left(1.0 / d_{x,i}^2 \right) \right\}}{\sum_{i=1}^{i=n} \left(1.0 / d_{x,i}^2 \right)} \qquad (6\text{-}3\text{-}1)$$

where: P = precipitation amount
x = station being estimated
i = estimator station
n = number of estimator stations, and
d = distance

The estimator stations are the closest stations with available data in each quadrant surrounding the station being estimated. The quadrants are based on the roughly 4 km HRAP grid system. Estimated amounts are not used to estimate other amounts. The same weights, i.e. $1/d^2$, are also used to time distribute accumulated amounts. Equation 6-3-1 assumes that the mean precipitation at all the estimator stations can be assumed to be the same as the station being estimated, which is the basic assumption of the non-mountainous area procedure.

Station Weights

The assumption for a non-mountainous area is that the mean precipitation is essentially the same at all locations with the basin, not only where there are stations, but also at all points in between. Thus, station weights can be determined merely by the location of the gages within the boundaries of the area. This requires that basin boundaries, i.e. latitude, longitude points that define the outline of the basin, are available. There are two methods in the MAP program for calculating station weights using gage and boundary locations, Thiessen weights and grid point weights. Both are generated within the program by overlaying the HRAP grid over the area. For Thiessen weights, the weight for each HRAP grid point is assigned to the closest station. For the grid point method, the weight for each grid point is divided based on the $1/d^2$ technique to the closest station in each quadrant. For both methods, the weight assigned to each station from all grid points is then summed and normalized to obtain the station weights. These weights will always sum to 1.0. Both methods produce fairly similar weights, though the grid point method assigns some weight to stations outside the boundaries that will not receive any weight with the Thiessen method. The Thiessen method is the preferred way to compute station weights in most cases.

Mountainous Area Precipitation Estimates

Introduction

As discussed earlier, in a mountainous area the mean precipitation is spatially variable. Thus, the estimation of missing data must account for the fact that the precipitation catch at the estimator stations may be different than at the station being estimated. Also, station weights cannot be computed based on the location of the gages with the boundaries of the area. The procedure outlined in this section was primarily developed to make it relatively easy to minimize bias between historical (calibration) estimates of MAP and subsequent operational estimates. For model parameters determined during calibration to perform in a similar manner during operational use, it is imperative that there is not a bias between the precipitation values used as input to the two applications of the model. The procedure relies heavily on isohyetal analyses of long-term precipitation patterns. Good isohyetal maps are essential to the success of the method. By using an isohyetal analysis as the basis for computing areal estimates, the typical precipitation pattern defined by the isohyetal maps is used to interpolate between station values and used to extrapolate into ungaged portions of the area. In many mountainous watersheds there are no stations at the highest elevations, thus the relationships defined by the isohyetal maps are used to estimate the amount of precipitation falling above the highest gages. This works well over the long term, but can result in errors for individual storms due to changing dynamics from one storm to the next. Since model parameters are based on the simulation of many events, these random errors for individual events should not affect the determination of parameter values during calibration. Operationally it would be beneficial to have a procedure that would account for the changing storm dynamics and still maintain the same long term pattern defined by the isohyetal maps used during calibration.

The suggested steps for computing MAP in mountainous areas is as follows:

1. Determine mean monthly precipitation for each station for the period of record,

2. Determine if station weighting should be done on an annual or seasonal basis,

3. Determine the mean seasonal or annual precipitation for the historical data period for each area within the river basin where MAP time series are to be generated after checking the isohyetal maps to determine if adjustments are needed for incompatibilities in the periods of record, differences between station data and isohyetal map estimates, or disparities based on a water balance analysis,

4. Determine annual or seasonal station weights for each MAP area by adjusting subjectively determined relative weights so that the correct areal average precipitation for the historical period of record will be generated, and

5. Compute MAP time series for each area within the river basin for the period of record using the predetermined station weights determined in step 4.

The remainder of this section describes recommended procedures for each of these steps.

Determining Mean Monthly Precipitation for Each Station

If all stations had complete data for the entire period of record, monthly, seasonal, and annual mean values could be computed directly from the observed data. However, in most cases the periods of observed data vary from station to station and when data are collected, there are times when the values are missing. Since the amount of precipitation can vary considerably from one time period to another, mean values need to be computed for all stations for the same period. This is the main function of the Precipitation Preliminary Processing (PXPP) program of NWSRFS. PXPP runs on a monthly basis. If any data for a station are missing for a given month, the data value for the month is estimated. When missing values have been estimated for all months with missing data, PXPP computes mean precipitation values for each station for the entire period of record from the observed and estimated monthly values.

The basic equation for estimating missing precipitation data in mountainous areas is:

$$P_x = \frac{\sum_{i=1}^{i=n} \left[\left(\overline{P_x} / \overline{P_i} \right) \cdot P_i \right] \cdot w_{x,i}}{\sum_{i=1}^{i=n} w_{x,i}} \qquad (6\text{-}3\text{-}2)$$

where: P = precipitation amount ($\overline{P}$ signifies mean value)

x = station being estimated
i = estimator station
n = number of estimator stations
w = weight applied to each estimator, usually $1.0/d^2_{x,i}$ and,
d = distance

PXPP solves this equation by establishing a base station and then computing the ratio of precipitation at each station to that caught by the base station for each month of the year. The base station must be a station with very little missing data during the period of record and should be a station that is naturally consistent (i.e. consistency corrections are not needed). A station with a low catch, as compared to the other stations, should not be used as the base station. The base station should be one with a moderate to high catch to avoid very high ratios being computed during months with little or no precipitation. The location of the base station within the region being used in the analysis is not critical. The ratio of precipitation at a given station to that caught at the base station for each month is computed using all years for which both stations have complete data for the month. If both stations have overlapping data for about 5 or more years, a fairly stable relationship can be determined. PXPP then substitutes the ratio to base for the appropriate month for station x and station i in place of the mean precipitation values into Eq. 6-3-2 in order to determine the average ratio of station x to station i. In this way Eq. 6-3-2 is used to compute precipitation for months with missing data.

Reasonable estimates of mean monthly precipitation cannot be computed by PXPP for stations with little or no overlapping data with the base station. Since the base station is usually selected partly because it has data for essentially the entire period of record, these are usually stations with very short records or with frequent missing data such that monthly totals seldom can be calculated. Typically such stations are not adding much information to the precipitation analysis and thus would be discarded. The only other option is to remove such stations from the PXPP runs and then estimate their mean monthly values based on other stations and monthly averages derived from the isohyetal maps so that they can be included in the MAP program.

Within the PXPP program, determining mean values over the period of record and making consistency checks and corrections is an iterative process. Consistency corrections will increase or decrease the observed values for some portion of the period of record and thus change not only the computed mean values for the station being corrected, but also the mean values for stations using the corrected data to estimate missing months.

After consistency corrections and mean monthly values (referred to as characteristics in some cases) have been computed using PXPP, the MAP program should be run to verify that the consistency plots are similar between the two programs. Typically the plots look basically the same, but sometimes the plots are slightly different for one or two of the stations, generally hourly stations, due to differences in the computational sequence in the two programs. In MAP, as mentioned in the non-mountainous area section of this section, hourly stations can only be estimated from other hourly stations, whereas in PXPP each station can be estimated from all the

others. If inconsistencies show up in MAP that were not present in the final PXPP run, these can sometimes be removed by changing or adding consistency corrections to the affected stations in the MAP run. Changing consistency corrections in the MAP program will not automatically change the mean monthly values (characteristics) being used by the program (adjustments to the mean monthly values could be made by the user if the change in the consistency correction is significant).

In Alaska there is very little hourly precipitation data available from the NCDC climatological network. In fact only a handful of stations are available over the entire state and most of these have a short period of record. Even those with the longest record, Anchorage and Fairbanks, don't go back to when many of the daily digital records began in water year 1949. Some hourly data can be obtained from real time stations for use in the historical analysis, but these are generally very short records and many of the sites have considerable missing data. The result is that only a few of the hourly stations have sufficient data for use in PXPP and in many cases these gages are located far from the river basin being processed. However, the MAP program requires that hourly stations be available. One solution to this problem is to remove any hourly stations that don't have sufficient data overlapping with that of the base station from the PXPP runs. Then estimate the mean monthly values for these stations based on nearby stations and the isohyetal maps so that they can be included in the MAP program. However, rather than locating the stations at their actual location which can cause problems when estimating missing data, position the stations at locations surrounding, but far outside, the other stations being used for the river basin. These hourly stations will then be used for time distribution of daily amounts whenever they have observed data, but their $1/d^2$ values will be very small and will not affect the estimation of missing data for the daily stations being used.

Annual versus Seasonal Station Weights

The MAP program allows for station weights (predetermined) in mountainous areas to be specified on an annual or seasonal basis. Seasonal weights are beneficial in regions where the spatial pattern of precipitation is greatly different in one season than in another. This is typically caused by different types of storms predominating at different times of the year. The region where seasonal weights are typically needed is the intermountain west. In this region large scale storms with significant orographic effects dominate during the winter, while small scale convective events cause most of the summer precipitation. In most other regions of the United States the precipitation patterns do not vary sufficiently enough from one season to the other to justify the use of seasonal weights. In these regions annual weights are generally fully adequate.

Since the MAP program allows for only a single definition of the winter and summer seasons, the seasons used for station weighting and for consistency checks must be the same. In most cases, especially in the intermountain west, the seasons to use for these two features are the same for all practical purposes so this doesn't present a problem.

There are two types of plots that can be constructed to help in deciding whether seasonal station

weights need to be used. The first, is to plot the seasonal relationship between a high elevation station and a nearby low elevation station. Ideally the two stations selected should have the same orographic exposure, i.e. both on the same side of a major divide and not one on the windward side and the other on the leeward side. This high-low precipitation plot is generated by plotting the monthly ratio of the average catch based on the PXPP program results at the two stations. If the monthly ratio of precipitation from the two stations is quite similar throughout the year, there is little benefit to be gained by using seasonal station weights, whereas if there is a significant variation and a distinct seasonal pattern in the ratio, then seasonal weights should be used.

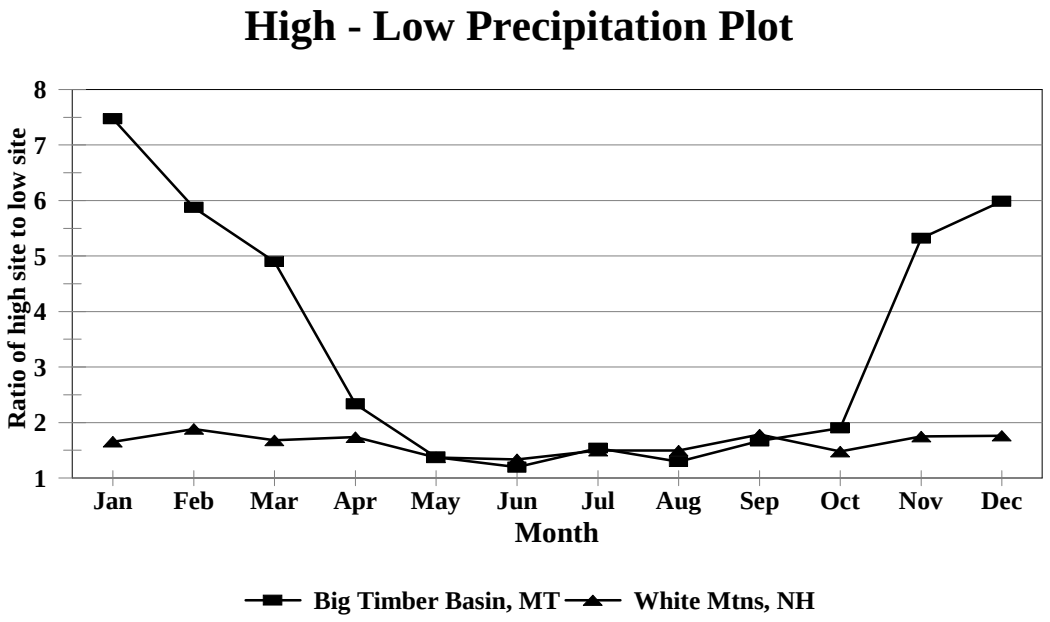


Figure 6-3-1. High-low precipitation plots for locations in the intermountain west and New England. (The Big Timber watershed plot is based on the Monument Peak SNOTEL site at 8850 feet and the Big Timber climatological station at 4100 feet. The White Mountain plot is based on climatological stations at Mount Washington (elev. 6262 feet) and Bretton Woods (elev. 1621 feet).)

Figure 6-3-1 shows such plots for two gages within the Big Timber watershed just north of Yellowstone National Park in Montana and for two gages in the White Mountains of New Hampshire. In the Big Timber case there is a very large difference in ratios from mid-winter when large orographic storms dominate to mid-summer when most of the precipitation is produced by small scale convective events. Seasonal weights are certainly justified in this region. The typical winter season in the intermountain west is from October to April and the summer season from May to September. Based on the Big Timber plot a winter season from November through March might be better for station weights, though the October to April season is probably more appropriate as far as when snowfall predominates for making consistency

adjustments. For the White Mountains annual station weights can be used since the variation in the high-low precipitation ratio is similar throughout the year.

The idea for using plots of high elevation to low elevation precipitation stations to get an insight into seasonal changes in precipitation patterns was introduced by Eugene Peck, former Director of the NWS Hydrologic Research Laboratory, [1964]. By examining pairs of high and low elevation stations scattered throughout the western mountains he showed that there were significant seasonal variations over much of the intermountain west, but little seasonal change in the ratios west of the Pacific Crest.

The second type of plot that is helpful in deciding whether to use seasonal station weights is a plot of precipitation versus elevation for each season. If the plots for the two seasons are dissimilar, this would indicate that the types of storms producing the precipitation are different and thus, seasonal station weighting would be appropriate, whereas if the relationship between precipitation and elevation is similar throughout the year, annual station weighting is sufficient. The high - low plot is best for defining the seasons to use for station weighting, but it only shows the relationship between two stations in the basin. The precipitation versus elevation plots are used to confirm whether the precipitation patterns are different for the entire network. Figure 6-3-2 shows precipitation versus elevation plots for an area in the upper Yellowstone River basin.

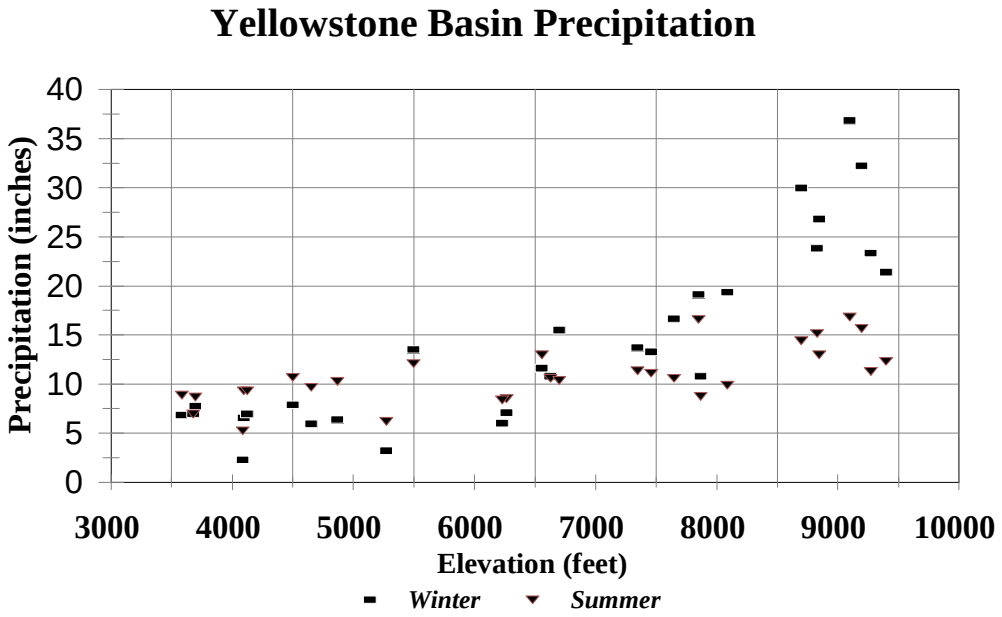


Figure 6-3-2. Precipitation versus elevation plots for the Upper Yellowstone River basin in Montana and Wyoming. Winter season is October through April, summer season is May through September.

This plot shows much more variation in precipitation with elevation, especially at the higher

elevations, in the winter season than during the summer, i.e. the orographic effects are much larger during the winter months. This clearly suggests that seasonal station weighting would be appropriate. Such differences in the precipitation versus elevation relationships are common throughout the intermountain west.

Determination of Average Mean Areal Precipitation for Each MAP Area

The next step for computing MAP in mountainous areas is to determine the average mean areal precipitation on a seasonal or annual basis for the historical period of record for each of the MAP areas within the river basin. This is done by averaging the precipitation shown on the isohyetal maps over each of the MAP areas either on an annual or seasonal basis and then applying any necessary adjustments to the result. Adjustments can be needed for several reasons.
First, if the period used to produce the isohyetal map differs from the period of historical record being used, an adjustment may be needed. Second, the mean precipitation values determined for the stations being used to compute the MAP time series should be compared against the estimate for these sites from the isohyetal maps to determine if adjustments are needed to portions of the isohyetal analysis and whether the station data makes sense. Third, water balance computations need to be made to determine if runoff information suggests adjustments to the areal mean derived from the isohyetal maps.

The first check to perform is whether the periods of record used to produce the isohyetal maps are the same as the period of record being used for the historical analysis. It is most likely that they are not the same. In that case, one needs to determine if adjustments need to be applied to the isohyetal analysis so that it reflects the historical period. The most straight forward way to do this is to take a number of the stations with long data records and compute the ratio of the average annual precipitation over the historical data period to the same value for the isohyetal map period of record. The average station precipitation determined by the PXPP program would be used to compute these ratios. If these ratios differ significantly from 1.0, then an adjustment is needed. If the ratios for all stations are fairly similar in magnitude, then the average ratio can be used to adjust the isohyetal maps. If there is a pattern to the ratios over the river basin, a different adjustment could be used for different parts of the basin. While annual precipitation should be sufficient for computing this adjustment, one could compare seasonal ratios if the MAP analysis was being done on a seasonal basis to confirm if the adjustment is similar for each season.

The second check is to compare the mean values for the stations being used for calibration as determined by the PXPP program with the mean values for these sites from the isohyetal analysis (any period of record adjustment determined in the first check should be applied to the station values picked off the isohyetal maps). This should be done on a seasonal basis for those river basins where it has been determined that seasonal station weights are appropriate for computing MAP. While the isohyetal analyses have generally been done in a careful manner with proven techniques, the results cannot be treated as the ‘true’ definition of what actually occurred. The data being used for calibration are typically more complete than the data used to generate the isohyetal maps. Generally the isohyetal maps use only stations with a long period of record,

whereas in calibration we also use stations with shorter records to better define the spatial variability. Also all the calibration stations should have been checked for consistency and adjusted if needed, whereas, this is probably not the case for all the stations used to generate the isohyetal maps. The best way to visualize differences between the station data and the isohyetal analysis is to plot the ratio of the value picked off the isohyetal map at each station location divided by the PXPP value for the station on a map of the river basin. Ideally the ratios should vary around 1.0. Isolated values that differ significantly from 1.0 could represent a small scale problem with the isohyetal analysis or could indicate a problem with the data for that station. If there are portions of the river basin where the ratios are quite different from 1.0 and exhibit a definite pattern, this most likely indicates that the isohyetal maps are over or under estimating the amount of precipitation in these areas; thus the long-term average values for the affected MAP areas derived from the isohyetal analysis will need to be adjusted. Before making any corrections it is best to do a water balance analysis to confirm that adjustments are needed for watersheds in these portions of the river basin.

Most isohyetal analyses are done using only station precipitation data. The techniques being used then interpolate and extrapolate these station data using elevation, aspect, prevailing storm direction, and other factors to estimate the average precipitation pattern over the region being studied. One way to determine if the average volume of precipitation shown on an isohyetal map is realistic is to do a water balance analysis. The water balance equation for a watershed can be expressed as:

$$P - R - \Delta S - L = ET \qquad (6\text{-}3\text{-}3)$$

where: P = average annual precipitation,
R = average annual runoff,
ΔS = average annual change in storage,
L = average annual losses or gains due to transfers across watershed boundaries or losses to deep groundwater aquifers, and
ET = average annual evapotranspiration.

Over a sufficiently long period the change in storage term should be essentially zero except for watersheds with a significant glaciated area. The watersheds used for a water balance analysis would be headwater areas with a value of the L term equal to zero and few complications such as large reservoirs or substantial irrigated acreage. Local areas could also be included as long as the area is relatively large enough so that a reliable estimate of the local runoff can be determined by subtracting upstream flows from downstream flows. The precipitation used would be that derived from the isohyetal analysis adjusted to the historical period of record. The runoff would be the mean annual runoff for this same period. In many cases the period of record of the streamflow data used to compute average annual runoff is less than the period of record being used for historical data analysis. For these watersheds MAP time series should be generated for all subareas using the mean areal average computed directly from the isohyetal analysis (adjusted for calibration period of record). Then the average annual precipitation for use in the water

balance computation can be calculated for the same years as the runoff data. In these cases, which is typical for many watersheds, one would complete the remaining steps in the MAP computational procedure so that time series would be available to determine the proper annual precipitation to use in the water balance equation and then if further isohyetal map adjustments were needed, the steps would be repeated using adjusted values of the mean areal averages.

The average annual ET determined from the water balance computations for as many watersheds as possible over the river basin should be tabulated for comparison. Another good way to see if the computed ET values make sense is to plot them against elevation. The variation in ET over the basin and with elevation should make sense. If there are watersheds that deviate significantly from a reasonable variation of ET over the basin, it is likely that the isohyetal map needs to be adjusted for these areas. If these watersheds correspond to the portions of the basin where the ratio of the station data to the isohyetal map estimate for the station were substantially different from 1.0, this further indicates that there are problems with the isohyetal estimates of precipitation over those areas. Adjustments should then be determined so that a more reasonable estimate of the average mean areal precipitation, annual or seasonal, can be used for computing each of the MAP time series. For watersheds included in the water balance analysis, the adjustments are based on both the water balance and patterns in the plots of the ratios of observed station data to the isohyetal estimates. For other drainages in the river basin, the adjustments are based on the patterns shown in the ratio plots.

Water balance computations in regions with large glaciated areas is more complex. In these cases, one needs to have a general idea as to whether there was a significant gain or loss of ice stored in the glaciers over the period being analyzed. Then either an attempt can be made to estimate the ΔS term in Eq. 6-3-3 so that ET variations could be computed directly or the value of $ET + \Delta S$ can be computed. Then ideally one could compare the $ET + \Delta S$ values for watersheds with and without glaciers, or at least with different amounts of glacial coverage, to see if the variation makes sense.

To illustrate these last two checks, we will look at a portion of the Upper Missouri River basin near Yellowstone National Park. Figures 6-3-3 and 6-3-4 show plots for both the winter and summer seasons of the ratio of the estimate picked off the isohyetal maps at each station location divided by the PXPP precipitation estimate for each station. For illustration purposes two watersheds, as well as their two elevation zones, are outlined on these plots; the Gallatin River near Gallatin Gateway, Montana and the Madison River near West Yellowstone, Montana. While the ratios tend to scatter around 1.0 over much of this region, there are definite places where there is a pattern of values greater than 1.0. Ratios greater than 1.0 appear over the drainage area of both of the outlined watersheds suggesting that the PRISM precipitation estimate is likely too large. Table 6-3-1 shows the results of water balance computations for a number of watersheds in this region. The ET values from Table 6-3-1 are plotted against the the mean elevation of each watershed in Figure 6-3-5. This plot shows a reasonable variation in ET versus elevation for most of the watersheds, however, the values for the Gallatin and

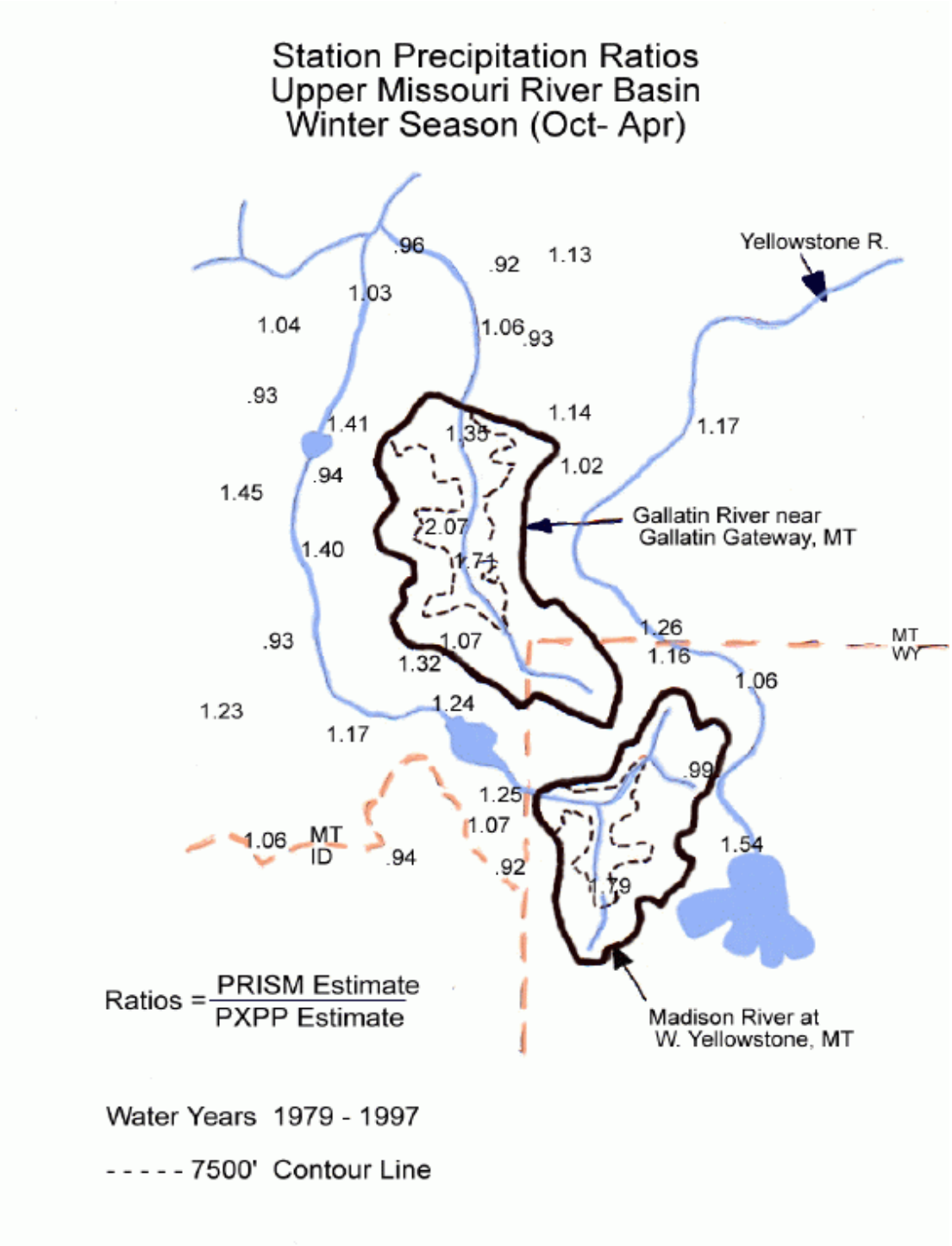


Figure 6-3-3. Winter season precipitation ratios for the Upper Missouri River basin.

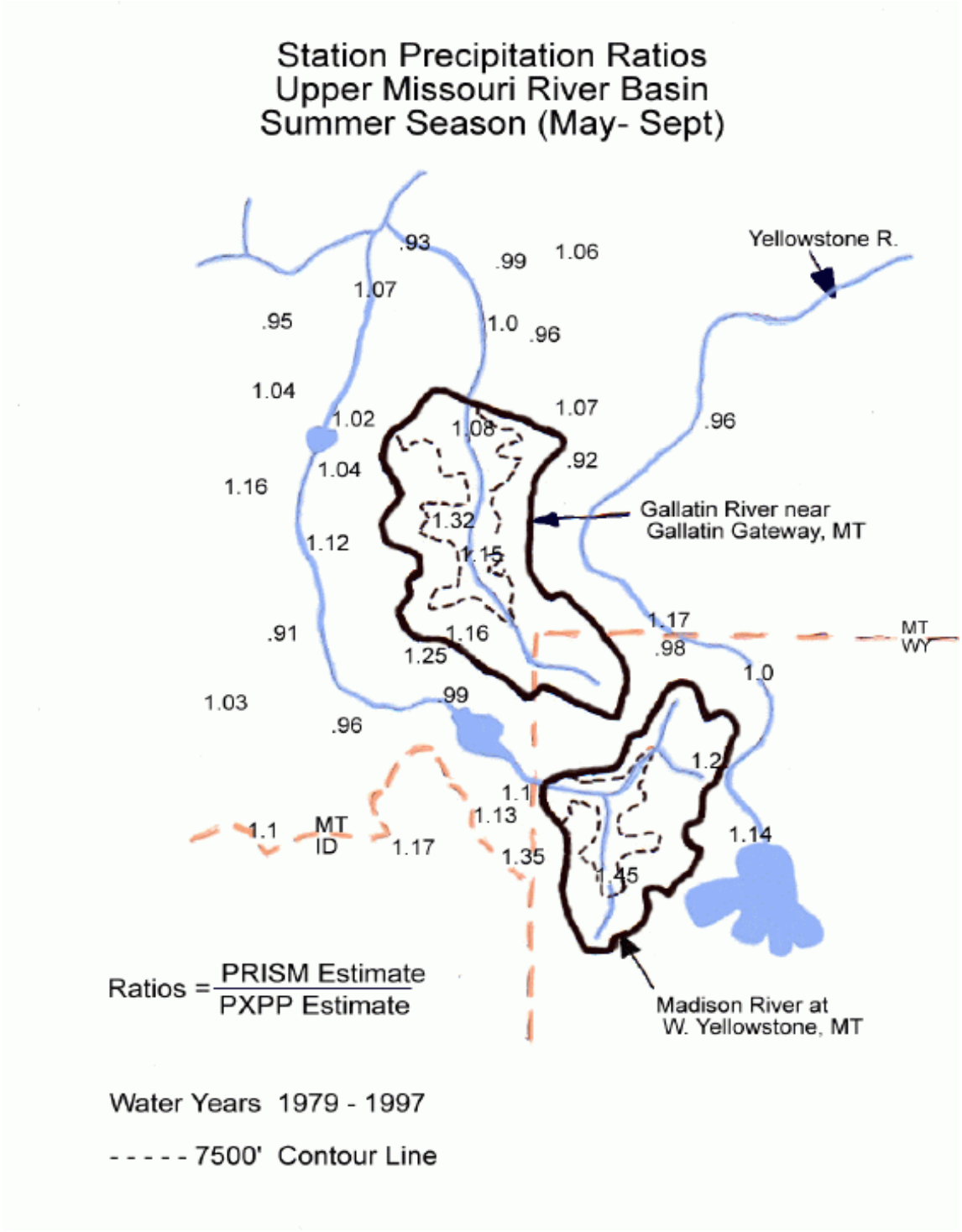


Figure 6-3-4. Summer season precipitation ratios for the Upper Missouri River basin.

Watershed	Period	Elevation	Pcpn	Runoff	ET
Stillwater R. nr. Absarokee,MT	wy 79-95	7186.	29.5	12.2	17.3
Clark Fk. nr. Belfry, MT	wy 79-97	7463.	27.7	9.5	18.2
Big Hole R. nr. Wisdom, MT	wy 89-97	7373.	24.8	4.5	20.3
Boulder R. nr. Boulder, MT	wy 85-97	6661.	22.6	4.0	18.6
Boulder R. nr. Big Timber, MT	wy 79-97	7533.	32.0	13.2	18.8
Gallatin R. nr. Gallatin Gateway, MT	wy 85-97	7880.	37.0	13.8	23.2
Madison R. nr. W. Yellowstone, MT	wy 90-97	7884.	44.3	16.8	27.5
Ruby R. ab. Ruby Reservoir, MT	wy 80-97	7247.	23.6	4.7	18.9
Lamar R. nr. Tower Falls, WY	wy 89-97	8366.	36.2	20.2	16.0

Table 6-3-1. Water balance for watersheds in the Upper Missouri and Upper Yellowstone River basins (Units for elevation are feet and for precipitation, runoff, and ET are inches).

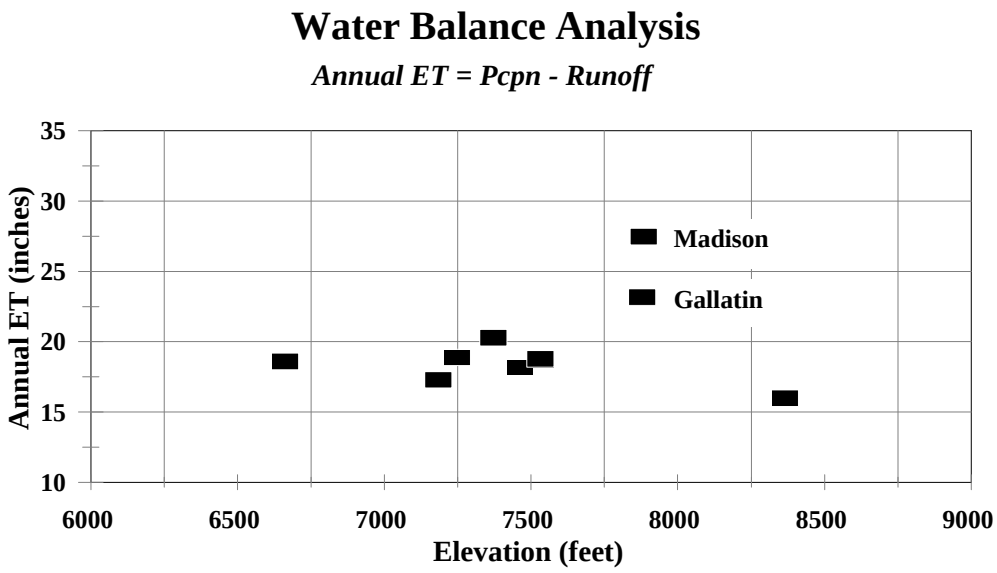


Figure 6-3-5. Annual ET derived from water balance computations versus elevation for watersheds shown in Table 6-3-1.

Madison Rivers do not conform to the relationship defined by the other watersheds. Thus, the water balance computations confirm that the PRISM precipitation estimates are too high in the portions of the river basin where these watersheds are located. For the Gallatin River, based on these 3 plots, adjustments of 0.625 and 0.87 (correspond to average precipitation ratios of 1.6 and 1.15) were applied to the winter and summer seasons, respectively, for the lower elevation zone (below 7500 feet) and an adjustment of 0.91 (corresponds to an average precipitation ratio of 1.1) was applied to both winter and summer for the upper elevation zone. These adjustments are applied to the average mean areal precipitation for each zone derived from the isohyetal maps after any corrections for differences in the period of record were applied. This resulted in an average annual precipitation of 31.9 inches for the water year 1985-1997 period and an annual ET of 18.1 inches which appears quite reasonable. Similarly adjustments to the Madison River PRISM estimates of precipitation could be computed so that a reasonable water balance estimate of annual ET could be generated. The precipitation ratio patterns would also be used to compute adjustments for areas not included in the water balance analysis, such as downstream locals on the Gallatin and Madison Rivers.

Determination of the Average Mean Areal Precipitation in Data Sparse Regions

In some areas the historical precipitation gage network is so sparse that it is very unlikely that isohyetal maps generated from these very limited data will indicate the true magnitude of the average precipitation over the region. This is generally the case with much of Alaska. The best that can be hoped for in such an area is that the PRISM analysis will reflect a reasonable relative distribution of the average precipitation pattern over the region. In most of Alaska a different approach must be taken to determine the average mean areal precipitation over each MAP area for the historical period of record, since it is highly unlikely that averaging the values from the isohyetal maps over each MAP area will provide a usable value. The suggested approach for these extremely data sparse areas involves starting with the runoff data which integrates the volume of water being produced by a watershed. Then the average amount of annual ET is estimated and added to the runoff to get an estimate of the average mean areal precipitation for the watershed. If multiple subareas are being used for the watershed, the average mean areal value for the watershed can be distributed into each subarea based on the pattern suggested by the isohyetal maps. This approach requires that the runoff data are corrected for any diversions across the boundaries of the watershed and complications such as irrigation or deep recharge are not affecting the water balance.

Suggestions for estimating the annual ET for use in such computations are included in Section 6-5 in the section “Determining Precipitation and Actual ET from Runoff, Evaporation, and Vegetation Information.” It is very important in this approach that ET estimates are done in a consistent manner from one watershed to another so that the resulting pattern of actual annual ET values makes sense when taking into account differences in climatic conditions, elevation, and vegetation cover. This approach may require a number of iterations since the initial guess at the ratio of actual ET to ET Demand may not turn out to be what actually occurs when the models are run. Also certain changes to model parameters can change this ratio and thus, alter the water

balance. It may also turn out that the relative distribution of precipitation between the subareas suggested by the isohyetal analysis may not give reasonable results when subarea variables such as the amount of snow cover or change in glacier volume are compared to observations of these variables. This will require modifications to the assumed distribution of precipitation over the watershed. Differences in the period of record of the runoff data and the historical data period being used for calibration can be adjusted for by comparing the ratio of the catch at precipitation stations with long records for the two periods.

After this procedure is applied to several watersheds in a region, the ratios of the precipitation amounts actually used to produce reasonable simulation and water balance results to the isohyetal map amounts can be calculated. These ratios can then be used to adjust the isohyetal map values for nearby local drainages or for watersheds with complications that don't allow the runoff data to be used to derive the areal average precipitation. The end result should be a more realistic estimate of the amount of precipitation than can be determined from isohyetal estimates derived from a very sparse precipitation network.

Determining Station Weights

The last step before computing the time series for each MAP area within the river basin is to determine the weight to be applied to each station. For most MAP areas within a mountainous region predetermined station weights should be used. As opposed to Thiessen or grid point weights which always sum to 1.0, predetermined weights generally will not sum to 1.0 because the precipitation pattern is not uniform over the area. Thiessen or grid point weights can sometimes be used for a few MAP areas in a mountainous region. These weights are appropriate in a flat portion of a mountainous river basin where the average precipitation is basically uniform over the area and all the stations that will receive weight have the same average precipitation as the MAP area. Predetermined station weights can be applied on an annual or seasonal basis as discussed earlier in this section. The recommended procedure for deriving station weights involves first subjectively assigning relative weights that sum to 1.0 to the appropriate stations and then adjusting the relative weights such that the summation of the final weight times the average precipitation for the each station will equal the average mean areal precipitation for the MAP area as determined from the isohyetal analysis and any needed adjustments.

The selection of which stations should be weighted for each MAP area and the assigning of relative weights to each of these stations is currently a subjective process. Hopefully in the future some objective techniques can be developed to assist in this process. The primary factors to consider when selecting the stations to be weighted and assigning relative weights is where the station is located relative to the MAP area and how representative is the catch at the station to other portions of the area. The location is a function of latitude, longitude, and elevation. In general, stations within or close to the area should be given most, if not all, the weight. This is especially true in regions or seasons when there is significant spatial variation in the amount of precipitation during storm events. If the region or season can be characterized by large scale storms with substantial orographic components and thus the typical spatial variation in

precipitation amounts follows a fairly well defined orographic pattern, then stations further outside the MAP area that reflect the orographic exposure of portions of the area can be given some weight. One needs to be careful not to give weight to stations too far outside the MAP area such that the resulting time series represents the precipitation over an area much larger than the area that you are trying to represent. Assigning weight to stations a considerable distance outside the MAP area will tend to cause the model response to individual storms to be damped out. This is more of a concern in basins where there are significant runoff events caused by rainfall than in basins where almost all the runoff comes from the melting of the seasonal snow cover, though even in snowmelt dominant basins variations in the amount of snow accumulation could be damped out by assigning weights to stations far outside the MAP area.

There are several displays that can be examined to help in determining which stations to weight and what relative weight to assign to each station in addition to the users knowledge of the climatic conditions in the region. An option in the PXPP program will generate tables which show the correlation of monthly precipitation at each site to all the other sites on a seasonal or annual basis. In these tables known seasonal patterns as reflected by variations in each stations monthly ratios to the base station are removed prior to computing the correlation. For a given station the correlation coefficients for all other stations relative to that station can be plotted on a topographic map. This map can then be examined to see if the correlations are primarily based on distance, elevation, prevailing storm pattern, or some other factor. Figures 6-3-6 and 6-3-7 show correlation plots for portions of the Yellowstone River basin for the winter and summer season. These figures tend to show that the correlation pattern in the winter season is related to both elevation and distance while the summer pattern appears to be primarily related to distance. This is typical of the intermountain west where large scale storms with substantial orographic components dominate in the winter and small scale convective storms cause most of the summer precipitation. This would suggest that stations further outside the MAP areas that reflect the orographic pattern of portions of the areas could given some weight in the winter (e.g. high elevation stations some distance outside the upper elevation zone could be weighted), but that in the summer the weights should be assigned to stations within or near the boundaries of the each area.

Another display that can be helpful when deciding on which stations to weight and assigning relative weights is an anomaly map. This map is created by drawing a line on the precipitation elevation plot (plots like figure 6-3-2) and then computing and plotting the deviation of each station from that line. The line should represent the general trend of how precipitation varies with elevation, but it doesn't precisely matter where the line is drawn. Positive deviations indicate areas where the precipitation is greater than what is expected for a given elevation and negative deviations show areas that are sheltered from the prevailing orographic pattern. Anomaly maps are primarily of value for understanding the orographic pattern over the basin produced by large scale storms, thus either the winter season is typically examined in the intermountain west or an annual anomaly map is generated in other areas where storms with substantial orographic components predominate. An understanding of the typical orographic pattern should help in selecting which stations should represent various portions of an MAP area.

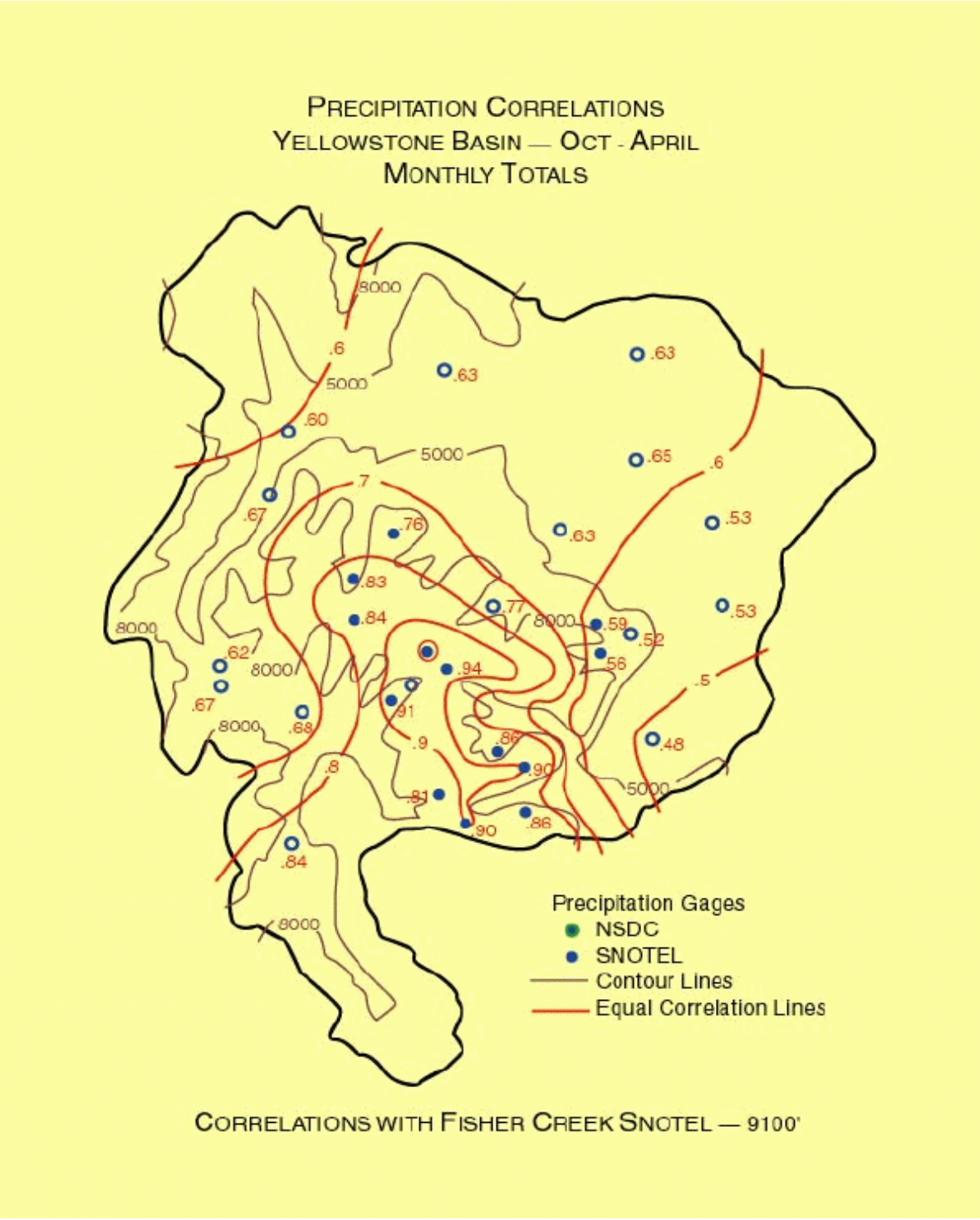


Figure 6-3-6. Winter season correlation pattern for Upper Yellowstone River basin around the Fisher Creek SNOTEL station

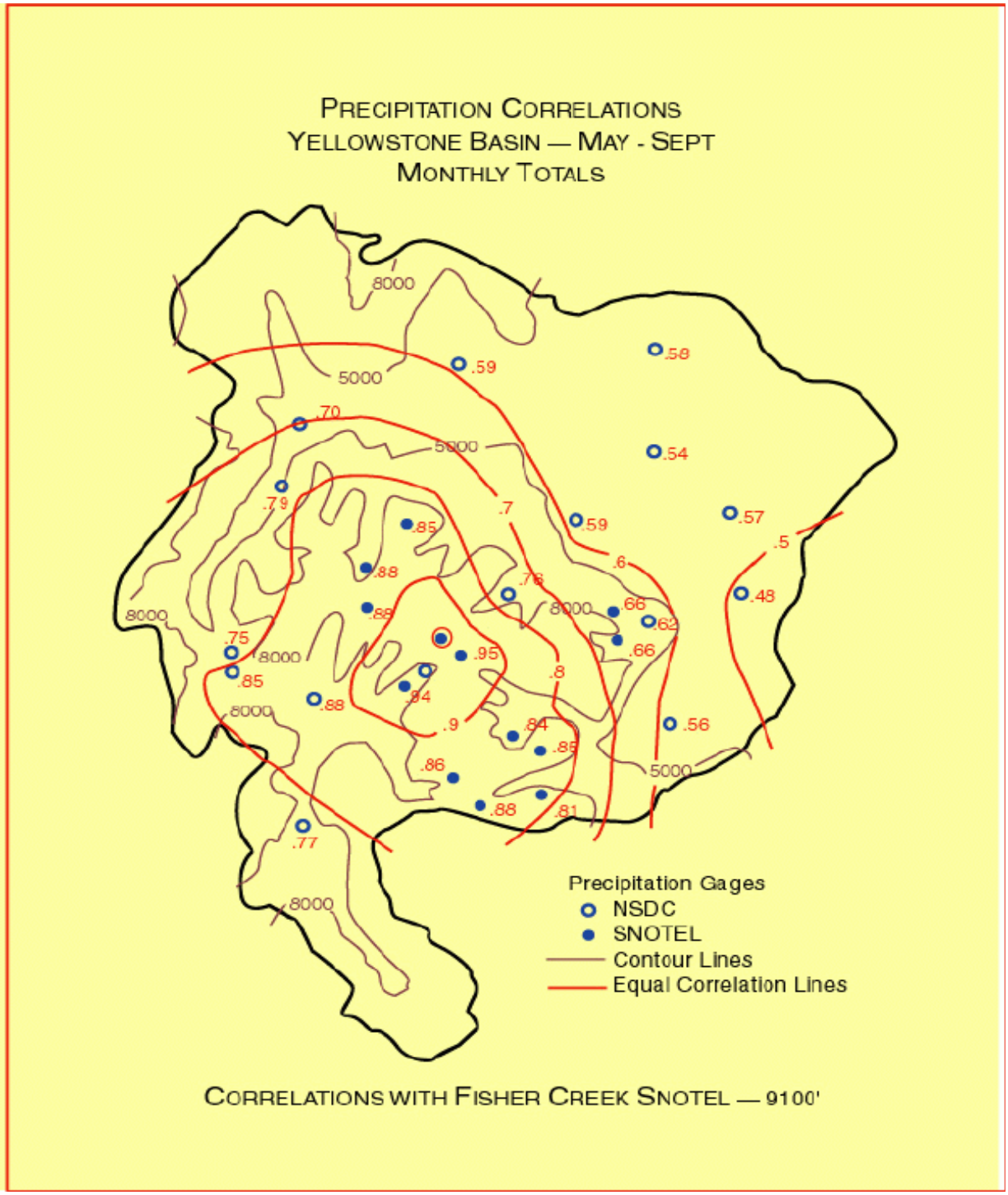


Figure 6-3-7. Summer season correlation pattern for Upper Yellowstone River basin around the Fisher Creek SNOTEL station

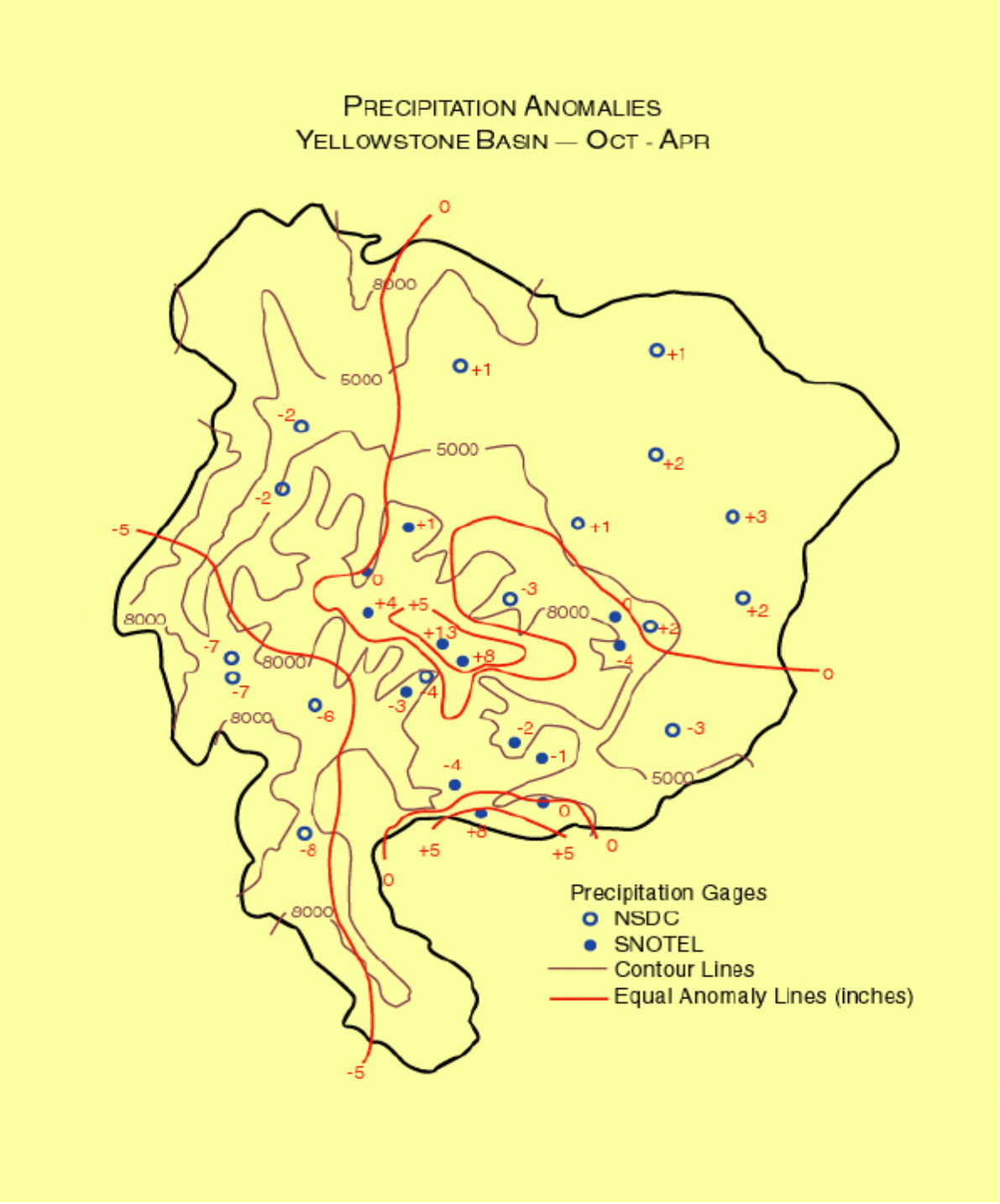


Figure 6-3-8. Winter season precipitation anomaly map for the Upper Yellowstone River basin.

Figure 6-3-8 shows an anomaly map for the winter season for the Upper Yellowstone River basin (Figure 6-3-2 shows the variation in precipitation versus elevation for this basin). This plot shows positive deviations in the northeast quadrant of the basin with the largest orographic effects at higher elevations of the Beartooth Range.

Once the stations to be weighted are selected and the relative weights chosen, then the relative weights are adjusted using Equation 6-3-4 for each MAP area to get station weights that will generate the average mean areal precipitation that has been determined to be the most likely estimate for the historical period of record. These are the predetermined weights used to compute the MAP time series for each area.

$$W_{i,s} = \frac{\overline{A}_s}{\sum_{i=1}^{i=N} \left(\overline{S}_{i,s} \cdot R_{i,s} \right)} \qquad (6\text{-}3\text{-}4)$$

where: i = station whose weight is being computed
s = season of the year
N = total number of stations with weight
W = station weight
R = relative station weight
$\overline{A}$ = average mean areal precipitation
$\overline{S}$ = station mean precipitation

Computation of the MAP Time Series

The final step in the historical precipitation analysis process for mountainous areas is to run the MAP program and compute the mean areal precipitation time series for each MAP area within the river basin for the period of record being used. The sequence used when estimating missing values and distributing accumulated amounts is the same described in the Background section under Non-Mountainous Area Precipitation Estimates in this chapter except that Eq. 6-3-2 is used to estimate missing data. The MAP program will generate summary tables for each area that show the total precipitation for each month, as well as the annual and seasonal mean values over the run period. The mean values computed should agree with the annual or seasonal averages determined from the adjusted isohyetal analysis. If not, the station weights were not computed correctly and need to be corrected.

Section 6-4

Computation of Mean Areal Temperature

Introduction

Temperature time series are needed for model calibration whenever snow is to be included in the computations. Temperature time series are also needed when the frozen ground options are to be used in the Sacramento Model and possibly for some of the other NWSRFS operations such as the Consumptive Use irrigation model to estimate evaporation. As was pointed out in chapter 6, the snow model is very sensitive to temperature. Small differences in temperatures can have a significant effect on the timing of snowmelt and will alter the model's determination of whether precipitation is in the form of rain or snow. Thus, it is very important that the MAT estimates are as unbiased as possible with regard to what actually occurred in nature. Bias can easily occur if one is not careful when extrapolating low elevation temperature measurements to high elevation portions of the basin. Such bias can seriously affect the simulation results and the proper determination of snow model parameter values.

At this stage of the process it is assumed that all the temperature data are available in the proper form and that the data have been checked for consistency and adjusted if necessary. It is also assumed that the areas where MAT estimates are to be generated have been determined based on the definition of the flow points needed for calibration and the possible subdivision of the drainage areas into elevation zones or subareas specified by other means. It is also assumed that it has been determined as to whether mountainous or non-mountainous area procedures should be used to compute areal estimates. Generally the mountainous area procedures need to be used for temperature only when there are definite variations in elevation over the river basin.

Limitation of Current NWSRFS Historical MAT Program

The current NWSRFS historical data MAT program uses only daily maximum and minimum temperature data to compute 6 hour mean areal values. In order to generate 6 hour estimates from daily max/min data, the program must make assumptions as to when during the day the max and min values occur and the shape of the diurnal variation in temperature. The program assumes that the max temperature occurs in the early afternoon and the min temperature occurs around 6 a.m.. The observation time of the max/min values is used to determine the day to which each value is assigned. If a station has a p.m. observation time, it is assumed that the max and min occur on the day that the observation is taken. If a station has an a.m. observation time, it is assumed that the recorded max temperature occurred the previous day and the min temperature occurred that morning. Thus, it is important to specify changes in observation time whenever there is a change from afternoon to morning or vice-versa for a station. The diurnal temperature pattern that is used in the MAT program is based on typical spring time patterns at the Central Sierra Snow Laboratory near Donner Pass in California and the NOAA/ARS snow research station near Danville, Vermont.

As would be expected, the daily temperature pattern produced by the MAT program is in error whenever the actual max and min temperatures occur at times considerably different from those assumed. However, errors also occur just due to the use of daily max/min values. For an a.m. observing station, rather than the recorded minimum temperature occurring that morning as assumed by the program, the recorded value could have occurred the previous morning. For a p.m. observing station, rather than the recorded maximum temperature occurring that afternoon as assumed by the program, it could have occurred the previous afternoon. Thus, even though the max and min values occur at the times assumed by the program, they may be assigned to the wrong day. Both of these problems, i.e. values occurring at times other than assumed and values being assigned to the wrong day, are illustrated in Figures 6-4-1 and 6-4-2. In Figure 6-4-1 not only are the 6 hour temperatures in error on days 5 and 6 due to the max actually occurring in the early morning hours, but some of the 6 hour values on days 2 and 3 are also off because the min recorded on the 3rd actually occurred on the morning of the 2nd. Similarly in Figure 6-4-2, the 6 hour MAT values on days 5 and 6 are in error for the same reason as the previous figure and some of the 6-hour values on day 4 are off because the max recorded on the afternoon of the 4th actually occurred on the afternoon of the 3rd. Both these problems could be eliminated by using instantaneous temperature data to determine the diurnal pattern and the time of occurrence of the max and min values. Such an enhancement would significantly reduce problems of mistyping the form of precipitation and would improve computations of the timing of snowmelt in many regions of the country. Instantaneous temperatures are used in the operational MAT program.

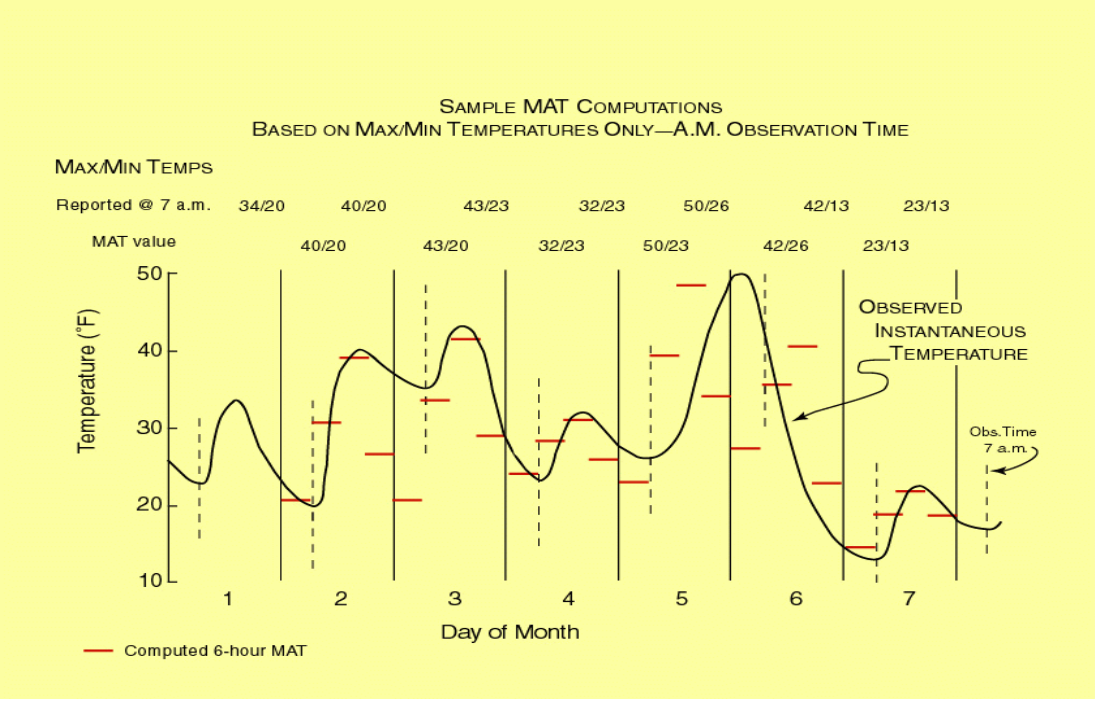


Figure 6-4-1. Sample MAT computations using only max and min temperatures and an a.m. observation time.

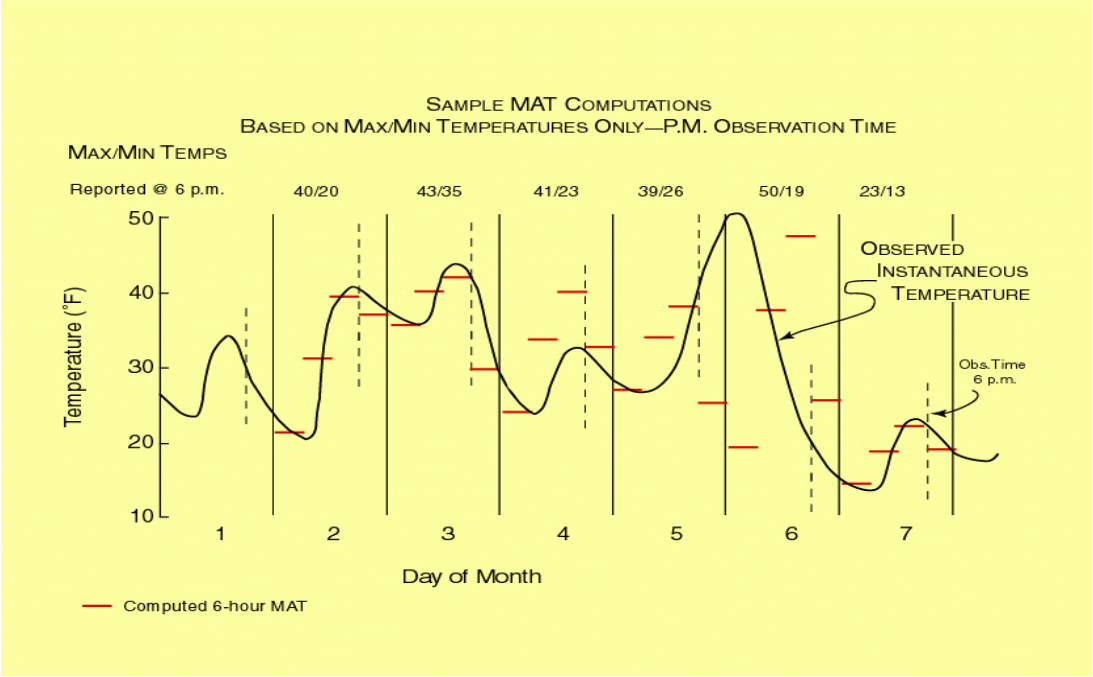


Figure 6-4-2. Sample MAT computations using only max and min temperatures and a p.m. observation time.

Estimation of Missing Max/Min Temperatures

Unlike with precipitation, mean monthly temperatures are used to adjust estimator stations to be consistent with the station with the missing values in non-mountainous, as well as mountainous areas. Missing temperature values, either maximum or minimum, are estimated in NWSRFS by the equation:

$$T_x = \frac{\sum_{i=1}^{i=n}\left[\left(\overline{T}_x - \overline{T}_i\right) + T_i\right] \cdot w_{x,i}}{\sum_{i=1}^{i=n} w_{x,i}} \qquad (6\text{-}4\text{-}1)$$

where: T = max or min temperature value ($\overline{T}$ signifies mean value),
x = station being estimated,
i = estimator station,
n = number of estimator stations, and
w = weight applied to each estimator; computed as:

$$w_{x,i} = \frac{1.0}{d_{x,i} + F_e \cdot \Delta E_{x,i}} \qquad (6\text{-}4\text{-}2)$$

where: d = distance,
ΔE = elevation difference, and
F_e = elevation weighting factor.

In non-mountainous areas the elevation weighting factor is zero. In mountainous areas F_e is used so that elevation, as well as distance, is used to determine which estimator stations are used. This is important since elevation has such a dominant effect on temperature in the mountains. The "TEMPCK" option in the MAT program can be used to investigate the effect of different F_e values on the accuracy of the estimates for a selected station. Normally, default values of F_e are used. The default values suggested are 20 (miles/1000 feet) for English units and 100 (kilometers/1000 meters) for metric units. As with precipitation, the station with the largest weight and an observed data value in each quadrant is used as an estimator. The quadrants are again based on the HRAP 4 kilometer grid system. Also, as with precipitation, estimated values are not used to determine missing values.

The TEMPCK option in the MAT program allows the user to select various values of F_e to estimate all the maximum and minimum temperatures at a specified station and then statistically compare the estimated and observed values. The TEMPCK option was used in the White Mountains of New Hampshire and the Mogollan Rim area of Arizona to investigate how the accuracy of the estimates varied with F_e. A value of F_e that gave nearly the best statistical result in both regions was selected as the default. It should be noted that for any given station the optimal value of F_e depends on the locations and elevations of the surrounding stations in the network. The most significant changes in the accuracy of the estimates occur when the use of a different F_e causes a change in the estimator station in one or more quadrants. If the closest stations in terms of distance are also the nearest in elevation, changing F_e doesn't have much affect on the accuracy of the estimates. It is recommended that the default value of F_e be used unless the TEMPCK option clearly shows that another value would improve the estimation of missing data. It is likely that the TEMPCK option would only be worthwhile trying for a station with a lot of missing data that had a large weight for an important area..

Once all the missing max and min values have been estimated for all stations, then the assumed typical diurnal temperature variation is applied in order to compute 6 hour mean temperature values at all the stations. If all stations have missing values for max or min temperatures on a given day, several resulting MAT values will be set to missing.

Computation of Station Mean Temperature

Unlike precipitation, where the PXPP program computes monthly mean values in a consistent manner for the entire period of record, there currently is no preliminary processing program for temperature in NWSRFS. The initial mean monthly max and min values are generally computed directly from the observed data for the station and then altered when consistency corrections are applied as mentioned in Section 6-2 (under the section titled "Guidelines for Making Consistency Corrections"), however, the mean values are still based only on the period of observed data for the station. Inconsistencies exist when different stations use different periods of record to compute the mean monthly values. Since there is no preliminary processing program for temperature, the stations used should have relatively long periods of record (at least 70% of the historical data period as mentioned in Chapter 3) so that the means computed from the data are close to the mean for the entire MAT computational period. However, in mountainous areas there are situations when parts of the region, especially high elevation areas, do not have any stations with that length of data and thus, stations with relatively short periods of record need to be included in the analysis. For stations with relatively short periods of record, the monthly means should be computed in a manner similar to that used by the PXPP program for precipitation. This procedure involves the use of a base station with a long, consistent record. The elevation of the base station should also be as close as possible to the elevations of the stations with short periods of record. For temperature data this procedure must be done manually. The steps in the process are as follows:

1. Select a base station and get its mean monthly max and min values for the MAT computational period from a MAT run for the entire period.

2. For each station with a short period of record, make an MAT run for the period that the station has data and get the mean monthly values for both that station and the base station for this shorter period. Compute the difference in mean monthly max and min temperatures between each station and the base station for the period when both stations have data.

3. For each station with a short record, apply the difference computed from the period when both it and the base station have data to the mean values obtained in step 1 for the base station. This results in an estimate of the mean values for the entire period for the stations with short records. These estimates will be the initial mean monthly max and min values to use for these stations.

4. As consistency corrections are applied and new mean values are computed, do <u>not</u> substitute the updated means computed by the MAT program for the stations with short records as the updated mean values computed by the program are based solely on the period of observed data for each station. If consistency corrections are applied to the stations with short records, the mean monthly values for those stations will need to be manually adjusted based on the magnitude of the correction and the portion of the observed data period over which it is applied.

Table 6-4-1 contains an example of the process for computing the appropriate means for a station

with a short record. In this example MAT values are to be generated for the historical period from 1949 to 1998. Station B has a consistent and essentially complete (i.e. very little missing data) record for the entire historical period. The table shows the mean monthly max and min values computed for this station for the 1949-1998 period. Station X is a high elevation station that was not available until 1985. Station B is the highest elevation station with a long record in the vicinity of Station X. By running the MAT program for the 1985-1998 period, mean monthly values can be generated for both stations for this portion of the historical record. These values are shown in the table (note that the means for Station B for 1985-1998 are generally higher than those for the entire historical period indicating that the later years were somewhat warmer). The average difference in temperatures between Stations X and B are then computed based on the period when both had data (note that in order to get a good estimate of the average difference between the stations, the record for both stations should be as complete as possible during the overlapping period -- the program computes the mean of all observed values for each station, whereas in reality the difference should be computed using only those months when both stations have no missing data) . These differences are then added to the means for Station B for the entire historical period in order to estimate the means for Station X that are appropriate for use during the 1949-1998 period.

Sta./Yrs.	value	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Sta. B 49-98	max	30.3	31.7	38.4	45.7	52.5	60.3	68.7	69.4	54.8	42.9	35.6	31.3
	min	2.8	5.6	11.2	21.7	28.6	31.7	35.7	36.2	30.6	20.4	10.9	6.4
Sta. B 85-98	max	33.6	34.8	41.3	47.7	54.3	63.7	70.4	70.2	57.1	45.8	38.2	33.8
	min	5.8	9.5	12.8	24.1	30.8	33.6	37.6	35.8	33.4	24.1	14.2	8.9
Sta. X 85-98	max	26.8	27.6	34.1	39.3	45.1	53.9	60.0	60.1	47.5	37.4	30.9	27.4
	min	9.5	13.7	15.4	25.3	30.4	31.2	34.2	32.6	31.5	23.8	16.0	12.3
Diff. (X-B) 85-98	max	-6.8	-7.4	-7.2	-8.4	-9.2	-9.8	-10.4	-10.1	-9.6	-8.4	-7.3	-6.4
	min	3.7	4.2	2.6	1.2	-0.4	-2.4	-3.4	-3.2	-1.9	-0.3	1.8	3.4
Sta. X 49-98	max	23.5	24.3	31.2	37.3	43.3	50.5	58.3	59.3	45.2	34.5	28.3	24.9
	min	6.5	9.8	13.8	22.9	28.2	29.3	32.3	33.0	28.7	20.1	12.7	9.8

Table 6-4-1. Example of Computing Mean Temperatures for a Station with a Short Record.

This procedure should certainly be used for any stations with data for less than 25 percent of the computational period. It should be considered for all stations with data for less than about 70% of the computational period. Ideally a preliminary processing program should be written for temperature data so that the proper means for all stations for the entire historical period will be computed correctly and automatically.

Non-Mountainous Areas

Determination of Station Weights and Generation of MAT Time Series

For non-mountainous areas the stations to be weighted and the weight assigned to each station are based on the location of the station relative to the boundaries of the area for which MAT is being computed. In the NWSRFS historical data MAT program, grid point weighting is used in non-mountainous areas. Basin boundaries are specified by latitude/longitude points. The 4 km HRAP grid is overlain over each MAT area and 1.0/d weights are determined for each grid point within the area based on the closest station in each quadrant around the point. The weights for all the grid points in the area are then normalized so that the sum of the station weights is 1.0. These weights are then applied to the 6 hour mean temperature values that have been computed for each station to get the 6 hour MAT time series for each area.

Mountainous Areas

Introduction

In a mountainous area the average temperatures vary considerably over each MAT area with most of the variation due to differences in elevation. Thus, the estimation of the mean areal temperature in mountainous regions is based on the typical relationship between temperature and elevation. The temperature-elevation relationship serves the same function for the recommended mountainous area temperature procedure as an isohyetal analysis does for the precipitation procedure. Such relationships are generally regional in nature, thus one relationship can be used for a number of MAT areas to estimate the average temperature at the mean elevation of the area.

Temperature-Elevation Analysis

The temperature-elevation analysis should be done on a regional basis, not for individual watersheds. The regions for which temperature-elevation relationships are developed need to be carefully selected based on climatic factors. Typically for temperature the major factors are latitude and distance from major water bodies, especially oceans. Other climatic factors may also need to be considered. It is up to the user to determine the regions over which the relationship of temperature with elevation can be considered the same.

The typical relationship between temperature and elevation varies seasonally due to differences in heating caused primarily by differences in the length of days and the amount of solar energy. Also the variations of maximum and minimum temperature with elevation are not typically the same. The lapse rate during the middle of the day when the max temperature generally occurs is generally steeper than the lapse rate in the early morning when temperatures are typically at the minimum values. Thus separate temperature-elevation relationships are developed for max and min temperatures on a monthly basis to reflect these variations.

In order to develop the relationships, the mean max and min temperatures for all stations in the region (after consistency corrections have been made) are plotted versus elevation for each month of the year. In NWSRFS, the TAPLOT program can be used to generate these plots. TAPLOT uses the letters of the alphabet to plot each station, thus a maximum of 26 stations are allowed. If there are more than 26 stations in a region, some of the stations that represent similar elevations can be removed before generating the plots. Once the plots are generated, lines can be drawn for each month to represent how max and min temperatures typically vary with elevation. When drawing the lines there are several factors that should be considered:

1. Though one straight line is normally drawn, the rate of change of temperature with elevation can vary in some regions. Also during certain months in some areas, lapse conditions may persist at higher elevations, while there is an inversion at the lower elevations. This is especially true during winter months in interior Alaska where even max temperatures can show inversions at low elevations during months with little or no sunlight.

2. There may be some stations that experience local effects, like cold air drainage, and are thus not representative of the elevation where they are located

3. The lines drawn should represent physically realistic lapse and inversion conditions. For example, lapse rates are typically in the range of about 0.3 to 0.9 °C/100 m for most areas, though they can be less, especially for min temperatures, in the winter at northern latitudes. A typical lapse rate used in many snowmelt modeling studies throughout the world is 0.6°C/100 m (3.3°F/1000 ft) for mean daily temperature (the max temperature lapse rate would be greater than this value and the min temperature lapse rate would be smaller). It can be helpful to draw a 0.6°C/100m lapse rate line on each plot to assist in making sure that the relationship selected is physically realistic.

4. The seasonal variation in lapse rates should exhibit a smooth transition from month to month. Typically lapse rates are greatest in the summer and smallest in the winter. In some regions minimum temperature lapse rates will show a double peak. In such regions the min lapse rate will peak in the spring and again in the fall when humidity is lower and it is easier to cool the atmosphere and decrease during the humid summer months. It is a good idea to compute the lapse rates from the temperature versus elevation relationships that are developed and plot the rates on a monthly basis to make sure the seasonal changes are reasonable and that there are not abrupt variations from one month to the next.

The temperature versus elevation relationships are primarily used to extrapolate temperatures from lower to higher elevations due to a lack of high elevation data in most basins. Since most snow accumulation, and thus the majority of the runoff, comes from higher elevations, it is critical that the extrapolation is physically realistic. Improper lapse rates can result in biased MAT values being generated at higher elevations. Biased temperatures will result in unreasonable snow model parameter values and poor simulations of the snow accumulation and ablation process.

Figure 6-4-3 shows max and min temperature versus elevation plots for the month of March for the Merrimack River basin in Massachusetts and New Hampshire. This river basin was divided into 2 regions as far as temperature versus elevation relationships. This was partly due to differences in latitude from south to north in the basin and partly due to the influence of the ocean on the southern portion of the basin. During the summer months the relationships were essentially the same, as the cooling effects of the ocean in the south offset the increased latitude of the northern portion. In the winter, temperatures were significantly cooler at a given elevation in the north than in the south where the ocean had a moderating effect. The lapse rates for max and min temperatures for each month were the same for both regions, only the intercepts varied. It should be noted that in addition to the max and min temperature versus elevation lines for the two regions, a line showing the 0.6°C/100m lapse rate is also drawn on the plots to assist in making sure that the lapse rates are realistic.

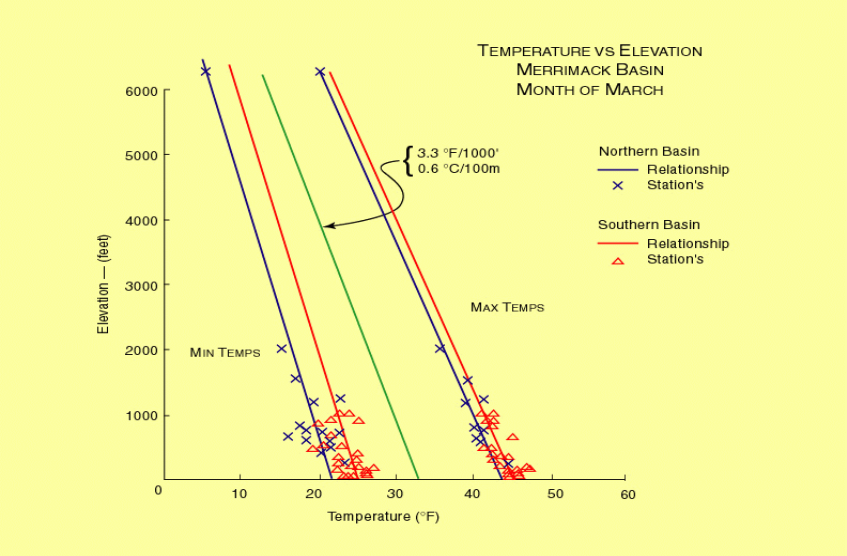


Figure 6-4-3. Temperature-elevation plots for March for the Merimack River basin.

Figure 6-4-4 shows a plot of the seasonal variation in lapse rates for the Upper Missouri River basin. This figure shows the lapse rates computed from the initially drawn temperature versus elevation lines and adjusted lapse rates. These adjustments were made so that the seasonal lapse rate pattern made a reasonably smooth transition from one month to the next. The adjusted lapse rates were used to slightly modify the initial temperature versus elevation lines. It is very important to compute and plot the lapse rates for each month as a check that the values are realistic and that the transition from month to month is reasonable.

Determination of Station Weights

The procedure recommended for temperature computations in mountainous areas for use with

NWSRFS involves establishing a synthetic or "dummy" station at the mean elevation of each MAT area and giving all the weight to the synthetic station. A synthetic station is one with no observed data; i.e. all the data values are estimated from surrounding real stations. In order to define a synthetic station, mean monthly max and min temperatures and a location are needed. The mean monthly max and min temperatures are picked off the regional temperature versus elevation plots for the mean elevation of the MAT area that the synthetic station represents. The location of the synthetic station is subjective, but should be selected so that the best possible estimator stations will be used to generate the synthetic data values. The location, elevation, and completeness of data for potential estimator stations are the major factors to consider when locating the synthetic station. Ideally the elevation of the estimator stations should be as similar as possible to the elevation of the synthetic station. Typically the synthetic station should be located so that several good estimators will be used to generate the data values.

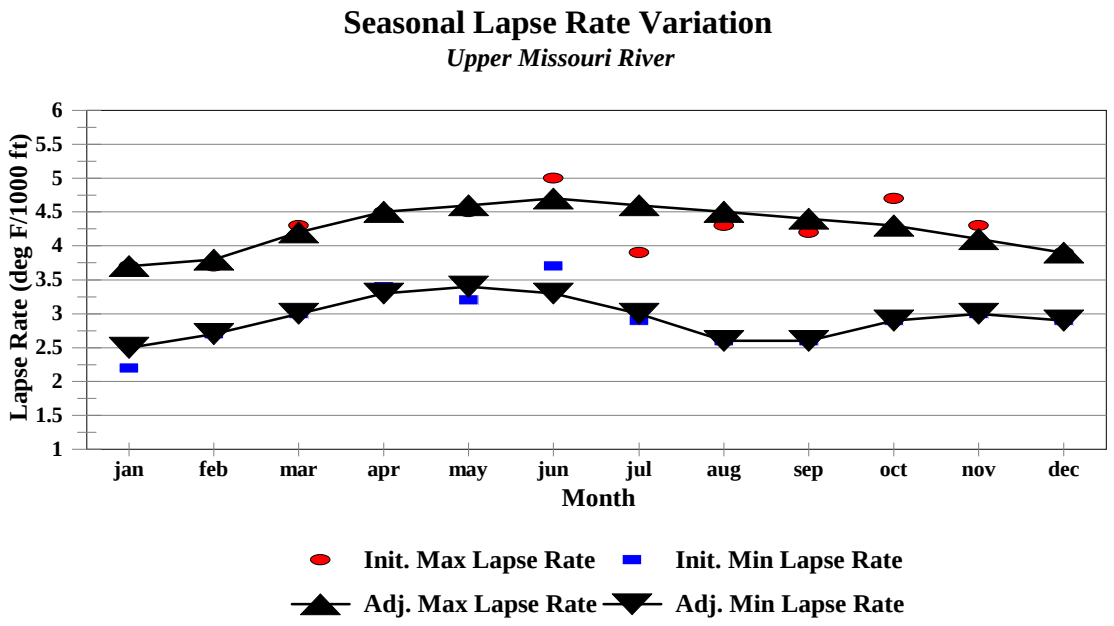


Figure 6-4-4. Seasonal variation in lapse rates for the Upper Missouri River basin.

Computation of MAT

The final step in the historical temperature analysis process in a mountainous area is to input the synthetic station information and the station weights and run the calibration MAT program to compute the time series for each MAT area for the entire period of record. For each MAT area, the synthetic station associated with that area is given a predetermined weight of 1.0 and all other stations, real and synthetic, are given zero weight. Besides generating 6-hour MAT time series, the program will produce summary tables for each zone showing the mean temperatures for each month and year, as well as the overall mean for the period of record.

Section 6-5

Evaporation Estimates for Conceptual Modeling

Introduction

In order to produce a proper water balance, conceptual models require realistic and consistent estimates of evaporation as well as precipitation. Although evaporation data will generally not affect the simulation accuracy of individual storm events nearly as much as precipitation data, good evaporation estimates are critical to producing a reasonable seasonal and annual water balance. Like the other input data used to drive the models, evaporation estimates should correspond as closely as possible to what actually occurs in nature. Biased estimates of evaporation will cause model parameters to take on unrealistic values and will deteriorate the quality of the simulations.

At this stage of the process it is assumed that all the data needed to determine evaporation estimates for use by the models are available in the proper form and that the data have been checked for consistency and adjusted if necessary. It is also assumed that the areas where evaporation estimates are to be generated have been determined based on the definition of the flow points needed for calibration and the possible subdivision of the drainage areas into elevation zones or subareas has been specified by other means. It is also assumed that it has been determined whether mountainous or non-mountainous area procedures should be used to compute the estimates. As with temperature, generally the mountainous area procedures only need to be used for evaporation when there are definite variations in elevation over the river basin.

Terminology

There are a number of terms that are used in conjunction with evaporation that sometimes take on different meanings to different people, thus, these terms need to be clearly defined as to how they are used in this manual.

- Potential Evaporation - Potential Evaporation (PE) is defined as the evaporation from a well wetted (i.e. the moisture supply is not limiting the evaporation), actively growing grass surface. The PE rate for grass under these conditions is calculated from the current meteorological conditions, i.e. temperature, humidity, wind, radiation, atmospheric stability, etc..

- Free Water Surface Evaporation - Free Water Surface (FWS) evaporation is the evaporation from a water surface with no heat storage. Evaporation from lakes and reservoirs involves some energy being used to warm the water body or being released as the water body cools and thus differs from FWS evaporation. Studies have shown that FWS evaporation is essentially the same as PE from a grass surface.

• Evapotranspiration (ET) Demand - ET Demand is the term used in the Sacramento Model to specify the evaporation that occurs when the moisture supply is not limiting given the type of vegetation that exists and the activity level of that vegetation. Thus while PE is defined for an actively growing grass surface, ET Demand is based on the actual vegetation in the area and how active that vegetation is given the time of the year and other factors.

• Seasonal PE Adjustment Curve - The PE adjustment curve modifies PE values on a seasonal basis for both the type of vegetation and the activity level of the vegetation. Thus, it is the seasonally varying ratio of ET Demand to PE or expressed in another way, ET Demand is equal to PE times the seasonal PE adjustment curve.

• Actual ET - Actual ET is the amount of evaporation loss from a watershed or subarea given the ET Demand and the current moisture conditions and snow cover. The actual amount of ET is limited by the amount of moisture in the soil and the evaporation rate is generally suppressed when snow covers the vegetation. Actual ET is less than or equal to ET Demand.

• Pan Evaporation - Pan evaporation is the amount of evaporation as measured with an evaporation pan. Pan evaporation differs from FWS evaporation in that there is some change in heat storage due to changes in the temperature of the water in the pan and because energy transfer occurs through the sides and bottom of a pan. The evaporation pan used most commonly in the United States is called a Class A pan.

• Pan Coefficient - The pan coefficient is the average ratio of FWS evaporation to the evaporation measured by a pan. Pan coefficients are less than 1.0 since the evaporation measured by the pan is greater than FWS evaporation over the long-term. The pan coefficient for a Class A pan is generally said to be about 0.7. The May-October pan coefficient actually varies from 0.66 to 0.88 over the contiguous United States as shown on map 4 in NOAA Technical Report NWS 33 [*Farnsworth and Peck*, 1982].

It is important to clearly understand these terms before reading the rest of this section and in order to properly prepare and use evaporation data with conceptual models. The Sacramento Model computes actual ET starting with an ET-Demand that is derived from PE and a seasonal PE adjustment curve. The ET-Demand values input to the model can either be in the form of climatological averages (defined at the mid-point of each month) or daily estimates (calculated by applying an average seasonal PE adjustment curve to a daily PE time series).

Determination of PE

For use in deriving evaporation estimates for application with conceptual models, PE can be in the form of daily time series or average monthly values.

• Daily PE is generally computed from meteorological factors using a Penman type equation. The most common equations use air temperature, dew-point, wind speed, and solar radiation

to compute a daily value of PE since atmospheric radiation and the data to derive stability are seldom available. If direct measurements of solar radiation are not available, the amount of solar energy can be estimated from percent sunshine using the method of Hamon, Weiss, and Wilson [1954] or sky cover using the method of Thompson [1976]. Values can be computed only at synoptic stations or at other climate stations that include instrumentation to measure these variables. Temperature, dew-point, and especially wind data must be adjusted to a specified instrument height. Lindsey and Farnsworth [1992] reported that using percent sunshine to estimate the amount of solar radiation produces a reasonably unbiased estimate of PE. This study also indicated that using sky cover to get solar radiation will result in a biased estimate of evaporation. The NWSRFS synoptic data program includes a correction for removing this bias. It should also be noted that with the introduction of ASOS the sky cover data are no longer the same as the manual observations of sky cover that the Thompson method uses to compute solar radiation.

Daily values of PE could also be obtained from daily pan evaporation measurements using the appropriate pan coefficient to adjust the measured values. To be more precise the measured pan values should be adjusted for changes in heat storage using daily measurements of the water temperature in the pan. Even then the pan coefficient is only the average relationship between pan evaporation and FWS evaporation and actually varies somewhat from day to day based on meteorological conditions.

• Monthly average PE can be obtained from computations using meteorological factors or from average monthly pan evaporation adjusted by the appropriate pan coefficient. NOAA Technical Report NWS 34 [*Farnsworth and Thompson*, 1982] gives monthly average pan evaporation computed from meteorological stations for selected stations in each state. These values need to be adjusted by the appropriate pan coefficient to get average monthly PE. Technical Report 34 also contains tables of monthly average evaporation from pan data for the period 1956 to 1970. Averages of pan evaporation for other periods can be gotten directly from daily climatological stations that have pan measurements. Again the pan data need to be adjusted by the appropriate pan coefficient in order to get average monthly PE. In addition, in many parts of the country pan data are not available in the winter since the water in the pans will freeze. For these stations the ratio of annual FWS evaporation to May-October FWS values as determined from Technical Report NWS 33 can be used to determine the average amount of evaporation from November through April. Then a nearby meteorological station from report 34 can be used to prorate this total into monthly values.

Seasonal PE Adjustment Curves

The seasonal PE adjustment curve accounts for the type of vegetation in the area and the activity level of that vegetation. The curve is defined at the mid point of each month. When used with daily estimates of PE, linear interpolation is used between each point on the curve to determine the proper adjustment for each day. When monthly ET Demand is being input to the Sacramento Model, then the average monthly PE is multiplied by the PE adjustment for that month to get the

ET Demand at the mid point of the month. Again linear interpolation is used to determine the ET Demand for a given day.

Studies have shown that the evaporation rates from most actively growing vegetation differs from the rate for grass on which PE computations are based. In addition, the evaporation rates can vary considerably throughout the year as the vegetation goes through various stages from dormancy to active growth. A number of studies have shown that the total evaporation loss from a forest subjected to frequent rain, i.e. with leaves kept wet much of the time, can be 50% or more larger than the evaporation rate from grass (e.g. [*Calder, 1976]*. Under dry conditions the losses from a forest are much less due to stomatal control of water loss. For agricultural areas estimates of the seasonal PE adjustment curve can be computed for various crops using an FAO report [*Doorenbos and Pruitt*, 1977]. This report gives coefficients for various crops and climatic conditions that can be used to construct the seasonal adjustment curve based on the times of germination, attainment of full ground cover, start of maturing, and full maturity or harvest. For the crops listed, the PE adjustment during the active growing period varies from 0.8 to 1.25. There is also evidence that moss and lichen plants in tundra regions, such as Alaska and high elevations in the lower 48, use water at slower rates than does vegetation in warmer regions [*Patric and Black, 1968]*. Tundra plants function more like mulch than like transpiring vegetation. In such areas, evaporative demand, as specified by PE, can considerably exceed precipitation, but the soil almost always remains wet, even saturated, throughout the short summer season.

The PE adjustment curve used with the Sacramento Model is based solely on the time of the year and doesn't include any dynamics that account for differences in past and current climatic conditions from one year to another. For example, the activity level of vegetation in June could vary from a year with plenty of moisture and thus lush, green vegetation to a drought year when the vegetation has turned brown. The shape of the curve is also the same from year to year and thus any variations in the time when leaves come out in the spring or begin to turn colors in the fall are not taken into account. The PE adjustment curve used for hydrologic modeling is an integration of all the vegetation types and activity levels that exist over a watershed, elevation zone, or other area definition. Thus, some the information in the literature can serve as a guideline when coming up with a PE adjustment curve, but much of the process is subjectively based on a general knowledge of the type of vegetation in the area and the timing of changes in the activity levels.

Figures 6-5-1, 6-5-2, and 6-5-3 show seasonal PE adjustment curves that have been used for watershed calibrations at various locations around the country. Most of the curves in these figures exhibit a seasonal pattern of activity levels. This is because most of the watersheds contain considerable deciduous vegetation or crops that have a seasonal growth pattern. The curve for the watershed in western Washington shows essentially no seasonal variation. This watershed is covered predominantly with conifers, and the climatic conditions are such that the vegetation should be able to transpire throughout the year.

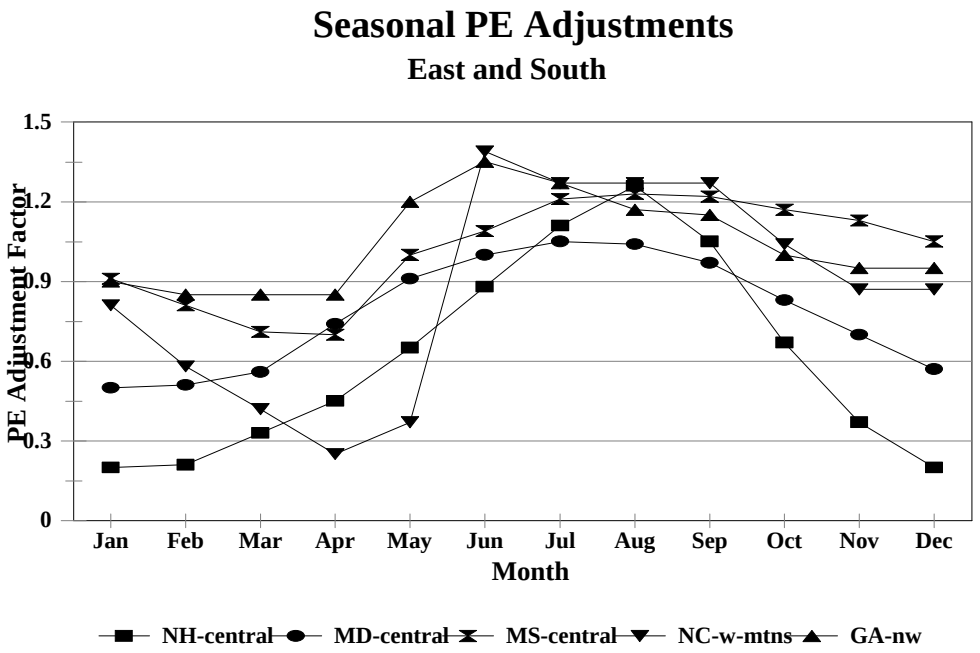


Figure 6-5-1. PE adjustment curves for watersheds in the east and south.

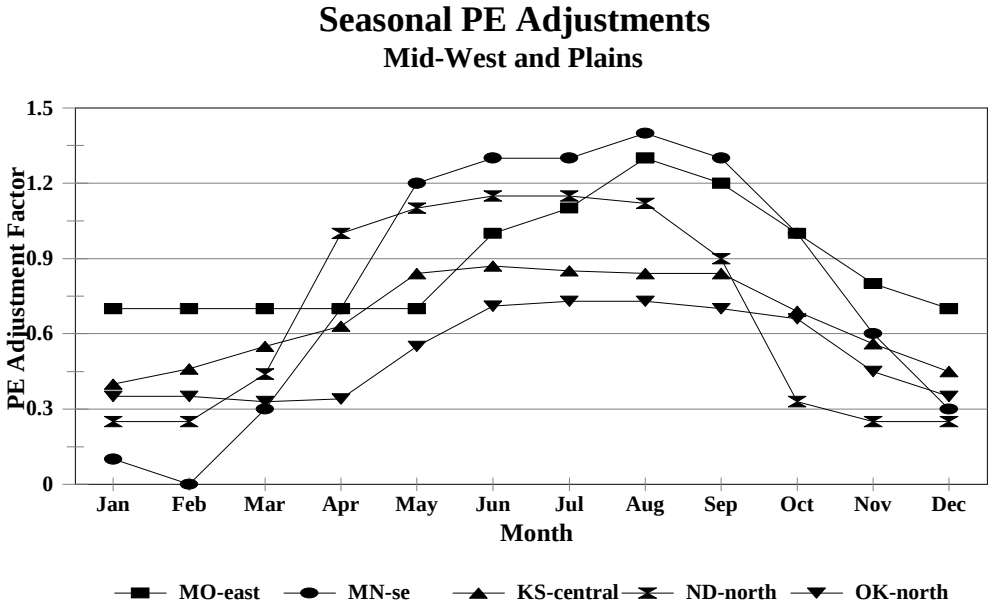


Figure 6-5-2. PE adjustment curves for watersheds in the mid-west and plains.

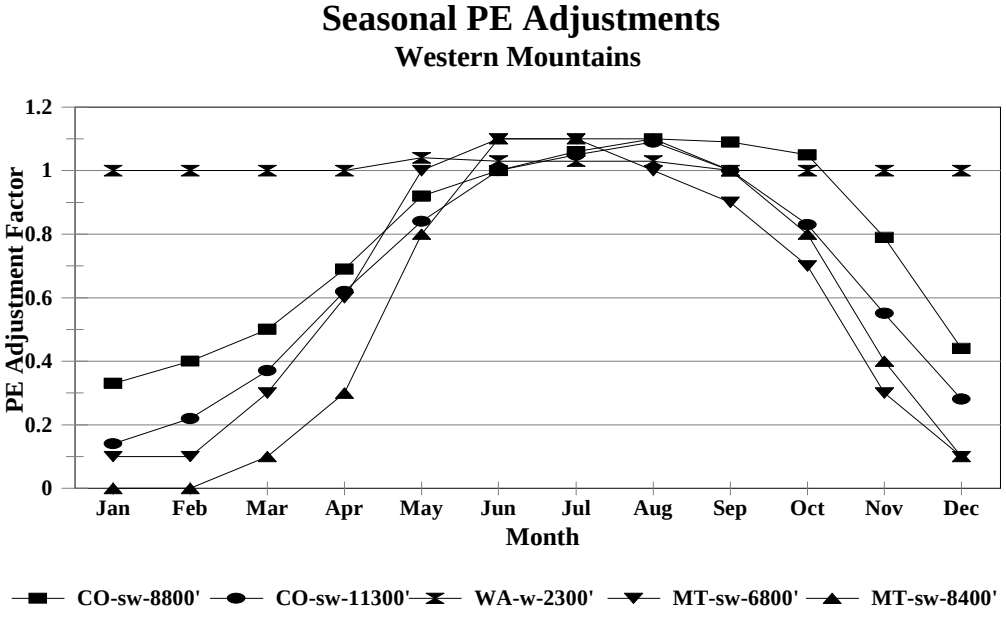


Figure 6-5-3. PE adjustments curves for watersheds in the western mountains.

Koren et al. [1998] compared NDVI green vegetation fraction values extracted from NCEP global data sets to PE adjustment curves from watersheds representing a northern mixed forest, southern mixed forest, southern Appalachian highlands, and the southwest (these curves were included in a 1970's prepared handout on deriving initial parameter values for the Sacramento Model). A reasonable correlation was found between the green vegetation fraction and the PE adjustment factors. This resulted in an equation for estimating the PE adjustment factors from the NDVI green fractions. This method can be used to derive a seasonal PE adjustment curve, but the method may not be applicable to the whole country. This method is included in CAP.

Non-Mountainous Area ET Demand Estimates

Procedure

For non-mountainous areas of the United States the map of annual FWS evaporation (map 3) in NOAA Technical Report NWS 33 has been adopted as the standard for long-term average PE values. The annual FWS evaporation values from this map provide an estimate of annual PE which when properly prorated to monthly values and combined with a reasonable seasonal adjustment curve gives ET Demand estimates that generally produce a realistic water balance. Thus, all PE estimates used in non-mountainous areas should first be adjusted to conform to the annual FWS values given in Technical Report 33.

The steps to follow for determining initial ET Demand values for each modeling area in a non-

mountainous region are:

1. Obtain monthly estimates of PE as described earlier in this section based on meteorological variables or pan data and adjust the annual total to correspond to the value picked off map 3 of Technical Report 33. The PE may be in the form of daily values or monthly averages. The PE estimate can be based on a single station, a weighted average of several stations, or computed by using the MAPE program. The MAPE program can combine daily PE computed from meteorological variables and adjusted daily pan measurements to get a mean areal value. The MAPE program contains the option to make consistency checks and corrections.

2. Determine a seasonal PE adjustment curve as described earlier in this section. This curve is always defined at the mid point of each month.

3. Compute ET Demand. If climatological average values of ET Demand are to be used, the average PE for each month is multiplied by the PE adjustment for the month to get values of ET Demand that are assigned to the mid point of each month. If daily estimates of ET Demand are to be used by the model, the input is a daily time series of PE and the seasonal PE adjustment curve defined at the mid point of each month.

The only available guidance regarding the use of daily PE values as opposed to mean monthly estimates comes from a brief HRL study [*Anderson and Farnsworth,* 1981]. In this study simulations from three watersheds using both daily PE values with a seasonal PE adjustment curve and mean monthly ET Demand estimates were compared. The average monthly ET Demand values in both cases were kept exactly the same. The results are shown in Table 6-5-1. There was very little difference in simulation results in North Carolina, a little improvement when using daily PE values in Mississippi, and more improvement, especially in regard to monthly volume errors, for the Oklahoma watershed. In general, two conclusions can be made from this study. First, PE data has more effect on volume computations than on daily flow calculations. Second, the use of daily PE data improves simulation results more in areas where there is a significant variation in PE amounts from one year to another. The greatest variability in annual PE values in the United States is in the southern plains.

During calibration the amount of ET Demand may need to be adjusted in order to obtain a good water balance. Estimates of long-term average precipitation should be quite reliable in a non-mountainous region, thus variations in ET Demand from one watershed to another should be related to differences in vegetation, as long as one doesn't neglect water balance terms such as diversions and deep aquifer recharge. When making adjustments in ET Demand, it is best to monitor how the seasonal PE adjustment curve is changing to make sure the transition from one month to another is realistic and that the differences in relationships from one watershed to another make sense.

	Daily RMS error (cms)		Monthly Vol. RMS error (mm)		
Watershed	Daily PE	Monthly Means	Daily PE	Monthly Means	Standard Deviation of Annual PE (mm)
Bird Ck. nr Sperry, OK	18.02	18.97	3.97	5.13	131.3
Leaf R. nr Collins, MS	15.76	16.02	7.54	7.91	89.3
Neuse R. nr Northside, NC	6.73	6.74	5.61	5.71	23.1

Table 6-5-1. Statistical comparison of streamflow simulations using computed daily PE values versus simulations using mean ET Demand estimates.

Mountainous Area ET Demand Estimates

Introduction

In mountainous regions the areal estimates of average precipitation are more likely to be in error than in non-mountainous areas. The isohyetal analysis used may reflect the general trend in how precipitation varies over the region, but at the watershed scale there can easily be random errors. Thus, if one treats the precipitation derived from the isohyetal maps as "true" and forces ET Demand adjustments to create a good water balance, it very likely that the resulting ET Demand variation over the region will not make physical sense. There should be a reasonable variation in ET Demand with elevation since PE and vegetation are highly correlated with elevation. Thus the underlying assumption for the recommended procedure for determining evaporation in mountainous areas is:

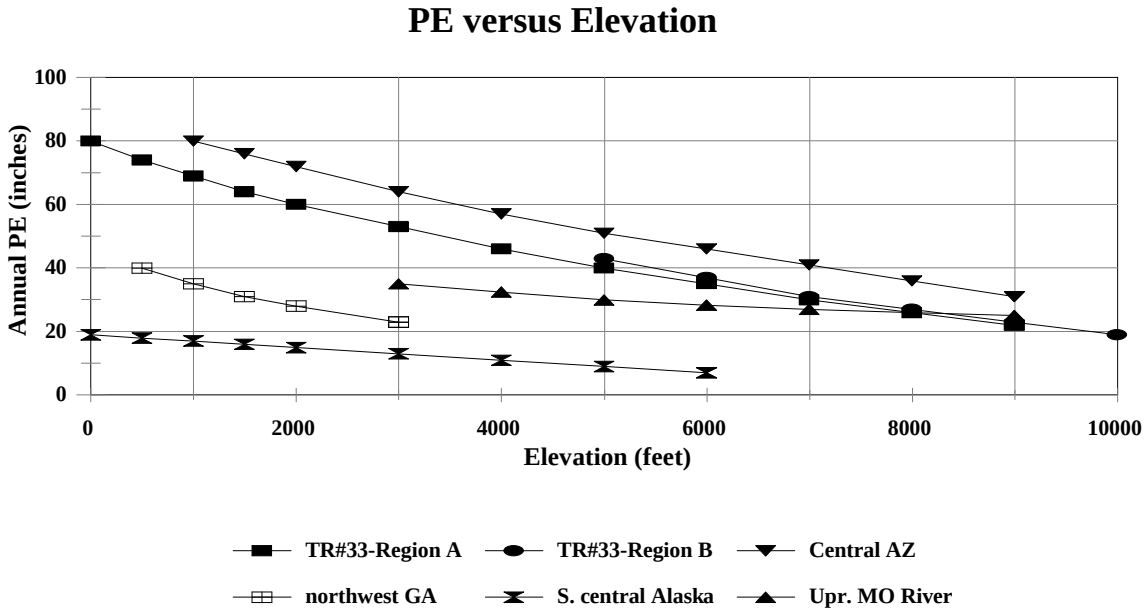
> *ET Demand variations are more spatially predictable and contain less error than precipitation variations, thus it is more logical to determine a reasonable ET Demand versus elevation relationship and then adjust MAP if needed than to assume MAP is correct and adjust ET Demand to obtain the correct water balance.*

Two approaches are given in this manual for determining the relationship between ET Demand and elevation. Which approach is used depends on whether the precipitation amounts from the available isohyetal analysis can be used to generate a reasonable water balance over most watersheds in the region. If so, water balance computations can be used to determine how actual ET varies with elevation and can serve as a basis for determining an ET Demand versus elevation relationship. If areal precipitation estimates derived from the available isohyetal maps do not produce a reasonable water balance over most watersheds in the region, then it is best to derive the variation in ET Demand with elevation from information on the variation of evaporation and vegetation over the region and combine this with runoff data to determine the amount of precipitation needed to produce the proper water balance. One way to judge the quality of the

precipitation estimates from an isohyetal map is to compare the resulting actual ET values to PE estimates for the region. Actual ET is generally less than PE in most regions, though the ratio will vary depending on climate. The ratio of actual ET to PE is lowest in an arid region. In a forested area with frequent rains, actual ET could be greater than PE as mentioned earlier in this section. A snow cover over substantial portions of the year will tend to suppress ET, thus reducing the ratio.

Before describing methods of obtaining an ET Demand versus elevation relationship, we should first look at how PE can vary with elevation as a guide. In mountainous areas the annual FWS map in Technical Report 33 is not drawn to a sufficiently small spatial scale to define PE at the watershed level. That map shows the general trend of PE variations in mountainous areas, but not at the detailed level needed for hydrologic modeling. The amount of PE at a given location is affected by the climate of that location. In the mountains there can be small scale variations in PE due to terrain effects on radiation and wind, however, at the watershed scale much of the variation in PE can be explained by elevation differences. PE decreases with elevation with the rate of decrease becoming less as elevation increases. Figure 6-5-4 shows the variation in PE with elevation for several regions of the United States. The two relationships from Technical Report 33 were produced by plotting pan measurements at various elevations over the regions and then using the average pan coefficient for the region to convert the values to FWS evaporation. The Alaska relationship was generated from values computed by the Thornthwaite method for stations in south central Alaska by Patric and Black [1968]. The Upper Missouri River basin relationship was determined from a few pan sites and values picked off map 3 of Technical Report 33. The other two relationships were determined by plotting PE derived from

Figure 6-5-4. PE versus elevation for various regions of the United States.



pan measurements and meteorological variables. The magnitude of PE varies from one region to another due to climatic differences, but the general shape of the curves are similar.

In many regions the shape of the ET Demand versus elevation relationship should be somewhat similar to the shape of a plot of PE versus elevation. This will occur in regions where there is significant vegetation at all elevations and the transition from one type of vegetation to another is not dramatic. Such regions are those in the eastern part of the country and west of the Pacific crest. In other regions the variation in vegetation types and thus PE adjustments are so great that the shape of a ET Demand versus elevation plot may not be similar to the PE versus elevation relationship. Watersheds in such regions may include various subsets of transitions from arid climate vegetation such as cactus and sagebrush to forested zones to alpine tundra to extremely high elevations with no vegetation at all. Such variations can occur in the intermountain west and Alaska. In the northern portions of the intermountain west, the ET Demand versus elevation relationship is likely similar to a PE versus elevation plot over the range in elevations where most of the runoff is generated. In addition to being difficult to determine, it probably is not that important to know the exact ET Demand versus elevation relationship over elevation ranges where little or no runoff is generated.

Determining ET Demand from a Water Balance Analysis

In order to use this approach to determine a ET Demand versus elevation relationship for a river basin, water balance computations for drainages within the basin should produce a plot of actual ET versus elevation that shows a clear relationship. There may be a few watersheds within the basin that deviate from the relationship defined by the others due to possible errors in the isohyetal analysis (as in Figure 6-3-5), but the results from most of the drainages should show a clear and realistic relationship. If there is no clear relationship or if the resulting actual ET values are not reasonable, then the second approach (i.e. using evaporation, vegetation, and runoff data) should be tried. The water balance approach works best for relatively wet regions where the ratio of actual ET to ET Demand is greater than about 0.75 or if the ratio is less, it is due to a long period of snow cover. In semi-arid regions where this ratio is lower, it is more likely that there will be considerable scatter in the actual ET values derived from water balance computations.

The steps to follow when trying to derive a ET Demand versus elevation relationship from the results of water balance computations are as follows:

1. Perform water balance computations for appropriate headwater and local drainages within the river basin to determine estimates of actual annual ET. Guidelines for water balance computations are in Section 6-3 under the section “Determination of Average Mean Areal Precipitation for Each MAP Area.”

2. Plot actual annual ET values derived from the water balance computations in step 1against the mean elevation of each drainage area (Figures 6-3-5 and 6-5-5 are examples of such

plots). If most of the points define a clear relationship that is physically realistic, then draw a line that defines the actual annual ET versus elevation relationship for the basin. If the plot exhibits a lot of scatter or the values are physically unreasonable, then try the second approach to determining an ET Demand versus elevation relationship.

3. If possible, use pan data or estimates based on meteorological variables to determine annual PE values for sites at different elevations in the vicinity of the basin. Plot these points and draw a PE versus elevation line. This step is optional, but can be helpful as a guide for determining the shape of the ET Demand versus elevation relationship.

4. Obtain a point on the ET Demand versus elevation relationship by calibrating the initial headwater basin (recommended strategy for calibrating the drainages in a river basin is discussed in chapter 7) to determine the average annual ratio of actual ET to ET Demand. This is done through an iterative process.

a. Make an assumption as to the average annual ratio of actual ET to ET Demand to determine an initial estimate of the annual ET Demand for the watershed. If the headwater has multiple elevation zones, the ratio generally increases with elevation when the typical length of snow on the ground is not substantially different between zones. When one zone has snow for a much longer period than others, its ratio will typically be smaller due to suppression of evaporation when a snow cover exists.

b. Using estimates of monthly PE and the seasonal PE adjustment curve generate monthly ET Demand values (could use daily PE data, if available) for this drainage.

c. Adjust the seasonal PE adjustment curve until the resulting annual ET Demand is the same as the ET Demand which was based on the assumed ratio of actual ET to ET Demand. Spreadsheets are an ideal tool for these calculations.

d. Calibrate the headwater area using these monthly ET Demand values. As the calibration proceeds, check the resulting average annual ratio of actual ET to ET Demand and revise the estimates of ET Demand based on the actual value of the ratio. This will typically have to be done once or twice.

e. Finally, take the resulting annual ET Demand value and plot it versus elevation on the graph produced in step 2.

5. Using the value of annual ET Demand from the initial headwater area, draw an initial estimate of the curve showing how ET Demand should vary with elevation over the river basin. The actual annual ET and PE versus elevation plots can be helpful in drawing this curve. Also how the ratio of actual ET to ET Demand likely varies with elevation should be taken into account.

6. Verify the ET Demand versus elevation curve by using annual ET Demand from this relationship to calibrate other headwater areas at different elevations. Adjust annual ET Demand and thus the curve if the resulting ratio of actual ET to ET Demand is unrealistic.

7. Use the resulting annual ET Demand versus elevation relationship when calibrating all of

the other drainages within the river basin. Seasonal PE adjustment curves used to prorate the annual ET Demand into monthly values should reflect changes in the vegetation types and activity patterns with elevation. Some scatter about the defined relationship is allowed based on the location of the drainage within the basin (e.g. areas in the northern part of a river basin might be assigned a slightly lower annual ET Demand for a given elevation than an area in the southern part of the basin). During the calibrations, corrections to the long-term water balance should be made by adjusting the amount of precipitation rather than changing the ET Demand curve.

This procedure may be somewhat difficult to follow, thus it will be illustrated step by step using the Merrimack River Basin in New Hampshire and Massachusetts.

Step 1 – Water balance values were computed for all appropriate headwater areas and a couple of local areas within the Merrimack Basin. The WATERBAL operation was used to calculate the water balance components (requires the inclusion of the SNOW-17 and SAC-SMA operations -- nominal parameter values were specified). The MAP time series were based on the PRISM estimates of annual precipitation. The PXADJ and SCF model parameters were set to 1.0 so that there would be no changes made to the MAP values. The water balance values are shown in Table 6-5-2.

Step 2 – The actual ET values generated by the water balance analysis were plotted against the mean elevation of each watershed as shown in Figure 6-5-5. There is a reasonable relationship between actual ET and elevation. Some of the scatter is undoubtedly caused by climate differences resulting from variations in latitude and distance from the coast and some is due to sight variations in vegetation from one area to the next, however, a major portion of the scatter could be caused by inaccuracies in the PRISM isohyetal analysis.

Step 3 – There were two pan evaporation sites and two locations with evaporation computed from meteorological factors within the basin. All the sites were at similar elevations, thus it was not possible to construct a PE versus elevation relationship. The available data were used to derive an estimate of mean monthly PE that was thought to be realistic for a 500-1000 foot elevation in the central part of the basin. This seasonal variation in PE is shown in Figure 6-5-6. This PE corresponds to an annual total of 27.3 inches.

Step 4 – The Smith River near Bristol, NH (BRSN3 - mean elevation of 1117 feet) was chosen as the initial headwater area to be calibrated for the Merrimack Basin. This watershed has a single elevation zone. As far as arriving at a final ET Demand estimate for BRSN3, the steps were as follows:

4.a – A value of 0.75 was used as the initial estimate of the ratio of actual ET to ET Demand for BRSN3. This was just a guess.

4.b& c – An initial guess of the shape for the seasonal PE adjustment curve was made for BRSN3. This curve was then adjusted so that when the values were multiplied by the

monthly PE shown in Figure 6-5-6 to get ET Demand, the resulting annual ratio of actual ET to ET Demand was about 0.75. This initial PE adjustment curve is shown in Figure 6-5-7. The resulting monthly ET Demand values are shown in Figure 6-5-6 and represent an annual total of about 25.2 inches.

Watershed	Period(WY)	Pcpn	Runoff	ET
BRSN3	49-77 49-93	41.6 41.9	23.1 23.2	18.5 18.7
PTRN3	49-77	43.8	23.4	20.8
DVSN3	49-78	42.9	22.6	20.3
SOUN3	52-77 52-87	39.7 40.4	19.5 20.0	20.2 20.4
SOHN3	49-76	44.0	23.0	21.0
WGTM3	49-77 49-92	43.8 44.8	22.4 23.4	21.4 21.4
GFFN3	49-78	42.3	21.0	21.3
NCIN3	49-70	42.5	20.6	21.9
FBGM3	73-93	48.4	26.3	22.1
SAXM3	81-93	47.1	25.0	22.1
MAYM3	49-77 49-93	45.0 45.7	22.1 22.5	22.9 23.2
WLMM3	65-93	44.7	21.8	22.9
RUMN3	49-77	39.0	24.2	14.8
WOON3	49-77	49.7	36.5	13.2
HENN3(local)	49-77	42.0	23.8	18.2
PLMN3(local)	49-77	45.2	28.1	17.1

Table 6-5-2. Water Balance Analysis for the Merrimack Basin (values are in inches).

4.d – The BRSN3 watershed was calibrated following the guidelines given in chapter 7. Early on in the calibration it became clear that the ratio of actual ET to ET Demand for this watershed had been underestimated. It was evident that the ratio was more in the range of 0.85-0.9 rather than the initial guess of 0.75. Thus, the PE adjustment curve was

altered and new ET Demand values computed. When the calibration was completed, the ratio of actual ET to ET Demand was 0.87. The final PE adjustment curve is shown in Figure 6-5-6. The final monthly ET Demand values, which resulted in an annual total of 21.7 inches, are shown in Figure 6-5-7.

4.e – The annual ET Demand for BRSN3 was plotted versus elevation on the plot produced in step 2 as shown in Figure 6-5-5.

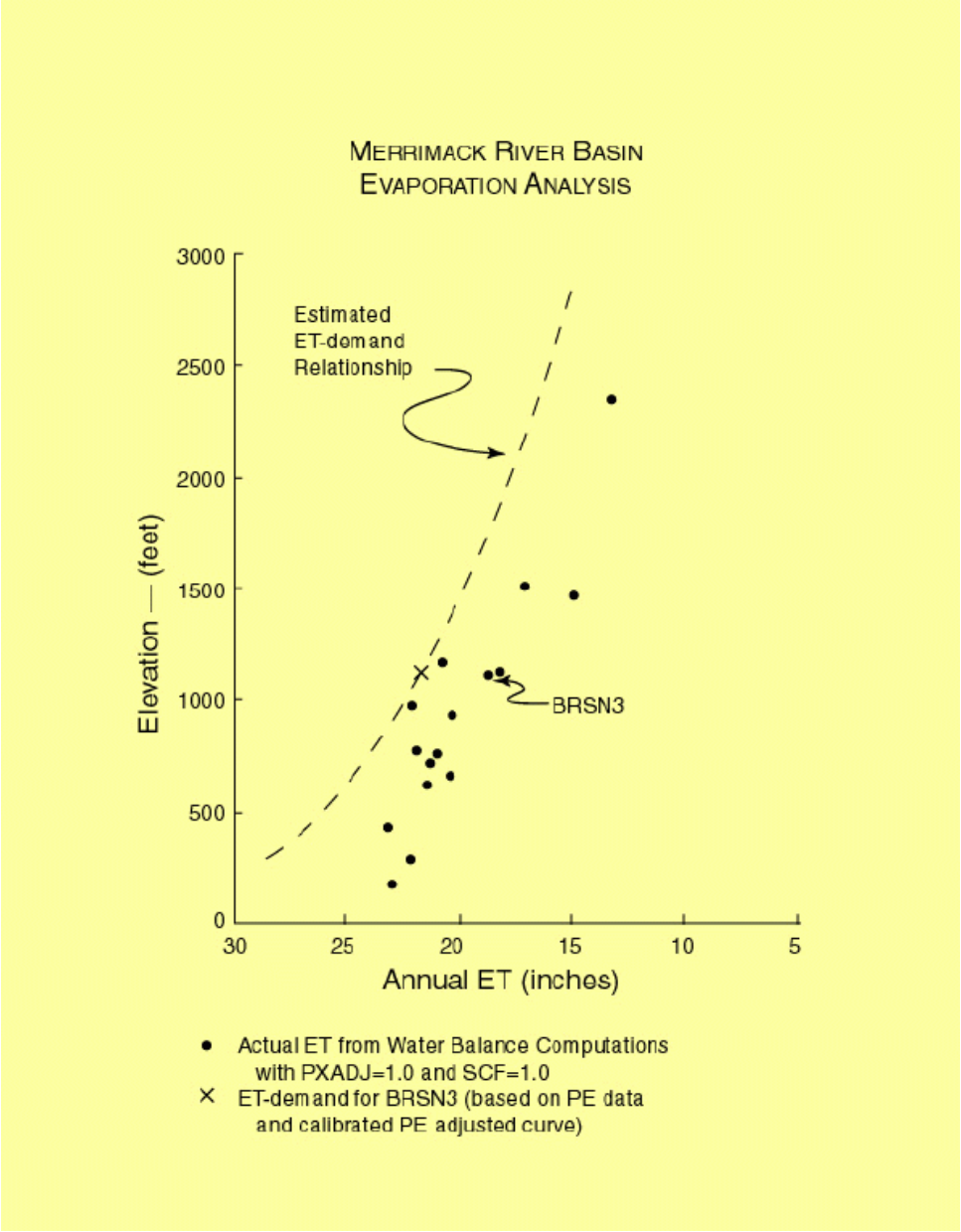
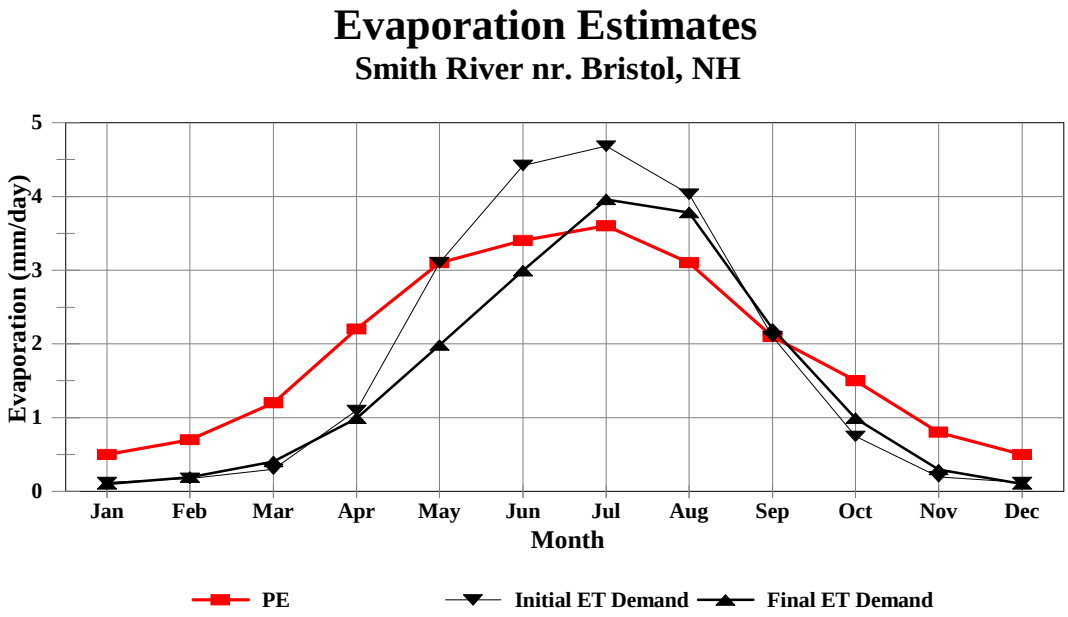
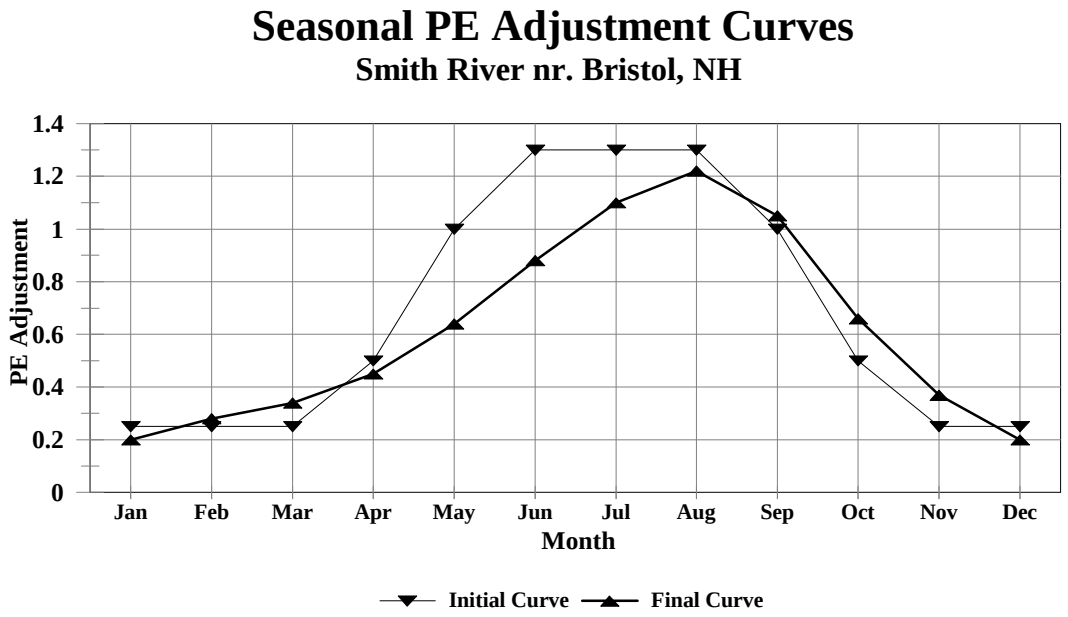


Figure 6-5-5. Evapor





ation versus elevation relationships for the Merrimack River basin.

Figure 6-5-6. Evaporation Values for the Smith River near Bristol, NH (BRSN3).

Figure 6-5-7. Season PE Adjustment Curves for BRSN3.

Step 5 – An estimate of the relationship between ET Demand and elevation was drawn for the Merrimack Basin using the trend in the actual ET versus elevation plot as a guide. This relationship is shown in Figure 6-5-5.

Step 6 – To test the validity of the estimated relationship between ET Demand and elevation derived in step 5, model simulation runs were made for two other headwater areas at different elevations. These were the Pemigewasset River near Woodstock, NH (two subareas with mean elevations of 1465 and 2846 feet) and the Assabet River near Maynard, MA (mean elevation of 448 feet). Annual ET Demand values for these watersheds were picked off the curve shown in Figure 6-5-5 and prorated to monthly values. The proration was based on modifying the shape of the seasonal PE adjustment curve for BRSN3 for perceived variations in vegetation activity and type from one area to another and adjusting the values so that when multiplied by the monthly PE shown in Figure 6-5-6, the resulting ET Demand had the correct annual total (variations of PE with elevation are implicitly absorbed into the seasonal adjustments). In deciding whether the estimated relationship between ET Demand and elevation was realistic, the higher and lower elevation watersheds were not calibrated, but instead the parameters determine for BRSN3 were used for the simulations. This seemed reasonable at this point in the calibration process since an assessment of hydrologic variability had suggested that model parameters should be quite similar over the entire Merrimack Basin (see Figure 4.2). These simulations resulted in variations in the ratio of actual ET to ET Demand to vary from around 0.85 at the lowest elevation to slightly over 0.9 at the highest elevation. This seemed quite realistic.

Step 7 – The annual ET Demand versus elevation relationship shown in Figure 6-5-5 was used to calibrate the other watersheds within the Merrimack Basin. Some slight deviations off this relationship were made to account for the location of each watershed within the basin (e.g. slightly different ET Demand values were used when watersheds at similar elevations were located at different latitudes). Figure 6-5-8 shows the actual annual ET Demand values used in the final calibration of each watershed. The figure also indicates those watersheds for which precipitation values needed to be adjusted in order to obtain the proper water balance (this was done by using PXADJ in the SNOW-17 operation). As mention previously it is more reasonable to establish a realistic relationship between ET Demand and elevation and then adjust precipitation if needed than to treat the isohyetal map as the truth and alter the ET Demand to get the proper water balance. The isohyetal analysis was quite good for this river basin such that none of the MAP time series had to be adjusted by more than 5%.

Determining Precipitation and Actual ET from Runoff, Evaporation, and Vegetation Information

This approach uses estimates of how ET Demand varies with elevation to determine the average amount of precipitation and actual ET over a river basin. This approach is primarily for use in areas where the isohyetal estimates of annual precipitation are not considered realistic. This is frequently the case in regions with very sparse precipitation networks such as Alaska (see section "Determination of the Average Mean Areal Precipitation in Data Sparse Regions" in Section 6-

3). This approach could also be tried in regions where there is a lot of scatter in the actual ET versus elevation plot determine d from water balance computations. This approach relies on using runoff data and an estimate of actual ET to determine the amount of average annual

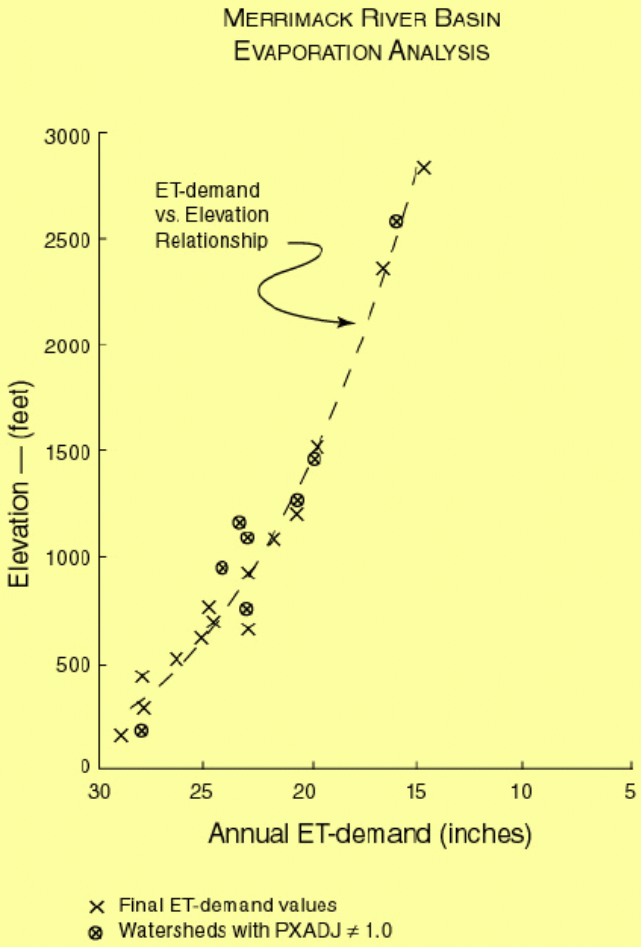


Figure 6-5-8. Final ET Demand values used for the Merrimack River basin.

precipitation. In Alaska this method can give more reliable estimates of average annual precipitation than estimates based on precipitation gage data because the precipitation network is sparse and the amount of actual annual ET is relatively small as compared to annual runoff. If this approach was used in an area where evaporation is much greater than runoff, such as the central Arizona mountains, it would probably not produce precipitation estimates that were as good as those determined by an isohyetal analysis using only precipitation gage data even though the water balance estimates of actual ET using the isohyetal map information might show considerable scatter. In such an area it might be possible to use a blend between the two approaches to estimate how ET Demand should vary with elevation and to refine the precipitation pattern over the region.

The steps to follow when deriving precipitation and ET Demand versus elevation estimates from evaporation and runoff data are as follows (again a spreadsheet is ideal for doing these computations):

1. Determine a relationship between annual PE and elevation.

2. Select watersheds within the river basin with good runoff data and adjust for any diversions or other gains or losses.

3. For the initial headwater area to be calibrated determine the annual PE at the mean elevation of the drainage. If multiple elevation zones are being used, determine the annual PE value at the mean elevation of each zone.

4. Prorate the annual PE amounts into average monthly values.

5. Subjectively determine values for the seasonal PE adjustment curves based on the vegetation types and activity patterns within the watershed or each elevation zone. These curves may vary considerably with elevation if the vegetation cover and climate changes substantially with elevation.

6. Multiply the average monthly PE times the seasonal PE adjustments to get monthly ET Demand estimates at each elevation. From these monthly values compute the annual ET Demand.

7. Make an assumption as to the ratio of actual ET to ET Demand for the watershed or each zone.

8. Estimate the actual ET by multiplying the ratio from step 7 times the annual ET Demand

values from step 6. If there are multiple elevation zones, weight each actual ET value by the portion of the watershed that it represents to get the annual actual ET for the entire watershed.

9. Add the annual actual ET to the average annual runoff to get an estimate of the average annual precipitation. If there are elevation zones, this precipitation can be prorated to each zone by using the precipitation pattern shown in the isohyetal analysis. MAP time series can then be generated that contain these average annual precipitation amounts by using the mountainous area precipitation procedure described in Section 6-3.

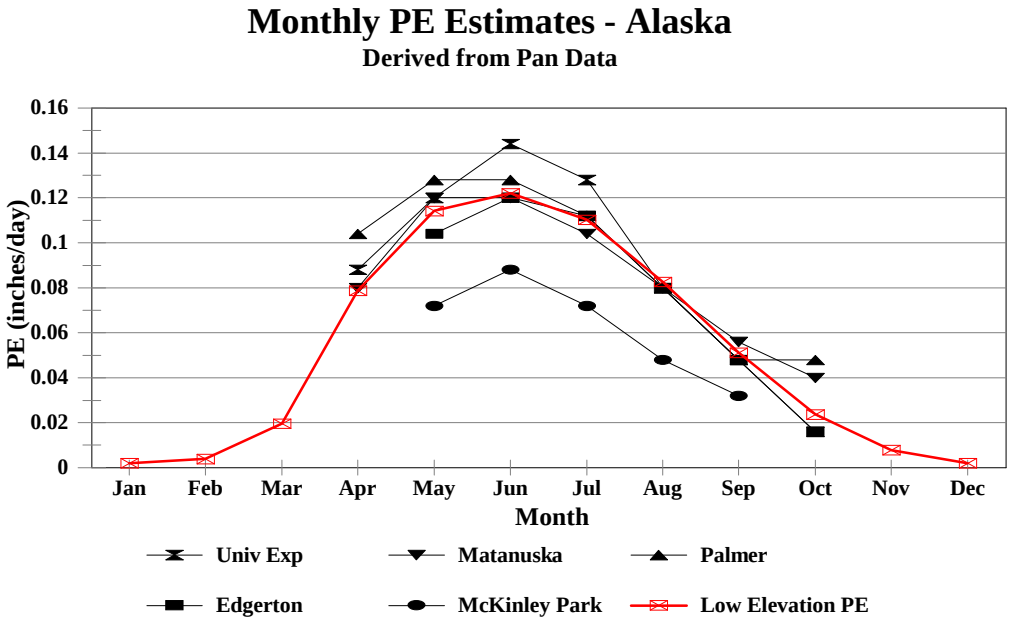
10. Calibrate the watershed using the MAP time series and the ET Demand values from step 6. The MAP time series will typically have to be regenerated a couple of times during the calibration process as the model values of the ratio of actual ET to ET Demand are more precisely determined. Also the distribution of precipitation with elevation may need to be altered if there is evidence that the initial pattern is unrealistic. The pattern of how precipitation is distributed with elevation can often be verified by comparing the amount of snow generated in various zones with available snow observations or by comparing modeled glacial changes with reported trends.

11. Calibrate other watersheds in the river basin using steps 3 through 10. Utilize the values used and determined during previous calibrations when working on new drainages. A single ET Demand versus elevation relationship for the entire river basin should emerge, as well as estimates of average annual precipitation for various watersheds and elevation zones. The ratio of these precipitation estimates to those obtained from the isohyetal analysis should be able to be used to adjust the isohyetal maps for portions of the basin where calibration is not possible.

To illustrate this approach we will use the Chisana River near Northway, Alaska. This 3280 mi^2 watershed is located on the north side of the Wrangell Mountains and varies in elevation from about 1700 to over 10000 feet with an average elevation of about 3500 feet. About 6% of the watershed is covered by glaciers. There is one precipitation station with a long record at the outlet and 2 others with short records at lower elevations in the watershed and about a dozen other such stations, all below 3200 feet and many with short records, scattered over a large area surrounding the watershed. A comparison of the PRISM estimate of precipitation for the watershed of 21 inches as compared to measured runoff of 9.7 inches results in an actual ET of 11.3 inches. This actual ET value was considered to be somewhat high given the length of the snow cover period over much of the watershed, thus it was decided to derive an estimate of average annual precipitation from runoff data and evaporation estimates. This was done in the following manner:

1. Monthly PE values were derived for low elevations in south central Alaska from pan data and an estimate of PE during the winter months. This seasonal PE curve and the pan sites used to create the curve are shown in Figure 6-5-9. The annual PE represented by this curve is 19 inches.

2. Using the shape from this low elevation seasonal PE curve and the relationship between



PE and elevation shown for south central Alaska in Figure 6-5-4, monthly PE curves were constructed for the mean elevation of the lower and upper zones in the Chisana watershed (glacier zone is considered to have no evaporation). These curves are shown in Figure 6-5-10. The annual PE for the lower zone with a mean elevation of 2400 feet is 14.3 inches and the annual PE for the upper zone with a mean elevation of 6000 feet is 6.7 inches.

3. Seasonal PE adjustment curves were subjectively estimated for each elevation zone as shown in Figure 6-5-11. The lower zone is primarily muskeg, willows, and black spruce. The upper zone is above the tree line with a mixture of tundra and barren areas.

4. Seasonal ET Demand curves were then calculated by multiplying the average daily PE for each month times the PE adjustment for that month. The resulting ET Demand curves are shown in Figure 6-5-12. The total annual ET Demand based on these curves is 8.5 inches for the lower zone and 2.0 inches for the upper zone.

Figure 6-5-9. Low elevation seasonal PE curve for south central Alaska.

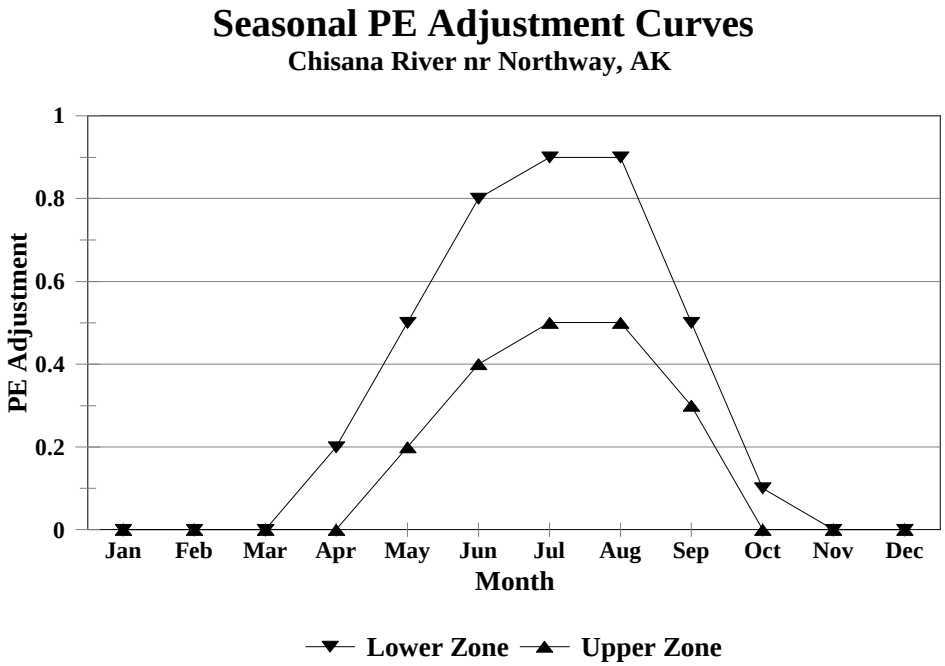


Figure 6-5-10. Seasonal PE curves for Chisana watershed elevation zones.

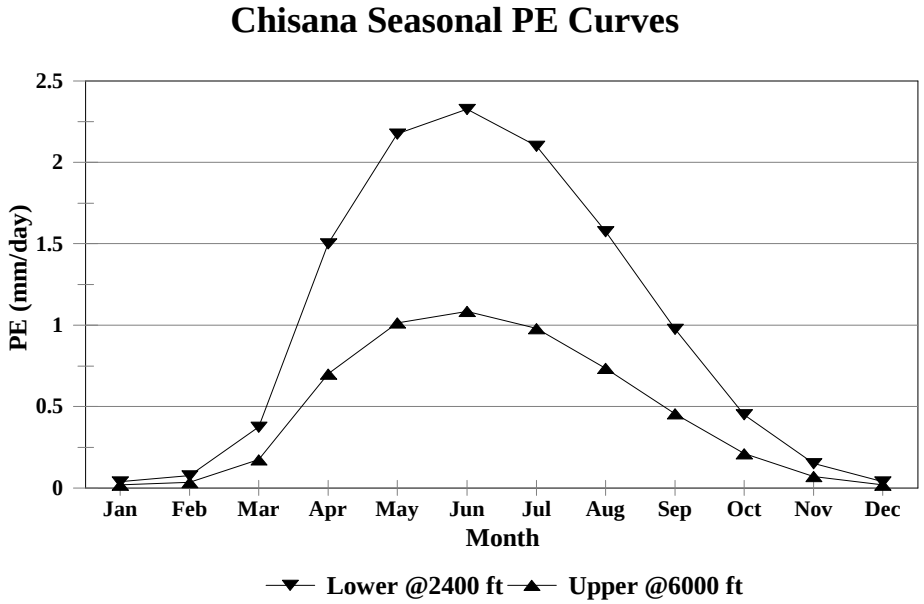
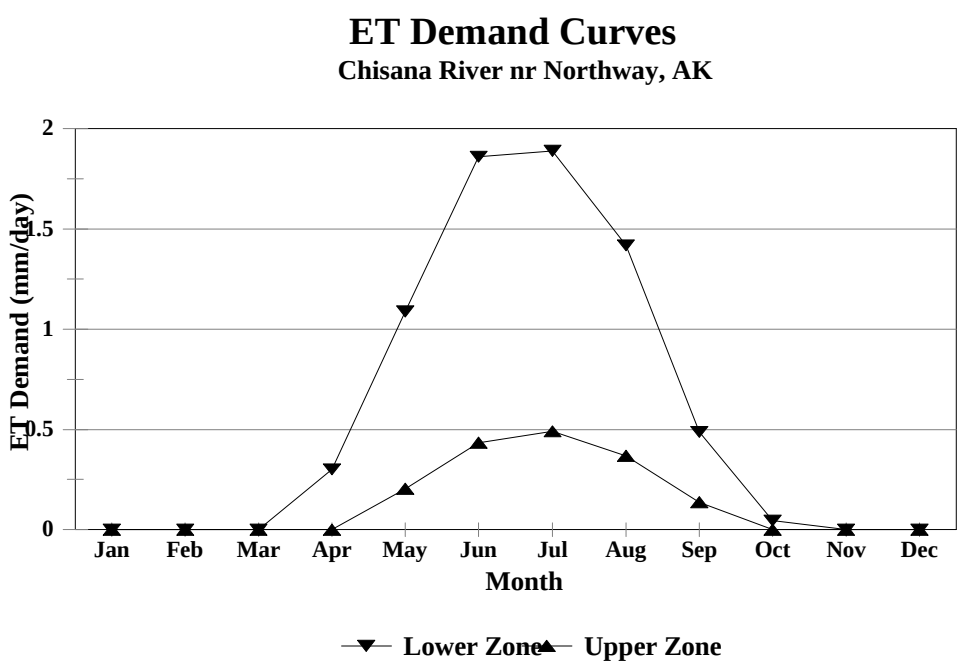


Figure 6-5-11. Seasonal PE adjustment curves for the Chisana River.

Figure 6-5-12. ET Demand curves for the Chisana River.



5. The next series of steps were to estimate the ratio of actual ET to ET Demand for each zone, use those values to calculate annual actual ET for the watershed, add annual actual ET to the average annual runoff to estimate the annual average precipitation,

generate MAP time
series based on this precipitation amount, and calibrate the watershed using this time series and ET Demand curves from step 4. Based on the calibration, the estimates of the ratio of actual ET to ET Demand were revised and new precipitation time series generated. The initial guess as to the ratio of actual ET to ET Demand was 0.75 for each zone. Based on the initial calibration, the ratio was adjusted to 0.8 for the lower zone and 0.6 for the upper zone. These values of the ratio gave annual actual ET values of 6.8 and 1.2 inches for the lower and upper zones, respectively, and a watershed average actual ET of 5.2 inches (lower zone comprises 72% of the area, the upper zone 22%, and the glaciers 6%). The average annual runoff was 9.7 inches which when added to the annual actual ET gave a watershed average precipitation of 14.8 inches (PRISM estimate was 21 inches). Elevation zone values of annual precipitation were determined by multiplying the watershed average by the ratio of zone precipitation to watershed precipitation from the PRISM analysis. These ratios were 0.73, 1.53, and 2.29 for the lower, upper, and glacial zones, respectively, which resulted in average annual precipitation values for the zones of 10.8, 22.7, and 34.0 inches. These values produced a reasonable simulation of streamflow and seasonal snow cover for the watershed and changes over time to the glaciers seemed realistic.

This approach was also used in the Upper Kenai River basin in Alaska to get annual precipitation estimates that would produce a reasonable water balance and actual ET values. For the Trail River near Lawing this approach resulted in an actual annual ET of 4.2 inches and an annual precipitation of 69.5 inches. The PRISM annual precipitation for the watershed was 96.7 inches.

Mountainous Area Evaporation Summary

No matter which of the two approaches in this manual are used to get evaporation estimates in a mountainous region or whether some other approach is followed, it is essential to end up with evaporation estimates that are physically realistic in both magnitude and spatial variability over the river basin. With the real potential for errors in precipitation amounts derived from any isohyetal analysis at the watershed scale, one cannot assume that the precipitation amounts are correct and then adjust evaporation to whatever values are needed to produce a near zero water balance. This approach is guaranteed to produce unrealistic estimates of evaporation. Therefore an approach must be followed that uses evaporation values that are realistic and vary logically with elevation. Any adjustments that may be needed to produce a near zero water balance should be made to the precipitation estimates.

Section 6-6

Mean Daily Flow Data Checks and Adjustments

Consistency Checks

Like precipitation, temperature, and evaporation data that are used to estimate input time series for model calibration, the streamflow data that are used to verify the simulation results should also be checked for consistency. Such a check can serve two purposes. First, a consistency check can reveal inconsistencies caused by factors that are not known to the user. Second, the check can determine whether known changes within the basin have a significant effect on the amount of runoff measured at the gage. Inconsistencies in streamflow data can occur for a variety of reasons including such things as changes in land use (e.g. rural to suburban, agricultural to forest, drainage of wetlands), the construction of new control structures or modifications in the operation of these structures, changes in the amount of water diverted into or out of the watershed, the effect of large forest fires, alterations in agricultural practices (e.g. changes in crops being grown, addition of drain tiles, modifications to irrigation practices), constantly shifting rating curves that aren't updated as often as they should, and changes to measurement methods or practices. Consistency checks could reveal information on both the magnitude and timing of these effects and determine whether the data may need to be adjusted or whether only portions of the period of record should be used for determining model parameters.

Consistency checks of mean daily flow data would be done using double mass analysis as it is for precipitation data. Prior to generating the double mass plots, the daily flow data needs to be converted to depth of runoff or at least the values from all the gaging sites need to be scaled to a common area. As with precipitation, the deviation of the accumulated runoff from the group average for each station would be plotted against the accumulated average runoff for the group. Unlike precipitation, only selected stations should be included in computing the group average. The group average should be computed using only stations that have consistent data, i.e. stations with known significant changes in land use, diversions, control structures, agricultural practices, etc. should not be part of the group average. Stations should be grouped geographically on individual plots as there can be shifts in the runoff relationships between watersheds due to changes in climatic patterns. Such changes are real and no adjustments are needed. Such real shifts would be indicated by having several stations in the same general area showing similar patterns of change in their double mass plots.

In most cases inconsistent runoff data would not be corrected by applying a simple multiplying factor as is done with precipitation. Generally the double mass plots would indicate that an inconsistency might exist and give an idea of its magnitude and timing. Then the reason for the inconsistency needs to be determined by looking into the history of changes that have occurred within the watershed. Once the reason is known, then a decision can be made as to how to adjust the data or whether only a portion of the data record can be used for calibration.

Even though the direct capability for making consistency checks on mean daily flow data doesn't currently exist within the NWSRFS software, there is a way that such checks might be made with some limitations using the existing programs. This would involve the following steps:

1. Convert the mean daily flow data to depth of runoff and create time series. This can be done by using a LOOKUP operation for each streamflow site. The LOOKUP operation allows the user to specify the relationship between any two variables without having to maintain the same dimension or units. Two points are sufficient in this case to define the relationship (e.g. the 0.0,0.0 point and a point that indicates the flow volume that represents 1.0 inch of runoff). The resulting runoff time series would be declared as OUTPUT time series.

2. Use these runoff time series as input to the PXPP program which will generate the information needed to produce a double mass plot (plot displayed either by PXPP or IDMA). The PXPP program will treat these time series as if they were precipitation data. In general this shouldn't matter as far as the double mass analysis is concerned except that the procedure in PXPP for estimating periods of missing data may not be totally appropriate for runoff data, thus one needs to avoid missing data periods when drawing conclusions regarding the consistency of the streamflow data. The missing data estimation procedure in PXPP may work in regions where most of the runoff comes from rainfall, but will undoubtedly have more problems in regions where the timing of snowmelt varies considerably from one gaging site to another. Runoff consistency checks in these regions may be possible using PXPP only during periods when all stations included in the analysis have observed data. Another problem in using PXPP to check the consistency of runoff data, is that PXPP doesn't currently have the capability to eliminate certain stations from the computation of the group average (this feature exists in the MAP program, but the MAP strategy for estimating missing data involving both hourly and daily stations will not work with runoff data derived from mean daily streamflow records). Thus, discharge stations with known consistency problems should not be included as input to PXPP.

Streamflow Data Adjustments

There are two general cases that need to be dealt with regarding possible adjustments to streamflow data. The first case is where there are changes in runoff generation that occur slowly over time, such as those resulting from changes in land use or agricultural practices or from natural events like a large forest fire (fire occurs suddenly, but recovery is slow). The second case is where there are modifications to natural flow conditions caused by man-made changes or natural transfers of water across watershed divides.

When the amount of runoff generated changes slowly over time, it is very difficult to make adjustments to the data. In these cases, the normal solution is to use only the period of record that most closely reflects the current state of the watershed when determining model parameter values via calibration. Generally this is the most recent period, though in some cases the most

recent period could be an aberration. For example, if a large forest fire occurred a few years ago and now the watershed has just recovered from the effects of the fire, it would be more appropriate to use the period prior to the fire for calibration.

When there are modifications to flow conditions due to man-made control structures or natural transfers across watershed divides, there are two general alternatives. The first option is to model all of the flow modifications so that the simulation results can be compared directly to the observed streamflow data. This would include modeling of such things as reservoir operations, effects of large lakes, diversions across watershed divides whether controlled by man or by nature, irrigation usage and return flow, and municipal or industrial use within the watershed boundaries. The capabilities for modeling a number of these exist within NWSRFS. If modeling these flow modifications is attempted during calibration, the data required should be available not only historically, but also in real time so that the calibration results can be directly applied operationally. If the real time operation of a control structure differs from the rules used during calibration, the changes would need to be included in the operational setup. There are definite advantages to modeling the flow modifications whenever possible. One advantage is that during operational use there is much less reliance on agencies that operate reservoirs or diversions to provide information on releases or diversion amounts. This results in much more timely forecasts at downstream locations. Another advantage is that probabilistic predictions can be extended further downstream, though the uncertainty of each modeled feature needs to be reflected in the total uncertainty of the prediction at each location.

The second option, when there are man-made or natural flow modifications, is to adjust the historical streamflow data to reflect natural flow conditions. Then the calibration results are compared to this “observed” natural flow. Operationally either the forecast can be natural flow values, or the simulated natural flow must be adjusted in some way based on real time information that is not available historically. On many western rivers where the regulation of flows is very complex based on current irrigation, municipal, industrial, fishery, and power generation demands, the reservoir and diversion operators want a prediction of the amount of natural flow generated from rainfall or snowmelt. Based on these predictions, they determine how the control structures are to be operated. This control information is then used by the RFC to create actual flow forecasts. Also in many western areas extended predictions of natural flow are used to make advance plans on items such as which acreage can likely be planted with crops and the amount of seed, fertilizer, and other agricultural products that probably will be needed. In other regions there are reservoirs and other control structures whose operations would be difficult to model historically. Operationally the RFC computes natural flow from the headwaters and locals and relies on the operating agencies to provide release information.

Some adjustments to historical streamflow data and possible methods of making these adjustments to generate natural flow time series for use when calibrating the models are listed below.

1. Diversions - these are diversions of water across watershed divides for various uses. The

ideal solution is to obtain a time series of daily diversion volumes that can be directly added or subtracted from the mean daily discharge at the downstream point. In some cases it may be appropriate to prorate how much of the diversion is applied to each day due to travel time from the diversion location to the river gage. Daily diversion time series are available in some cases along with the other streamflow data provided by the USGS. In other cases it may be necessary to request this information from whoever operates the diversion. In cases where daily diversion data are not available, there may be tabulations of monthly volumes that are published or easier to obtain from the diversion operator than daily data. These monthly volumes can be prorated into daily time series so that the mean daily flow data at downstream points can be adjusted. While the adjustments are not correct on a daily basis, at least the correct volume adjustment is made to the river data.

2. Reservoirs and lakes - this requires either the calculation of the natural inflow to the pool or the adjustment of downstream flows for storage changes and direct gains or losses such as those from rain or evaporation. The water balance equation for a lake or reservoir can be expressed as:

$$I - O + P \cdot A - E \cdot A + D = \Delta S \qquad (6\text{-}6\text{-}1)$$

where: I = Inflow,
O = Outflow,
P = Precipitation on water surface,
E = Evaporation from the water surface,
A = Water surface area,
D = Diversions into or out of the pool other than at the main outlet, and
ΔS = Change in Storage (each term is in volume units).

In some cases the amount of precipitation and evaporation may be small compared to the inflow and outflow volumes and can be neglected. Many reservoirs and most lakes don't have any canals or pipes that divert water into or out of the pool or these volumes may be small compared to the main inflow and outflow and thus can be ignored. In other cases diversion data or estimates should be used, as well as estimates of precipitation and evaporation. An MAP time series can be generated to estimate the precipitation falling on the water surface. If the surface freezes during the winter, a rain plus melt time series may be more appropriate. Evaporation losses could be estimated from daily FWS evaporation, ideally adjusted for changes in heat storage, or from calculations of climatological mean lake evaporation.

Natural inflow is computed from Equation 6-6-1 by measurements or estimates of the other quantities. Generally, at least outflow data must be available along with either pool elevation or storage data to compute the ΔS term. To adjust streamflow data at downstream locations for the effect of reservoirs or lakes would require the computation of the combined ΔS, P, E, and D terms in the equation. This quantity would be added to the measured discharge at

downstream points. In many cases the daily values of this quantity should be prorated to reflect routing effects between the reservoir or lake and the downstream locations.

3. Springs - a few watersheds in some parts of the country contain large springs which contribute a significant amount, at least at lower flow levels. Some of the water discharged by such springs undoubtedly originates within the watershed boundaries, but a large fraction may come from outside the immediate drainage. Where such large springs exist there is frequently some information on the magnitude and seasonal variation of the discharge. Sometimes it is periodic flow measurements just downstream from the spring published in the USGS Water Resources Data reports. It is also likely that the outflow from these springs varies from year to year based on changes in precipitation patterns. If sufficient information is available, either a daily time series can be created to represent the outflow from the spring or the CHANLOSS operation can be used to at least account for the general magnitude of the outflow and how it varies seasonally.

4. Flood overflows - in some locations there are overflows across drainage divides during high flow periods. This can occur when two rivers come very close to each other, when flood control dams are built near watershed boundaries, and in regions with very flat terrain. Such occurrences are usually mentioned in the remarks section of the USGS Water Resources Data reports. These reports may indicate when such overflows occurred and may provide information that would be helpful in calculating the magnitude of the flow that crosses the divide.

5. Irrigation Losses - the net loss from irrigation, i.e. evaporation loss minus return flow, is generally difficult to determine. There may be many places within a watershed where water is removed from the river for irrigation and records of withdrawals are difficult to obtain and time consuming to process if they can be obtained. There are at least three general ways for dealing with irrigation losses. The best method is to try to model the losses and the return flows with procedures like the CONS-USE operation. Another technique is to account for the general magnitude of the irrigation effects by subtracting or adding flows on a seasonal basis even though the timing and magnitude of the losses and return flows are not the same every year. This can be done with a feature like the CHANLOSS operation or even mimicked with the Sacramento Model RIVA parameter. The other method is to just recognize that irrigation exists within a basin and not alter model parameters to match observed flows during periods with irrigation effects.

Chapter 7

Step 5 - Calibration of Hydrologic Models

Introduction

Now that all the needed information is available, the variability of physical and climatological factors have been assessed, the calibration locations and periods of record to use were selected, and the data have been checked, analyzed, and put in the proper form, it is finally time to calibrate the hydrologic models. In order to simulate conditions over an entire river basin many models and procedures must be used. Most of these contain variables that must be determined. Some can be determined directly by analyzing physical or experimental data, such as reservoir storage - elevation relationships and spillway rating curves or the drainage area of a particular watershed or local area. Other variables, which constitute the majority, are model parameters that vary from one area to another based on changes in physical factors and climatology. Methods exist for estimating the parameters for many models in an *a priori* fashion based on various information. Such parameter estimates may be satisfactory for some applications, but seldom provide the accuracy needed for river forecasting. To use the models to produce reliable forecasts both in the short term and for extended periods into the future, requires that a thorough calibration be conducted to determine the appropriate values of the parameters.

This manual will primarily focus on strategies and procedures for calibrating the NWSRFS SNOW-17 and Sacramento soil moisture models, however, in order to compute the flow in the rivers, a number of other models must be used. Even for a headwater drainage with no complications, a model of the channel system is required to convert the runoff into the channel network to discharge at the streamgage location. For downstream local areas, some kind of routing model is needed to translate the flows from upstream to the downstream locations. For points with more complexity within their drainage area, models of reservoir operations, irrigation demands, glacial effects, etc. could be required. These other models will be mentioned in this chapter though not in the same detail as the snow and Sacramento models.

There are many references in the literature to calibration procedures and techniques. The vast majority of these deal with the calibration of models to an individual drainage area, usually a headwater area with few complications. While this manual includes strategies and procedures for calibrating an individual drainage, the primary emphasis is the calibration of an entire river basin which eventually leads to the calibration of the entire area of responsibility of an RFC. To calibrate the models needed for an entire RFC area for river forecasting applications, requires a calibration process that is efficient, that results in spatially consistent parameter values, and that accurately simulates streamflow and other variables under a full range of climatic conditions. If calibration of an entire RFC area is conducted as a series of largely independent efforts with various individuals working on one watershed at a time with only minor coordination, the results will not only fall short of the objectives, but will take much longer to complete than necessary. The strategies and procedures given in this chapter are aimed at fully meeting calibration

objectives in the minimum amount of time.

Objectives

There are three basic objectives when calibrating conceptual hydrologic models to an entire river basin for river forecasting applications.

1. Produce a good reproduction of the observed hydrograph at each individual point on the river system. The aim is to achieve a fit that contains the minimum amount of bias possible, i.e. all errors are random. This includes all types of bias including overall bias, bias related to the magnitude of flow, seasonal bias patterns, and bias related to specific snow and soil moisture conditions such as during an abnormally large snow accumulation year or after a long dry spell. Also intermediate variables such as snow water equivalent and soil moisture deficits should compare realistically to any observations of these variables. The amount of random error should be largely a function of the random error associated with the input variables, especially precipitation. Errors in the amount of precipitation, as categorized by the typical spatial variability of this input variable, are the primary reason that lumped models do not produce satisfactory results in some areas of the country as was discussed back in Chapter 1 and illustrated in Figure 1.1.

2. The parameters of the models should function as they are intended. Both the SNOW-17 and Sacramento models are conceptual models which represent, although in a simplistic fashion, the main physical processes that occur in nature. These models were designed to have a physical basis and the parameters control portions of the models that represent specific components of the overall process. The parameters of the Sacramento model were designed to represent items such as the timing and maximum contribution of various runoff components, the maximum soil moisture deficits that can occur, and the rate of the movement of water within the soil profile with changing moisture conditions. The snow model parameters represent such items as the seasonal variation in melt rates when the area is completely snow covered, the areal depletion pattern as the snow melts, and the amount of liquid water that can be held within the snow cover. The effects of each parameter are designed to be reflected in specific sections of the simulated hydrograph under specific soil moisture or snow cover conditions. In order to be consistent with the physical basis of the models and to produce results that will not only best reproduce the full range of historical observations, but also be most likely to extrapolate correctly beyond what was observed in the available historical record, each parameter of the models should be used as it was intended.

3. There should be a realistic variation in parameter values from one area (headwater, local, or subdivision within a drainage) to another within the river basin and with areas just across the divide in adjacent river basins. Changes in parameter values from one area to the next should be explainable based on changes in physiographic factors, climatic conditions, or hydrograph response. Not only is this objective reasonable from a physical point of view, but

if adhered to, makes it much easier to monitor and understand operational variations and run time adjustments to state variables.

One question that is often asked, especially by those first learning how to calibrate conceptual hydrologic models, is "when am I done." Basically the answer is that you are done when the objectives have been met, i.e. when all possible bias has been removed, when the portions of the hydrograph controlled by each parameter are checked to make sure that the parameter is acting as intended, and when any changes in parameter values from one area to another are consistent with the assessment that was made of the spatial variability of hydrologic factors across the region.

When trying to judge whether the objectives have been met, it is important to remember that there is often considerable noise in the input data being used for calibration, especially the precipitation input. The amount of noise varies with the region of the country and the gage network. There may not only be considerable noise in the data for a given watershed, but the amount of noise can vary from one watershed to another depending on the number of gages available, as well as their location and accuracy of the measurements. Noise in the input data can make it not only difficult to determine the appropriate parameter values for a given watershed, but the variation in the amount of noise from one watershed to another can affect the spatial consistency of the results. This is why the strategy recommended in this chapter starts with calibrating the watershed that has the least noise in the data record. This is also one of the reasons why a sufficiently long period with considerable climatic variability is needed for calibration. Such a period should minimize "curve fitting", though there still may be considerable uncertainty in the value of parameters that control portions of the models that are seldom activated. Verifying the results on independent data periods should help to reduce this uncertainty and further minimize any "curve fitting". After the initial watershed in the basin is calibrated, one must be careful that realistic spatial consistency patterns are not destroyed by "curve fitting" during the calibration of the other drainages in the basin. In order to achieve the proper balance between the calibration objectives, given the amount and variation of noise in the data, the reproduction of the observed hydrograph, at least as measured by goodness of fit statistics, may have to be sacrificed somewhat in order to achieve spatial consistency of the parameters.

It is stated in objective one that all possible bias should be removed. When working with a lumped model, there are certain types of bias that are inherent in how the model is applied. In addition, there are certain model limitations that will cause trends in the results. These factors result in bias that typically cannot be removed including:

- An under simulation of the highest flows. A lumped application of a model uses the average amount of precipitation over an area, whereas in nature the amount of precipitation is seldom uniform. Since the rainfall-runoff process is non-linear, those portions of an area that have rainfall or snowmelt amounts that are greater than the mean areal value will produce relatively more runoff than parts of the area where the amounts are below the mean. This results in the actual runoff being greater than what would be produced by applying the mean

value to the entire area. Adjustments to some model parameters, especially the percolation curve in the Sacramento model, can partly adjust for this tendency, but especially in regions where there is typically a large variation in intensity levels during storms over individual drainage areas, a lumped application of a model will under compute runoff during high flow events.

• A over simulation of low flows. When most models are applied in a lumped fashion it is typically assumed that baseflow is being generated over the entire area based on the contents of the groundwater storages, at least that is the case with the Sacramento model. Under the lowest flow conditions, this is likely not the case in nature. Thus, there is a tendency for a lumped application of the Sacramento model to over simulate the lowest flow levels.

• In mountainous areas where snowmelt dominates runoff production, the simulated spring snowmelt typically occurs too early during years with a much below normal snow cover. This occurs because the snow primarily only covers the highest portion of the upper elevation zone, whereas the lumped application assumes the snow is distributed over the entire zone. This situation is described further in Section 6-1 and illustrated in Figure 6-1-3.

• Some biased results when snowmelt is not occurring over the entire area. The model assumes that either melt is taking place over the entire snow covered area or it is not. Especially in mountainous areas, early in the melt season snowmelt may only be occurring at the lowest elevations and on south facing slopes. This can result in some bias in simulating runoff during the first week or so of the snowmelt period.

• The largest snowmelt runoff events are typically under simulated, particularly in regions where high winds and dew-points are associated with major snowmelt situations. This is partly due to the lumped application of the model, but primarily due to using an index to compute snowmelt. In the SNOW-17 model air temperature is used as the sole index to snowmelt. While temperature is a good indicator of melt under most conditions, during some extreme snowmelt situations the typical relationship between temperature and melt doesn't hold. Especially in the northeast, major snowmelt events are associated with high dew-points and wind speeds. This causes large amounts of latent and sensible energy exchange and alters the normal relationship between air temperature and melt. There are other situations when the relationship between temperature and melt varies from the normal, but these are not associated with a particular level of melt and thus tend to randomly affect flow interval bias computations.

• Rainfall events that occur late in the snowmelt season on watersheds with a prolonged snow depletion period, generally mountainous areas, typically are over simulated. In these situations the soil has dried out in portions of the area that have been bare of snow for sometime, whereas as long as the areal average melt computed by the model, which is coming from only a small part of the area, exceeds the evaporation rate, the soil will remain wet. This can be minimized by using additional elevation zones, but the use of too many

zones can cause operational difficulties (see Section 6-1).

• Baseflow recharge can't be modeled consistently when there is a lower zone tension water deficit in the Sacramento model. This is caused by the model assuming that a constant fraction of the area (PFREE parameter) contributes to recharge during this situation when in reality the fraction should undoubtedly vary depending on the size of the deficit, i.e. the ratio of LZTWC/LZTWM. If the ratio is 0.0, little recharge should occur. As the lower zone becomes more saturated, the amount of recharge should increase.

Calibration Methods

There are two basic methods used for the calibration of hydrologic models. The first is a guided trial and error procedure where the users knowledge of the model and how each parameter affects the results are used to control changes to parameter values. Decisions as to which parameters to change are made primarily by comparing simulated versus observed values, especially hydrograph plots. This procedure is most effective when interactive, graphical software is available to view the results and make parameter changes. The calibration is finished when the user subjectively determines that the objectives have been met.

The second method is automated calibration [*Gupta et al.*, 2000]. In this method various computer algorithms are used to achieve the best simulated reproduction of observed values, typically mean daily discharge. The algorithms contain strategies for varying the values of user specified parameters in an attempt to obtain an optimal fit. Typically the user can apply limits on the range over which parameter values can varying in the hope of obtaining more physically realistic results. The quality of the reproduction is often determined by a single statistical objective function, such as minimizing the daily root mean square error. Sometimes a series of steps are used where different groups of parameters and different objective functions are used at each step (e.g. the objective function used for parameters affecting low flows may differ from the function used for parameters that primarily control storm runoff) [*Hogue et al.*, 2000]. In some approaches multiple objective functions are used to try to find a group of parameter sets that will produce good results based on several criteria [*Gupta et al.*, 1998]. Then the user can choose subjectively from this group of parameter sets. Automatic optimization has been primarily used for the calibration of individual watersheds, mainly headwater drainages. There are limited strategies available for using automated optimization over entire river basins.

Table 7-1 summarizes some of the features of the interactive trial and error method versus automated optimization of parameters (this table was prepared for a calibration training video developed by the NWS in conjunction with the Hydrologic Research Center [*Hydrologic Research Center*, 1999]). The biggest difference between the two methods is that the interactive method allows for the user to maintain the physical basis of the models, whereas the automated method relies on various algorithms to achieve a statistical best fit determination of the parameter values. NWSRFS contains software for both methods of calibration. The Interactive Calibration Program (ICP) allows for the user to make parameter changes and view the results graphically.

The automatic optimization program (OPT) contains 3 algorithms and several objective functions for use in determining best fit parameter sets. The OPT program was quite useful prior to the time when ICP was first available. Back then trial and error calibration was being done in a time consuming batch mode.

INTERACTIVE vs. AUTOMATED CALIBRATION

INTERACTIVE	AUTOMATED
• Emphasis on Component Process Representation	• Emphasis on Overall Model Fit to Data
• Requires Good Knowledge of Physical Model Basis	• Treats Model as Nonlinear Regression
• Person Intensive	• Low Personnel Requirements
• Use of a Multitude of Performance Criteria	• A Small Number of Statistical Criteria
• Less Affected by Data Quality Problems	• Sensitive to Data Quality
• Requires Well Designed Graphical Interfaces	• Requires Robust Optimization Methods
• Likely to Produce Parameter Estimates Which Would Allow Reliable Simulations of Future Events	• Likely to Produce Parameter Estimates with Uncertain Value for the Simulation of Future Events

Table 7-1. Comparison of Interactive and Automated Calibration Methods.

It is this author's belief, based on personal experience, that automated optimization methods cannot be used to meet the 2nd and 3rd objectives listed earlier in this chapter for the calibration of conceptual hydrologic models. Because of this assessment, the author has not used automated methods enough in recent years to feel qualified to include a discussion of strategies for using such procedures for calibrating conceptual models in this manual. Other references, such as those mentioned earlier in this subsection, are available for those who want to try automated methods. The biggest obstacle for the successful use of the interactive trial and error method is the time required to develop the knowledge of the model structure and how to isolate the effects of each parameter. While automated methods can in many cases achieve a good reproduction of streamflow for an individual watershed quicker than the interactive trial and error method, by following the procedures and strategies in this chapter, interactive calibration should produce not only a good reproduction of observed conditions, but preserve the physical nature of the models and do so in a very efficient manner for a large area like an entire river basin. After the calibration of the initial headwater area within a river basin, the strategies given in this chapter

should produce parameter sets for subsequent drainages that meet all the listed objectives at least as fast as automated methods can meet only the first objective.

Interactive Calibration Program (ICP)

The software used to perform interactive trial and error calibration within NWSRFS is the Interactive Calibration Program (ICP). ICP provides an interactive, graphical interface for the Manual Calibration Program (MCP). MCP contains all of the models and other procedures that are used to generate a simulation and output the results. The input for MCP specifies all of the time series that are to be used, the sequence of operations (i.e. models and other procedures including displays), and parameters and initial state variables for each operation. ICP allows the user to select a watershed, run MCP, display the results, make parameter changes, save specified simulated time series, and resubmit a new MCP run with modified parameter values. This process is continued until the calibration is judged to be finished.

ICP currently contains two graphical displays. These are linked to the WY-PLOT and PLOT-TS operations of MCP. The WY-PLOT display allows the user to graphically plot mean daily flow time series, both simulated and observed. The WY-PLOT display also includes panels that show many of the state variables and internal computations for the SNOW-17 and Sacramento models. These panels are described in detail at the beginning of section 7-7 for the snow model and 7-8 for the Sacramento model. Pull down, tear off menus allow the user to rapidly switch between the snow and Sacramento panels, as well as from one subarea to another for watersheds with multiple zones. The same method can be used to quickly go back and forth between multiple WY-PLOT displays if they are specified in the MCP control file. The user can scroll through the period of record, change the length of the period displayed, alter the range of the plots, and switch from arithmetic to semi-log scales. A single simulated time series can be saved for display after a subsequent run. This allows the user to compare the results from one run to another.

The PLOT-TS display allows the user to stack from one to six plots on top of each other, all showing the same period. The time series included on each plot must have the same units, but can have different time intervals. The flexibility of the PLOT-TS display allows the user to construct plots to help in visualizing a variety of situations. Just like the WY-PLOT display, the user can scroll through the run period, change the length of the period shown at any one time, alter the range of the plots, and switch from arithmetic to semi-log scales. One time series on each plot can be saved and replotted after a subsequent run.

Other output generated by the operations in MCP can be viewed via ICP in text format. This includes the statistical computations generated by the STAT-QME operation. Graphical displays likely will be added for some of these operations in the future.

ICP contains interactive windows for changing values of selected parameters for the snow, Sacramento, and unit hydrograph operations. Plots are provided for changing multiple valued

parameters like the ET-Demand curve for the Sacramento model, the areal depletion curve for the snow model, and the shape of the unit hydrograph. Other changes to the MCP control file are made by editing the file directly.

River Basin Calibration Strategy

In order to meet the calibration objectives in an efficient manner it is necessary to have a clear strategy for calibrating each of the headwater and local areas within a river basin. The recommended strategy is as follows:

1. First calibrate the headwater area with the best data and least complications. The aim of calibration is to determine the model parameter values that will provide the best possible forecasts of future conditions. Also the strategy recommended for a river basin ties the parameter values for all the other drainages within the basin to the parameters determined for the initial headwater area. Therefore, it is critical to start with an area where one has the best possible chance of determining appropriate parameters for the snow and Sacramento models. Complications such as reservoirs, diversions, irrigation, power plants, etc. and noise in the input data make it much more difficult to determine proper parameter values. The more noise caused by errors in the input data and complications that affect the observed output signal, the more difficult it is to see through the noise and determine the best values for the model parameters. Thus, the first area to calibrate should be the one with the best data and least complications. Section 7-1 describes in detail the selection criteria and the strategies and procedures that are recommended for use in calibrating the initial headwater area.

2. Calibrate other areas with minimal complications. These are other headwater areas, including reservoir inflows where the inflow hydrograph doesn't contain significant noise, as well as local areas for which a good definition of the local contribution can be determined. It includes areas where the flows have been adjusted to natural flow conditions and those where the flows were not adjusted but the man-made controls don't have a dominating effect. Once the local natural flow contribution is separated out by subtracting routed flows or correcting for man-made features, parameters for these areas can be determined in a similar fashion as the initial headwater area. The difference is that during the calibration of these areas *only the parameter values that clearly needed to be changed should be altered*. It is important not to make parameter changes for these areas merely to improve goodness of fit statistics or to make changes based on single events. Parameter changes should be based on clear evidence over many periods or events. If this philosophy is followed, the result should be a realistic variation in parameter values across the river basin. The assessment of spatial variability done in step 2 of the calibration process (see chapter 4) should provide an insight into the magnitude and types of parameter changes that should be expected. Section 7-2 describes details for determining local area hydrographs and strategies to follow when calibrating these drainage areas.

3. Determine parameters for the remaining headwater and local areas. These are areas where

a good definition of the local natural flow cannot be obtained due to the effect of man-made controls or excessive noise in the hydrograph after subtracting routed upstream flows from the discharge at the downstream point. In these areas a true calibration of the snow and soil moisture models is not possible. This step also includes areas for which the data are not available to make adjustments for or model the effect of control structures, thus a historical simulation is not possible. Parameters in both of these cases are generally obtained from nearby areas that were calibrated. Only a few simple adjustments are then made (if even possible) to remove any bias. Section 7-3 describes procedures to follow for determining the appropriate parameters for these remaining drainage areas.

By following this strategy one should be able to meet the calibration objectives in an efficient manner. The calibration of the initial headwater area must be done carefully as the parameters for all the other areas are tied to the values determined for this area. The calibration of this area should take longer than any of the other areas. The calibration of the other areas with minimal complications should go quite quickly as long as one knows how to identify the portion of the hydrograph affect by each model parameter and thus can fairly easily determine which parameters clearly need to be altered. The time required for the completion of the remaining headwater and local areas is primarily a function of how long it takes to determine routing parameters and/or remove or model man-made complications.

Statistics to Monitor

Decisions regarding which parameters need to be changed and the overall reproduction of the observed conditions are primarily determined by examining graphical plots of simulated and observed time series that are produced by the WY-PLOT and PLOT-TS displays of ICP. In addition to viewing these plots, there are various statistics that are helpful to monitor when doing interactive trial and error calibration. The currently available statistics are computed from mean daily flow time series by the STAT-QME operation and are not in graphical form (eventually it would be beneficial to be able to generate statistics for any data type, display many of the results graphically, and compute a wider variety of statistics than are included in STAT-QME). STAT-QME will optionally generate statistics on a water year basis in addition to the multi-year statistics for the entire run period. It is recommended that the user focus on the total run period statistics in order to see trends that may exist in the overall simulation. Yearly statistics are seldom of value. Of the currently available multi-year statistics there are several that are very helpful to periodically check during a calibration (Figures 7-1 and 7-2 show how these statistics appear in the output -- in the ICP program the STAT-QME output is included in the 'Edit Wide Listing' display). These are:

- Seasonal and overall bias – These are shown in terms of both percentage and depth of runoff. Both columns are useful. The overall bias shows whether the current data and parameters are generating a near zero water balance or whether the models are producing too much or too little overall runoff. For the calibration period the overall bias should at least be less than 5%. The monthly figures indicate how these quantities vary seasonally. The aim is

to have a random variation of reasonably small deviations from zero. A definite pattern in the seasonal bias indicates that something needs to be adjusted. Several months in a row of a double digit percent bias in one direction likely indicates a problem.

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MULTIYEAR STATISTICAL SUMMARY

FRENCH BROAD-ROSMAN AREA (SQ KM) = 176.00 WATER YEARS 1954 TO 1965

MONTHLY	SIMULATED MEAN (CMSD)	OBSERVED MEAN (CMSD)	PERCENT BIAS	MONTHLY BIAS (SIM-OBS) (MM)	MAXIMUM ERROR (SIM-OBS) (CMSD)	PERCENT AVERAGE ABSOLUTE ERROR	PERCENT DAILY RMS ERROR	MAX MONTHLY VOLUME ERROR (MM)	PERCENT AVG ABS MONTHLY VOL ERROR	PERCENT MONTHLY VOL RMS ERROR
OCTOBER	5.644	5.493	2.75	2.299	42.470	13.25	47.10	15.322	6.40	8.17
NOVEMBER	4.425	4.579	-3.36	-2.267	-11.530	14.79	25.55	-15.238	11.49	13.38
DECEMBER	6.120	6.064	0.92	0.847	-17.333	12.80	26.10	14.508	6.77	8.36
JANUARY	6.568	6.380	2.94	2.856	-15.996	13.19	23.62	21.007	7.88	9.50
FEBRUARY	8.370	8.202	2.04	2.301	-8.448	11.83	17.65	12.042	4.56	5.31
MARCH	8.773	8.618	1.79	2.351	5.101	11.68	16.26	17.723	6.33	7.07
APRIL	10.218	10.594	-3.55	-5.535	-21.689	10.66	18.01	-36.210	5.48	8.13
MAY	7.310	7.404	-1.27	-1.429	-7.120	11.85	17.46	-13.731	6.12	7.29
JUNE	5.725	5.687	0.68	0.566	5.939	14.34	20.32	26.752	11.70	15.71
JULY	4.560	4.557	0.08	0.054	-5.751	16.22	23.76	-25.436	13.62	18.12
AUGUST	4.762	4.740	0.47	0.336	6.125	12.60	19.76	18.309	7.95	10.39
SEPTEMBER	4.366	4.292	1.74	1.098	11.797	16.24	27.37	26.773	12.99	17.26
YEAR AVG	6.369	6.350	0.31	3.477	42.470	12.88	23.79	-36.210	7.80	10.18

DAILY RMS ERROR (CMSD)	DAILY AVERAGE ABS ERROR (CMSD)	AVERAGE ABS MONTHLY VOL ERROR (MM)	MONTHLY VOLUME RMS ERROR (MM)	CORRELATION COEFFICIENT DAILY FLOWS	LINE OF BEST FIT OBS = A + B*SIM A	B
1.510	0.818	7.403	9.663	0.9696	0.4717	0.9229

FLOW INTERVAL	NUMBER OF CASES	SIMULATED MEAN (CMSD)	OBSERVED MEAN (CMSD)	PERCENT BIAS	BIAS (SIM-OBS) (MM)	MAXIMUM ERROR (CMSD)	PERCENT AVG ABS ERROR	PERCENT RMS ERROR
0.00 - 2.00	305	1.712	1.634	4.75	0.0381	0.931	13.64	18.71
2.00 - 4.00	1213	2.966	2.979	-0.41	-0.0060	2.120	13.82	17.46
4.00 - 6.00	916	4.881	4.917	-0.73	-0.0177	4.169	14.00	18.10
6.00 - 9.00	968	7.250	7.303	-0.73	-0.0261	5.268	12.68	16.23
9.00 - 14.00	464	11.169	10.813	3.29	0.1745	7.041	11.90	16.76
14.00 - 30.00	212	18.751	18.440	1.69	0.1526	7.290	11.38	14.29
30.00 AND ABOVE	32	44.073	46.414	-5.04	-1.1492	42.470	15.20	23.90

Table 7-1. Illustration of Overall (Year Avg. Percent), Monthly (Percent and Runoff Depth), and Flow Interval Bias (Percent) Output from the STAT-QME Operation.

• Flow interval bias – This table indicates whether mean daily discharges are over or under computed for 7 intervals. The preferred option is for the user to specify the intervals (possibly using physically meaningful values such as bankfull and flood flow), though the STAT-QME operation will generate interval ranges if user specified values are not entered. Again percentage and depth of runoff columns are included, but only the percentage column is worth examining. The aim is to have a random variation of reasonably small deviations from zero, though as mentioned under the objectives section of this chapter there is a

tendency with lumped models to under simulate the highest flows and over estimate the lowest flows.

• Accumulated flows and errors – This optional table tabulates the accumulated simulated and observed flows and their difference on a quarterly basis (produced when the 'QUAR' input field is included). Ideally the errors for each period should be random and the accumulated error should meander back and forth around zero. Trends in the quarterly accumulated errors indicate that the relationship between simulated and observed discharge is changing over time. This may indicate an inconsistency in the input or discharge data or a physical change within the watershed that is not being considered. This table doesn't need to be monitored as frequently as the overall, seasonal, and flow interval bias values.

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ACCUMULATED FLOW IN MM

PERIOD		YEAR	OBSERVED	SIMULATED	ACC ERROR	ERROR THIS PERIOD
OCTOBER	TO DECEMBER	1953	168.06	160.61	-7.45	-7.45
JANUARY	TO MARCH	1954	542.81	525.73	-17.09	-9.63
APRIL	TO JUNE	1954	799.86	780.28	-19.58	-2.50
JULY	TO SEPTEMBER	1954	881.93	865.58	-16.35	3.23
OCTOBER	TO DECEMBER	1954	987.47	966.18	-21.28	-4.93
JANUARY	TO MARCH	1955	1211.08	1214.44	3.36	24.64
APRIL	TO JUNE	1955	1593.08	1584.05	-9.04	-12.40
JULY	TO SEPTEMBER	1955	1838.30	1789.53	-48.77	-39.73
OCTOBER	TO DECEMBER	1955	1961.83	1894.51	-67.32	-18.55
JANUARY	TO MARCH	1956	2220.75	2139.16	-81.58	-14.27
APRIL	TO JUNE	1956	2490.30	2378.58	-111.72	-30.14
JULY	TO SEPTEMBER	1956	2622.37	2484.45	-137.92	-26.20
OCTOBER	TO DECEMBER	1956	2792.93	2644.84	-148.09	-10.17
JANUARY	TO MARCH	1957	3146.74	2982.87	-163.88	-15.79
APRIL	TO JUNE	1957	3607.68	3428.88	-178.80	-14.92
JULY	TO SEPTEMBER	1957	3806.87	3642.86	-164.01	14.79
OCTOBER	TO DECEMBER	1957	4262.06	4084.87	-177.19	-13.18
JANUARY	TO MARCH	1958	4653.86	4462.38	-191.48	-14.30
APRIL	TO JUNE	1958	5062.24	4857.98	-204.25	-12.77
JULY	TO SEPTEMBER	1958	5272.48	5084.24	-188.24	16.01
OCTOBER	TO DECEMBER	1958	5397.99	5212.71	-185.28	2.95
JANUARY	TO MARCH	1959	5671.75	5510.14	-161.60	23.68
APRIL	TO JUNE	1959	6081.86	5935.07	-146.79	14.82
JULY	TO SEPTEMBER	1959	6303.80	6162.87	-140.93	5.86
OCTOBER	TO DECEMBER	1959	6657.06	6541.30	-115.75	25.18
JANUARY	TO MARCH	1960	7095.50	6988.56	-106.94	8.82
APRIL	TO JUNE	1960	7432.55	7322.72	-109.83	-2.89
JULY	TO SEPTEMBER	1960	7638.30	7518.04	-120.25	-10.43
OCTOBER	TO DECEMBER	1960	7825.32	7691.68	-133.65	-13.39
JANUARY	TO MARCH	1961	8138.75	8026.83	-111.92	21.73
APRIL	TO JUNE	1961	8484.82	8369.68	-115.14	-3.22
JULY	TO SEPTEMBER	1961	8878.04	8742.62	-135.43	-20.28
OCTOBER	TO DECEMBER	1961	9243.02	9120.68	-122.34	13.09
JANUARY	TO MARCH	1962	9689.95	9611.56	-78.39	43.95
APRIL	TO JUNE	1962	10068.47	10002.94	-65.52	12.87
JULY	TO SEPTEMBER	1962	10193.77	10156.20	-37.58	27.95
OCTOBER	TO DECEMBER	1962	10349.19	10341.75	-7.44	30.14
JANUARY	TO MARCH	1963	10611.38	10629.90	18.53	25.97
APRIL	TO JUNE	1963	10833.53	10862.24	28.71	10.18
JULY	TO SEPTEMBER	1963	10980.88	10996.57	15.68	-13.02
OCTOBER	TO DECEMBER	1963	11120.88	11123.67	2.79	-12.89
JANUARY	TO MARCH	1964	11546.26	11537.09	-9.17	-11.96
APRIL	TO JUNE	1964	11951.35	11912.78	-38.57	-29.40
JULY	TO SEPTEMBER	1964	12240.48	12260.09	19.61	58.18
OCTOBER	TO DECEMBER	1964	12811.19	12850.57	39.38	19.77
JANUARY	TO MARCH	1965	12811.19	12850.57	39.38	0.00
APRIL	TO JUNE	1965	12811.19	12850.57	39.38	0.00
JULY	TO SEPTEMBER	1965	12811.19	12850.57	39.38	0.00

Table 7.2. Sample Accumulated Flow and Error Table from the STAT-QME Operation.

While these statistics don't indicate which parameters or data need adjustments, they do give some insight into trends that exist and possibly the parameters and model components that should be examined. Many of the other statistics that are produced by the STAT-QME operation only give an indication of the overall goodness of fit and may be useful when evaluating the results, but they aren't very helpful in providing insights into making changes to parameters.

Flows to Route Downstream

After calibrating each point within the river basin, time series need to be produced and saved for use in calibrating and evaluating the simulation at the next location downstream. In order to best calibrate these downstream locations and to properly evaluate the ability to reproduce historical conditions at downstream gages, the following two time series should be generated at each point:

- Adjusted instantaneous discharge – these time series are used when calibrating the downstream local areas. Ideally a quality controlled observed instantaneous discharge time series whose volume matched that of the USGS mean daily flows would be routed downstream, but such a time series is rarely available. As a substitute an adjusted instantaneous discharge time series can be generated with the ADJUST-Q operation. This operation uses the simulated instantaneous and observed mean daily flows to create adjusted instantaneous discharge values. Periods of observed instantaneous discharge can also be included. The adjusted instantaneous discharge time series uses the shape information from the simulated time series, but adjusts the daily volume to match the observed mean daily flow. These adjusted instantaneous flow time series are then routed to the next downstream point and subtracted from the observed flow at that location to get the local area contribution which is then used to calibrate the local area. By adjusting to the observed volume, errors in upstream simulations are not propagated downstream. The calibration of the local areas are then based on observed discharges just like the headwater drainages. If observed daily flows are not available for some periods at an upstream gage, the adjusted time series will contain just the simulated values. Such periods should be avoided when calibrating the local area.

- Simulated instantaneous discharges – these time series are used to evaluate the ability to forecast at downstream locations. When forecasting, whether it be a short term or extended forecast, there are no observed data in the future (forecast values may be modified somewhat based on the last available observations of streamflow). To determine how well one can reproduce conditions at downstream points in the forecast mode, simulated discharge for the entire drainage area above the gage needs to be compared to observed flow. If only adjusted discharges are routed downstream, errors and bias from upstream locations will be removed from downstream simulations. Relatively minor errors and bias at upstream points could add up to larger errors and bias downstream.

These two time series are generated and saved after completing the calibration at each location. For headwaters, both time series can be created in the same run. For downstream points, two runs are needed. To generate the adjusted time series, adjusted flows should be routed down

from upstream points in addition to using the ADJUST-Q operation at the downstream location. This insures the best possible simulated instantaneous discharge at the downstream point and will use as much observed volume data as possible if there are periods of missing data at that location. To generate the simulated time series, simulated flows must be routed down from upstream points so that all the flow at the downstream location is made up of simulated values.

At some locations it is impossible to perform a historical simulation due to the lack of the necessary data to adjust or model the effect of control structures. At these locations the sequence of routing simulated flows downstream is broken. Observed flow data at these locations can be used to create instantaneous streamflow values to route downstream (generally instantaneous flows are calculated from observed mean daily discharge data using the CHANGE-T operation and, when available, observed instantaneous flows can be used) for use in calibrating the next downstream location, however, a full historical simulation of the entire drainage area below such points is not possible.

Recalibration

Over time there will undoubtedly be a need to recalibrate all or part of a river basin. Reasons for needing to recalibrate were given in chapter 2. Whether a recalibration is necessary can be the user's choice in some cases and mandatory in others. Cases when the user must decide if a recalibration is needed include:

- calibration expertise and knowledge has increased and as a result past calibrations, as well as operational performance, are reviewed to determine at what locations a recalibration is likely to improve results (if the increase in expertise and knowledge will likely result in improved data estimates, then the affected historical data record should also be regenerated).

- operational results are unsatisfactory under certain circumstances -- after determining the situations that cause problems, the user must decide if these were the result of cases that were missing, overlooked, or modeled improperly during the calibration – if so, a recalibration is a possible solution (in many of these cases the historical data record must first be extended to include the situations that didn't occur within the previous calibration period), and

- physical changes (such as new agricultural practices, large forest fire, or modifications to land use) have occurred within portions of the basin and based on decreased operational performance it is decided that the effects are significant – if so, a recalibration is needed for the affect portions of the basin (in this case the historical data must first be extended and then only the period after the change was well established would be used to modify the model parameters).

- climatic changes have occurred and it appears that the operational output exhibits trends such as over estimation of runoff during certain seasons which might be attributed to increased evaporation rates - if so, a recalibration should be explored using an extension of

the historical record. Adjustments, such as those to climatological average ET-Demand values might not require recalibrating every watershed, but only selected drainages to determine the size of the adjustments to apply over the entire area.

In the first of these cases it is probably best to totally redo the calibration, i.e. calibrate the headwater area with the best data first and then move on to the other watersheds following the recommended guidelines. Trying to start with the parameters from the previous unsatisfactory calibration will only interfere with and negatively influence the process. In the other two cases it is recommended that one start with the existing parameter values and change only those that clearly need to be modified.

Situations when a recalibration is mandatory include:

- existing operational models are being replaced with new models (such as replacing an event API type rainfall runoff model with a conceptual soil moisture accounting model) in an attempt to improve forecast results (the existing historical data may be able to be used to calibrate the new model, however, in many cases the new model may require new data types or the data may need to be in a different form (e.g. continuous as opposed to event data)),

- historical data have been reanalyzed using different procedures (such as switching from the non-mountainous area technique to a mountainous area procedure) resulting in input time series that are biased as compared to those used for the previous calibration (in this case new model input data must be generated for the entire historical data period),

- new methods which should improve model input (such as determining precipitation estimates from a combination of gage, radar, and other data), but likely produce values that are biased compared to that used in the current calibration, have generated a sufficiently long historical record to be used to adjust the model parameters (in this case the recalibration can use only the period for which consistent data can be determined using the new method), and

- new forecast points are to be established at locations within the basin that were not part of the current calibration thus resulting in a subdivision of existing drainage areas and changes to the routing reaches (frequently the historical data must be extended in order to have as much historical streamflow data as possible at the new locations).

In the first of these cases a complete calibration of the new models is required since the new models are being applied to the basin for the first time. In addition it may be necessary to modify parameters or at least check the performance of other models that were part of the previous calibration. For example, if the rainfall runoff model is being replaced, the parameters for the snow and channel response (e.g. unit hydrograph) models may need to be modified. Even though the snow model should be independent from the rainfall runoff model, its parameter values may have been affected by how the parameters and data for the rainfall runoff model were determined in the previous calibration. The same thing holds true for the channel response model and, in

addition, the function of the channel model may vary depending on the structure of the rainfall runoff model (e.g. the Sacramento model requires a different unit hydrograph than an API type model – see section 7.6). As long as the other models were properly calibrated previously, only adjustments that are clearly needed should be made to the parameters for these models.

In the other cases when a recalibration is mandatory, a complete new calibration is generally not required. It should be possible to utilize the parameters from the previous calibration as a starting point and then modify only the values of those parameters that need to be changed. For the snow and soil moisture models changes should be substantial only when there is a significant bias between the new and previous input data. Guidelines for the specific case of using radar based precipitation estimates for operational forecasting are included in chapter 8.

Section 7-1

Calibration of Initial Headwater Area

Selection Criteria

The following criteria, not in any particular order, should be used for selecting the initial headwater area to calibrate within a river basin:

- The period of observed streamflow data should be as long as possible, ideally covering most or all of the period of record used for the historical data analysis. This long period is needed for two reasons. First, besides needing a period to calibrate the models at the location, it is also important to have some independent periods to validate the results. Second, many times it is helpful to make comparisons between calibrations at other locations within the basin and the initial headwater area and since the streamflow data for the other locations may have varying periods of record, the initial headwater point needs a long record to insure that there are overlaps with the other sites.

- The physical characteristics of the drainage area should have been basically stable over time, i.e. no substantial land use, vegetation cover, or agricultural changes and the streamflow data should be consistent. In addition, the initial headwater should be physiographically representative of the total river basin, or at least the portions of the basin that produce significant runoff.

- The networks used to estimate the model input variables, especially precipitation, should have good coverage. Noise in the input data, especially precipitation, makes it more difficult to determine the proper parameter values, thus the mean areal inputs need to have a minimal amount of random error.

- If there is a wide variation in the amount of annual runoff over the river basin, especially if there are drainages with small amounts of runoff, the initial headwater area should be a drainage with average to above average runoff relative to the other parts of the river basin.

- There should be minimal complications. There should be no significant reservoirs or large lakes that dampen out the hydrograph response. Diversions, if any, should be small and should have observed data that can be used to adjust the streamflow to natural conditions. The amount of irrigated acreage should have little, if any, effect on flow. There shouldn't be any power plants or other controls that cause substantial noise at low flows. Glacier contribution to runoff should be avoided if possible.

Using these criteria, all of the headwater gages within the river basin should be evaluated and the one that comes the closest to meeting the criteria should be used as the initial area for calibration.

Periods of Record

As mentioned in Chapter 7, besides calibrating the models for the initial headwater area, a validation should be done on a separate period of the record at that location. If a careful validation is done on the initial headwater, it is probably not necessary to run separate validations at the other locations within the river basin. Guidelines for selecting the period to use for calibration are as follows:

- The calibration period should contain as much variety in hydrologic conditions as possible. There should be events with very high flows and there should be extended dry periods with very low flows. Ideally there should be significant runoff events at all times of the year and under different initial soil moisture conditions. There should be years with both much above and considerably below normal volumes of runoff. If snow is a factor, there should be years with both very large and abnormally small amounts of snow cover and ideally snowmelt periods at various times during the snow season. However, it is probably best to not include the flood of record, the minimum flow of record, or other extremes within the calibration period. It would be better to save these events and periods for validation so that you can see if the calibrated models can extrapolate beyond the conditions experienced during calibration.

- The calibration period typically needs to be about 10 years long, at least in the areas where lumped, conceptual models generally provide satisfactory results. Experience has shown that a period of about 10 years is needed in order to determine stable values of the model parameters. When the models are calibrated using shorter periods, the parameters are more likely to vary depending on the period used. In regions where lumped models give marginal results, the calibration period may need to be longer in order to get a sufficient number of events and variety of conditions to determine the parameter values. In regions where the results are typically unsatisfactory, even the full period of record is likely not going to be adequate for determining the parameters with any degree of certainty.

- If possible, it is a good idea to have the calibration period for the initial headwater overlap with periods when most of the other streamgages within the basin also have observed daily flow records. In this case the same period can be used at most locations for calibration and observed discharges will be available to adjust the instantaneous flows that are routed downstream.

The periods to use for validation should be as long as possible and, as mentioned above, ideally will contain some extreme events and situations to test the extrapolation capabilities of the calibrated models. Essentially the entire period with observed mean daily flow data, except for the calibration period, will be used for validation.

Initial Model Parameter Values

Initial parameter values for the snow and Sacramento models can be obtained in one of two ways

for the initial headwater area.

- If an adjacent river basin has been already calibrated, the initial snow and Sacramento model parameters should be obtained from the headwater area within the previously calibrated basin that is the most hydrologically similar to the initial headwater in the current river basin. This nearby headwater should have good data, minimal complications, and its calibration results have met the objectives so that you are quite confident in the parameter values. In this case the strategy for calibrating this initial headwater will be the same as for other headwaters and locals with minimal complications within the river basin, i.e. only change those parameters that clearly need to be altered. By following this procedure parameter values should not only end up being consistent from one area to another within a river basin, but should show realistic variations across the entire RFC area.

- If no adjacent river basin has been previously calibrated or if there is a considerable difference in hydrologic conditions between this initial headwater area and any drainage that has been previously calibrated in the region, then the initial snow and Sacramento model parameters can be derived from available information. Guidelines and methods have been developed to derive initial parameter values by analyzing hydrographs and from physical or climatic information. Section 7-4 contains guidelines for determining initial parameter values for the SNOW-17 model. Section 7-5 describes techniques for deriving initial parameter values for the Sacramento model. Even if the initial parameters are obtained from a previously calibrated drainage, it is a good idea to read and understand the material in these sections. The information included should help in understanding the function of each parameter, assist in knowing how to isolate the effects of the parameters, and give some insight into a reasonable range of values.

Initial parameter values for the channel response model that is used to convert the runoff entering the channel system to a discharge hydrograph at the gaging location should always be determined directly for each drainage area. The channel network for each drainage is unique and the response function for one area should not be used for another. Section 7-6 discusses how the unit hydrograph technique is used in conjunction with the Sacramento model and methods for deriving an initial estimate of the unit hydrograph ordinates.

Calibration Strategy

Once initial parameters have been determined for each of the models, the next step is to adjust the parameter values so that the simulation results meet the calibration objectives. The calibration of a conceptual model by making interactive adjustments to model parameters is completely dependent on having the proper knowledge of the function and how to isolate the effects of each parameter. The function of each parameter is determined by understanding the structure of the model. Knowing the model structure will also help in understanding what conditions must exist in order to select the portions of the simulation results to examine to determine if changes should be made to the value of a given parameter. It is critical for a person

to gain a reasonable understanding of how to isolate the effects of each model parameter if they are going become competent at calibrating conceptual models. Without a reasonable proficiency in knowing the effects of each parameter on model response, interactive calibration becomes a very inefficient process with little probability of the user ever determining the proper parameter values. Without this knowledge the user would be better off using an automatic calibration method. Section 7-7 contains a discussion of how the major snow model parameters affect model response. Section 7-8 contains a similar discussion for the Sacramento model. An understanding of these 2 sections and the structure of each model is an absolute prerequisite to interactive calibration.

When using the interactive trial and error method of calibration, it is best to follow a proven strategy for determining which parameters need to be modified rather than just randomly looking at various portions of the record and making parameter changes. Even though a step by step strategy is outlined, it cannot be followed in a cookbook fashion. The steps give a general pattern to follow, but the user must remain somewhat flexible. Some general items to consider concerning calibration in general and in particular when using the recommended strategy are:

- Be reasonably bold when making parameter changes. Finding the proper value will take less time if the changes to parameter values are fairly large. If you overshoot, it is much easier to estimate your next trial value than if you only make a sight incremental change.

- Model parameter values are selected to produce the best results over a number of events or occurrences of a given situation and should not be assigned based on a single event. The random errors that occur for individual events during calibration can hopefully be minimized operationally by improved data estimates or run-time modifications to model computations.

- Remove large errors in parameter values whenever they are detected. The first step in the strategy involves removing large errors that exist at that point in the process, but during later steps significant errors in other parameters may become apparent. When these errors are causing enough noise in the simulation results that they make it difficult to determine the proper value of the parameters currently being worked on, the parameters causing this noise must be modified, at least to the degree that their effect is not interfering with the current step.

- One should periodically return to previous steps to recheck the results. Parameter changes in subsequent steps may necessitate adjustments, usually small, to parameters values determined during an earlier step.

- Remember to periodically check the statistics mentioned in chapter 7. These statistics help to identify trends in the simulation and assist in determining the periods, and possibly the parameters, to examine.

- Change the duration and scale of the ICP displays depending on which flow components

and parameters are being examined. Long durations, typically one or two years, and a semi-log scale are used when working on low flow components and parameters, while shorter periods, generally in terms of months, and an arithmetic scale are used when examining storm events or snowmelt runoff periods.

The recommended strategy for calibrating an individual watershed is as follows:

1. Remove large errors. If certain initial parameters or some of the input data are considerably in error, there will be a large discrepancy between the simulated and observed hydrograph. These problems need to be corrected before proceeding in a more step by step fashion through the parameters. The amount of noise caused by these errors make it very difficult to isolate the effect of individual parameters and to determine which parameters need to be modified and by how much. The aim at this point is not to totally correct these errors, but to get the simulation results at least in the right ballpark.

Large timing errors should typically be corrected first, since they can also be a source of volume discrepancies. The most common large timing problems that exist at this point in the process are:

- Large error in the percolation rate for the Sacramento model such that there is way too much storm runoff and not nearly enough baseflow or vice versa. This problem can usually be corrected by changing the LZFSM and LZFPM parameters by the same ratio such that the PBASE term in the percolation equation is increased or decreased until the split between storm runoff and baseflow is more reasonable.

- Large error in the amount of surface runoff generated by the Sacramento model such that major storm events are way over or under simulated. This problem is generally corrected by changing the UZFWM parameter upwards or downwards.

- Improper channel response function is being used for the Sacramento model (see discussion in Section 7-6). This occurs when the unit hydrograph contains the timing effects of interflow, as well as surface runoff. This problem is corrected by removing interflow from the unit hydrograph being used.

There could also be large timing errors associated with snowmelt runoff, however, if the initial parameter guidelines are followed and the temperature data are reasonably unbiased, this should not be a problem. There could be timing problems with the snowmelt during the largest snow years due to the approach recommended for determining the SI snow model parameter, but this is to be expected at this point.

In general, for most watersheds the overall volume error should be no more than about 10%. For watersheds with small amounts of annual runoff, especially those with less than an average of 5 inches per year, the initial volume error may be greater. If the initial volume

error is larger than expected, it is important to look carefully at the components of the water balance to determine if there is a problem that needs to be corrected before proceeding with the calibration. The most likely problems are biased precipitation or evaporation data or streamflow data that haven't been corrected for diversions or other gains or losses. In some cases the tension water capacities of the Sacramento model could be off so far that the ratios of contents to capacity remain too large or too small and thus result in an error in the amount of computed actual evaporation. This is most likely in a arid or semi-arid region. In some cases the problem is merely that the unit hydrograph ordinates do not represent the specified drainage area (this should result in a warning from the UNIT-HG operation if the correct drainage area was entered). Any significant volume errors need to be corrected before proceeding with the calibration.

In some regions, especially the northeastern United States, there may be frequent errors in determining the form of precipitation, i.e. rain versus snow. This not only affects the model response at the time of the precipitation event, but also will affect the volume of snow available for melt at a later date. The amount of noise in the simulated results caused by this problem will make it very difficult to determine appropriate parameter values. When this occurs, the data for individual events, typically temperature, need to be modified so that the form of precipitation is generally correct before proceeding with the calibration. Suggestions for how to correct the form of precipitation are included in Section 7-4 under the "Form of Precipitation" section.

2. Obtain a reasonable simulation of baseflow. In order to obtain the best results from the Sacramento model, it is important to properly calibrate all components of the model. At many forecast points one may be primarily interested in high flows, however, the structure of the model makes it impossible to maximize the ability to predict high flows when low flows are not reasonably modeled. Contrary to the thoughts of some people, the model doesn't need multiple sets of parameters (e.g. one for simulating floods and another for extended predictions involving all flow levels). The simulation of the various flow components are all interconnected, thus the percolation of water into the lower zones and the computation of baseflow must be correct in order to get a good reproduction of storm runoff. A proper calibration should yield parameter values that will not only reasonably simulate low flows, but will also provide the best simulation of high flows.

Since many of the low flow parameters are involved in the percolation equation, it is necessary to concentrate on this part of the model first to obtain a good foundation before focusing on upper zone soil moisture and storm runoff. The aim of this step initially is not to finalize the parameters that control baseflow, but to get the simulated hydrograph to generally match the observed under low flow conditions. Then as the calibration proceeds, you will periodically return to this step and make refinements to the parameter values.

When working on getting a reasonable reproduction of low flows, it is best to start with primary baseflow. Section 7-5 discusses the importance of determining in advance which

portions of the observed hydrograph are to be modeled with each of the available runoff components and that the identification of primary baseflow is critical to a successful simulation. That section also describes situations when it is difficult to isolate primary baseflow. The periods when primary baseflow is the only source of runoff or at least predominates are examined to determine if the LZPK and LZFPM parameters need to be modified. Next supplemental baseflow periods are examined to check the values of LZSK and LZFSM. Changes to all these parameters will alter the PBASE term in the percolation equation and thus not only affect the amount of water going to the lower zone and baseflow, but also modifying the amount available for storm runoff. Thus, if the overall volume is in the right ballpark, the total volume of storm runoff should become more reasonable as the simulation of baseflow improves. Another parameter to examine at this point is PFREE. In some cases it may be necessary to make some crude adjustments to the shape of the percolation curve, i.e. parameters ZPERC and REXP, at this stage of the process, but the refinement of the shape of the percolation curve occurs in a later step.

When working on getting a reasonable reproduction of low flows, one should recognize whether there are periods when baseflow is drawn down due to evaporation from riparian vegetation or from irrigation withdrawals. Both of these causes of low flow draw down have a similar effect on the hydrograph. Knowledge of the basin is required to decide which is most likely the cause. Such periods should not be used when making adjustments to the main parameters that control low flow. After other low flow periods are being simulated in a reasonable fashion, one can then try various values of the RIVA parameter or introduce the Consumptive Use operation (CONS_USE) to see how well the draw down periods can be modeled. RIVA should then be set back to zero or the CONS_USE operation removed and these periods ignored until the final step.

3. Adjust major snow model parameters, if snow is included. Once the baseflow simulation is reasonably good, the major snow model parameters should be checked to determine if adjusments are needed to the volume and timing of snowmelt. The major snow model parameters are MFMAX, MFMIN, SCF, UADJ, SI, and the areal depletion curve. The minor snow model parameters generally should not need to be adjusted during calibration, but in a few cases modifications to these parameters may be necessary. It is most important to check the snow model parameters at this point in areas where snowmelt runoff is significant and especially when there is an extended melt season. If snowmelt events are infrequent and generally occur over a short period, it may be necessary to wait and adjust snow model parameters at the same time as the Sacramento model parameters that primarily affect storm runoff (step 5). If form of precipitation problems were not corrected in step 1 and it is now apparent that noise resulting from the mistyping of winter events is making it difficult to determine the proper parameter values, then the data should be corrected as described in Section 7-4. If mistyping of precipitation is infrequent and random, it should not be necessary to correct the data in order to determine the snow model parameter values.

If it is necessary to use snowmelt parameters (primarily MFMAX and MBASE) that are

considerably different from those suggested in Section 7-4 or if snowmelt occurs consistently early or late in spite of the changes that are made to the parameters, it is quite likely that the temperature estimates are invalid. There may be other reasons for discrepancies at the beginning of the snowmelt season as discussed near the end of Section 7-7, but significant overall timing problems or unreasonable parameter values are a good indication that the computation of the MAT values should be reexamined using the guidelines in Section 6-4. Especially in mountainous areas, the initial MAT estimates may need to be redone because of difficulties in determining the temperature versus elevation relationship due to scatter in the station data or the use of improper lapse rates.

4. Adjust tension water capacities. This step involves finding periods that isolate the effect of tension water deficits in both the upper and lower zones of the Sacramento model, i.e. the UZTWM and LZTWM parameters. The idea is to determine whether the deficits are generally too large, too small, or about right, thus indicating how the parameters should be modified. During this step is also a good time to check the value of the PCTIM parameter. Fast response runoff during periods when upper zone tension water deficits exist can only be modeled by constant impervious area runoff.

5. Adjust parameters that primarily affect storm runoff. This involves altering the value of UZFWM to get the proper division between surface runoff and interflow, changing UZK to get the correct timing of interflow, determining if ADIMP is needed and if so, finding the best value, and refining the shape of the percolation curve over a large range of LZDEFR values primarily by adjusting the ZPERC and REXP parameters. When making adjustments to the percolation rates in the Sacramento model, it is best to look at the entire curve and determine what changes are necessary and then select parameter values that will produce the curve that is needed. A procedure for evaluating the entire percolation curve is described in Section 7-8.

6. Make final parameter adjustments. This typically includes looking at the following:

- If riparian vegetation evaporation effects exist, determine the final value of the RIVA parameter or if there are irrigation withdrawals, add the CONS_USE operation and make the necessary adjustments to its parameters.

- Refine the timing of major peaks, mainly those that produce surface runoff, by modifying the shape of the channel response function (unit hydrograph).

- Adjust ET-Demand curve values to improve the seasonal bias pattern (if it can be deduced that ET errors are the cause of any trend in the seasonal bias). When making modifications to mean monthly ET-Demand values, the monthly PE adjustment curve should be examined to make sure that the values are realistic and that abrupt changes do not occur from one month to another.

• Raising or lowering the entire percolation curve to improve the flow interval bias pattern by changing LZFSM and LZFPM by the same ratio.

Indeterminate Parameters

It needs to be recognized that depending on the conditions that exist within a given watershed and the types of events that occur within the calibration period of record, there may be some parameter values that cannot be reliably determined. This occurs when the parameter is never or rarely activated during the calibration period. It is important to understand when this occurs since it may affect the operational ability of the model to extrapolate to conditions outside the range of what was included in the calibration period. Parameters for which reliable values sometimes cannot be determined are:

Snow Model:

• SI – If bare ground occurs as soon as snowmelt begins during every year, i.e. there is no significant period when the area remains at 100 percent snow cover, it is impossible to determine the value of SI other than to know that it is greater than the maximum average areal water equivalent that occurred during the calibration period.

• UADJ – If few or no significant rain-on-snow events with warm temperatures occur, it is not possible to obtain a reliable value of the UADJ parameter.

Sacramento Model:

• UZFWM – If surface runoff never occurs, it is impossible to determine the proper value of UZFWM other than to say that it has to be great enough that the model will never generate surface runoff. When there are only a few events with surface runoff, the value of UZFWM generally contains much uncertainty in watersheds where the distribution of rainfall is highly variable during these large storms.

• LZTWM, UZTWM, and PFREE – In very wet regions where significant soil moisture deficits never occur, at least during the calibration period, it is not possible to determine the value of the tension water capacities or PFREE. In some areas there are sufficient dry spells to estimate the UZTWM parameter, but none of sufficient length to determine LZTWM or get a good estimate of PFREE. Also in semi-arid regions, there may never be sufficient moisture to fill the lower zone tension water, thus it is not possible to know the proper value for this parameter. In very dry regions there is not enough runoff to be confident in any of model parameter values.

• ZPERC and REXP – If the vast majority of events occur over a limited range of lower zone moisture conditions, it is very difficult to obtain unique values for ZPERC and REXP. Various combinations of these parameters can result in similar percolation rates

over a small range of soil moisture. This is especially common in wet regions.

Special Situations

There are some special situations that can exist when calibrating watersheds that need some additional discussion. This includes watersheds with multiple zones, glaciers, and frozen ground.

• Multiple Zones – It is important to maintain a realistic relationship between parameter values from one zone to another within a watershed whether the drainage is subdivided based on elevation zones, travel time zones, or some other breakdown. There are 2 basic recommended approaches:

- Start with the same values for all parameters for each zone. Then keep the parameter values the same as the calibration progresses except when there are events that clearly allow one to determine unique parameter values for a given zone. These could be rain events where the precipitation only occurs over one of the zones or snowmelt runoff periods where all or most of the melt is coming from one zone.

- Start with parameter sets that have different initial values for some parameters and then maintain the ratio or difference in these parameters between the various zones as parameter values are modified. The initial differences can be based on relationships, either objective or subjective, between parameters based on soils, vegetation, or climatic conditions or based on differences in parameter values from previously calibrated watersheds that most likely represent one zone or another. The initial relationship between the parameters for each zone are only modified during the calibration if there are a number of events that clearly allow for a unique determination of certain parameter values for a given zone. ICP contains a feature (included with the Selected Parameters option under the Edit menu) that allows the user to maintain the ratio or difference (fixed for each parameter) between zones when changing parameter values for the snow and Sacramento models. When this feature is on, the parameters in all zones are altered to maintain the ratio or difference whenever the value for any one zone is changed.

• Glaciers – A separate zone is typically used to model glacier effects. One glacier zone should be sufficient unless possibly the glacier covers most of the drainage and covers a wide range of elevations. Two general modeling approaches have been used to simulate the streamflow response from the glaciated area:

- Use the snow model to simulate the accumulation and melt that occurs on the glacier surface and the GLACIER operation to model the time delays that take place as the water moves through the glacier. The GLACIER operation allows for a variable withdrawal rate of any liquid water passing through the glacier that is a function of the amount of inflow at the glacier surface in the recent past. This results in little outflow and a storage

buildup when rain or meltwater first enters the glacier. As warmer weather persists and passages within the glacier open up, there is a much faster response to surface inflow and the water in storage is released. This results in both a delay and dampening effect.

- Use the snow model to simulate the accumulation and melt at the surface, plus some of the delay that occurs when water passes through the glacier and use the Sacramento model to handle the variation in attenuation rates of this water. In this case a user specified seasonal melt factor variation is used in the snow model to artificially produce some of the delay between when rain or melt occurs and when the water reaches the streamgage. Only certain portions of the Sacramento model are used, primarily those that control the attenuation of water through the system by dividing it into the various components, surface, interflow, and supplemental and primary baseflow. The following Sacramento model parameters can be set to zero for a glacier; PCTIM, ADIMP, RIVA, PFREE, and the ET-Demand values. The tension water capacities can be set to a non-zero value and the storage initially set to the capacity, which will not change during the run since the evaporation rate is zero.

In both cases the initial water equivalent of the snow cover should be set to a very large value (typically 30,000 mm has been used) and the initial liquid water storage should be equal to the PLWHC parameter multiplied by the initial water equivalent (i.e. liquid water storage should be full at the start of the run). During the calibration it is important to monitor the change in the water equivalent to insure that any increase or decrease in glacier ice is reasonable compared to any available mass balance studies for glaciers in the region.

The main advantage of using the GLACIER operation is that it is a more conceptually correct approach. The snow model is being used as it was intended and the GLACIER operation is attempting to model the effects of the glacier on storing and attenuating the surface melt and rain water. The main advantage of using the Sacramento model is that it allows for more runoff components. The GLACIER operation allows for a variable withdrawal rate which can produce values that are similar to interflow and supplemental and primary baseflow rates in the Sacramento model. The GLACIER operation cannot pass water through the glacier as quickly as surface runoff occurs in the Sacramento model. High intensity rains on some glaciers in the late summer when passages through the ice are fully open show a surface runoff type response. The GLACIER operation cannot mimic this type of response.

• Frozen Ground – Extensions to the Sacramento model are used in an attempt to model the effects of frozen ground. The algorithms consist of an indicator of the amount of frost in the soil (either a frost index as in the preliminary frozen ground model or frost depth in a new procedure being developed) and then a modification of portions of the Sacramento model, typically percolation and interflow withdrawal rates, based on the amount of frost. Frozen ground mainly has a significant effect on streamflow in somewhat open regions where there are cold periods with little snow cover during portions of the winter and percolation rates are quite different for frozen and non frozen soil. Regions where dense forests and/or significant

snow cover exist provide enough insulation to prevent substantial frost from developing except at far northern latitudes where permafrost occurs. Regions where percolation rates are very low even when the soil is not frozen or where permafrost exists (percolation rates remain similar throughout the year), typically don't require the use of the frozen ground algorithms.

When deciding whether to include the frozen ground algorithms, it is a good idea to start the calibration without considering frozen ground and concentrate on events during the summer and fall, plus spring events during years with a substantial snow cover over most of the winter accumulation period. The frozen ground algorithms can be included and the amount of frost displayed, but the effect of the frost on the Sacramento model should be turned off (done in the preliminary frozen ground model by setting the SATR parameter to zero). Thus, one can see when the algorithms could change the response of the Sacramento model and a decision can be made as to whether the inclusion of frozen ground is needed and should likely improve the results.

Use of Other Data during Calibration

The primary observations used to compare against model simulated results during calibration are mean daily flow data, however, there are other observations that should also be used when they are available. This includes other types of flow data and snow observations.

• There are two types of streamflow data that should be used whenever possible besides mean daily discharges. These are instantaneous discharges and peak flows.

- Instantaneous discharge data are very helpful for faster responding watersheds and when there is a significant diurnal variation in flows during snowmelt periods. The instantaneous discharges are not needed on a continuous basis, just during selected storm events and snowmelt periods. When the storm hydrograph from a watershed doesn't peak for 2 to 4 days or more, generally mean daily flows are adequate for determining the parameters that control storm events (primarily UZFWM and possibly ADIMP) and the shape of the unit hydrograph, however, for areas that peak sooner, instantaneous discharges are often needed to refine these parameters. The same is true of watersheds with snowmelt periods that last for more than a few days. Mean daily flow data will not indicate the magnitude of diurnal fluctuations that may exist. This can only be determined by examining instantaneous discharge data.

To illustrate the situation involving diurnal variations in streamflow during snowmelt we will use data from the Central Sierra (CSSL) and Upper Columbia (UCSL) Snow Laboratories. These data were collected in the late 1940's and early 1950's as part of the Snow Investigations conducted by the Corps of Engineers and the Weather Bureau [*Snow Hydrology*, 1956]. The watersheds used are both small (CSSL watershed is 3.96 mi^2 and the UCSL watershed is 8.09 mi^2), thus any damping of the response due to the channel

system is minor. The original calibrations for both of these watersheds were done using only mean daily flow data.

Figure 7-1-1 shows the simulation of mean daily flows for the 1950 snowmelt period for CSSL. This figure includes the original calibration, based only on daily flows, which had no surface or variable impervious runoff during snowmelt periods, and the final calibration, which used observed 6 hour instantaneous discharges to determine model parameters and as a result generated both of these runoff components. As can be seen the mean daily flow simulations in both cases are quite good. Figure 7-1-2 shows the instantaneous flow simulations from the original calibration that show a much more damped response than what actually occurred. Figure 7-1-3 shows a much more realistic reproduction of the instantaneous flows after the UZFWM value was reduced to produce surface runoff during high intensity snowmelt periods and the ADIMP parameter was used to generate fast response runoff when snowmelt rates were lower. The simulated instantaneous discharges are delayed somewhat from the observed because a 6 hour time interval is being used, whereas a hourly interval would be more appropriate for such a small, quick responding watershed, however, the amplitude is quite reasonable. Without instantaneous flow data one would not know which parameter set was most appropriate.

Figure 7-1-4 shows the simulation of mean daily flows for the original calibration from the UCSL. Figure 7-1-5 shows the simulation of instantaneous discharges at a 6 hour

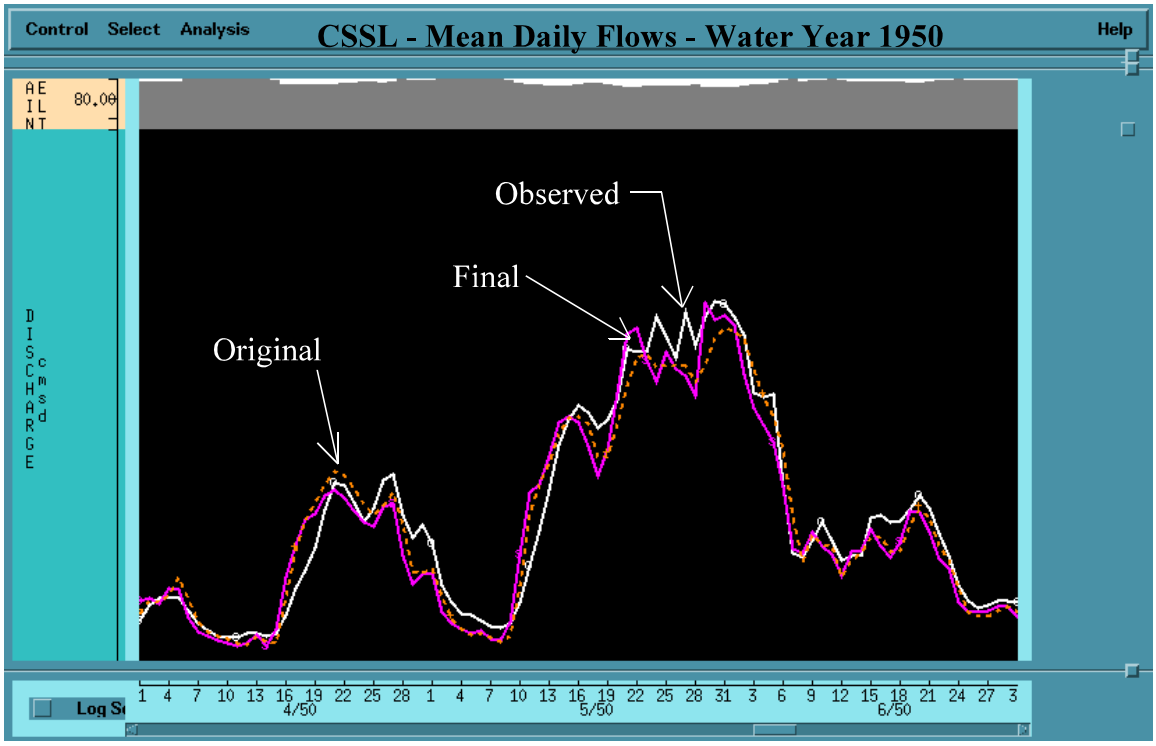


Figure 7-1-1. Simulation of mean daily flows for CSSL.

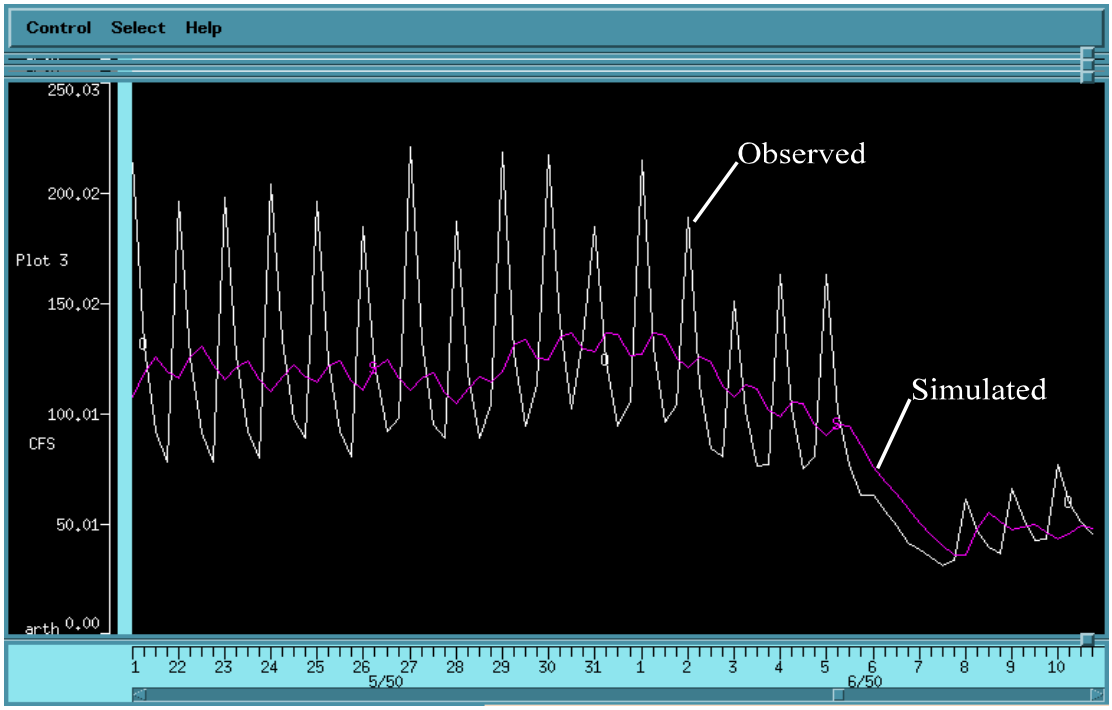


Figure 7-1-2. Simulation of instantaneous flows from the original calibration for CSSL.

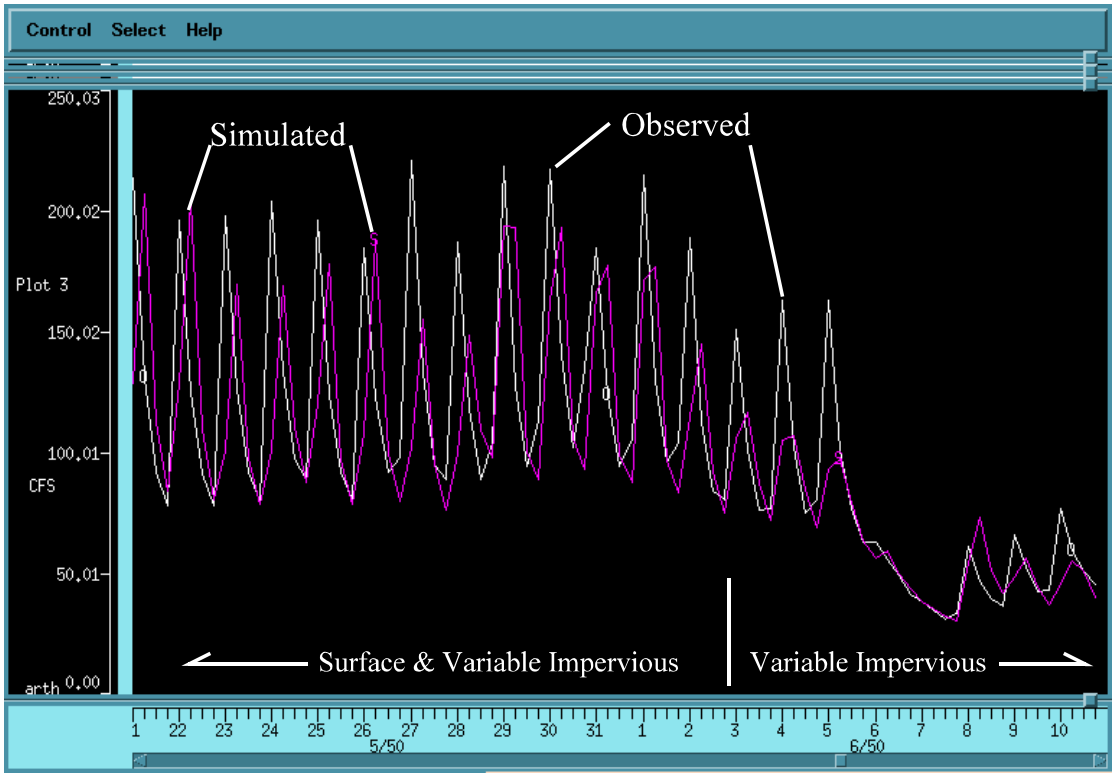


Figure 7-1-3. Simulation of instantaneous flows from the final calibration for CSSL.

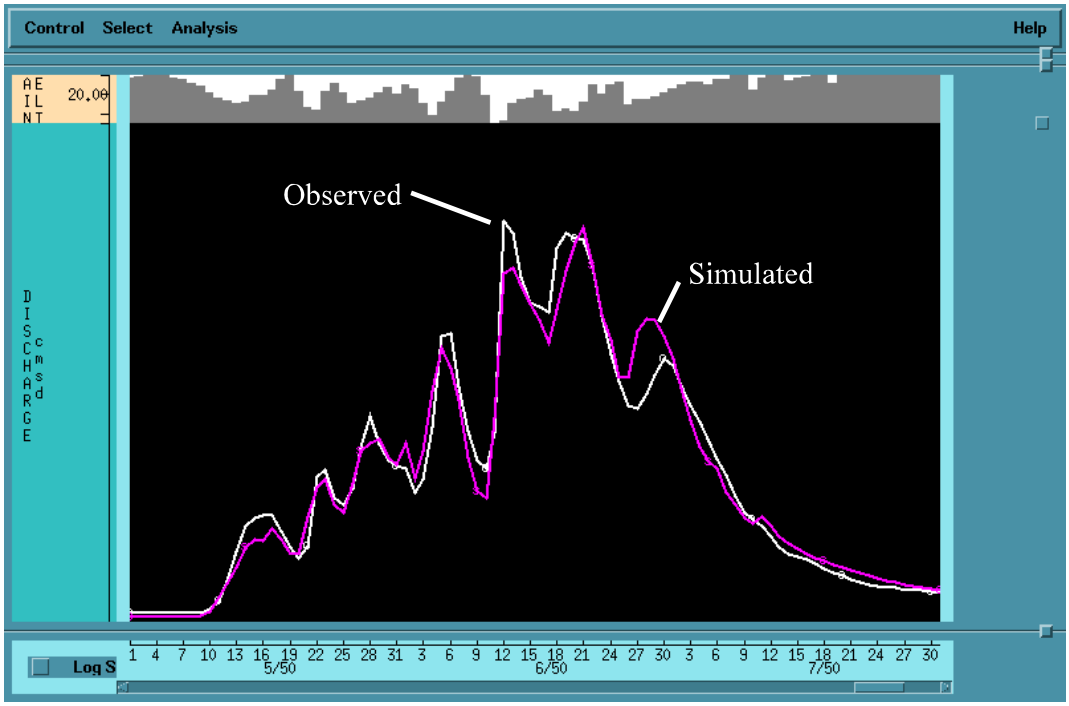


Figure 7-1-4. Mean daily flow simulation for the original UCSL calibration.

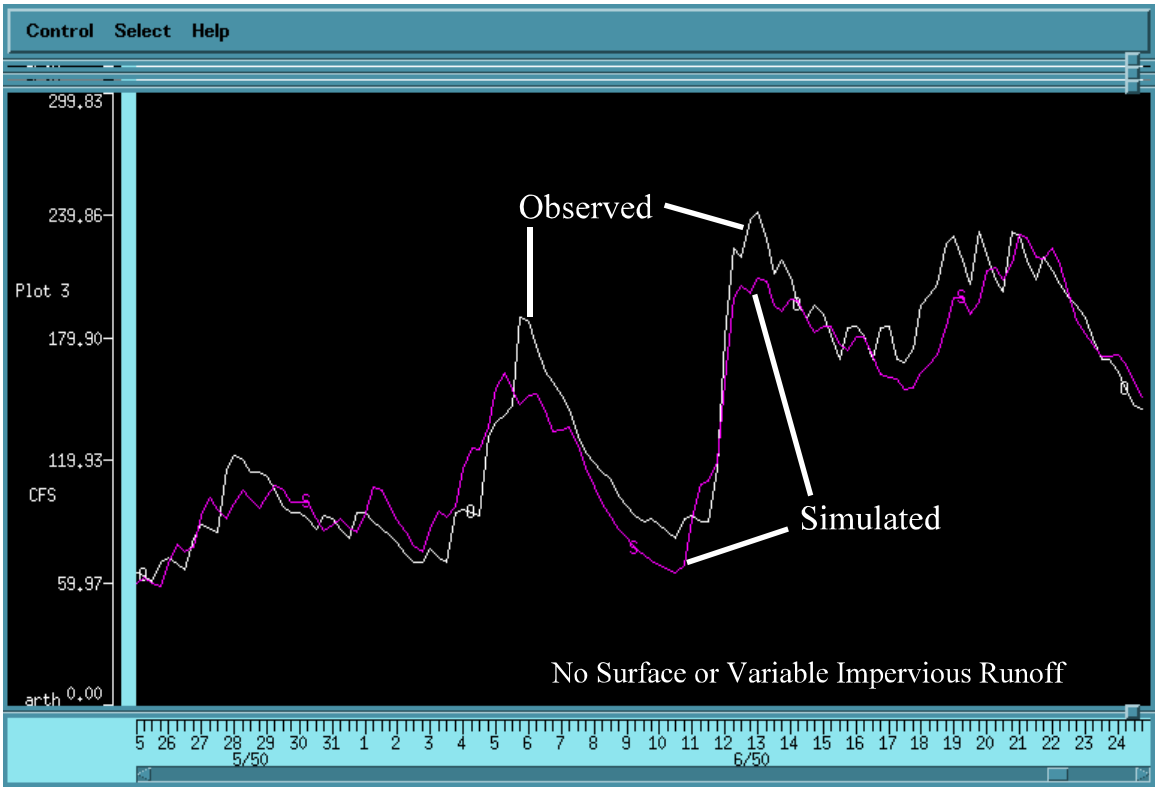


Figure 7-1-5. Instantaneous discharge simulation from the original UCSL calibration.

interval for this same parameter set. In this case the instantaneous flow simulation is quite realistic based on only using mean daily flows for calibration, however, one would not expect to find such a close correspondence in all cases. Unless one has observed instantaneous data available for examination.

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/home/eaa/calb_report/figures/peakflow.temp line 4, col 61, 3421 bytes

PEAKFLOW DISCHARGE AND TIMING ERROR SUMMARY

OBSERVED PEAK Q(CFS)	H(FT)	DATE	FLAG Q	H	SIMULATED PEAK Q(CFS)	DATE	TIMING ERROR(DAYS)	DISCHARGE ERROR(CFS)	DISCHARGE RATIO (SIM/OBS)
3620.0	8.1	3/19/1968	6		3150.0	3/20/1968	-1	-470.0	0.87
3420.0	8.1	1/26/1979	6		2610.0	1/27/1979	-1	-810.0	0.76
2580.0	7.2	4/ 7/1987	6		2910.0	4/ 7/1987	0	330.0	1.13
2300.0	6.7	6/ 7/1982	6		2300.0	6/ 8/1982	-1	0.0	1.00
1960.0	6.4	4/ 2/1993	6		1890.0	3/31/1993	2	-70.0	0.96
1780.0	6.3	6/ 3/1984	6		1690.0	6/ 1/1984	2	-90.0	0.95
1780.0	6.1	2/12/1970	6		1530.0	2/12/1970	0	-250.0	0.86
1690.0	6.0	3/27/1969	6		1620.0	3/27/1969	0	-70.0	0.96
1590.0	5.9	3/23/1972	6		1530.0	3/24/1972	-1	-60.0	0.96
1420.0	5.6	3/29/1978	6		1470.0	3/29/1978	0	50.0	1.04
1360.0	5.5	3/16/1986	6		1340.0	3/16/1986	0	-20.0	0.99
1260.0	5.5	3/13/1983	6		1470.0	3/13/1983	0	210.0	1.17
1250.0	5.4	3/15/1977	6		1540.0	3/15/1977	0	290.0	1.23
1160.0	5.2	2/ 4/1973	6		1370.0	2/ 4/1973	0	210.0	1.18
1030.0	5.0	1/28/1976	6		1590.0	1/29/1976	-1	560.0	1.54
1030.0	5.0	2/26/1975	6		1160.0	2/26/1975	0	130.0	1.13
1020.0	4.8	3/23/1980	6		1240.0	3/23/1980	0	220.0	1.22
899.0	4.5	3/17/1971	6		614.0	3/18/1971	-1	-285.0	0.68
857.0	4.5	2/27/1981	6		1430.0	2/27/1981	0	573.0	1.67
830.0	4.7	3/23/1974	6		1140.0	3/23/1974	0	310.0	1.37
779.0	4.3	4/23/1991	6		703.0	4/23/1991	0	-76.0	0.90
775.0	4.3	4/ 5/1990	6		717.0	4/ 5/1990	0	-58.0	0.93
739.0	4.3	5/13/1989	6		688.0	5/13/1989	0	-51.0	0.93
692.0	4.1	3/28/1988	6		960.0	3/28/1988	0	268.0	1.39
614.0	4.0	11/24/1991	6		870.0	11/24/1991	0	256.0	1.42
403.0	3.4	2/13/1985	6		484.0	2/14/1985	-1	81.0	1.20
MEAN: 1417.0	5.4				1462.0		-0.1	45.3	1.09

(OBSERVED DISCHARGE EVENTS ONLY)

"*" INDICATES SIMULATED PEAK ON SEARCH WINDOW BOUNDARY, WHICH PROBABLY IS NOT THE TRUE PEAK.

DISCHARGE RMS ERROR = 296.500 (CFS)
TIMING RMS ERROR = 0.760 (DAYS)
AVERAGE PERCENT ERROR (AVEOBSQ-AVESIMQ)/AVEOBSQ = 3.2 %
CORRELATION COEFFICIENT (DISCHARGE) : R = 0.936
BEST FIT LINE: OBSQ = A + B * SIMQ : A = -227.600(CFS) B = 1.120

Figure 7-1-6. Sample display from the PEAKFLOW operation.

- Peak flow data are helpful to determine how well the magnitude of the instantaneous peaks are reproduced, especially when instantaneous discharge data are not available at all or only for a small portion of the calibration period. These data are most valuable for fast responding streams, but can also be helpful when diurnal flow variations occur during snowmelt. The current PEAKFLOW operation tabulates a comparison of observed and simulated peaks by calculating the difference between the values and their ratio as shown in Figure 7-1-6. There is also a timing comparison, but only in terms of which day the peak occurred since the archived records don't contain the time of the observed peak during the day. It is somewhat difficult to use this tabular summary directly, since if mean daily flows are over or under simulated, the peaks on those days should also be similarly affected (i.e. if the highest daily flows are somewhat under simulated, it should be expected that the peaks should show a similar tendency, however, there is no tabulation of the daily flow bias just for the days with observed peak flow data). It would be better to compare the ratios of the observed peak to the observed mean flow for each peak flow day to the same ratios for simulated discharges. This would more clearly indicate whether the instantaneous flow simulations had a similar diurnal pattern as the observed discharge data.

• Snow observations can be used to verify the snow model computations by assisting to discern the actual form of the precipitation and to confirm whether the simulation of the snow cover is reasonable. Snow data that can be used for these purposes are snowfall, water equivalent and depth, and possibly areal extent of cover.

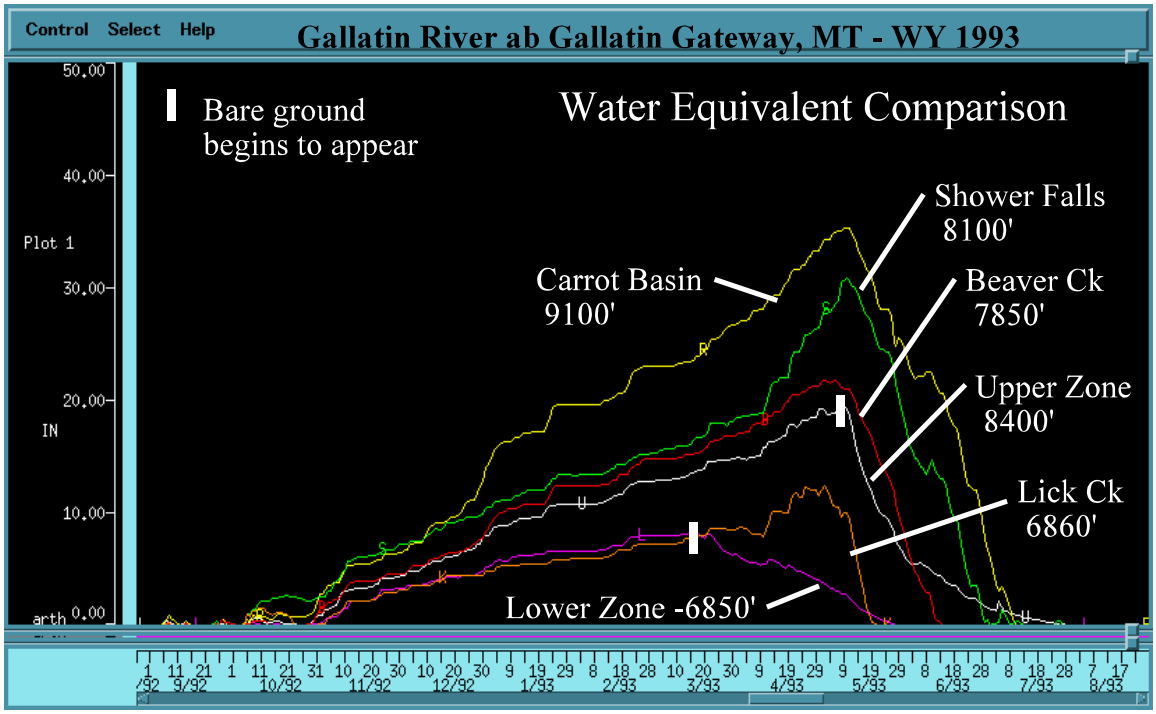
- Snowfall data – Data on new snowfall can be helpful in the determining if the form of precipitation selected by the snow model is correct or whether it needs to be modified. Section 7-4 includes of discussion of problems associated with the model determining the correct form of precipitation under certain conditions. This section also shows how

snowfall data can be used along with streamflow response and possibly water equivalent or depth data to check whether the form of precipitation needs to be changed.

- Water equivalent and depth data – Comparisons can be made between the mean areal water equivalent computed by the model and point or flight line observations of water equivalent to determine if the model computations are reasonable. There is always the problem that point or even flight line measurements only represent a small portion of a drainage area and thus there will seldom be a one to one correspondence between computed and measured values. Even so, observations of water equivalent can provide information as to how realistic are the model calculations. It is especially helpful to have several observations of water equivalent scattered through the area and at a range of elevations in the mountains. Generally the more observations, the more likely that one can realistically assess the model performance. In the mountains the model computations should typically show a pattern that is similar to snow course or SNOTEL observations during the accumulation period. During the melt period the observation sites should go bare before the mean areal water equivalent goes to zero. The model water equivalent will exhibit a more gradual decrease once bare ground begins to show up in portions of the watershed, while point measurements will go abruptly to zero. In flatter terrain where the time from complete cover to no snow conditions occurs over only a few days, this effect is less apparent. Figure 7-1-7 shows a water equivalent comparison for the Gallatin River above Gallatin Gateway, Montana. Model water equivalents for the upper and lower zones are plotted along with data from 4 SNOTEL sites.

If no water equivalent data are available, comparisons can be made between modeled water equivalent and observed depth of snow on the ground. When doing this one must remember that the density of the snow cover changes throughout the snow covered period, thus the relationship between water equivalent and depth is ever changing. In spite of this, depth data can be helpful to assess whether the model results are reasonable.

Figure 7-1-7. Water equivalent comparison for Gallatin River ab. Gallatin Gateway, MT.



- Areal extent of snow cover – Areal extent observations derived from satellite data are available for many western mountain areas for more recent years. The satellite estimates are available for entire watersheds and specific elevation bands. These observations can be compared against the areal extent of snow cover computed by the model, however, one must be aware that there should not be a one to one correspondence between the model areal extent and direct observations. The reason for this is that the areal depletion curve in the model implicitly includes other factors than just the areal snow cover as discussed in Section 7-4. The model should be using an areal extent that is less than that observed during the period when bare ground exists, with the largest discrepancies occurring in areas with the most rugged terrain. Thus, if one takes this difference between model and actual areal extent into account, the observations should be helpful in assessing whether the model results are reasonable.

Validation of Results

As indicated previously it is a good idea to validate the calibration results for the initial headwater watershed on other portions of the period of record. If done properly, this will test the extrapolation capabilities of the calibration results and determine if “curve fitting” occurred during the calibration (i.e. parameters were tweaked just to improve goodness of fit statistics). As mentioned earlier, the validation period should ideally include events that are outside the range of what occurred during the calibration period, i.e. the flood and low flow of record would

be in the validation period. Ideally statistics for the calibration and validation periods should be similar, though typically the calibration period statistics are slightly better as some degree of "curve fitting" is hard to avoid. A variety of statistics can be used to compare the results during the calibration period and one or more validation periods. These can include:

- Root mean squared errors, both daily flows and monthly volumes,
- Differences and standard deviations of monthly mean flows to test seasonal variations,
- Autocorrelation functions of observed flows and simulation residuals,
- Frequency distributions of observed and simulated flows, and
- Histograms of high flow peaks and their simulation errors.

Some possible causes of validation problems, i.e. differences in results between the calibration and validation periods, besides "curve fitting" include:

1. Calibration period is not adequate either because the validation period contains events that excite certain model components for the first time or events that seldom occurred during calibration occur during the validation period under somewhat different conditions (e.g. surface runoff may have occurred once during calibration with a certain spatial rainfall pattern, but occurs with a different rainfall pattern during the validation period). Also certain problems present, but unnoticed in calibration, may be amplified in validation.

2. Factors external to the models are changing over time, i.e. the observed flow data are not consistent due to changes in factors such as reservoir releases, agricultural practices, and vegetation and land use changes.

If these problems occur the suggested actions are as follows:

Case 1. Use the entire period of record to refine the calibration so that all possible model components are used and so that the maximum number of occurrences of specific situations are included in determining the most likely value of each parameter. This should only involve modifications to those parameters whose values were based on few, if any, events during calibration. Parameters whose values are based on only a few events or control components that were never used will contain a high degree of uncertainty. Parameters that sometimes fall into this category were mentioned in the section on Indeterminate Parameters in this section. For all parameters, but especially these, a subjective estimate of the degree of uncertainty (e.g. ± percentage of the parameter value to represent the standard deviation of the uncertainty) should be made for operational use.

Case 2. The observed flow data should be made consistent over both periods of record or

another watershed should be selected for the initial calibration. This case should not occur if the selection criteria for the initial headwater area were followed.

If a thorough validation is done for the initial headwater area within the river basin, validation is probably not necessary for the other drainages in the basin. Using the strategy recommended in this chapter, the parameter sets for all the other drainages in the river basin will be closely tied to the parameters determined for the initial headwater area.

Section 7-2

Calibration of Other Headwaters and Locals with Minimal Complications

Introduction

This section describes procedures and strategies to follow when calibrating other drainages within a river basin for which a good definition of the local flow contribution can be determined. The calibration for these areas will be closely tied to the results from the initial headwater area after taking into account information determined earlier about the spatial variations of physical, climatic, and hydrologic conditions over the basin. In order to determine if a drainage area falls into this category, it is necessary to examine the hydrograph produced from the area. In order to obtain this hydrograph it may only be necessary to look at the observed streamflow data or it may be that some computations must first be performed to remove the effect of control structures or upstream flows. Once the local drainage area hydrograph is established, an evaluation can be made as to whether model parameters can be reliably determined or whether there is so much noise in the hydrograph that a full calibration is not possible. This decision is subjective, but guidelines for making the choice include:

- Compare the hydrograph to the streamflow record for the initial headwater area or other headwaters that have no complications caused by such things as reservoirs, large lakes, diversions, or irrigation. Determine whether the general pattern is similar or, at least any differences are reasonable. For instance see if the hydrographs go up and down at the same time though the time to peak and amount of attenuation may vary, and if the magnitude and timing of snowmelt runoff makes sense.

- Examine baseflow periods to assess whether the general magnitude of the flow and the recession rate can be reasonably determined most of the time at these flow levels. A semi-log scale should be used when making this determination. Considerable scatter might occur at times when looking at low flow periods, but if such periods are few or the effects minor enough so that the baseflow pattern can be reasonably ascertained, a calibration of low flow parameters is possible.

Derivation of the Hydrograph for the Drainage Area

As mentioned previously for many headwater areas the observed mean daily flow data defines the hydrograph for the drainage area. In other cases computations may be necessary in order to determine the hydrograph that can possibly be used for calibration. This includes situations where the flow data need to be adjusted to natural flow conditions by adding or subtracting diversions, the inflow to an upstream reservoir is computed from outflow and storage data, and downstream local area contribution is derived by subtracting routed upstream flows from the observed flow at the gaging station. In a few cases the computations may involve a combination of these situations, e.g. a downstream local area may contain some diversions.

Adjustment to Natural Flow

In order to determine natural flow conditions to use for calibrating the snow, soil moisture, and channel models, the streamflow data need to be adjusted for man-made effects. Primarily this includes adjustments for diversions into or out of the drainage area. It could also include adjustments for the effects of reservoirs within the area, a large spring, or flood overflows across drainage divides. Section 6-6 discusses adjustments to mean daily flow records. Once the adjustments are completed, an evaluation can be made as to whether the resulting hydrograph can be used for calibration or whether too much noise exists.

Computation of Reservoir Inflows

The inflow hydrograph for a reservoir can be computed, if sufficient data are available, and possibly used to calibrate the models to the drainage area. Inflow hydrographs to any reservoir can be calculated though it is highly unlikely that the derived flows can be used for calibration other than the case of a headwater reservoir, i.e. a dam with no upstream calibration points. The water balance equation for a reservoir, as was shown in Section 6-6 (Eq. 6-6-1), is:

$$I - O + P \cdot A - E \cdot A + D = \Delta S \qquad (7\text{-}2\text{-}1)$$

where: I = Inflow,
O = Outflow,
P = Precipitation on water surface,
E = Evaporation from the water surface,
A= Water surface area,
D = Diversions into or out of the pool other than at the main outlet, and
ΔS = Change in Storage (each term is in volume units).

In order to compute the inflow, the outflow and the other terms must be known. At a minimum the outflow and change in storage terms are required. The change in storage can be computed from reservoir storage data using the DELTA-TS operation. If pool elevation measurements, rather than storage, are available, then they must first be converted to storage using the elevation-storage curve for the reservoir with the LOOKUP operation. Precipitation and evaporation could possibly be ignored if the water surface area of the reservoir is very small as compared to the drainage area above the reservoir, i.e. if the volume of water transferred at the water surface is small compared to the inflow. If precipitation and evaporation are included, the computations are relatively straight forward when the water surface area can be assumed to be a constant (LOOKUP operation used to convert depth of precipitation or evaporation to a volume based on the surface area). If the water surface area changes substantially over the range of pool elevations that occur, then this needs to be accounted for in the computations. In that case the volume of water transferred is a function of both the amount of precipitation and evaporation and the amount of storage or the pool

elevation. Unfortunately the 3 variable relationship operation in NWSRFS, LOOKUP3, is currently only programed to solve for depth as a function of discharge and depth or 2 discharge variables and is not in a general form for any 3 variables (this application would require solving for volume as a function of depth and volume or 2 depth variables).

In many cases when calculating the flow from the drainage area above a reservoir, errors in the various terms used to compute the inflow will result in considerable variability in the resulting time series. This is especially true at low flow levels and when the reservoir surface area is quite large. Also noise can result from problems caused by wind effects on the measurement of storage or pool elevation. High winds will tend to pile water up on the downwind edge of the lake resulting in inaccurate values of the change in storage term when the wind speed and direction differ greatly from one day to the next. For these reasons, in many cases it is not possible to derive a reservoir inflow time series that is of good enough quality to be used for model calibration.

Computation of Local Flows for Downstream Locations

The typical sequence for a downstream location is to route the upstream flows to the downstream gage and then subtract the routed flow from the total downstream flow to get the local area contribution. The resulting local area hydrograph can be examined to determine if a separate calibration can be done for this drainage or whether there is so much noise that it will minimize the chance of obtaining model parameter values with any degree of confidence. The flow to be routed downstream should be the adjusted instantaneous discharge, as discussed in Chapter 7, so that simulation volume errors from upstream locations are not carried downstream. In some cases, primarily based on where the local drainage network enters the main channel, it may be more realistic to add all or a part of the local area contribution to the upstream flow before routing. This situation is illustrated in Figure 7-2-1. In this case it would be most realistic to add the flow from Areas 1 and 2 to the flow at point A before routing since these areas enter the main channel at the very upstream

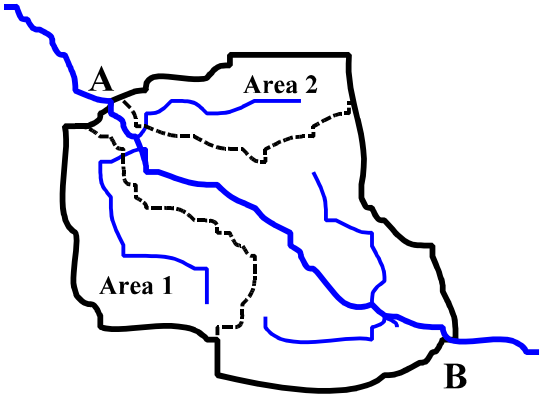


Figure 7-2-1. Routing

example.
end of the local drainage. In this situation a separate local area hydrograph is not obtained and a full calibration of the local area is not possible.

NWSRFS contains a number of routing models that can be used to move upstream flows to a downstream location. Generally the parameters of these routing models can best be determined from instantaneous discharge data for a number of storm events with different magnitudes of flow. In some cases routing model parameters can be adequately determined using mean daily flow data. This occurs mainly when the response time of the watersheds is relatively slow and there is a considerable amount of lag and attenuation in the river reach. The verification of the routing model parameters can be accomplished by using a long continuous record with a wide variety of flow conditions. An evaluation of the resulting local area hydrograph is one of the best methods of determining whether any further refinement of the routing model parameters is necessary. The steps in this process are as follows:

1. Generate adjusted instantaneous discharge hydrographs at each upstream location. If simulated instantaneous flows are not available at an upstream point, i.e. the area above that location has not been modeled, then adjusted instantaneous flows can be derived from just observed mean daily discharge data, rather than using the ADJUST-Q operation with both observed daily and simulated instantaneous flows. An option within the CHANGE-T operation allows for the generation of an instantaneous discharge time series from mean daily flow data. The shape of the resulting hydrograph should be quite realistic for areas with a slow response time, i.e. peaks in terms of days, but cannot be guaranteed for fast responding streams.

2. Route these adjusted instantaneous upstream flows, either separately or in some combination, to the downstream point.

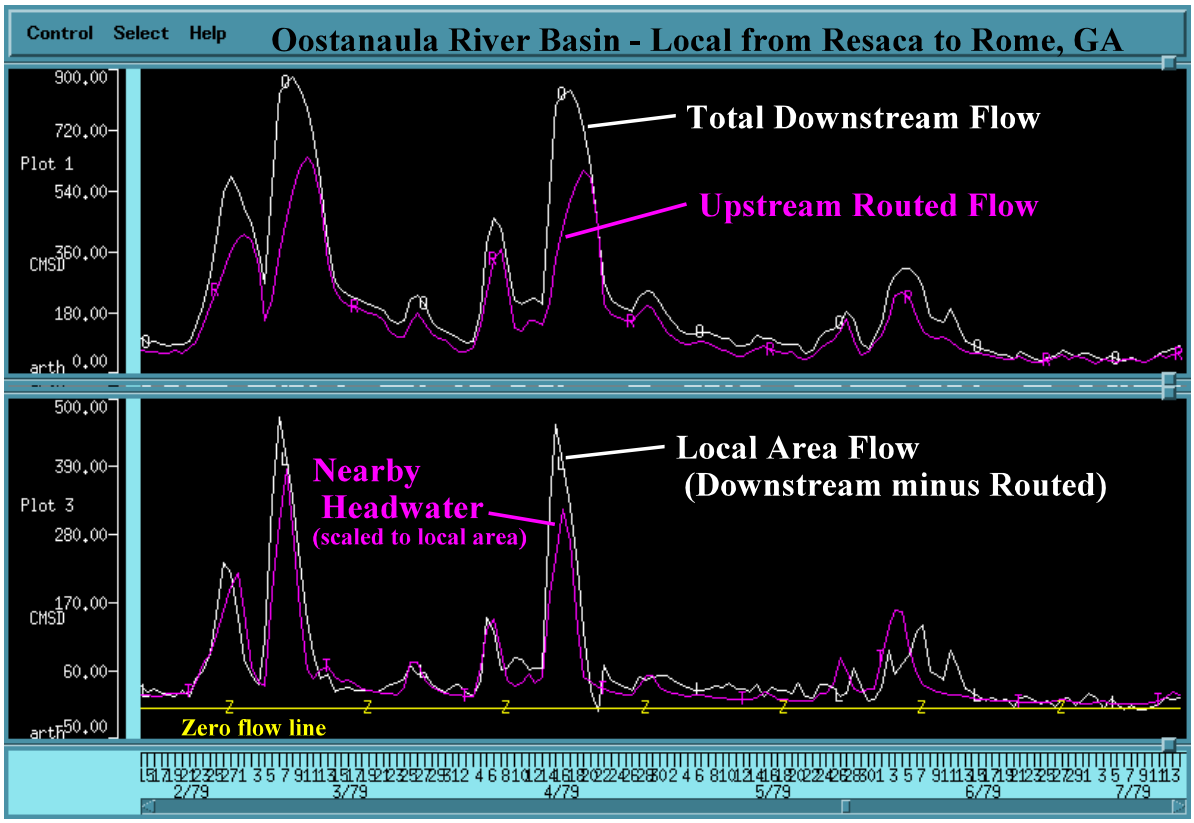
3. Use the MEAN-Q operation to produce a mean daily routed time series.

4. Subtract the routed mean daily flow from the observed mean daily flow at the downstream location to generate a mean daily hydrograph for the local area. This also can be done for instantaneous discharges for periods when observed instantaneous flow data are available at the downstream point.

5. Produce two daily flow plots to evaluate the results. One plot should contain the total routed and downstream observed discharges. It can also contain the local area flow, upstream inflows (before routing), and individual routed flows. The second plot would contain the local area hydrograph and discharges from one or more nearby headwater areas with minimal complications. The hydrographs for the nearby headwaters should be scaled to the local drainage area before plotting. These plots can be produced using two WY-PLOT operations or a single PLOT-TS operation. Using PLOT-TS has two

advantages. First, both plots can be viewed at the same time and second, negative values can be seen (generated when routed flow is greater than the downstream total flow - a zero flow line can be shown by having a time series included for which all values have been set to zero using the CLEAR-TS operation). The results are evaluated by seeing if the local area flow plot looks like it should by comparing it to the hydrographs for nearby headwater areas. Peaks should occur at the same time and the recessions should be smooth without any sudden dips caused by the routed flow being close to or exceeding the total downstream discharge.

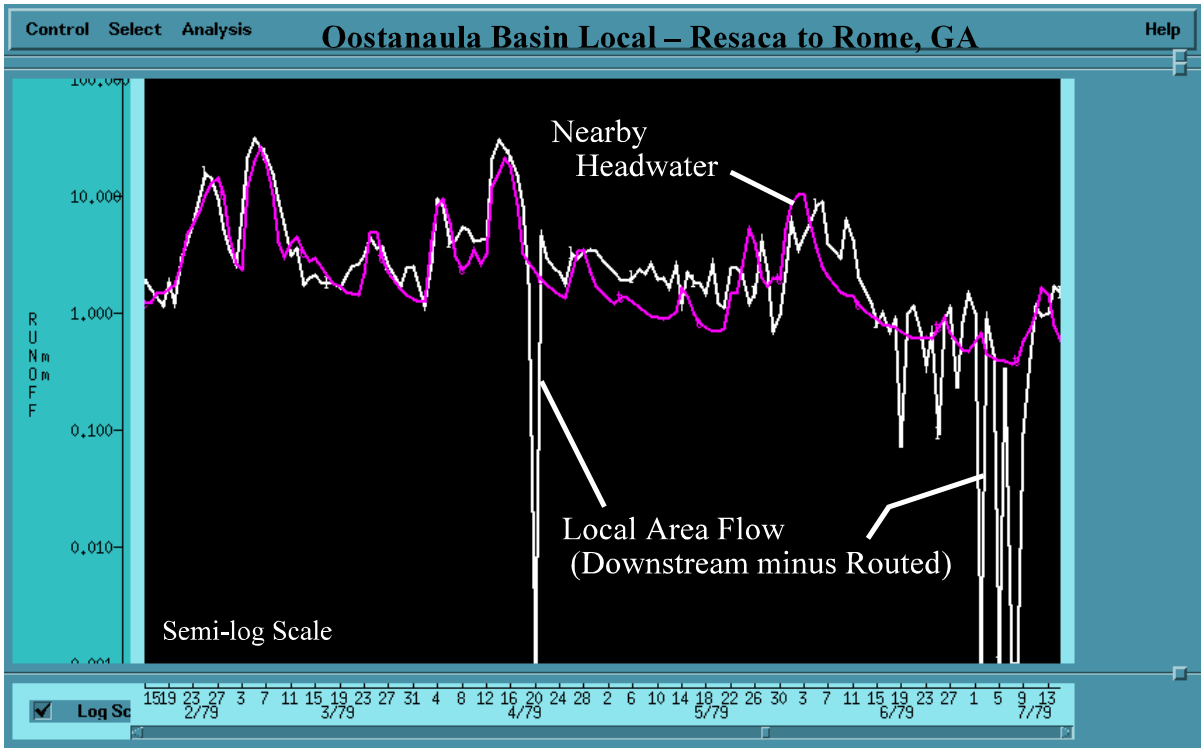
6. Modify the routing model parameters if needed to try to improve the resulting local area hydrograph and rerun the computations. Once the best possible results are obtained, decide whether the local area hydrograph is of sufficient quality to use for a full calibration. If a full calibration can be done, the local area daily flows can be saved to a file for use while calibrating (routing operations would not be included while working on the local area) or the local area flows can be recomputed on each run, rather than being read from a file, by including the routing computations. During calibration the simulated and "observed" (total minus routed) local flows would be used to determine and evaluate changes to model parameters. At the end, both the local and total area flows should be



examined before finalizing the parameter values.

Figure 7-2-2. Sample PLOT-TS display for deriving a local area hydrograph.

Figures 7-2-2 and 7-2-3 show examples of the types of plots used to evaluate the routing model parameters. Figure 7-2-2 is a PLOT-TS display that has two daily flow plots. The top plot shows the routed and total downstream flows. The bottom plot includes the local flow (total downstream minus routed), the hydrograph from a nearby headwater that is scaled to the local drainage area, and a time series of all zeros. In general the derived local area flow looks fairly reasonable though there are times when the local flow dips to zero or below due



to the routed flow being equal or exceeding the downstream discharge. There also appears to be more variability in the derived local flow than in the hydrograph for the nearby headwater at lower flows.

Figure 7-2-3. Sample WY-PLOT semi-log display for deriving a local area hydrograph.

Figure 7-2-3 shows a semi-log display of the lower panel of the previous figure using the WY-PLOT operation display. This plot accentuates the dips and the flow variations in the local area flow. While this derived local area hydrograph is probably of sufficient quality to make some limited parameter adjustments, it is not good enough for a full calibration.

Strategy to Follow

The recommended strategy to follow when calibrating these other headwater and local areas with minimal complications and for which a good definition of the hydrograph can be determined is:

1. First calibrate the other headwater areas with minimal complications. These headwaters should not contain any large discharge adjustments or be reservoir inflow points. These should be the next best drainages for determining parameter values. Start with a watershed that is close to the initial headwater and then move across the river basin. The steps to follow for each of these watersheds are:

a. Determine the calibration period – It is best to use the same period to calibrate the remaining watersheds as was used to calibrate the initial headwater area. In some cases this is not possible because the other headwater areas may not have observed streamflow data for this period or there may have been changes within the drainage area over time that dictate that another period be used for calibration.

b. Assign initial parameter values – Initial values for the snow and Sacramento models should be those determined for the most hydrologically similar area that has been previously calibrated. Different values could be used for some of the minor snow model parameters if there are significant differences in the typical amount of snow from one area to the next. Differences in soils and vegetation could be used to alter some of the Sacramento model parameters if there is sufficient confidence in the relationship between a given parameter and the physiographic information. Also, the ET-Demand curves could vary based on elevation and vegetation cover as described in Section 6-5. A unique channel response function should be derived for each new watershed as drainage areas and the channel network will vary from one watershed to the next.

c. Determine parameter adjustments -- Follow the strategy described in Section 7-1 for the initial headwater area with the exception of *only changing those parameters that clearly should be altered*. Possible changes to parameters that affect certain components of the models may be suggested based on the assessment of spatial variability that was described in Chapter 4, but parameter values should not be actually altered until closely examining the simulation results in a step by step manner. Modifications for large errors should involve changing as few parameters as possible. Then as you go through the various groups of parameters, i.e. starting with parameters that control low flows, then major snow model parameters, then tension water capacities, then storm runoff parameters, and at last the final adjustments, locate those portions of the hydrograph where the effects of each parameter can be isolated and then only change the parameter value when there is clear evidence that a change is justified. The magnitude of the parameter change doesn't have to be large, but there must be no doubt that the change is needed based on looking at multiple situations when the parameter's effects can be isolated. The statistics mentioned in Chapter 7 should be monitored during the calibration and may assist in determining which parameters may need to be modified. Also additional data, such as instantaneous and peak flows and snow observations should be used when available just like for the initial headwater area.

Making only those parameter changes that are clearly justified, will result in a consistent

and realistic variation in parameter values across the river basin. It is important to base any changes on a consistent pattern involving a large number of cases when the effect of a given parameter can be isolated. Parameter changes shouldn't be based on one or two events or a sight improvement in some 'goodness of fit' statistic. It is more important to achieve spatial consistency in the parameter values than to change values merely to improve some statistical value. By starting with parameters from a previously calibrated watershed and then looking at each parameter in an organized manner should also result in significantly reducing the time needed for calibrating the remaining drainages within the river basin, thus making the whole calibration process much more efficient.

2. Calibrate the remaining drainages in this category for which it has been determined that the hydrograph is of sufficient quality to support a full calibration. This includes locations with large flow adjustments, headwater reservoir inflow points, and downstream local areas. The sequence of steps and the strategy for making parameter changes should be the same as was just described for the other headwater areas. Many of these drainages will contain somewhat more noise in the "observed" discharge hydrograph. The more noise that exists, the harder it generally is to justify parameter changes, thus these drainages will typically have fewer parameter modifications than the main headwater areas. The time needed to calibrate these locations is largely dependent on the time needed to make flow adjustments or derive the hydrograph for the drainage area. The time needed for making snow and soil moisture model parameter adjustments shouldn't take very long if done in an organized manner.

Section 7-3

Model Parameters for other Headwaters and Local Areas

Introduction

This section discusses procedures for determining parameters for drainage areas for which a full calibration is not possible. These are portions of the river basin where the hydrograph for the drainage area, either headwater or local, contains considerable noise and those areas where there is not even sufficient information to derive the flow contribution from the area, e.g. the area above a reservoir that doesn't have storage or outflow data. For such areas the general approach is to assign snow and soil moisture parameters from the most similar previously calibrated watershed and then see if it is possible to make any adjustments to those parameter values to remove bias in the resulting simulation. In some cases it may not even be possible to produce a historical simulation for these drainage areas.

Strategy

The recommended strategy to follow for these drainage areas is as follows:

1. Determine the period to be used for possible calibration. It is best to use the same period as was used for the initial headwater area and all, or at least most, of the other locations within the river basin that have previously been calibrated, but in some cases this is not possible. Needed observed data may not be available for this area for that period or there may have been changes within the drainage area over time that dictate that another period be used for calibration. For downstream local areas it is best to use a period for which not only downstream flow data are available, but also observed upstream flows so that errors in upstream calibrations are not propagated downstream.

2. Attempt to derive a hydrograph for the drainage area as discussed in Section 7-2 to determine to what extent a calibration can be performed. In some cases it will not be possible to calculate the flows for the drainage area due to a lack of sufficient information. Examples include reservoirs lacking storage (or pool elevation to compute storage) or outflow data and areas with large diversions for which data are not available. If a hydrograph for the drainage area can be derived, make a judgement as to what level of calibration is possible based on the amount of noise in the calculated flows. This may range from a full calibration (in that case refer to Section 7-2 for the strategy to follow), to the possible adjustment of selected parameters, to only the chance of a crude adjustment to remove any bias in the simulation results at the downstream location. If a full calibration is not possible based on your judgment proceed to the next step.

3. Assign initial parameter values to the models used to simulate the local area contribution to the flow. Initial values for the snow and Sacramento models should be those determined

for the most similar area for which a full calibration was possible. Similarity is generally a subjective decision based on the assessment of spatial variability of physiographic features that was described in Chapter 4. When the hydrograph for the drainage area can be calculated with sufficient reliability, at least at certain flow levels, parameter sets from several calibrated areas can be tried in order to see which best represents this area. Different values could be used for some of the minor snow model parameters if there are significant differences in the typical amount of snow from one area to the next. Soils or vegetation information could possibly be used to alter some of the Sacramento model parameters if there is enough confidence in the relationships between the parameters and these data. Also, the ET-Demand curves could vary based on elevation and vegetation cover as described in Section 6-5. A unique channel response function (unit hydrograph) should be derived for each new watershed as described in Section 7-6 as drainage areas and the channel network will vary from one area to the next.

4. There are two basic cases at this point. The first is that a historical simulation is possible and calculated values can be compared with computed, adjusted, or observed data. The second is that a historical simulation is not possible due to a lack of information.

a. Historical Simulation Possible – In this case two basic situations exist. The first is that the natural flow from the drainage area, either headwater or local, can be determined. The second is that the effect of the control structures will be included as part of the historical simulation as opposed to just modeling natural flow.

When natural flow can be calculated it could be for a headwater reservoir where inflows are derived from storage, outflow, and possibly precipitation and evaporation data; for a headwater location where observed discharges have been adjusted to natural flow conditions to correct for diversions, changes in storage, and other factors; or for a downstream location where both local (total minus routed) and total flow, possibly first adjusted for the effect of control structures, are available. In this situation simulated streamflow is compared to these computed “observed” natural flow values. Simulating the flow should only involve the use of snow, soil moisture, and channel response models for the drainage area and routing models if this is a downstream location.

When the effect of the control structures are being modeled, total simulated discharge is compared directly to the total measured observed flow data which contains the effect of the structures. This involves modeling everything of significance that occurs within the drainage area. In addition to snow, soil moisture, channel response, and routing models; reservoir operations, irrigation usage, and diversions will be modeled.

In both of these situations the aim is produce an unbiased estimate of flow taking into account errors and noise in the data. Water balance components should be examined to make sure they are consistent with adjacent drainages. Precipitation and evaporation values should be reasonable if the guidelines in Sections 6-3 and 6-5 were followed when

the MAP time series and evaporation estimates were generated. Computed runoff values will contain considerable uncertainty when there are large diversions, reservoir effects, or irrigated areas or when the local area is small relative to the upstream area. Changes are generally not made to individual snow and soil moisture model parameters unless the hydrograph for the drainage area being modeled can be derived and is of sufficient quality to clearly indicate that such change is justified. When this occurs the magnitude of the change should be significant since the hydrograph is not of sufficient reliability to make minor adjustments. Even if the hydrograph is not well enough defined to isolate the effects and make changes to individual parameters, some general adjustments can often be made to correct for any bias that exists between the computed and "observed" values. This should only be done if there is reasonable confidence that the long term runoff is realistic. In that case, the following types of parameter changes can be considered:

- For volume bias, either overall or seasonal, adjustments can be made to the overall amount of precipitation, the snowfall correction factor, and/or the amount or seasonal variation in ET-Demand.

- For bias related to the magnitude of the flow (e.g. under simulation of low flows and over simulation of high flows), the Sacramento model percolation curve can be shifted upwards or downwards. This is done by changing LZFWM and LZFPM by the same ratio.

If any of these changes are made, it is important to make sure that the resulting values are physically reasonable compared to those for nearby drainage areas.

In this situation most of the effort is involved in calculating the "observed" flow to use and/or in determining parameters for the routing, reservoir, irrigation, or diversion models. The data are not of sufficient quality to allow for many changes to the snow, soil moisture, and local channel response models.

b. Historical Simulation Not Possible – A historical simulation is not possible when there is insufficient information to compute inflows to a reservoir, adjust discharges to natural flow conditions, or model everything of significance that occurred within the drainage area. In many of these cases the information and data necessary to simulate streamflow are available for operational forecasting use, but not historically. In a few cases the data may exist, but the method of operating the control structures are not included in any available modeling operation. One option in these cases is to add a new operation to NWSRFS that will handle the situation. When historical simulation is not possible at a given location within the river basin, the snow, soil moisture, and channel model parameters that were assigned to the drainage area cannot be adjusted. Also, simulated and adjusted instantaneous flows cannot be computed in the normal manner for routing to subsequent downstream locations. In these cases, instantaneous discharges must be generated from mean daily observed flows at the downstream location using the special

feature in the CHANGE-T operation for use in calibrating the next point down the river system.

Section 7-4

Initial Parameter Values for the SNOW-17 Model

Introduction

This section provides guidelines for assigning initial parameter values for the SNOW-17 snow accumulation and ablation model. These guidelines can also be used as an indication of whether parameter values being used for a given watershed or subarea are physically realistic. The snow model parameters are related to the physical and climatic features of the area being modeled. The main physical features that are important are vegetation cover and terrain features such as slope and aspect. The main climatic features are the typical amount of snow cover and the prevailing winter temperatures.

This section doesn't attempt to define the structure or the algorithms of the model. Readers are referred to section II.2-SNOW-17 of the NWSRFS Users Manual for a complete model description.

Form of Precipitation

The form of precipitation, i.e. rain or snow, can be specified in one of three ways for the SNOW-17 model. The methods of determining the form of the precipitation are:

1. By a threshold temperature which is the parameter PXTEMP. If the air temperature for the current time interval is less than or equal to PXTEMP, then the precipitation is assumed to be snow. If the temperature is greater than PXTEMP, then the precipitation is treated as rain.

2. By a rain-snow elevation time series. This time series specifies the elevation above which snowfall is occurring and below which the precipitation is in the form of rain. In this case an area-elevation curve is input to the snow model so that the fraction of the area with snow and the fraction with rain can be computed based on the rain-snow elevation value for each time interval. The rain-snow elevation time series can be generated by the RSNWELEV operation in NWSRFS. This operation uses either an air temperature time series for a specified elevation or a freezing level time series along with a user specified lapse rate and the PXTEMP parameter to calculate the elevation at which the PXTEMP temperature value occurs. The equation used by the RSNWELEV operation is:

$$E_{rs} = E_v + ((T_v - PXTEMP) \cdot (100/L_p)) \qquad (7\text{-}4\text{-}1)$$

where: E_{rs} = Rain-snow elevation (m),
E_v = Elevation associated with T (m),
T_v = Temperature at elevation E (°C),
L_p = Lapse rate during precipitation (°C/100m).

When air temperature is the input variable, T_v is the temperature for the time interval and E_v is the elevation associated with the time series. When freezing level is the input variable, E_v is the freezing level elevation and T_v is 0°C. The lapse rate used is typically the saturated adiabatic lapse rate (0.55°C/100 m or about 3°F/1000 ft) since the only relevant time periods are those when precipitation is occurring. If the value of the rain-snow elevation is missing for a given time interval, then method #1 is used to specify whether all the precipitation is rain or snow.

Generally when calibrating an MAT time series is used as the input to the RSNWELEV operation since historical freezing level data are seldom available. For watersheds with multiple elevation zones there is a question as to which MAT time series should be used to determine the rain-snow elevation for each zone. Physically there is only one rain-snow elevation for the watershed for a given time interval. This would suggest using the MAT time series for only one of the zones to compute a single rain-snow elevation to be used by all the zones. However, for example, if the rain-snow line was near the mid-point of the upper elevation zone and the MAT for the lower zone was being used to calculate the rain-snow line, the lapse rate would be applied over a considerable elevation range which would likely result in a less accurate estimate of the rain-snow line than if the upper area MAT was being used. This would suggest using the MAT time series for each elevation zone to compute the rain-snow elevation for that zone to minimize the elevation range over which the lapse rate was being applied. In this case when the actual rain-snow elevation is within a given elevation zone, the MAT for that zone would be used to estimate the elevation and although different rain-snow elevations would be calculated for the other zones, they would generally be outside the elevation range of those zones and not affect the form of the precipitation. Which MAT time series are used to estimate the rain-snow elevation for watersheds with multiple elevation zones is also dependent on the elevations of the stations used to compute the MAT values for each zone. If a zone has no temperature stations within its elevation range, the MAT has been calculated by applying seasonally varying lapse rates over a large elevation difference, and not the saturated adiabatic lapse rate. In this case, the MAT could be more likely to be in error during periods when precipitation occurs.

Based on the discussion in the previous paragraph, the following recommendations are offered for computing rain-snow elevations for watersheds with multiple elevation zones for use during calibration.

- If there are actual temperature stations within an elevation zone that were used to compute MAT, then the MAT time series for that zone should be used to calculate the rain-snow elevation for the zone.

- If there are no temperature stations within an elevation zone that were used to compute the MAT for the zone, then the MAT time series for the zone that has the smallest elevation difference to the zone in question and has actual temperature stations within its elevation range should be used to compute the rain-snow elevation.

Operationally a single rain-snow elevation should be used for all elevation zones so that only one value has to be changed to alter the division between rain and snow.

3. A time series can be used to specify the fraction of the precipitation that is in the form of snow during a given time interval. This time series is referred to as a percent snowfall time series. The values assigned to this time series are generated outside of the operations table. Historically the information needed to specify the form of precipitation might only be available for a research location. Operationally a procedure involving meteorological data, observations of the form of precipitation, and forecaster interaction could be devised to produce a percent snowfall time series. If the percent snowfall value is missing for any time interval, then method #1 is used to determining if all the precipitation is rain or snow.

Most studies have indicated that a reasonable threshold temperature for separating rain from snow is around 1 - 1.5°C. The Snow Hydrology manual indicates about 45% snow, 25% mixed, and 30% rain at 34°F and 33% snow, 17% mixed, and 50% rain at 35°F. The Snow Hydrology manual also shows cases of rain occurring when the surface temperature is 29°F and snow at 40°F. In the intermountain west, at least at higher elevations, winter precipitation is almost always in the form of snow even when the temperature is near 40°F. Usually in this region the temperature drops fairly quickly after the onset of snowfall, but it may be fairly warm when the event begins. It is clear that air temperature measured at ground level is not a perfect indicator of the form of precipitation. In NWSRFS the problem is compounded by only using max/min air temperature to compute historical MAT time series. The assumptions of when the max and min occur and the use of a fixed diurnal temperature variation result in incorrect temperature values during certain periods (see section titled “Limitation of Current NWSRFS Historical MAT Program” in Section 6-4). This most commonly occurs when temperatures are changing substantially from one day to the next which typically occurs with a frontal passage which is also when precipitation is most likely. From experience it is clear that the use of historical MAT values computed from only max/min temperature results in more cases of the form of precipitation being in error than the use of a threshold temperature to separate rain from snow. The use of the threshold temperature in the snow model to determine the form of precipitation should be less of a problem operationally because the OFS MAT preprocessor uses both instantaneous and max/min temperature data during the observed data period. The instantaneous temperatures are used to determine the daily temperature variation for stations that only report max/min values.

It is recommended that a value of PXTEMP in the range of 0.5 to 2.0°C be used with the SNOW-17 model. Generally a value of 1.0°C should be adequate. In many intermountain west areas it may be necessary to use a considerably higher value, in the range of 3-5°C in order to minimize errors in the form of precipitation (in this region temperatures frequently go well above freezing during portions of days with snow showers, however, the temperature drops back to near freezing whenever the precipitation intensifies and generally in the winter and early spring all precipitation is in the form of snow). It is also recommended that once a value of PXTEMP is established for a given region, that it should be used for all areas within the region. It is not

worth the effort to try to varying PXTEMP for each area to try to get a near equal number of cases when rain should have been snow and vice versa. Since most of the problems are caused by the use of only max/min temperatures, varying PXTEMP is generally not very productive. The use of only max/min temperatures also results in a bias in the form of precipitation determination in many regions, e.g. errors in the daily temperature variation result in more cases of rain being typed as snow than vice versa. Thus, there is a built-in bias that varying PXTEMP cannot correct. It is best to assign a reasonable value to PXTEMP and then correct the form of precipitation for individual events as needed. Also operationally it is much easier to deal with a single value of PXTEMP for all areas in a given region than to have to remember or look up a unique value for each area.

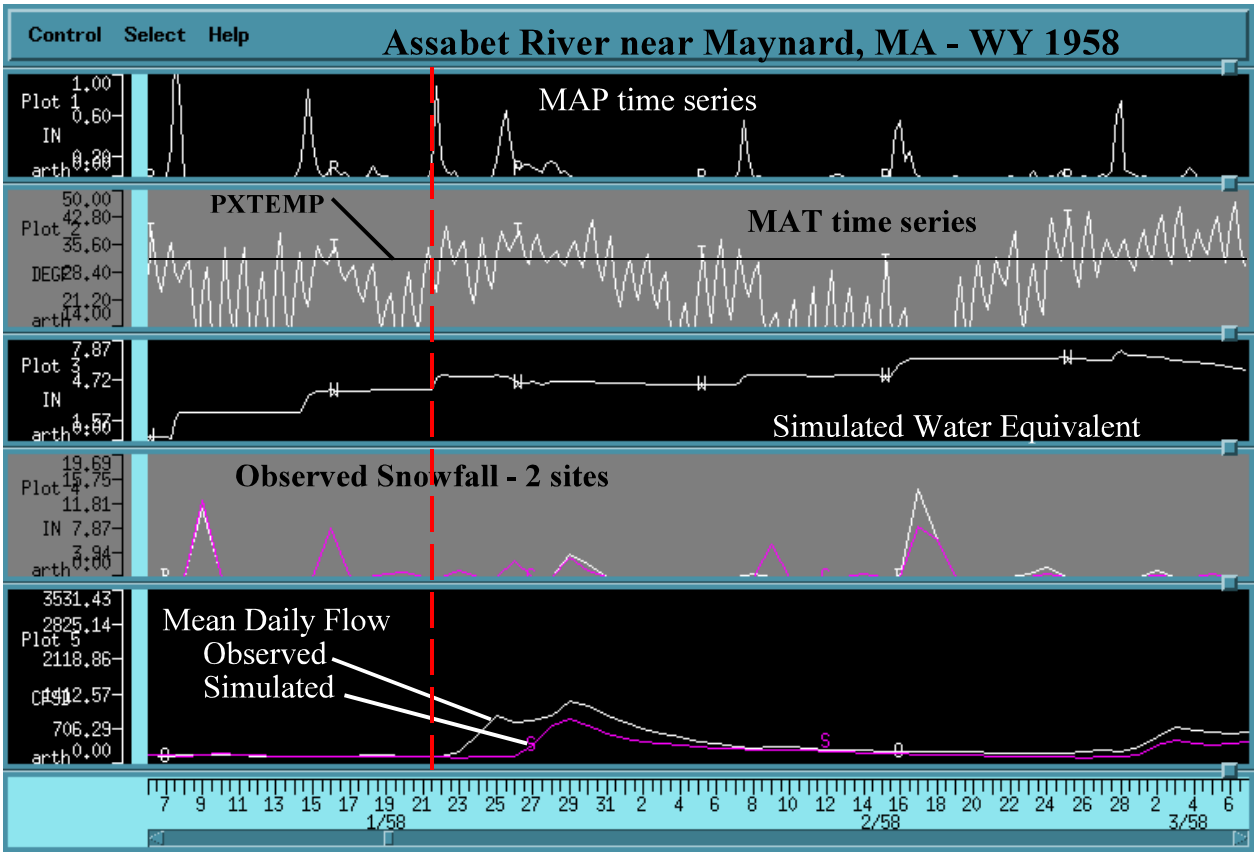
From a calibration standpoint, it is only necessary to correct the form of precipitation for individual events when the mistyping causes so much simulation noise that it is impossible to determine the other model parameters with confidence. Obviously having the correct form of precipitation for all events will produce the best results and minimize goodness of fit statistics, however, the aim of calibration is to determine the proper parameter values for operational use and not to minimize certain statistics. Thus, the form of precipitation only needs to be corrected for individual events when errors interfere with determining the proper values of other snow model parameters and parameters for the other models being used. It is not cheating to correct the form of precipitation when it interferes with determining parameter values. Operationally there should be much more information available, besides air temperature, to indicate whether it is raining or snowing at a given location. In many regions, especially those with temperatures below freezing during most precipitation events such as the north central states, the intermountain west, and Alaska, it is necessary to correct the form of precipitation for few, if any, events. Only events with a large of amount of precipitation that is in the wrong form typically need to be corrected in these regions. In other regions, where the temperature is frequently near freezing during periods of precipitation, there can be many events where the form of the precipitation is mistyped. This case is very common in the northeastern part of the United States. There can also be problems with the division of rain and snow west of the Cascade Range in the Pacific northwest due to errors in the diurnal temperature variation. In this region the rain-snow elevation option is generally used to separate rain from snow since rain occurs during most winter events at lower elevations whereas it is snowing higher up in the mountains.

Correcting the form of precipitation is also needed when the historical time series are to be used for ESP applications. For calibration, significant errors in the form of precipitation only need to be corrected for the period used to calibrate the models. For ESP applications, the entire historical period of record should be corrected. If not, bias produced by an improper assignment of the form of precipitation will influence the probabilistic distribution of the ESP products. It is probably only necessary to correct larger events that are in error, just like calibration. Errors in small events shouldn't affect the probability distributions generated by an ESP application. It is especially important to correct the form of precipitation for watersheds where there is a tendency for the errors to be dominantly in one direction (e.g. many more cases of rain treated as snow than vice versa).

When attempting to correct the form of precipitation it should be done in a systematic manner and with as much information as possible to support any changes that are made. The form should not be changed just to get a better fit between simulated and observed flow values. Generally the form of precipitation is altered by changing the MAT value for one or more time intervals. The information that can be used to support the decision includes streamflow response, snowfall observations, and water equivalent and depth data. In data sparse regions, such as portions of Alaska, the decision may have to be solely based on streamflow response since the other data are not available. In that case changes should only be made when the streamflow response clearly indicates that the form of precipitation is most likely in error.

The use of the PLOT-TS display in ICP can be very helpful when trying to determine when the form of precipitation is in error and what MAT values need to be changed to correct the problem. All of the data needed to decide whether the form of precipitation as determined by the model is correct or whether it needs to be changed can be shown on a single display. Figure 7-4-1 shows an example of such a PLOT-TS display. In this case 5 plots are included. The top plot shows the 6 hourly MAP time series. The next plot down contains the 6 hourly MAT values. The next plot is the 6 hourly simulated water equivalent. The next plot is daily snowfall observations from 2 climatological stations near the watershed. The bottom plot shows the mean observed and simulated

Figure 7-4-1 Sample display for hydrologic analysis



observed mean daily flow. The exact value and time of any plotted value can be displayed by clicking on the display. By examining all of this information it can be decided as to whether the form of precipitation needs to be altered. For example, in the figure during the first period on the 22nd of January a large amount of precipitation is typed as snow since the MAT value is well below PXTEMP. This causes the simulated water equivalent to increase. However, the climate observers reported very little snowfall during the event and the observed streamflow showed a considerable increase in the subsequent days, both indicating that the precipitation was most likely predominately rain. Thus, the MAT value was changed to be greater than PXTEMP with an editor. The MAT was most likely in error during this case due to the minimum temperature recorded on the morning of the 22nd actually occurring on the previous morning. By proceeding through the period of record using such a display, problems caused by the form of precipitation being in error can be systematically corrected before continuing with the calibration.

Major Snow Model Parameters

The major parameters of the SNOW-17 model are the snow correction factor SCF, the non-rain melt factors MFMAX and MFMIN, the areal depletion curve and the related SI parameter, and sometimes the average wind function during rain-on-snow events UADJ. These are the parameters that typically have the greatest effect on model results and are the only ones generally modified during calibration.

SCF – SCF is the only parameter in the snow model that has a significant effect on the volume of water available for snowmelt runoff. It is used to adjust any precipitation that is typed as snow. The main reason for SCF is that most precipitation gages under catch snowfall as shown in Figure 7-4-2. The under catch is related to the wind speed at the orifice of the gage and the effect can be reduced by the installation of a windshield. The wind speed at the gage is influenced greatly by the exposure of the site. Most climatological stations are situated where there is protection from the wind whenever possible. A site with a good exposure doesn't need a windshield. Of course the MAP time series generally involves the weighing of a number of gages with different exposures and possibly some with and some without windshields.

The SCF parameter is related to the average wind effect on snowfall catch over all the events during the period of record. The catch deficiency varies from event to event, and even within an event, as the wind speed at the orifice changes. If there are a large number of events involved in the build up of the snow cover, the effect of variations in catch deficiency from event to event tend to cancel out, though there may be some years when there is more wind during snow storms than other years.

Besides explicitly trying to account for the average catch deficiency during snowfall events, the SCF parameter implicitly includes other factors that affect the accumulation of a snow cover that are not included in the snow model. This includes sublimation losses, both from

the snow surface, from snow intercepted by the forest, and from wind blown snow, and snow

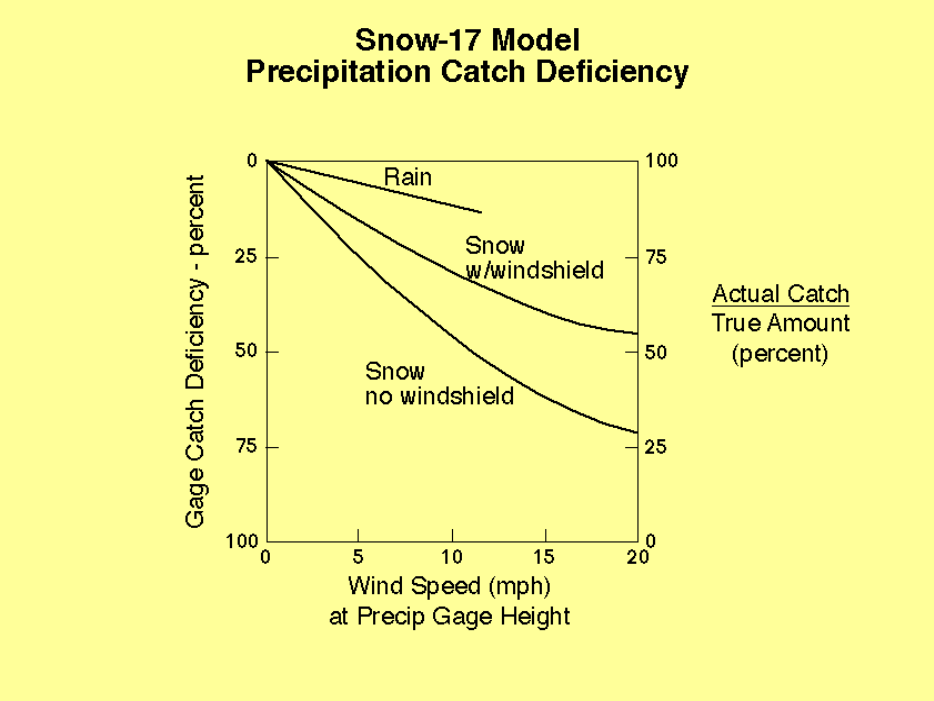


Figure 7-4-2. Typical precipitation catch deficiency due to wind.

blown across the watershed divide. Sublimation losses tend to be small in most areas, but can be moderately significant in places with low humidity and high wind speeds during the winter. Generally the amount of snow transported from one watershed to another is negligible in most regions.

After considering all these factors, a reasonable initial value for the SCF parameter is in the range of 1.1 - 1.2. Higher values would be expected in watersheds where there are few gage locations with good exposures such as in the northern plains. In such a region SCF values could start in the range of 1.3 - 1.6. In regions where all the precipitation stations have good exposures and there are significant sublimation losses, SCF could be equal to 1.0 or even slightly less without being physically unrealistic.

The value of SCF can also be affected by whether water balance computations were used to determine the magnitude of the MAP input for the model. When mean annual precipitation

is estimated directly from the water balance as in the method described in Section 6-3 under "Determination of the Average Mean Areal Precipitation in Data Sparse Regions", the processes related to SCF are implicitly absorbed in the MAP values. In this case in order for the precipitation used by the model to be the same as the amount computed by the water balance, SCF should be initially set to 1.0. Since gage catch deficiencies are not considered when deriving the average annual precipitation with this method, the net result could be not enough snowfall and too much rain. When a water balance analysis is used primarily to identify portions of the basin where the isohyetal analysis needs to be modified as described in Section 6-3 under 'Determination of Average Mean Areal Precipitation for Each MAP Area", there is less chance that the value of SCF will be affected. In any case SCF should not be used in regions dominated by snowmelt runoff to correct for volume problems resulting from an inadequate precipitation analysis.

MFMAX and MFMIN – These are the maximum and minimum values of the seasonally varying non-rain melt factor. MFMAX occurs on June 21st and MFMIN on December 21st. In regions where the snow cover generally builds up throughout the winter and doesn't melt until mid or late spring, MFMAX dominates the computations of melt. In regions where snowmelt periods can occur anytime throughout the cold season, both factors are important.

The non-rain melt factor determines the melt rate when the area is completely covered by snow, thus it has a controlling effect from when melt first begins until significant bare ground exists. This melt factor is based on the average relationship between air temperature and melt at the surface of the snow cover over the entire period of record. It is to be expected that there will be situations when the actual melt rate is greater or less than the average. Adjustments can be made during these situations operationally as discussed in Section 8-1, but during calibration it is the average melt rate that is sought. The average value of the melt factors are physically related to forest cover, slope and aspect, and typical meteorological conditions. Some items to consider when choosing or evaluating values for the melt factors are:

- If an area had an extremely dense forest cover, i.e. so dense that no sunlight reached the forest floor and calm conditions always prevailed such that only longwave radiation exchange between the forest canopy and the snow below caused melt, a lower limit for the melt factor could be computed. This value would be about 0.32 mm/°C/6 hr and there would be no seasonal variation. Such a situation doesn't exist for an area of any reasonable size.

- If a region, such as the north Pacific coast, has persistent cloud cover most of the time, this will act similarly to a dense forest cover. It would be expected that the maximum melt rate would be relatively low and there wouldn't be a pronounced difference between MFMAX and MFMIN. If cloud cover generally persists for some months, but not for others, the seasonal melt factor variations built into the model (normal sinusoidal or Alaska, northern latitude, pattern) may not be appropriate. In this situation there is an

option in the SNOW-17 operation for the user to specify the seasonal variation pattern. This option should be used only when it can be substantiated that the seasonal variations built into the model are not adequate.

• The melt rate should be the greatest for a large open area in a region where dew-points typically exceed 0°C and windy conditions prevail during periods of substantial melt. In such an area all forms of energy exchange could contribute to melt. In general, in an area with low humidity, solar radiation will be the dominate cause of melt as atmospheric radiation will be less than that emitted by the snow surface due to generally clearer skies and negative latent heat exchange will tend to offset positive sensible heat exchange. This would result in a lower average melt rate than for an area with generally higher humidity for the same amount and type of forest cover.

• Open zones in the mountains, such as above tree line, generally have lower melt rates than a flat open area due to variations of slope and aspect caused by the rugged terrain. This is especially true early in the melt season when the non-rain melt factor controls the computations as melt is typically not occurring over an entire mountainous zone.

•In northern and high elevation regions where the snow ages slowly and melt seldom occurs during the mid-winter period, the MFMIN parameter will tend to have lower values than in regions where substantial melt frequently occurs at various times throughout the winter.

• The largest ratio of MFMAX to MFMIN should occur in mostly open areas in regions with generally low humidity levels and infrequent mid-winter melt periods, such as high elevation zones of the intermountain west, portions of interior Alaska, and other places where MFMIN tends to have low values. The smallest ratios of MFMAX to MFMIN are expected in heavily forested areas in regions with more humid climates and periodic mid-winter melt, such as coastal areas and much of the eastern continental United States.

Based on these considerations and prior calibration results the following recommendations are offered for selecting initial values for the MFMAX and MFMIN parameters.

Description of Area	MFMAX	MFMIN
Dense conifer forest or persistent cloud cover	0.5 - 0.7	0.2 - 0.4
Mixed cover - conifer, deciduous, open	0.8 - 1.2	0.1 - 0.3
Mostly deciduous	1.0 - 1.4	0.2 - 0.6
Mostly Open flat terrain mountainous terrain	 1.5 - 2.2 0.9 - 1.3	 0.2 - 0.6 0.1 - 0.3

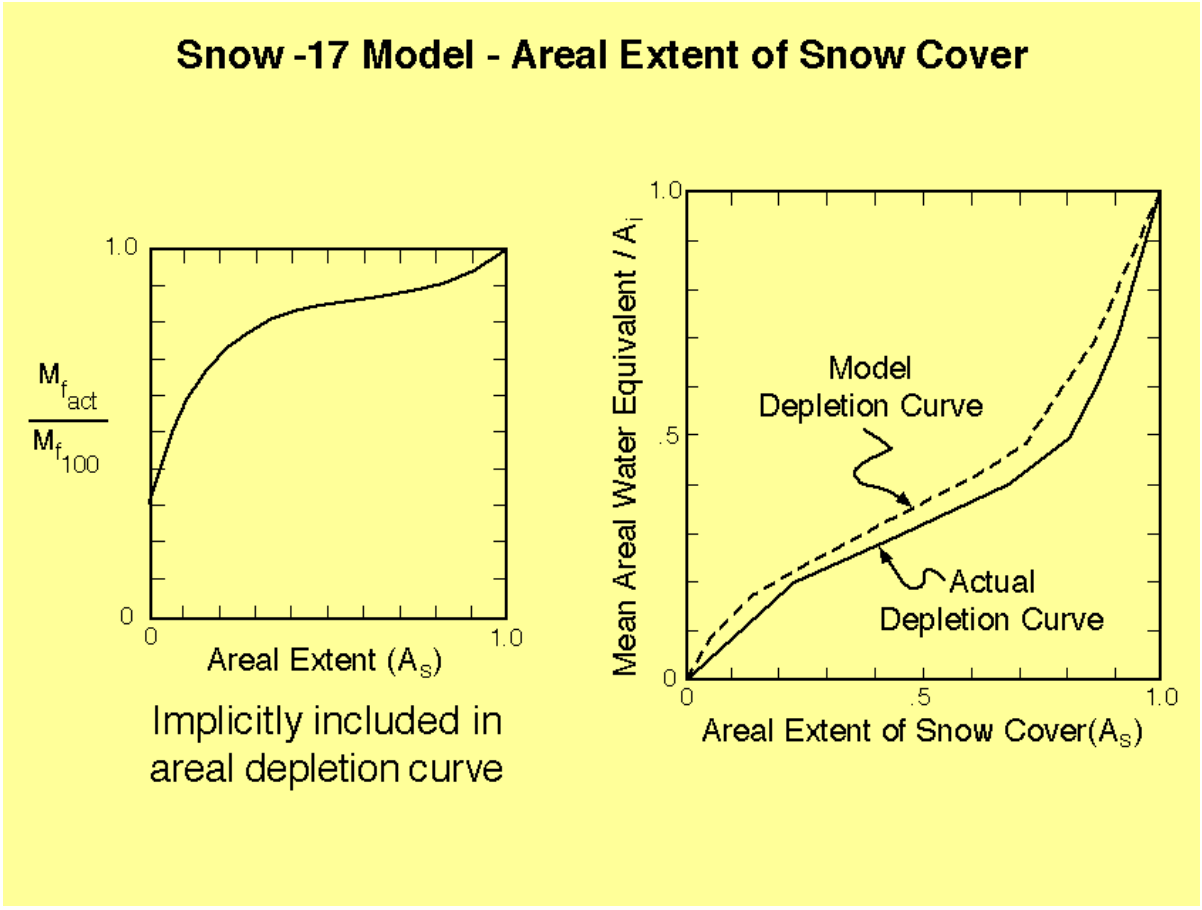
Table 7-4-1. Suggested initial values for MFMAX and MFMIN.

These guidelines can also be used to decide if the parameter values determined during calibration are reasonable. If the calibrated values fall significantly outside these ranges, it could be due to biased temperature data, i.e. the MAT values being used are warmer or cooler than what actually occurred in nature.

Areal Depletion Curve -- The areal depletion curve controls the melt rate when only portions of the area being modeled are covered by snow. The idea behind the areal depletion curve is that the pattern of snow accumulation and ablation is similar from one year to the next. For example, during the accumulation period the variation in the amount of snow on the ground, including the location of drifts and shallow places, is similar each year due to terrain and vegetation effects and prevailing storm and wind directions. Also, the melt pattern is similar with lower elevation, south facing, bare slopes melting first and higher elevation, north facing, forested or sheltered areas retaining snow the longest. The main function of the depletion curve is to account for how much of the area is covered by snow, however, like in any conceptual model, everything that is occurring in nature is not explicitly being modeled. Besides accounting for the areal extent of the snow cover, the depletion curve also absorbs variations in the melt rate over the snow covered portion of the area. The non-rain melt factor determines the melt rate when there is 100 percent cover, but as the snow cover depletes the melt rate over the area that still has snow is reduced. This occurs because the portions of the area with the highest melt rates, such as south facing slopes, typically go bare first. The areas where snow generally remains the longest are forested or sheltered north facing slopes with much lower melt rates. This variation in melt rates ends up being

Figure 7-4-3. Model versus actual snow cover depletion curve.

implicitly included in the areal depletion curve. Thus, the areal depletion curve is really calculating an "effective" areal extent of snow cover and not the actual areal extent. This is illustrated in Figure 7-4-3. This figure shows a possible variation in the ratio of the actual melt rate $\left(M_{f_{act}}\right)$ to the 100 percent cover rate $\left(M_{f_{100}}\right)$ as a function of the areal extent of the snow cover and how that relationship would cause the depletion curve needed by the model to differ from that which would be constructed based on detailed measurements of water equivalent and the fraction of the areal actually covered by snow. The areal depletion curve for the model should indicate a lower areal extent of snow cover than would be determined by satellite or other aerial observations.



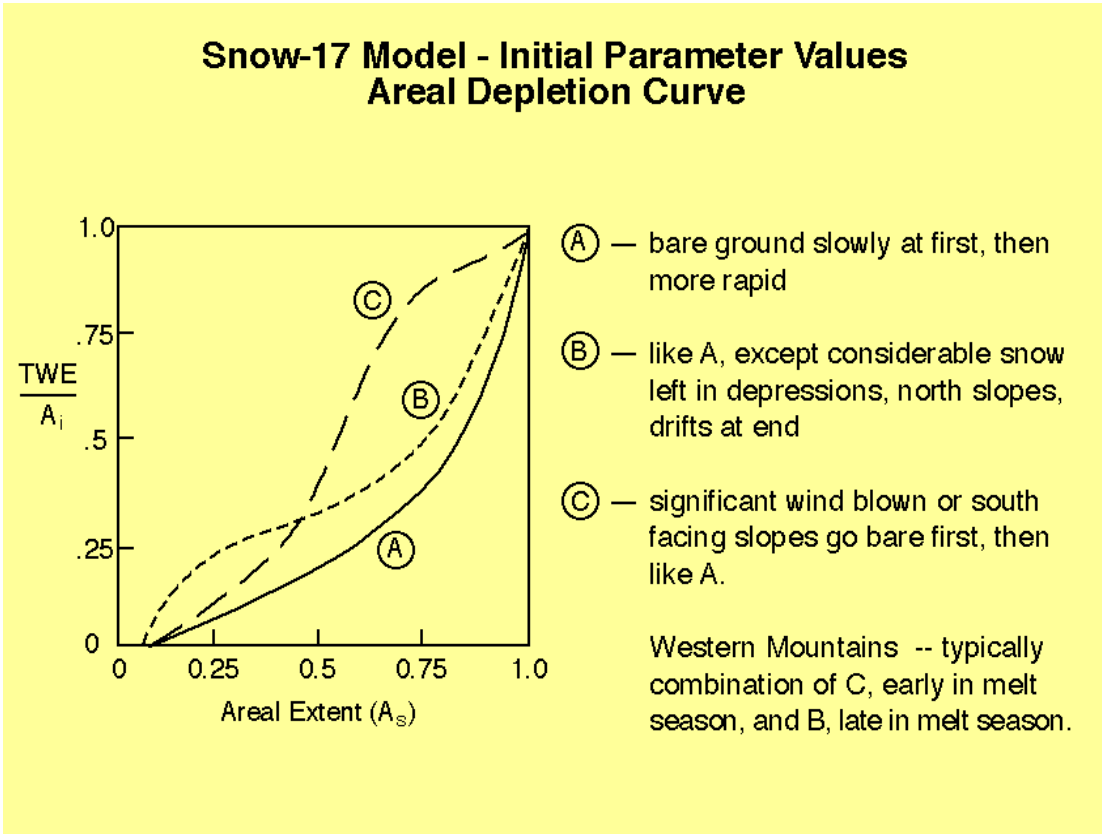


Figure 7-4-4. Typical shapes of areal depletion curves.

The factors that affect both the pattern of areal snow cover and the variation in melt rates as the snow cover depletes are primarily controlled by terrain, vegetation, and climatic conditions. Thus, the shape of the areal depletion curve for a given area can be estimated based on a knowledge of these factors. Figure 7-4-4 shows some typical shapes for areal depletion curves. The “A” curve represents an area with generally flat to hilly terrain, either uniform or randomly mixed vegetation cover, and not an excessive amount of redistribution of snow due to blowing and drifting. In this case bare ground shows up slowly at first and then more rapidly as the snow cover depletes. Curve “B” represents a flat to hilly area where

the terrain and redistribution pattern results in portions of the area having more snow, typically in sheltered depressions or drifts, than other parts. These parts of the area with disproportionally more snow are typically the last portions to go bare. Curve "C" represents an area that has windblown portions with little snow cover or significant south facing slopes that go bare fairly quickly after melt begins. Thus, the areal cover depletes rapidly at first and then the remaining portion of the area has a depletion pattern similar to the "A" curve. Mountainous regions, especially in the western states and Alaska, typically have a depletion curve that is a combination of the "B" and "C" curves. In the mountains there are frequently portions of the area with south facing slopes or shallow cover due to wind effects that go bare quickly after melt begins and also sheltered depressions and conifer forested, north facing slopes that still retain snow well after most of the area is bare.

SI – The SI parameter is the mean areal water equivalent above which the area essentially has 100% snow cover. The easiest way to determine the appropriate value for SI is to set the initial value greater than any average areal water equivalent that ever occurs during the period of record. Then during calibration the years with the greatest amount of snow are examined to determine if the areal cover needs to remain at 100% for some time period after the beginning of melt. By analyzing these years as described in Section 7-7 one can arrive at the value to assign to the SI parameter. Typically the initial SI is set to 999 mm or 9999 mm depending on the amount of water equivalent expected. By using all 9's for the initial value, it is easy to later tell whether a value for the SI parameter was assigned during calibration or whether bare ground begins to show as soon as melt starts every year, i.e. the area doesn't remain at 100% cover once melt starts and thus the value of SI is left as all 9's.

UADJ – The UADJ parameter represents the average wind function during rain-on-snow events. The wind function is involved in the latent and sensible heat, i.e. turbulent transfer, terms in the energy balance. The UADJ parameter will not affect the longwave radiation and heat from rain water terms in the rain-on-snow melt equation used in the model. Based on energy balance studies [*Anderson*, 1976] the wind function parameter can be computed as:

$$UADJ = 0.002 \cdot u_1 \qquad (7\text{-}4\text{-}1)$$

where: u_1 = 6 hr. wind travel in km at a 1 meter height above the snow surface.

Thus, if one can estimate the average wind speed over the entire area being modeled during significant rain-on-snow events at one meter above the snow surface, then an initial value for UADJ can be computed. Significant rain-on-snow events are those that occur when the air temperature is well above freezing. Clearly it is difficult to estimate the average wind speed at one meter for a forested, mountain watershed, but generally a rough guess is sufficient for an initial value. Also the UADJ parameter is not real sensitive for most regions. Typical values for UADJ range from about 0.05 (2.5 mi/hr wind speed) to 0.20 (10 mi/hr wind speed). Forested areas will tend to have lower wind speeds than mostly open areas.

Minor Snow Model Parameters

The remaining parameters in the snow model are assigned based on climatic information and are typically not changed during the calibration process. These parameters generally do not have a very significant effect on the simulation results as long as the value is in the right ballpark.

TIPM – The TIPM parameter is used to compute an antecedent temperature index that is intended to represent the temperature inside the snow cover but near the surface. The gradient defined by the antecedent index and the air temperature (used to estimate the temperature at the snow surface) determines the direction of heat flow during periods when melt is not occurring. The lower the value of TIPM the more weight is assigned to the temperatures from previous time intervals and the antecedent index then represents the temperature of the snow further below the surface. For areas with generally shallow snow cover the antecedent index should represent a temperature fairly close to the surface in order to best estimate the direction of heat flow. For regions that typically have a deep pack, the antecedent index should represent the temperature further below the snow surface. It is recommended that a value of TIPM = 0.05 be used for areas that experience a deep snow cover (greater than 3 feet maximum depth during most years). A value of TIPM = 0.20 is appropriate for areas with shallow (generally less than 1 foot depth) or intermittent snow cover. Intermediate values should be used for other areas.

NMF – The negative melt factor determines the amount of energy exchange that occurs when melt is not taking place at the snow surface. The NMF parameter is the maximum value of the negative melt factor. The seasonal variation in the negative melt factor is the same as for the non-rain melt factor. A seasonal variation is needed since density is the primary factor that affects the thermal conductivity of the snow and the snow density generally varies in a seasonal manner. The density is lowest during periods of accumulation and increases as the snow ages during melt periods. Simplified heat transfer calculations indicate that a reasonable value for NMF is around 0.15 mm/°C/6 hr. This value is generally independent of the typical amount of snow that occurs in an area. It is based on a maximum snow density of 0.3 for shallow snow cover and 0.5 for a deep pack. If an area generally has a maximum density less than these values, NMF should be decreased and if the maximum density is typically greater than these values, NMF would be increased. A maximum reasonable range for the NMF parameter is from 0.05 to 0.30.

MBASE – The MBASE parameter is the base temperature used to determine the temperature gradient for non-rain melt computations. When modeling an area with a variety of vegetation cover, slopes, and aspects, the value of MBASE is almost always 0°C. This includes most watersheds where snowmelt runoff is significant. Generally if a non 0°C value of MBASE is needed for most watersheds in order to properly simulate snowmelt, there is a good chance that there is a bias or error in the MAT computations. Thus, before using a non 0°C value for MBASE, in most cases, one should carefully check and reevaluate the MAT calculations. There is a chance that a MBASE greater than 0°C is needed in high elevation, open areas

with generally clear skies and relatively low humidity during the melt season. Under such conditions negative longwave radiation and latent heat exchange can offset solar radiation and sensible heat exchange at temperatures slightly above 0°C during the day and refreezing of the water in the upper snow layers occurs at night even though the temperature is above freezing. MBASE values greater than 0°C are frequently needed when applying the snow model to a open, high elevation point location such as a snow course site. Non 0°C values are also frequently needed when modeling other point locations due to site specific factors that control the relationship between measured air temperature and melt amounts.

PLWHC – The PLWHC parameter controls the maximum amount of liquid water that can be retained within the snow cover expressed as a decimal fraction of the amount of ice in the snow cover. The PLWHC parameter should be selected based on ripe snow cover conditions, i.e. the snow is well aged, isothermal at 0°C, and has its liquid water capacity full. Fresh (i.e. newly fallen, low density snow) snow can hold more liquid water than ripe snow, however, the existence of the liquid water will cause metamorphism to begin and within hours, if sufficient water is applied, the snow will become ripe. Most studies indicate that the maximum liquid water holding capacity for ripe snow is in the order of 2 to 5 percent. Thus, the PLWHC parameter should be assigned a value in the range of 0.02 to 0.05 with generally the lowest values being used for areas with very deep snow covers. In addition to snow retaining liquid water throughout the pack, a slush layer generally builds up at the snow-soil interface. For a deep snow cover the amount of water held in this slush layer is inconsequential, but for a shallow snow cover it can be significant. To account for the slush layer, it has been found that in areas with shallow snow covers, especially in the plains and open agricultural areas of the midwest, that a higher value is needed for the PLWHC parameter in order to properly delay the onset of melt water entering the soil. In such areas values for PLWHC in the range of 0.1 to 0.3 are quite common.

DAYGM – The DAYGM parameter controls the amount of melt per day that occurs at the snow-soil interface. This is a constant amount of melt that takes place whenever there is a snow cover. The following values are recommended for the DAYGM parameter:

DAYGM = 0.0 for areas with generally frozen soils under the snow, and
DAYGM = 0.3 for areas with intermittent snow cover or with fairly temperate climates, such as the Sierra Nevada mountains in California, during the winter.

Other areas would use DAYGM values somewhere in between these limits.

Section 7-5

Initial Parameter Values for the Sacramento Model

Introduction

In most cases initial parameter values for the Sacramento Model are obtained from a nearby previously calibrated watershed. This can be a watershed within the river basin being worked on or one in an adjacent river basin that has previously been calibrated. However, when calibrating the initial headwater area in a river basin where no nearby watersheds, or at least none with physiographic features at all similar to the current area, have been previously calibrated, initial parameter estimates should be derived. Procedures are available for obtaining initial parameter estimates from an analysis of the observed daily flow hydrograph and from soils information.

Besides providing recommendations for deriving initial parameter values, this section also contains information on the physical basis for some of the parameters. This information should help in better understanding of how the model works and provide a basis for possibly altering parameters from one area to another based on physical characteristics. The portions of the section that describe how to determine parameter values from a hydrograph analysis should also help in understanding what to look for in order to isolate the effects of many of the parameters. Section 7-8 focuses on how to isolate the effect of each parameter, but the material in this section should provide some added emphasis to understanding this critical aspect of interactive calibration of a conceptual model. This section doesn't attempt to describe the structure and algorithms of the Sacramento model. For that information the reader is referred to Part II.3-SAC-SMA of the NWSRFS Users Manual.

Before describing the techniques for deriving initial parameter estimates, we first need to talk about how the various runoff components of the Sacramento model are going to be used to reproduce the hydrograph for the watershed. It is very important to take some time prior to starting the calibration to examine the various runoff time delay segments that are represented in the hydrograph and decide which runoff component is to be assigned to each segment. This effort should provide the best chance for obtaining good results in the least amount of time. If the runoff components are not assigned properly at the beginning, considerable time can be wasted modifying the parameter values once one realizes that the runoff components are not being used correctly.

Assigning Runoff Components

The Sacramento model contains 4 basic runoff components with various time delays that can be used to represent the various portions of the hydrograph. These are primary baseflow, supplemental baseflow, interflow, and surface runoff. The model is designed so that the longest time delay, usually in terms of months or years, is assigned to primary baseflow. Supplemental baseflow generally has a time delay in terms of weeks or months and interflow typically has a

time delay in terms of days. Surface runoff has no time delay in the Sacramento model, i.e. it becomes inflow to the channel during the same time interval as the rain or melt that produced it. The Sacramento model can also generate constant and variable impervious area runoff components. Both of these respond immediately just like surface runoff. Variable impervious runoff can be produced from low intensity rain or melt when the watershed is quite saturated. Constant impervious runoff can occur whenever there is rain or melt no matter what the soil moisture conditions. An analysis of the hydrograph may determine if these two additional components will be needed, but primarily at this stage of the calibration we are trying to decide which portions of the hydrograph will be assigned to each of the 4 main runoff components.

The first step in this process is to identify when primary baseflow is the only or at least the dominant source of runoff. Primary baseflow typically sustains flow in the channel long after any events that produced storm runoff have occurred and after perched, or supplemental, aquifers have been drained. Primary baseflow is used to represent the flow segment with the longest time delay. It is critical to properly identify what portion of the hydrograph represents primary baseflow because the time delays for the other runoff components are all going to be based on this determination. The segment of the hydrograph with the next slowest time delay will be assigned to be modeled with supplemental baseflow and then the next slowest with interflow. Surface runoff can only be used to produce immediate storm runoff from high intensity events. Periods of surface runoff can usually be identified by comparing the immediate amount of storm runoff to the amount of rain plus melt (the period for determining the amount of immediate storm runoff is dependent on the response time of the channel system as represented by the unit hydrograph -- initial snow model parameters are used to get an idea as to the amount of melt). If the amount of immediate storm runoff is around 50% or more of the rain plus melt, then it is likely that surface runoff needs to be generated for these events. Identifying what runoff component will be used to represent each portion of the hydrograph prior to beginning the calibration should insure that all the components are used properly and that the various time delays that occur can each be modeled.

In some regions it is quite easy to identify which runoff component will be used to model each portion of the hydrograph. If the time delays for each runoff component are close to typical values, there are sufficiently long dry periods after major events to allow for a clear identification of interflow and supplemental baseflow recessions, and there are also some even longer dry periods when primary baseflow becomes the only component with no distortions of the flow during these periods, the runoff components can be fairly easily identified. However, in many regions, probably the majority, there are complications that make it more difficult to clearly identify what portion of the hydrograph represents each component of runoff, especially to properly identify primary baseflow. Primary baseflow becomes more difficult to identify in very wet regions, when frequent small rains occur during low flow periods, when the supplemental baseflow recession is very slow, and when riparian vegetation evaporation draws down the flow during dry periods.

Figure 7-5-1 illustrates the case of assigning runoff components in a very wet region. In such a

region rainfall or snowmelt occurs frequently and soil moisture conditions remain quite wet. Dry periods that exist for a long enough time so that primary baseflow predominates only occur on

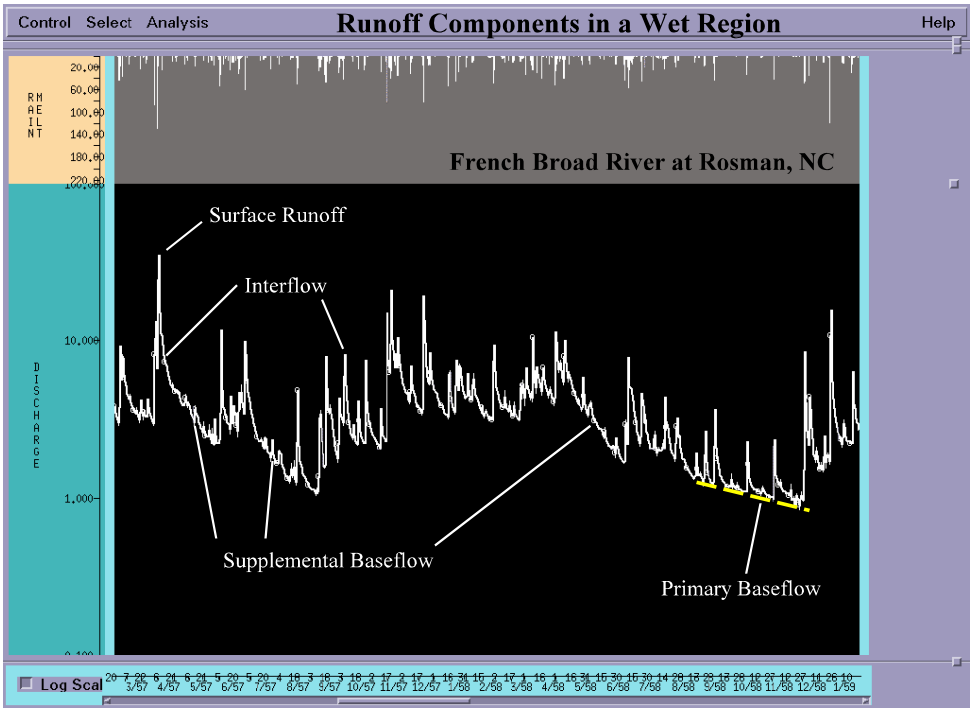


Figure 7-5-1. Assigning runoff components in a wet region.

the average about once every 5 to 10 years. Even then there are frequent small rain events that occur during these periods. Thus one must look very carefully at a long period of record to properly determine the appropriate primary baseflow recession. Once primary baseflow is identified, generally the times when the other components can be isolated fall into place. In regions with high percolation rates such as the one in this figure, surface runoff seldom occurs, if at all. If surface runoff does occur, it is associated with only the largest flood events. Interflow produces most of the rises from rain or melt periods, as well as the early part of the recession for the flood events that generate surface runoff.

Figure 7-5-2 illustrates the case when there are frequent small rain events during the low flow periods when primary baseflow dominates. In such regions, periods when primary baseflow dominates occur during the majority of the years, but the recession doesn't show up as a nice straight line on a semi-log plot due to the frequency of constant impervious runoff produced by small rain storms. One must again determine the primary recession rate by the general slope of the semi-log plot during periods when no recharge occurs, i.e. there is no other runoff than that from constant impervious areas. As with most cases, once primary baseflow is identified, the other runoff components fall into place.

Figure 7-5-3 illustrates the case where the supplemental baseflow recession is very slow. In

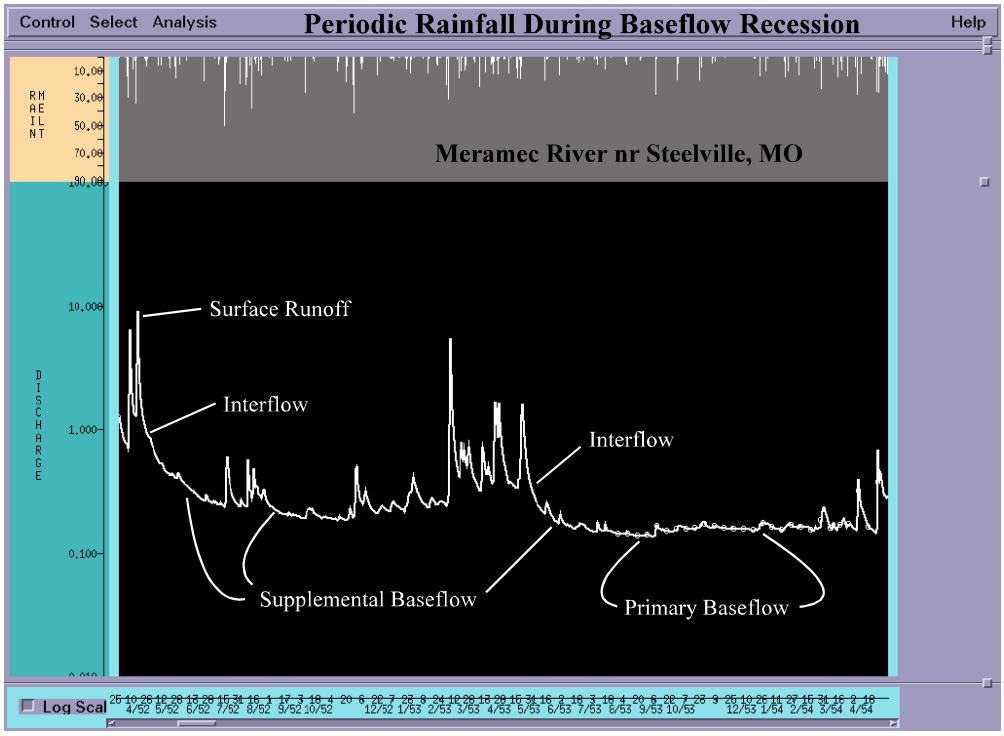


Figure 7-5-2. Runoff components in regions with frequent small events during baseflow periods.

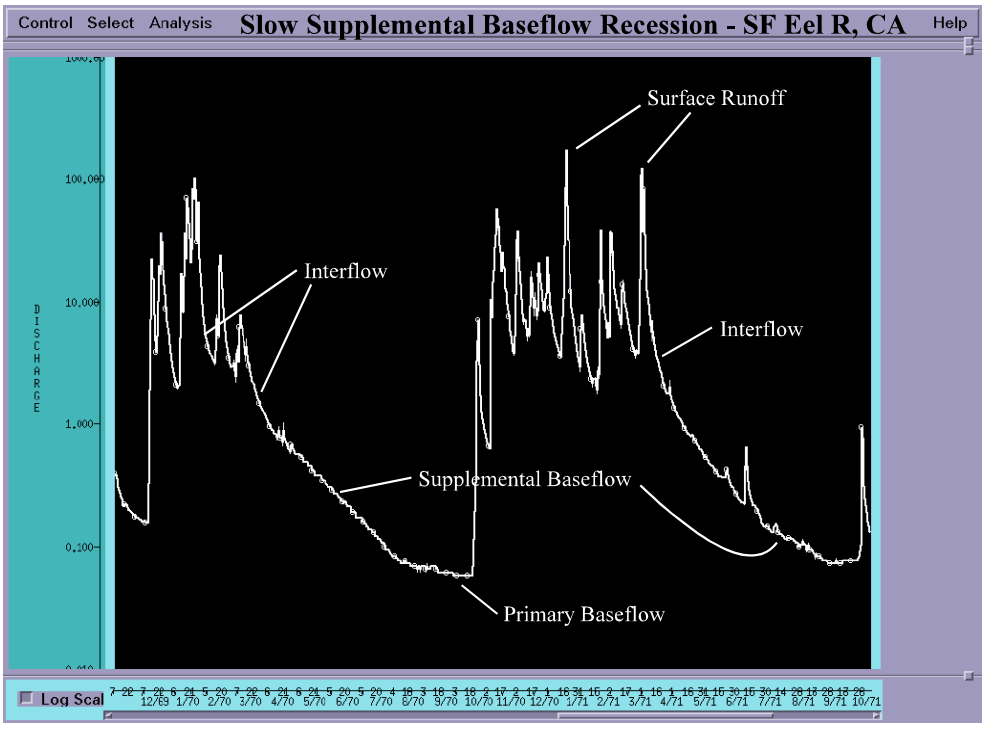


Figure 7-5-3.

Watershed with a very slow supplemental baseflow recession.

these situations it may take many months for the supplemental aquifers to drain, thus requiring quite a long dry period before there are situations when primary baseflow predominates. Periods with only primary baseflow may only occur once every 10 years or more, especially if this is in a region with year around rainfall. This is probably the most difficult case for determining how to allocate the runoff components. There is a tendency to not look carefully at a sufficiently long record to find the few cases when primary baseflow dominates. Instead the periods labeled as supplemental baseflow are modeled as primary causing simulated baseflow to drop off too rapidly when primary baseflow actually predominates. When this is done, it is likely that supplemental baseflow and interflow will end up with similar withdrawal rates.

Figure 7-5-4 illustrates the case when there is a large amount of evaporation from riparian vegetation during summer and early fall dry periods. These are also the periods when primary baseflow predominates. The evaporation from the riparian vegetation draws down the water table near the stream causing a rapid decrease in baseflow. In such regions if dry conditions persist into the late fall and winter, the baseflow level will increase with no recharge occurring due to the reduction in riparian vegetation effects caused by a decrease in evaporation and the vegetation becoming dormant. It is these late fall and winter periods that must be used to determine the proper primary recession rate. When modeling such a watershed, the simulated baseflow should follow the dashed line shown in the figure when the RIVA parameter is set to zero, i.e. evaporation from riparian vegetation is not being included. It is important to be able to recognize when riparian vegetation affects baseflow and ignore these periods when deriving baseflow recession rates.

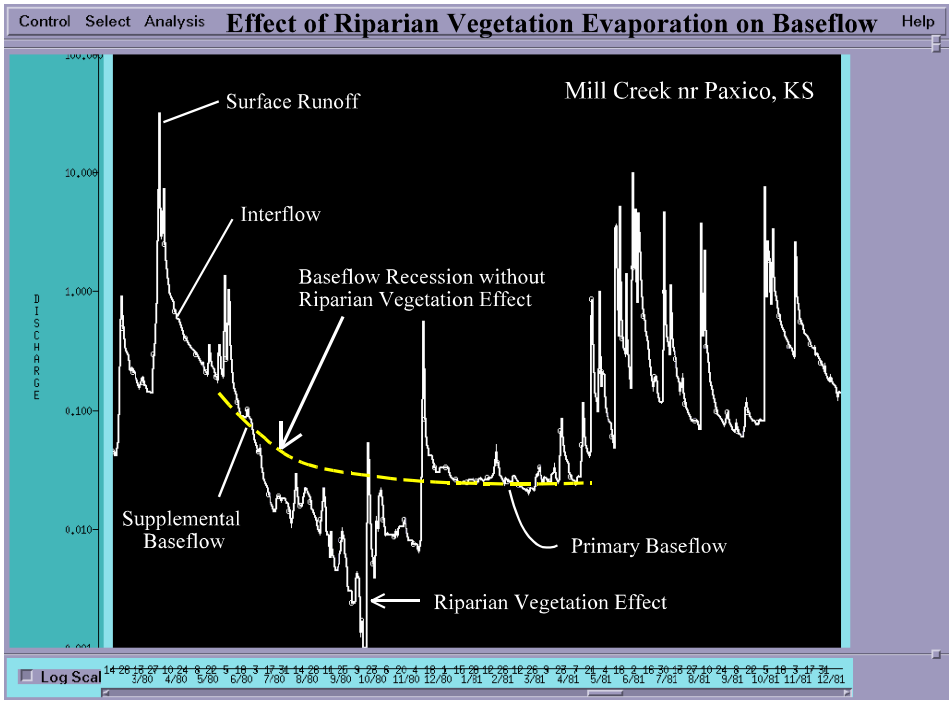


Figure 7-5-4. Assigning runoff components in a region with riparian vegetation.

It is very important to understand these complications when determining how the runoff components of the model are going to be used to simulate the various time delay segments of the hydrograph. It is also critical to take the time to do a careful analysis of the observed hydrograph over a long period in order to understand how the runoff components of the model should be used. Doing this should simplify the calibration and produce the best possible results in the least amount of time.

Deriving Initial Parameter Values

The Sacramento model was designed so that each model parameter serves a particular function and the effect of the parameter controls the hydrograph response under a specific set of circumstances. In most cases this effect can be isolated when the proper situation occurs and thus, the appropriate value of the parameter can be estimated by analyzing the observed hydrograph over a period of many years. Thus, hydrograph analysis is the most direct method of deriving initial estimates for most parameters of the model. For some parameters a numerical derivation is not possible, but an analysis of the hydrograph contains information that suggests the general range of values. In other cases the specific set of circumstances needed to derive the parameter value doesn't occur during the period of available record. In some of these cases a lower limit for the parameter can be computed based on less than ideal circumstances and in other cases the initial value must be assigned using general guidelines or another method.

Procedures have been developed to derive *a priori* estimates of the model parameters based on physiographic information. The procedures proposed up to this point in time have relied solely on soils information to estimate the values for most of the Sacramento model parameters. The most recent of these procedures [*Koren*, 2000] is included in CAP. This procedure utilizes STATSGO soil texture data in 11 soil layers to derive 11 of the 16 Sacramento model parameters. In reality many of the Sacramento model parameters are related to factors in addition to soils information, including vegetation, geology, terrain, and man-made features such as farm ponds and agricultural drain tiles. This means that methods using only soils data typically have to rely on assumptions that may not be valid for all regions or even for all watersheds in a given region.

One of the possible uses of *a priori* parameter estimates is as a source of initial values for model calibration. More importantly, if such a procedure can be shown to produce good results in a given region, it could be used to determine how parameter values should vary from one watershed to another, to obtain parameter estimates for ungaged areas, and to specify how

parameter values vary within the boundaries of a watershed. Such estimates would be helpful when applying the models to a large region, when trying to apply models to small ungaged watersheds for flash flood forecasting, and for distributed applications of a conceptual model. An *a priori* method could be used to directly obtain parameter values or it could be used to determine differences in parameters from one watershed to another or within the drainage boundaries. In the second case a calibration would be used to determine the appropriate parameter values for a watershed and then the parameter variations within the drainage or with adjacent watersheds would be obtained by adjusting the calibrated values by the relationships between *a priori* estimates for these areas.

Model calibration involves determining appropriate model parameter values based on data sets that contain noise, both in the input and output data. When the amount of noise becomes excessive, it becomes very difficult to determine the values of many of the parameters. As discussed previously in this manual, a sufficiently long period of record is generally needed to filter out the effect of the noise during individual events. Since the amount of noise will vary from watershed to watershed, primarily based on the number and location of precipitation gages, differences in calibrated parameter values from one watershed to another may not be totally reflective of physical differences between the areas, but could be influenced by differences in the amount of noise in the data. This is why the procedure recommended in this manual stresses only changing those parameters that can be clearly justified when moving from the initial headwater area to other watersheds in the river basin. If an *a priori* parameter estimation procedure can be shown to produce realistic values for a given region, it could be a useful tool to achieve spatial consistency in model parameters. Given the noise in the data, it is important to have physically realistic variations in parameter values across the river basin even if it means less accurate overall fit statistics.

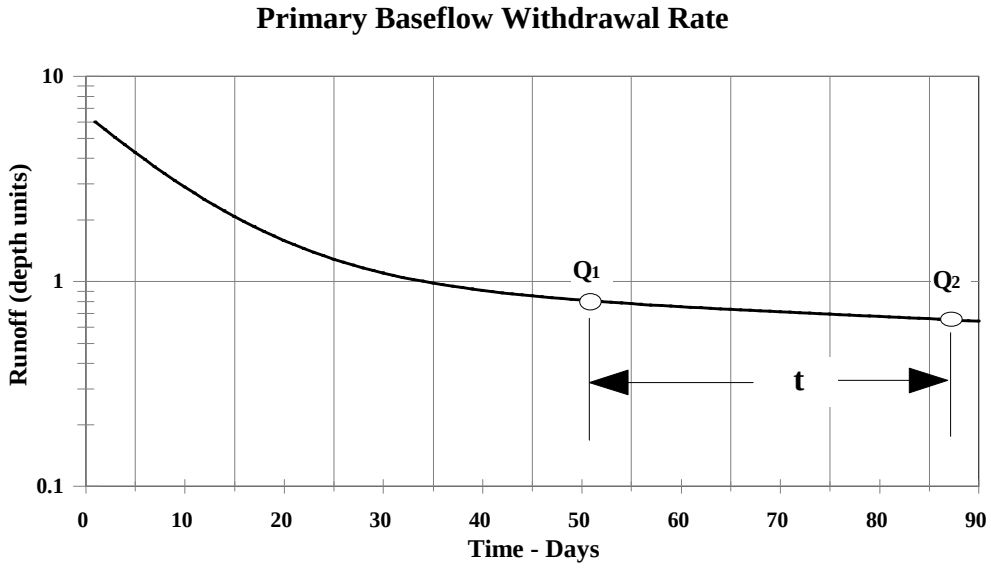
An evaluation of any *a priori* parameter estimation procedure for a conceptual model like the Sacramento model should examine how each derived parameter behaves during times when the effect of the parameter can be isolated. Comparisons of overall 'goodness of fit' statistics for simulations based on *a priori* parameter estimates to those based on calibrated values don't reveal whether the procedure results in physically realistic parameter values. It is suggested that before an *a priori* parameter estimation procedure is used in a given region that comparisons be made between simulations using the *a priori* estimates and observed hydrographs for gaged headwater areas. These comparisons should concentrate on those portions of the hydrograph where each of the parameters can be best isolated to determine if the *a priori* estimation procedure is deriving realistic parameter values. Comparisons should ideally be made for several watersheds in the region, especially cases where the spatial analysis conducted in step 2 (see Chapter 4) suggests that some of the model parameters are quite different. In order to take full advantage of an *a priori* parameter estimation procedure, the method should be able to detect these differences. Section 7-9 illustrates such an evaluation for the Koren *a priori* parameter estimation procedure for several regions scattered around the country.

On the following pages recommendations are given for deriving the initial value of each of the

Sacramento model parameters for use in model calibration. The recommendations are primarily based on deriving the values from a hydrograph analysis, but also utilize estimates from *a priori* parameter estimation procedures in certain situations. Also for a few of the parameters, this author's opinion as to what the parameter physically represents is included. As mentioned in the introduction to this section, all this information should be helpful in understanding the structure of the model and how to isolate the effects of the various parameters even if it is seldom used to derive initial values.

LZPK

The LZPK parameter reflects the slowest baseflow recession rate which occurs after there has been no groundwater recharge for a period that is typically in terms of months. The only method for obtaining a reasonable value for LZPK is from a hydrograph analysis. The identification of periods when primary baseflow dominates the hydrograph is discussed in the section "Assigning Runoff Components" in this section. It is critical to properly identify when primary baseflow is dominate before deriving the value of LZPK. Once the proper periods are found, the values needed to compute LZPK are obtained as shown in Figure 7-5-5 from the straight line portion of a semi-log plot. The straight line segment will have to be estimated when complications such as riparian evaporation, small power plants, and diversions cause fluctuations in the flow.



Figu
re 7-5-5. Hydrograph values needed for deriving LZPK.

Using time, t, in days, the primary recession rate, K_p, can be computed as:

$$K_p = \left(\frac{Q_2}{Q_1} \right)^{1/t} \qquad (7\text{-}5\text{-}1)$$

and the withdrawal rate (1.0 minus the recession rate), i.e. the LZPK parameter value, is:

$$LZPK = 1.0 - K_p \qquad (7\text{-}5\text{-}2)$$

During winter periods when there is snow on the ground, the computed withdrawal rate could be less than the value of LZPK due to small amounts of melt at the snow-soil interface which can be providing a steady slow recharge to the lower zone storages.

LZSK

LZSK is the withdrawal rate from the supplemental baseflow aquifers. As with LZPK, the only method for obtaining reasonable values for LZSK is by analyzing the observed daily flow hydrograph. The effect of this parameter can be isolated by finding relatively dry periods when no recharge is occurring beginning several days after intervals of significant runoff and recharge. During such periods both supplemental and primary baseflow are being generated. Thus, in order to determine the supplemental recession, the primary baseflow component must first be removed. When primary baseflow is subtracted from the total flow, periods of supplemental baseflow should plot as a straight line on a semi-log plot. This is illustrated in Figure 7-5-6.

Supplemental Baseflow Withdrawal Rate

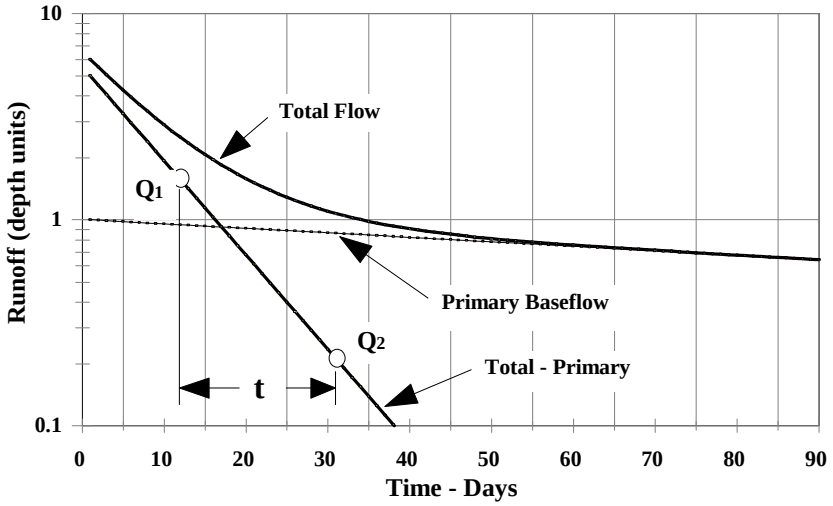


Figure 7-5-6.

Hydrograph values needed for deriving LZSK.

Two points can then be picked off this straight line and by using time, t, in days, the supplemental recession rate can be computed as:

$$K_s = \left(\frac{Q_2}{Q_1} \right)^{1/t} \qquad (7\text{-}5\text{-}3)$$

and the withdrawal rate, i.e. the LZSK parameter value is:

$$LZSK = 1.0 - K_s \qquad (7\text{-}5\text{-}4)$$

As one gains experience, the two runoff values needed to compute the supplemental recession can be obtained by subtracting primary baseflow from total flow for two days during the supplemental recession without having to generate a plot of total flow minus primary baseflow.

LZFPM

There are two ways of computing an estimate of LZFPM from a hydrograph analysis. The first is to find a nice recession period following several months of significant runoff and recharge during which the primary baseflow aquifer has a chance to accumulate a substantial amount of storage. The largest primary contents will typically occur during the very wettest years. If the recession period is of sufficient length to reach the point where only primary baseflow remains, then the primary recession can be extrapolated backwards to find the amount of primary baseflow at the end of the recharge period, Q_x, as shown in Figure 7-5-7.

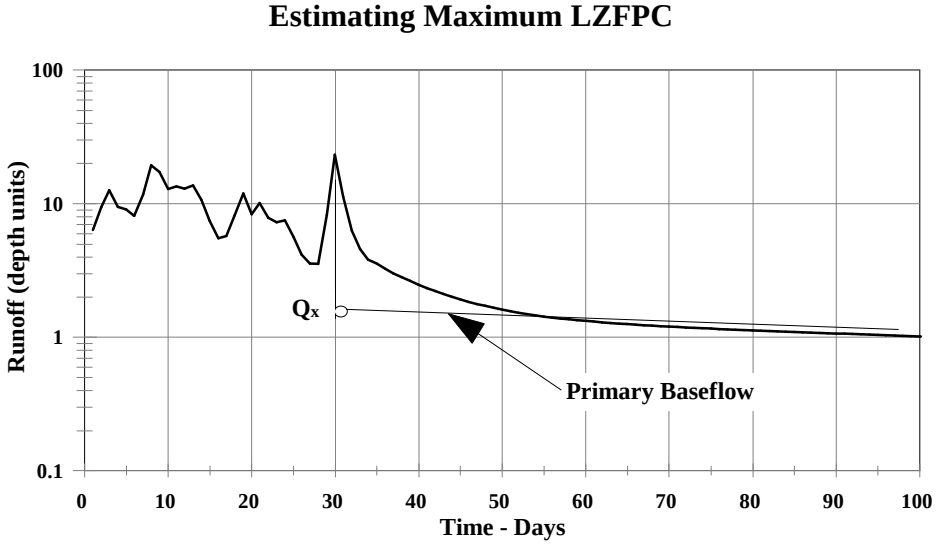


Figure 7-5-7. Back extrapolation to estimate maximum LZFPC.

The initial estimate of LZFPM can then be computed from:

$$\text{LZFPM} = \left(Q_x + \left(\varepsilon \cdot Q_x\right)\right) / \text{LZPK} \qquad (7\text{-}5\text{-}4)$$

where ϵ is a decimal fraction typically in the range of 0.1 to 0.25. The ϵ term is needed since the contents of the lower zone primary storage never actually become completely full. Generally the greater the withdrawal rate, LZPK, the larger the value of ϵ since it is harder to fill a storage the faster it drains.

If a backwards extrapolation is not possible during any of the wet years to get an estimate of Q_x, then the second way to estimate Q_x and then LZFPM is to attempt to sketch on a semi-log plot the primary baseflow component during some of the wettest years. This can be done if periods when primary baseflow is dominant exist prior to and after a prolong period of recharge. With these tie in points and having an estimate of LZPK, one can sketch the primary baseflow contribution and come up with a value of Q_x to use in Equation 7-5-4.

LZFSM

An estimate of LZFSM can also be derived by back extrapolation. This can be done when the baseflow recession can clearly be determined after a very large storm event. If the supplemental withdrawal rate, LZSK, is relatively slow, then a recession period after a series of storms is likely needed to have as much water as possible in supplemental storage. Figure 7-5-8 illustrates how to estimate the maximum amount of supplemental runoff, Q_x, by extrapolating the supplemental recession back to when the contents of this storage are at their fullest. It is best to subtract primary baseflow from total flow in order to clearly find the supplemental recession, though with experience the plotting of the total minus primary line is not needed to estimate a reasonable value for Q_x

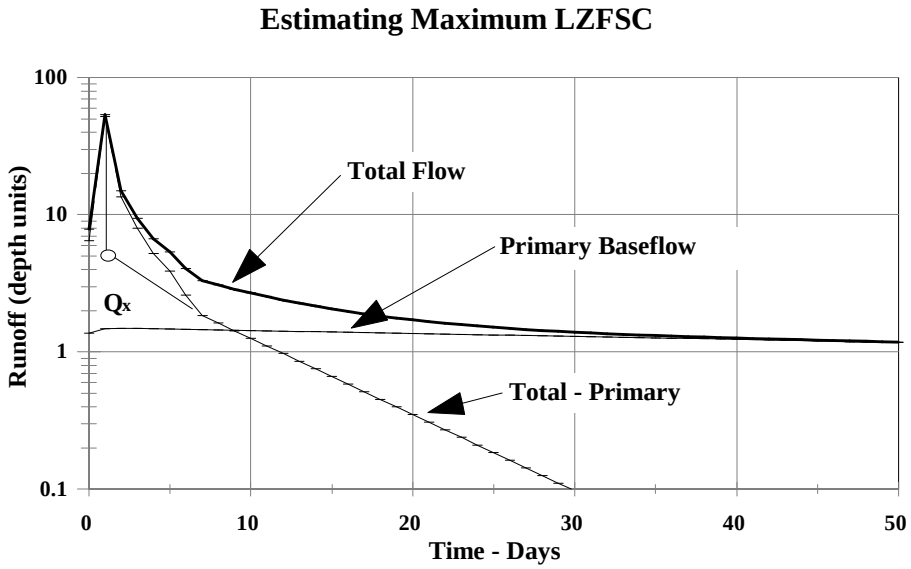


Figure 7-5-8.
Back extrapolation to estimate maximum LZFSC.

Once Q_x is determined an estimate of LZFSM can be computed from:

$$\text{LZFSM} = \left(Q_x + \left(\varepsilon \cdot Q_x\right)\right) / \text{LZSK} \qquad (7\text{-}5\text{-}5)$$

where ϵ in this case typically varies from about 0.5 to 1.0. Larger values of ϵ are needed for supplemental baseflow than for primary since the supplemental storage drains faster, thus the supplemental contents seldom get close to capacity. Also the value of ϵ should be greater for larger values of LZSK.

PCTIM

The PCTIM parameter represents the part of the area that always produces some runoff no matter what soil moisture conditions exist. The value is generally not the same as the portion of the area covered by surfaces such as pavement, roofs, and rock outcrops since runoff from many of these surfaces encounter areas of soil before reaching the stream channel. PCTIM represents impervious areas that are directly connected to the channel system.

An good estimate of the amount of constant impervious area can normally be derived from a hydrograph analysis. The conditions needed to derive a value of PCTIM are a week or two of dry weather during late spring or summer that produce a significant upper zone tension water deficit, followed by a moderate rain event (typically 0.25 to 0.75 inches) which is not sufficient to fill this deficit (no recharge occurs). The amount of runoff produced during such events is then computed as shown in Figure 7-5-9.

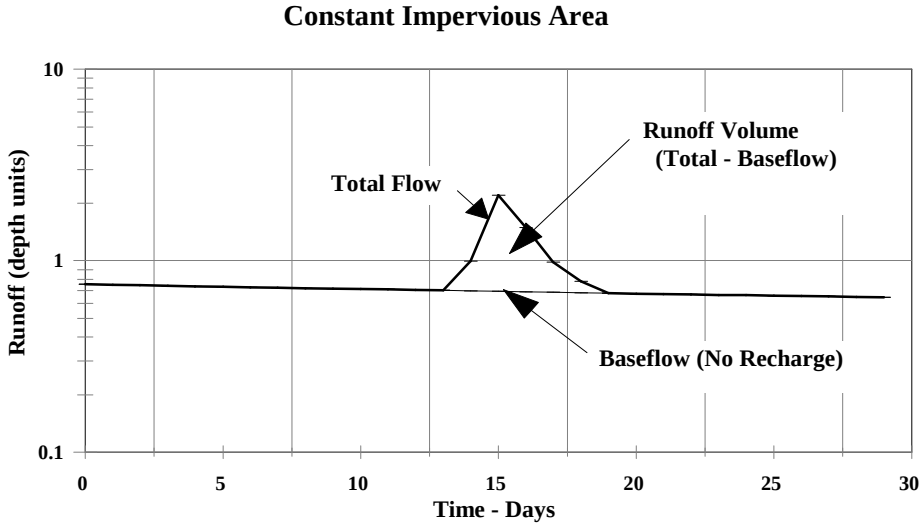


Figure
7-5-9. Runoff volume from the constant impervious area.

Then an estimate of the PCTIM parameter is computed by dividing the runoff volume, R_{imp}, by the amount of precipitation, P, as shown in Equation 7-5-6.

$$PCTIM = R_{imp} / P \qquad (7\text{-}5\text{-}6)$$

It is best to compute PCTIM for a number of events since rainfall under such conditions can be more uncertain since it is often from convective events with considerable spatial variability. Then take the average of these events after throwing out any with a significantly greater runoff percentage than the majority of the cases. The upper zone tension water may have filled during such events and thus, they could contain some interflow, or even variable impervious runoff, in addition to runoff from areas that always act as impervious.

UZTWM

The UZTWM parameter indicates the amount of rain that must fall after a long dry period before any runoff, other than that from constant impervious areas, is produced. The upper zone tension water comprises water held in the pervious surface soil, plus interception (by vegetation and forest litter) and depression storage. In agricultural regions with many farm ponds, the effect of these ponds is typically implicitly absorbed by the UZTWM parameter.

If the right conditions occur, a good estimate of UZTWM can be computed from an analysis of the hydrograph. The conditions needed are several weeks or more of quite dry weather in the late spring or summer, followed by a significant rain event that generates a small amount of recharge and more runoff than specified by PCTIM to indicate that the upper zone tension water deficit has been filled. Such an event is shown in Figure 7-5-10.

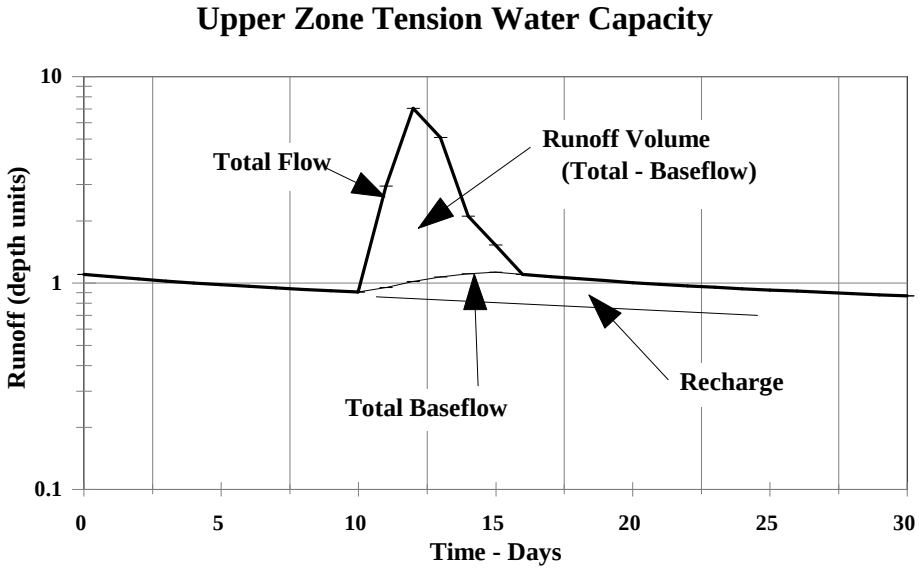


Figure 7-
5-10. Runoff from event that just fills UZTW storage.

The amount of rain that falls during such an event can be used as an estimate of UZTWM. Rainfall amounts that produce no additional runoff after a summer dry period can also be used as a lower limit for UZTWM. A number of events should be examined.

In regions where such situations don't occur, typically wet regions with frequent summer precipitation, it would probably be best to use the soil based value of UZTWM as an initial value. This should produce spatial consistency in UZTWM values over such a region. It should be verified that the soil based value is greater than any rainfall amounts that produce little or no recharge or additional runoff after the limited dry periods that may occur in these regions.

LZTWM

The LZTWM parameter indicates the maximum moisture deficit that can occur in the lower soil layers. Tension water is only removed from the lower zone by evapotranspiration via the vegetation in the watershed. Thus, this parameter is primarily a function of the depth of the root zone and not of the depth of the soil layer, though in shallow soils, the root zone may be controlled by the depth to bedrock. This was clear when modeling a watershed that had been transformed from a rural, forested landscape to a mostly suburban area with the primary vegetation being grass. The main parameter change needed to account for the effects of this transformation was to significantly reduce LZTWM. The grass cover produced runoff much earlier in the fall after a dry summer than when the watershed was forested.

The LZTWM parameter can be derived from a hydrograph analysis when the right conditions occur. The conditions needed are very dry conditions from late spring to late fall followed by a 2 to 3 week period with sufficient rain to fill the soil moisture deficit that has been generated. The water balance equation can then be used on the period from just after UZTW fills to just after LZTW fills (can be detected by a large increase in recharge) to compute the lower zone deficit that existed prior to the rain. Such a period is illustrated in Figure 7-5-11.

Estimating LZ Tension Water Deficit

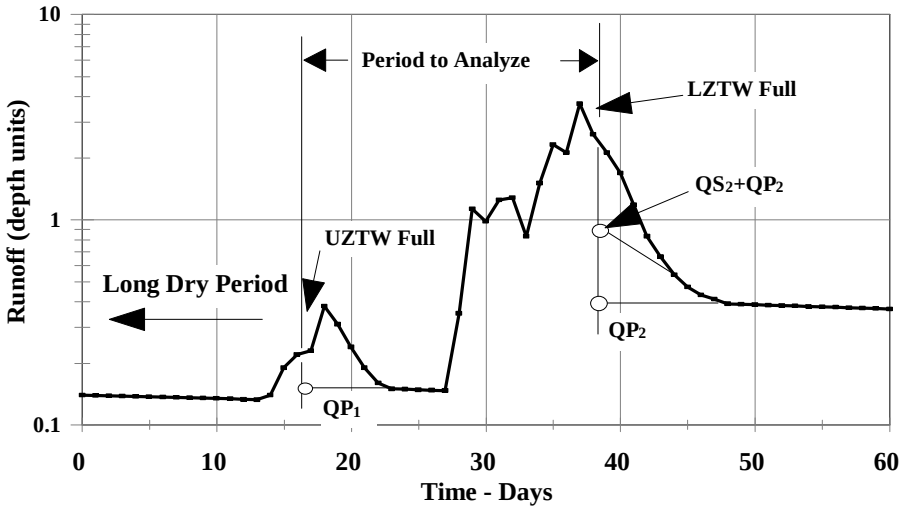


Figure 7-5-11. Period for computing LZTW deficit via water balance.

By assuming UZTW is full and UZFW is empty at the beginning and end of the period, that LZFSC=0.0 at the start, and that deep recharge equals zero, the LZTW deficit is computed as:

$$\Delta LZTWC = P - R - ET - QS_2/LZSK - (QP_2 - QP_1)/LZPK \qquad (7\text{-}5\text{-}7)$$

LZTWM (Continued)

where: P = precipitation, R = runoff, and ET = evapotranspiration. The ET for each day during the period should be close to the ET-Demand rate since UZTW should remain full or nearly full due to the periodic rainfall that generally occurs and the fact that evaporation rates are low at this time of year. The initial estimate of the LZTWM parameter should be somewhat greater than the maximum deficit that occurs after such a long dry period. If the LZTW deficit was ever equal to LZTWM it would indicate that the wilting point was reached throughout the watershed which is an unlikely situation.

In regions where a sufficiently long dry period never occurs and thus the LZTW deficit never approaches its maximum, the initial value of LZTWM should probably be obtained from the soil based derivation. It would be an improvement if this derivation also took vegetation type and coverage into account, however, the use of the current soil based value would insure some spatial consistency in wetter regions. When large lower zone deficits never occur, the effect of this parameter is difficult to isolate during calibration and fairly large variations in its value can be compensated for in many cases by reasonable changes to other parameters and the ET-Demand curve.

UZK

The withdrawal rate from upper zone free water, i.e. the UZK parameter, cannot be derived like the lower zone withdrawal rates, LZPK and LZSK, because water percolates from this zone in addition to draining out as interflow. It is possible in many cases when doing a recession analysis for a major storm event to compute the overall recession rate for the upper zone free water. The steps in a recession analysis are described in Section 7-6 as part of the procedure for deriving a unit hydrograph for use with the Sacramento Model. This overall recession rate is based on both the interflow withdrawal rate and the percolation rate. The percolation rate of course varies with soil moisture conditions. If the recession analysis is being done for a storm during a time when soil moisture conditions are close to saturation and the watershed has a very low percolation rate under these conditions, the upper zone free water recession rate derived from the analysis should be only slightly greater than the interflow withdrawal rate and can be used to estimate the initial value of UZK. In a case with such low percolation rates, the watershed should produce very little baseflow in general and almost all the storm runoff from the event should be surface runoff. The combination of all these conditions is quite rare, thus normally it is not possible to derive an good initial value for UZK from a hydrograph analysis.

When UZK cannot be derived from a hydrograph analysis, it is recommended to start with a nominal value of UZK = 0.3. The soil based derivation of UZK assumes that the interflow withdrawal rate is related to soil texture, with the more clay particles, the slower the withdrawal rate. Texture is indexed by the ratio of field capacity to porosity with the more clay, the higher the ratio. This seems logical, however, for the watersheds used by the author to test the soil based parameter derivations, the empirical equation used to compute UZK from this ratio produced unrealistically high UZK values in most cases. For this reason it is suggested that the nominal value of 0.3 be used as an initial estimate of UZK.

UZFWM

When a reasonable initial estimate of the UZK parameter can be derived from a hydrograph storm recession analysis as mentioned on the previous page, then an estimate of the capacity of the UZFW storage can also be computed. The maximum amount of interflow runoff can be determined by extrapolating the total minus baseflow line segment back to the time when the hydrograph peak occurred to obtain Q_i in depth units. Since surface runoff must be generated from such an event in order to determine a reasonable value for UZK, the UZFWC must be completely full at this point. Thus, UZFWM can be computed as:

$$UZFWM = Q_i / UZK \qquad (7\text{-}5\text{-}8)$$

If UZFWM cannot be derived from a hydrograph analysis, which is the most common case, the initial value of the parameter can be based on some general guidelines depending on how frequently surface runoff occurs. Surface runoff can be detected prior to running the model by comparing the amount of immediate storm runoff to the amount of rain+melt for a given event. When the immediate storm runoff exceeds about 50% of the rain+melt, it is likely that surface runoff occurred. The guidelines are given in Table 7-5-1. Estimates of UZFWM derived from soil data could also possibly be used as initial estimates. The soil based UZFWM values from the evaluation by this author were generally in the same ballpark as the calibrated values for most watersheds, though most of the watersheds examined generated surface runoff infrequently. For the one watershed that had frequent surface runoff (woon3 - 6-10 times/year), the soil based value of UZFWM was over twice as large as the calibrated value.

Table 7-5-1. Guidelines for initial estimate of UZFWM.

Frequency of Surface Runoff	Suggested Initial Value of UZFWM
Every moderate to heavy rainfall event (i.e. very frequently)	10 - 20 mm
Every large rainfall event	15 - 30 mm
Only during the largest flood events	30 - 60 mm (upper end of range for very wet regions)
Never or only during a record flood event	40 - 100 mm (upper end of range for very wet regions)

ADIMP

The ADIMP parameter indicates the maximum amount of variable impervious area within the watershed. These are portions of the watershed that become completely saturated and thus act as impervious areas as the soil moisture increases. The variable impervious area portion of the model is related to the variable contributing area concept that has been described in hydrology literature. The portions of the watershed that can be modeled using this feature are areas adjacent to the stream channels and areas along ravines that drain directly into the channel system. If such areas generate fast response runoff when the soil is wet no matter what is the rainfall intensity, then the ADIMP parameter is needed.

Generally it is recommended to initially set ADIMP to 0.0 and then determine during the calibration if a non-zero value is needed. However, in some cases the need for ADIMP and the computation of an initial estimate can be determined from a hydrograph analysis. What is required is a watershed with a quick response time for the channel system (i.e. unitgraph peaks in 6-12 hours). In this case an estimate of ADIMP can be made by examining moderate intensity rainfall events that occur when the soil is very wet (generally use moderate intensity storms that occur within a few days after a major event). Such a case is shown in Figure 7-5-12.

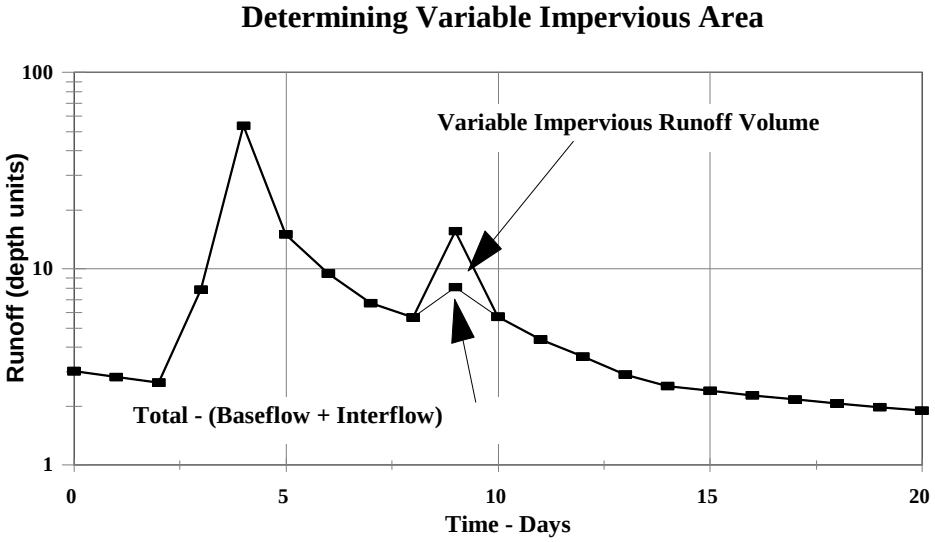


Figure 7-5-12. Determination of variable impervious area runoff volume.

An estimate of ADIMP can be computed from the runoff volume, R_v, and the precipitation, P, for this event as:

$$ADIMP = \left(R_v / P\right) - PCTIM \qquad (7\text{-}5\text{-}9)$$

PFREE

The PFREE parameter specifies what decimal fraction of the percolated water goes directly to lower zone free water storages when the lower zone tension water is not full. While this parameter would not be required if one were modeling a column of soil, it is needed when the model is being applied to a watershed. Over an entire drainage area the capacity of the lower zone tension water varies due to variations in soil properties and the depth of the root zone. This results in the tension water storage filling in some parts of the watershed before the capacity is reach over the whole area. In reality the fraction of percolation going to free water should not be a single value, but should be a curve with no percolation going to free water storages when LZTWC=0.0 and nearly all the percolation recharging baseflow as LZTWC approaches LZTWM. However, the developers of the model decided that given the simplified nature of the algorithms that a single value was adequate.

A general idea of the initial value of PFREE can be obtained by examining the hydrograph when moderate rains occur during otherwise dry periods or when the baseflow is first being recharged after a long dry summer. The amount of recharge that occurs during these periods is an indication of the magnitude of PFREE. This is illustrated in Figure 7-5-13.

PFREE Parameter

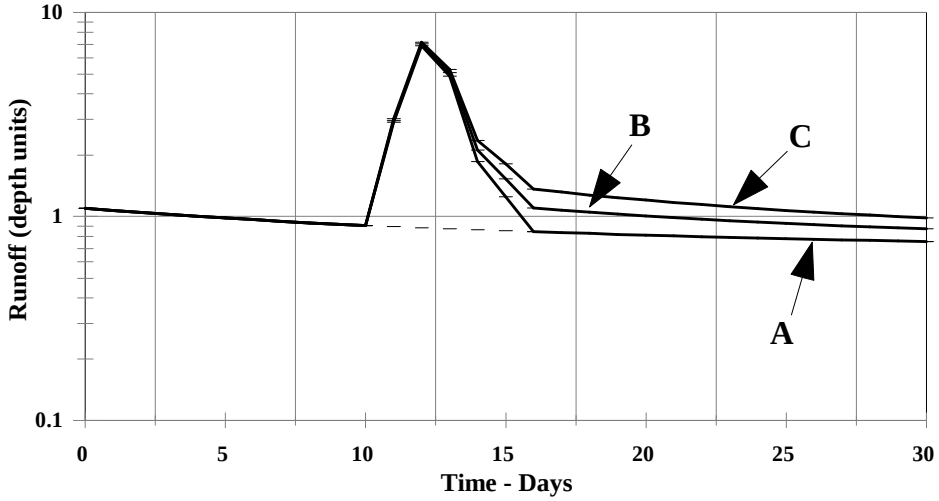


Figure
7-5-13. Evaluation of baseflow recharge during dry periods to estimate PFREE.

No baseflow recharge occurs in case A indicating that either UZTW never filled during the event or that PFREE should be 0.0. Case B shows some recharge occurring and would suggest using an initial value of PFREE in the range of 0.1 to 0.25, while case C has more recharge and would indicate setting PFREE to a value in the range of 0.3 to 0.5.

ZPERC and REXP

When considering the variation in percolation rates from dry to wet conditions, one should be thinking in terms of what the curve should look like instead of dealing with ZPERC and REXP as individual parameters. The shape of the percolation curve should be based on the type of soil that exists over the area. For example, a predominately sandy soil would have a large permeability when wet, but the ratio of dry to wet percolation rates would be relatively small and there wouldn't be much curvature to the relationship, whereas a clay soil would have a much larger permeability when dry than the low percolation rate that would exist when the soil was wet and there would be much more curvature to the relationship. This is illustrated in Figure 7-5-14.

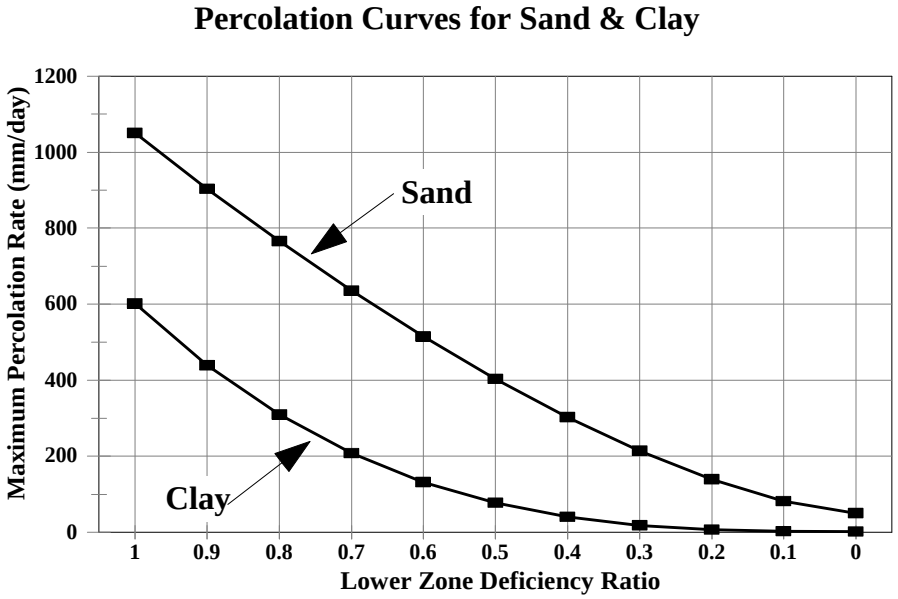


Figure 7-5-14. Sample percolation curves for sand and clay soils.

Hydrograph characteristics, as well as soils information, can indicate the general soil type and its permeability and thus can be used to obtain an initial estimate of the ZPERC and REXP parameters. Guidelines for estimating these parameters are given in Table 7-5-2. These guidelines are based primarily on logic considerations and not on actual calibration results. The relationship between ZPERC and REXP values from calibrations and soil types or hydrograph characteristics is not well defined. This is because for most watersheds the range of lower zone deficiency ratios for most significant runoff events is quite small (generally in the order of 0.2 to 0.3). Over such a small range, various values of ZPERC and REXP can produce similar percolation rates. When using real data with most of the significant events occurring at similar moisture levels, the ZPERC and REXP parameters are not very sensitive, thus the calibrations end up with a variety of values.

ZPERC and REXP (Continued)

The value of REXP derived from soil properties using the Koren method should give similar values for REXP as shown in the table. Soil based values of ZPERC may not be as useful due to questions concerning the assumption that the maximum daily percolation rate is equal to the combined capacity of all the lower zones, both tension and free water, as well as uncertainty in the capacity of these zones derived from soil data. From a physical standpoint, the value of REXP should always be greater than 1.0 as the shape of a percolation curve should always be concave.

General soil type	Hydrograph characteristics	Initial ZPERC and REXP
Clay	Frequent surface runoff, Little baseflow (max of 1 mm/day), PBASE: 2 - 4 mm/day	ZPERC: 150 - 300 REXP: 2.5 - 3.5
Silt	Some surface runoff - especially during larger storms, Moderate amount of baseflow (max of around 2 mm/day), PBASE: 4 - 8 mm/day	ZPERC: 40 - 150 REXP: 1.8 - 2.5
Sandy	No surface runoff or only during the very largest storm events, Considerable baseflow (max greater than 2.5 mm/day), PBASE: greater than 8 mm/day	ZPERC: 20 - 40 REXP: 1.4 - 1.8

Table 7-5-2. Guidelines for initial values of ZPERC and REXP.

OTHER PARAMETERS

Suggestions for initial values of the other Sacramento model parameters are as follows:

• RIVA – the need for RIVA should be determined when examining the hydrographs to decide how to assign runoff components as discussed earlier in this section, but the initial value should be set to 0.0. The final value of RIVA will be determined near the end of the calibration as discussed in Section 7-1.

• RSERV – this is a very insensitive parameter and in almost all cases a value of 0.3 is reasonable.

• SIDE – the initial value of SIDE should be set to 0.0.

• Seasonal PE adjustment or ET-Demand curve – recommendations and guidelines for these curves are given in Section 6-5.

• EFC – this parameter is only used when the snow model is included and an areal extent of snow cover time series is passed from the snow model to the Sacramento model. In that case the ET-Demand is modified when snow is present using the equation:

$$D_s = \left(\mathrm{EFC} \cdot \mathrm{D_{ns}}\right) + \left[\left(1 - \mathrm{EFC}\right) \cdot \left(1 - \mathrm{S_c}\right) \cdot \mathrm{D_{ns}}\right] \qquad (7\text{-}5\text{-}10)$$

where: D_s is the ET-Demand with snow, D_{ns} is the ET-Demand without snow, and S_c is the decimal fraction areal extent of the snow cover. The initial value of EFC should be the portion of the area covered by conifer trees times the average cover density of the conifers expressed as a decimal fraction.

• Preliminary (Frost Index) frozen ground model – based on limited use of these algorithms the following initial values are suggested:
  ◦ CSOIL - 0.1 for open areas and 0.05 for forested areas
  ◦ CSNOW - 0.08
  ◦ GHC - 0.1
  ◦ RTHAW - 0.0 (don’t use unless clearly needed)
  ◦ FRTEMP - -3.0
  ◦ SATR - 0.0 (with SATR=0.0 the frost index will have no effect on interflow withdrawal and percolation rates – start with SATR=0.4 once it is determined that the use of the frozen ground model may be helpful)
  ◦ FREXP - 8.0

Typical Range for Parameter Values

Based on experience with the Sacramento model over a wide range of physiographic conditions, Table 7-5-3 gives the typical range of values for each of parameters. In some cases the value could fall outside this range, but if so, there needs to be clear evidence that such a value is required. For example, in some watersheds the value of LZPK may be greater than the upper limit shown in the table, but in most cases when LZPK is greater than this limit, it is very likely that what should be modeled as supplemental baseflow is being treated as primary baseflow.

Parameter	Lower Limit	Upper Limit
LZPK	0.001	0.015
LZSK	0.03	0.20
LZFPM	40.	600.
LZFSM	15.	300. (highest values associated with low values of LZSK)
UZTWM	25.	125.
LZTWM	75.	300.
UZK	0.2	0.5
UZFWM	10.	75.
PFREE	0.0	0.5
PCTIM	0.0	0.05
ADIMP	0.0	0.20
ZPERC	20.	300.
REXP	1.4	3.5
RIVA	0.0	0.20

Table 7-5-3. Typical range of values for the Sacramento model parameters.

Section 7-6

Unit Hydrograph Application for the Sacramento Model

Introduction

A number of methods could be used to convert runoff generated by a rainfall-runoff model into a discharge hydrograph at a location on the channel system. The only current method included in NWSRFS is the unit hydrograph procedure. This section describes how a unit hydrograph should be applied with the Sacramento model and methods for determining an initial estimate of the unit hydrograph for a drainage area. Also discussed are items to consider when using a unit hydrograph including the time interval of the ordinates, application to watersheds that are divided into multiple zones, and how to handle watersheds where the shape of the response varies considerably depending on the flow level.

Function of a Unit Hydrograph for Use with the Sacramento Model

The Sacramento model generates runoff into the channel system. Delays that occur as water moves through the soil, either as interflow or baseflow, are accounted for within the model. Time delays associated with surface and impervious runoff are not included in the Sacramento model as these runoff components are added to the channel runoff during the same period as the rain or melt occurs. Thus, when a unit hydrograph is used with the Sacramento model, the unit hydrograph must account for the delay and attenuation that occurs as surface and impervious runoff moves over the land surface and, most importantly, the delay and attenuation that occurs as the water moves through the channel system. Typically the response time of the channel system is much more significant than the overland flow response time and thus dominates the unit hydrograph.

In more traditional applications of the unit hydrograph method, such as with an API type rainfall-runoff model, the procedure must account for storm runoff delays both through and over the soil, as well as the effect of the channel system on watershed response. For a watershed where surface runoff dominates storm events and there is very little interflow produced, the function of a traditional unit hydrograph is basically the same as the unit hydrograph that would be used with the Sacramento model. However, for a watershed where interflow, or possibly even supplemental baseflow, is a major contributor to storm runoff, the delay time for the water moving through the soil needs to be included in the unit hydrograph and thus a traditional unit hydrograph is not appropriate for use with the Sacramento model. Figure 7-6-1 shows a traditional, i.e. API model, unit hydrograph for a watershed that has a considerable interflow contribution during most storm events. The figure also shows the interflow component of the API unit hydrograph and the appropriate Sacramento model unit hydrograph for this watershed. The Sacramento model unit hydrograph is derived by subtracting the interflow component from the API unit hydrograph and adjusting the ordinates so that they add up to the correct sum.

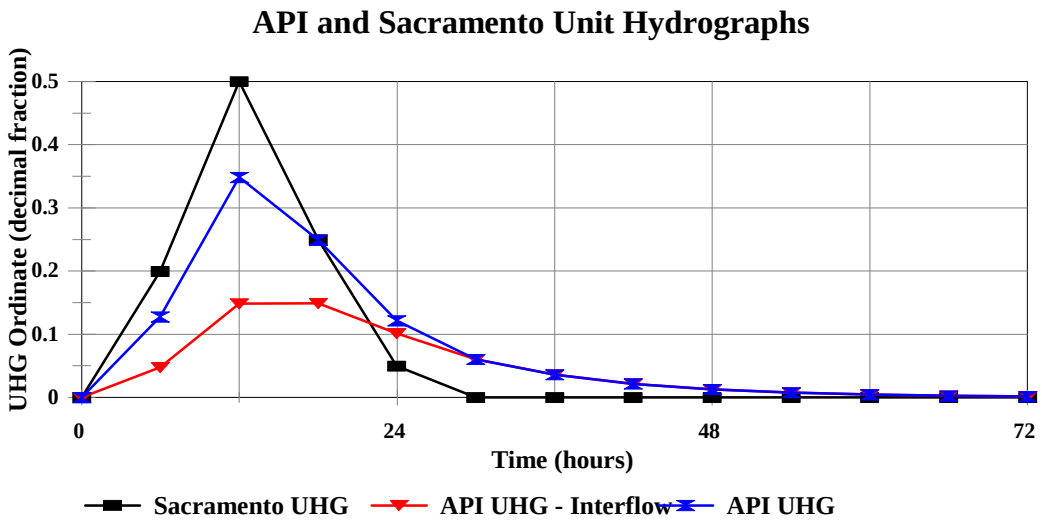


Figure 7-6-1. Unit hydrographs for a watershed with considerable interflow.

When using the unit hydrograph method with an API model, the shape of the storm response is always the same no matter how much storm runoff is generated since the unit hydrograph is a linear function and the storm runoff is all applied to the same time interval when the rain or melt occurs. When the unit hydrograph is used with the Sacramento model, even though the unit hydrograph is a linear function, the watershed response is nonlinear because the breakdown of total storm runoff into interflow and surface runoff varies depending on the amount of rain, the rainfall intensity, and the soil moisture conditions. With the Sacramento model only the fast response runoff, i.e. surface and impervious, is all produced at one time. Interflow and baseflow contributions are made available to the unit hydrograph over a number of time intervals after the rain or melt occurs. Using the unit hydrographs shown in Figure 7-6-1, the next two figures illustrate how the Sacramento model response differs from an API model response for synthetic storm events of different magnitudes. Figure 7-6-2 shows a watershed response to various amounts of storm runoff when an API model is used. The shape of the response is the same for all three cases. Figure 7-6-3 shows the response of the watershed when using the Sacramento model for the same three events. The event with 8.15 mm of storm runoff only generates interflow, while the 21.23 mm event has 57% surface runoff and 43% interflow, and the 50.1 mm case is produced by 81% surface runoff and only 19% interflow. When comparing these two figures one can see that by using the unit hydrograph to model only the channel system and not interflow response with the Sacramento model, that small events can have a lower peak and more damped response, while large events will have a higher peak and quicker response than when using an API model with a unit hydrograph.

It is very important to clearly understand what the unit hydrograph is being used to represent when working with the Sacramento model. If unit hydrographs derived for use with an API

model are to be used with the Sacramento model, any interflow or baseflow contribution must first be removed so that the resulting unit hydrograph only represents the delay and attenuation of water through the channel system and over the land surface.

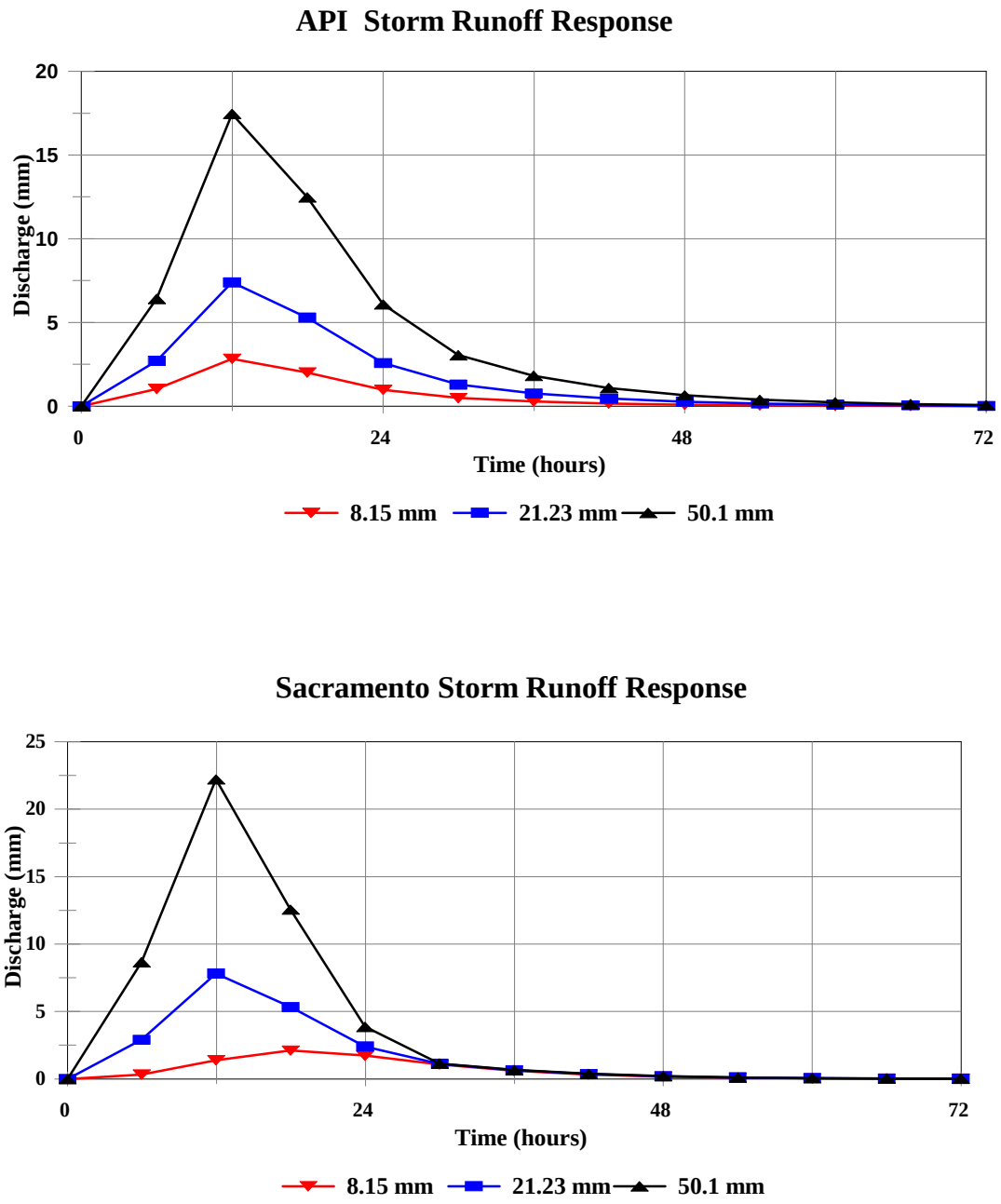


Figure 7-6-2. API model response to different size storm events.

Figure 7-6-3. Sacramento model response to different size storm events.

Derivation of an Initial Unit Hydrograph for the Sacramento Model

There are two basic methods for deriving an initial estimate of the unit hydrograph for use with the Sacramento model. The first is to use data from actual storm events for the watershed and the second is to use one of a number of techniques to derive a synthetic unit hydrograph. If the proper conditions exist it is much better to derive the unit hydrograph using actual storm data.

Derivation of a Sacramento Model Unit Hydrograph from Storm Events

The procedure to derive a Sacramento model unit hydrograph from actual watershed data involves separating the fast response runoff contribution, i.e. surface and impervious, from the other runoff components prior to doing the unit hydrograph computations. In order to do this the following conditions must be met:

- the storms used must produce a significant amount of fast response runoff,
- there must be a reasonably clean recession period for some time after each event, typically in order of several weeks, and
- the precipitation causing the storm response should occur within a few time intervals; all within one period is best, but up to 3 or 4 time intervals are okay (complex storms and snowmelt periods should be avoided as when deriving a traditional unit hydrograph).

In order to separate out the fast response runoff contribution so that the effect of the channel system and overland flow can be isolated, a recession analysis is used. The recession analysis should allow one to remove the baseflow and interflow contribution. This is analogous to separating out baseflow when deriving a traditional unit hydrograph only in this case the interflow contribution must also be removed. Once the watershed response to surface and impervious runoff is determined, then the remainder of the process is exactly the same as when computing unit hydrograph ordinates in the traditional case. If one is not familiar with these steps, they are described in many textbooks.

In order to perform a recession analysis, a semi-log plot is generated for the receding portion of the hydrograph. Such a case is shown in Figure 7-6-4. This figure contains an idealized case generated with synthetic data so that the steps in the procedure can easily be described. The steps in the recession analysis are:

- Identify the primary baseflow component if possible. This component will plot as a straight line on the semi-log plot after interflow and supplemental baseflow storages have been drained completely. In the figure, the plot is not extended all the way out to this

point. Extend the primary baseflow contribution back to under the peak of the hydrograph and then subtract the primary flow component from the total flow. The result is shown as ‘total - primary’ in the figure.

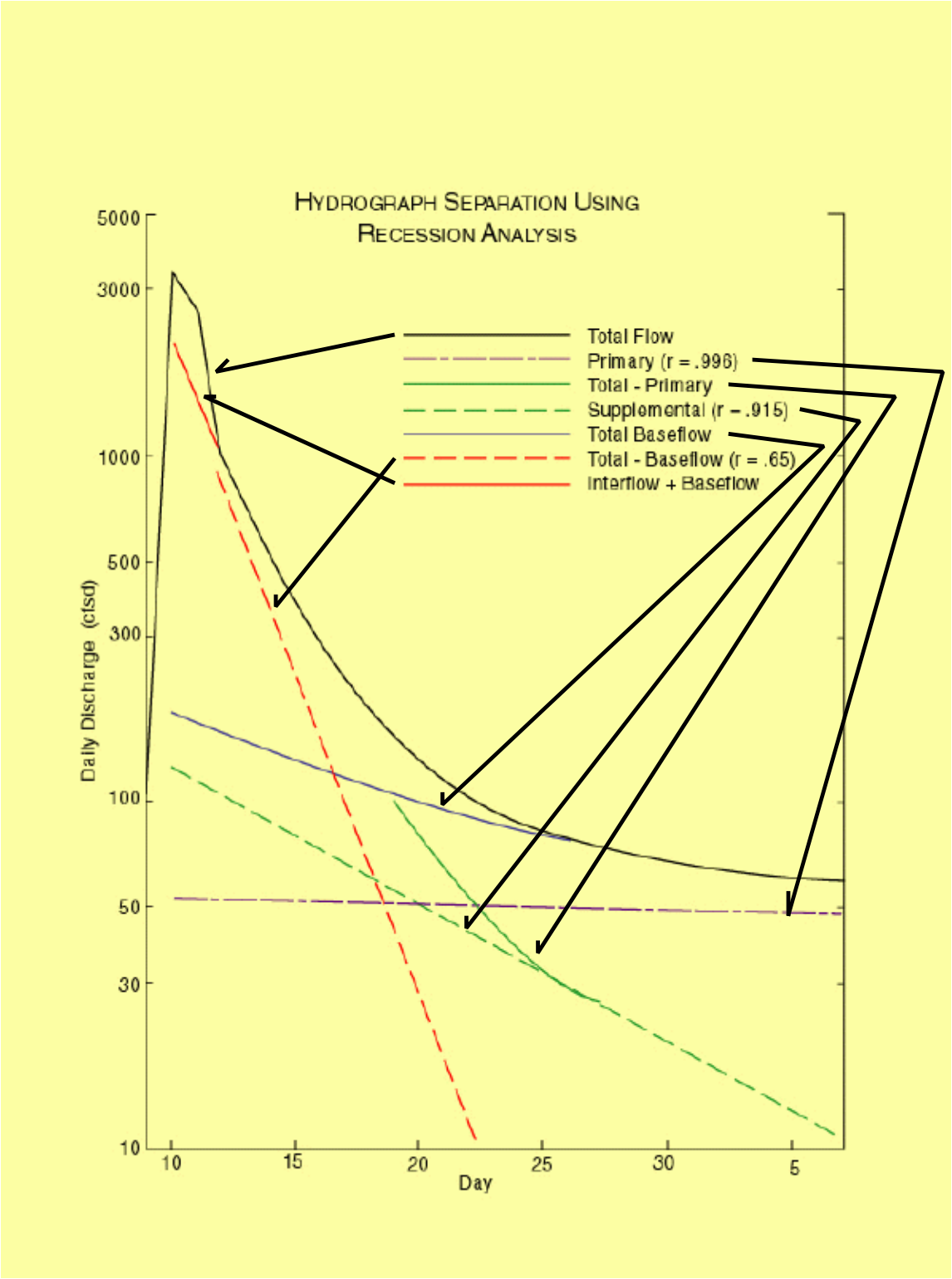


Figure 7-6-4. Illustration of a recession analysis using synthetic data.

• The straight line portion of the total minus primary line on the plot identifies the supplemental baseflow component. Extend the supplemental baseflow contribution back to under the hydrograph peak and then add the supplemental baseflow to the primary amount to get the total baseflow for this event. Then subtract the total baseflow from the total flow to get the 'total - baseflow' contribution. The total minus baseflow portion is shown on the plot starting at about the point where fast response runoff ceases though it could be extended back under the peak. In many real cases it is difficult to separate out primary and supplemental baseflow contributions, especially when the recession period is not very long or the supplemental withdrawal rate is quite slow. In these cases one would try to identify the total baseflow contribution, rather than dealing with primary and supplemental separately. The total baseflow recession will not plot exactly as a straight line on a semi-log plot since it is comprised of flow from two aquifers with different recession rates.

The 'total - baseflow' line will normally plot as a straight, or nearly straight, line on the semi-log plot. A recession rate can be computed for this segment. This is not the interflow recession rate, but instead an indication of how fast water is draining from the upper zone free water storage both as percolation and interflow. If soil moisture conditions are nearly saturated and this is a watershed with very low percolation rates under wet conditions, the amount of percolation should be small and thus, the total minus baseflow recession rate would be only slightly greater than the interflow recession rate and could be used to calculate an initial estimate of the Sacramento model UZK parameter. In most cases the 'total - baseflow' recession rate will be considerably greater than the interflow recession rate and can only be used as an extreme upper limit for UZK.

• The straight line portion of the 'total - baseflow' amount indicates the interflow contribution to the storm. This straight line segment is then extended back under the hydrograph peak and the total baseflow is added to the interflow contribution to obtain the amount of 'baseflow + interflow' for the event. This amount is only less than the total flow for a short time after the rain that produced the rise. During the remainder of the recession all the flow is either interflow or baseflow.

• The last step is to plot the total baseflow and the baseflow plus interflow amounts on an arithmetic plot as shown in Figure 7-6-5. Some subjectivity is needed when drawing the recharge portion , i.e. the period from when the hydrograph starts to rise until these flow components start to recede. The upper zone free water storage and thus interflow should typically peak a time interval or two after the storm peak, while baseflow recharge generally will take longer and thus, the baseflow contribution will not peak until a day or two after the storm peak.

The sum of interflow and baseflow can then be subtracted from the total instantaneous discharge amount to get the fast runoff, i.e. surface plus impervious, contribution to the storm. The surface plus impervious discharge is then used to derive the unit hydrograph .

to be used with the Sacramento model using textbook methods.

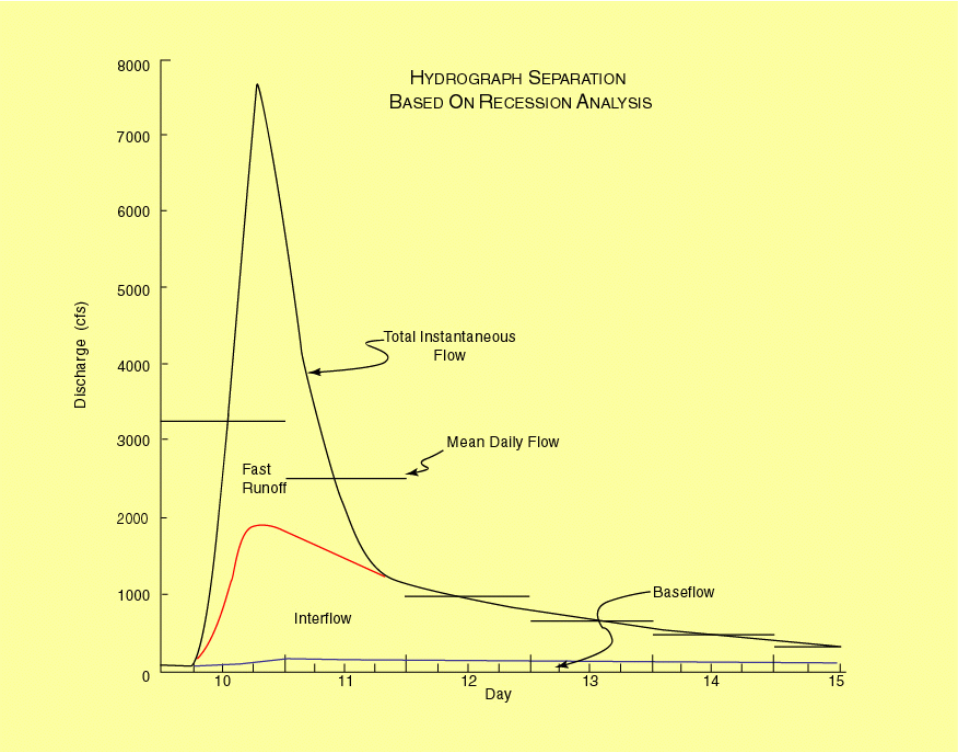
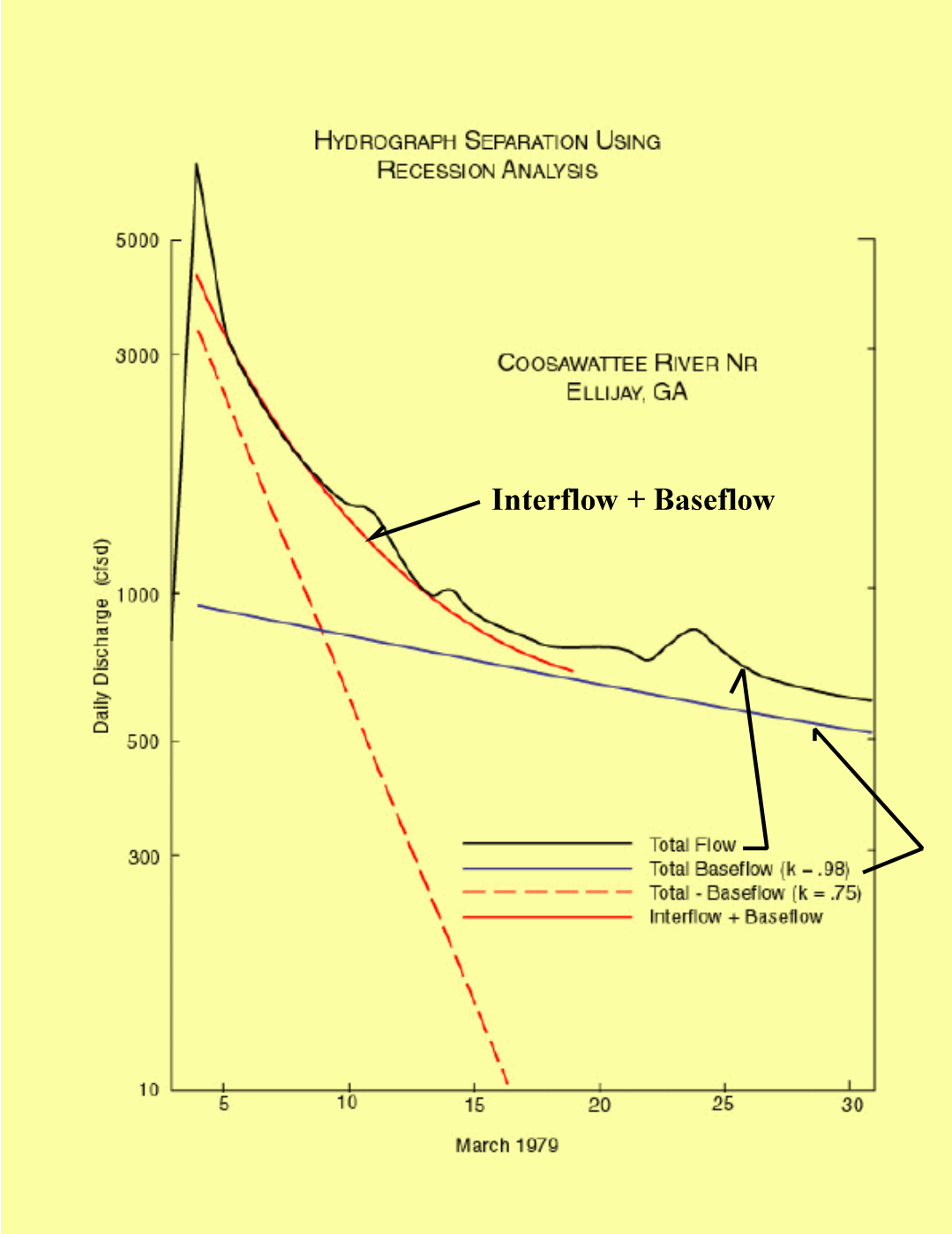


Figure 7-6-5. Separation of fast resonse runoff for deriving a unit hydrograph.

Figure 7-6-6 shows a recession analysis for an actual watershed. In this case there are several small events that occur in the period after the main storm, thus making it more difficult to determine the baseflow contribution. By looking at a low flow period further out in time than shown on the plot, the total baseflow contribution could be reasonably estimated. Then baseflow was subtracted from total flow to obtain the 'total - baseflow' amount and the straight portion of this line indicates the interflow contribution. 'Interflow + baseflow' was then computed and plotted for the storm event. Differences between total flow and 'interflow + baseflow' for the storm event that occur during the recession are the result of interflow and baseflow recharge from the subsequent small rainfall events. The total baseflow and 'interflow + baseflow' amounts are then plotted on an arithmetic plot as shown in Figure 7-6-7 to determine the fast runoff contribution to the storm hydrograph. Observed instantaneous discharge data were not available for this event so the total instantaneous flow was estimated by knowing the mean daily discharge amounts and the instantaneous peak flow. The rainfall for this event is also shown in the figure. The period of rain that generated most of the runoff was between 6 and 12 hours in length. Thus, the Sacramento model unit hydrograph computed from the fast runoff contribution to this event will be between a 6 and 12 hour unitgraph. An "S" curve analysis could be used to try and determine the appropriate time interval and then to compute a 6 hour unitgraph to be used with the model. If the storm event

doesn't contain any fast response runoff, which was the case most of the time for the Ellijay watershed (surface runoff only occurs 3 times during the period of record), then the recession analysis should show that the entire rise is primarily from interflow and the event can't be used to determine an estimate of the unit hydrograph to use with the Sacramento model.



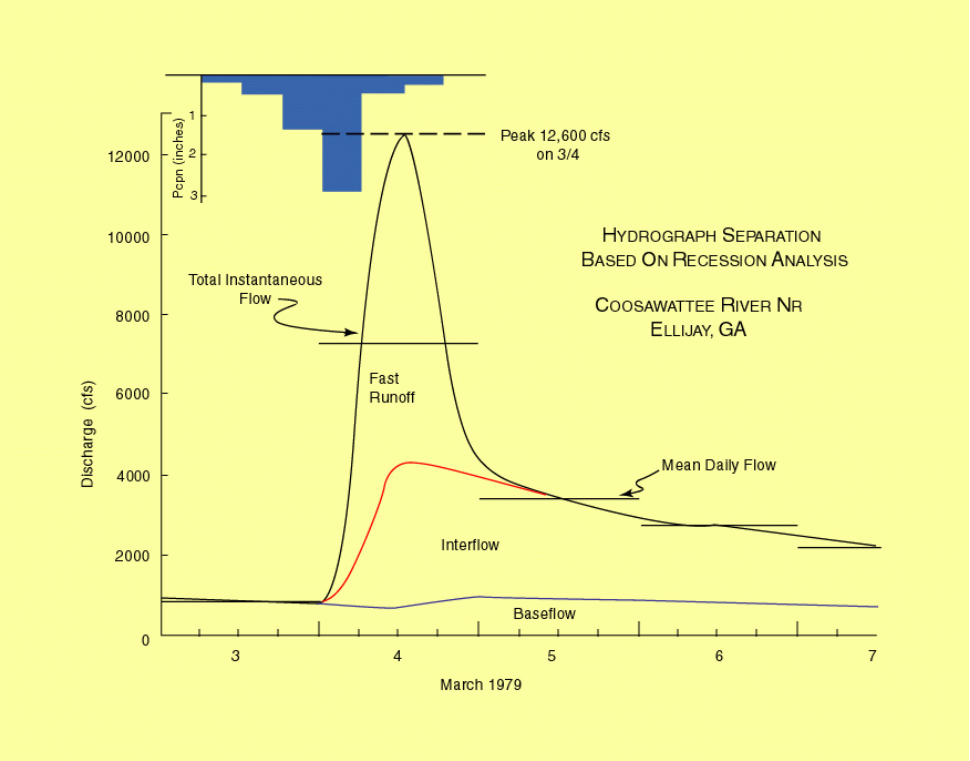
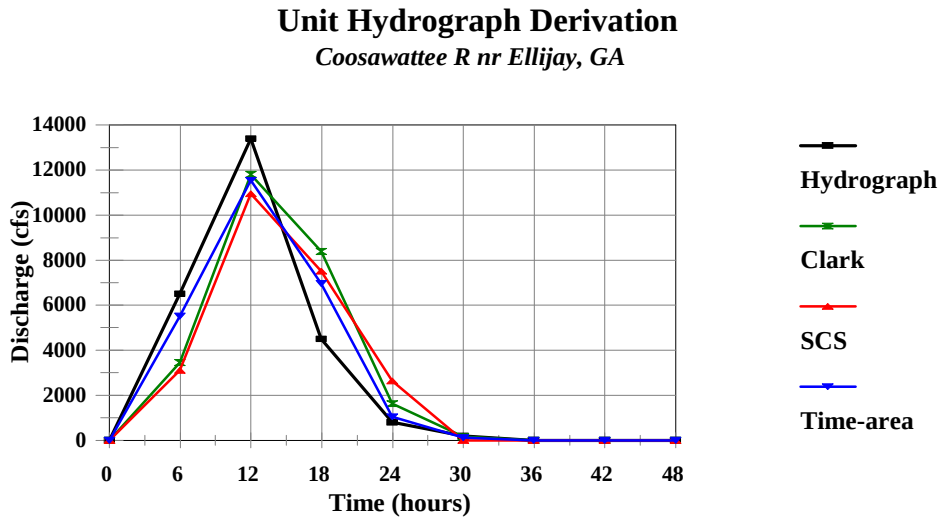


Figure 7-6-6. Recession analysis for the Ellijay watershed for March 1979 storm.

Figure 7-6-7. Fast runoff contribution to March 1979 storm for the Ellijay watershed.

Derivation of an Initial Unit Hydrograph using Synthetic Methods

If a unit hydrograph for use with the Sacramento model cannot be derived from precipitation and discharge data for the watershed, then there are a number of synthetic methods that have been proposed that are available. These methods generally fall into two categories. First



there are time of travel methods which compute the portion of the watershed contributing each time interval based on travel time estimates. Second there are methods that use watershed geometry to estimate the peak discharge, the time to peak, and the base of the unit hydrograph. These methods contain coefficients that are determined from unit hydrographs derived from actual watershed data though in many cases the coefficients may only represent a specific region and could be more appropriate for a traditional, i.e. API model, unitgraph. In addition to using a synthetic method, in some cases it is adequate to subjectively estimate the initial unit hydrograph to use with the Sacramento model based merely on typical time to peak information for major events, though this requires a certain amount of experience.

This author is not an expert on synthetic methods for deriving a unit hydrograph. Thus this manual will not include a discussion of the pros and cons of each procedure. The only warning offered when using a synthetic method is for slower responding watersheds where the storm hydrograph doesn't typically peak until days, rather than hours, after the rainfall or snowmelt occurs. For fast responding watersheds, most of the synthetic methods will give similar results, but for slower responding watersheds, none of the methods may produce the proper time delay. Figure 7-6-8 shows initial unit hydrographs generated from several methods for the Coosawattee River near Ellijay, Georgia which peaks quite quickly. Three

Figure 7-6-8. Unit hydrographs derived with different methods for Ellijay, GA.

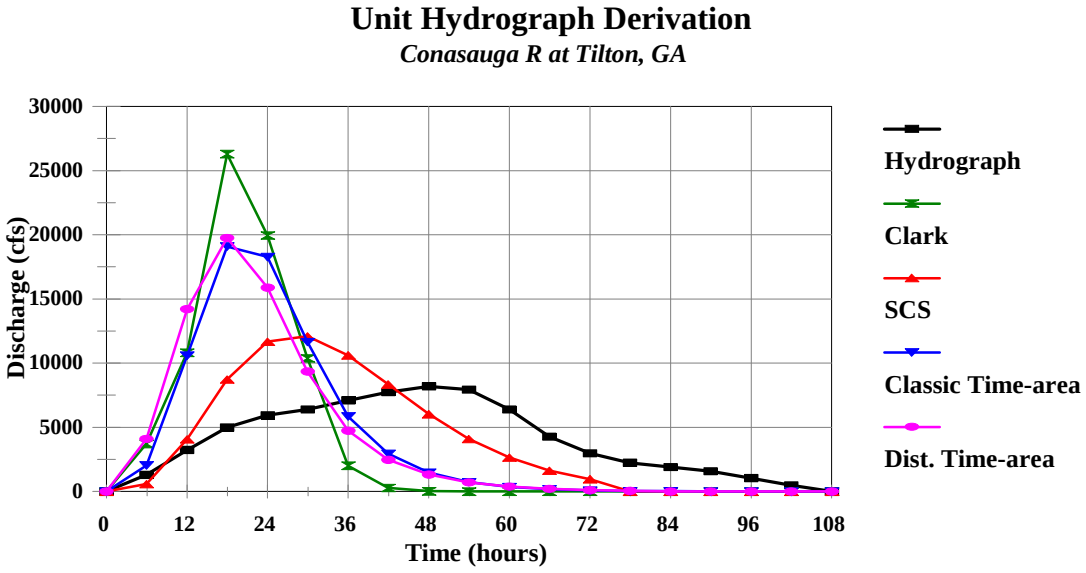
Figure 7-6-9. Unit hydrographs derived with different methods for Tilton, GA.

synthetic unit hydrographs are compared to one derived from actual hydrograph data. For this watershed that peaks in 12 hours, all the unitgraphs are quite similar. Figure 7-6-9 is for the Conasauga River at Tilton, Georgia. For this watershed that takes a couple of days to peak, the synthetic methods all produce unit hydrographs that peak much quicker than the one derived from actual storm data. The unitgraph generated by the SCS method is more delayed than the other synthetic ones, however, the person who generated the synthetic unitgraphs was warned prior to running the SCS method that the time delay was quite a bit longer than the other synthetic methods were showing and thus the coefficients were changed based on this information to produce a longer delay. For watersheds that don't have a unit hydrograph with the typical shape associated with most drainage areas, generally due to the geometry of the drainage area, the time-area methods should have a better chance of having the proper shape than other synthetic methods, though there still may be problems with the timing for slow responding watersheds.

Other Considerations when Deriving Initial Unit Hydrographs

There are several other items that should be considered when deriving and using unit hydrographs. These include:

Watersheds with Multiple Zones - When a watershed is divided into several zones, typically



either based on elevation or travel time, for rainfall-runoff computations, there are two basic

sequences that can be followed when applying a unit hydrograph. First, the runoff from each zone can be weighted and the resulting mean watershed runoff can be applied to a single unit hydrograph for the drainage area. Second, the runoff from each zone can be applied to a separate unit hydrograph and then the results combined to produce the instantaneous flow hydrograph for the watershed. If separate unitgraphs are to be used for each zone, it is generally not possible to derive the response for each zone from a hydrograph analysis unless events can be found where all of the runoff comes from only one zone per event. Typically in this case one of two approaches are followed. One approach is to derive a total watershed unit hydrograph from storm data and then divide it into unitgraphs for each zone in such a way that the combined response is the same as the unit hydrograph derived for the total area taking into account that the runoff contribution from each zone may not have been the same. Another approach is to use a synthetic method to derive the unitgraph for each of the zones individually. For zones where no runoff reaches the designated river location for some period of time, i.e. there is some distance between the river location and the closest part of the zone, either zero ordinates can be inserted at the beginning of the unitgraph or the LAG/K operation can be used to lag the unitgraph by the required amount.

For watersheds that are divided into multiple zones based on travel time, separate unit hydrographs should be used for each zone or it defeats the purpose for subdividing the watershed. For watersheds with multiple elevation zones, it depends largely on the distribution of the elevation zones as to whether single or multiple unit hydrographs should be used. When the elevation zones are significantly related to distance from the river location, then separate unitgraphs are recommended for each zone. However, when the higher elevation zone has sort of a horseshoe pattern around the watershed divide (thus some portions of the upper zone are closer to the gage than a large part of the lower zone) then the use of a single unit hydrograph for the watershed is generally adequate. Also, in these cases it is typically very difficult to derive separate unitgraphs that can be verified as to their timing and shape.

Time Interval of Ordinates – Unit hydrographs are designated with a time interval that represents the time interval associated with the runoff being applied to the unitgraph. Thus, if 6 hour runoff data are being calculated and passed through a unit hydrograph to obtain discharge at a point on the river, a 6 hour unit hydrograph must be used. The time interval associated with the unit hydrograph has nothing to do with the ordinate spacing that is used to define the unitgraph. A unit hydrograph is a continuous curve that defines the instantaneous hydrograph produced by one unit of runoff over a specified time interval. The ordinates that are used to define that curve can be at whatever time interval is necessary to adequately define the shape of the curve. Thus, for example, if a 6 hour unit hydrograph for a watershed typically peaks in 15 hours, 3 hour ordinate spacing should be used to define the unitgraph so that the peak will occur in 15 hours and not in 12 or 18 hours as would be required if the ordinates were defined at 6 hour intervals.

Variable Channel Response – For some watersheds the shape and timing of the unit

hydrograph is related to the flow level. This normally occurs when the channel width increases significantly as the flow goes over the bank and there is considerable roughness in the flood plain. This is the same physical situation that exists for channel reaches that require variable routing parameters at different flow levels. This situation is quite common in areas with only a slight amount of channel slope, a fairly broad flood plain that contains a lot of vegetation, and flood levels that can be considerably higher than the flood plain elevation. Typically for such watersheds the channel response has much more attenuation when the stream first goes out onto the flood plain and then the attenuation decreases as the flow level increases and the effect of the flood plain roughness is diminished. This effect can be modeled by using a routing procedure that allows a variable amount of attenuation, such as a variable K in the LAG/K operation, as a function of the flow level immediately after the unit hydrograph is applied. The unit hydrograph defined is for the flow level with the minimum amount of attenuation and then attenuation is added as needed at other flow levels with the routing model.

Section 7-7

Effect of Snow Model Parameters on Model Response

Introduction

A critical aspect of interactive trial and error calibration is to understand the unique effect that each parameter has on model response and how to isolate that effect. This understanding allows the person doing the calibration to make educated decisions as to which parameter values should be changed as opposed to random guesses. Without this understanding, interactive calibration becomes an exercise in futility.

This section describes the unique function of each of the major snow model parameters and also discusses things to examine when deciding which, if any, parameters should be changed when simulated and observed flows don't agree at the beginning of the snowmelt season. The unique aspects of each parameter are illustrated by hydrograph plots for the appropriate time periods with different values of the parameter being considered. These plots are generated by running the models with everything exactly the same except for the value of the parameter being discussed. Under such conditions the unique effect of each parameter can clearly be shown. The actual problem of deciding on which parameters to change becomes much more difficult during calibration because of noise in the data and the fact that typically multiple parameters need to be adjusted. The combination of the effects of multiple parameters not having the proper value make it much more difficult to determine which parameter values need to be altered. To have any hope of making the proper changes in a reasonable amount of time, one must know what periods to examine and what to look for to isolate the effects of each parameter. It is also helpful to have a systematic strategy for checking the parameters for possible changes. Such a strategy is outlined in Section 7-1.

Besides plots showing the effect of changing each of the parameters, panels that show how the model states are changing and other results of the model computations are included in many of the figures. These panels are extremely helpful in determining when the conditions occur that are needed to isolate the effects of the parameter. An explanation of the items included in these panels is in the next section. Even though two hydrograph plots are included on each figure to show the difference in response with different values of the parameter, the panels showing the model states and computations can only be included for one of the parameter values for a single zone. This value and zone are indicated next to the panels on each figure.

Explanation of Panels Showing Model States and Computations

Figure 7-7-1 illustrates the panels which are part of the ICP WY-PLOT display for the SNOW-17 operation. The plots on these panels are for a daily time interval. Instantaneous values such a water-equivalent are for the end of the day and mean, average, or total values such as energy exchange and air temperature are the mean, average, or total for the day.

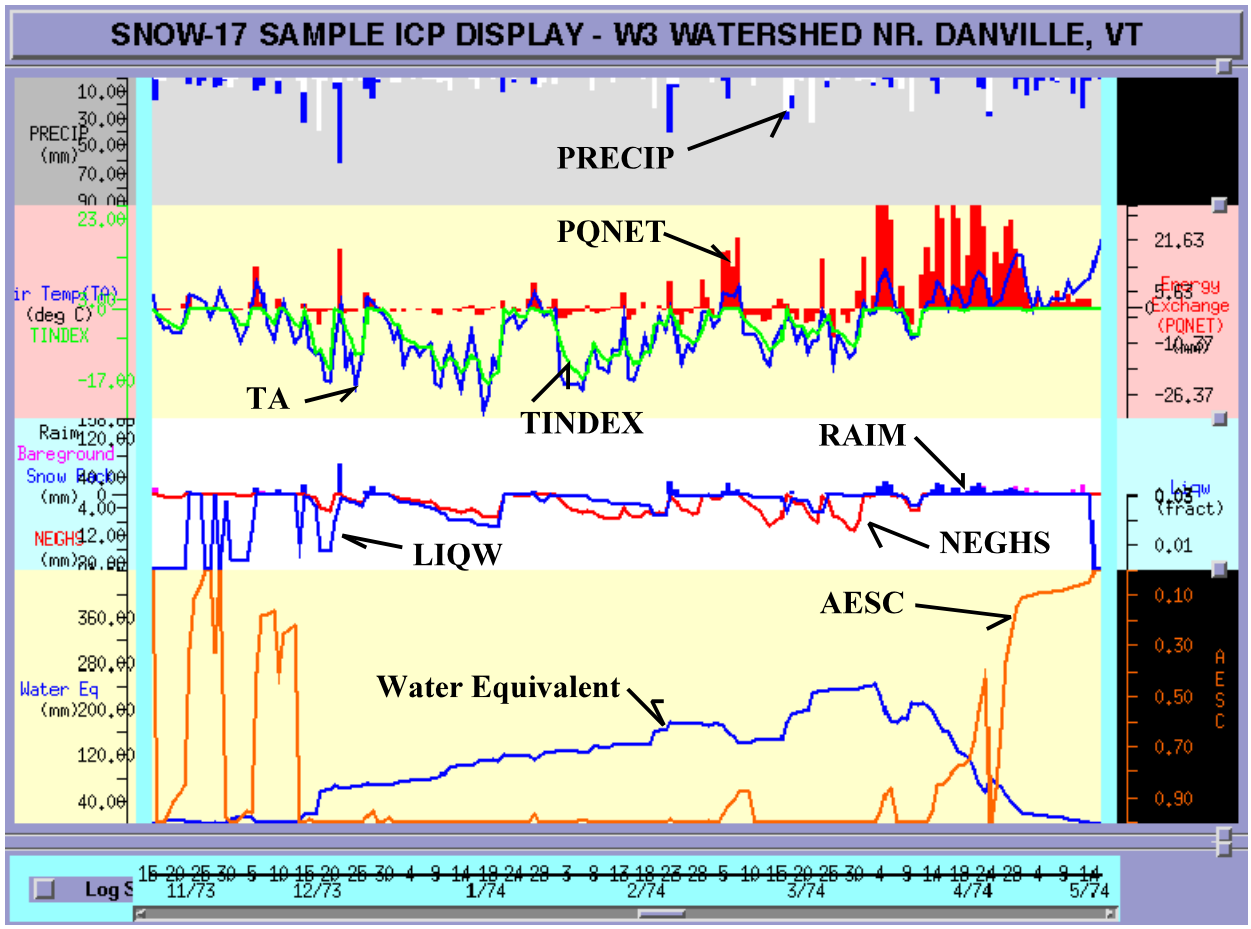


Figure 7-7-1. Sample ICP display for the SNOW-17 model.

• Top Panel – This panel shows the total amount of precipitation (PRECIP) for each day with the scale on the left side. The portion in blue indicates that the precipitation was rain, while the portion in white indicates snow. If the rain-snow elevation option is used, this panel will also contain the average elevation of the rain-snow line if it is within the elevation range of the zone. The rain-snow elevation scale will be on the right side.

• Second Panel from Top – This panel shows the average daily air temperature (TA) in blue and the average daily snow cover temperature index (TINDEX) in green with the scale on the left side. The gradient defined by these two variables indicates the direction of heat flow when the air temperature is below freezing. The panel also contains the total daily net energy exchange (PQNET) in red with the scale on the right side. The units of the energy exchange are mm (that is the energy needed to melt 1 mm of snow). The energy exchange is the combination of non-rain melt, rain-on-snow melt, and changes in the heat deficit of the snow cover during the day.

• Third Panel from Top – This panel shows the amount of liquid water (LIQW) in the snow

cover in blue with the scale on the right side. The maximum amount of liquid water is controlled by the PLWHC parameter. The brown line shows the heat deficit or negative heat storage (NEGHS) in the snow cover at the end of each day with the scale on the lower left side. The units for NEGHS are mm, i.e. the amount of melt or rain that is needed to return the snow cover to isothermal conditions at 0°C. This panel also shows the total amount of rain plus melt (RAIM) for each day with the scale on the upper left side. When the RAIM is in blue it indicates water that came from the snow cover (either melt or rain water that passed through the snow), while when RAIM is in magenta, it indicates rain on bare ground.

Bottom Panel – This panel shows the total amount of water equivalent in blue with the scale on the left side. This is the sum of the ice portion of the snow cover, any liquid water, and water moving through the snow at the end of each day. The panel also shows the computed areal extent of the snow cover (AESC) in brown with an inverted scale on the right side (bottom of plot is 100% cover and the top is bare ground).

Major Snow Model Parameters

This section describes the unique effect of each of the major snow model parameters. Examples are included for a watershed with a single zone and for most parameters, examples are also shown for watersheds with multiple elevation zones.

SCF – The snow correction factor is the only snow model parameter that controls the amount of snow that accumulates and thus the volume of snowmelt runoff. This is assuming that significant errors in the form of precipitation have been corrected. The effect of SCF shows up in the later part of the snowmelt season, i.e. when either too much or too little snow may remain and thus continues the snowmelt period beyond when it should end or cause melt to stop too soon. This is the period to examine to determine if there is a problem with the amount of snow. The higher the value of SCF the more snow will accumulate and the longer snow will remain and vice versa.

Figure 7-7-2 shows the effect of changing the value of SCF on the volume of snowmelt runoff for a watershed with only one zone. There is no difference during the period when there is 100% cover with both values of the parameter. The difference occurs during the later part of the melt season when the increase in the amount of snow with SCF=1.4 results in more runoff volume. Once the snow is gone, the difference in response between the two values disappears over time. There is some difference in runoff from the large rain event near the end of May, since the soil is more saturated at that time when SCF=1.4 due to a longer period of snowmelt, but by the rain event in mid June, the difference is minimal.

Figure 7-7-3 shows the effect of changing SCF for just the upper elevation zone of a mountain watershed for a year with a large enough amount of snow that the upper area remains at 100% cover for some period of time after melt begins and there is no significant rain during the snowmelt season. The figure also shows what portion of the runoff is

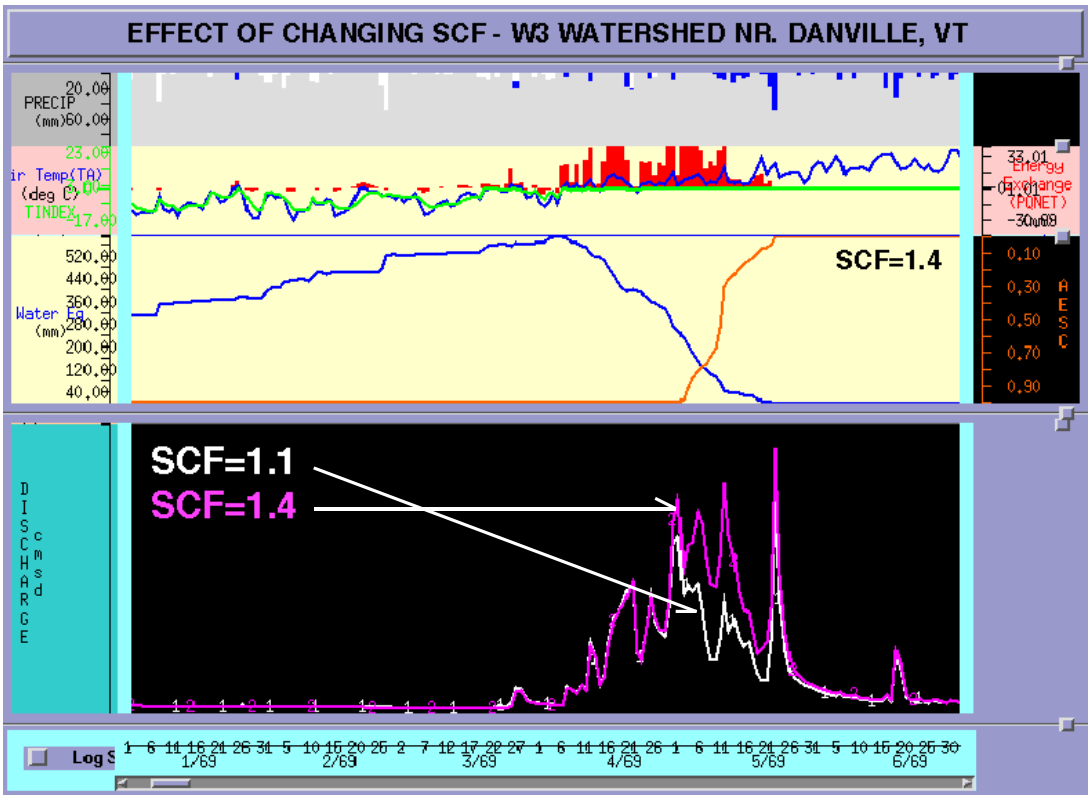


Figure 7-7-2. Effect of changing SCF for a watershed with one zone.

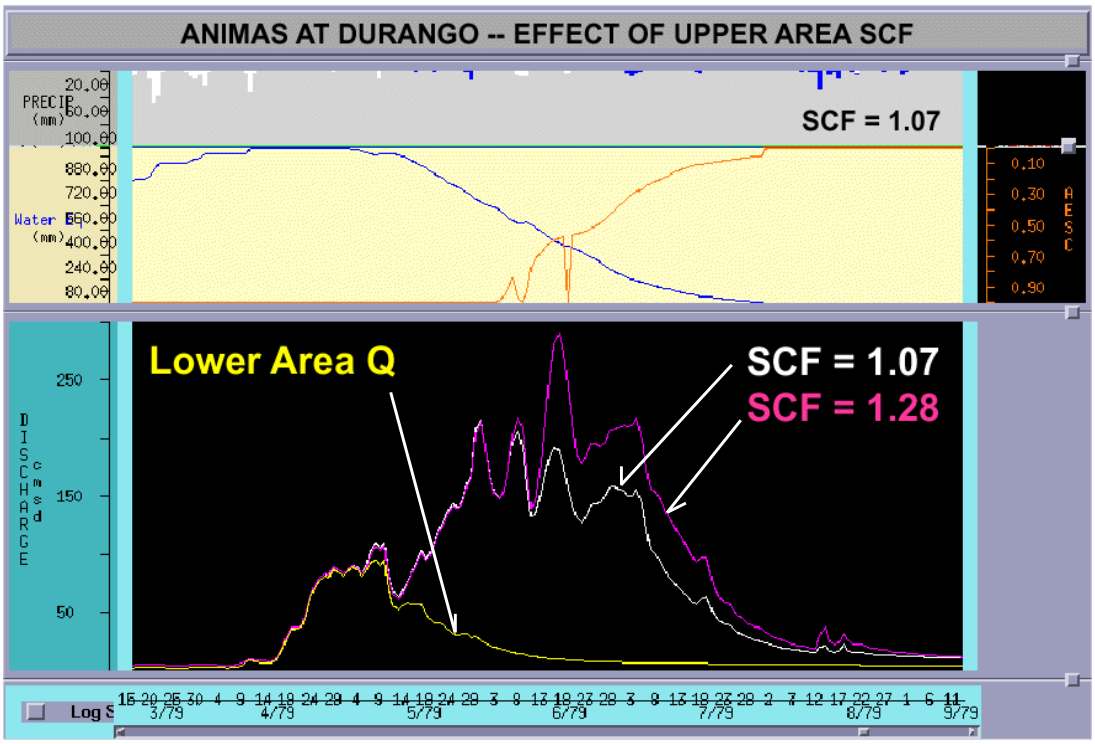


Figure 7-7-3. Effect of changing SCF for a watershed with two zones.

e 7-7-3. Effect of changing SCF for an upper elevation zone.

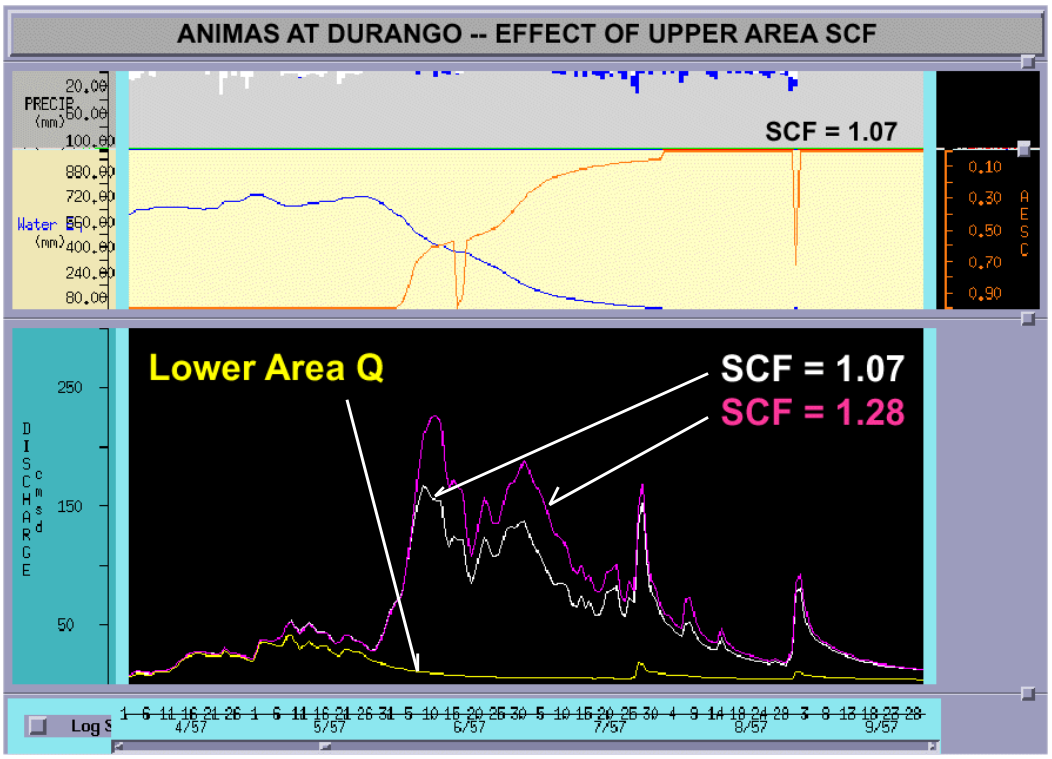


Figure 7-4. Effect of changing SCF within the upper zone F h near end of melt.

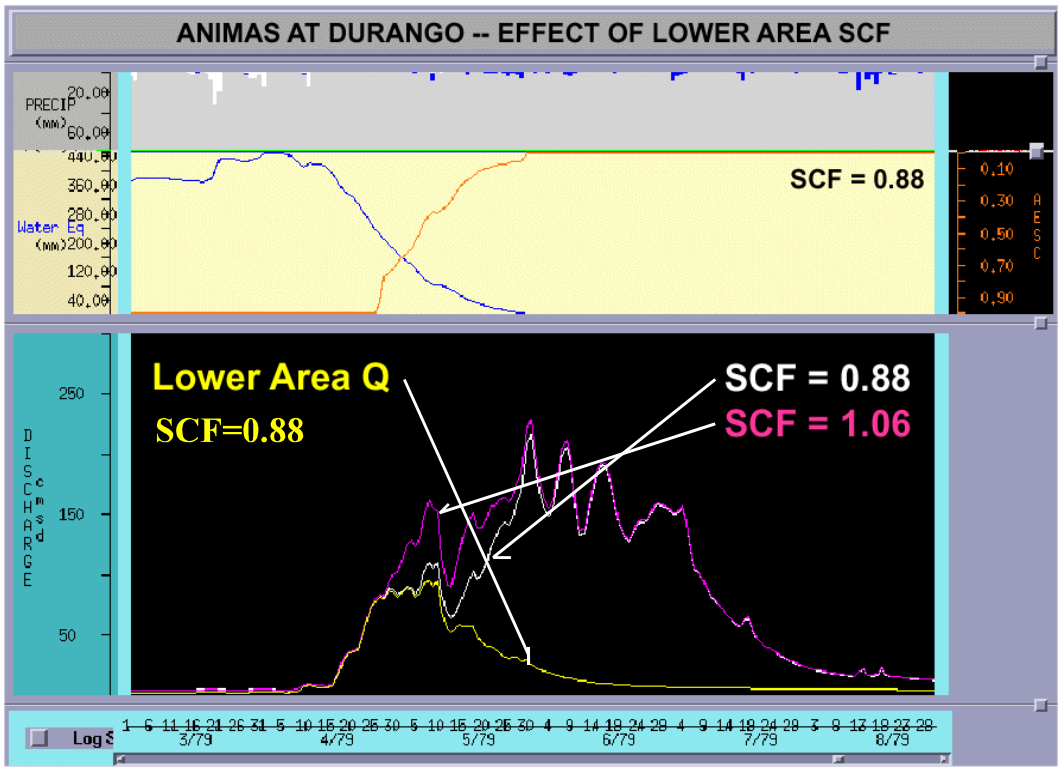


Figure 7-7-5. Effect of changing SCF for a lower elevation zone.

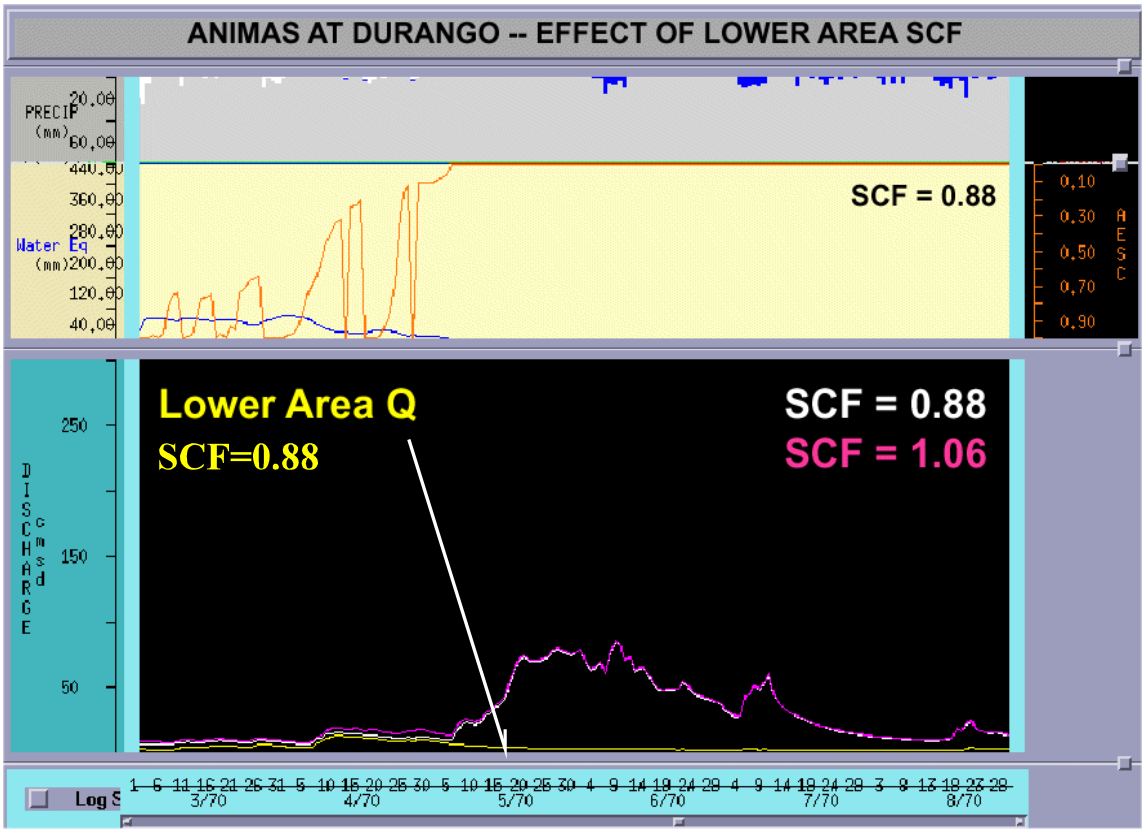
produced by the lower zone. It is very important when working with watersheds with multiple elevation zones to known the contribution of each zone. The response to snowmelt is the same as in the previous figure since the upper zone melts off last and there are several weeks with 100% cover after melt begins. Figure 7-7-4 shows the same case for another year for the same watershed. During this year the upper zone only remains at 100% cover for a short period after melt begins and the lower zone contribution is much less than in the previous example. Also during this year there is a significant amount of rain late in snowmelt season and for some time thereafter. Again the primary effect of changing SCF is to alter the amount of snow accumulation and thus the volume of runoff from snowmelt. There is some effect on the amount of runoff from the rain events, but this is small and diminishes with time. Figure 7-7-5 shows the effect of changing SCF for the lower zone for the same year as shown in Figure 7-7-3. This is a year when there is substantial volume of snowmelt runoff from the lower zone. Again by increasing the value of SCF, the volume of snowmelt runoff is increased, however, since the lower zone melts off well before the upper zone, the effect shows up toward the early part of the snowmelt season for the entire watershed. Figure 7-7-6 illustrates the effect of changing the lower zone SCF value on a year with little snowmelt runoff from the lower zone. In this case since the snowmelt runoff volume from the lower zone is minimal, the effect of changing SCF for this zone has little effect on the hydrograph.

Figure 7-7-6. Effect of changing lower zone SCF during a low snow year.

To repeat, the unique effect of the SCF parameter is to control the amount of snow accumulation and thus the volume of runoff from snowmelt. It is the only snow model parameter whose primary effect is to control the runoff volume.

Under the SCF parameter initial value discussion in Section 7-4 it was mentioned that when average annual precipitation is based on the water balance (determined by adding an estimate of actual annual ET to mean annual runoff) that the result could be too much rain and not enough snow. If this appears to be the case for such watersheds, the PXADJ factor in the SNOW-17 operation can be decreased below 1.0 to reduce the amount of rain and then SCF increased as necessary to obtain the correct amount of snow cover runoff.

MFMAX and MFMIN – These two parameters determine the maximum and minimum value of the non-rain melt factor. MFMAX has more effect when most of the melt occurs after March 21st, while MFMIN has more effect on melt that occurs prior to that date. The non-rain melt factor is used to determine the melt rate when the area is completely covered by snow, thus the effect of these parameters can be isolated when non-rain melt is occurring and



there is 100% or nearly complete areal snow cover. SCF and the areal depletion curve have

no effect when there is 100% cover and UADJ is only used when there is rain-on-snow melt occurring. A variety of factors can affect when the onset of melt occurs and the response during the very early part of the melt season (these will be discussed later in this section), thus it is best to examine the period after melt is well underway and there is still complete or nearly complete areal cover to determine if the non-rain melt factor should be altered. Since the non-rain melt factor varies seasonally and melt can occur at different times, one technique that may prove useful is to plot the seasonal melt factor variation and then note when the results would be improved by changing the melt factor up or down and when it should be left the same. After examining a number of years, hopefully there will be a pattern to the suggested changes and a new seasonal melt factor curve can be drawn and new values of MFMAX and MFMIN determined for the next iteration.

Figure 7-7-7 shows the effect of changing both MFMAX and MFMIN. In this case MFMAX has the most effect since almost all the melt occurs after March 21st. The overall effect of changing these melt factors is to cause more melt to occur early in the melt season and thus less later because the snow will be gone sooner. To isolate the effect of these parameters the period to examine is from after melt begins to when bare ground begins to show, i.e. from late March until about the 21st of April. There is a difference in the hydrographs later in the melt season, but this is not the period to use to determine if the non-rain melt factor needs to be adjusted as other snow model parameters can also affect this later period.

Figure 7-7-8 shows the effect of changing MFMAX for both elevation zones of a mountain watershed during a big snow year when there is significant snowmelt runoff from the lower elevation zone. Again the overall effect of increasing MFMAX is to cause more melt to occur early in the melt period, in this case for both elevation zones. Figure 7-7-9 shows the same change for a small snow year when there is little snowmelt runoff from the lower

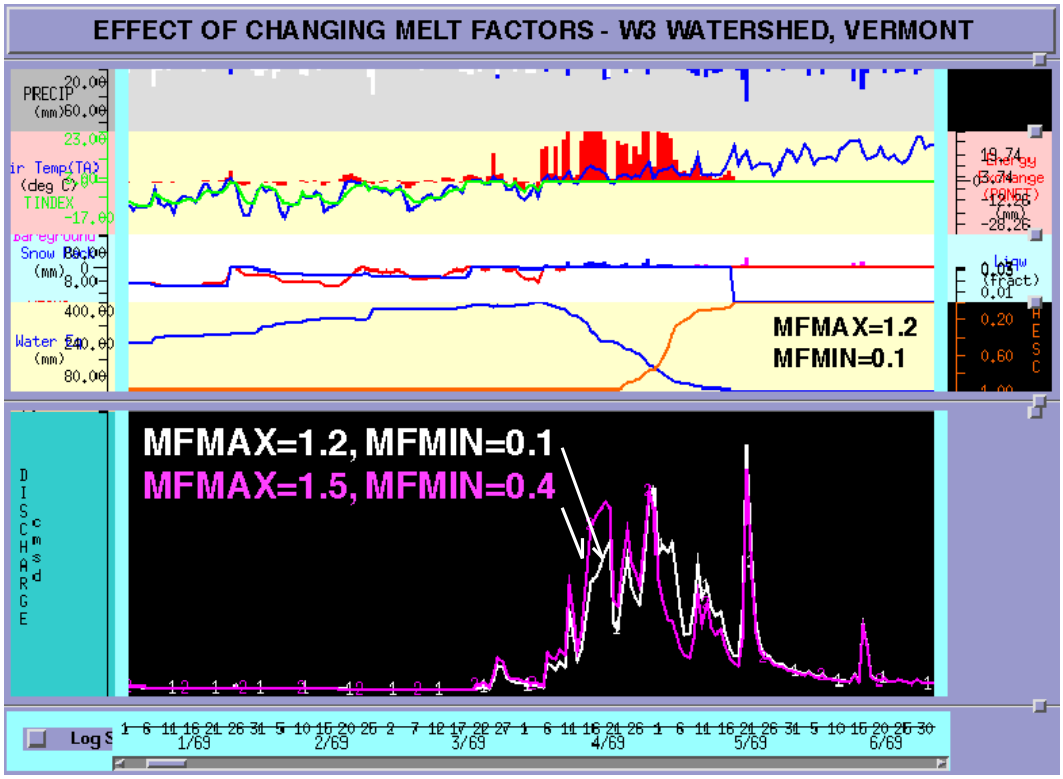


Figure 7-7- Effect of changing MFMAX and MFMIN watershed zone.

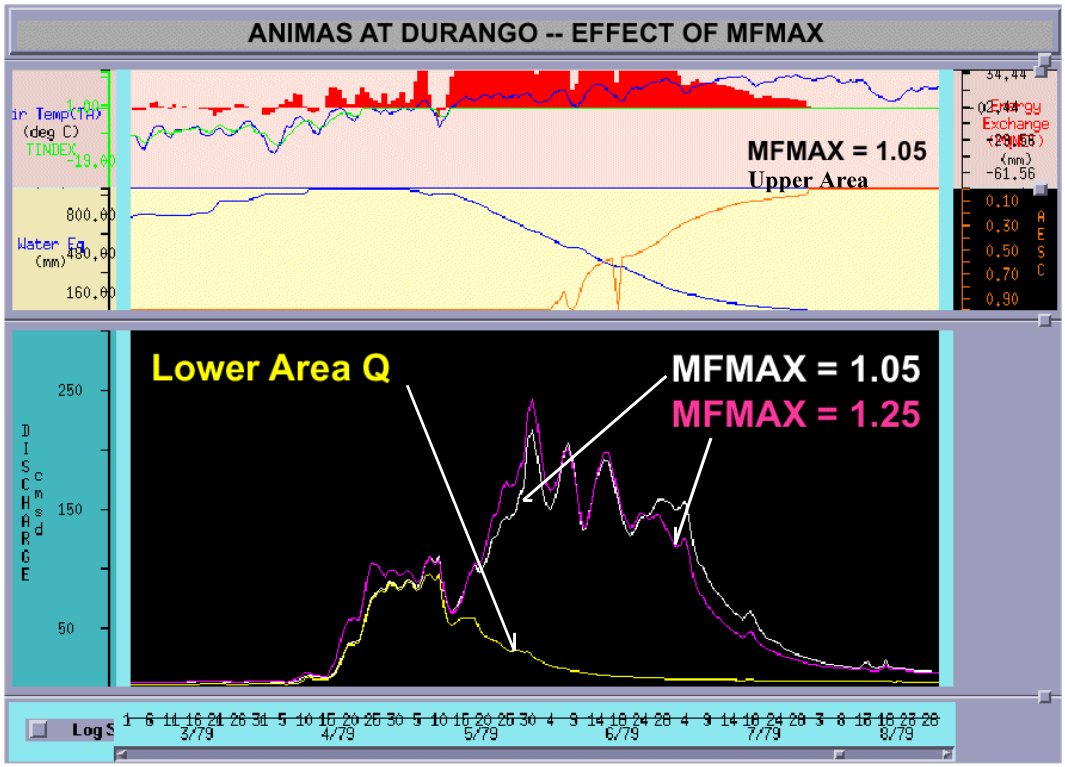


Figure 7-7-8. Effect of changing melt factors for 2 elevation zones during a big snow year.

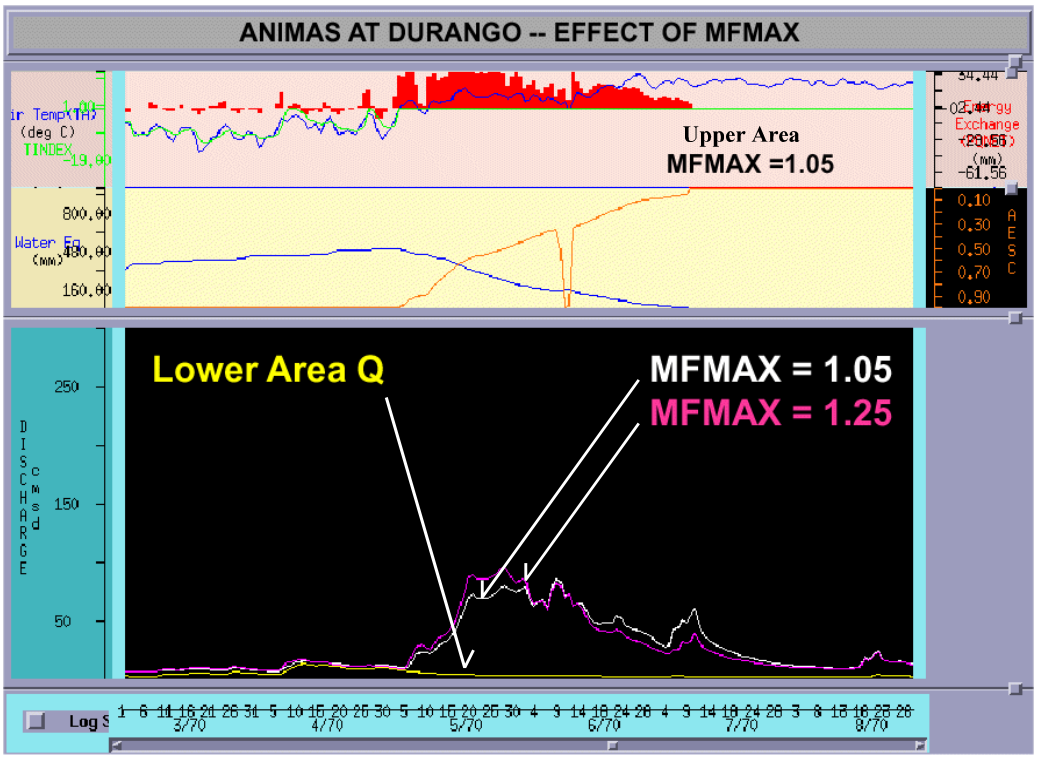
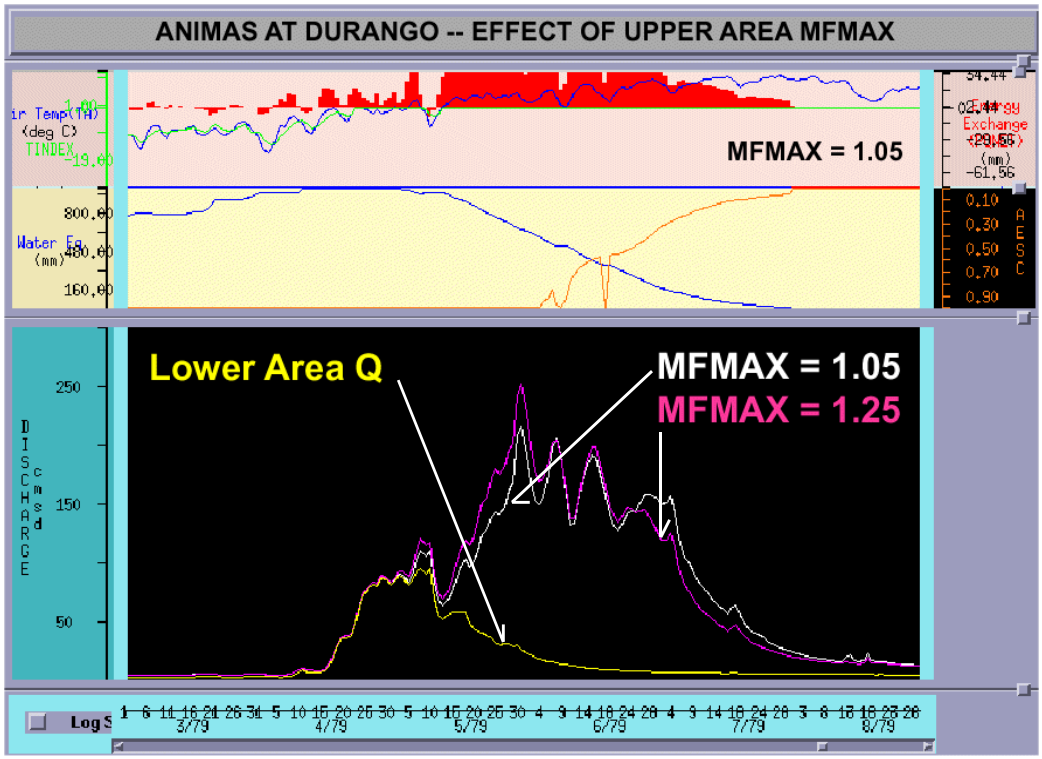


Figure 7-9. Effect of changing MFMAX for 2 elevation zones for a snow year.



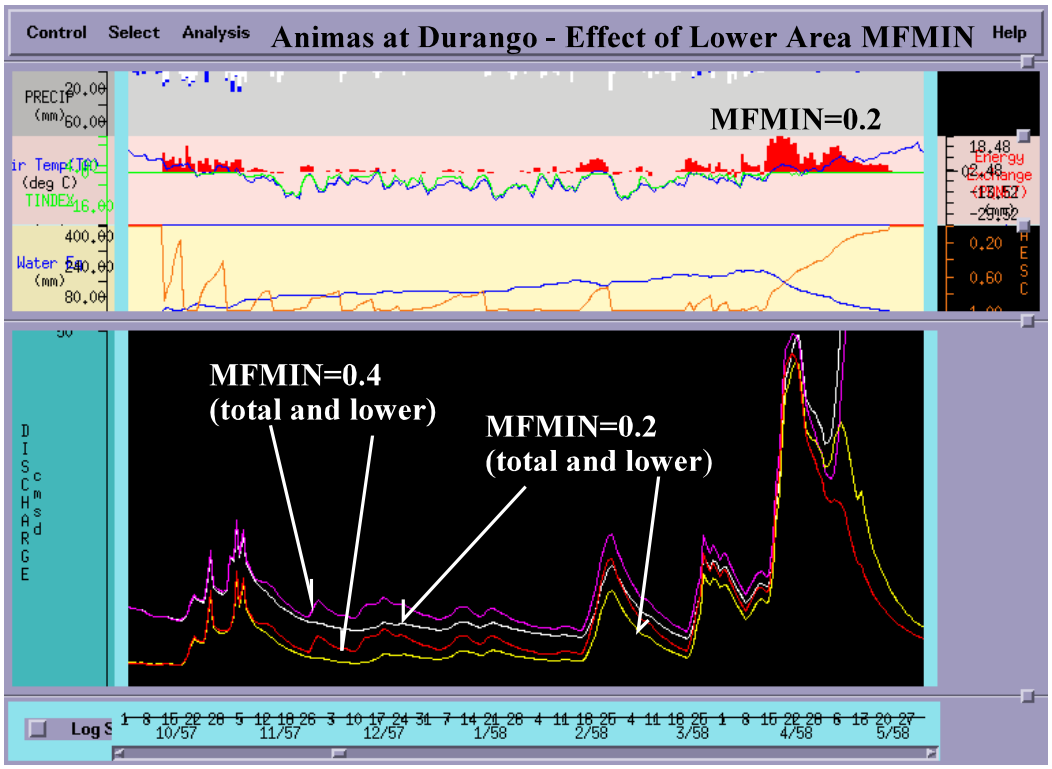


Figure 7-10. Effect of changing MFMIN for the lower elevation zone. MFMAX for the upper elevation zone only.

Figure 7-7-11. Effect of changing MFMIN for the lower elevation zone.

elevation zone. In this case the effect of increasing MFMAX for the lower zone is not noticeable since this zone produces little runoff. For lower elevation zones the effect of changing snow model parameters will only be visible during years when these zones have sufficient snow to produce a significant amount of runoff. Figure 7-7-10 shows the effect of changing only the MFMAX parameter for the upper elevation zone during a large snow year. In this case the timing of the melt from the upper zone is affected which again underscores

the importance of knowing where the runoff is coming from when multiple zones are being used. Figure 7-7-11 shows the magnified effect of changing the MFMIN parameter for the lower elevation zone for this same watershed. Increasing the MFMIN value causes more melt during mid winter periods when the temperature goes above freezing at the lower elevations. Changing MFMIN for the upper zone wouldn't affect these periods since the temperatures remain below freezing at the higher elevations. In many watersheds in the intermountain west, MFMAX is the most important melt factor parameter since most melt occurs in the spring, however, in some cases there is some winter melt at lower elevations which allows for checking the value of MFMIN for the lower zone. The value of MFMIN for the upper zone is fairly insensitive for these watersheds.

To repeat, the unique characteristic of the non-rain melt factor parameters is that they determine the melt rate when there is 100% snow cover, thus the effects of MFMAX and MFMIN can be isolated once melt has really started and the area is completely, or nearly completely, covered by snow.

SI – The SI parameter determines when and for how long the area remains at 100% areal cover. During small snow years, typically bare ground shows up as soon as significant melt occurs. During large snow years, most areas remain at or near 100% areal cover for some period of time after melt begins. Thus, the effect of SI can generally be isolated by examining the snowmelt patterns during years with different amounts of snow accumulation. In Section 7-4 it was recommended that the initial value of SI be set to a value greater than any mean areal water equivalent that will occur during the period of record. If this is done, the initial run will show bare ground showing up as soon as melt begins for all years. Then one can examine the larger snow years to see if more melt is needed for some period into the snowmelt season. By tabulating the water equivalent values above which more melt is needed, one can determine a reasonable guess at a value of SI to try.

Figure 7-7-12 shows the effect of SI on a large snow year. When SI=999, bare ground shows up as soon as the snow begins to melt. When SI=200, the watershed remains at 100% areal cover until the mean areal water equivalent drops below 200 mm. This causes the entire area to contribute melt until the water equivalent (WE) drops below 200 mm. Figure 7-7-13 shows the same parameter combination for a small snow year. In this case there is almost no change in the response since the mean areal water equivalent barely exceeds 200 mm and thus, in both cases bare ground shows up soon after melt begins.

Figure 7-7-14 shows the effect of SI for the upper zone on a watershed with 2 elevation zones during a large snow year. When using a value of SI=600 mm, the melt continues over 100% of the area until the water equivalent reaches the SI value near the middle of June. When SI=999, the areal cover begins to deplete as soon as outflow is initiated from the snow cover in the upper zone in early May. This causes the mean areal melt to be less than what would occur if the zone had remained completely covered. Figure 7-7-15 shows the effect of SI for the upper zone during a small snow year. In this case there is no difference in the response

since the mean areal water equivalent never exceeds 600 mm. Figure 7-7-16 shows the effect of using a value of SI that causes both zones to remain at 100% areal cover for a significant time after melt begins during a large snow year. This increases the snowmelt contribution from both zones during this period.

To repeat, the effect of the SI parameter can be isolated by examining the early part of the melt season, i.e. the same period as used to isolate MFMAX and MFMIN, only in the case of SI one is looking for differences in response between large and small snow years. SI allows for 100% of the area to contribute for some period during years with a large snow accumulation, but not during small snow years.

Areal Depletion Curve – The areal depletion curve controls the amount of snowmelt once bare ground begins to appear. Thus, it is the timing of snowmelt whenever the mean areal water equivalent is less than SI that is examined to determine if the shape of the areal depletion curve should be altered. The best way to determine how the curve should be changed is to plot the areal depletion curve and then go through each of the years in the

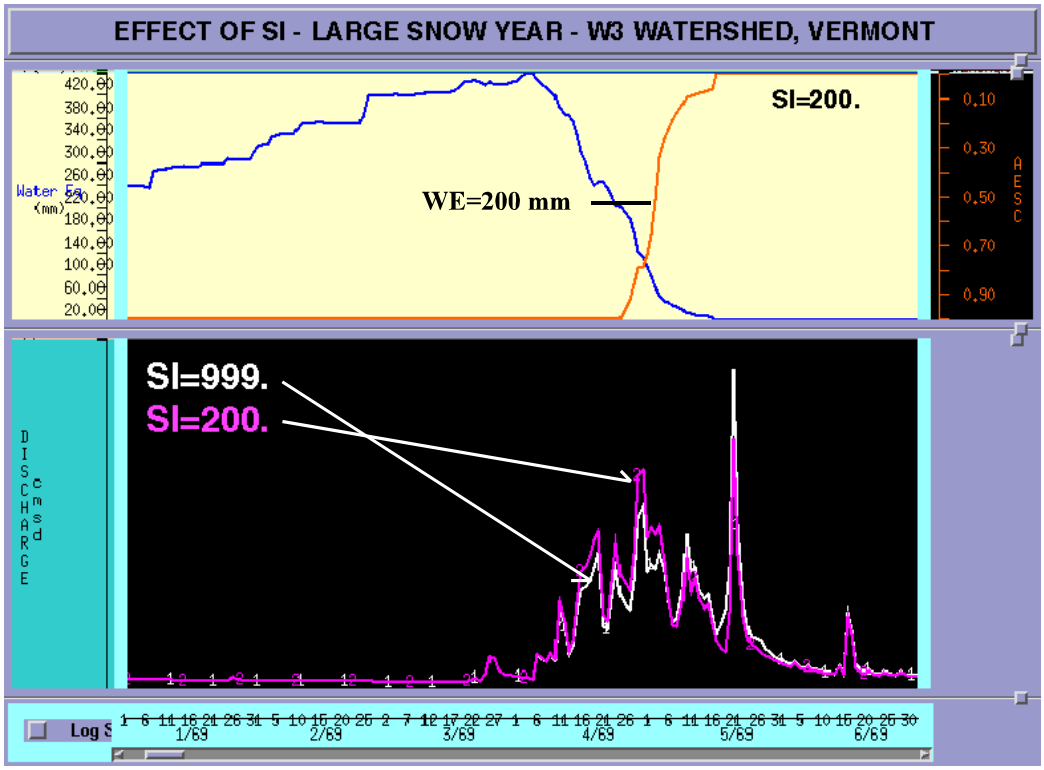


Figure 7-12. Effect of SI on large snow year single zone for a watershed.

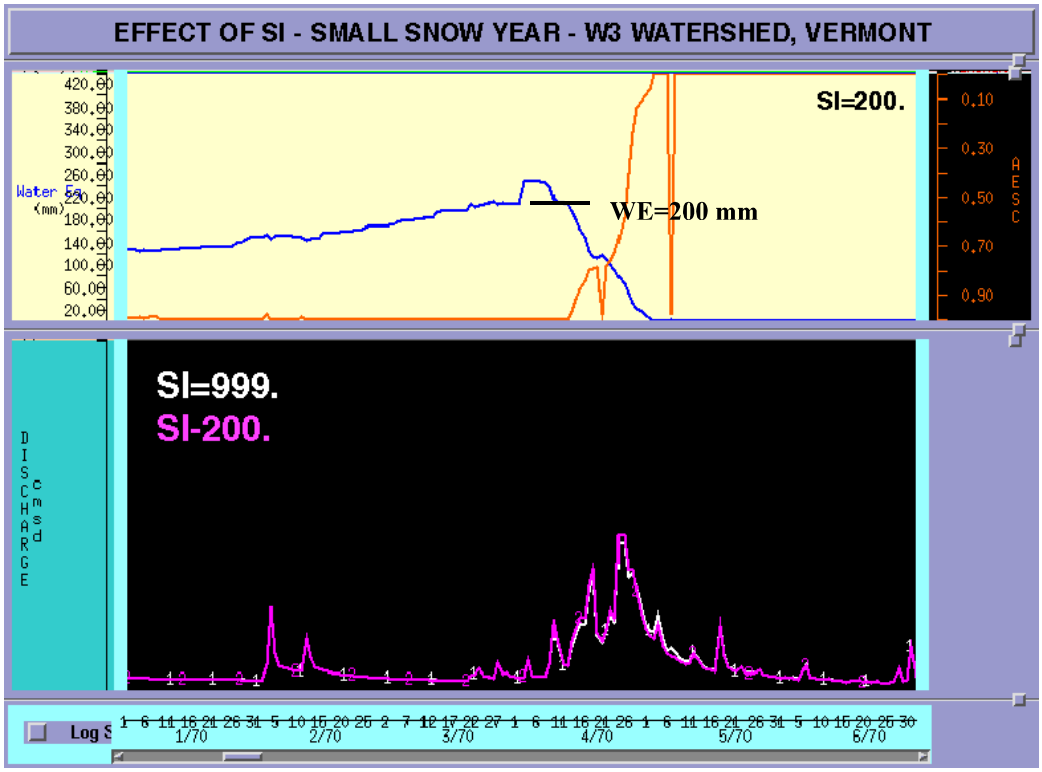


Figure 7-7-13. Effect of SI on a small snow year for a single zone watershed.

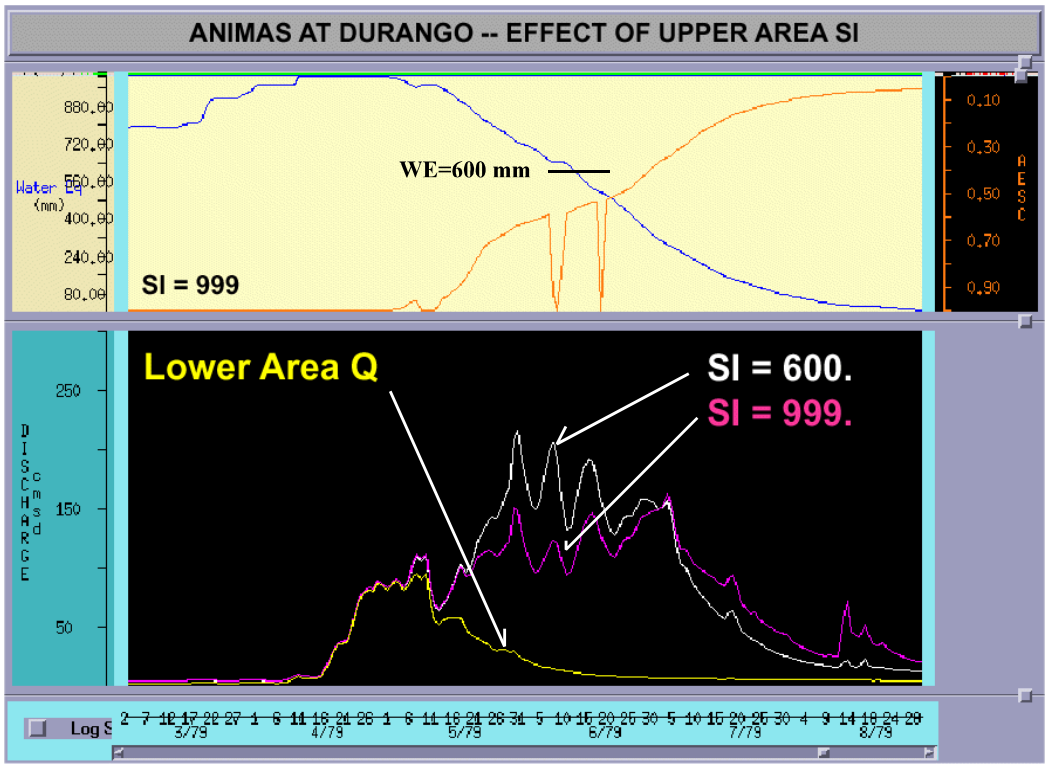


Figure 7-14. Effect of changing SI upper zone for a large snow year.

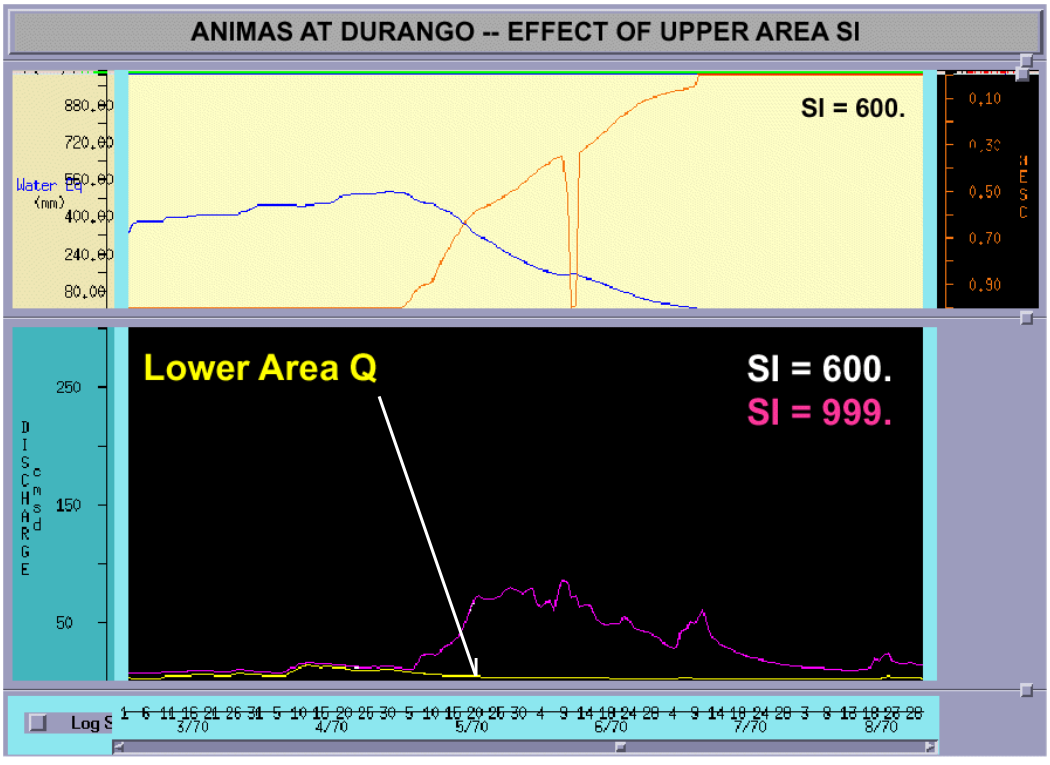


Figure 7-7-15. Effect of changing SI for a upper zone for a small snow year.

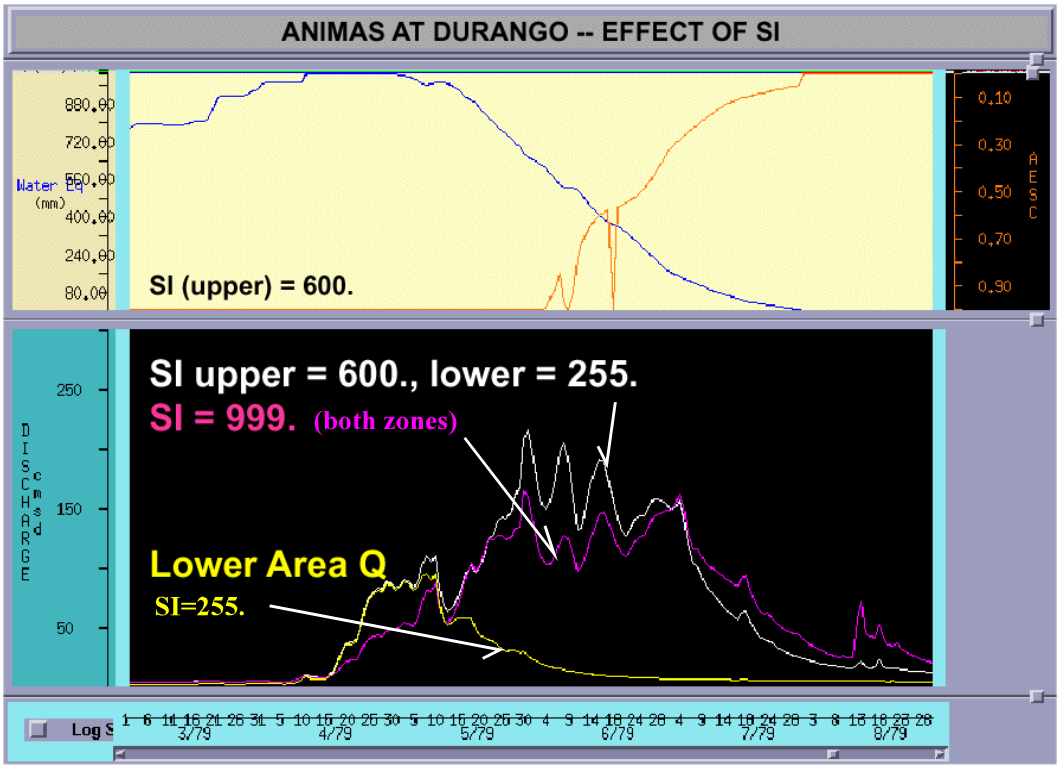


Figure 7-7-16. Effect of changing SI for both elevation zones during a large snow year.

calibration period and note what portion of the curve is being used at different times when some bare ground exists. If more melt is needed at a given time, it suggests that the areal cover should be increased for that WE/A_i ratio and vice versa. Hopefully a pattern will develop that indicates how the overall shape of the curve should be changed. In the following examples the sample initial value curves, curves B and C as shown on Figure 7-4-4, are used to illustrate the effect of changing the shape of the areal depletion curve.

Figure 7-7-17 shows the effect of changing the depletion curve for a watershed with no elevation zones. There is no difference in the hydrographs prior to when bare ground first begins to show. After that the 'B' curve melts off the snow quicker because the areal extent of the snow cover is significantly greater than for the 'C' curve for WE/A_i ratios above 0.4.

Figure 7-7-18 illustrates the effect of changing the areal depletion curve for the upper zone of a mountain watershed during a large snow year. Again the hydrographs are the same until bare ground begins to show up in the upper area. Again the 'B' curve melts off the snow quicker at first than the 'C' curve, however, due to the shape of the 'B' curve when only a little snow remains, there is still some snow left in this case in mid August to augment the summer rains that occur then. Figure 7-7-19 illustrates changing the areal depletion curve for both elevation zones for the same large snow year. Both zones remain at 100% cover for some period after melt begins. Changes to the depletion curve for the lower

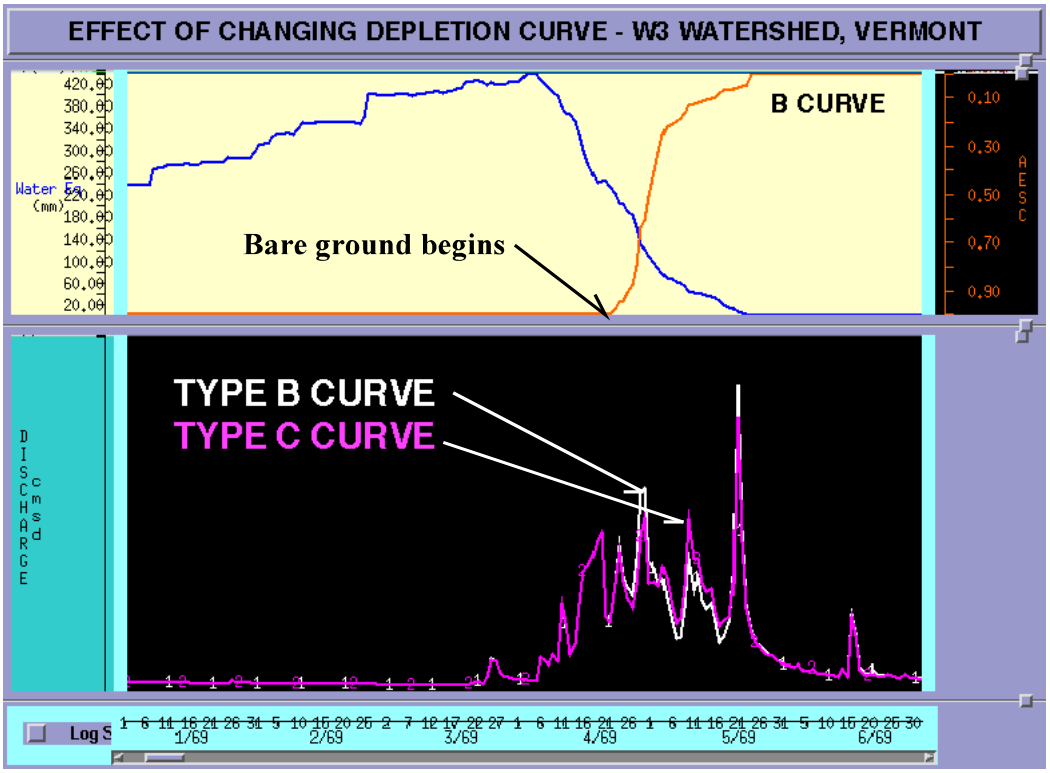


Figure 7-7-17. Effect of different depletion curves on a watershed with single zone.

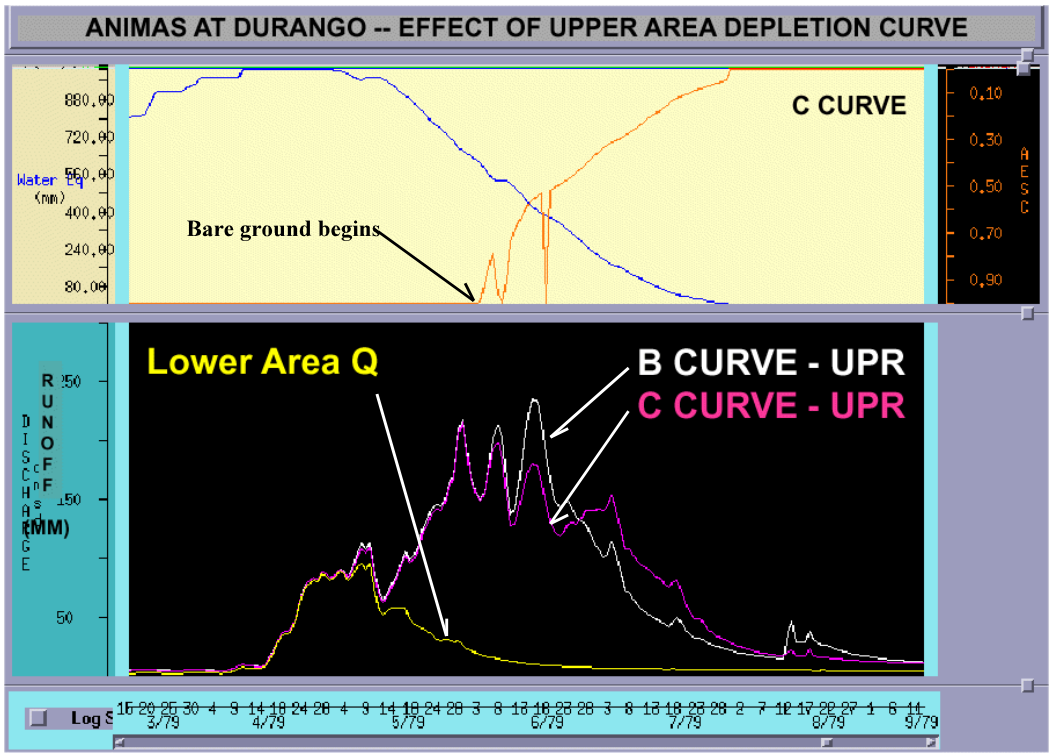


Figure 7-7-18. Effect of changing depletion curve for an upper elevation zone.

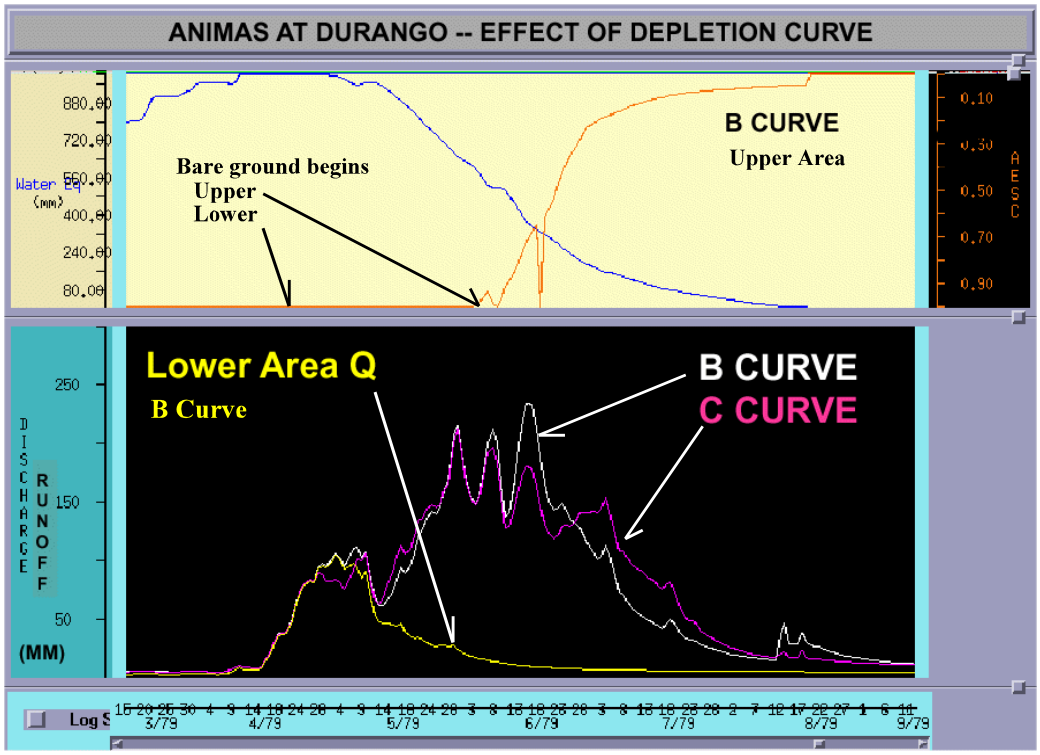
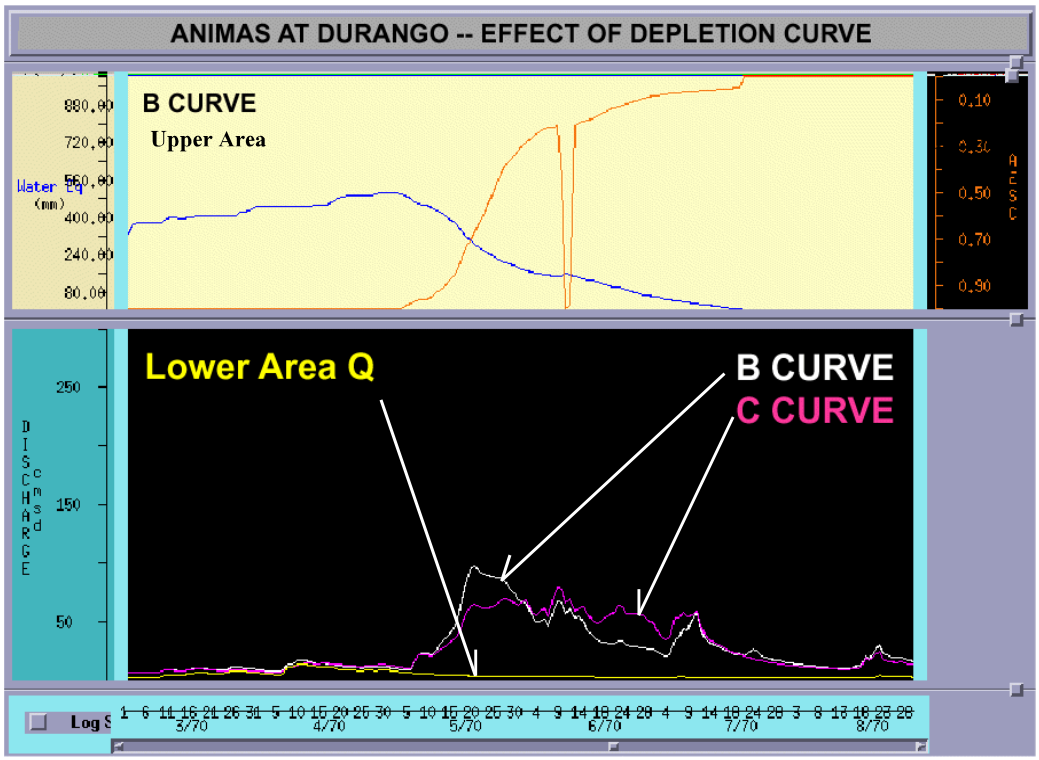


Figure 7-7-19. Effect of changing depletion curve for both elevation zones.



Figur e 7-

7-20. Effect of changing depletion curve for both zones for a small snow year.

zone persist long enough so that the portion of the hydrograph when the upper area is completely covered is affected. Figure 7-7-20 shows the effect of changing the depletion curve for this same watershed for a small snow year. During this year bare ground begins to show up as soon as active melt begins since the water equivalent for neither area exceeds SI, thus the entire melt season is affected. Only the upper zone shows much effect to the depletion curve changes because very little runoff is generated from the lower area.

UADJ – The rain-on-snow melt equation is only used when the rainfall rate exceeds 2.5 mm in a 6 hour period. Thus, to isolate the effect of the UADJ parameter, the rainfall rate must exceed that value. Also to have a significant amount of melt produced during rain periods, the air temperature must be well above freezing. If the temperature during rain is just barely above freezing, very little melt will be generated. It should be emphasized that the amount of rain is not a factor in isolating the effect of UADJ, but only that sufficient rain occurs to use the rain-on-snow melt equation when the air temperature is well above freezing.

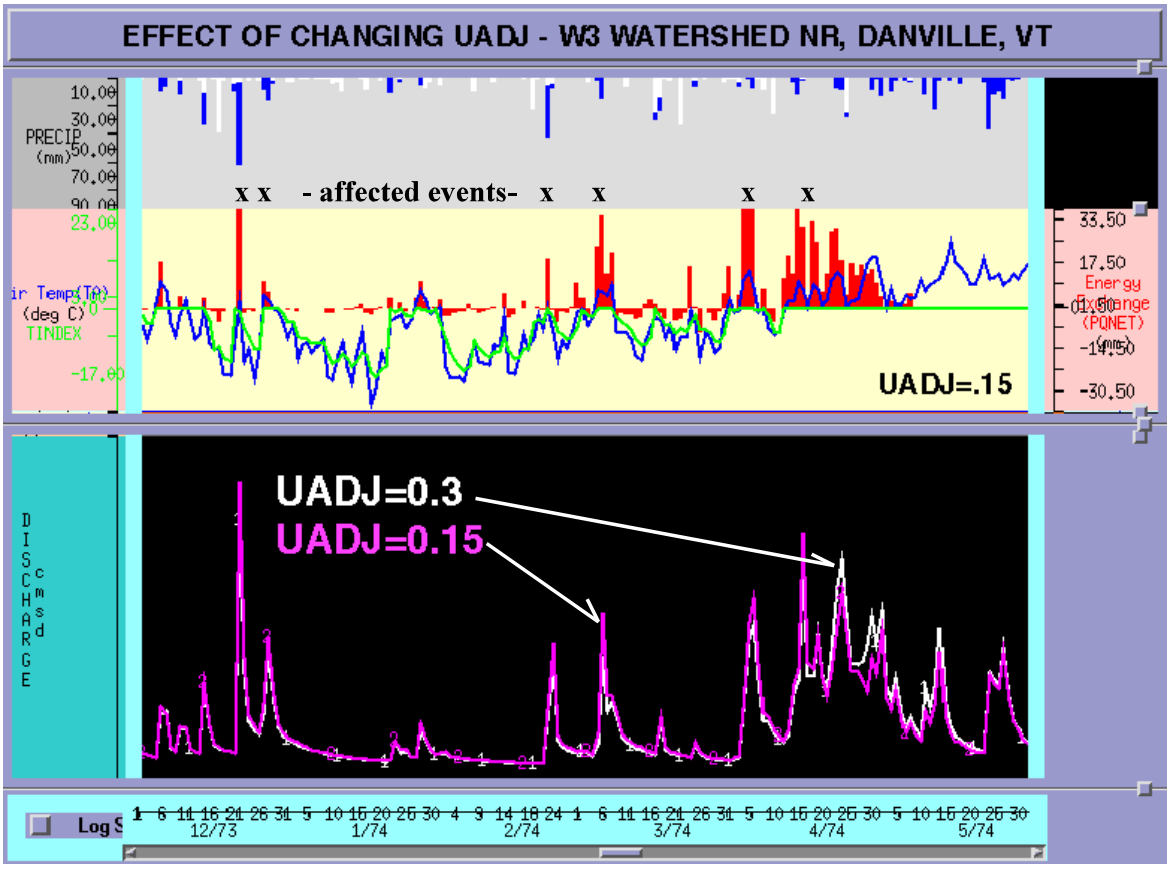
Figure 7-7-21 shows the effect of UADJ during a year with a number of rain-on-snow events. Increasing the value of UADJ resulted in more melt during many of the rain events as noted on the figure. Since these events were mostly in the winter, the additional snow

Figure 7-7-21. Effect of changing UADJ.

melted with UADJ=0.15 resulted in less snow being available for the main melt season in April. For a few of the rain-on-snow events there was little difference in the amount of melt computed because the air temperature was barely above freezing. These were the 3 events occurring between the 6th and 16th of December and the 2 events in late January.

Table 7-7-1 summarizes the primary effect of each of the major snow model parameters. It is critical to understand the primary effect of each of these parameters in order to isolate the periods to examine to determine if changes are needed to their values.

Parameter	Primary Effect
SCF	Controls volume of snowmelt runoff
MFMAX, MFMIN	Controls melt when at or near 100% areal cover
SI	Controls when 100% cover exists during large snow years



Depletion curve	Controls melt when areal cover less than 100%
UADJ	Controls melt when rain falls and the temperature is warm

Table 7-7-1. Summary of primary effect of major snow model parameters.

Discrepancies at the Beginning of the Snowmelt Season

Various parameters and conditions can affect when the onset of snowmelt runoff occurs after a long accumulation period. This section attempts to discuss those items that should be examined when snowmelt runoff starts too early or too late. The things to consider include:

• Snow Cover Heat Deficit – If the size of the heat deficit is too large or too small prior to a snowmelt period, this can result in the onset of runoff to occur a few days late or early. Typically the size of the heat deficit is only sufficient to cause the onset of runoff to be off by a period of days and not weeks. A significant heat deficit will only exist prior to melt when there has been an extended period of very cold weather followed by a quick warm up to melt conditions. A significant heat deficit cannot develop when temperatures are not well below freezing or when the temperature warms up slowly, even though remaining below freezing, after an extremely cold period. Also a large heat deficit cannot build up in a shallow snow cover because there is not enough heat storage capacity. The size of the computed heat deficit is controlled by the TIPM and NMF parameters. To develop a significant heat deficit, the value of TIPM should be fairly low, typically 0.05 or 0.1, in order to maintain a large gradient between the air temperature (assumed snow surface temperature) and the antecedent temperature index (TINDEX, indicates the temperature at some distance within the snow cover) during a long cold period. When the temperature at the snow surface is considerably less than the internal snow temperature, then the larger the value of NMF, the faster the heat deficit will grow. Thus, if the size of the heat deficit is thought to be the cause of problems at the onset of snowmelt runoff, then one should try changing the TIPM and NMF parameters. It should also be recognized that even though small TIPM values and large NMF values will cause a significant heat deficit to develop during an extremely cold period, this same combination will cause the heat deficit to more rapidly diminish during a warm up. Thus, in order to have a large heat deficit that will affect melt computations and the timing of runoff, there must be a rapid transition from an extremely cold period to melt conditions.

Liquid Water Storage – The amount of liquid water that goes into storage within the snow cover when significant melt first occurs can have an effect on the onset of snowmelt runoff. If too much of the melt is held within the snow cover as liquid water, the runoff will be delayed and vice versa. For regions with a relatively deep snow cover during most winters, it is physically unrealistic to use a liquid water holding capacity for ripe snow (PLWHC parameter) outside the range of 0.02 to 0.05. In these regions liquid water storage is seldom

the cause of problems related to the onset of snowmelt runoff. However, in regions with generally shallow snow cover, especially in the plains and north central regions, the overall amount of liquid water that can be held in the snow can be considerably greater than 5% due to the slush layer that typically builds up at the snow-soil interface. In these regions a PLWHC value of 0.15 to 0.30 is quite common and the PLWHC parameter is a likely candidate for correcting problems at the onset of runoff from melt.

Ice Blocking the Channel System – In some regions, especially portions of the north central area and parts of Alaska, ice may build up and block portions of the channel system, especially drainage outlets and culverts. This results in the initial melt water ponding in fields and low lying areas until the ice melts and frees the outlet. By neglecting this situation, computed snowmelt runoff appears sooner than what is observed. Currently there is no explicit way to model this effect. A separate procedure or modification to a channel response or routing model would be the most physically realistic way to address the problem. The use of a larger than normal PLWHC may implicitly account for some portion of this condition.

Soil Moisture Conditions – Soil moisture conditions can have a significant influence on runoff at the beginning of a snowmelt period in many regions. Two soil moisture model situations typically can cause problems during such periods. First, the magnitude of the tension water deficits when melt begins will have a large effect on the timing and the amount of the initial runoff. In some regions it is common during years with a dry summer and fall for large tension water deficits to still exist when melt begins. During other years the deficits will be full going into the winter and remain full until melt begins. If the tension water deficits are too large when melt starts, generally the onset of runoff will be delayed longer than it should and vice versa. Thus, it is important to check the size of the tension water deficits prior to snowmelt especially if problems exist with the onset of runoff from the melt.

Percolation problems can also affect the timing of runoff at the onset of snowmelt though these problems typically continue well into the melt season. If the percolation rate is too large when melt begins, too much water will initially go into lower zone storages, while if the rate is too low, the water will run off too quickly. Generally if a reasonable baseflow simulation is obtained prior to working with the snow model parameters this will not be a significant problem though if the lower zone deficiency ratio in the Sacramento model moves over a large range during the melt season, there could be a problem with the percolation rate for the drier conditions that exist when melt begins.

Temperature Bias – If snowmelt occurs well before or after the onset of observed runoff (typically a week or two) and none of the previously mention situations will explain the problem, it is very likely that the temperature values being used are too high or too low for the area they are representing. There can be two reasons for this situation. First, in some cases, especially in lower elevation zones of mountain watersheds, the elevation of the MAT

time series is generally not the same as the mean elevation of the snow cover. The elevation of the MAT time series typically is selected as the mean elevation of the area being modeled, whereas in lower elevation zones, the snow may predominately be in the upper portion of the area. Thus, the mean elevation of the snow covered area is greater than the mean elevation of the entire zone. This situation can also occur during very low snow years in the upper elevation zone as described in Section 6-1. The solution to this problem is to use the lapse rates in the snow model to adjust the MAT time series values to the proper elevation. This is done by setting the mean elevation of the area to the mean elevation of the snow covered portion of the zone. Then the lapse rates will be used to adjust the MAT values by the elevation difference between that defined for the temperature data and that specified for the zone.

The second reason for the temperature values being off is that the MAT computations are producing biased values, typically due to improper monthly temperature-elevation relationships being used to extrapolate observed station data to the mean elevation of the area being modeled. Generally the problem involves extrapolating low elevation observations to higher elevations zones where most of the snow and runoff occur. In a few cases the problem is caused by an input error to the MAT program that causes the wrong stations to be weighted for one or more zones thus producing a biased temperature estimate. In order to see if adjustments to the temperature data will correct the problem, an elevation difference and lapse rate can be used to adjust the values or a non 0°C MBASE parameter value can be tried, however, once it is determined that the temperature data are biased, the MAT time series should be regenerated using the proper temperature-elevation relationships and station weights.

Section 7-8

Effect of Sacramento Model Parameters on Model Response

Introduction

It is critical to understand the most important function of each of the model parameters when performing interactive trial and error calibration. It is also essential to know what to examine to find the periods when the effect of each parameter can be isolated. This understanding allows the person doing calibration to make educated decisions regarding which parameters should be altered as opposed to random guesses. Without this knowledge, interactive calibration becomes an exercise in futility.

This section describes the most important function of each of the Sacramento model parameters and how to determine which periods to examine to isolate the effects of each parameter on model response. Besides their major function, many of the Sacramento model parameters also influence other aspects of the model computations. This is especially true with all the parameters that are involved in the percolation computations. The percolation equation in the model involves 8 parameters, ZPERC, REXP, LZFSM, LZFPM, LZSK, LZPK, LZTWM, and UZFWM. Two of these parameters are unique to the percolation computations, ZPERC and REXP, while all of the others are involved in other aspects of the model's algorithms. Thus, while these other parameters have a major function of their own, they also affect the percolation computations. Because of this, when the values of these parameters are changed, there are multiple effects. The user needs to anticipate these effects, but should concentrate on the major function of each parameter when performing interactive calibration.

The major function of each Sacramento model parameter is illustrated in this section by using mean daily flow hydrograph plots showing the response of two different values of the parameter for a period when the effects of the parameter can be isolated. The plots are generated by running the model with everything exactly the same except for the values of the parameter being considered. Under such conditions the unique effect of each parameter can clearly be shown. The actual problem of deciding on which parameters to change becomes much more difficult during calibration because of noise in the data and the fact that generally multiple parameters, perhaps involving several models, need to be adjusted. To have any hope of making the proper changes during calibration, one must know what periods to examine and what to look for to isolate the effects of each parameter. It is also helpful to have a strategy for checking parameters for possible changes. Such a strategy for the Sacramento model is outlined in Section 7-1.

Most of the parameters are discussed individually, however, the percolation curve and the parameters that primarily affect it are lumped together. When making changes to the percolation rate it is important to think in terms of the entire curve as opposed to the individual parameters that influence the curve. In addition to the hydrograph plots, panels that show how the model states are changing and other results of the model computations are included in each of the

figures. These panels are extremely helpful in determining when the conditions occur that are needed to isolate the effects of each parameter. Even though two hydrograph plots are included on each figure, the panels showing the model states and computations can only be included for one of the parameter values. This value is specified next to the panels on each figure.

Explanation of Panels Showing Model States and Computations

Figure 7-8-1 illustrates the panels which are part of the ICP WY-PLOT display for the SAC-SMA operation. The plots on these panels are for a daily time interval. Instantaneous values such as zone contents are for the end of each day.

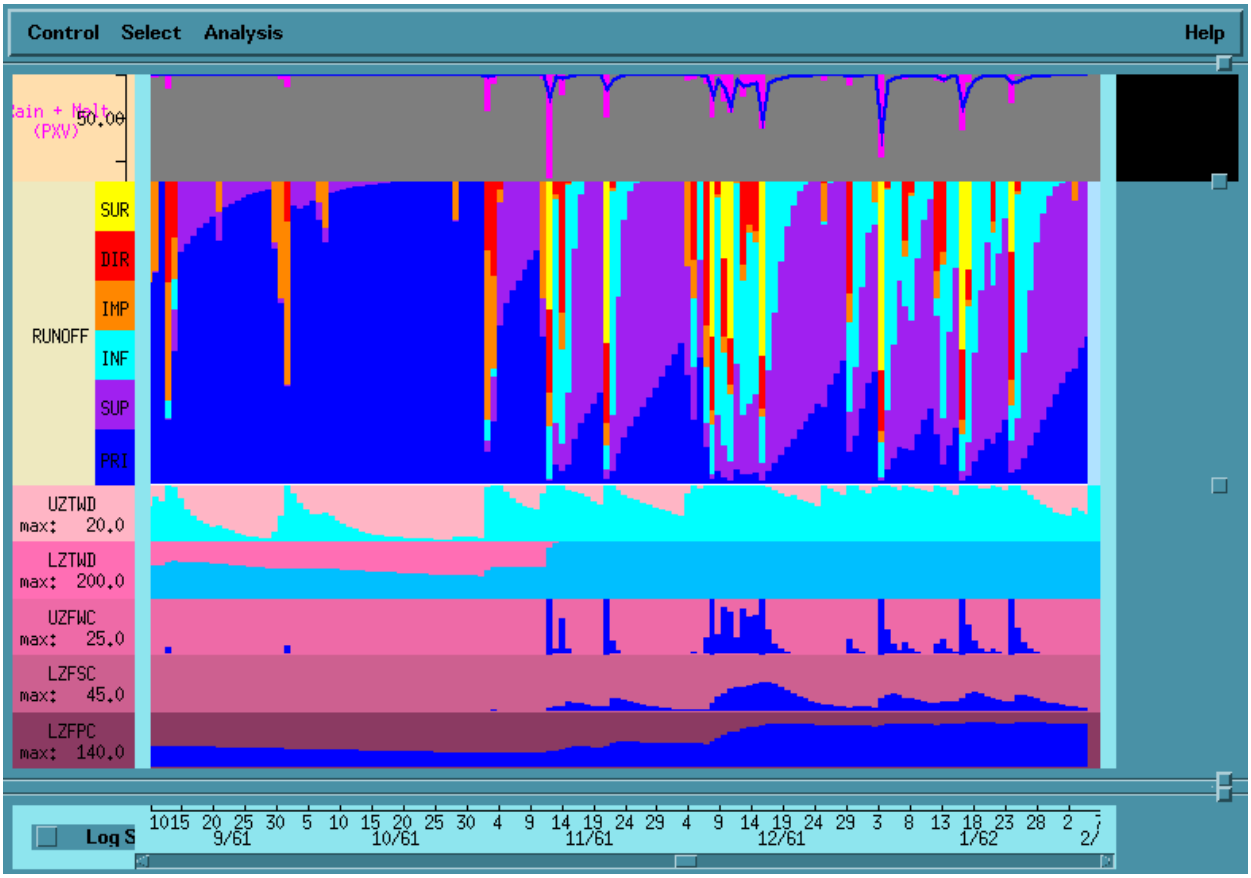


Figure 7-8-1. Sample ICP display for the SAC-SMA operation.

Top Panel - This panel shows the total amount of water coming into the Sacramento model for each day as a bar value in magenta with the scale on the left. These can be rain+melt values from the snow model or the direct input of rainfall when snow computations are not included. The dark blue line in this panel is the total amount of channel inflow, i.e. runoff, generated by the model for each day. This is the sum of all the runoff components. The scale is the same as for the precipitation input.

Middle Panel - This panel shows a breakdown of the runoff into components for each day. The total height of the panel is divided in proportion to the fraction of each runoff component that occurs on any day. Surface runoff is displayed in yellow, direct runoff (i.e. from the additional impervious area) in red, constant impervious runoff in orange, interflow in light blue, supplemental baseflow in purple, and primary baseflow in dark blue. This panel allows the user to see relatively how much runoff is being generated by each of the components on a daily basis.

Bottom Panel - This panel shows the states of each of the upper and lower zone moisture storages at the end of each day. The capacities of each zone are labeled on the left. The blueish tones represent contents and the pinkish tones deficits (i.e. capacity minus contents). If for a given day a zone is all blue, it indicates that the zone is completely full of water. Vice versa if the a zone is all pink, it indicates that the zone is empty. Generally when viewing this panel, the tension water storages are being thought of in terms of deficits, i.e. how much water is needed to fill the storage, while the free water storages are being thought of in terms of contents, i.e. how much water is available for runoff, or in the case of the upper zone free water storage, both for interflow and percolation.

Sacramento Model Parameters

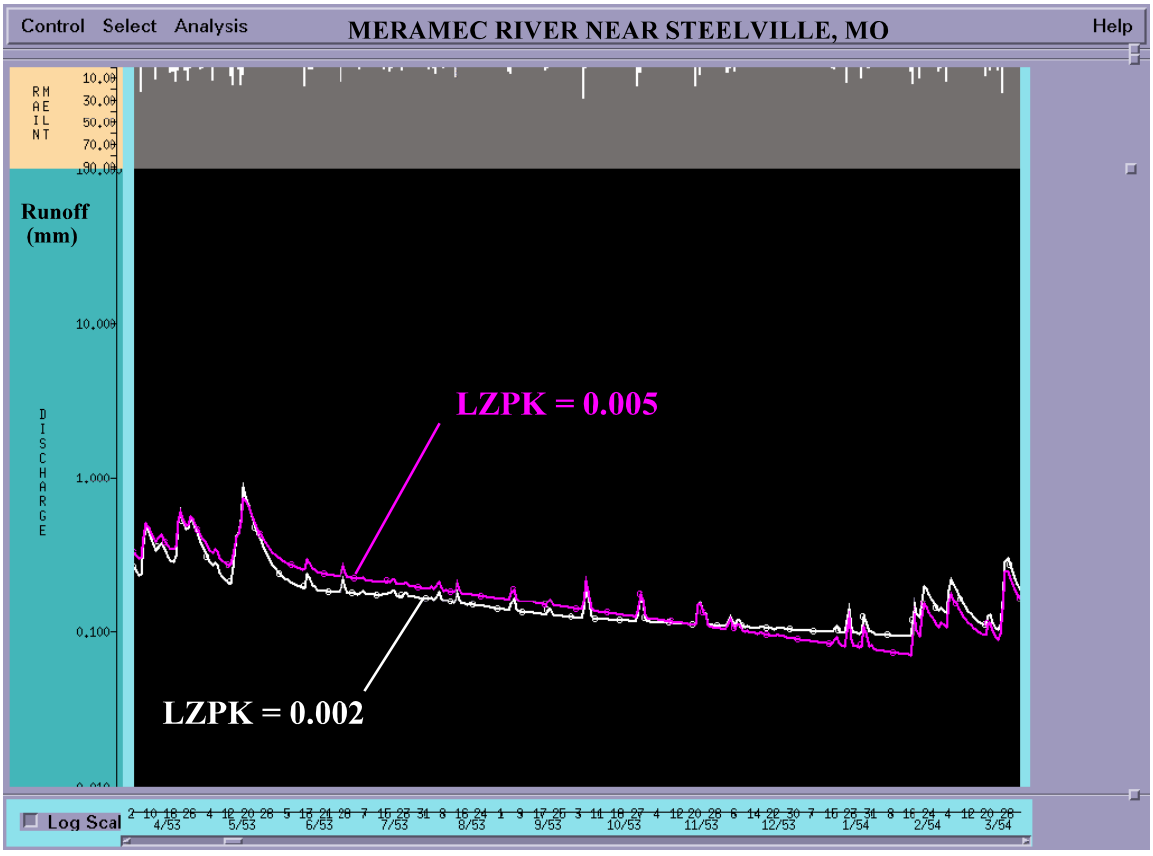
On the remaining pages in this section the major function of each parameter is described, as well as how to isolate the effect of the parameter. The order that the parameters are discussed is based on the strategy for examining the parameters during a calibration as suggested in Section 7-1. The parameters that primarily affect baseflow are described first, followed by tension water storages and PCTIM, storm runoff parameters, and lastly those that primarily affect evapotranspiration. Table 7-8-1, at the very end of the section, summarizes the primary function of each of the Sacramento model parameters.

Besides this material, the portions of Section 7-5 that describe how to derive estimates of parameter values from an analysis of the hydrograph provide considerable information that should assist in determining how to isolate the effect of each model parameter. That material should be reviewed even if one is not going to derive initial parameter values.

LZPK

The primary function of the LZPK parameter is to specify the slope of the primary baseflow recession. LZPK is not to be used to correct for problems with the magnitude of primary baseflow. As long as the simulated slope of the recession during periods when only primary baseflow is contributing to runoff is parallel to the observed recession, LZPK shouldn't be modified even if though the amount of primary baseflow is considerably high or low. Figure 7-8-2 illustrates how changing the value of LZPK alters the slope of the recession. For this illustration the percolation term PBASE was kept the same for both values of LZPK.

The value of LZPK is checked by examining periods when primary baseflow is basically the only source of runoff. Small amounts of constant impervious runoff may occur during such periods from rain that is not sufficient to fill the upper zone tension water storage, but no other runoff components should contribute significant amounts. Complications that can affect periods when primary baseflow dominates are discussed under the 'Assigning Runoff Components' portion of Section 7-5. Periods with riparian vegetation evaporation should be avoided when determining whether the value of LZPK should be changed. Also, during periods with a snow cover it should be recognized that ground melt may cause the primary recession to be slower than during snow free periods. Thus, it is best to use snow free periods whenever possible.



Figu

re 7-8-2. Effect of changing LZPK on model response.

LZSK

The primary function of LZSK is the timing of supplemental baseflow runoff. This can be evaluated in terms of the slope of the supplemental recession or in terms of how long significant supplemental baseflow exists after the storage is recharged. The longer that supplemental baseflow exists, the slower the withdrawal rate. LZSK shouldn't be used in an attempt to change the amount of supplemental baseflow, but only the timing. The periods to use for determining if the value of LZSK should be changed are recessions after significant groundwater recharge has occurred. There should be no major runoff producing events during the recession in order to best evaluate LZSK. Such an ideal case may be difficult to find for watersheds with a very slow supplemental recession rate (see Section 7-5 for further information).

Figure 7-8-3 illustrates the how changing the value of LZSK alters the timing of the supplemental recession. The value of PBASE in the percolation equation is kept the same for both values of LZSK. The panel at the top is for LZSK=0.20. It shows for the October 1964 recession period that the supplemental baseflow contribution is essential gone by the end of the month. For the LZSK value of 0.10, supplemental baseflow lasts until the middle of November.

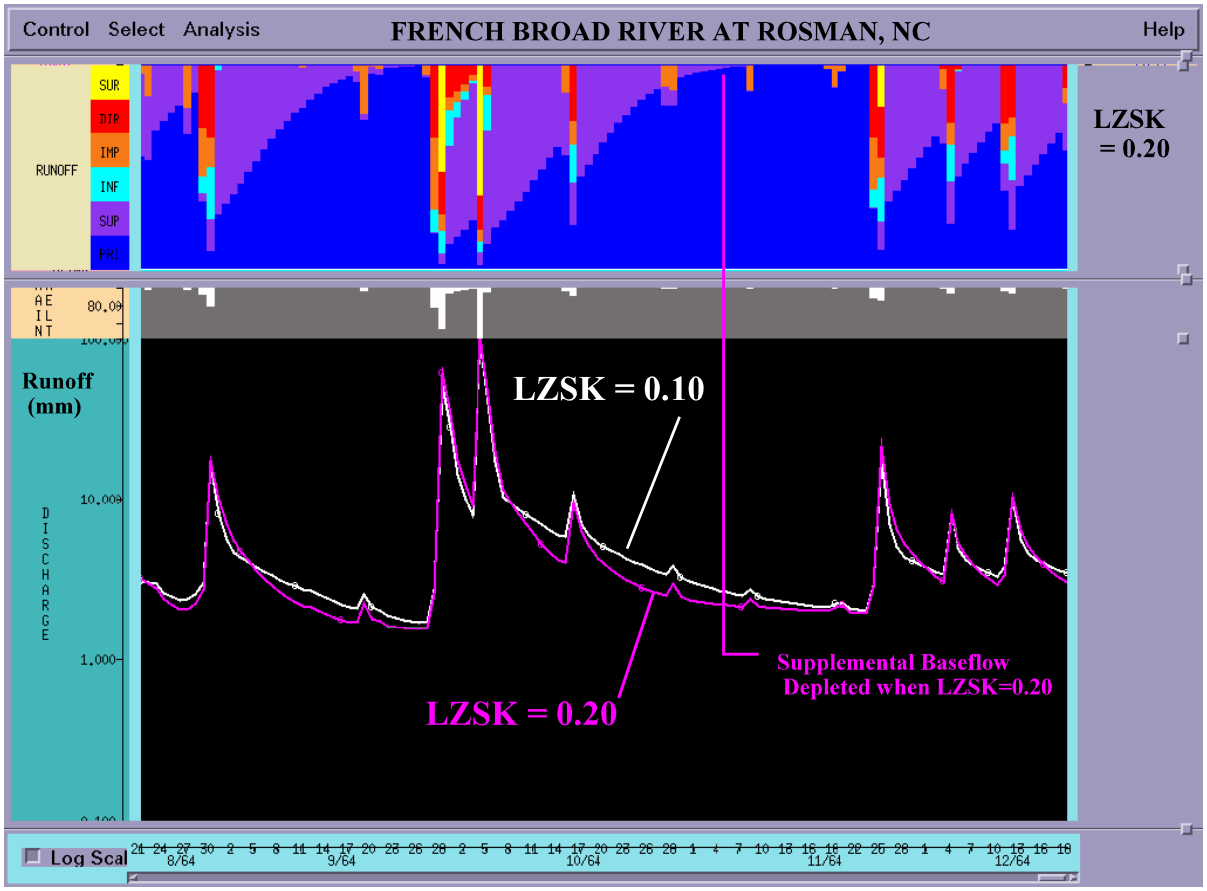


Figure 7-8-3. Effect of changing LZSK on model response.

LZFPM

The primary function of LZFPM is to control the amount of primary baseflow runoff. Whereas LZPK controls the withdrawal rate, i.e. the slope of the recession, LZFPM controls the magnitude. The greater the value of LZFPM the more water that can be stored in the primary storage, thus a greater overall contribution from primary baseflow. If the total amount of baseflow from both primary and supplemental contributions is too great or too small, the percolation rate will need to be decreased or increased to correct the problem. Once the total amount of baseflow is reasonable, then LZFPM can be changed so that the proper proportion is assigned to primary baseflow runoff.

The periods to examine for LZFPM are the same as were used to evaluate LZPK, i.e. periods when primary baseflow is the major source of runoff. Figure 7-8-4 illustrates how changing the value of LZFPM alters the magnitude of primary baseflow runoff. When LZFPM is altered, very little change occurs during periods when other runoff components dominate.

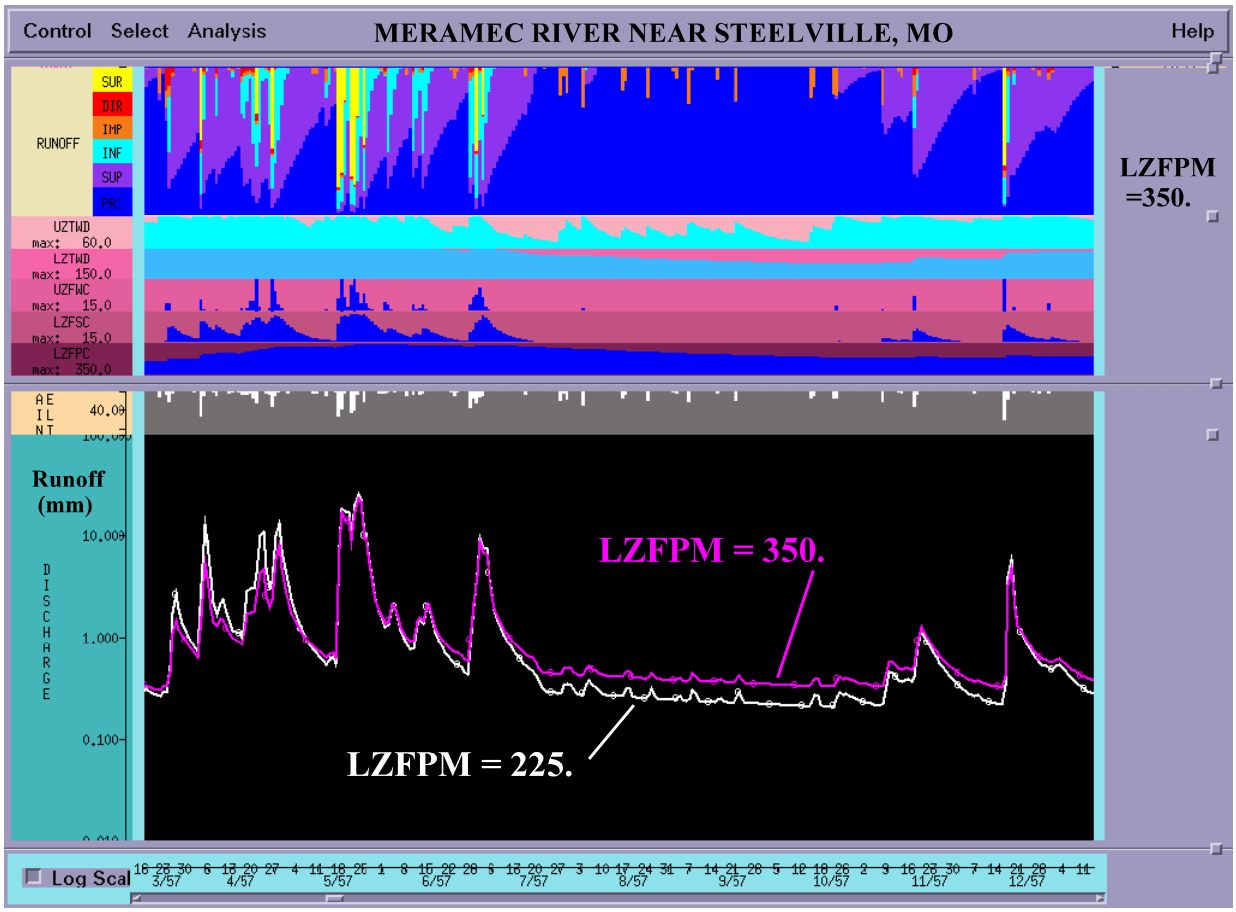


Figure 7-8-4. Effect of changing LZFPM on model response.

LZFSM

The LZFSM parameter is unusual in that it has 2 primary functions. The first, which is described on this page, is to control the amount of supplemental baseflow. The other is its effect on the percolation rate. This function is discussed, along with the role of other parameters on percolation, in the 'Percolation Rate' portion of this section.

Analogous to primary baseflow, LZSK controls the timing of supplemental baseflow and LZFSM controls the magnitude. The greater the value of LZFSM, the more water can go into supplemental storage. The periods to examine to isolate this effect of LZFSM are the same periods used to evaluate the LZSK parameter, i.e. recession periods after significant recharge when there is considerable supplemental baseflow contribution and very little storm runoff. It is especially important to examine recession periods after major events when the maximum amount of supplemental baseflow should occur. Figure 7-8-5 illustrates the effect of changing the value of LZFSM. It can be seen that besides affecting the magnitude of supplemental baseflow, altering LZFSM also changes the response during and immediately after the storm events. This change is due to the effect of LZFSM on the percolation rate. The larger the value of LZFSM, the greater the percolation rate and thus less storm runoff (i.e. surface runoff and interflow).

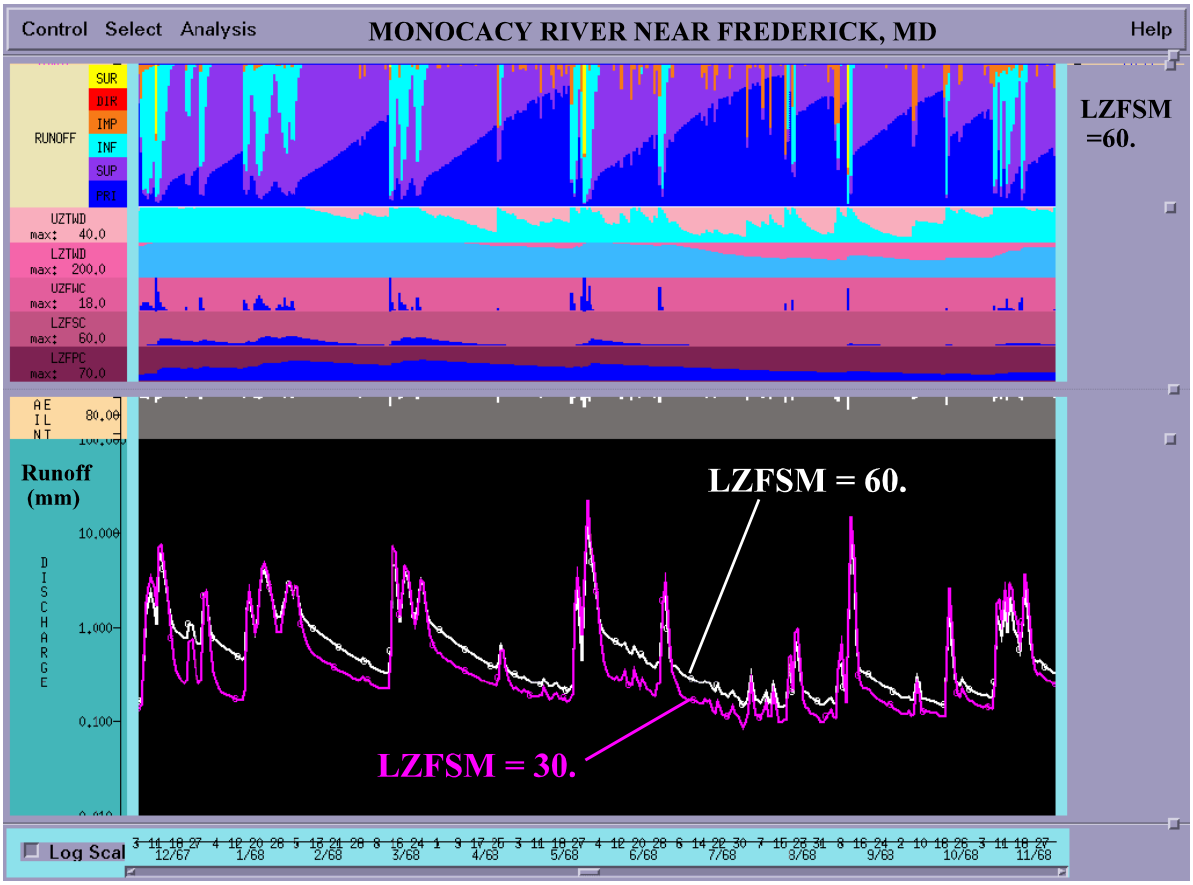
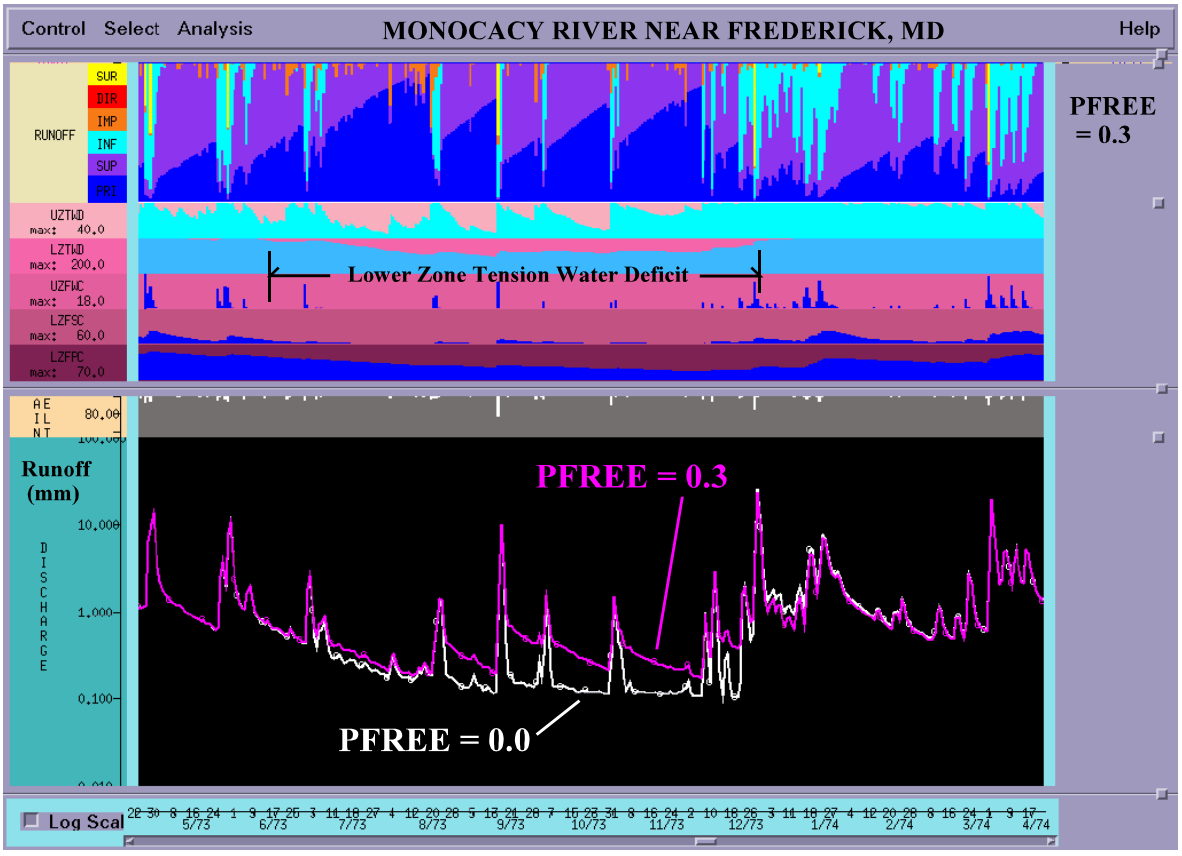


Figure 7-8-5. Effect of changing LZFSM on the magnitude of supplemental baseflow.

PFREE

The function of PFREE is to determine what fraction of the percolation goes directly to free water storages, i.e. how much baseflow recharge occurs, when there is a lower zone tension water deficit, i.e. LZTWC is less than LZTWM. Thus the only periods to examine when evaluating the value of PFREE are those when there is a lower zone tension deficit. Whenever lower zone tension water is full, PFREE has no affect. The amount of recharge is determined by how much the baseflow increases after an event that produces storm runoff (the upper zone tension water must fill in order to have any percolation or storm runoff). When there is a lower zone tension deficit, the relative amount of recharge will likely vary depending on the size of the deficit as in reality PFREE should be function of the dryness of the lower zone as discussed in Section 7-5. Generally it is preferable to use a PFREE value that works well during extended periods with a substantial lower zone tension deficit than to set the value of PFREE based on the shorter transition periods when the deficit first appears or dissipates, though this decision is left up to the user. Figure 7-8-6 illustrates how PFREE affects baseflow recharge after storm events when there is a lower zone tension deficit. Note that PFREE has essentially no effect when the lower zone tension water is full. In very wet regions lower zone deficits seldom occur. In such areas it may not be possible to determine the value of PFREE. Even though PFREE has little effect in such areas, spatial consistency in the parameter value should be maintained.



F

igure 7-8-6. Effect of changing PFREE on model response.

SIDE

The function of the SIDE parameter is to have a portion of the withdrawal from baseflow storages go to deep groundwater recharge rather than appearing in the stream. The SIDE parameter allows the user to get rid of water by removing it from baseflow when storm runoff is being reproduced in a reasonable fashion and increasing the amount of evaporation or decreasing the precipitation seems physically unrealistic. The SIDE parameter should only be used when there is some collaborating evidence that significant deep groundwater recharge is an important part of the water balance.

Figure 7-8-7 illustrates the effect of using a non zero value for SIDE. The same fraction of the withdrawal from lower zone free water storages is removed every day from the water balance. Since the amount of withdrawal varies with the contents of the primary and supplemental storages, the amount of deep recharge is greatest when the amount of water in these storages is the greatest.

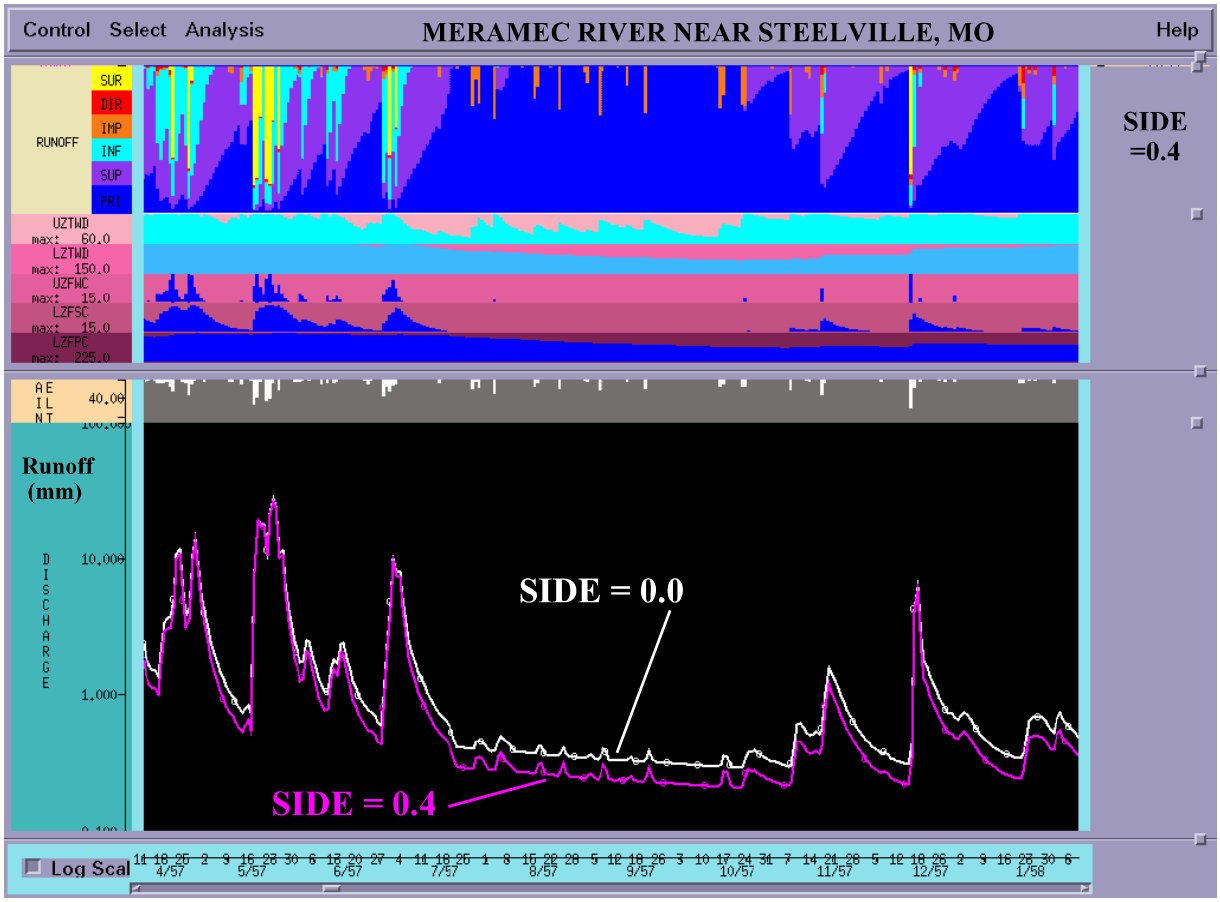
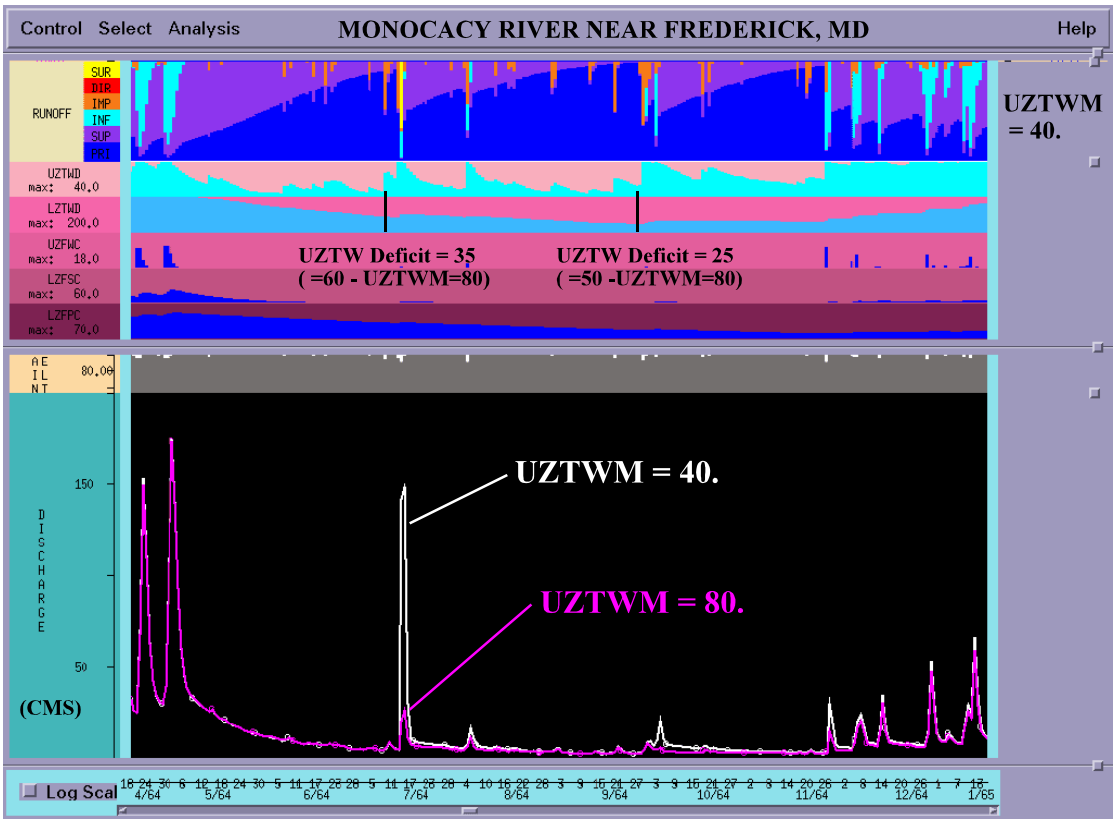


Figure 7-8-7. Effect of using SIDE on model response.

UZTWM

The primary function of UZTWM is to control the maximum upper zone tension water deficit that can be created. Remember no storm runoff, other than that from constant impervious areas, can occur when the upper zone tension water is not full. The value of UZTWM controls the time when the other storm runoff components (surface, interflow, and direct runoff) are suppressed during a dry period. Typically dry periods after a storm event during the summer are used to determine if changes are needed to UZTWM. Such periods allow for a sizeable upper zone tension deficit to develop. An evaluation can then be made as to whether the deficit is too large or not large enough based on the response from moderately large rain events that occur when the magnitude of the deficit approaches the value of UZTWM. Figure 7-8-98 illustrates the effect of changing UZTWM on the amount of storm runoff produced after a large upper zone tension deficit is generated. The deficits that exist prior to two storm events are noted on the figure. The first event had a much larger hydrograph response when UZTWM was equal to 40 partly due to more rainfall, but also the intensity was sufficient to generate surface runoff.

In very wet regions a significant upper zone tension deficit may never develop, thus it is impossible to determine the value of UZTWM. In such areas it is important to maintain a realistic spatial consistency in the values of UZTWM. Soil based estimates as discussed in Section 7-5 can provide this consistency in such regions.



Figure

7-8-8. Effect of changing UZTWM on model response.

PCTIM

The PCTIM parameter is evaluated at the same time as UZTWM since the only source of new runoff during periods when an upper zone tension deficit exists is from constant impervious areas. During periods when there is a substantial upper zone tension deficit, the response from small to moderate rain events is examined to determine if the value of PCTIM should be changed. As with any parameter, changes should be based on the trend over many events. The periods used to evaluate PCTIM are typically in the warm season when convective rainfall is predominate, especially for smaller events. Since the MAP values from such events are quite uncertain, it is even more important in the case of PCTIM to base any decision on the response of many events. Figure 7-8-9 illustrates how changing the value of PCTIM affects the watershed response during periods when a upper zone tension deficit exists. PCTIM has only a small effect on the hydrograph during other periods unless the value of PCTIM is very large.

In very wet regions where significant upper zone tension deficits seldom exist, it may be very difficult to determine a unique value for PCTIM. In many cases in such regions it is best to set PCTIM to zero if constant impervious runoff can't clearly be shown to exist..

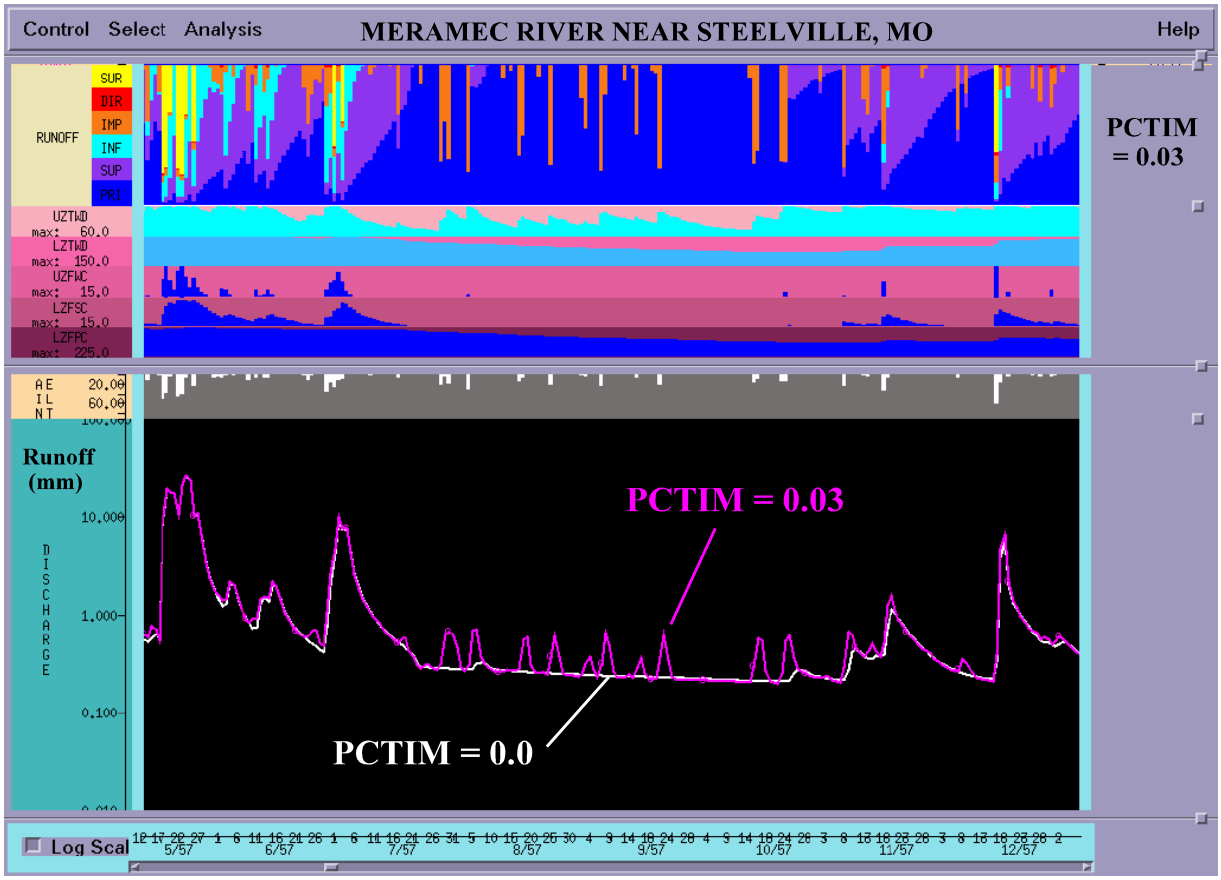
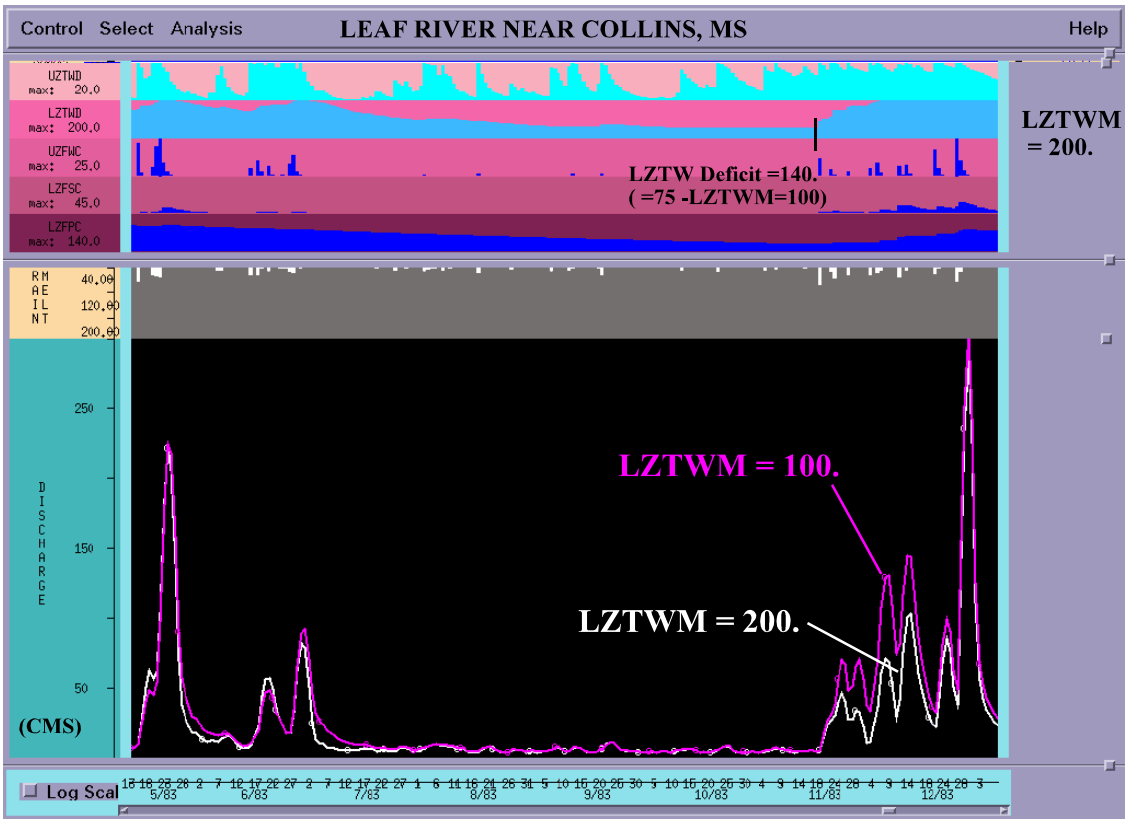


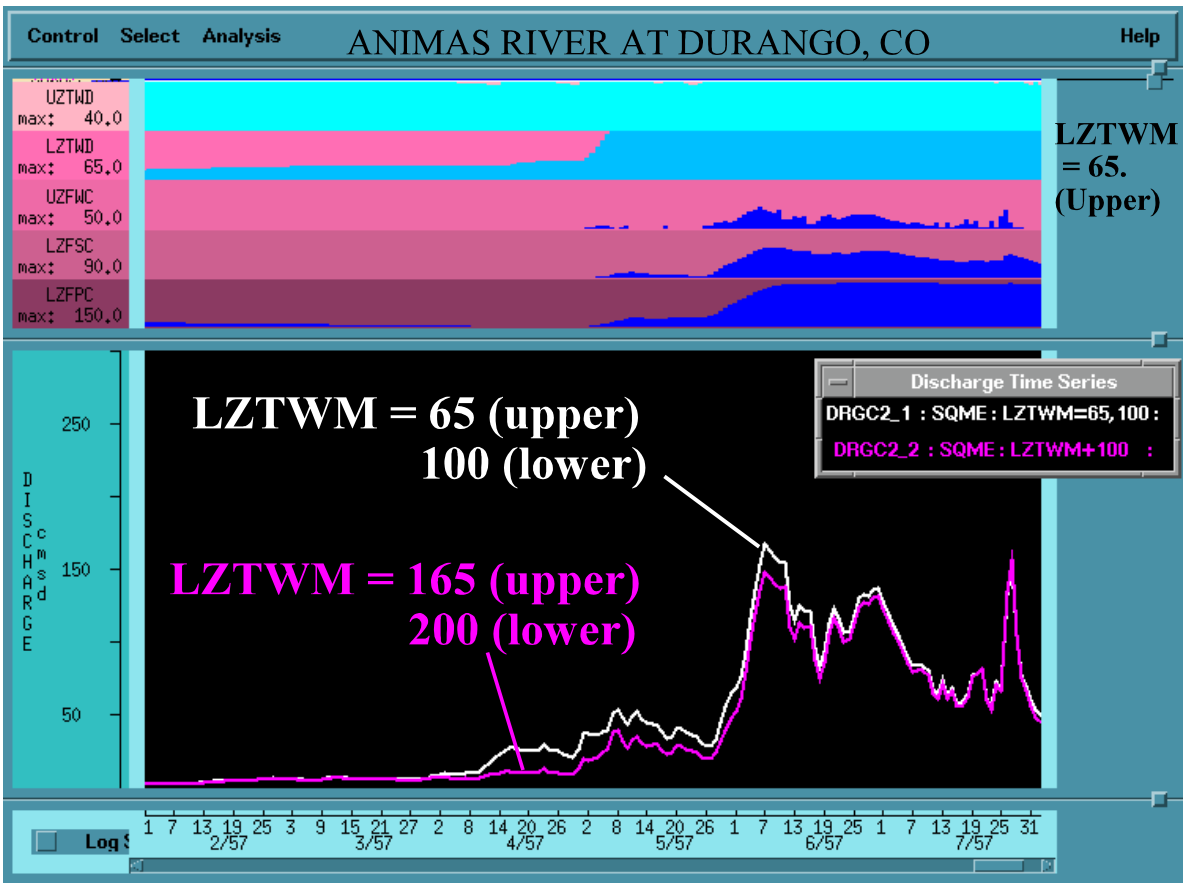
Figure 7-8-9. Effect of changing PCTIM on model response.

LZTWM

The primary function of LZTWM is to specify the maximum lower zone tension water deficit that can occur. The maximum possible deficit will occur after a long dry period when the vegetation withdraws all the moisture from the soil that the roots can reach. If the absolute maximum deficit occurred, all of the vegetation in the watershed would be wilted. Since such a situation seldom exists in nature, the lower zone tension water should never dry out completely. To determine if the value of LZTWM should be changed, the hydrograph response after long dry periods when the lower zone tension deficit approaches the value of LZTWM should be examined. The amount of runoff that occurs as the deficit is filled will indicate if the deficit was allowed to be too large or not big enough. In most areas, a large deficit that is built up over a very dry summer will be filled by fall rains, however, in watersheds where snowmelt dominates, such as in the intermountain west, the deficit is usually not filled until the spring melt season. In these cases the response in the early part of the melt season should be examined. Several factors can affect the model response at the beginning of the snowmelt season as discussed in the last part of Section 7-7. The value of LZTWM should only affect the response during years when a large deficit is carried through the winter. Figure 7-8-10 illustrates the type of period to use and the effect of changing LZTWM for the case when the deficit built up during the summer is filled by fall rains. Figure 7-8-11 illustrates the effect of LZTWM when a deficit is carried through the winter and not filled until the early part of the melt season.



Figure



7-8-10. Effect of changing LZTWM for an area with fall rains.

Figure 7-8-11. Effect of changing LZTWM on a predominately snowmelt basin.

If the size of the lower zone tension deficit never approaches the value of LZTWM for watersheds where long dry periods periodically occur from late spring into the fall resulting in hydrologic drought conditions, the value of LZTWM should likely be reduced so that the maximum deficit is closer to the value of LZTWM. During extreme dry periods it is physically realistic for the deficit to approach its limiting value. In wet regions where significant lower zone tension deficits never occur, a true value for LZTWM can't be determined by calibration. In these regions changing LZTWM will alter the amount of computed ET since the LZTWM is part of the equation used to calculate evaporation from the lower zone tension water, however, LZTWM must be change by quite a bit to produce a significant change in ET (i.e. the parameter is not very sensitive in this situation). In wet regions it is probably best to use the soil based estimates of LZTWM to maintain spatial consistency though it would be better if the estimates were based on vegetation, as well as soils data. In very dry regions it may also be impossible to reliably determine the value of LZTWM since the lower zone tension deficit may seldom be filled. Simulation results using a lumped application of a conceptual model are generally poor such regions as shown in Figure 1-1.

Percolation Rate (ZPERC, REXP, LZFSM)

It is best to evaluate the entire percolation curve when determining whether changes are needed to the parameters that control amount of percolation. This is especially true in the case of ZPERC and REXP. Since LZFSM has two primary functions, changes can be made to alter the magnitude of supplemental baseflow without doing a full evaluation of its affect on the percolation curve. The thought process recommended when evaluating the percolation curve is to first determine how the curve needs to be changed at different levels of lower zone wetness and then decide on the combination of parameter values that will produce the desired curve. It is important to evaluate the whole curve before deciding on any parameter changes since percolation rates may be too high under some moisture conditions and too low under others. Only an examination of events that occur at various lower zone moisture conditions will indicate how the curve should be modified.

As indicated ZPERC, REXP, and LZFSM are the 3 parameters whose primary function is to control the percolation curve (LZFSM also affects the amount of supplemental baseflow). REXP determines the shape of the curve, ZPERC the percolation rate when the soil is completely dry, and LZFSM dominates the rate when the soil is saturated). Other parameters also affect the percolation rate, but that is not their primary function. This includes UZFWM, LZFPM, LZTWM, LZSK, and LZPK. The PBASE term in the percolation equation, which specifies the rate when the lower zone is saturated, is calculated based on the fact that water can't enter the lower zone under saturated conditions faster than the lower zone free water storages can drain. PBASE is computed as:

$$\mathbf{PBASE = LZFSM \bullet LZSK + LZFPM \bullet LZPK}. \qquad (7\text{-}8\text{-}1)$$

The withdrawal rates, LZSK and LZPK, should never be changed in order to modify PBASE. Their only function is to specify the slope of the two baseflow recessions. Changes to LZFPM will alter the value of PBASE, however, generally the LZFSM • LZSK term of the equation is an order of magnitude greater than the primary baseflow term, thus LZFSM becomes the dominate parameter when changing the saturated percolation rate. When the division between supplemental and primary baseflow is correct, but changes to PBASE are indicated, both LZFSM and LZFPM can be changed in a manner that keeps their ratio the same.

Before evaluating the percolation curve, it is important to first know what to look for to decide whether a change in the percolation rate is needed. Figure 7-8-12 illustrates the model response that indicates there is an error in the percolation rate. When LZTWC is equal to LZTWM, i.e. lower zone tension storage is full, none of the percolated water can be stored in tension water. In that case the percolation curve controls the division between storm runoff, surface plus interflow, and baseflow, supplemental plus primary. A decrease in the percolation rate at any given lower zone moisture level will produce more storm runoff and less baseflow recharge. An increase in the percolation rate will do the opposite. When LZTWC is less than LZTWM, then much of the percolated water will go into tension water storage, thus reducing the amount of recharge (all of

the percolated water will go into lower zone tension storage when PFREE is zero - some recharge will occur if PFREE is greater than zero). Thus, in this case, changing the percolation rate primarily alters the amount of storm runoff with only a minimal effect on recharge depending on the value of PFREE.

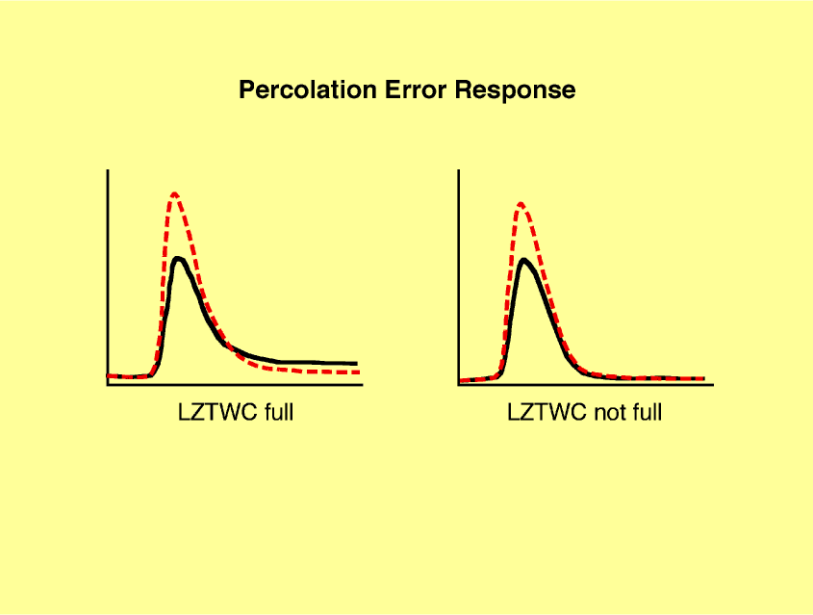


Figure 7-8-12. Illustration of model response to changes in the percolation rate.

The ICP has a percolation analysis feature built into the WY-PLOT display. This tool allows the user to go through many events, hopefully over a wide range of lower zone moisture conditions, and indicate the relative magnitude that the percolation curve should be altered to improve the simulation. If this analysis shows any trends in how the overall shape of the curve should be modified, then the parameters can be changed by trial and error until the desired curve is obtained. The steps to follow to perform a percolation analysis are:

1. In the main ICP window select ‘Water Year Plot’ under the ‘Display’ menu. Then choose ‘SAC-SMA’ under the ‘Select’ menu of the WY-PLOT display (the Sacramento model zone contents and runoff component breakdown must be included on the WY-PLOT display for the percolation analysis option to be active). Then choose ‘Percolation’ under the ‘Analysis’ menu. This will cause the current percolation curve to be displayed in one window and the current values of ZPERC, REXP, LZFSM and LZFPM to be displayed in another window.

2. In the window that shows a plot of the percolation curve choose ‘Points’ under the ‘Select’ menu and then choose ‘Build’. After this is done, whenever the right mouse button is clicked within the hydrograph plot portion of the WY-PLOT display, a dotted, red vertical line will appear on the percolation curve plot to indicate the lower zone deficiency ratio, LZDEFR, at that time and thus where you are on the percolation curve.

3. Go through the calibration period and select those events when the model response indicates a percolation error as shown on Figure 7-8-12. Events where other types of errors predominate, such as volume or timing errors, should be ignored. The SAC-SMA display is used to determine whether a lower zone tension deficit exists. Then determine where you are on the percolation curve during the event as described in step 2. You can determine where you are just prior to the event or can get the average percolation rate by checking on days just before and after the event. The main consideration is to be consistent in what you do from one event to another. For each event selected, click on the percolation curve plot where the vertical line appears to indicate whether the percolation rate should be higher, lower, or remain about the same. This will put a small red 'x' on the plot to indicate the relative change in the percolation rate that is needed to improve the simulation of that event (e.g. increase the rate significantly, decrease the rate slightly, or keep the rate the same). The amount that the curve needs to be changed is subjective. Over time you will become more skilled at anticipating the amount of change needed.

4. After going through the entire calibration period there should be many red x's on the percolation curve plot. If there is a trend in the suggested changes, it indicates that a new percolation curve is likely needed. If the points are scattered around the plotted curve, it indicates that no changes are needed to the current parameter values. If a new curve is suggested, the idea is to select changes to the parameter values to produce a curve that best fits the points. This is done by trial and error. Under the 'Select' menu of the percolation curve plot window choose 'Parameters'. This will bring up a new window that lists the current value of the 4 parameters. Make changes to the parameters and then click on 'Compute'. A new curve will be drawn on the percolation plot. Keep doing this until you get a curve that fits the points to your satisfaction. Then use those parameter values for the next calibration run. There is also a 'Solve' option under the 'Points' option of the 'Select' menu that will automatically fit a curve to the points. It is not recommended that this option be used.

In many situations going through the steps in the percolation analysis will not result in significant changes in the simulation results. By the time that a percolation analysis is performed, adjustments typically have been made to get a reasonable baseflow simulation and an overall water balance that is within a few percent. In order to for this to be the case, the overall amount of storm runoff should be realistic. Thus, most events will already have a reasonable percolation rate. The percolation analysis is primarily helpful in determining if the percolation curve adequately handles the full range of soil moisture conditions. Thus it is very important when doing the analysis to include as wide of range of LZDEFR values as possible. This is because over a certain range of LZDEFR values there are many combinations of PBASE, ZPERC, and REXP that will produce very similar curves. This is illustrated in Figure 7-8-13. For LZDEFR values from about 0.3 to 0.6 all of the curves on this figure look very much alike. For this example it would be very important to have events that occur under very wet conditions and especially cases when significant rain occurs when the soil is very dry in order to adequately define the percolation curve. In wet areas there may not be a wide range of LZDEFR values and

thus it is impossible to get unique values for ZPERC and REXP. In these situations it is best to follow the guidelines in Section 7-5 and maintain spatial consistency. In some watersheds even though a wide range of LZDEFR values occur, the rain events that take place when the soil is dry are not of sufficient size to produce enough runoff to determine if the percolation curve needs to be modified. In these cases, if a good evaluation of the percolation curve can be achieved for some watersheds in the region, the values of ZPERC and REXP for those watersheds should be applied throughout the region with some possible modification for changes in soils.

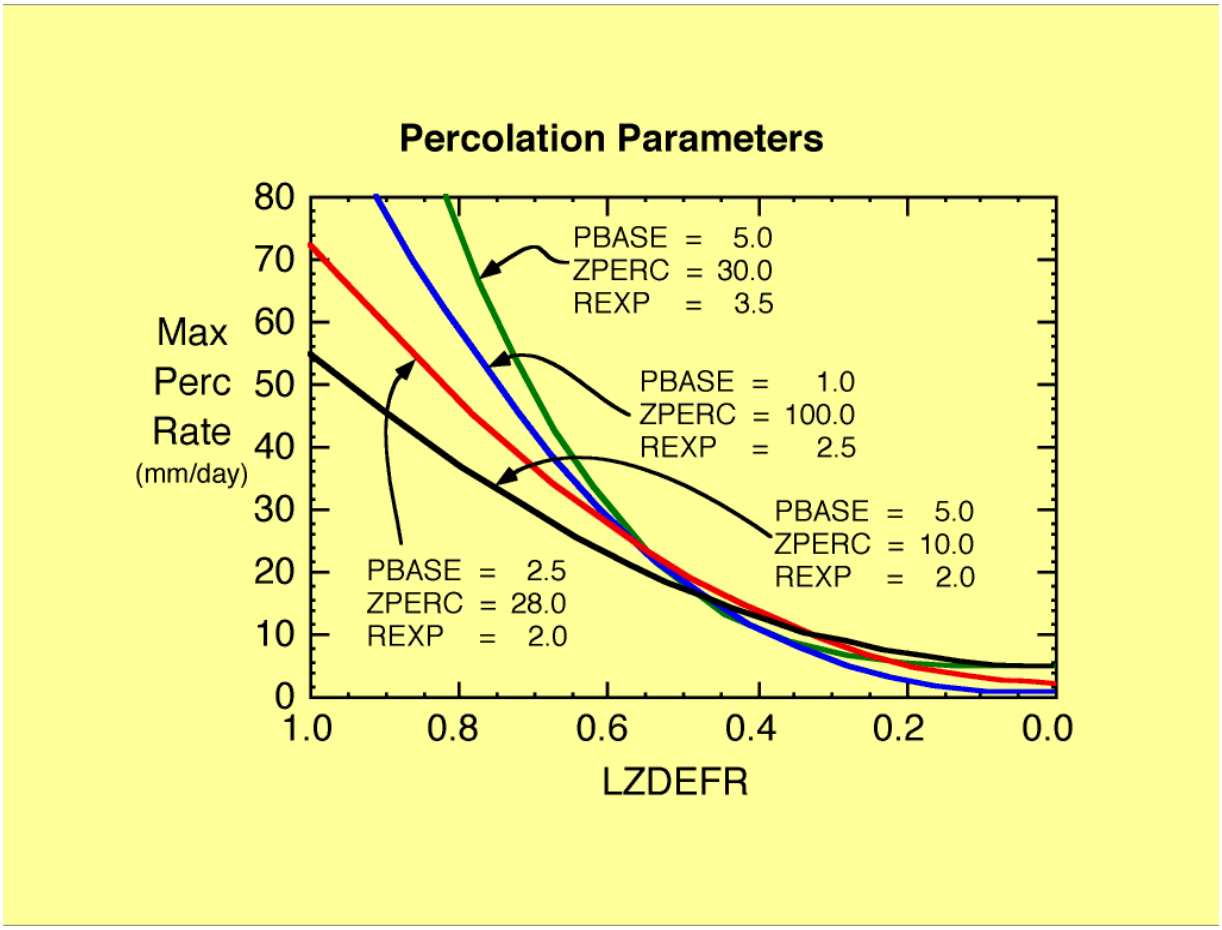


Figure 7-8-13. Illustration of how various parameter values produce similar percolation curves.

Figures 7-8-14 through 7-8-18 illustrate how changes to PBASE, ZPERC, and REXP affect model response. These figures should also reinforce what you are looking for when determining whether the percolation rate needs to be changed. In most of these figures the percolation analysis window that shows the percolation curves is included along with the main WY-PLOT display.

Figures 7-8-14 and 7-8-15 show the effect of PBASE on the model response. An arithmetic scale is used for the first figure to emphasize the high flow response and a semi-log scale is used for

the second figure to accentuate the effect on baseflow recharge. When PBASE is 7.3, there is

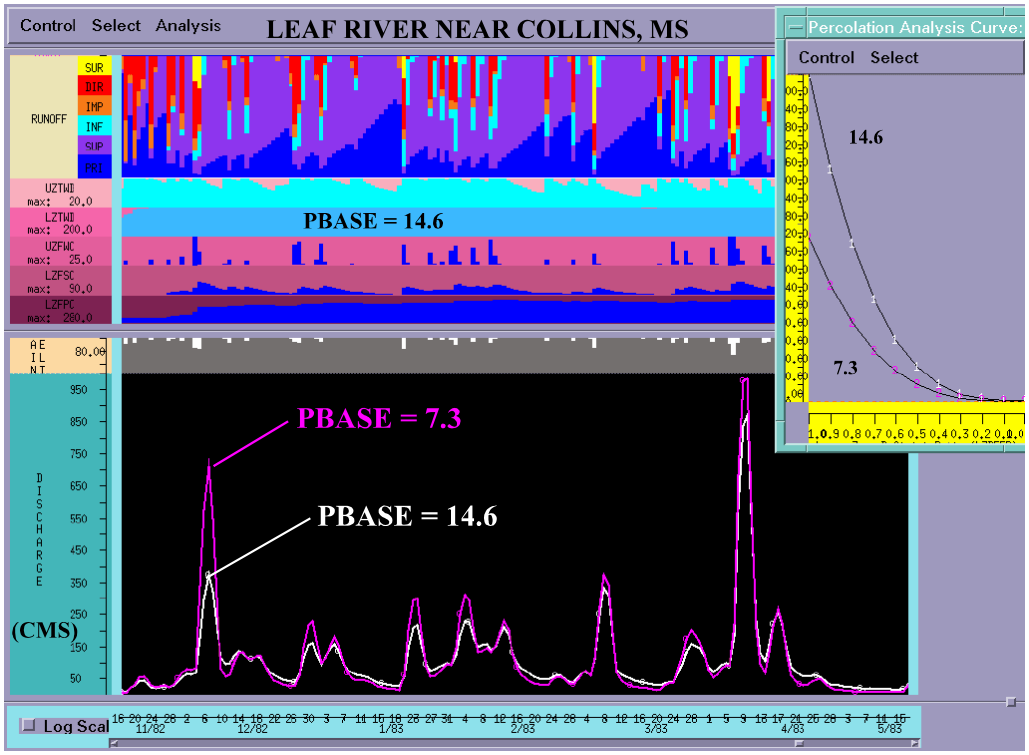


Figure 7-8-14. Effect of changing PBASE on response (arithmetic scale).

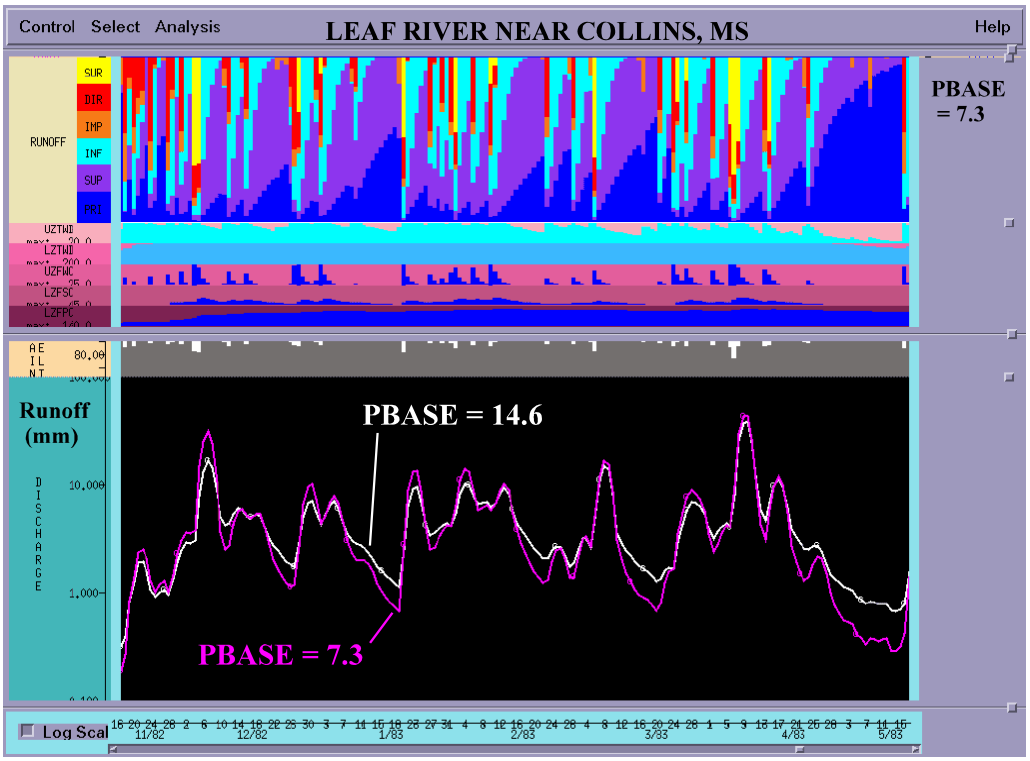


Figure 7-8-15. Effect of changing PBASE on model response (semi-log scale).

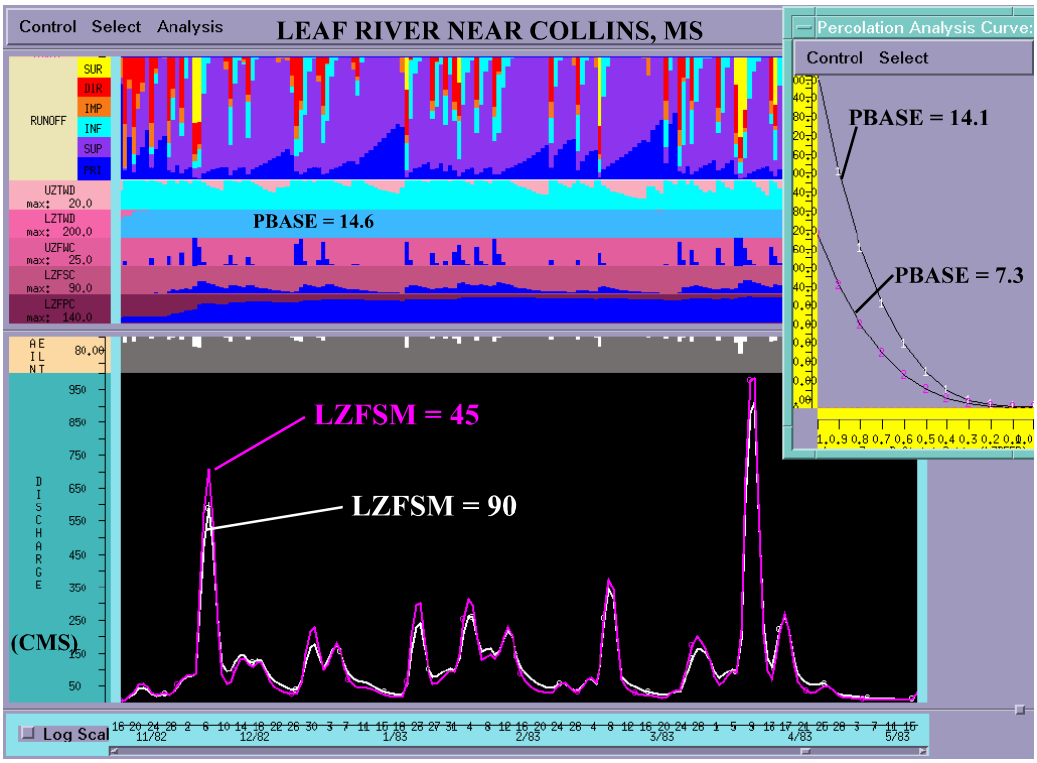


Figure 7-8-16. Effect of changing LZFSM on the percolation rate.

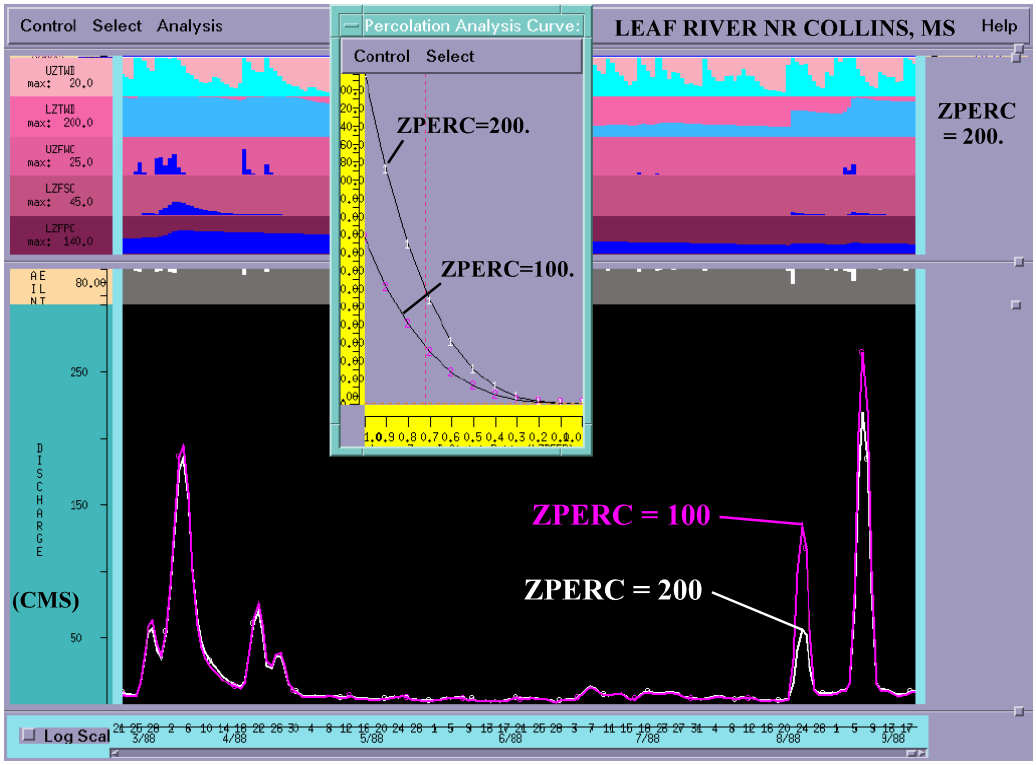


Figure 7-8-17. Effect of changing ZPERC on model response.

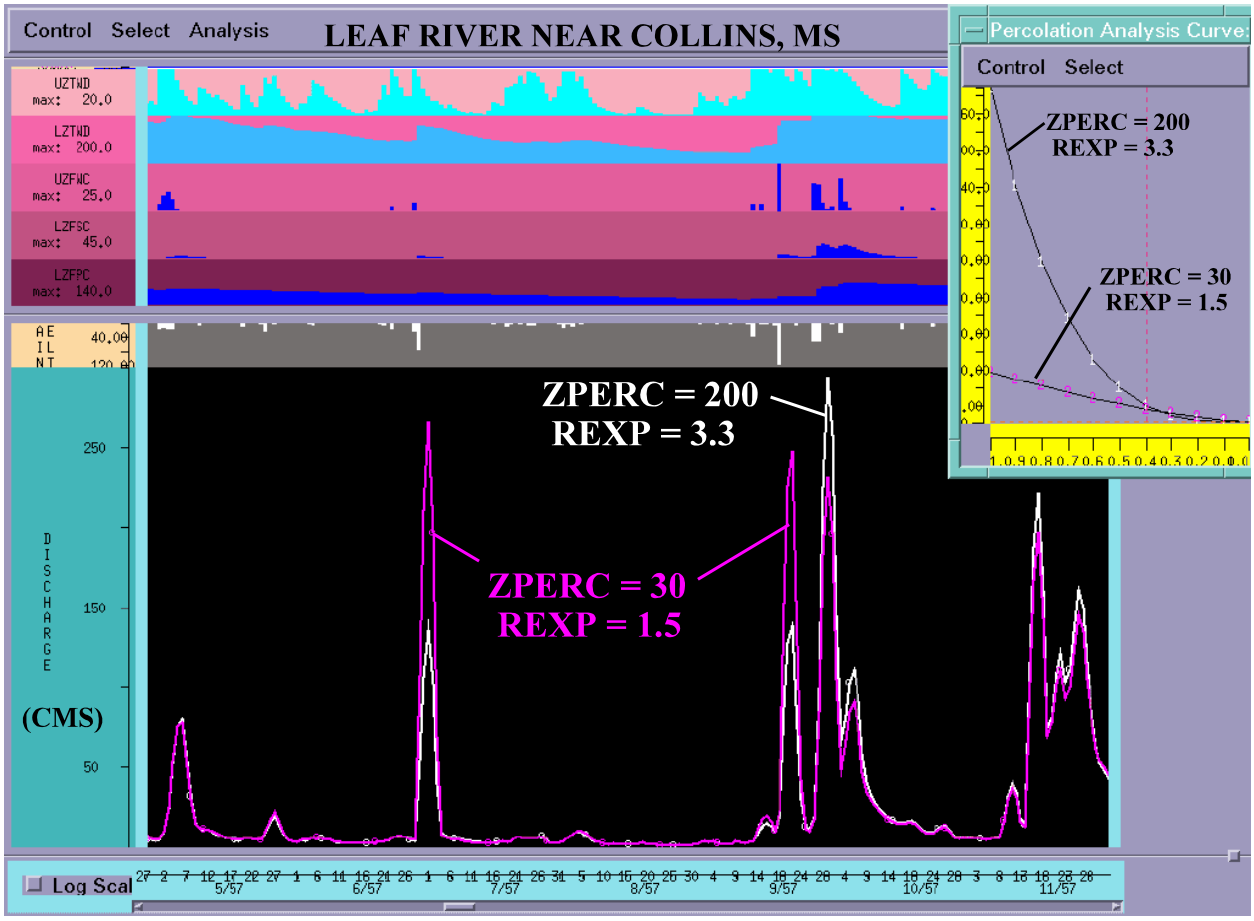


Figure 7-8-18. Effect of changing ZPERC and REXP on model response.

more storm runoff and less baseflow. This occurs for all events because a change in PBASE causes the entire percolation curve to increase or decrease by the ratio of the new PBASE to the previous value. Figure 7-8-16 is the same as Figure 7-8-14 except that only LZFSM is changed (in the previous 2 figures both LZFSM and LZFPM were altered by the same ratio in order to change PBASE). Doubling LZFSM from 45 to 90 mm causes PBASE to change from 7.3 to 14.1, thus showing how LZFSM dominates the PBASE value (doubling both LZFSM and LZFPM resulted in a PBASE of 14.6). The effect on percolation of doubling LZFSM is almost the same as the effect when both lower zone free water storages are doubled.

Figure 7-8-17 illustrates the effect of changing only ZPERC. Doubling ZPERC results in the maximum percolation rate under dry conditions to almost double while the rate under saturated conditions is not changed. Thus, the events in August and September, when the LZDEFR is quite large (i.e. dry soil due to a substantial lower zone tension deficit and little water in free water storages), are significantly more affected by changing ZPERC than the events in March and April when the soil is wet.

Figure 7-8-18 illustrates how events can be affected in different ways when the percolation curve is modified. In this case the two curves cross at a LZDEFR value of about 0.4. The maximum percolation rate is lower for the ZPERC=200, REXP=3.3 curve when the soil is wet, but greater when the soil is dry. This causes the model response to change as the soil becomes more saturated during the period shown. During the summer, the soil is quite dry, thus the ZPERC=30, REXP=1.5 curve produces much more storm runoff for the early July and mid September events. Right after the mid September event the LZDEFR value drops below 0.4, therefore the response of the remaining events is the opposite, i.e. the 30,1.5 curve produces the least amount of storm runoff. This example emphasizes why it is important to work with the entire percolation curve and to analyze the rate over a wide range of moisture conditions.

UZFWM

The primary function of the UZFWM parameter is to control when surface runoff occurs. Surface runoff can only occur when the intensity rate of the rainfall or rain+melt is sufficient to fill the upper zone free water storage. It is much easier to fill this storage when the percolation rate is low. This implies that surface runoff most frequently occurs under wet conditions and in regions where the soils have low permeability (baseflow amounts are low). Thus, the events to examine to determine if a change is needed to the value of UZFWM are those where surface runoff is already taking place with the current value of the parameter and those with large amounts of precipitation and near saturated soil conditions. Surface runoff can also occur for very high intensity events under dry conditions for watersheds with low permeability. The immediate response from these events is evaluated to determine whether more or less surface runoff is needed. The most important events to analyze are those that are just producing a little surface runoff and those that are currently not generating any, but the upper zone free water is close to being filled. An analysis of these events should suggest whether UZFWM should be increased or decreased. It should be noted that the UZFWC values shown on the ICP Sacramento display cannot always be used as an indication of how close an event was to producing surface runoff since the value shown is for the end of the day and free water contents can change quite rapidly for the upper zone.

Figures 7-8-19 and 7-8-20 illustrate how the model response is affected by changing the amount of surface runoff that is generated as the value of UZFWM is modified. The hydrograph plots on these two figures are exactly the same. The difference in the figures is the panel showing the runoff components. On Figure 7-8-19 the components are shown for the case when UZFWM=15. Figure 7-8-20 shows the components when UZFWM=30. When UZFWM=15 there are 4 events generating surface runoff. The first 3 produce quite a bit of surface runoff, while the early May event only generates a little. When UZFWM is changed to 30, the May event doesn't generate any surface runoff and the other events produce much less. As UZFWM is increased and surface runoff decreases, the amount of interflow increases since there is more room to storage potential interflow runoff in the upper zone free water. Thus an effect of altering UZFWM is to change the relative division of storm runoff into surface and interflow. When less surface runoff occurs as in the case of UZFWM=30, not only is the immediate response reduced, but the timing is also modified. When there is more interflow, the peak occurs later because surface runoff enters the channel immediately while interflow is released from storage over some period of time.

For regions with highly permeable soils and those that never experience high intensity precipitation, at least at the spatial and temporal scales being used, surface runoff will never occur. In these cases it is impossible to determine the value of UZFWM. If UZFWM can be reasonably determined for a couple watersheds in such regions, these values could be used throughout the region. If not, a value great enough to surpress surface runoff should be selected and spatial consistency should be maintained.

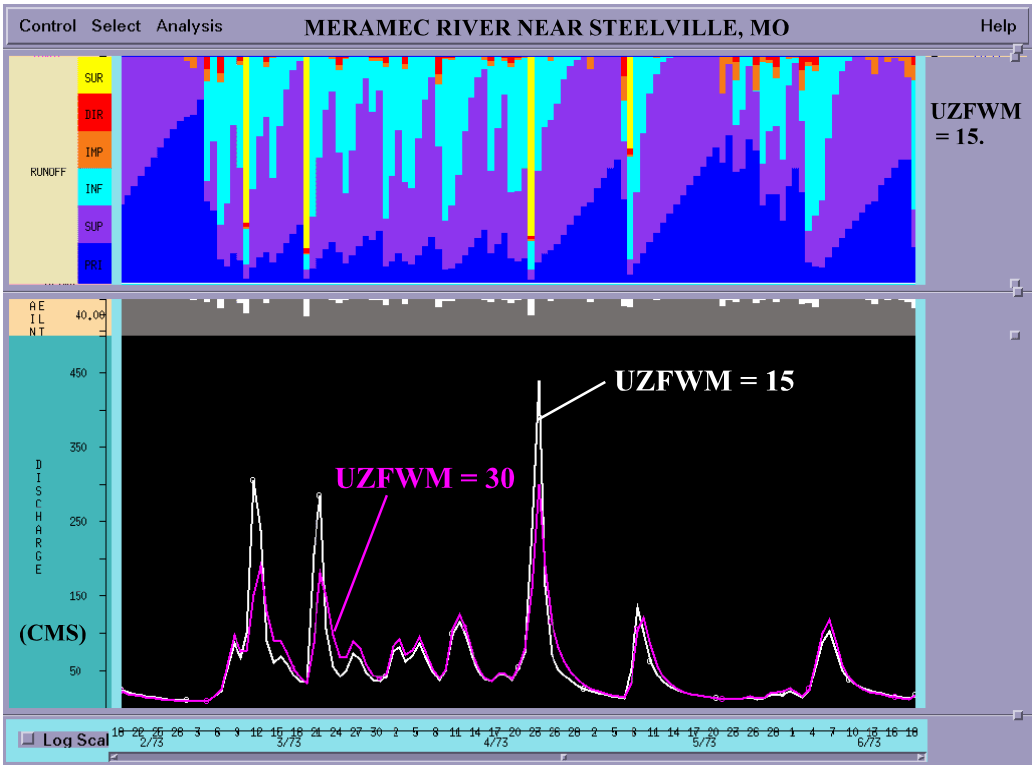


Figure 7-8-19. Effect of changing UZFWM on response model (UZFWM=15 components).

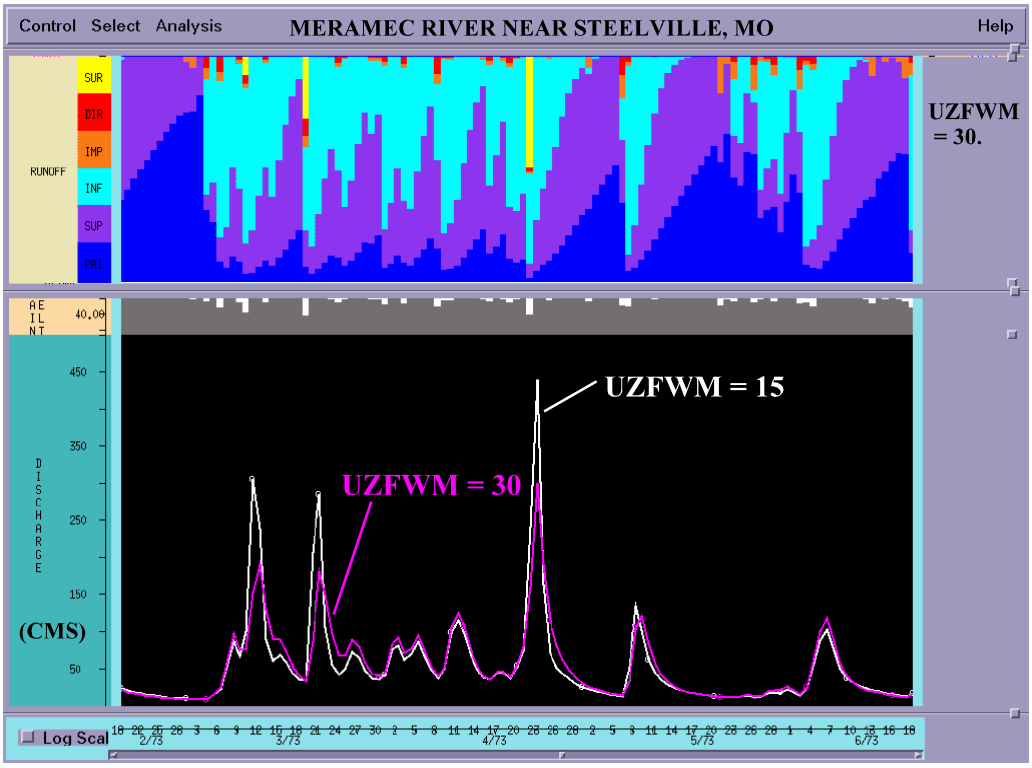


Figure 7-8-20. Effect of changing UZFWM on model response (UZFWM=30 components).

UZK

The primary function of the UZK parameter is to control the timing of interflow runoff. For most watersheds, interflow dominates small to moderate rises in the hydrograph. These are the events to examine to determine if the value of UZK should be modified. For watersheds with highly permeable soils and little or no surface runoff, interflow will dominate the response from all storm events. In these cases the timing sometimes appears to vary with the magnitude of the event, i.e. a greater UZK is indicated for large events than for small rises. Since only a single value can be used, it is generally best to try to match the response of the larger events when this situation occurs, though sometimes ADIMP can be used to produce a faster response from the larger events without affecting the smaller events very much.

Figure 7-8-21 illustrates the effect of changing UZK on model response. Changing UZK primarily affects the smaller rises where interflow is basically the only source of storm runoff. When surface runoff is significant, modifying UZK doesn't have much affect on the response. It should be noted that there is a large difference in the two UZK values used in this figure. This indicates that UZK is not as sensitive as some of the other model parameters. The effect of small changes in the value of UZK may be difficult to see.

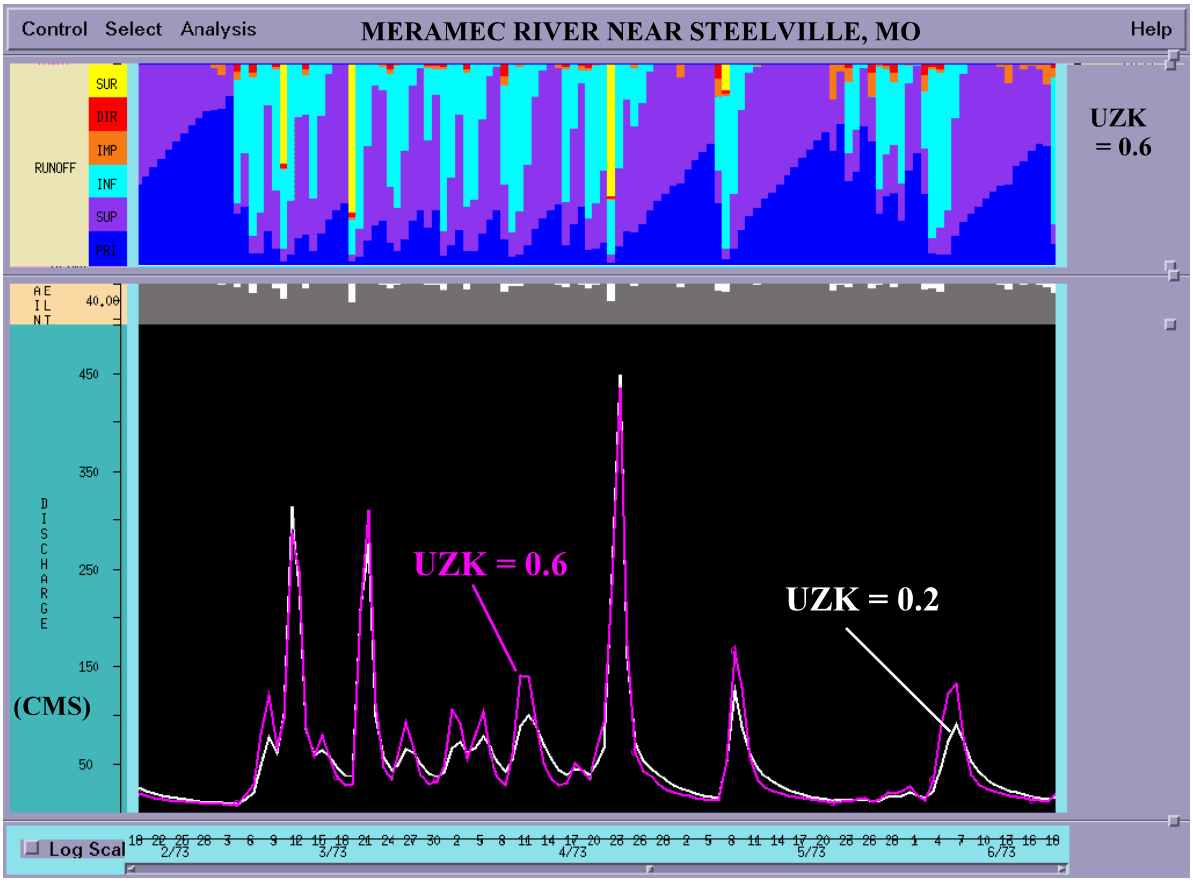


Figure 7-8-21. Effect of changing UZK on model response.**ADIMP**

The primary function of the ADIMP parameter is to produce some immediate response runoff (direct runoff) in cases when precipitation intensities are relatively low or the soils are very permeable. These are events when interflow dominates and there is little or no surface runoff. This suggests that for watersheds where surface runoff occurs frequently ADIMP may not be needed and that ADIMP will have the most noticeable effect on watersheds where surface runoff seldom or never occurs. It should be remembered that direct runoff can only occur when upper zone tension water is full, i.e. during events that produce some storm runoff.

Figure 7-8-22 illustrates the effect of ADIMP on a watershed that generates surface runoff from most large rain events. For the large events where surface runoff predominates, adding direct runoff by using some ADIMP produces little difference in model response. For smaller events where there is no surface runoff, the use of ADIMP generates a more rapid response though the semi-log scale masks some of the effect. Figure 7-8-23 illustrates the effect of ADIMP on a watershed that seldom produces any surface runoff. In this case adding some direct runoff produces a more rapid response from all the storm events.

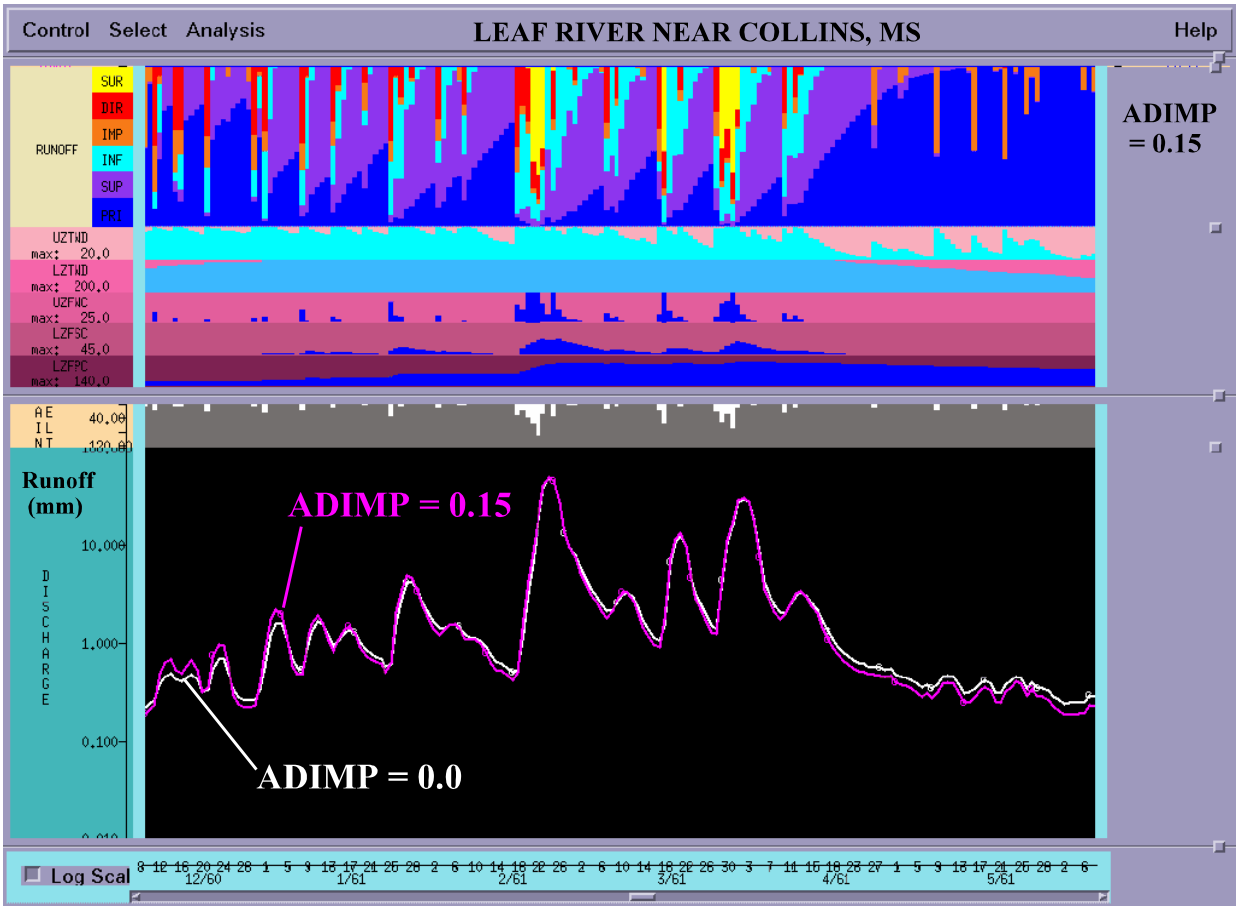
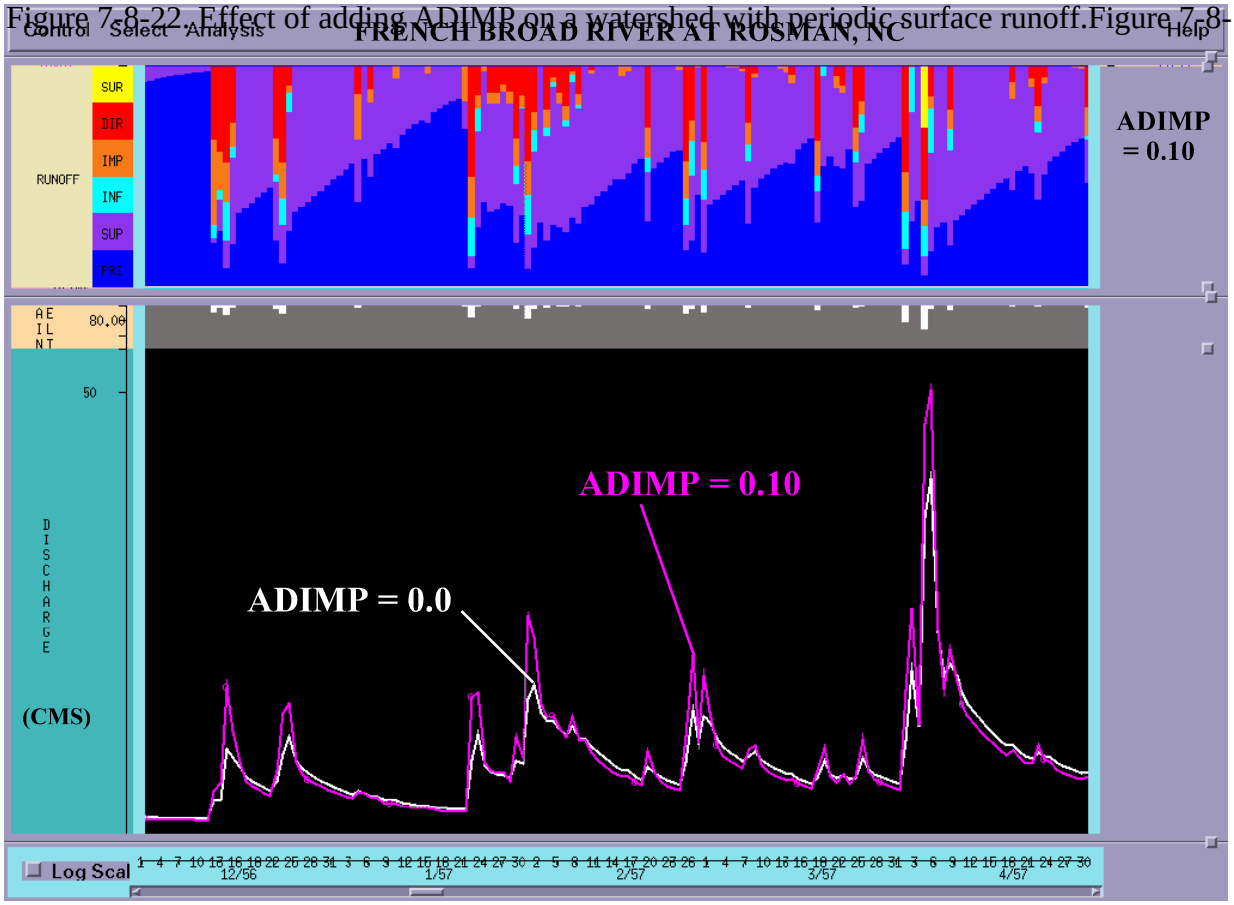


Figure 7-8-22. Effect of adding ADIMP on a watershed with periodic surface runoff.Figure 7-8-



23. Effect of adding ADIMP for a watershed with little surface runoff.

As indicated in Section 7-5 it is best to start with ADIMP set to zero unless a value can easily be derived from a hydrograph analysis. Then during the calibration decide whether ADIMP is needed. There is no reason to add complexity to the model if it is not clearly needed.

RIVA

The primary function of the RIVA parameter is to draw down the baseflow during very dry periods when ET-Demand rates are significant. These are the periods when riparian vegetation doesn't have adequate moisture near the surface to satisfy its needs so it withdraws water from groundwater before it reaches the channel system. Thus, the periods to examine to determine if the value of RIVA needs to be changed are those when there is a substantial upper and lower zone tension water deficit during a time of the year when significant evaporation is occurring. Typically the most likely time for riparian vegetation evaporation to occur is in the later part of the summer and early fall during a dry year. Figure 7-8-24 illustrates how adding RIVA will affect the model response during a dry period. In this case substantial tension water deficits exist from late May into October. The lower zone tension deficit doesn't fill until early January, but the riparian vegetation effect is over by mid November primarily because of the reduction in the evaporation rates and partly because of reduced upper zone tension water deficits.

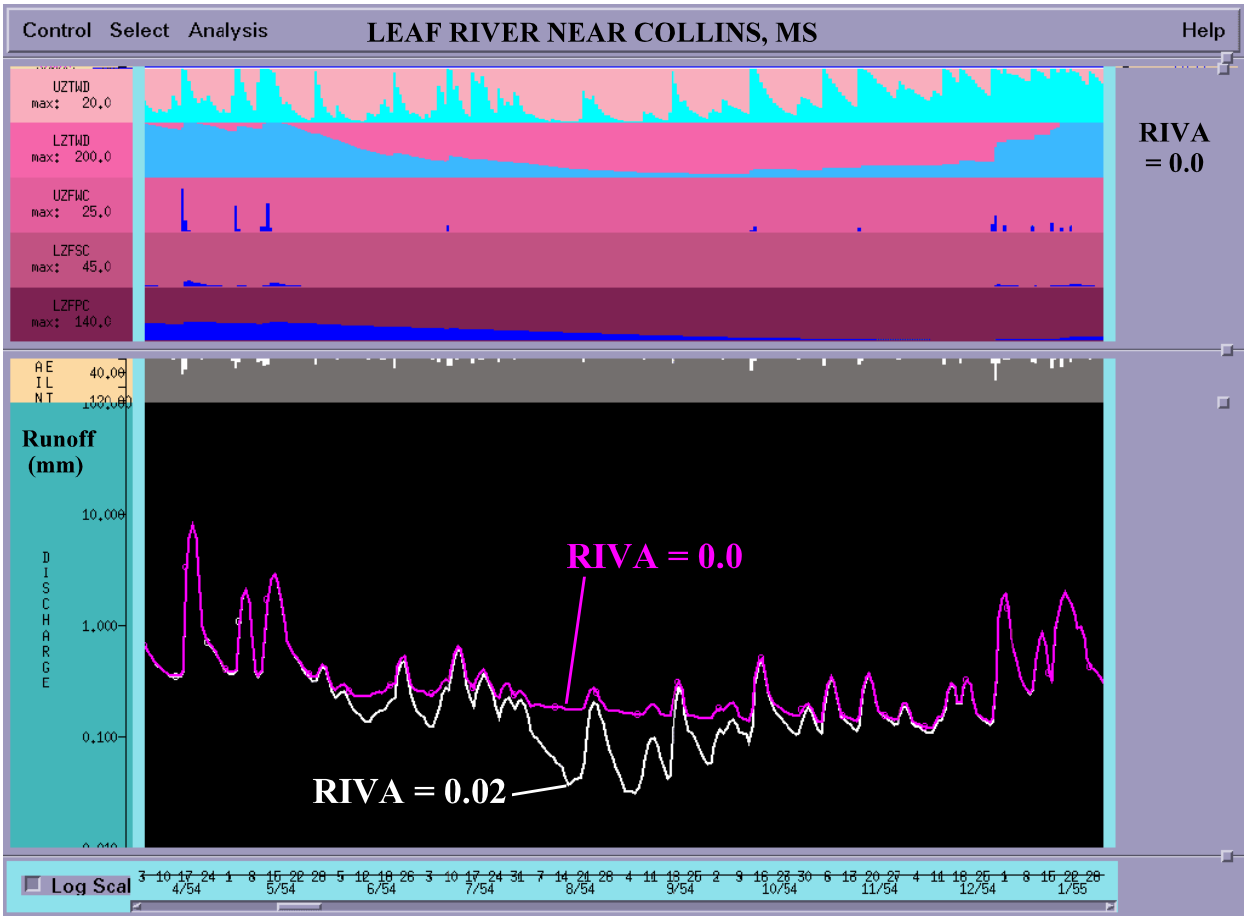


Figure 7-8-24. Effect of adding RIVA on model response.

PE Adjustment Curve

Changes to ET-Demand are indicated by a pattern in the seasonal, i.e. monthly, bias values on the statistical summary. Since other processes can affect the seasonal bias, such as the division between storm runoff and baseflow, baseflow withdrawal rates, and snowmelt timing, one shouldn't look at possibly changing the ET-Demand until problems with these other processes have been corrected. Errors in ET-Demand could be the result of improper estimates of PE, but are most likely due to an erroneous estimate of the seasonal PE adjustment curve. Thus, it is best to change the ET-Demand rate by modifying the PE adjustment curve. When using average monthly ET-Demand as input, the PE adjustment curve should be changed and then each months value multiplied by the average PE. This should insure that the shape of the PE adjustment curve remains physically realistic. Figure 7-8-25 illustrates how changes to the PE adjustment curve will affect the seasonal bias pattern. It should be noted that changes in the ET-Demand for a given month may affect the bias in subsequent months since tension water deficits produced by evaporation don't affect runoff until an event occurs.

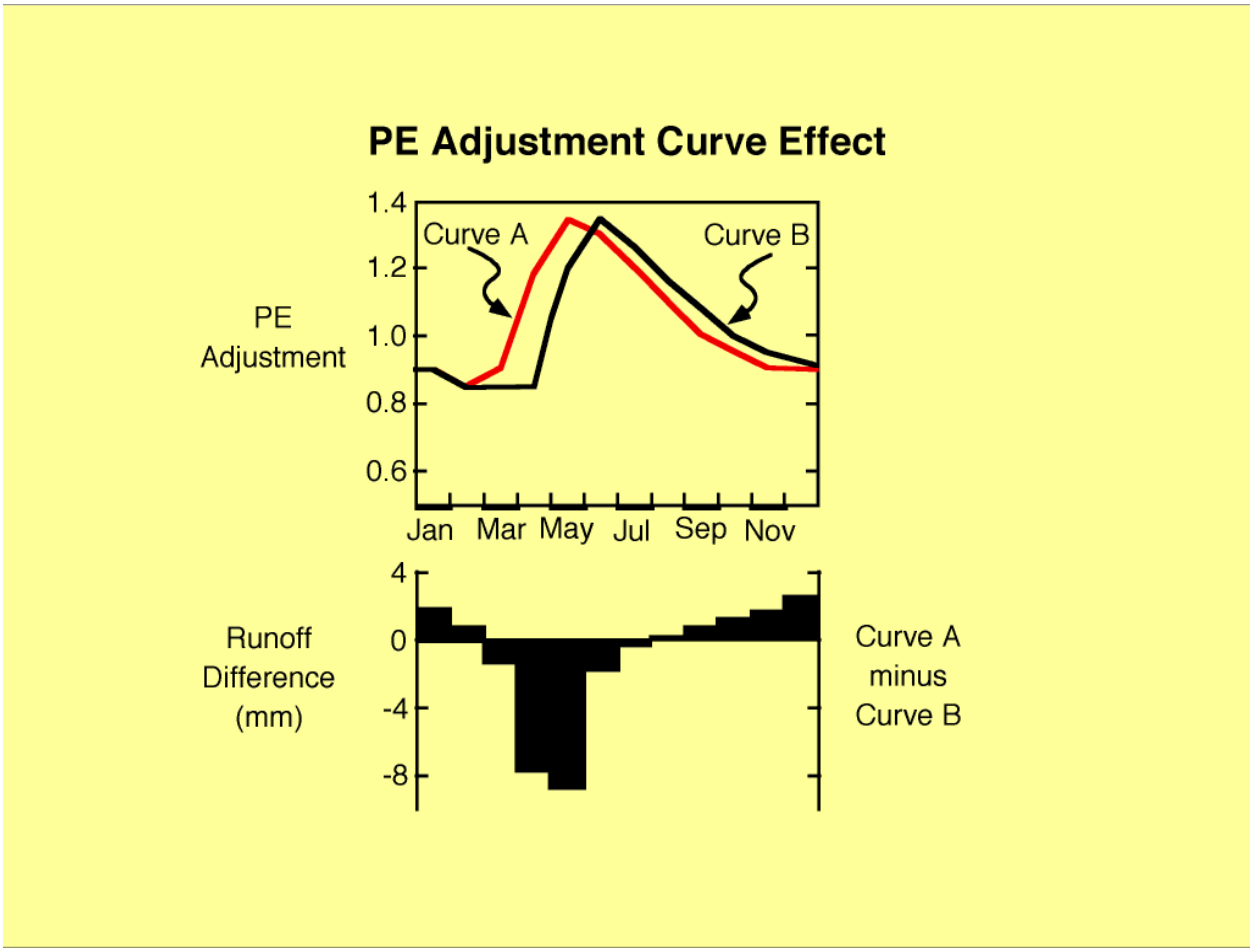


Figure 7-8-25. Effect of changing the PE adjustment curve on the seasonal bias pattern.

Parameter Effect Summary Table

Parameter	Primary Effect
LZPK	Controls slope of primary baseflow recession
LZSK	Controls timing of supplemental baseflow
LZFPM	Controls magnitude of primary baseflow
LZFSM	Controls magnitude of supplemental baseflow and percolation curve
PFREE	Controls baseflow recharge when lower zone tension deficit exists
SIDE	Determines if recharge to deep groundwater aquifers occurs
UZTWM	Controls maximum tension deficit that can occur in the upper zone (determines when storm and direct runoff and recharge will first occur after a dry period)
PCTIM	Controls the amount of fast response runoff when an upper zone tension deficit exists
LZTWM	Controls the maximum tension deficit that can occur in the lower zone
Percolation Curve	Controls the division between storm runoff and baseflow • ZPERC determines percolation rate when soil is completely dry • REXP determines the shape of the percolation curve • LZFSM controls the saturated percolation rate
UZFWM	Controls when surface runoff occurs
UZK	Controls the timing of interflow
ADIMP	Determines whether direct runoff occurs from low intensity events when the soil is wet
RIVA	Determines whether evaporation from riparian vegetation occurs when the soil is dry and evaporation is significant
PE Adjustment Curve	Controls the seasonal ET-Demand pattern along with PE - effect reflected in statistical seasonal bias pattern

Table 7-8-1. Summary of primary effect of each Sacramento model parameter.

Section 7-9

Evaluation of an *A Priori* Parameter Estimation Procedure

Introduction

This section illustrates how an *a priori* parameter estimation procedure can be evaluated for use in a given region. As mentioned in Section 7-5, is it important to evaluate the derived parameter values based on periods when the effects of each parameter on the hydrograph can be isolated rather than on overall 'goodness of fit' statistics. This is the only way to determine if the procedure is determining physically realistic values for the parameters. It is also helpful to determine if the *a priori* procedure can derive realistic spatial variations in parameter values over a river basin especially when some of the watersheds response quite differently than others or have significantly different values of key physiographic factors.

The Koren [2000] procedure is used in this section to illustrate the method for evaluating the derived parameter values. In this procedure total soil depth and porosity, field capacity, and wilting point information for 11 soil layers (from ground surface to 2.5 m beneath) derived from the STATSGO high resolution gridded soil data files are used to calculate estimates of 11 of the Sacramento model parameters. Tension water is determined as a function of the difference between the field capacity and wilting point and free water as the difference between porosity and field capacity, which is physically what these storages are intended to represent. The capacity of these storages are determined by dividing the soil profile depth into an upper and lower zone and further dividing the lower zone free water into primary and supplemental components. This assumes that the free water and tension water for each zone occupy the same space in the soil profile. This may not be the case. Certainly in most watersheds the primary baseflow, and probably the supplemental, aquifers are further down in the soil than the limits of the root zone which controls the lower limit of tension water storage. The free water withdrawal rates are calculated from empirical relationships. ZPERC is determined using the model percolation equation and assuming that the maximum daily percolation rate is equal to the contents of the all the lower zones, which may not be a correct assumption. The percolation curve shape parameter, REXP, is estimated using an empirical equation that produces low values, near 1.0, for sand and high values for clay soils, which are appropriate. The PFREE parameter is assumed to relate to water that follows paths through cracks, faults, etc. to escape the capillary demands of the soil and it is further assumed that clay type soils have more cracks. The PFREE parameter is actually more likely to be related to the distribution of lower zone tension water depths over the watershed and the fact that some portions of the area are saturated and thus recharging groundwater while other areas still have tension water deficits.

Evaluation

The suggested method of evaluating any *a priori* procedure for estimating model parameters is to compare the simulated hydrograph produced by the derived parameters to the observed

streamflow for the watershed. The objective is then to examine the periods when the effects of each parameter can be isolated to determine if the derived values are reasonable. In the following illustration of this technique, the parameters derived using the Koren method are referred to as soil based parameters.

In order to generate a hydrograph other parametric information is required in addition to the 11 Sacramento model parameters produced from the soils information. For the examples in this section these other parameters were determined as follows:

- Sacramento model parameters PCTIM, ADIMP, and RIVA are set to 0.0. Also RSERV is set to 0.3, its recommended default which is typically satisfactory for most watersheds, and SIDE is set to 0.0, which is the case for most areas.

- The ET-Demand curve was either obtained from a calibration of the watershed or from CAP estimates of mean monthly PE and the seasonal PE adjustment curve. If the overall bias was greater than a few percent, the PEADJ factor in the SAC-SMA operation was used to get the water balance within this limit.

- The unit hydrograph was obtained from a calibration or by estimating the channel response and then making adjustments to obtain a reasonable fit of the timing for major runoff events.

- If snow was included in the simulation, the snow model parameters were obtained from a calibration of the watershed.

Generally when evaluating an *a priori* parameter estimation procedure for the Sacramento model, ET-Demand values would be estimated as described in Section 6-5, the initial unit hydrograph would be obtained as discussed in Section 7-6, and the initial snow model parameters would be based on the guidelines included in Section 7-4. In this case calibrated values were used whenever possible to remove any variability generated by improper values of these parameters.

The first basin examined is the Oostanaula River basin in Georgia. Figure 7-9-1 shows a semi-log plot comparison of the soil based parameter simulation to the observed flow for the Conasauga River near Tilton (TLNG1) for a portion of the period of record. Figure 7-9-2 is an arithmetic plot for the same period. These figures indicate that at least most of the soil based parameter estimates are quite realistic for this watershed. Some slight adjustments are probably needed to some of the baseflow parameters, but overall the method derives values that produce a very reasonable simulation. In Chapter 4 a spatial assessment of hydrologic variability within the Oostanaula basin, based on hydrograph plots as shown in Figure 4-1, indicated that there were significant differences in how the Ellijay, Hinton, and Tilton watersheds responded. This assessment indicated that the percolation rate for Elijay and Hinton should be greater than for Tilton as those watersheds generate much more baseflow and considerably less fast response storm runoff than the Tilton watershed. In order to see if the soil based parameter estimation

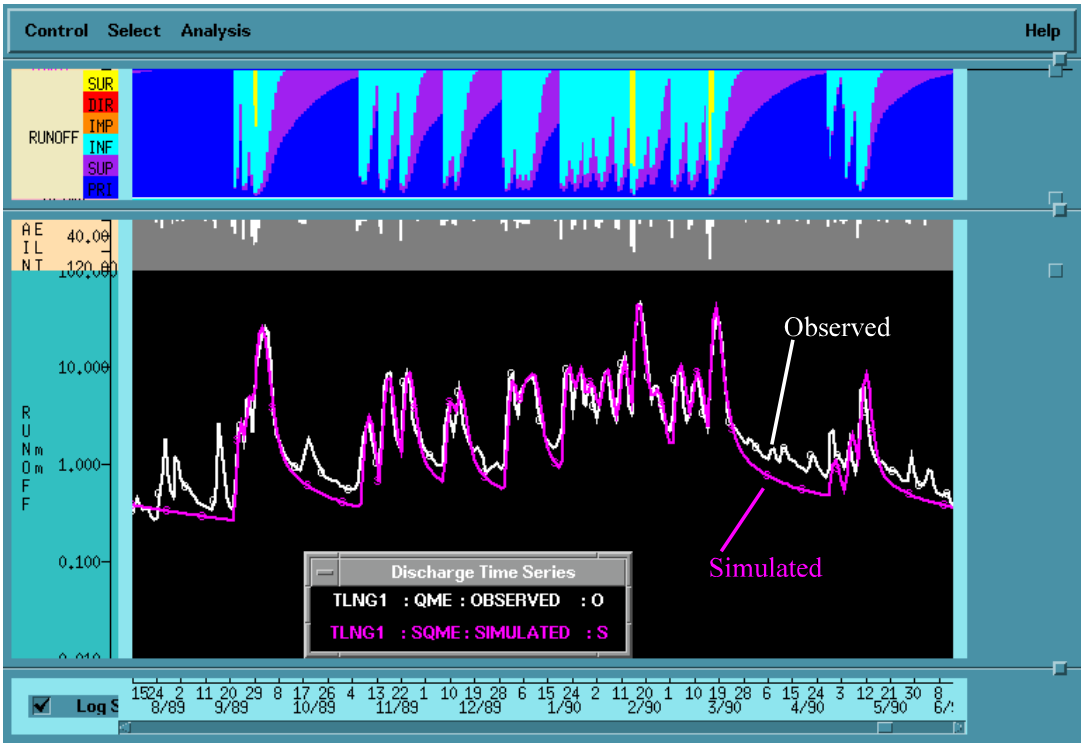


Figure 7-9-1. Base Simulation for Tilton semi-log plot.

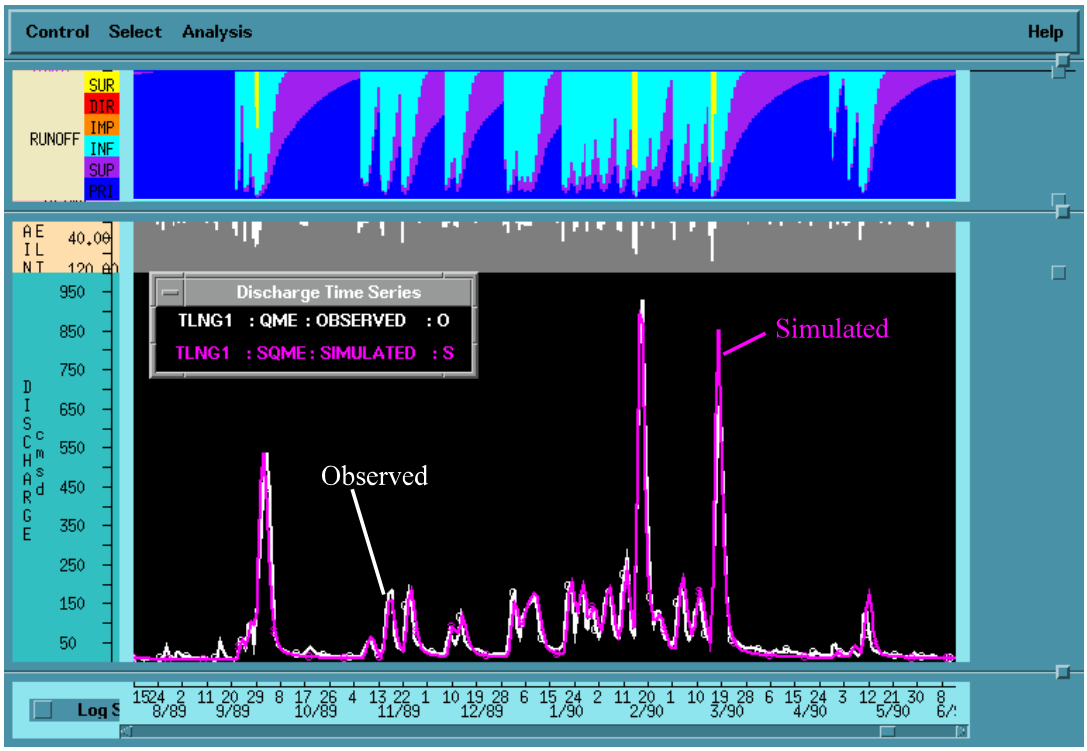


Figure 7-9-2. Soil Based Simulation for Tilton - arithmetic plot.

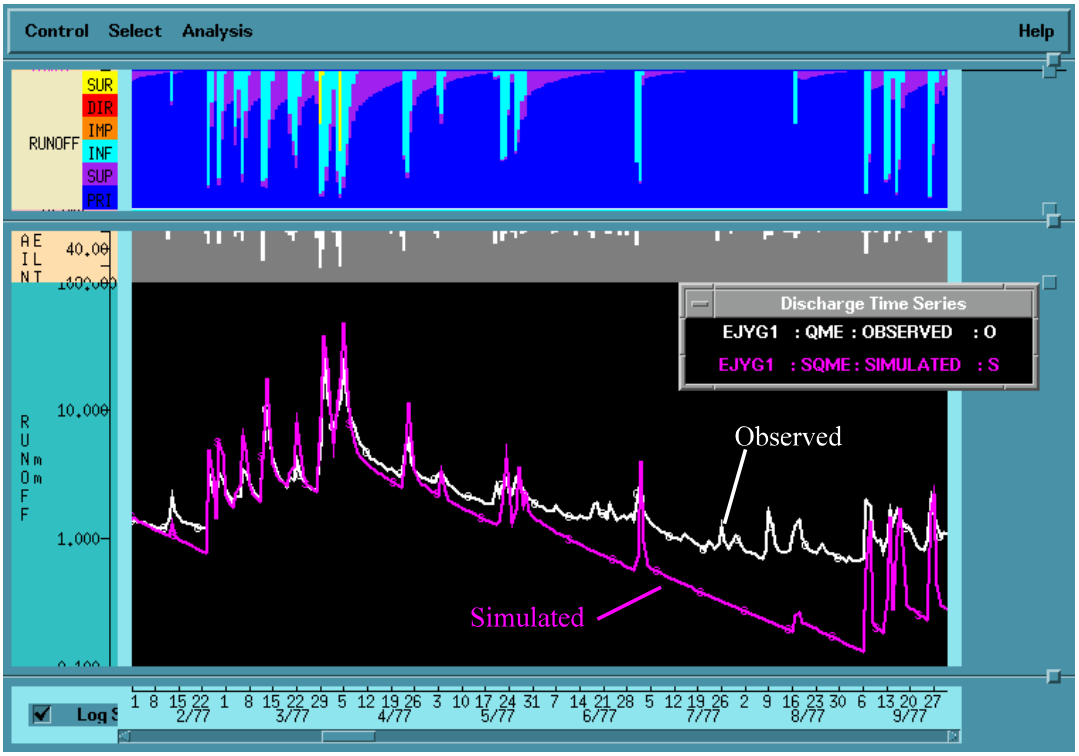


Figure 7-9-3. Soil Base Simulation for Elijay - semi-log plot.

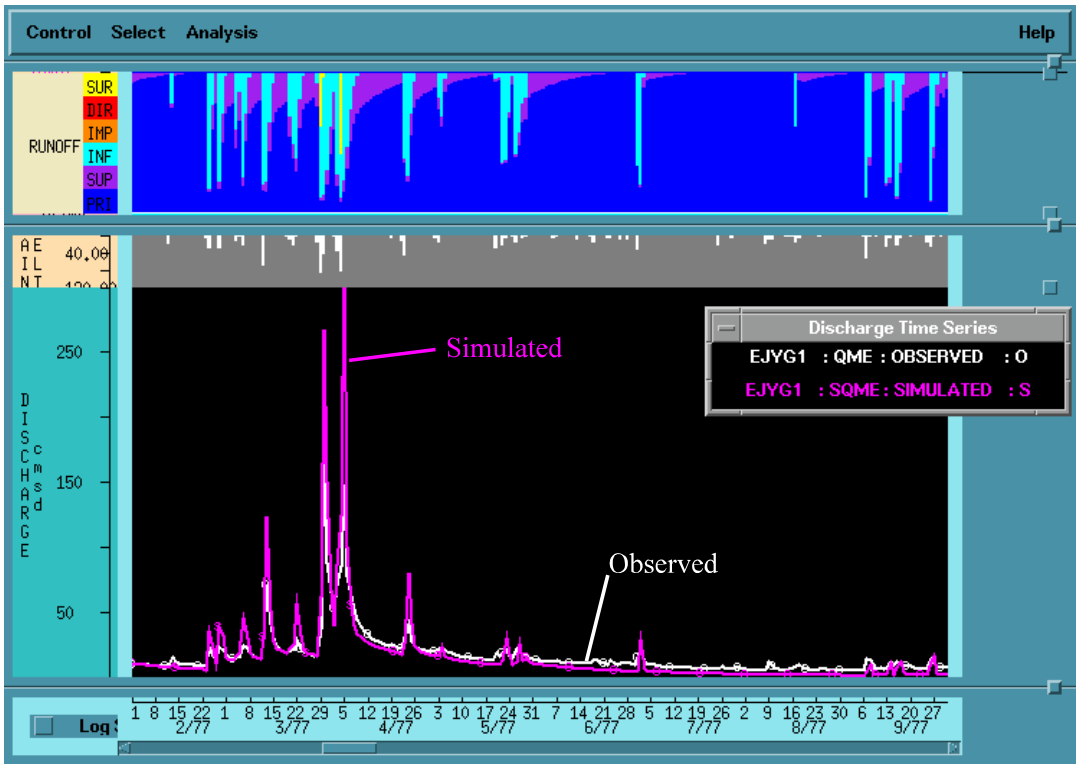


Figure 7-9-4. Soil Based Simulation for Elijay - arithmetic plot.

method could reproduce these differences, the method was applied to the Coosawattee River near Ellijay (EJYG1). Figures 7-9-3 and 7-9-4 show semi-log and arithmetic plots comparing the soil based simulation to the observed flows for this watershed. These plots indicate that the soil based procedure is not able to detect the variability in hydrologic conditions that exist between Tilton and Elijay. The soil based simulation for Elijay generates too much surface runoff and not nearly enough baseflow. The primary baseflow recession is much too fast and supplemental baseflow is practically non-existent.

The next basin examined is the Illinois River above Tenkiller Dam in Oklahoma. Figure 7-9-5 shows a semi-log plot of the soil based parameter simulation versus the observed flow for a portion of the period of record for the Barren Fork at Eldon (ELDO2). The figure shows that the soil based parameters are fairly reasonable and would at least be useful as initial estimates. Some adjustments are needed to the baseflow withdrawal rates and storage capacities. Also the interflow withdrawal rate appears to be too fast. Figure 7-9-6 shows a comparison of observed flows at Eldon and the Illinois River at Watts (WTTO2). This figure indicates that the response of these two watersheds is very similar. Figure 7-9-7 shows a comparison of the soil based parameter simulations for the same two watersheds. The parameters and the resulting hydrographs are basically the same. Thus, in this case where there is little difference in the response of two nearby watersheds, the soil based method correctly indicates that conditions are similar.

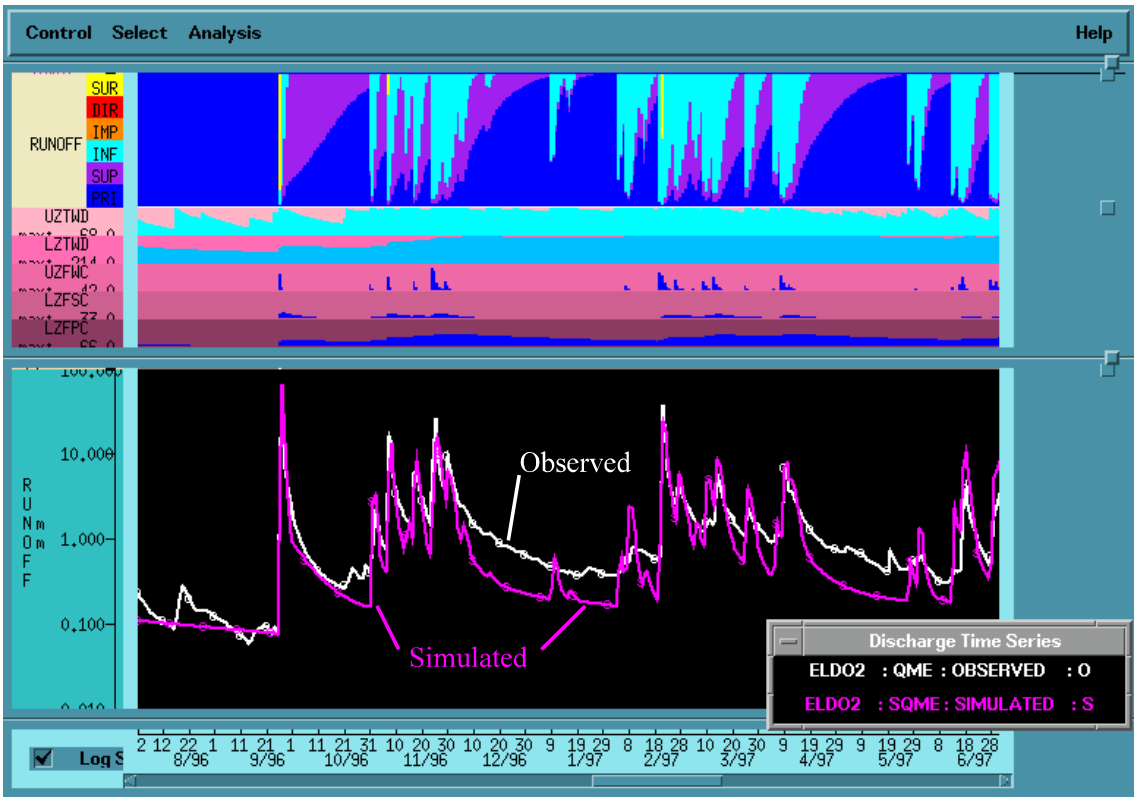


Figure 7-9-5. Soil sediment simulation on Eldorado Basin for selected plot.

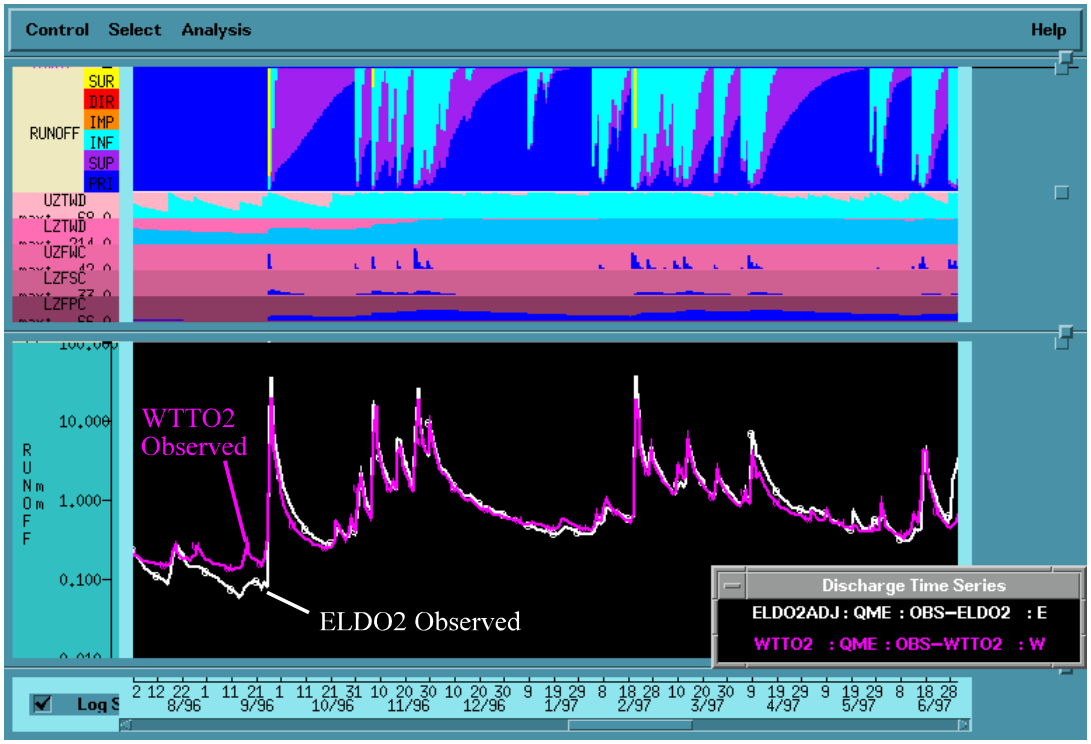


Figure 7-9-6. Comparison of Observed Flows for Eldon and Watts.

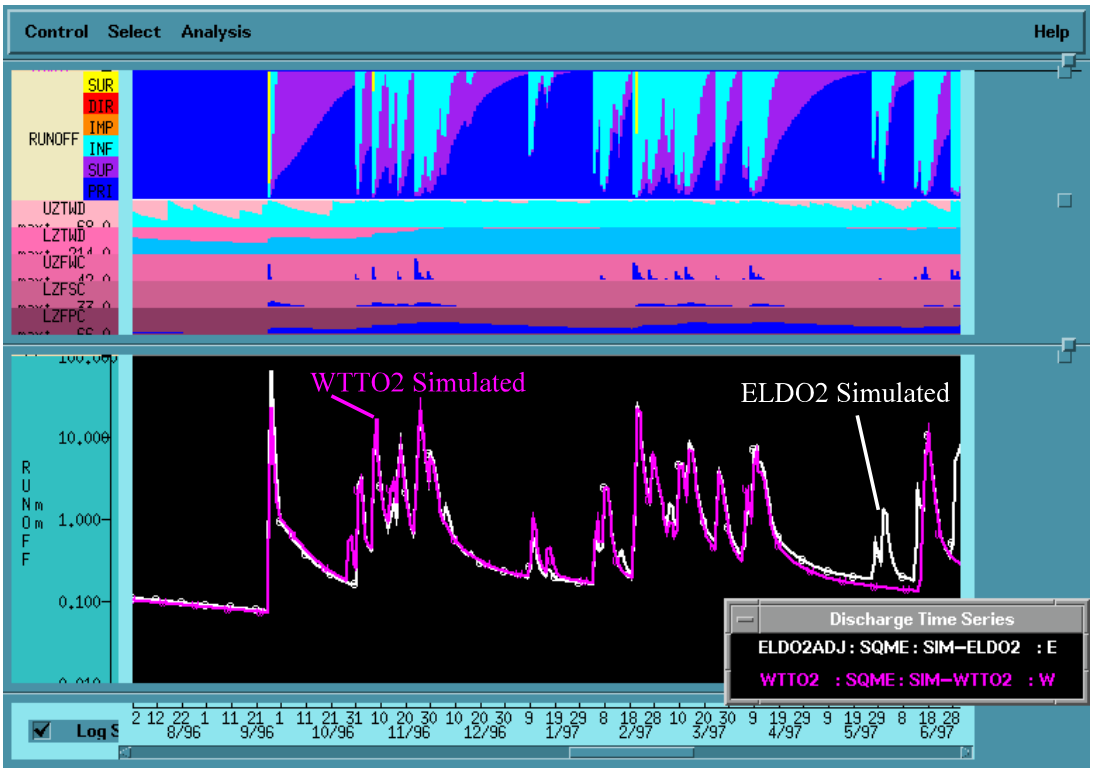
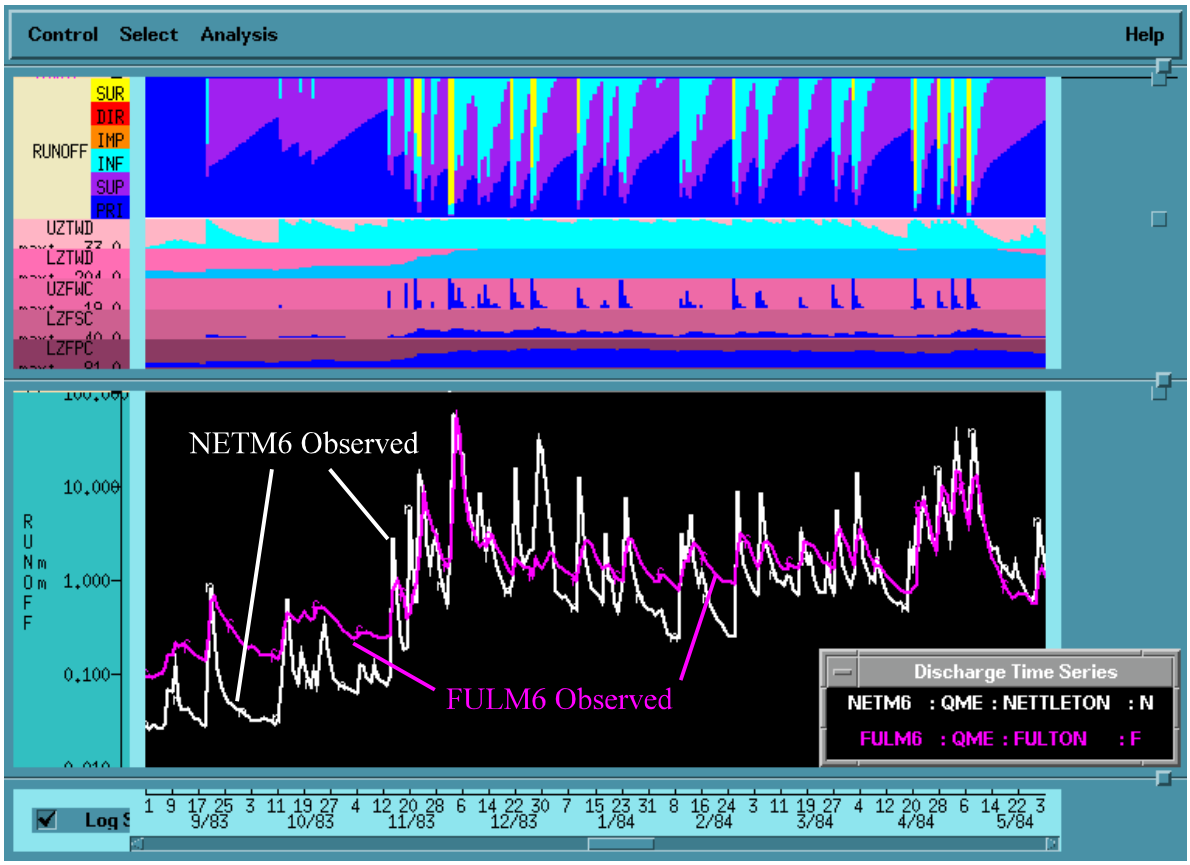


Figure 7-

9-7. Comparison of Soil Based Simulations for Eldon and Watts.

The next basin evaluated was the Upper Tombigbee River in northeastern Mississippi. Figure 7-9-8 shows a comparison of the observed streamflows for Town Creek at Nettleton (NETM6) and the Tombigbee near Fulton (FULM6). The response of these two adjacent watersheds is fairly different. Town Creek is mostly within the region called the Black Belt that extends from central Alabama into northeastern Mississippi and contains very black, fertile soil. The Fulton drainage is just outside of this region. The soil based parameter estimation method recognizes that the soils are different between the two watersheds in terms of generating somewhat different values for a number of the parameters (e.g. UZTWM - 33 mm vs. 42 mm, UZFWM - 19 mm vs. 33 mm, UZK - .30 vs. .46, ZPERC - 93 vs. 66, LZFPM - 81 vs. 125, LZSK - .08 vs. .12, LZPK - .010 vs. .020, and PFREE - .37 vs. .26 – the value for Nettleton is listed first), however, as shown in Figure 7-9-9 the simulations using the soil based parameters are quite similar. The simulated hydrographs are much more like the response of the Fulton watershed than Town Creek. Thus,



even though individual parameter values vary from one watershed to the other, the resulting soil based simulations do not reflect the differences between the two drainages.

Figure 7-9-8. Comparison of Observed Flows for Nettleton and Fulton.

The next basin tested was the Merrimack River basin in New Hampshire and Massachusetts. Figure 7-9-10 shows a semi-log plot comparing soil based simulated flows versus observed values for the Smith River near Bristol, New Hampshire (BRSN3). Figure 7-9-11 shows an

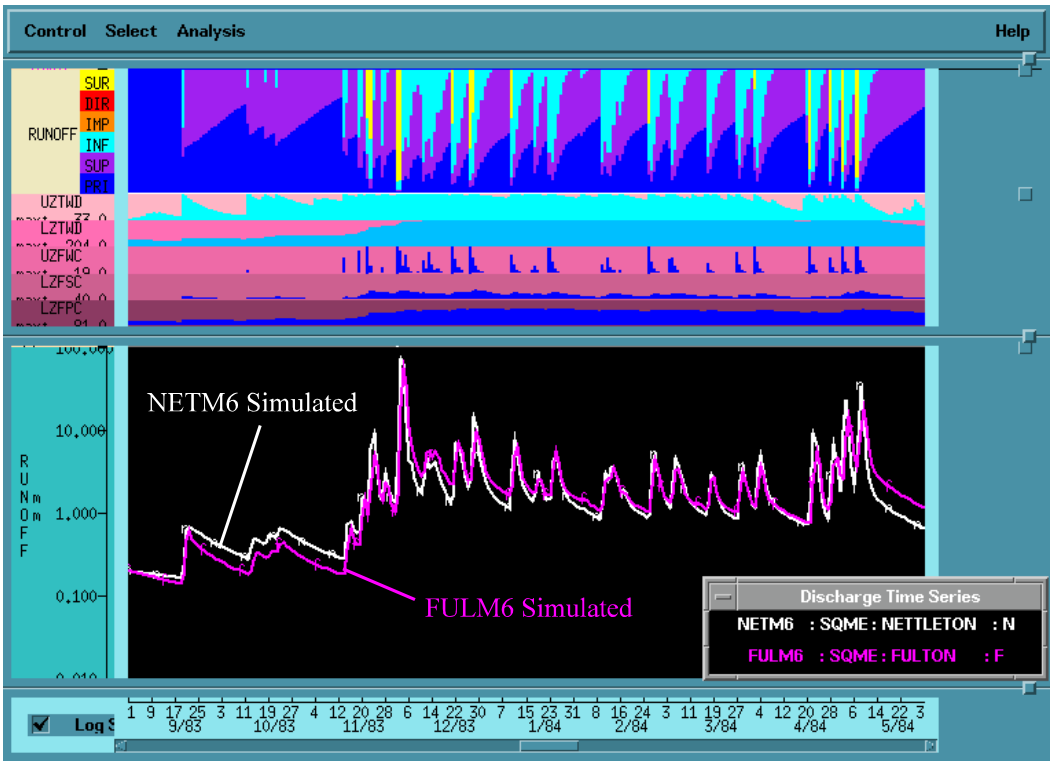
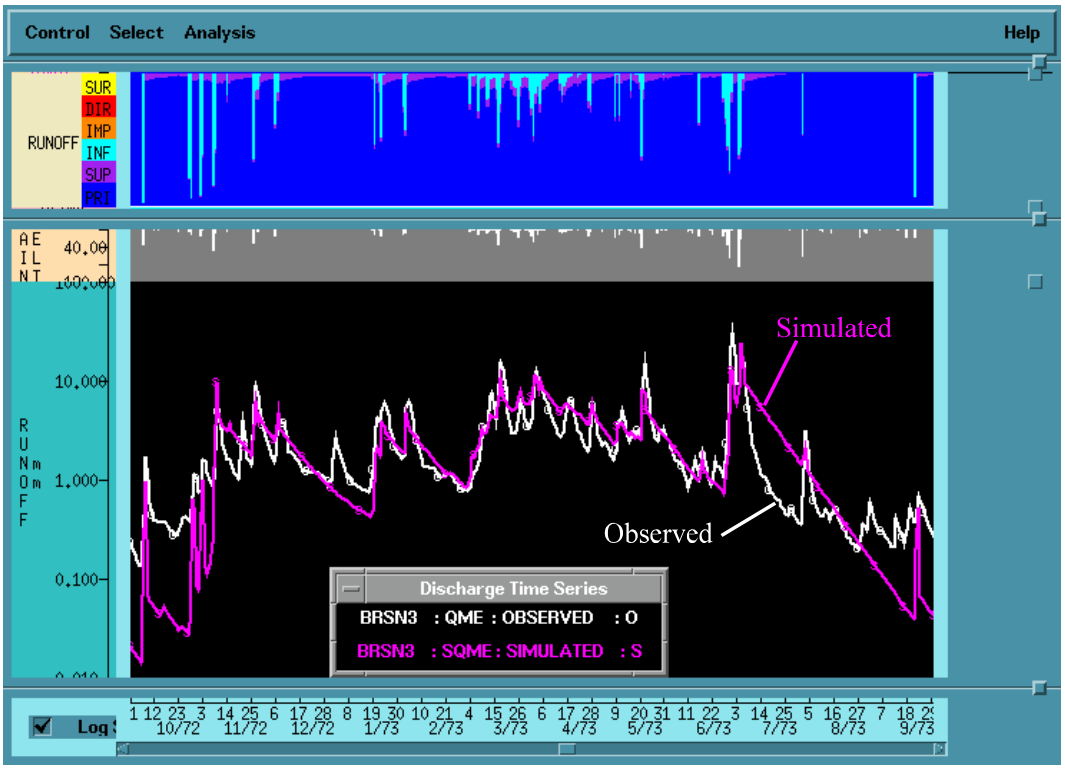


Figure 7-9-9. Comparison of Soil Moisture Simulations Based on Nettleton and Fulton.



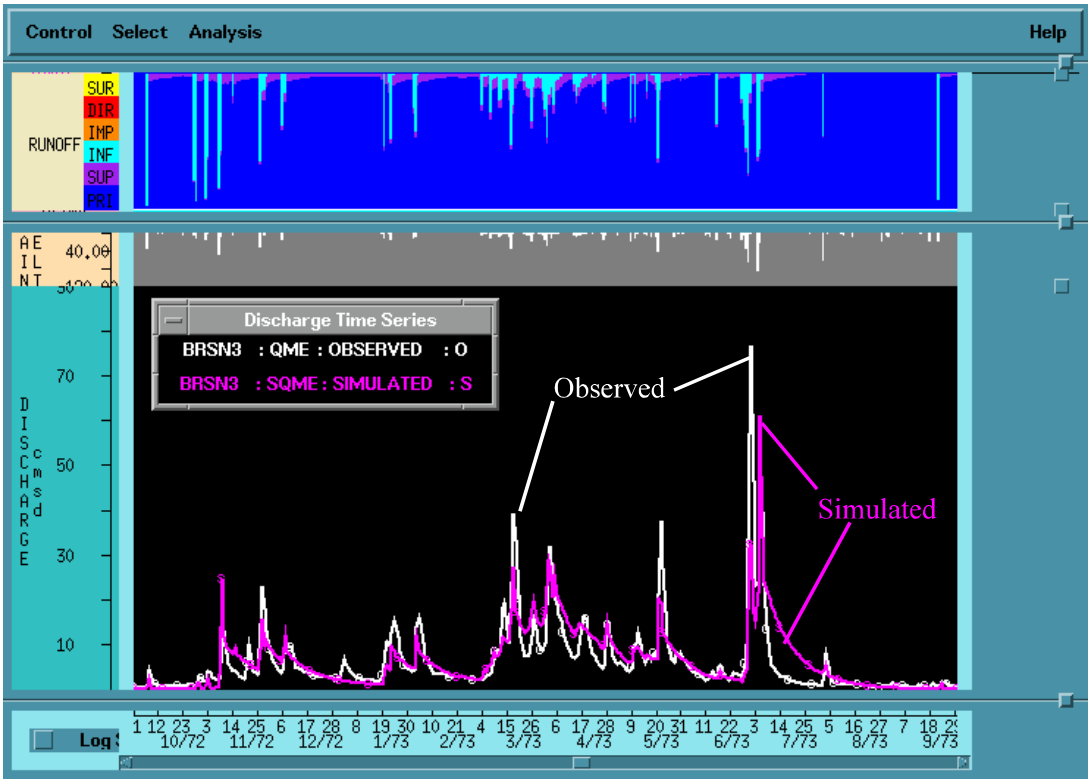


Figure 7-9-10. Soil Based Simulation for Bristol - semi-log plot.

Figure 7-9-11. Soil Based Simulation for Bristol - arithmetic plot.

arithmetic plot for the same period. For this watershed the soil based parameter values are not physically realistic. Almost all of the flow is primary baseflow with some fast interflow during storm events. In reality the primary baseflow withdrawal rate should be much slower, the supplemental baseflow contribution should be considerable, there should be much more interflow, though at a slower withdrawal rate, and the largest events should produce surface runoff. Similar results were obtained when applying the soil based method to two other watersheds in the Merrimack drainage; the Pemigewasset River near Woodstock, NH and the Assabet River near Maynard, MA. Thus, the soil based method did produce similar parameter

values throughout the Merrimack basin as would be expected based on Figure 4-2, however, the values were not reasonable.

The last watershed examined was the Gallatin River near Gallatin Gateway, Montana (GLGM8). Figure 7-9-12 shows a semi-log plot of the comparison between the soil based simulation and observed flows for this two elevation zone, western, snowmelt runoff dominated drainage. Figure 7-9-13 contains an arithmetic plot for the same period. As for the Merrimack basin, the soil based parameters for this watershed are not physically realistic. There is too much primary baseflow and the withdrawal rate is much too quick, there is not nearly enough supplemental baseflow, and the amount of interflow is too great and too fast.

For these five basins the results from the Koren soil based parameter estimation procedure are

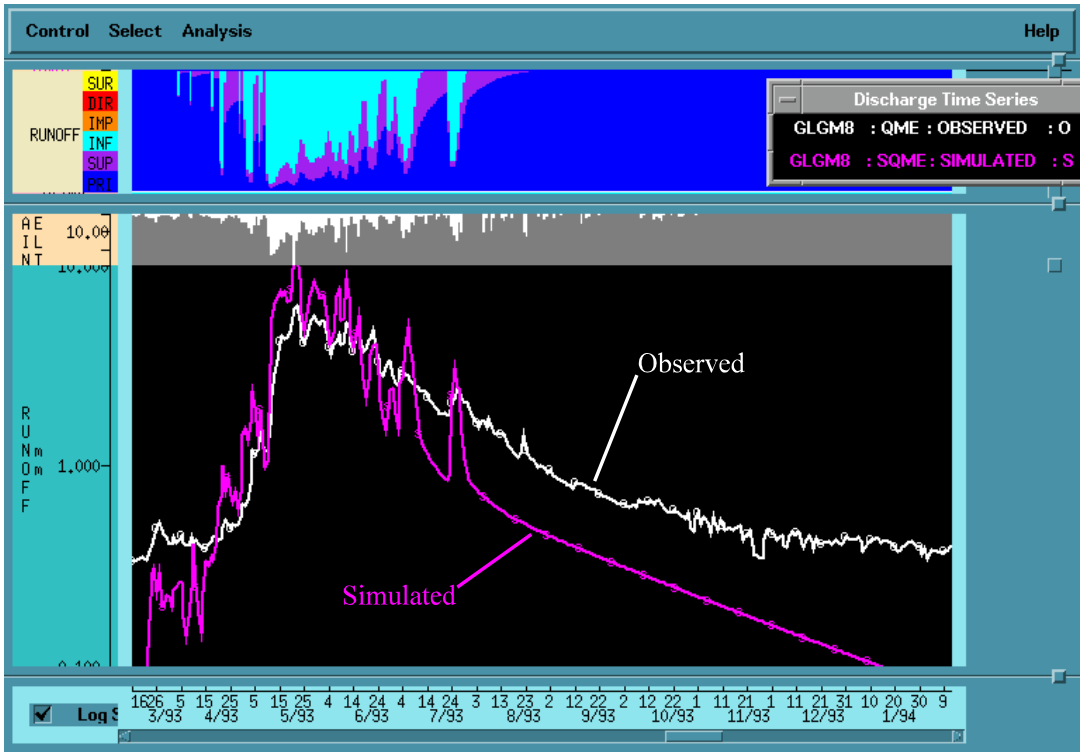


Figure 7-9-12. Soil Base Simulation for Gallatin Gateway - in-log semi plot.

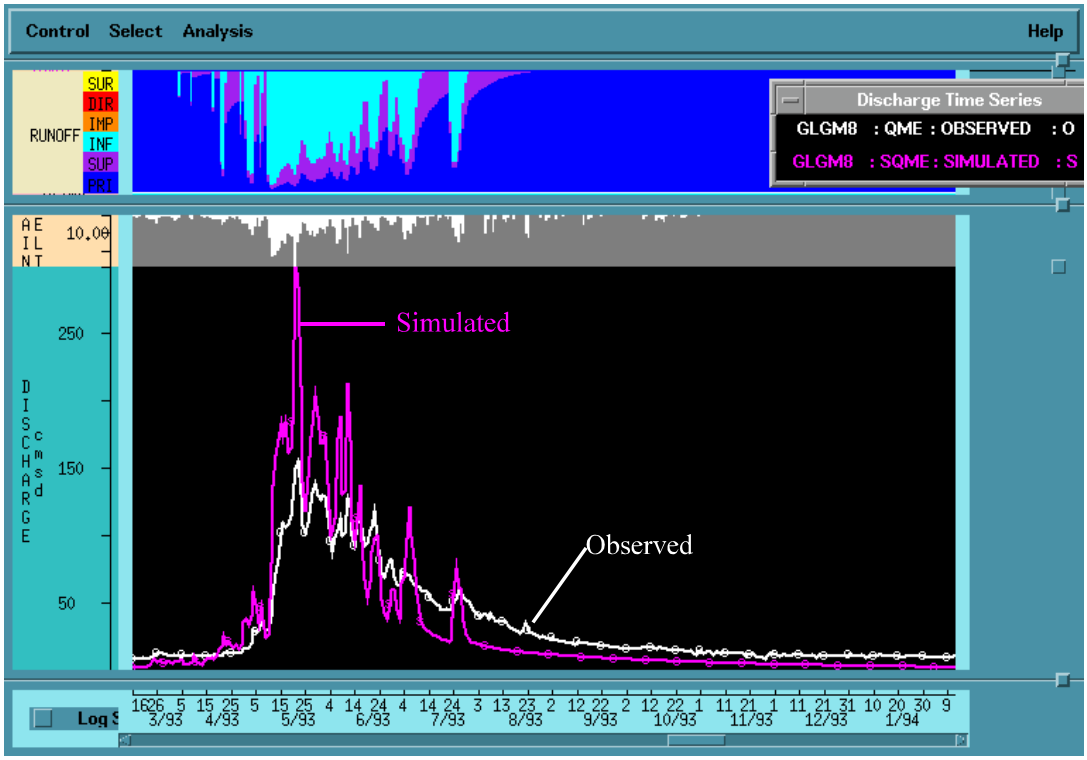


Figure 7-9-13. Soil Based Simulation for Gallatin Gateway - arithmetic plot.

mixed. In the three river basins scattered across the southern part of the country, the Oostanaula, the Illinois, and the Tombigbee, the soil based method generated values that at least for one watershed in each basin produced fairly realistic parameter values, however, in the two cases where there were differences between the response of adjacent watersheds, the simulations using the soil based parameters didn't reflect this variability. For the other two basins, the Merrimack and the Gallatin, the parameter estimates from the soil based method were not physically realistic and wouldn't be useful even as initial parameter values. It should be noted that, at least in this case, all of the watersheds that exhibited the worse results with the soil based parameters, the Gallatin, Merrimack, and Elijay, were in mountainous areas with considerable forest cover. This could just be coincidence. Overall, the greatest weakness of the soil based method seems to be in determining appropriate values for the free water storage parameters, both the upper and lower zone withdrawal rates and capacities. These are parameters for which reliable estimates can often be obtained by analyzing the observed hydrograph as described in Section 7-5.

Summary

It is critical to carefully evaluate any *a priori* parameter estimation procedure before using it in a given region. Such procedures may work well in some regions and not in others. It is also important to verify whether the procedure will serve the purpose for which it might be used. A given method might provide reasonable initial parameter values for a calibration for some watersheds, but not adequately reflect the variability in parameter values across the region. It is recommended that a given procedure be evaluated by comparing *a priori* parameter simulations to observed flows for several watersheds in the region. The emphasis of the comparisons should be to determine if the parameter values are realistic by looking at the portions of the hydrograph where each parameter can be isolated and not by comparing overall 'goodness of fit' statistics.

Chapter 8

Step 6 - Operational Implementation of Calibration Results

Introduction

A properly performed historical data analysis and model calibration should have significant benefits for the operational application of the models and procedures as mentioned in Chapter 1. To repeat, these are:

- Short term river forecasts should more closely track observations, thus improving forecast accuracy and lead time and requiring fewer adjustments by the forecaster.

- Models can be used to generate reliable extended probabilistic predictions.

However, these benefits can be substantially diminished if the operational system is applied in a manner that produces biased results compared to the calibration. There should be differences in the random variations between operational and calibration simulations (ideally the operational results will exhibit less random error), but there shouldn't be a bias.

The primary purpose of historical data analysis is to generate unbiased estimates of the model input data, as well as the output data that will be used to validate the simulations. The aim is also to have the minimal amount of random noise given the available historical data networks. These data are then used to calibrate the hydrologic models to simulate streamflow and other variables throughout a river basin. The primary purpose of calibration is to obtain the parameter values for each model that will generate the most unbiased simulation possible. Calibration also aims at producing parameter values that properly represent the physics that are occurring and are consistent from one watershed to another. The purpose of model calibration is not to produce the best possible value of some "goodness of fit" statistic, but instead to determine physically realistic and consistent parameter values that will allow the models to be used to generate the best possible results in an operational application.

The primary purpose of operational forecasting is to produce an unbiased forecast of what will occur in the future with minimal random variation. In order to use the calibration results to accomplish this goal, several things must occur.

- The operational data estimates must be unbiased compared to those used for calibration.

- The models are defined such that the basic simulation (i.e. the simulation produced using the same input data types and model configuration used for calibration with no alternative data analyses, modeling schemes, or real time adjustments included) will be unbiased compared to that obtained during calibration.

- Real time analysis procedures, that in many cases involve data not available or examined during the historical processing step, are applied in a way that reduces the random error in the data estimates, without producing a bias.

- Adjustments are applied to the models that reduce the random variations and unavoidable trends that were in the calibration simulations in a manner that doesn't produce bias in the forecasts.

This chapter describes how to use an operational forecast system so that the real time results will be unbiased compared to calibration simulations. There is some discussion of ways to reduce random error, but the emphasis is on minimizing bias. As with the rest of this manual, the concepts might apply to any forecast system that relies on conceptual hydrologic models, but the discussion is focused on the application of the NWSRFS Operational Forecast System (OFS). It should be clearly stated that this chapter is not intended as a complete guide on how to produce operational river forecasts, but only to describe what must be done to insure that the benefits of calibration are preserved.

The chapter will be divided into two main parts. The first part will indicate what needs to be done to define the basic operational system so that it will generate results that are unbiased compared to those produced during calibration. The amount of random variation will differ from that experienced during calibration primarily due to differences in the density and distribution of the data networks, primarily precipitation. The second part will discuss various procedures and techniques that can be used to reduce the random errors in the data input and model results. The focus will be on applying these procedures and techniques in a manner that will minimize adding any bias.

Operational Definition using Calibration Results

Introduction

In NWSRFS various information needs to be specified in order to use the Operational Forecast System (OFS). This includes the definition of :

- all stations with their available data types and related parametric information,

- all of the areas for which mean values of precipitation, temperature, and evaporation are to be computed along with their parametric information, and

- all segments with the sequence of operations (models and procedures needed to generate simulations of streamflow and other variables of interest) and the parametric information for each operation.

Most of the parametric information needed when defining the operational system are obtained

directly from the calibration results or are based on values derived during the calibration. The material presented in this part of the chapter describes guidelines for defining the operational system so that it should very closely mimic the results obtained during calibration. In this definition no changes are made to the time and space scale that the data analysis procedures and hydrologic models are applied or to the types of data used during calibration. Recommendations for using different data types and analysis methods or varying the time and space scales will be mentioned in the second part of the chapter. These may involve making changes to the basic definition of the operational system described in this part. Except for variations caused by differences between the density and distribution of the real time and historical networks, the operational system as defined by following the guidelines in this part of the chapter should produce, with no run time modifications applied, essentially the same simulation results as the calibration system.

Operational Station Definitions

Station definitions can be one of the major reasons for operational results to be biased as compared to calibration simulations. This is primarily due to differences in the real time and historical data networks. It is common for stations to be available in real time that were not included in the historical records and vice versa. Bias can occur due to an improper specification of parametric information for stations that are only available operationally, equipment at sites with only operational data being incompatible with the gages used for calibration, inconsistencies that occurred after the historical data period, and differences in the reporting characteristics of the networks. The best case scenario would be for the operational and historical networks to be comprised of exactly the same stations with no changes in location, equipment, or reporting characteristics, however, this is seldom the case. Thus, it is critical to make sure that all the real time data are compatible with the station data used for calibration.

When data for which monthly mean values must be provided operationally (mountainous area precipitation stations and all temperature stations) are checked for compatibility, the checks are always made based on the full period of record of the historical data, i.e. the period for which the data were analyzed and processed, not the period used to calibrate models at any specific location. This can present a dilemma when different historical periods of record are used for different portions of an RFC area. Generally the same historical data period is used for an entire river basin. Ideally all the river basins that utilize the mountainous area precipitation analysis procedure or use temperature data within the RFC area of responsibility would have the same historical data period. However, this is likely not the case due to such factors as differences in networks over time (e.g. the SNOTEL network record length might only impact a portion of the RFC area - see Chapter 5) and the length of time involved in calibrating an entire RFC area (i.e. last basin calibrated would have more historical data available than the first basin). Stations that are used only for computing mountainous area MAP and any MAT time series within a single river basin are not affected. The dilemma occurs for those stations that are near the divide between two or more basins that have different historical data periods. Mean monthly values for these stations can vary from one historical period to another. Hopefully the differences are not

too great. This will likely be the case if the historical data periods are quite long. The best that probably can be done when operationally defining these stations is to use the average, or a weighted average of the mean monthly values for each historical period depending on which basin the station has the most influence.

Though there are similarities in the methods of checking for compatibility from one data type to another, there are also differences, thus each major data type will be discussed separately.

Precipitation Data

Precipitation is the single most important input to the snow and soil moisture accounting models. As shown in Chapter 6 any bias in precipitation will be magnified in the runoff computations. Thus, it is critical that the precipitation observations used operationally be as unbiased as possible when compared to that used in calibration. In order to avoid such bias, all station precipitation data used operationally must be compatible with the historical station data. The methods for doing this depend on whether a given station is part of a mountainous or non-mountainous area procedure for computing MAP. Stations that can be used in both types of procedures should be checked using each method.

Some RFCs use either non-mountainous or mountainous area procedures over their entire area, however, in some cases the procedure used to process precipitation data varies from one part of an RFC area to another. This is especially true when there is a clear demarcation between the relatively flat terrain and mountains. During calibration either one procedure or the other is used for any group of stations in a precipitation analysis (i.e. either monthly means are defined for all stations or they are not). However, during operations, stations in the transition zone from flat to mountainous terrain could be used for areas that use both types of procedures. For operational precipitation station definitions, any station that can be assigned weight, or used to estimate missing data for a station that can be assigned weight, for an MAP area that uses the mountainous area procedure should have mean monthly precipitation amounts (often referred to as characteristics) included in its definition. Typically this means that all stations in the mountains will have mean monthly values defined, as well as any stations in the adjacent flat lands that can be used to estimate missing data at a mountain station. The mean monthly amounts should be those that are appropriate for the historical period of record used for the mountainous areas. If there is not a clear demarcation between the flat terrain and the mountains, it is usually easier to include mean monthly values for all stations.

- Non-Mountainous Areas - As defined in Chapter 6, a non-mountainous area is an area where the long term average values of the variable being analyzed are essentially the same at all locations. The compatibility checks to be performed for stations in non-mountainous areas depend on whether the station was part of the historical analysis. For real time stations that were part of the historical analysis, the user needs to verify that there have been no station moves or equipment changes since the end of the historical

analysis period. Equipment changes would involve substituting a different type of precipitation gage or adding or removing a wind shield. If this is so, then it is reasonable to assume that the real time data should be compatible with that used for calibration. If the station was moved or the equipment altered, then the consistency of the record should be checked by doing a double mass analysis against a group of other precipitation stations in the vicinity. If an inconsistency is found, the appropriate consistency correction should be applied in the operational station definition (i.e. a correction that will make the real time data consistent with the historical period - e.g. if the gage now catches more than during the historical period, the operational consistency correction would be less than 1.0). In the OFS real time precipitation corrections can be applied on an annual or seasonal basis.

For real time stations that were not part of the historical analysis, a check should be made that the long term average precipitation for those stations is essentially the same as the stations that were available historically. It might be assumed that since long term average precipitation is the same, or a general trend of slowly increasing or decreasing values exists, over the region, that any new station will have a catch that is consistent with the gages used in the historical analysis. This may not be true due to site exposure or the type of equipment used. Windy, open sites or locations with nearby obstructions might under or over catch precipitation. Especially in regions where snow is important, windy, open sites should be avoided unless the gage is properly shielded. In addition, various studies have shown that the catch from certain types of gages are not compatible with the catch from standard climatological stations. For example, it is recognized that heated tipping bucket gages catch less precipitation, especially snowfall, than a universal weighing gage. Different types of gages can also catch varying amounts of rain. A study by the USGS west of Chicago [*Straub and Parmar*, 1998] compared the catch from 5 NOAA climatological gages (8 inch nonrecording and universal weighing gages) and 10 USGS tipping bucket gages. Using only periods when rain occurred, the USGS gages caught only 86% of the amount recorded by the NOAA gages (the tipping bucket gages were adjusted during periods of intense rain as per manufacturer's specifications). When one gage of each type was installed at the same location for a short period, the USGS gage recorded 91% of the catch of the NOAA gage. The inclusion of a number of such gages in the operational network could result in long term MAP values being biased by 5-10%, depending on the weights assigned to each station, as compared to those used for calibration. This could result in a 10-25% bias in runoff computations.

Before using any stations operationally that were not part of the historical network, it should be verified that the catch from these stations is compatible with the historical data. The method used to do this is to compute the ratio of the catch at the new station to one or more nearby stations that were part of the historical network which haven't had any relocations or equipment changes. This should be done over some period of time in order to remove the effect of spatial variations within individual storms. Typically the ratios are computed using monthly totals from the gages. Generally in a non-mountainous area

one or two years of data should be sufficient to obtain reasonably stable values of the ratios. The ratios should be very close to 1.0. If not, either the new station shouldn't be used as part of the operational network or it should be corrected so that its precipitation measurements are compatible to the historical data. Hopefully a couple of years of past data can be obtained from the new station in order to do the comparison. If not, sufficient data should ideally be collected so that the data can be checked before defining the station as part of the operational network. In many regions there is a real deficiency of real time precipitation reports. In these cases, it may be judged that any new gages should immediately be used to compute areal values even if their use may create a bias. This may be a reasonable decision, but at least the data should be analyzed after a couple of years to determine if an adjustment is needed.

In the western parts of the United States the RFCs have always recognized that there are tremendous variations in amount of precipitation from one location to another in the mountains and have thus always taken this into account when generating model input. In the east, however, there has been a tendency in the past to ignore the effect of the mountains. Station weights were computed for both calibration and operations using weighting schemes such as the Thiessen method that only considers the station location. In many cases the historical and operational networks were somewhat different, though for many years most of the gages were in the mountain valleys and thus the overall effect on MAP computations wasn't too great. In addition, empirical rainfall-runoff models, such as API type models, used in the past were not as sensitive to precipitation bias as conceptual soil moisture accounting models such as the Sacramento model. With the switch to conceptual moisture accounting models and the advent of the IFLOWS network that covers much of Appalachia with gages on the mountain ridges, the continued use of the non-mountainous area procedure for mountain areas will likely cause a considerable bias between the operational and historical MAP values and result in over forecasting most flood events. The mountainous area procedure should be used in all such regions, both for historical and operational precipitation processing.

• Mountainous Areas - In a mountainous area mean monthly precipitation amounts must be defined for all stations. The monthly means used in the operational definition should be those that correspond to the period of record used for the historical data analysis. For real time stations that were part of the historical analysis, the mean monthly values determined using the PXPP program during the historical analysis should be used when the station is defined in the operational system. As with non-mountainous areas, these stations should be checked for consistency if any moves or equipment changes have occurred in the years since the historical analysis period ended. If there are any inconsistencies, corrections should be applied in the operational system so that the real time reports are adjusted to be consistent with the historical analysis period.

For operational stations that were not part of the historical network, mean monthly values must be determined prior to the stations being defined as part of the operational network.

The procedure for estimating mean monthly values for each of these new stations for the historical period of record is as follows:

- compute the ratio of precipitation at the new station for each month to that at one or more nearby stations that were used in the historical analysis and have well defined monthly means (i.e. the monthly means are based on a long period of observed data), and then

- multiply these monthly ratios by the mean monthly amounts for each of the nearby stations used in the historical analysis to obtain estimates of the mean monthly amount for the new station -- then average or weight the estimates if more than one nearby station was used.

Typically monthly totals are used to compute the ratios rather than daily or storm totals. The length of the period needed to get a stable estimate of the ratio for each month varies somewhat with the spatial variability of the precipitation in the region. In regions with strong orographic patterns during most events, such as the Pacific Coast, 3 years of data should be sufficient to obtain a good estimate of the ratios. In regions with considerable convective activity at times during the year, such as the intermountain west, experience with the PXPP program has shown that around 5 years of data are needed before a reasonably stable estimate of the mean monthly values are obtained. The data needed to compute the monthly means might already be archived or it may be that such data will need to be collected before the station can be defined and used in the operational system. If the user wanted to start using the new station immediately, estimates of the mean monthly precipitation could be obtained from the isohyetal analysis after it is adjusted to the historical period of record (see Section 6-3). This should only be done if the user has considerable confidence in the isohyetal analysis. Even then local exposure and equipment peculiarities could cause the isohyetal estimate to deviate from the actual value enough to cause a significant bias in the resulting MAP values. Operationally it is best to only use stations that were not part of the historical analysis after sufficient data are available to get a good estimate of the mean monthly precipitation at these sites.

For all types of areas, differences in how historical and real time precipitation data are reported can also produce a bias between the resulting MAP time series. In the past many non automated stations in the real time network used a 0.5 inch reporting criterion, i.e. the observer only was to begin reports after more than that amount of precipitation occurred during a storm. Especially during days with spotty convective activity, there was a definite tendency to significantly overestimate the missing amounts at those stations that didn't report using the few stations that did receive more than 0.5 inches of rain. Undocumented studies at the Southeast RFC and HRL showed that the average bias during summer months in regions with a lot of this type of convective activity and many criterion reporting stations could reach nearly 100%. It could not even be assumed that when a station didn't report that it must have had less than 0.5 inches of precipitation as surveys at two RFCs indicated that such stations

only reported about half the time when they should.

The 0.5 inch reporting criterion is no longer used, however, there still are differences, some perhaps very subtle, in how historical and real time data are reported that could produce biased operational precipitation estimates. A Forecast Systems Laboratory (FSL) study [*Tollerud*, 2001] and conversations with personnel from some RFCs indicate that there can be substantial differences in the number of reports from observers on days with light or scattered amounts as opposed to days when most stations receive some precipitation. It appears that many manual observing sites only report on days when precipitation occurs. The FSL study showed that there were dramatically different frequency distributions of the percent of rain days between daily reporting stations and automated sites. There were even some differences between automated real time sites and stations in the hourly climatological network. The study also showed that a sizable number of the automated stations have 1-4 missing hourly observations per day. Such real time reporting characteristics, especially having many stations not reporting on days with scattered convective activity, could easily result in biased estimates of station precipitation amounts. This is clearly something that needs to be carefully monitored and evaluated as a significant bias could be produced in many cases when there are substantial differences in how real time and historical precipitation data are reported.

Temperature Data

Biased temperature estimates will have a significant effect on the timing of snowmelt runoff as shown in Chapter 6. They will also have a negative effect on the determination of the form of precipitation. Since for all temperature stations the mean monthly maximum (max) and minimum (min) values must be defined in the operational system, the key, as far as the station definitions, is to make sure that the mean monthly values provided are compatible with the period used to perform the historical data analysis. It doesn't matter whether a given station is to be used for a mountainous or non-mountainous area or both.

For operational stations that were included in the historical analysis, the mean monthly max and min temperatures provided for the operational definition should be exactly the same as those used in the final MAT run for the historical period of record. If any of these stations have been relocated since the end of the historical data period, then the consistency of the record should be checked and if necessary, corrections applied in the operational definition to adjust the current data so that it is compatible with the data for the historical period.

For operational stations that were not part of the historical network, mean monthly max and min temperatures for the historical period need to be computed. The procedure for estimating these values for each new station is as follows:

- compute the difference in max and min temperature for each month between the new station and one or more nearby stations that were used in the historical analysis and have

well defined monthly means (i.e. the monthly means are based on a long period of observed data), and then

- add these differences to the mean monthly max and min temperatures for each of the nearby stations used in the historical analysis to obtain estimates of the mean monthly max and min temperatures for the new station -- then average or weight the estimates if more than one nearby station was used.

The differences should be computed using monthly average max and min temperatures to avoid the effects of different observation times and frontal passages. Only months with complete records should be used. Monthly differences, rather than long term, should be used because the relationship between stations can vary seasonally, even in non-mountainous areas. Typically a couple of years of data should be sufficient to determine stable temperature differences since temperature can be observed every day as opposed to precipitation which occurs only periodically and because temperature is not as spatially variable as precipitation.

Evaporation Data

Evaporation data only need to be defined in the operational system for stations where Potential Evaporation (PE) is computed on a daily basis from meteorological factors. If the models only use climatological estimates of mean evaporation rates, these values are supplied with the other model parameters in the segment definition. The primary cause for daily operational PE estimates to be biased compared to the values used in the historical analysis are differences in how solar radiation data are obtained. In addition, an incorrect specification of the anemometer height in the operational definition will also produce a bias. The user needs to make sure that the correct anemometer height is provided as part of the operational station definition and that this value is updated whenever the height is changed.

The only currently available historical daily PE estimates from NHDS are based on solar radiation being derived from manual observations of sky cover. Historical daily PE estimates can be obtained for a selected group of synoptic stations scattered across the country up until the time that each station's ASOS system was commissioned. Once ASOS is commissioned at a station, manual sky cover observations are no longer available. Some automated sky cover values are included in the variables observed by ASOS, but these are not compatible with the method used to estimate solar radiation from manually observed sky cover. In order to estimate solar radiation from manual sky cover observations a parameter, referred to as the B3 parameter, must be supplied. Also a correction factor is needed that adjusts the computed PE values so that they give the same annual lake evaporation as shown in NOAA Technical Report 33 for the location of the station (see section on "Determination of PE" in Section 6-5). These values are automatically provided when historical estimates of PE are produced using the NHDS. If manual sky cover observations were still available at a synoptic station, both the B3 parameter and the appropriate correction factor should be supplied in the PE

section of the OFS station definition.

There are other ways of obtaining daily solar radiation data to use for operational daily PE computations. As mentioned in Section 6-5, percent sunshine observations have been shown to produce estimates of solar radiation that will generate unbiased estimates of PE as compared to the Technical Report 33 annual lake evaporation. Thus, operational estimates of PE using percent sunshine to estimate solar radiation should be unbiased as compared to historical estimates of PE available from NHDS. Solar radiation can also be measured directly. Such data are not readily available at many locations. There is also a method of obtaining daily solar radiation from satellite data. Whatever data are used operationally to obtain solar radiation estimates for use in computing daily PE, the user needs to make sure that over the long term the resulting annual PE values are essentially the same as those shown in Technical Report 33. If not, the PE values being used operationally are likely biased compared to those that were used for calibration. As discussed in Chapter 6, biased PE values will result in a bias in runoff computations. Such a bias will especially have a significant effect on the soil moisture deficits that are produced after a dry period and thus, can have a major effect on the runoff produced by a subsequent storm event.

Operational Area Definitions

Generally bias in the operational estimates of areal inputs for the models is the result of problems with station data or incorrect information supplied when defining the stations. However, bias can also be produced if the areas are not defined properly. As with station definitions, each data type will be discussed separately.

Mean Areal Precipitation (MAP)

For non mountainous areas the operational MAP time series should be generally unbiased as compared to the historical values as long as the stations involved are checked for consistency and compatibility and the reporting characteristics of the network don't result in biased station estimates, especially on days with spotty convective activity. A slight bias is difficult to completely avoid when there are differences in the composition of the two networks. However, if the station data are properly selected and corrections applied when needed, the resulting bias in the MAP values should be no more than a couple percent as long as the reporting characteristics of the network don't cause a more serious problem.

For mountainous areas, the operational MAP values shouldn't be biased if the appropriate mean monthly precipitation is determined for the stations that were not part of the historical analysis, the stations are checked for consistency when moves or equipment changes occur, network reporting characteristics don't result in biased station estimates on days when precipitation is not occurring over most of the area, and the areas are properly defined. For each MAP area that uses the mountainous area procedure, the exact same average mean areal precipitation that was used for computing historical station weights with Eq. 6-3-4 should be

used to compute operational station weights. This can be a seasonal or annual average depending on what was used during the historical analysis. Just as for the historical analysis, Eq. 6-3-4 is used to compute the operational station weights. If the operational network differs from the historical network, the relative weight assigned to each station may be different, but this will not cause a bias. The guidelines for selecting relative weights mentioned in Section 6-3 should also be used when determining relative weights for the operational system.

The amount of random error in the areal precipitation estimates will vary based on differences in the density and distribution of stations in the historical and operational networks. However, if the stations and areas are properly defined and the network reporting characteristics don't create a problem, this shouldn't result in a bias between operational and historical MAP values. As mentioned earlier, if either estimate is only based on a single station, the noise will typically be so great that proper application of the models is not possible. Also, in mountainous areas if there are great discrepancies in the distribution of the gages, especially in terms of elevation, from one network to the other, there may not be a long term bias, but the likely statistical differences in the variability of the estimates will cause problems.

Comparisons should be periodically made between the operational MAP values and what would be obtained using the historical network due to the importance of precipitation estimates and because of all the factors that could cause deviations between operational and historical MAP amounts. This requires that the operational MAP values be archived. Such a comparison can be done anytime that there is a sufficient overlap between the archived operational MAP estimates and the available climatological data needed to compute historical MAP values. An ideal time to perform an MAP comparison is when extending the historical data record (record extension discussed at the end of Chapter 6). If archived operational MAP data are available for the extension period, it would be very easy to compare the values. Besides bias comparisons, both annual and seasonal, other statistics could be computed (e.g. based on precipitation amount, frequency, and variability) to determine what differences exist between the operational and historical estimates. Each time series could also be used to produce simulations of streamflow and other variables to assess how any differences affect model performance. Unfortunately such comparisons have seldom been done in the past.

Mean Areal Temperature (MAT)

For non mountainous areas grid point weights are typically used for both historical and operational MAT computations. For any MAT areas where the station weights differ from the operational to the historical network, areal mean monthly temperatures should be computed and compared. This is done by multiplying the weight for each station by its mean monthly max and min temperature to get the weighted mean monthly max and min temperature for the area. This should be done for all months that the MAT values impact

model computations, typically months that can have snow. The weighted mean monthly temperatures should be essentially the same for both networks. If not, either the weights could be modified by using predetermined weights operationally or stations whose mean values deviate significantly from the other stations could be removed from the computations. It is essential that the weighted areal mean temperatures are basically the same for both networks to avoid biased temperature data used in the model computations.

For mountainous areas, the typical procedure is to define a synthetic or "dummy" station that represents the average temperature over the area or the portion of the area that is generally affected by snow. The synthetic station is then assigned a weight of 1.0. When defining mountainous MAT areas in the operational system, it is critical to use the exact same mean monthly temperatures for the synthetic station as were used in the historical analysis. The location of the synthetic station can vary from the historical to the operational system, but not the mean max and min temperatures assigned to the synthetic station. The location of the synthetic station in the operational system should be based on obtaining the best possible estimate of the areal value based on the available real time temperature data. That is, the same logic is used for locating the synthetic station operationally that was used for the historical analysis (see Chapter 6-4).

The historical MAT time series are generated by using only max and min temperature data. As shown on Figures 6-4-1 and 6-4-2 in Section 6-4 the use of just max/min data can produce erroneous MAT estimates not only on days when the diurnal variation differs from the assumed pattern, but also on other days based on when the max and min values are observed. Even though the problem only occurs on certain days, it can cause the historical MAT data to have an overall bias as compared to what would be computed using both instantaneous and max/min values. The NWSRFS operational system uses both types of temperature data. For stations with an a.m. observation time, this causes the historical estimates to be somewhat greater than the operational values on the affected days. For stations with a p.m. observation time, the result is just the reverse. The overall difference for any area will depend on whether a.m. or p.m. observation times prevail in the historical network. In addition, as mentioned in Section 6-4 the diurnal variation used to obtain 6 hour means from max and min values when computing historical MAT is based on sites in northern Vermont and the central Sierras of California. This assumed typical daily temperature pattern might not be representative of all regions, especially Alaska where the daylight period can be quite a bit longer during melt periods than in the lower 48. It would be very difficult to operationally correct for the bias caused by using only max/min data historically. The real solution to this problem is to modify the historical MAT preprocessor so that it uses the same data, i.e. both instantaneous and max/min, as the operational MAT preprocessor.

Just as with precipitation, comparisons should periodically be made between the operational MAT estimates and those that would be generated from the climatological network to insure that there is minimal bias. Again the ideal time to make such comparisons is when the historical data record is being extended.

Mean Areal Potential Evaporation (MAPE)

The operational MAPE definition should use exactly the same synoptic stations and weights as used for the historical estimates. The average monthly PE values required for the operational definition should be those that were computed by the calibration MAPE program for the historical data period (table of monthly averages displayed at the end of the calibration MAPE program output). Though the calibration MAPE program allows for the use of daily pan data to be used for MAPE computations, this is seldom done. The NWSRFS OFS doesn't include the option of using daily pan observations. If pan data were used historically, any bias between the operational and historical estimates should be minimal if all the computed station PE values are adjusted to conform to the lake evaporation given in NOAA Technical Report 33 (see Section 6-5).

While it would be good to compare archived operational MAPE estimates to values computed from climatological data, this is currently not an option since the data used historically are no longer available with the advent of ASOS. The best that can be done is to verify that the operational estimates of PE at each station are compatible to the annual lake evaporation from report 33.

Forecast Data

Estimates of precipitation, temperature, and evaporation out into the future are clearly important for river forecasting. Any bias in future data will not have a cumulative effect, but it will decrease forecast accuracy. Bias in observed data estimates affect the model state variables (in NWS terminology carryover) that are saved in order to initiate subsequent forecast runs, thus not only affecting the current period, but also later events.

- Future Precipitation - Future estimates of MAP are input directly into OFS and are not computed based on any defined information. In the past Quantitative Precipitation Forecasts (QPF) that were used to determine areal estimates of future precipitation for the next few days were generally biased on the high side, especially during convective storm periods. Much work has been done over the past decade to reduce this bias. QPF is now routinely used at most RFCs for short term forecasting. It is still important to verify that the future MAP values being used are unbiased compared to the data used for calibration. In order to properly do this, future MAP should be compared to MAP values generated from climatological data, i.e. the same data used for calibration. Comparisons between future MAP and operational MAP values don't prove that the QPF estimates are unbiased since, as discussed earlier, the operational MAP estimates can be biased.

- Future Temperature - Future estimates of MAT are generated in OFS using the station and area definitions that are used for the observed data period. The major difference is that for the future period the computations are based only on forecast max and min temperatures, typically for only a subset of the observational network. Even though this appears to be the

same procedure that is used to generate historical MAT estimates, there are some differences which could result in a bias. First, the equations used to compute 6 hour means from max and min values, i.e. represent the typical diurnal temperature pattern, are somewhat different than those used in the historical MAT program. The future MAT equations are also based on Z time rather than local time, thus they produce a slightly different diurnal pattern for each time zone. Second, at least some of the forecast temperatures used by the RFCs are the mid-day and early morning predictions. Thus, the "max" could be lower than the "min" if the temperature was forecast to drop throughout the day. Also, the problem described in Section 6-4 of the wrong max or min being used on days when temperatures are generally increasing or decreasing should not occur with forecast values. Third, the meteorological forecasts may not result in the same long averages used in the historical analysis since they are produced using a different procedure. For all these reasons it is important to compare the future and historical MAT values periodically. It would also be more consistent if all the MAT computations in NWSRFS used the same method. The most accurate estimates should come from a procedure involving the use of both max/min and instantaneous temperature values.

• Future Evaporation - Future MAPE values are calculated in OFS from the average monthly values defined for each area. The last computed value for the observed data period is blended over a specified number of days to the average curve. The average curve is determined by linear interpolation between the average monthly values. Thus, if the average monthly values are properly defined, the future MAPE values shouldn't create a simulation bias.

• Data for Extended Predictions - The historical time series of precipitation, temperature, and evaporation are currently used as the input for extended predictions. Near future estimates, as just described in the preceding paragraphs, are typically used for the first few days and then blended into the historical data in an attempt to minimize the effect of events in the historical record that are very unlikely to occur based on the current weather forecast. Since the same input is being used for extended predictions as was used for calibration, there shouldn't be a bias introduced other than that contained in the model states for the start of the run and in the forecast data for the first few days. However, given the meteorological outlook, it could be that the data traces for each of the historical years are not equally likely to occur this year. Methods to include long range weather outlooks in ESP runs have been subjective in the past.

Operational Segment Definitions

All parametric information contained in the active part of the basic segment definitions (i.e. the operations that perform the actual simulation computations as opposed to operations that convert or display data) should be exactly the same in both the operational and calibration systems. This includes the areas that are being modeled and the time interval of all data and model computations. Changes may need to be made to some of this parametric information if the space and time scales used operationally are modified in an attempt to improve forecasts by using new data types and processing methods that were not available historically. Such changes will be discussed in the second part of this chapter. This section will concentrate on the proper

definition of the segments to reproduce the results obtained during calibration utilizing similar data with no real time modifications to input values or model states.

Operational segment definitions need to reflect the different data types that are typically used for real time computations, such as river stage observations that need to be converted to discharge with the STAGE-Q operation as opposed to historical streamflow data that has been already converted by the agency responsible for collecting the data. Operational segments will contain direct input of reservoir and diversion data, that were used historically to compute natural flow conditions, when operational forecasts need to reflect what is actually occurring within a river basin. In other cases, especially in the west where water supply is the major concern, the operational segments will continue to generate natural flow as during calibration. The operational segments will also include different operations to produce and display the output needed for forecasting instead of the WY-PLOT and PLOT-TS displays used to verify simulation results during calibration. However, the bottom line is that the basic segment definitions should conform to the same areas and time intervals that were used for calibration and all the parameters for the models that are performing the basic hydrologic computations should be exactly the same as those determined during the calibration.

When first defining a segment, initial values of the state variables need to be provided for many of the models. In many cases, such as for channel response, routing, and reservoir models, these values can easily be estimated from current river and pool levels or default values can be used for operations where the values will only affect the computations for a short period. For the snow model, the initial values of the model states can be defaulted to zero when no snow cover exists. If a significant amount of snow is present when the model is initiated, then it is critical to input a realistic estimate of the current mean areal water equivalent. If a greater amount of snow existed earlier in the season, then the maximum water equivalent since snow began to accumulate also needs to be provided. Both of these values can be obtained from available snow measurements as described under the .WECHNG MOD in Section 8-1. The heat deficit, antecedent temperature index, and liquid water content are estimated based on the temperature pattern in the recent past. For the Sacramento model, estimates of the lower zone free water contents should be able to be derived from the hydrograph by determining the current supplemental and primary flow contributions. These runoff amounts are then divided by the appropriate withdrawal rates to get the current contents. Upper zone free water contents are typically set to zero since the effect of this variable will only last for a few days. Tension water contents can be reasonably estimated by comparing the general precipitation pattern over the last few months with a similar period from the calibration simulation. The initial value of ADIMC is set to the sum of the upper and lower zone tension water contents.

Introduction

The major objective during forecasting is to utilize all available information so that simulations closely mimic observations and forecasts have maximum lead time with minimal uncertainty. Random fluctuations that were tolerable during calibration, in that they didn't drastically interfere with the determination of the proper values of the model parameters, now need to be reduced to a minimum to provide the best possible operational forecasts. There are several methods for improving real time forecasts over what would be obtained by just using the simulated results generated from the basic OFS station, area, and segment definitions (i.e. the definitions just described in this chapter utilizing the same types of station data and modeling schemes used for the historical analysis and calibration).

- Use data analysis procedures that incorporate the dynamics of what is occurring, both by new measuring techniques and physical modeling, to improve the input to the hydrologic models over that provided by the historical techniques which often rely on climatological patterns. Such procedures are capable of providing better spatial and temporal resolution as well as improved areal estimates.

- Apply the models in a distributed manner at a finer spatial and temporal resolution than was possible during calibration by utilizing the output from the new data analysis procedures.

- Adjust the model states and computations, either automatically or manually, based on real time observations or the output from external modeling systems.

Besides these methods for improving river forecasts which will be discussed in more detail in this chapter, forecasts can also be improved for many users by quantifying the uncertainty in the predicted values. This has been done in the past for extended predictions. In the future probabilistic forecasts will be possible for all lead times. In order to make accurate probability statements for short term forecasts, the uncertainty in the models themselves and the model states must be taken into account, in addition to the uncertainty in future precipitation, temperature, and evaporation. The ensembles used to generate probabilistic predictions will need to incorporate the trends and uncertainty in meteorological forecasts from short term QPF to long range outlooks. Whatever is done in terms of producing the ensembles, it is important that the statistical properties of the historical data and the historical model simulations based on the calibrations be preserved, including subtle features like the amount of snow verus rain and the frequency of surface runoff.

Before discussing each of the listed methods for improving forecasts a couple of general items should be noted . First, in most cases there is not a one-to-one relationship between new data estimates and the historical values used for calibration or between observations or values computed by external modeling systems and values produced by conceptual models. Thus in order to avoid any bias being introduced, the relationship between the estimates must be determined and corrections applied when necessary. Second, most new data analysis or

hydrologic modeling systems evolve over some period of time. Even though a new method is tested thoroughly during the development phase, further modifications to the algorithms and data processing techniques are frequently needed when the procedure is put into operational use. Thus, the relationship between values produced by the new method and the historical calibration may change for some period of time before becoming stable.

The intention of the following discussion of each of the methods for improving forecasts is to indicate how such methods can be applied in a manner that preserves the benefits of calibration. The emphasis is on how to use such procedures without creating a bias compared to the calibration simulations. It is not the intention of this manual to describe these methods in any detail or to offer guidelines for their use other than how they can be applied in an unbiased manner.

Improved Data Analysis Procedures

The only procedure that is currently available operationally for improving model input utilizes radar, raingage, satellite, and other data to produce hourly estimates of precipitation on a 4x4 km grid. This procedure is currently referred to as the Multisensor Precipitation Estimator (MPE) [*Seo et al.*, 2000]. Since the MPE procedure generates estimates with a finer spatial and temporal resolution than generally can be obtained from just raingage data, it opens up the possibility of not only providing improved precipitation values for use by lumped applications of hydrologic models, but of allowing for the real time use of distributed models for river forecasting.

Other possible real time data analysis procedures that might be developed in the future to improve operational input include:

- mountainous area precipitation analyses that utilize some combination of orographic modeling, gage reports, radar information, and other data to account for the storm dynamics,

- mountainous area temperature analyses that utilize a combination of meteorological modeling, temperature observations, upper air soundings, and other data to account for the dynamics of how temperature varies with elevation, and

- direct estimates of ET-Demand that account for vegetation dynamics on evaporation rates.

Experience with radar-based estimates of precipitation have indicated that in many cases the estimates are biased and not consistent over time [*Jonhson et al.*, 1999]. This is partly due to changes that have been made to the algorithms and data processing techniques. There have especially been difficulties with obtaining reasonable precipitation estimates from radar-based methods in mountainous areas and when snow is occurring. Thus, before using these data as either lumped or distributed input to the hydrologic models for real time applications, it is essential that the estimates are carefully evaluated in a systematic manner. This evaluation

should determine whether radar-based precipitation estimates can be used to produce unbiased simulation results and if so, how they can be applied in order to improve operational forecasts. The application could be to provide lumped 6 hour values or lumped hourly estimates of precipitation. Beyond that radar-based precipitation estimates could allow the models to be applied in a distributed manner.

One possible procedure of evaluating whether improvements are possible by using radar-based precipitation estimates is described in this subsection. In the discussion of the procedure 3 types of mean areal precipitation time series are utilized. These are referred to as:

- MAPX - areal mean precipitation produced from the radar-based estimates by averaging the grid points that fall within the area being modeled.

- MAPH - areal values produced from climatological data, i.e. values generated using the same precipitation gage data and methods as used for the historical analysis.

- MAPO - areal values produced using real time precipitation reports with the OFS MAP preprocessor function based on the station and area definitions discussed previously in this chapter.

The steps in the procedure for a given watershed are as follows.

1. Calibrate the watershed following the methods described in this manual. This will insure that a sufficiently long record is used so that a stable set of parameter values can be obtained considering the noise in the input data.

2. Evaluate the archived radar-based estimates of precipitation for consistency and potential bias. There probably needs to be in the order of 5 years of reasonably stable radar based estimates to obtain a good evaluation. One way to do this is to compare the monthly MAPX values to overlapping MAPH estimates. The pattern of the ratio of MAPX to MAPH values should be consistent over time. There will be scatter since each method will compute different values for each storm. Trends in the ratio with time generally would indicate changes in how the radar based values were processed. Ideally the average ratio should be reasonably close to 1.0. For a non mountainous area this would indicate that the radar based procedure is producing values over the long term at each point that are the same as would be measured by a raingage. However, it is more critical that the ratio is reasonably consistent over time than being equal to 1.0, though typically if there is a significant bias, there may also be a lot of variability associated with the radar-based estimates.

This would also be a good time to compare archived MAPO values with overlapping MAPH estimates. If the station and area definitions for OFS are compatible with those used for the historical analysis, the monthly and annual ratios of these time series should be very close to 1.0. If not, it indicates that either the OFS station or area information is not correct or

something else, such as different reporting characteristics, causes these time series to differ.

3. If the evaluation in step 2 looks reasonable, then generate 6 hour MAPX time series (assuming that was the time interval used for calibration) for the watershed and use these as input to the hydrologic models. Compare the results to those obtained using the MAPH estimates as model input. If the long term ratio of MAPX to MAPH is 1.0, the overall simulation bias should be similar from both time series. If the ratio is not 1.0, then the MAPX values need to be adjusted, typically by a multiplying factor (PXADJ in either the SNOW-17 or SAC-SMA operation). It is preferable at this point to use PXADJ to correct for any bias as opposed to modifying model parameter values because the simulation bias is caused by differences in the MAPX and MAPH amounts. If the MAPX estimates are not consistent over time, this will result in the model simulations using those data to have more of a time trend than those produced by the MAPH time series.

Once the MAPX values are adjusted so that the overall bias is the same for both precipitation time series, an evaluation can be made of the results. Ideally the MAPX values will contain less random error and thus provide an improved simulation capability. If so, it indicates that at least the radar based precipitation can be used to provide a lumped 6 hour input for operational forecasting that should improve results. If the comparison of the simulation results indicates that there is a pattern, perhaps seasonal, in terms of which MAP produces the best results, then rules can be devised for when to use the radar based values operationally and when to use MAPO estimates (assumes that the MAPO values are compatible with the MAPH estimates). Such a situation will likely exist in regions where snow predominates during the winter months. In this case the MAPX values may provided improved input during warm months, but the MAPO estimates would produce better results during the cold season (the MERGE-TS operation is used in OFS to switch between alternative time series). The MAPX and MAPH time series for the test period could be merged based on the proposed rules to determine how this should affect the operational simulation results.

4. If the use of MAPX time series indicates an improvement in step 3, then generate 1 hour MAPX time series, adjust model parameters for the change in the time scale if necessary, and evaluate the results. Models that are non linear, such as the Sacramento model, are scale dependent, i.e. the results can vary depending on the spatial and temporal scale that the model is applied [*Finnerty et al.*, 1997]. Thus, in order to get similar results at different scales, some of the parameter values will likely need to be altered.

Hourly data should have the most chance of improving simulation results for fairly fast responding watersheds that have significant amounts of surface runoff and variations in rainfall intensity during storm events. For these watersheds instantaneous streamflow data, preferably hourly, is needed in order to make the parameter adjustments and to properly evaluate the results. The PXADJ value determined in step 3 should continue to be used to correct for the overall bias in the MAPX time series. The most likely Sacramento model parameter modifications are to the percolation rate and UZFWM. The percolation curve

would typically need to be adjusted slightly upwards. This should be done by changing the LZFSM and LZFPM parameters by the same percentage. This will change the PBASE value, but retain the shape of the curve and not alter baseflow recession rates. UZFWM should need to be increased somewhat. The unit hydrograph would also need to be converted to a one hour runoff duration. It is also likely that there would be a need to modify the shape of the unit hydrograph somewhat based on the finer temporal resolution of the computations and the observed flow data.

If doing the model computations on a hourly basis improves the simulation results, it should primarily be noticeable in the simulation of storm peaks. This would especially be the case for major surface runoff events when high intensity rainfall occurs for just a few hours within the 6 hour intervals. Watersheds that don't generate much or any surface runoff or have fairly uniform rainfall rates during major storms, shouldn't show any real improvement in simulation results by using hourly precipitation input. During seasons when snowmelt runoff predominates, it is unlikely that using hourly precipitation input will improve the results. The melt rates, especially when using a temperature index model, will not vary significantly from hour to hour and the snow cover will have a damping effect during rain-on-snow periods.

By going through such an evaluation for several watersheds scattered throughout an RFC area, it could be determine whether the radar-based estimates would provide unbiased values of precipitation for use with the calibrated conceptual models. A determination could be made of whether radar-based estimates could be used all the time or only for certain seasons or types of storms. It could also be determined whether adjustments were needed to the model parameters in order to avoid a simulation bias, especially when a change is made to the temporal scale of the model computations. Radar-based estimates of precipitation clearly have the potential of improving this critical model input, at least in non mountainous areas. However, before they are used operationally it should be verified that a simulation bias is not be introduced.

Distributed Modeling

The Hydrology Laboratory of NWS has had a major research effort for a number of years involving the use of distributed models for river forecasting [*Smith et al.*, 1999]. This effort is aimed at evaluating the data needed for such an application, examining various options for how models can be applied in a distributed manner, investigating the appropriate models to use, and ultimately to come up with a set of guidelines for how to use distributed models and under what circumstances they are needed and should produce improved forecast products.

The use of hydrologic models in a distributed mode offers the potential for improvements in simulating the response from basins with spatially varying precipitation patterns and allows for creating forecast products at a finer spatial resolution. Forecasts could not only be generated at the watershed outlet, but possibly also at interior points. This would allow for significant improvements in specifying the locations of possible flash flooding. The greatest potential for benefits from distributed modeling is in basins where runoff typically only occurs over a portion

of the drainage area and in areas where the location of intense rainfall and thus fast response runoff varies from event to event. In such watersheds the magnitude and timing of the hydrograph response is highly dependent on where the runoff is being produced. Distributed modeling should allow for the successful simulation of streamflow in regions where lumped applications provide unsatisfactory or marginal results (see Figure 1-1). Even in regions where lumped applications generally provide reasonable results, distributed models could improve the forecasts for certain events. The most improvement should be noticed for events with considerable surface runoff concentrated over only a portion of the drainage. Elongated watersheds are more likely to have a different hydrograph response in such cases than a more oval basin. The more dampening of the response either due to infiltration and percolation, snow cover, or the channel system, the less likely that a distributed application of the models will result in improved simulation results.

In order to apply models in a distributed mode it is essential that quality precipitation estimates are available at the appropriate spatial and temporal scale. This is why the validity of gridded precipitation estimates, such as those from radar-based procedures, must first be verified before attempting to use the models in a distributed manner. In addition to gridded estimates of precipitation, some of the rainfall-runoff model parameters will have to be modified when going to a finer spatial resolution. This is because of the scale dependency of many such models, including the Sacramento model [*Finnerty et al.*, 1997]. These factors imply that some period of stable gridded precipitation estimates need to be available for testing the distributed application of the model and to determine what parameter changes are needed to insure that the results are unbiased compared to the lumped calibration. This could be done by applying the model in a distributed mode for a period with stable precipitation estimates and comparing the simulation to the lumped calibration results for the same period (calibration simulation would most likely use the areal average of the gridded precipitation, but could use MAP computed from historical gage data). Parameters should be adjusted so that the overall bias of both simulations are the same. Also such a comparison would allow for an evaluation of the potential benefits of distributed modeling.

For watersheds in regions where the lumped application of conceptual models generally yields unsatisfactory results (see Figure 1-1), the comparison and parameter adjustment procedure just described is not really relevant. In such regions it is impossible to obtain appropriate model parameter values by applying the models in a lumped mode and the climatological records do not contain a sufficient gage density to attempt a distributed calibration. In such regions reasonably reliable calibrations and operational application are only possible after a sufficiently long record of stable radar-based precipitation values have been archived. These data would then be used in an attempt to calibrate the models in a distributed mode.

Distributed modeling could also be attempted for snow accumulation and ablation. Most likely this would be done by using an energy budget approach for computing the snow cover energy exchange rather than by trying to figure out how to realistically vary melt factors from one subarea to another. There are potential difficulties in determining how precipitation and the

energy balance input vary in mountainous areas as a function of elevation, aspect, slope, and vegetation cover. Even though a energy budget based distributed snow model will hopefully remove the trend of under simulating major melt events, a comparison of distributed model simulation results with a lumped application of a temperature index model should be able to determine if the distributed results are reasonably unbiased and provide an overall improvement compared to the historical calibration.

It also should be noted that changes to the spatial and temporal scales that the models are applied operationally has an impact on ensemble streamflow prediction (ESP) techniques used to generate probabilistic forecasts. Short term ESP applications will most likely use ensembles that are generated based on uncertainties in QPF and other meteorological forecasts. These ensembles should be able to be generated for the same spatial and temporal scale as the models are being applied operationally. However, for extended ESP applications the ensembles have been the historical precipitation, temperature, and evaporation time series that were produced during the historical data analysis step of the calibration process. These time series are at the spatial and temporal scale that was used for model calibration. Thus, in order to make extended ESP runs, the models must, at least currently, be defined operationally at these same scales. It would be quite difficult for an RFC to maintain multiple sets of compatible operational segment definitions at different spatial and temporal scales for short term forecasting and extended predictions. Thus, until some solution to this dilemma is developed, basins for which extended ESP applications are important will need to be defined operationally at the same spatial and temporal scale as used during calibration.

Adjustment Procedures

One of the most common methods for reducing the variability of the operational simulations is to adjust the input data, model computations, or model state variables based on a variety of observations or the output from an external modeling system. Adjustments can be made manually by the forecaster, generated by an external analysis of relevant data, or through the use of an automated procedure. The aim in all cases is to reduce variability and remove trends so that the models can be used to generate an unbiased forecast with minimal uncertainty. It is important when applying such adjustments that a bias is not introduced as compared to the simulation that would result if no adjustments were applied, other than the removal of some of the modeling deficiencies mentioned in Chapter 7.

Automated updating procedures include Kalman filter based procedures such as that utilized by the SS-SAC operation in the OFS [*Georgakakos et al.*, 1988] and a variational assimilation approach [*Seo et al.*, 2002]. The application of these automated procedures have thus far been limited to non-snow, lumped, single area headwater basins. The procedures use streamflow observations to adjust precipitation and evaporation inputs to the Sacramento model and state variables for the Sacramento and channel response models. The procedures are primarily intended to make adjustments during runoff events. The SS-SAC updating procedure uses a state-space version of the Sacramento model. Since this version doesn't exactly duplicate the

regular Sacramento model computations, there is some potential for bias to be introduced. Both of these procedures have primarily been tested by comparing short lead time forecasts using the updating procedure to forecasts produce using a simple blending or persistence technique. In evaluating such procedures it would be important to determine which variables are being altered and by how much. It is possible that state variables could be changed to physically unrealistic values based on current streamflow observations that would not impact a short lead time forecast, but could produce a substantial bias for a longer lead time or especially for an extended prediction.

The primary method for making real time adjustments for all types of watersheds and modeling configurations with the OFS has been the use of run time modifications (MODs). The MODs in the OFS allow the user to modify time series values, model state variables, model parameters, and rating curves. Input data can be changed, as well as time series generated by any of the operations in the segment definitions. The user can specify where in the operations table the change is made and which values are modified. Model state variables can be altered at the beginning of a run or at a specified time within the run. MODs that alter model parameters are primarily used to adjust the computations on a temporary basis for such things as abnormal snowmelt rates or particular precipitation patterns that affect the channel response. Rating curve modifications are intended for locations with constantly changing ratings.

MODs can be specified by the forecaster via the Interactive Forecast Program (IFP) in NWSRFS at run time or they can be generated externally. When making MODs with IFP, the forecaster should not only rely on deviations between observed and simulated streamflow, but on a variety of other information such as: radar data, meteorological data and analyses including forecasts, observer reports, snow cover information, recent climatic trends (above or below normal conditions), satellite measurements, and river ice reports. MOD images are generated externally when an objective analysis procedure is used to compute the adjustments (e.g. procedures are available to compute snow water equivalent estimates based on snow course or areal gamma flight line measurements).

The main concern from the point of view of this manual is that MODs are not applied in a manner that will result in a biased forecast compared to what would be generated without any modifications. For example, it is difficult to determine the cause of any deviation between simulated and observed values and the magnitude of any needed adjustments early in a storm event. After the event is over, it is much easier to decide the cause (e.g. a volume problem can clearly be distinguished from a timing error after the event, however, early in the event the hydrograph response may look the same in both cases). The sooner the proper adjustment can be made, the longer the forecast lead time. However, if the wrong adjustment is made, it will typically result in the need to make further modifications which will alter the forecast and decrease the time available for action. In addition, the wrong adjustment might cause the state variables to take on values that could result is a considerable bias during the subsequent recession period or the next event and could especially impact extended predictions.

The chance of making a modification that could bias subsequent forecasts is even greater when there are several sources of error. For example, during the widespread flooding in the northeastern United States in late January 1996 [*Office of Hydrology*, 1998], the meteorological situation clearly indicated extreme snowmelt conditions, but there was also evidence of an underestimation of precipitation due to pronounced orographic effects. In addition, there was uncertainty in the initial snow cover and soil moisture conditions. If all the error was attributed to a single factor, it would be quite likely that some of the model states would not be correct and would result in subsequent forecasts being biased. In such cases it would be extremely valuable for the forecaster to have access to tools that would provide insight into potential errors and estimate a reasonable magnitude for each likely problem.

Besides storm events, there are other times when MODs are used to modify model states. During low flow conditions adjustments can be made to baseflow storages based on recent deviations between simulated and observed streamflow. Prior to the snowmelt season, model estimates of areal water equivalent are modified based on measurements of the snow cover. Soil moisture tension water deficits could be altered based on soil moisture measurements or model response after a dry period. Such updates can not only have an effect on the next significant runoff event, but more importantly they will alter the results of any extended predictions and can produce a substantial bias if not applied properly.

If the models being used for forecasting are well calibrated and properly applied operationally, two things should occur in regard to real time adjustments. First, there should be less need to make updates as the model simulations should more closely track observed conditions. Except for trends caused by model limitations or the spatial scale of the application (see Chapter 7), the errors from a well calibrated and properly applied model should be random. Models that are poorly calibrated or improperly applied will produce biased results and thus will need to be adjusted more frequently. Second, the proper adjustment to make and its magnitude should be able to be based on a logical assessment of potential sources of error. For example, a scientific evaluation of rainfall volume or snowmelt rates should be able to be used directly to determine if adjustments to these values are needed and the appropriate magnitude. Whereas, with a poorly calibrated or improperly applied model, there may be no logical explanation for the adjustments needed to cause the simulated conditions to reasonably match observations.

Section 8-1 provides some guidance for using the MODs directly associated with the SNOW-17 and SAC-SMA operations in a manner that reduces the chance of generating biased forecasts. General recommendations for using MODs are more in line with an operational forecasting, rather than a calibration manual, and are thus not included.

Operational Implementation Summary

If the objectives are met, i.e. unbiased simulation with minimal random error, parameters that

properly reflect the function of each component of the models, and spatially consistent parameter values, then calibration results should significantly improve operational forecasts. The models should more closely track observations, fewer real time adjustments should be needed, and the choice of which modifications to apply and their magnitude can be based on logical decisions using all available information. In addition, besides producing unbiased short term forecasts with minimal uncertainty, the models can be used to generate accurate extended probabilistic predictions. However, all of these benefits are only possible if the calibration results are properly implemented in the operational environment. This requires that the results produced with the operational parametric definitions must be unbiased compared to the calibration results. It also requires that alternative sources of model input, changes to the spatial and temporal scales that the models are applied, and real time adjustments be applied in a manner that while reducing random error and model deficiencies doesn't generate biased results.

While the guidelines and recommendations offered in this chapter should produce forecasts that contain the benefits of a proper calibration, the user needs to periodically verify the results and carefully evaluate alternative inputs and methods of applying the models. Comparisons and consistency checks should be made every few years to verify that real time data are indeed generating operational input and simulation results that are unbiased compared to the historical input and calibration output. New data analysis procedures, changes in the spatial and temporal scale that the models are applied, and real time adjustments need to be carefully evaluated to make sure that their use results in not only unbiased simulations, but produce real forecast improvements.

Section 8-1

Guidance for Using MODs with the SNOW-17 and SAC-SMA Operations

Introduction

This section provides some guidance for using the main MODs that are directly associated with the SNOW-17 and SAC-SMA operations in a manner that reduces the chance of adding a bias to future forecasts and thus preserving the benefits of calibration. For the SNOW-17 operation the MODs discussed are:

- .MFC (used to change the melt rate under non-rain situations)
- .UADJ (used to change the melt computations during rain-on-snow periods)
- .WECHNG (used to change the mean areal water equivalent)
- .AESCCHNG (used to change the areal extent of snow cover)

For the Sacramento model the MODs included are:

- .SACBASEF (used to change the amount of water in lower zone free water storages)
- .SACCO (used to change any of the Sacramento model state variables)

Though this discussion provides a brief description of exactly what happens when each MOD is applied and provides recommendations for their use, the focus, as mentioned, is on using these MODs in a manner that doesn't bias the state variables and thus cause problems during subsequent forecasts, including extended predictions.

.MFC MOD

The Melt Factor Correction (.MFC) MOD is used to multiply the non-rain melt factor by a constant for the period specified. This MOD can be used to change the melt rate for computational intervals within both the observed data period and out into the future. The primary reason for the .MFC MOD is that the relationship between air temperature and surface energy exchange varies depending on meteorological conditions. The normal seasonal variation in this relationship is specified when the SNOW-17 model is calibrated. However, the scatter about the seasonal melt factor curve is ignored during calibration as only the average relationship is being established. Operationally there is a need to remove this variability in order to improve forecast accuracy. During most periods the calibrated melt factor provides a very reasonable estimate of snowmelt, however, under certain situations the actual melt rate can be substantially different than the normal rate. Comparisons between energy balance melt computations and the SNOW-17 model have identified those meteorological conditions that produce the largest errors in temperature index based melt calculations [*Anderson*, 1976]. These include:

- Periods with high dew-points and high wind speeds - During such periods the turbulent transfer terms in the energy balance, i.e. the latent and sensible heat terms, transfer more energy than under normal conditions causing the melt rate to increase. In this case a .MFC value greater than 1.0 would be needed.

- Periods with much above normal temperatures and calm winds - Under these conditions the turbulent transfer terms are negligible, but the calibrated melt factor is computing more melt than would normally be expected at that time of year. This results in too much melt and thus the need for a .MFC value of less than 1.0.

- Periods with clear skies, but much below normal temperatures during the height of the melt season - Under these conditions there is a significant amount of solar radiation and the snow surface has aged so that the albedo is low enough for much of the radiation to be absorbed, however, the air temperature is indicating that not much melt should be occurring. In this case the .MFC value would need to be greater than 1.0.

Even though the forecaster can recognize these situations and realize that the melt rate needs to be modified, the problem is to determine the magnitude of the adjustment. If an error in the melt rate is the only cause for the simulated hydrograph to differ from the observed, the forecaster can interactively determine a correction that will bring the two hydrographs into line. However, even in this case there may be difficulties as the meteorological conditions could be changing from one time interval to the next. Thus, it may be possible to determine the average correction over several intervals, but the variation from one interval to another would be difficult to assess. The problem is compounded when there are multiple causes for the simulation errors or when trying to decide the correction to use when abnormal melt conditions are forecast for the next few days. This all indicates that a tool is needed to provide the forecaster with guidance as to when the .MFC MOD is needed and what is a physically realistic value for any corrections.

While it is difficult to obtain the data necessary to calibrate and apply an energy balance model on a routine basis, especially for complex forested areas in the mountains, it is likely that energy balance computations could be used to provide guidance for using the .MFC MOD. One possible procedure for doing this can be demonstrated using the January 1996 flood event in the northeastern United States [*Office of Hydrology*, 1998]. This event was characterized by extreme melt rates caused by high dew-points and winds and significant orographic precipitation. For this event energy budget melt values were computed for selected synoptic stations in the northeast. Dividing the melt computed from the energy balance by the air temperature minus MBASE gives the melt factor needed for each interval to reproduce the energy balance melt amount. The normal melt factor could be determined by calibrating the SNOW-17 model for each of these stations using observed temperatures and water equivalents. Then the ratio of the energy balance melt factor to the normal melt factor, which is the .MFC value, could be computed. Figure 8-1-1 shows how the ratio of the actual (energy balance) melt factor over the typical (calibrated) melt factor, i.e. the .MFC MOD, would have varied throughout this event for several synoptic stations in the northeast (in this case the normal melt rate was assumed to be 0.7 mm/°C/6 hr rather than based on a calibration of the sites - a typical melt rate for an open site in this part of the country for mid January).

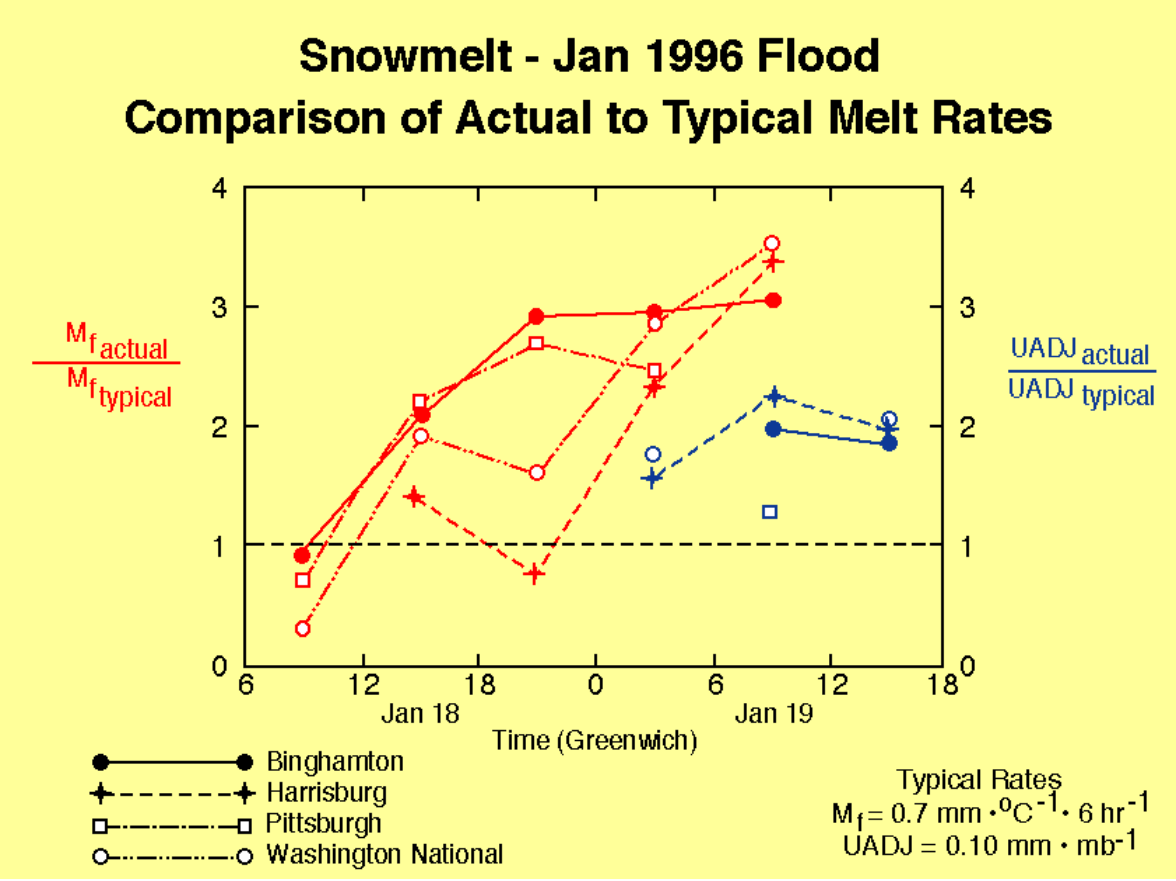


Figure 8-1-1. Variation of .MFC and .UADJ MODs for the January 1996 flood.

The next step in the procedure would be to map the .MFC values computed for each synoptic station (any station with the data needed for energy balance computations and observed water equivalent data could be used). This should give an indication of how the correction varies spatially. Isocorrection lines could be then drawn using meteorological observations or predictions as a guide. This is illustrated in Figures 8-1-2 and 8-1-3 for two successive 6 hour periods during the 1996 event. By 0Z high winds and dew-points existed in West Virginia,

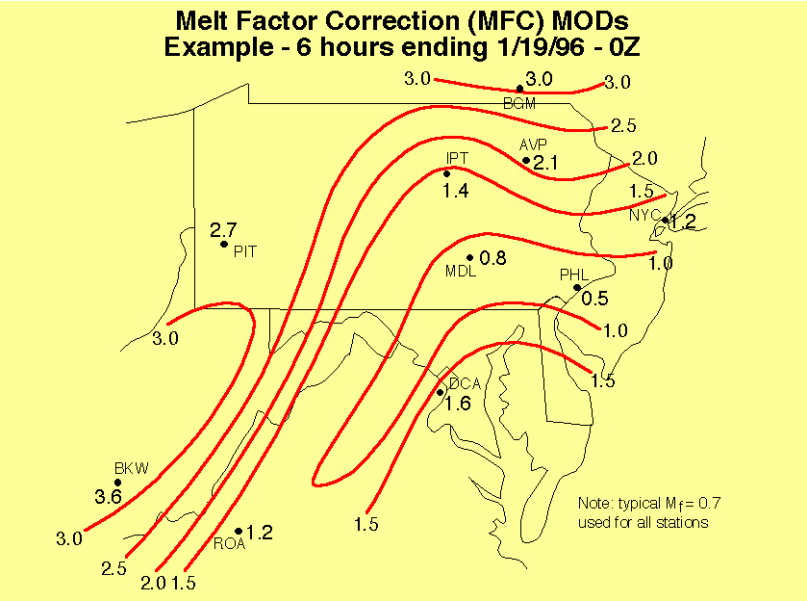


Figure 8-1-2. Variation in .MFC values for period ending at 0Z on Jan. 19,1996.

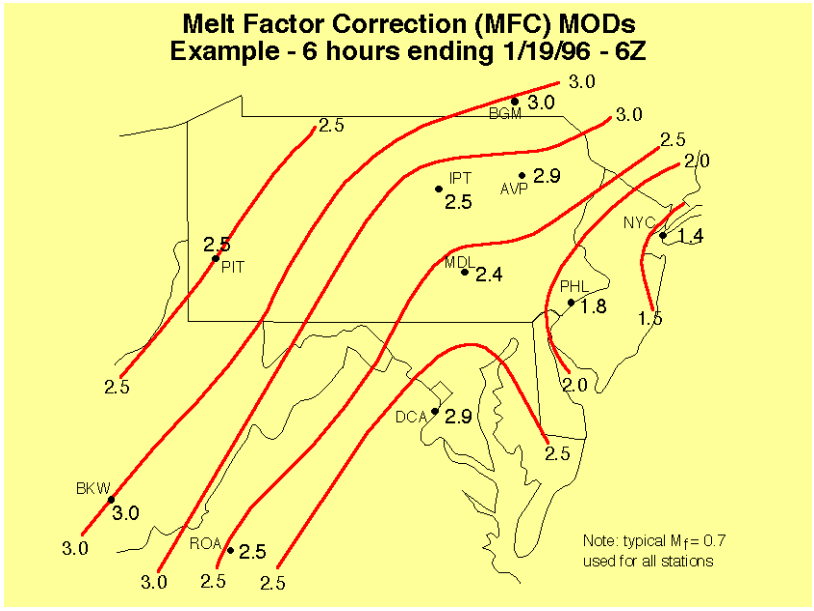


Figure 8-1-3. Variation in .MFC values for period ending at 6Z on Jan. 19,1996.

western Pennsylvania, and New York resulting in much above normal melt rates for these areas. Just east of the mountains conditions were very stagnant, though quite warm and humid, producing below normal melt conditions. By 6Z the high winds covered most of the region. The last step in the procedure would be to extract .MFC values for any particular watershed. This assumes that the .MFC values computed for an open site are appropriate to use for an entire watershed with mixed cover. This seems like a fairly reasonable assumption.

In the future estimates of the appropriate melt factor corrections could possibly be obtained from the Snow Data Assimilation System (SNODAS) being developed by NOHRSC [*Carroll et al.*, 2001]. SNODAS utilizes downscaled fields of meteorological data from numerical weather models and gridded precipitation estimates to drive a distributed energy and mass balance snow model. The resulting snow model states can be updated using satellite, airborne, and ground-based snow observations. The procedure was first run for the central and eastern U.S. in the spring of 2001 and continuously for the conterminous U.S. for the winter of 2110-2002. SNODAS is still in the evolutionary stage. Adjustments are being made to many of the algorithms and data fields. It will probably take a few more years for the system to reach a stable configuration. A period would then be needed to develop the relationships between melt factors derived from the SNODAS energy budget calculations and those from the SNOW-17 model. These relationships could then be applied to SNODAS derived melt factors to identify when abnormal melt rates are likely to occur and to obtain realistic estimates of the magnitude of any corrections. Alternatively, the relationships might be used to adjust SNODAS energy exchange estimates prior to directly overriding the SNOW-17 computed values.

If some method is not used to determine a physically realistic estimate for melt factor corrections, there is good chance that the correction applied will be too large or too small. In either case, the resulting water equivalent at the end of the correction period will be off (assuming it was correct to begin with) and subsequent forecasts will contain too little or too much snow, thus producing a bias.

.UADJ MOD

The wind function correction (.UADJ) MOD is used to multiply the UADJ parameter (wind function term) in the rain-on-snow melt equation by a constant for the period specified. This MOD can also be applied to future intervals as well as the observed data period. It should be noted that the .UADJ value only affects the turbulent transfer terms in the rain-on-snow melt equation. The longwave radiation and heat of rainwater terms are not altered.

The primary reason to use the .UADJ MOD is to account for variations in wind speed during rain-on-snow periods when the temperature is enough above freezing so that significant melt occurs. Since .UADJ is only used when rain is occurring, there is always the question as to whether simulation errors are due to melt computations or whether the estimate of the amount of rain is in error. As with the .MFC MOD, energy budget computations could be used to determine a reasonable value for the wind function correction. Figure 8-1-1 shows .UADJ values derive by comparing the wind function derived from the energy budget to a typical value for an open site for several synoptic stations when rain was occurring during the 1996 northeastern U.S. flood event. Computations of .UADJ values at synoptic stations, or other sites with the data needed for energy balance calculations, could be used as a guide for determining watershed corrections. However, wind adjustments might be more difficult to extrapolate in regions with substantial terrain and forest cover than melt factor corrections. As with the .MFC MOD, improper wind function adjustments will affect the resulting water equivalent and cause some bias in subsequent forecasts. In the future relationships could be derived to possibly adjust SNODAS computations of melt during rain-on-snow periods for use within the SNOW-17 model.

.WECHNG MOD

The Water Equivalent Change (.WECHNG) MOD is used to change the mean areal water equivalent at a specified time during the period of observed data. When the water equivalent is modified with this MOD, the areal extent of snow cover is assumed to remain the same. This results in the areal depletion curve being shifted when the change is made when some bare ground exists. This is illustrated in Figure 8-1-4. In order to change both the water equivalent and the fraction of the area cover by snow, both the .WECHNG and .AESCCHNG MODs must be used. In this figure TWE is the total water equivalent, i.e. the sum of ice and liquid water in the snow cover, and A_s is the areal extent of the snow cover.

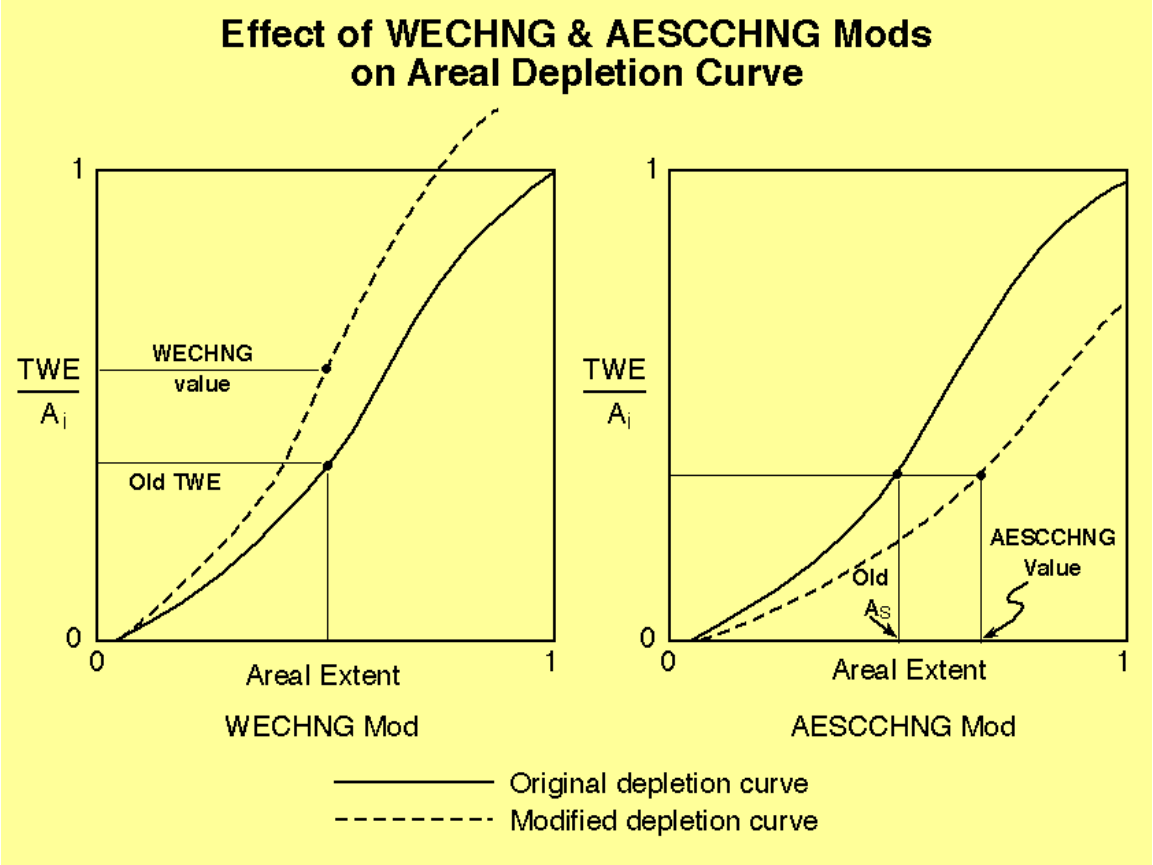


Figure 8-1-4. Effect of .WECHNG and .AESCCHNG MODs on the areal depletion curve.

The water equivalent typically needs to be modified because the snow cover accumulated by the model can be in error due to inaccurate precipitation input. This is caused by variations in catch deficiencies from normal and by storm dynamics that produce a different areal distribution than the typical pattern indicated by the isohyetal analysis. Also snowmelt computations may have been in error due to abnormal lapse or melt rates. The updated value of the water equivalent is normally based on water equivalent observations from snow courses, airborne gamma flight lines, or special surveys. The updated areal water equivalent derived from the observations can

be directly substituted for the model value or each estimate can be given some weight. This depends on the relative uncertainty of each value. In many cases this is a subjective decision.

In flat terrain it may be okay to assume that there is a one-to-one relationship between the observations and the model's areal estimate. This is more likely if there are sufficient observations scattered over the area. The problem that can occur in very flat areas, like portions of the upper midwest, is that during the melt period significant water can pond in fields before ice at the outlet melts and allows the water to drain. In this case airborne gamma observations may include both snow and ponded water. If such observations are used directly, it could resulted in a biased estimate of areal water equivalent.

In mountainous terrain it is highly unlikely that even a weighted average of the observations can be used directly without first determining the relationship between the weighted average and model estimates. Generally the relationship between available observations and model values will vary from one time of the year to another. Thus, separate relationships need to be derived for each date that observations are typically available. In order to determine the relationship between the observations and the water equivalent generated by the model, first a reasonably long, stable historical record of observations must exist. Second, there needs to be a model simulation using historical gage data for an overlapping period.

The simplest updating procedure would be derived by plotting the observed water equivalent from a single site or a weighted average of several sites in and near the area against the model estimate for the same historical date. The average relationship derived from this plot could then be used to update the model whenever real time observations are available from those sites. The procedure can be evaluated by using the relationship to update the water equivalent during the historical period and then determine whether the resulting snowmelt runoff volume is an improvement over the simulation with no updates.

A more complex procedure is the Snow Estimation and Updating System (SEUS) based on the methodology proposed by Day [1990]. This system was in use by the western RFCs for awhile, but is no longer supported by the NWS. A replacement for SEUS was develop for use in the Pacific Northwest [*RTi*, 2000]. This procedure develops regression equations between fraction of normal values of observed and simulated water equivalents for various times during the snow season. The procedure is based on fraction of normal values to make it independent of the network of observations. This avoids having to recalculate regression equations every time a new station is added or deleted from the network. In addition, the procedure avoids having to estimate missing data since the estimated fraction of normal is based only on the stations which actually report. Based on tests on selected basins to evaluate the effect of this updating procedure on hydrograph simulations, it was decided to weight the estimate derived from the observations and the model simulated value equally.

In the future output from SNODAS may be used for determining water equivalent updates for SNOW-17. SNODAS could prove to be especially beneficial in the central and east regions of

the country where the historical water equivalent network is quite sparse. SNODAS has the capability to use snow depth values, as well as water equivalent, to update model states. As with most updating procedures, relationships will need to be established between the SNODAS mean areal water equivalents and those produced by the SNOW-17 model in order to make sure that a bias is not introduced when the water equivalent is modified.

The key is to carefully evaluate any procedure for updating water equivalent as improper changes can have a large effect on the volume of snowmelt runoff and thus result in a significant bias. At least for mountainous areas, an objective procedure must be based on derived relationships between historical observations and model estimates.

.AESCCHNG MOD

The Areal Extent of Snow Cover Change (.AESCCHNG) MOD is used to change the areal extent of snow cover at a specified time during the period of observed data. When the areal extent of the snow cover is modified with this MOD, the mean areal water equivalent is assumed to remain the same. This results in the areal depletion curve being shifted. This is illustrated in Figure 8-1-4. In order to change both the water equivalent and the fraction of the area cover by snow, both the .WECHNG and .AESCCHNG MODs must be used.

The areal extent of snow cover may need to be changed because of variations in accumulation and melt patterns from normal or due to errors in the model simulation of water equivalent or melt. The most routine observations of the areal extent of snow cover are those obtained from satellite data by NOHRSC. At least in the past, aircraft were used in some regions to observe the snow line in the mountains and thus derive an estimate of the areal cover. Again, as with most observations, areal extent of the snow cover observations generally can't be used directly to update the snow model. It could be that in open, flat terrain the observed and simulated values of snow cover have a one-to-one relationship, however, this is certainly not the case in the mountains and even in watersheds with reasonably level terrain, but varying vegetation patterns. The main reason is that the snow model implicitly uses the areal depletion curve to account for variations in melt rates and other factors in addition to areal cover as discussed in Chapter 7-4 and illustrated on Figure 7-4-3. Thus, in most areas a relationship between the areal extent observations and the model computed values would have to be derived before the satellite estimates could be used to update the snow model. This hasn't been done to this author's knowledge. Such a relationship may not be linear as suggested by Figure 7-4-3. In order to derive the relationship an overlapping period with stable satellite observations and historical model simulated values must be available. In mountainous areas with multiple zones, the satellite estimates must correspond to the elevations used to divide the watersheds. If satellite observations are used to update the areal extent in the snow model without first defining the relationship with the simulated snow cover values, it is very likely that a bias will result, especially in mountainous regions.

.SACBASEF MOD

The Sacramento Baseflow (.SACBASEF) MOD is used to multiply the contents of both lower zone free water storages, i.e. LZFSC and LZFPC, by the same constant. This MOD allows the user to increase or decrease baseflow and maintain the relative division of water between the two storages. Generally this is the preferred method of modifying the amount of baseflow. A proper calibration should provide for a reasonable allocation of baseflow between supplemental and primary components. When there are low flow errors in a historical simulation, it is typically because baseflow is too high or too low, not the relative magnitudes of the two components.

When the .SACBASEF MOD is applied, it will take some period of time before the baseflow reaches the desired level, i.e. the change in the amount of baseflow doesn't appear instantaneously, but gradually over a number of time intervals. This is because baseflow, like all runoff generated by the Sacramento model, must pass through the channel response function, typically a unit hydrograph, before reaching the drainage outlet. Thus, the transition time is dependent on the number of ordinates in the unit hydrograph.

The decision as to whether baseflow needs to modified is generally based on comparing the simulated and observed hydrographs during a low flow period. When applying the .SACBASEF MOD there are several things to watch out for and situations to avoid. These include:

- The only flow components making up the simulated hydrograph should be primary and supplemental baseflow (supplemental flow can be zero if this storage is completely drained). There should be no other sources of runoff prior to when the hydrograph is being analyzed for a period equal to the length of the unit hydrograph. Currently there is not a graphical display of runoff components in IFP analogous to the ICP display shown in Figure 7-8-1, though this information can be obtained in a tabular form (PRINTSMA technique). In many cases during a recession it is assumed that the simulated hydrograph can be brought in line with the observed by changing baseflow, when in reality, other runoff components are still contributing to runoff. This can especially be the case near the end of a long snowmelt season. It appears that the hydrograph is in a baseflow recession, when snowmelt from a small portion of the area is still producing interflow and possibly impervious runoff.

- Baseflow modifications shouldn't be made during periods when there is evaporation from riparian vegetation. Such evaporation occurs when the RIVA parameter is greater than zero, there is a reasonable level of ET-Demand, and tension water deficits exist. The algorithms in the Sacramento model for handling riparian evaporation are quite simple, thus the simulation errors during such periods are much more likely to be the result of problems with computing the loss due to riparian vegetation than the amount of water in the lower zone free water storages. You probably just need to live with the simulation errors that exist under these conditions.

- It is probably unwise to modify baseflow at locations where problems can exist with the

ratings, especially at low flows. This is certainly the case during the winter in northern regions. River ice can significantly affect the rating curves in these area, thus it may be very likely that the observed flow values are in error. Observed flows are also likely to be in error at other locations where the channel is not well defined at low water levels.

If care is taken when using the .SACBASEF MOD, this should be the best way to correct for simulation errors during baseflow periods. This MOD takes advantage of the calibrated relationship between supplemental and primary baseflow and thus, in general, is less likely to result in biased values of the related storages than if the lower zone free water contents were modified independently.

.SACCO MOD

The Sacramento Carryover (.SACCO) MOD is used to change the value of any of the Sacramento model state variables at a specified point in time. The contents of any of the 6 moisture storages, UZTWC, UZFWC, LZTWC, LZFSC, LZFPC, and ADIMC, can be modified. Tension water can be changed by either specifying a new content or deficit. The MOD also allows for modifying the value of the frost index used in the frozen ground algorithm associated with the Sacramento model.

One tool that can be helpful when using the .SACCO MOD is the tabulation summary of moisture contents that can be generated by the WATERBAL operation after a calibration is complete. This table shows the average, high, and low values for each of the 6 moisture storages for each month and thus can be used to avoid making changes to the contents that are outside of the range experience during the historical simulation.

Following are some guidelines for using the .SACCO MOD in a manner that will avoid generating biased values.

- LZFSC and LZFPC - The preferrable method for making changes to these contents is to use the .SACBASEF MOD. If only primary baseflow exists and the situations mention under the .SACBASEF discussion are not present, then the effect of using the .SACCO MOD to change the baseflow would be the same as using the .SACBASEF MOD, however, it is much easier to compute a multiplying factor than to calculate the content that would be needed to produce the desired amount of baseflow. When both supplemental and primary components are contributing to the baseflow, it very difficult to assess how to change the contents of each storage based on the short period of time that is typically displayed operationally (normally less than a week of observed data are being processed). This makes it is very difficult to change these two storages independently. It is much preferred to use the relationship between these two runoff components that was established during the calibration and rely on the .SACBASEF MOD to make run time modifications to the contents. On most watersheds the contents of the supplemental, and especially the primary, storage will affect the amount of low flow that is produced for a considerable time in the future.

- UZFWC - It is not recommended that the forecaster attempt to modify this value. The contents of the upper zone typically change quite rapidly, especially when water is entering the model. The UZFWC value could possibly be changed to increase or decrease the amount of interflow during the first few days of a recession, but even then it is difficult to determine a new value since the upper zone free water storage is losing water both as interflow and percolation. The only saving feature if UZFWC is modified, is that the effect will only last for a short time since this zone typically drains within a few days though the effect of altering the percolation amount would persist much longer.

- UZTWC, LZTWC, and ADIMC - Tension water deficits develop over some period of time

during periods of dry weather and substantial evaporation. Potential errors in the size of these deficits are caused, in part, by biased evapotranspiration estimates due to abnormal weather conditions or vegetation activity. The errors can also be due to biased estimates of precipitation for the events, frequently convective, that occur during the period that the deficits develop.

Streamflow observations can't be used to determine if a problem exists in the size of the tension water deficits until a substantial runoff event occurs. This could be a rain storm after a dry period or the beginning of snowmelt for a year when the summer moisture deficiency wasn't satisfied by fall rains. In both of these cases the tension water deficits may only be a portion of the reason why the streamflow response is not simulated correctly. In the case of the rain event, the amount of rainfall for the storm could be incorrect. In the case of the snowmelt situation a number of factors can affect the simulation at the beginning of the melt season as discussed at the end of Section 7-7. These factors need to be taken into account when making modifications to the conditions prior to such events.

Tension water deficits could also possibly be adjusted based on soil moisture observations. There have been studies that showed favorable comparisons between the Sacramento model tension water deficits and soil moisture measurements. In order to properly use the soil moisture data to update the tension water contents, relationships would have to be established using an overlapping period of stable soil observations and historical model simulations. The problem in many cases would be in finding a sufficient record of moisture observations to develop the relationships. The soil moisture data would not have to be available in real time in order to be used for updating as tension water deficits develop slowly. It would be sufficient if the data were available within a day or two after the observations were made.

If the tension water contents are suspected to be in error prior to any runoff events and there is not an objective procedure for updating the values, extreme care is advised when making any changes. This is because these deficits, especially the lower zone deficit, can affect the amount of runoff produced over a considerable time into the future. Thus, any change can have a significant impact on an extended flow prediction, especially in times of drought. The range in the contents during the historical simulation should certainly be used to avoid a completely unreasonable modification.

Typically if the upper or lower zone deficits are altered and the ADIMP parameter is being used to produce direct runoff, the value of ADIMC should also be adjusted.

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Article

Is snow in the Alps receding or disappearing?

BENISTON, Martin

Abstract

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BENISTON, Martin. Is snow in the Alps receding or disappearing? *Wiley Interdisciplinary Reviews*, 2012, vol. 3, no. 4, p. 349–358

DOI : 10.1002/wcc.179

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Focus Article

Is snow in the Alps receding or disappearing?

Martin Beniston*

Snow in a populated and economically diverse region such as the Alps plays an important role in both natural environmental systems, (e.g., hydrology and vegetation), and a range of socio-economic sectors (e.g., tourism or hydropower). Changes in snow amount and duration may impact upon these systems in various ways. The objective of this text is to assess whether the public perception that snow has been receding in recent decades in the European Alps is indeed upheld by observations of the behavior of the mountain snow-pack in the last few decades. This article will show that, depending on location—and in particular according to altitude—the quantity of snow and the length of the snow season have indeed changed over the past century. While a major driving factor for this is clearly to be found in recent warming trends, other processes also contribute to the reduction in snow, such as the influence of the North Atlantic Oscillation on the variability of the mountain snow-pack. This article ends with a short glimpse to the future, based on recent model studies that suggest that snow at low to medium elevations will indeed have all but disappeared by 2100. © 2012 John Wiley & Sons, Ltd.

How to cite this article:
WIREs Clim Change 2012. doi: 10.1002/wcc.179

INTRODUCTION

Snow is an important element of the natural environment in many mid- to high-latitude mountain regions around the globe, such as the European Alps, and its presence or absence can have a range of consequences for many socio-economic sectors.[1] Snow is probably the biggest single contributor to seasonal runoff[2] in hydrological basins like the Rhine or the Rhone rivers when the snow-pack releases water during the spring and summer melt. Its presence at high elevations up to the middle or the end of the summer ensures, along with seasonal glacier melt, a sustained discharge in most mountain rivers even during prolonged dry spells.[3] As a result, the Alps have often been referred to as 'the water tower of Europe'[4] because of the key contribution of alpine rivers to populated lowland water resources in France, Germany, northern Italy, and Central and Eastern Europe. In terms of ecosystem functioning, snow is a major determinant for many alpine plant species since the timing of snow-melt often signals to dormant plants the beginning of the annual vegetation cycle.[5] Plant survival largely depends on the duration of the snow season, because the protective cover of snow that insulates plants from extreme mid-winter cold conditions.[6] In economic terms, snow is an important resource for winter tourism, and in many mountain resorts income from the ski sector far outweighs other tourist activities.[7] Through its major role in the hydrological cycle, the presence of snow and ice in high mountains has a direct bearing on the energy sector, since hydropower infrastructure is ultimately dependent on the timing and amount of runoff during the period of snow-melt. This is important in countries like Austria or Switzerland that depend for more than 60% on hydropower for electricity generation.[8]

Because snow is central to so many different environmental and economic sectors, an assessment of past and future snow patterns can help assess how these sectors are capable of adapting to shifts in the quantity and seasonality of the mountain snow-pack. This article will begin by a short inventory of the changes in the determining factors for snow, namely temperature and precipitation, and how the evolution of these factors has influenced alpine snow cover. Following this first appraisal, more subtle mechanisms

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than just warming temperatures or changes in winter precipitation will be explored, such as the impact of changing weather types and modes of decadal-scale variability, in particular the North Atlantic Oscillation. Finally, this article will address issues related to the future behavior of snow in a warming climate, to assess whether recent changes in alpine snow may be a forerunner of future winters in the course of the 21st century.

CHANGES IN THE ALPINE SNOW-PACK IN THE COURSE OF THE 20TH CENTURY

Available Data

Most of the national weather services within the European Alps have archives of basic meteorological data that include snow parameters. For example, MeteoSwiss (the Swiss weather service) climate archive contains daily data in digital form spanning most of the 20th century for over 100 locations.[9] Snow data is archived in the form of depth and new accumulation over a 24-h period; hence seasonal statistics of snow accumulation in relation to other parameters such as temperature, precipitation, or pressure can be compiled with relative ease. Unfortunately, the coverage of regular recording stations becomes sparse as one enters regions of complex topography, so that the number of high-elevation stations or those in remote parts of the Alps is low compared to lowland regions, for reasons of access, maintenance, and cost. For the climate observing network of MeteoSwiss (the National Basic Climate Network—NBCN), 40% of the recording stations are located below 1000 m above sea-level, 45% between 1000 and 2000 m, 10% between 2000 and 3000 m (i.e., three stations), and just one station is found above 3000 m. The representativeness of stations is nevertheless reasonably well respected in terms of station coverage as a function of surface area for each altitude class. The quality of snow data, compared to more standard variables such as temperature and pressure may, however, be open to question, since the depth of the snow cover depends upon many complex factors, such as wind, radiation, infiltration of melt-water, or the difficulty of verifying the quality of an automatic measurement at a remote location on a regular basis. Despite these drawbacks, the available snow data is generally of sufficient quality to be of use to assess some broad, long-term trends and the manner in which these are influenced by weather and climate.

While much of the information used in the sections to follow is based on Swiss climate data, because of its high spatial and temporal density and its recognized quality, the conclusions reached here are in general valid for many other parts of the Alps, from the Savoie region in the northern French Alps, across the northern Italian Alps, into southern Germany and Austria.

In terms of regional climate model (RCM) simulations, use will be made in this paper of model results compiled in the context of two major European projects, 'PRUDENCE'[10] and 'ENSEMBLES'.[11] In these projects, a suite of state-of-the-art climate models were used to simulate 'current' climate (i.e., for the reference 1961–1990 period) and scenario climates over Europe based on a limited number of carbon emission pathways. The procedures to generate model-simulated data involves using global climate model (GCM) to provide the initial and boundary conditions for the RCMs, that operate at a much finer time and space resolution, but restricted to the European scale.[12] According to the regional climate model, snow characteristics are either directly simulated and available as gridded output data, or can be inferred from temperature, radiation and precipitation data using dedicated models that simulate the fine-scale behavior of snow accumulation or snow melt (as, e.g., the French SAFRAN/CROCUS model,[13] now used operationally for forecasting).

Observed Changes in Winter Temperature and Precipitation, and Consequences for Snow

Because temperature and precipitation are the principal weather determinants for snow, a brief overview is given here of the behavior of these key parameters in the Swiss Alps in the time-frame ranging from 1931 to 2010. The map in Figure 1 shows the location and altitude of the climatological stations used in the analyses conducted in the framework of this article. The locations for which temperature and temperature data is discussed in this article have not undergone any significant changes capable of biasing the results and trend estimates.[9]

Figure 2(a) shows the winter-time mean temperature anomalies for seven selected sites, computed by taking the average of daily data over the months of December, January, and February (DJF), and subtracting the 30-year mean value of the 1961–1990 reference period. These include here the north-eastern Swiss Plateau (Zurich, 556 m above sea level), the western Prealpine region (Château d'Oex, 985 m), the central Alpine Forelands (Engelberg, 1035 m), the eastern Swiss Alps (Davos, 1590 m), and high elevation mountain sites (Saentis, 2505 m, and Jungfraujoch,

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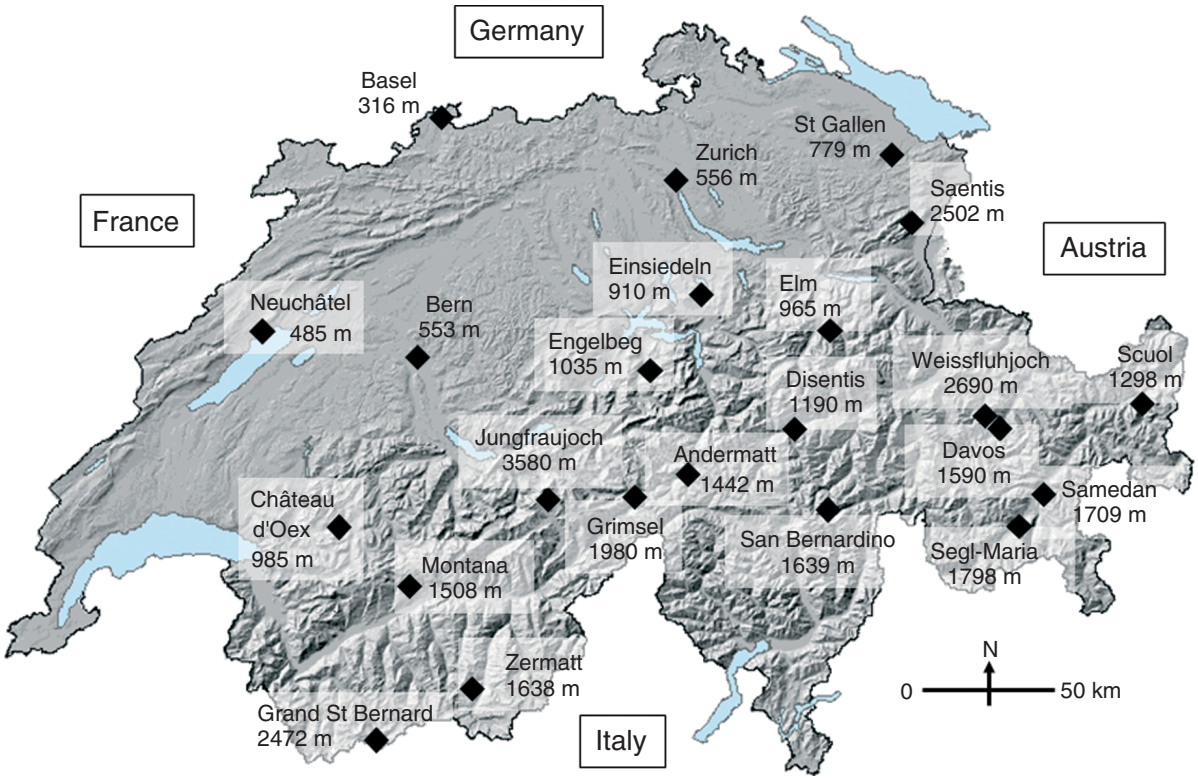


FIGURE 1 | Map of Switzerland showing the climate observation sites used in this article.

3580 m). Whatever their location, the temperature trends are remarkably in phase despite differences in altitude and geographic location (i.e., between the drier or wetter regions of the Swiss Alps), a feature observed for an additional 12 NBCN sites not shown here. Particularly conspicuous are the cold winter of 1962/1963 and the warm winters in 1989/1990 and 2006/2007. Decadal scale wintertime mean temperature trends for the 1931–2010 period range from 0.14 °C per decade in Zurich to 0.40 °C per decade at the Jungfraujoch, with no clear difference between trends at low/medium and those at high elevations. For example, trends at Château d'Oex are almost as large as at Säntis and larger than at Davos. Overall, temperatures in the Alpine region exhibit are far greater rates of change than the global average temperature increase,[14] a feature that has been observed in other mid-latitude mountain regions.[1] Reductions in snow cover and duration, that contribute to positive feedback effects between the surface and the atmosphere, may partly explain these observations. Use of the nonparametric Mann–Kendall statistical test to assess whether trends in time series are statistically significant shows that all temperature trends illustrated here are highly significant beyond the 99.9% confidence level. Figure 2(b) illustrates the changes in winter precipitation anomalies for the same locations as above (except for the Jungfraujoch, where precipitation measurements are absent in view of its extreme location). As for temperature, the precipitation records are well in-phase with each other, despite widely varying differences in site characteristics between stations. In general, precipitation increases with height, but there is also a decreasing precipitation gradient from north–south and west–east. Despite the large variability in the precipitation records, there is an overall decline over the 1931–2010 period of 15–25% in winter-time precipitation amount according to location except for the high-elevation Saentis site which exhibits an increasing trend of the 80-year record, although the last decade has seen a decline as for the other sites. At Engelberg, for example, precipitation has ranged from 234 mm/season (the decade beginning in the winter 2000/2001) to 320 mm/season (decade of the 1960s), a difference of almost 30%). Precipitation in the latter part of the record is comparable to that experienced in the 1970s, where a succession of drier-than-average winters was observed. These decadal-scale fluctuations are likely linked to long-term shifts in synoptic weather patterns between the North Atlantic, the Mediterranean, and continental

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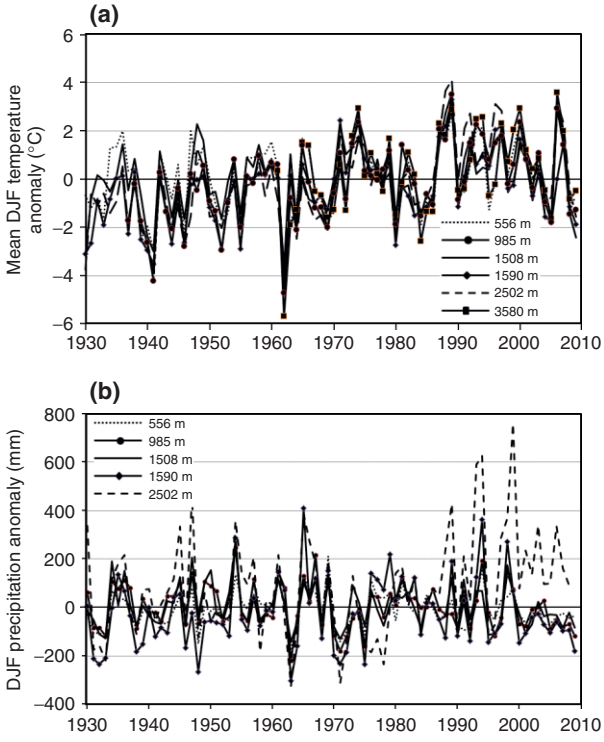


FIGURE 2 | (a) Evolution of mean winter (December–January–February, or DJF) temperature anomalies (based on the baseline period 1961–1990) from the winter 1930/1931 to the winter 2009/2010 (80-years) for six representative sites ranging from 550 to 3580 m above sea level (Zurich, 556 m; Château d'Oex, 985 m; Montana, 1505 m; Davos, 1590 m; Saentis, 2502 m; Junfraujoch, 3580 m). (b) As for (a), except for total winter precipitation (without the Jungraujoch site).

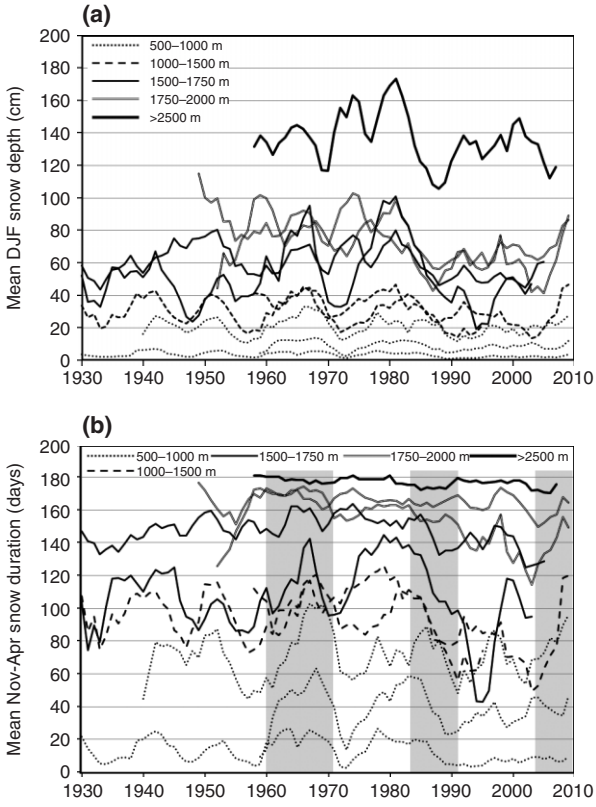


FIGURE 3 | (a) Mean winter snow thickness for the winter months December. January–February–April for 10 representative stations ranging from 500 to 2700 m above sea-level (Zurich, 556 m; St Gallen, 779 m; Château d'Oex, 985 m; Engelberg, 1035 m; Scuol, 1298 m; Montana, 1508 m; Davos, 1590 m; Segl-Maria, 1798 m; Arosa, 1840 m; Weissfluhjoch, 2690 m). Values have been smoothed with a 5-year running mean to remove the high noisiness of interannual snow variability. (b) As for (a), except for mean snow duration beyond a 10-cm depth threshold for the 6-month period November–April.

Europe. Unlike temperature, there are no significant precipitation trends, whatever the altitude considered.

The change in the behavior of the alpine snow-pack is provided in Figure 3(a) and (b) (changes in mean snow depth and duration of continuous snow cover), for 10 sites located at different altitudes ranging from 556 m in Zurich to 2690 m at Weissflujoch. For the purposes of clarity, a 5-year filter has been applied to the data to remove the high interannual variability of alpine snow cover and highlight possible long-term trends. There is an obvious altitudinal dependency of snow depth on altitude, but regional characteristics also modulate the snow-pack, such as the south-facing station of Montana in Valais or Segl-Maria in the Engadine (not shown in this graph), both located in a conspicuously dry part of the country. The figure shows that there are decadal-scale fluctuations in snow depth, with varying amplitude according to location. These fluctuations are reasonably in phase, whatever the location, such as the relative peaks in the mid-1960s and late 1970s and the trough in the early 1970s. All sites exhibit a decline in the average winter snow-depth that range from about 10% in Château d'Oex to over 50% in Zurich. The moister northern part of the Alps has experienced a lower decline (10–25%) than the drier parts of the country, such as the Engadine in south-eastern Switzerland, with reductions in snow depth of about 35% despite the higher elevation.

In Figure 3(b), the duration of the snow season (defined here as the number of days where snow exceeds a 10-cm thickness threshold) shows that there is a general shortening of the season at all levels, with declines stronger in the drier parts of the country (e.g., in the Engadine, not shown in this graph with about 15% less days over the period of record) and at low elevations (e.g., in Zurich where the average duration has been reduced by 50%). In most other locations, the general decline is between 5 and 15%, although there are marked decadal-scale changes, such as the

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increases in the 1960s (shaded column to the left), or the recent increases in snowy winters toward the end of the record (shaded column to the right). For example, in the south-facing station of Montana, the average decadal snow duration ranges from 79 in the 1990s to 126 in the 1980s, and snow amounts since 2000 are roughly the same as in the 1930s. In the more precipitation-prone location of Engelberg, the 1990s saw on average 65 days of snow per winter compared to 105 in the 1960s, a figure that is close to observed in the latest decade.

A decline is visible in the late 1980s (shaded column at the center of the figure), a feature that has been referred to as a regime shift by Marty,[15] who suggests that for many locations in Switzerland, there is a clear step-like decrease in the number of days with snow on the ground and unclear trends thereafter. In the last years of the record, an increase is observed as a result of the snow-abundant winter of 2008–2009 and the snowier-than-average winter that followed.

The noisiness of the snow record on annual-to-decadal time scales makes it difficult to assess visually whether an underlying trend exists. Applying the Mann–Kendall test to the trends over the period of observation shows that snow-depth and duration trends are not significant beyond the 95% confidence level at low to intermediate levels, and exhibit no significance at elevations beyond 2,000 m where atmospheric warming has not yet begun to disrupt the behavior of the mountain snow-pack.

Causes of Observed Change in the Behavior of the Alpine Snow-Pack

Temperature and Precipitation Trends

The most obvious explanation for the observed changes is to be found in the strong rise in mean, minimum and maximum temperatures recorded over the past century at all elevations. Temperature has a more important impact on snow below altitudes around 1500–2000 m than above, since even with the higher temperatures observed in recent years at high elevations, they nevertheless remain below the freezing point for much of the season and thus snow can accumulate and remain on the ground for extended periods of time. Any changes in snow behavior at high altitudes are thus for the moment a function of changes in precipitation. As seen in Figure 2(b), there have been varying amounts of precipitation over the 80-year period under consideration, with obvious impacts on total seasonal accumulation illustrated in Figure 3(a). At lower elevations, however, increases in precipitation have not always managed to compensate for increases in temperature, because in many instances precipitation falls in the form of rain and thus tends to undermine a pre-existing snow-pack, as seen for Château d'Oex in Figure 3. Although precipitation remains roughly the same for the 1940-1970 and the 1980–2010 periods, snow abundance is seen to decrease by 35% as a result of temperatures that are on average 1.8 °C higher for minimum temperatures and 1.2 °C higher for maximum temperatures. Table 1 summarizes the climatic influence on snow duration for three alpine sites, in the form of the change in the average number of days as a function of 1 °C shifts in minimum and maximum temperatures, 100 mm changes in precipitation, and 10 Pa changes in pressure. Each additional degree rise in temperature reduces the duration of snow on the ground by up to 12 days according to altitude and local site characteristics of the weather station (slope, exposure, etc.), whereas increases in precipitation by 100-mm increments contribute to increases in snow duration that is clearly altitude-dependent, with higher altitudes extending their snow season when precipitation is greater. High pressure periods, on the contrary, have a bearing on snow duration, through the associated reduction in precipitation that they imply. Lower elevations are much more sensitive to high pressure than higher altitudes, since the presence of high pressure can prevent the snow season from beginning or contribute to the melting of a thin layer of snow, whereas at higher levels, a period of persistent high pressure that intervenes once snow is already on the ground will not significantly change the duration of the snow season.

TABLE 1 | Average Changes in Snow Duration at Various Altitudes When using Winter Minimum or Maximum Temperatures, Precipitation or Pressure as Predictors for Snow Duration

	T_{min} Days/°C	T_{max} Days/°C	Precip Days/ 100 mm	Pressure Days/10 Pa
Davos (1590 m)	−9	−12	8	−8
Ch. d'Oex (980 m)	−6	−8	4	−16
Zurich (567 m)	−8	−7	2.5	−22

In Figure 3(b), the stations located below the 1500–2000 m level (i.e., all but the four uppermost curves) reflect the dominant control of temperature over precipitation in determining the amount of snow that ultimately accumulates on the ground, particularly in the 20-year period between the early 1980s and late 1990s. The temperature influence is greatest at the lower elevations where the number of freezing days is the lowest. Transitions from snow to liquid precipitation at increasingly higher elevations, under increased winter temperatures, may be one explanatory factor for these trends.

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Influence of Weather Types

It is possible to quantify the characteristics of winters not just in terms of the changes in temperature or precipitation taken separately, but in terms of the behavior of joint quantile distributions, as shown in earlier studies.[16] Particular combinations of temperature and precipitation have been shown to be closely related to the underlying atmospheric circulation patterns that are responsible for either snow-abundant or snow-sparse winters in the Alps.[17] This is because cold winters are not always associated with abundant snow, and dry winters can still have snow if prior to a dry spell snow has accumulated in sufficient quantities.

In order to define joint temperature/precipitation quantiles, use is made of the quantile thresholds defined in 2009 by Beniston,[18] where the frequency of days when temperature and precipitation are simultaneously above or below the 25% and 75% quantiles have been counted. Hence Cold/Dry conditions prevail when the joint quantiles are equal to or below $T_{25}p_{25}$, (the subscript refers to the threshold for each variable), Cold/Wet ($T_{25}p_{75}$), Warm/Dry ($T_{75}p_{25}$), and Warm/Wet ($T_{75}p_{75}$). Daily mean temperature and precipitation statistics for the winter half year, that is, from early October to end April, for the reference period 1961–1990 have been used to define the reference quantile thresholds.

Figure 4 illustrates the rather significant mode changes that have been observed at the Zurich site, by comparing the average number of days per winter for each decade since the 1930s. The most conspicuous change is that of the Warm/Dry mode that doubled in persistence between the 1960s and the 1990s and still remains a dominant mode even if it has declined since 2000. Less marked but just as important in its relative change is the Warm/Wet mode that has also doubled between the early and most recent parts of the record. The Cold/Dry and Cold/Wet modes have declined by roughly 50%. In other words, the dominance of warm modes is clear today, in sharp contrast to the period prior to the 1970s when the frequencies of cold and warm modes, whether dry or moist, were roughly equivalent.

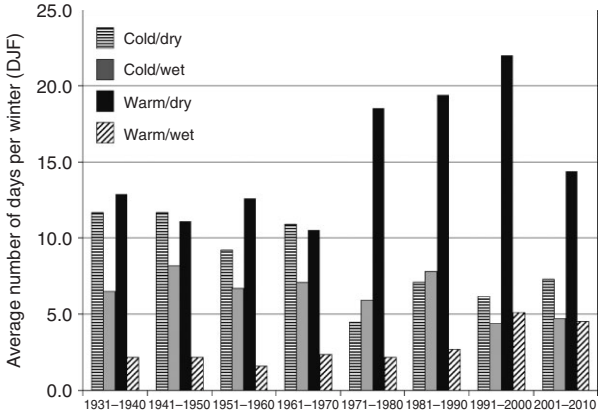


FIGURE 4 | Decadal changes in the four winter modes Cold/Dry, Cold/Wet, Warm/Dry, Warm/Wet for the Zurich measurement site.

Snow behavior is sensitive to these four modes. Clearly the transition toward a majority of warm modes implies that it is difficult for sustained snow cover to persist at low to medium elevations. At higher altitudes, the amount and duration of snow will be controlled by precipitation amounts, since whatever the mode, temperatures at height still remain below the freezing point for much of the season. In a recent study, Beniston and co-workers[17] showed that snow-abundant winters at most altitudes are not so much governed by, for example, the Cold/Wet mode, but by a reduced frequency of the Warm/Dry mode. If the number of winter days within this mode is 1 standard deviation or more below its average value for the 1961–1990 baseline, then the likelihood of having a snow-abundant winter is high; conversely, if the number of days is 1 standard deviation or more above the mean frequency, then the winter will be generally snow-sparse. Since the tendency for this weather type has increased substantially over the past 40 years, the number of snow-sparse winters has increased while snow-abundant winters have declined. Nevertheless, several winters since the 1990s have been endowed with much snow in the Alps, such as the 1998–1999 winter or the 2008–2009 season that were in some parts of the mountains record-breaking winters in terms of snow amount and duration, despite the overall context of a dominance of the Warm/Dry mode.

Influence of the North Atlantic Oscillation

The North Atlantic Oscillation (NAO) is widely recognized as one of the major decadal-scale modes of climate variability that influences features of winter weather from the eastern seaboard of North America to large portions of western and northern Europe.[19] The NAO is related to the relative intensities of two major centers of activity that influence the mid-latitude Atlantic, namely the Azores high pressure anticyclone and the Iceland low pressure zone. This pressure dipole determines the intensities and trajectories of storm tracks across the North Atlantic. The positive (or warm) mode of the NAO, is associated in general with drier and warmer than average conditions over southern Europe and the Mediterranean, and mild but wetter conditions in northern Europe,[20] whereas during the negative (or cold) phase of the NAO, the

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situation is reversed (hence the term 'oscillation'). The Alps are located in a pivotal region and often exhibit Mediterranean characteristics during positive NAO events and northern/continental European weather patterns during the negative phase. The behavior of the mountain snow-pack can thus be expected to be fairly sensitive to that of the NAO, in particular the wintertime (DJF) value of the index that will be used hereafter.

Numerous studies have investigated the links between the NAO and alpine climate characteristics.[20–22] One of the most significant findings is that a strongly-positive phase of the NAO is related to a strong persistence of high pressure fields over the Alps,[20] that in turn negatively affect precipitation since atmospheric perturbations are routed either far to the north of the Alps or well south in the Mediterranean region. In addition, strong correlations have been established between anomalously warm winter temperatures at most elevations during periods with a strongly positive NAO index. Other synoptic patterns have also been explored[21]; this particular study suggests that 50% of the variability of snow-pack behavior is related to the establishment of blocking patterns over Europe, and that the direct influence of the NAO on snow pattern is less obvious. It is indeed through a 'cascade' of consequences (NAO mode on pressure fields; pressure fields on precipitation; and precipitation/temperature thresholds on snow) that the NAO influence can be detected. For a study undertaken for the entire French alpine chain, Durand et al.[22] have not been able to detect a significant influence of the NAO on snow-pack behavior. It is worth noting, however, that the left-hand and right-hand shaded areas in Figure 3(b) correspond to periods of lower-than-average NAO index values, and more extended snow duration, while the central shaded area with the declines in snow duration correspond to some of the strongest positive modes of the NAO in the second half of the 20th century.

ALPINE SNOW COVER IN A 'GREENHOUSE CLIMATE': POSSIBLE TRENDS UP TILL 2100

Many studies have focused on possible trends of mountain snow-packs[23–26] because there is an interest by various sectors (e.g., ski operators or the hydropower companies) to know how this economically-important resource may change in the future. The Intergovernmental Panel on Climate Change (IPCC) has also explored these issues from both the physical perspective and that of the environmental and socio-economic impacts of changing snow cover.[27]

Most studies have focused on the end of the 21st century, as this is the period that has received most attention in the climate-modeling community, because by 2100 the anthropogenic climate signal is likely to be stronger than at any other time during the 21st century. This may not necessarily be the most useful time horizon for policy making and for formulating adaptation strategies, however, and an increasing number of studies and research projects are considering projections of change within the first half of the 21st century (e.g., the EU 'ACQWA' project[28] and its policy implications[29]).

Climate model projections *per se* are generally not sufficient to assess the future behavior of snow in complex topography. It is thus necessary to interface some form of downscaling technique that enables to 'translate' the grid-point values of climate parameters into an estimate of snow depth and duration at a very local scale. Statistical downscaling techniques[30] can be applied, but also physically based snow models[13,31] that use RCM-generated outputs as initial and boundary conditions for a snow and surface energy-balance model.

Whatever the technique used to generate estimates of snow depth/duration, all studies reach the same conclusion, namely that in a warming climate, snow will become increasingly rare below about 1500–2000 m altitude in the Alps (although this figure can vary according to the orientation and the location within the alpine domain, that experiences both north–south and east–west precipitation gradients). RCM projections for the alpine region[32] suggest a range of 3–5 °C increase in mean temperatures and an increase in seasonal precipitation by 5–20%. Using this information, Beniston et al.[33] focused upon the temperature-precipitation-snow depth relations and the changes in snow volume that these levels of climatic change would imply. Table 2 shows the

TABLE 2 | Estimates of Changes in Snow Volume in the Alps between the Baseline 1961–1990 Climate and a Scenario Climate Based on the IPCC 'A2' Greenhouse Emissions Pathways by 2100

Elevation m	Current Snow Volume 10^9 m^3	Future Snow Volume 10^9 m^3	Change in Volume Future-Current
500	6–8	0–0.5	98–100% loss
1000	15–20	0.2–1	95–98% loss
1500	28–35	8–12	65–70% loss
2000	40–50	20–28	50–60% loss
2500	30–38	18–22	40–45% loss
3000	8–10	8–10	<5% change
3500	2–3	2.5–3.5	15–20% gain
4000	0.8–1.2	1.0–1.4	15–20% gain

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range of snow volume for the baseline 1961–1990 climate, which reflects both geographical site characteristics and the variability of winter temperature and precipitation. On the basis of the plausible range of temperature and precipitation using the IPCC high emissions 'A2' scenario,[34] it is seen that by the end of the 21st century, low to medium elevations (below 1500 m above sea-level) will experience between 70 and 100% loss of snow volume, while with increasing height the loss diminishes and even increases above 3000 m. The conclusions reached here require caution in their interpretation, because RCM-simulated snow cover does not always correspond to reality. Indeed, simulated snow cover is seen to be highly sensitive to physical parameterization schemes, model grid resolution, and the type of atmospheric forcing that is considered.[35] The results discussed here simply attempt to show the type of changes that could intervene by 2100, but should in no way be considered to be predictions of future snow behavior.

A recent study by Uhlmann et al.[26] has shown that for many ski domains in Switzerland, the conditions for economically-viable skiing (i.e., at least 80–100 days with snow exceeding the 30-cm threshold[7]) will no longer hold for numerous resorts. Table 3 highlights the changes in the number of days that exceed the 30-cm threshold for a number of representative sites in the Jura Mountains, and the western, central, and eastern Swiss Alps. Reductions in snow duration from 30 to 100% are projected, and are dependent on altitude and geographic location. The numbers provided in Table 3 suggest that many resorts will need to diversify rapidly their tourism offer in order to be less dependent on the ski industry for the essential part of their financial revenues. Although many communities have installed snow-making equipment in order to buffer the effects of certain adverse snow seasons, this will become increasingly obsolete as winter temperatures exceed the critical thresholds above which snow-making equipment cannot be used for snow-making.

TABLE 3 | Changes in Snow Duration for Various Swiss ski Resorts between the Baseline 1961–1990 Climate and a Scenario Climate Based on the IPCC 'A2' Greenhouse Emissions Pathways by 2100

Resort, Altitude and Region	Current Duration	Future Duration
Chaux-de-Fonds, 1000 m, Jura Mountains	30	0
Adelboden, 1350 m, west-central Alps	60	20
Engelberg, 1100 m, central Alps	120	30
Disentis, 1100 m, eastern Alps	90	30
Davos, 1600 m, eastern Alps	100	70
Zermatt, 1600 m, south-central Alps	75	40

CONCLUSION

This overview of the behavior of snow has focused essentially on the case of the European Alps and more particularly on Switzerland, as there is a wealth of relevant data and long records that enables detailed studies to be undertaken.

It has been shown that, over the past 80 years, snow has exhibited considerable interannual and interdecadal variability. As a general rule, the fact that winter temperatures have risen over the intervening period by 2 °C or more according to location implies that, particularly at elevations below 1000–1500 m, the long term trends in snow point to a general decline in both duration and total accumulation. This linear trend since the early part of the 20th century is to some extent masked by decadal-scale fluctuations where snow thickness and duration does actually increase for a number of years, as seen in the most recent part of the record.

The conclusions reached in this article confirm other recent findings undertaken in Switzerland and in neighboring alpine countries.[36–38] For example, Schöner et al.[40] have shown for one of the longest climate records in Austria that '*both maximum snow-depth and winter season precipitation show a clear decreasing trend for interannual variability,*' a conclusion also reached by Hantel et al.[41] for an extensive study of the European Alps in which winter temperature can be considered to be a close proxy for snow duration. Changes in the alpine freeze–thaw mechanisms during winter in particular in the Italian Alps have been studied[41] to assess how the impacts on snow affect soil leaching processes and their influence on living organisms. An extensive study in the French Alps[22] shows that for a half-century period, '*Snow patterns in the French Alps are characterized by a marked declining gradient from the northwestern foothills to the southeastern interior regions. This applies mainly to both depths and durations, which exhibit a maximal latitudinal variation at 1500 m of about 60 days, decreasing strongly with the altitude.*'

Under current climatic conditions, snow cannot be considered to be 'disappearing' but, certainly, it is beginning to recede at low elevations. According to climate model projections, however, it is likely that at elevations below perhaps 3000 m, snow will recede increasingly quickly. This is because the higher levels of precipitation projected by many RCMs for the Alps will not compensate for the strong rise in

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winter temperatures. As a consequence, snow may well have disappeared in the lowland and foothill regions around the alpine chains by the end of the 21st century, and will have significantly declined at medium levels. The possible increases in snow amount above 3000 m represent only a fraction of the total surface area and snow volume that the Alps have experienced up till now.

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The Response of Northern Hemisphere Snow Cover to a Changing Climate*

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(Manuscript received 5 June 2008, in final form 14 October 2008)

ABSTRACT

A snowpack model sensitivity study, observed changes of snow cover in the NOAA satellite dataset, and snow cover simulations from the Coupled Model Intercomparison Project phase 3 (CMIP3) multimodel dataset are used to provide new insights into the climate response of Northern Hemisphere (NH) snow cover. Under conditions of warming and increasing precipitation that characterizes both observed and projected climate change over much of the NH land area with seasonal snow cover, the sensitivity analysis indicated snow cover duration (SCD) was the snow cover variable exhibiting the strongest climate sensitivity, with sensitivity varying with climate regime and elevation. The highest snow cover–climate sensitivity was found in maritime climates with extensive winter snowfall—for example, the coastal mountains of western North America (NA). Analysis of trends in snow cover duration during the 1966–2007 period of NOAA data showed the largest decreases were concentrated in a zone where seasonal mean air temperatures were in the range of −5° to +5°C that extended around the midlatitudinal coastal margins of the continents. These findings were echoed by the climate models that showed earlier and more widespread decreases in SCD than annual maximum snow water equivalent (SWEmax), with the zone of earliest significant decrease located over the maritime margins of NA and western Europe. The lowest SCD–climate sensitivity was observed in continental interior climates with relatively cold and dry winters, where precipitation plays a greater role in snow cover variability. The sensitivity analysis suggested a potentially complex elevation response of SCD and SWEmax to increasing temperature and precipitation in mountain regions as a result of nonlinear interactions between the duration of the snow season and snow accumulation rates.

1. Introduction

Snow cover represents a spatially and temporally integrated response to snowfall events, and the sequence of snowfall and melt events determines not just the quantity of water stored as snow but also snowpack condition (e.g., grain size and compaction), which in turn determines avalanche risk, energy required for melting, albedo of snow, and much more. Snowpack takes on special significance in mountain regions where snow stores enormous quantities of water, altering the ecologic and economic balance of regions far downstream by delaying the release of water months after precipitation events. Persistent changes in snow accumulation or melt can therefore have significant ecologic and economic consequences. Snow cover is also integrally linked with observed changes in global climate, especially for Northern Hemisphere (NH) land areas, through its role in modifying surface albedo. Observed monthly mean snow cover extent over the NH is strongly anticorrelated with air temperature, with the strongest relationships in the April–June period when extensive snow cover coincides with strong solar radiation (Déry and Brown 2007). Recently, Barnett et al. (2008) provided evidence of an anthropogenic signal in changes in snowpack and the hydrologic regime over the western United States.

Snow cover is anticipated to decrease in response to global warming, as snow cover formation and melt

* Joint Institute for the Study of the Atmosphere and Ocean Contribution Number 1584.

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DOI: 10.1175/2008JCLI2665.1

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are closely related to a temperature threshold of 0°C. However, as first noted by Groisman et al. (1993) and more recently by Räisänen (2007), the snow cover response to global warming is complicated by projected increases in precipitation, particularly over high latitudes. The snow cover response to warming could therefore vary with latitude and elevation, with potential for increased accumulation in high latitudes and high elevations where increases in precipitation are sufficient to offset reductions in the length of the accumulation season. During the process of developing a synthesis of historical snow cover trends for the recent Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (Lemke et al. 2007), the authors were struck by the large variability in the published literature—for example, snow cover trends in the mountain regions of Europe are characterized by large regional and altitudinal variations (e.g., Vojtek et al. 2003; Scherrer et al. 2004; Brown and Petkova 2007); snow cover over North America (NA) has increased in the fall half of the year but decreased in the spring (Brown 2000); snow cover trends in China exhibit strong regional variations (e.g., Zhang et al. 2006; Qin et al. 2006; Zhao and Moore 2006; Changchun et al. 2007; Yang et al. 2007) despite widespread warming, glacier wasting, and permafrost thawing (Li et al. 2008); and snow depth trends over northern Eurasia show contrasting increases in the northeast and decreases in the west (Popova 2007).

Published trends in various snow cover variables from in situ data are summarized graphically in Fig. 1. Note that spatial sampling is highly irregular, and Fig. 1 is meant primarily as an inventory of published research (some mentioned above and some not) rather than an integrated global study. The panels distinguish types of snow quantities and in some cases seasons. Although each panel has some data points with positive trends and some with negative trends, the fraction of negative trends is larger for snow cover duration (SCD) than for midwinter snow depth and larger still for spring snow depth (SD) and snow water equivalent (SWE).

Two important factors contributing to this variability are the different periods used in the various studies and atmospheric circulation patterns such as North Atlantic Oscillation (NAO), which plays a role in the contrasting trends in Eurasian snow depth reported by Popova (2007). Other sources of variability are related to measurement issues (e.g., the spatial representativeness of point snow depth measurements, changes in measurement procedures, and others) and to the fact that the different variables used to monitor snow cover respond differently to a changing climate. For example, SD and SWE are more sensitive to changes in snowfall amount

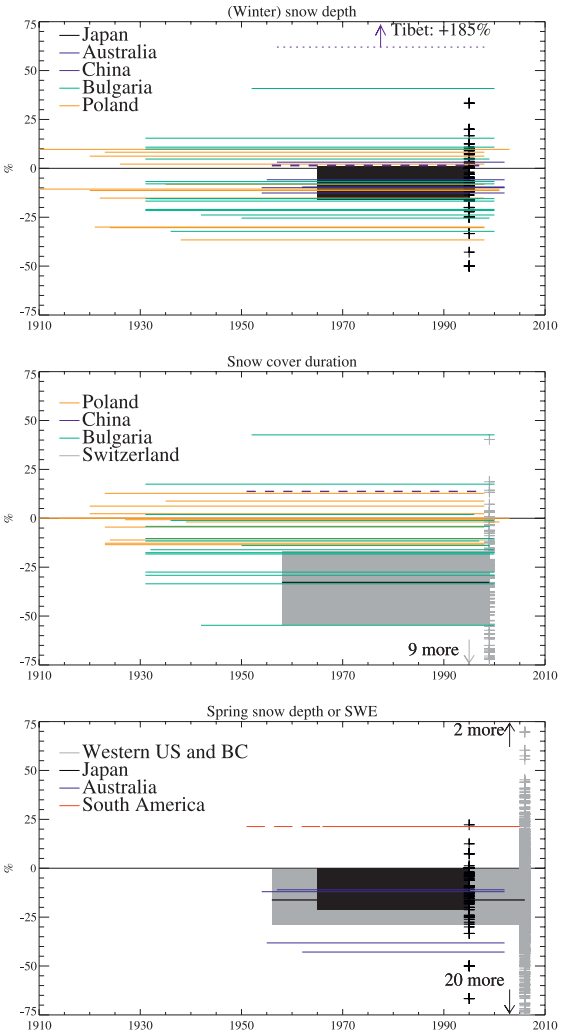


FIG. 1. Summary of published works describing trends in various snow quantities. The country shown is color-coded, and each line indicates the magnitude of the trend (*y* axis) and the duration (*x* axis). Dashed lines are estimated; thick lines are area averages; studies with large amounts of data (>50 points) are summarized with a colored box indicating the 25th and 75th percentiles, with a black line for the median; and pluses at the ending year indicate the trends at individual points. (top) Snow depth, which is computed as January–February, average for Japan (Ishizaka 2004), annual average for northwest China and Tibet (Qin et al. 2006); and maximum annual value for the four Australian stations (Hennessy et al. 2003), Bulgaria (Brown and Petkova 2007), and Poland (Falarz 2004). (middle) Annual snow cover duration for northwest China, Bulgaria, and Poland; and winter (December–February) duration for Switzerland (Scherrer et al. 2004). Note that nine Swiss stations had trends less than −75%, beyond the range of the plot. (bottom) SWE on 1 Apr for 11 western states and British Columbia (Mote 2006; note that 22 states/province lie beyond the range of the plot), March SD for Japan, 1 Sep SD for Australia, and an average of 6 snow courses (maximum annual SWE, usually in September) in Argentina and Chile (Masiokas et al. 2006); before 1966, only two to five snow courses were available, so the average is shown as a dashed line.

than snow-covered area (SCA) or SCD, which are closely controlled by air temperatures at the start and end of the snow cover season. SCA and SCD are also more closely linked to albedo feedbacks, which are stronger during the spring period (Groisman et al. 1993; Déry and Brown 2007) and over mountain regions (Giorgi et al. 1997; Fyfe and Flato 1999). The response of snow to climate change in mountain regions is further complicated by vertical gradients in temperature and precipitation. Typically, the largest relative changes are observed at lower elevations (e.g., Laternser and Schneebeli 2003; Scherrer et al. 2004; Mote et al. 2005; Mote 2006). However, there is evidence at some higher elevation sites of increased peak snow accumulation (e.g., Vojtek et al. 2003; Mote et al. 2005; Regonda et al. 2005) in response to increasing precipitation (Zhang et al. 2007).

The main purpose of this paper is to attempt to make some order out of the reported changes by placing observations of twentieth-century snow cover change into a framework of observed and expected changes. The expected changes were identified through (i) the sensitivity analysis, with a snowpack model to determine the response of different measures of snow to changes in temperature and precipitation; and (ii) the analysis of twentieth- and twenty-first-century changes in snow cover simulated by climate models to examine the spatial pattern and magnitude of snow cover response to warming. These were compared to observed changes in NH snow cover during the 1966–2007 period with the NOAA dataset (Robinson et al. 1993). The results from these analyses should provide some guidance on when snow-climate regions and elevation zones are likely to exhibit the earliest response to a changing climate and provide some context for interpreting published literature on changes in snow cover. This paper is organized as follows: a description of data sources used in the sensitivity analysis is provided in section 2; the methodologies for the snowpack model sensitivity analysis and the analyses of the NOAA dataset and the climate model output are presented in section 3, and the results are presented and discussed in section 4.

2. Data sources

a. Climate station data

Daily maximum and minimum temperature and total precipitation data for the 1961–1990 period from four different sites in Canada (Table 1) were used in the snowpack sensitivity analysis. These sites come from four of the main snow-climate categories identified over the NH by Sturm et al. (1995): (i) the *maritime* class, characterized by relatively mild winter temperatures and high precipitation (Tahtsa Lake, British Columbia, henceforth referred to as TL); (ii) the *prairie* class, characterized by low precipitation, high winds, and variable winter temperatures (Saskatoon, Saskatchewan, henceforth referred to as SK); (iii) the *tundra* class, characterized by low precipitation, high winds, and cold temperatures (Resolute Bay, Northwest Territories, henceforth referred to as RB); and (iv) the *taiga* class, characterized by relatively cold temperatures and higher precipitation (Goose Bay, Newfoundland and Labrador, henceforth referred to as GB). Although the GB region is identified as taiga in the Sturm et al. (1995) classification, it is influenced by maritime air masses during winter and accumulates more snow than taiga sites located further inland. This site is still useful, however, as it is typical of the snow climate of central Quebec and Labrador, which has the largest annual snow accumulation in North America east of the cordillera. RB lies inside the November–May[1] −20°C isotherm, where climate models show an increase in maximum SWE in response to climate warming (Räisänen 2007; refer to Fig. 12). The relatively mild, heavy precipitation environment of TL in the British Columbia coastal mountains is typical of the Pacific Northwest snow climate, which has been shown to be particularly sensitive to climate warming (e.g., Hamlet et al. 2005; Bales et al. 2006). At this site, nearly 60% of the mean daily air temperatures below freezing are warmer than −5°C (Table 1), compared with ~25% for GB and SK and <15% for RB.

b. NOAA satellite snow cover extent

National Oceanic and Atmospheric Administration satellite snow cover extent data were used to evaluate the snow cover climatologies of climate models from the Coupled Model Intercomparison Project phase 3 (CMIP3) multimodel dataset (Meehl et al. 2007) and to examine snow cover trends during the 1966–2007 period. The dataset is described in Robinson et al. (1993) and consists of weekly charts of the presence or absence of snow cover on a 190.5-km polar stereographic (PS) grid over the NH. The charting method changed in May 1999 with the introduction of the higher resolution (~25 km) daily interactive multisensor (IMS) snow cover product (Ramsay 1998). A pseudoweekly product has been derived from the IMS daily product by taking each Sunday map as representative of the previous week. This has resulted in obvious inconsistencies in snow

[1] Räisänen defined his −20°C isotherm for the November–March period, but the November–May period isotherm gives slightly better visual agreement with the zone where climate models indicate SWE will increase.

TABLE 1. Summary of climate stations used in the snow cover sensitivity analysis.

Climate stations	Location	Snow-climate class[a]	Elevation (m)	Mean annual snowfall (cm)[b]	Mean winter temperature (°C)[b]	Freezing temperature warmer than −5°C[c] (%)
Resolute Bay	74.7°N, 95.0°W	Tundra	67	83.8	−31.5	13.5
Saskatoon	52.2°N, 106.7°W	Prairie	501	113.1	−16.0	24.5
Goose Bay	53.3°N, 60.4°W	Taiga	46	445.2	−14.6	26.5
Tahtsa Lake	53.6°N, 127.7°W	Maritime	863	1041.2	−6.8	57.5

[a] From Sturm et al. (1995).
[b] From 1951 to 1980 published climate normals, Environment Canada.
[c] From observed daily climate data, 1961–90.

cover time series at 32 grid points in mountain and coastal areas, which were removed from the analysis. A homogeneity assessment of NOAA fall and spring snow cover duration with corresponding snow cover duration series from 133 stations across Canada indicated no evidence of a discontinuity in 1999 with the change in procedures, but there was evidence of a systematic improvement in the product with linear increases in the correlation between NOAA and corresponding surface observations in both the fall and spring snow seasons, and a statistically significant (0.05 level) reduction in the difference between NOAA and stations of −0.24 day yr^{-1} in the spring period. This is likely related to the increasing frequency and resolution of satellite coverage over time and to the increasing skill of analysts and was most marked over mountain regions. There was no change in the NOAA–station difference in the fall season, which may reflect the more rapid nature of snow onset versus spring melt (i.e., less potential for obscuring cloud and patchy snow). This problem requires more detailed analysis with in situ data from other regions of the NH before corrections can be applied. However, it appears from the Canadian evaluation results that the effect of improvements in snow mapping over time is an order of magnitude smaller than the observed trends in SCD from surface observations, so this is unlikely to change conclusions about where snow cover is exhibiting the largest changes. Monthly snow cover duration data were obtained from Rutgers University, which contain the corrections recommended by Robinson et al. (1991) and the Rutgers weighting scheme (Robinson 1993) to correctly partition weekly charts into appropriate months. Trend analysis of snow cover duration was carried out for the snow year (August–July) and for the fall (August–January) and spring (February–July) halves of the snow year using the Mann–Kendall method.

c. WCRP's CMIP3 multimodel dataset

Monthly snow cover and snow water equivalent output from models in the World Climate Research Programme's CMIP3 multimodel dataset (Meehl et al. 2007) were obtained to examine the simulated trends in twentieth-century snow cover and the projected changes in snow cover under the A2 scenario to determine the regions where snow cover signals first emerge. Models were only included if they had SWE or snow cover output for both the twentieth-century 20C3M simulation and the A2 emission scenario. Most modeling groups provided SWE output, but only about half the models had information on snow cover. SWE output was obtained from 14 models and snow cover output from 8 models (Table 2) through the Earth System Grid data portal (available online at ftp://ftp-esg.ucllnl.org). A number of these models are closely related to each other, but Räisänen (2007) found there could be substantial differences in climate between similar models. Although some models provided multiple simulations, distinguished only by different initial conditions, we only examine one simulation (''Run1'') per model, as not all models provided multiple runs.

Monthly snow cover fraction was multiplied by the number of days in a month to obtain estimates of annual and seasonal snow cover duration, as daily model output was not readily available for many of the models. Change in SWE was assessed by looking at the change in annual maximum monthly SWE during a snow year, which was defined as starting in August and ending in July. This takes account of potential shifts in the timing of peak SWE.

d. Northern Hemisphere SWE climatology

It was not the purpose of this paper to carry out a detailed evaluation of SWE simulated by the CMIP3 models. However, it is important to have some idea of how well the models simulate SWE and to identify obvious problem areas if the models are to be used as guidance for interpreting current trends. A high-quality global SWE dataset does not exist for evaluating climate models. Brown et al. (2003) developed a daily SWE analysis for NA for the second phase of the Atmospheric Model Intercomparison Project (AMIP2)

TABLE 2. Summary of climate models used in this study.

Model	Modeling Group	Global grid resolution (nlon × nlat)	Snow cover	SWE
CCSM3	National Center for Atmospheric Research, United States	256 × 128	x	x
CGCM3.1 (T47)	Canadian Centre for Climate Modeling and Analysis, Canada	96 × 48	x	x
CNRM-CM3	Météo France	128 by 64		x
CSIRO Mk3.0	CSIRO Atmospheric Research, Australia	192 × 96	x	x
CSIRO Mk3.5	CSIRO Atmospheric Research, Australia	192 × 96	x	
ECHAM5	Max Planck Institute for Meteorology, Germany	192 × 96		x
ECHO-G	University of Bonn, Germany and Korean Meteorological Agency, Korea	96 × 48		x
GFDL CM2.1	Geophysical Fluid Dynamics Laboratory, United States	144 × 90		x
GISS-ER	Goddard Institute for Space Studies, United States	72 × 46	x	x
HadCM3	Hadley Centre for Climate Prediction and Research, United Kingdom	96 × 73		x
HadGEM1	Hadley Centre for Climate Prediction and Research, United Kingdom	192 × 145		x
INM-CM3.0	Institute for Numerical Mathematics, Russia	72 × 45	x	x
IPSL-CM4	Institut Pierre Simon Laplace, France	96 × 72		x
MIROC32 (medres)	Center for Climate System Research, Japan	128 × 64	x	x
MRI- CGCM2.3.2	Meteorological Research Institute, Japan	128 × 64	x	x

period (1979–97) from historical snow depth observations, but this effort has not been duplicated in Eurasia, where the historical data are spread across many countries and organizations. Roesch (2006) generated a global SWE climatology from the U.S. Air Force (USAF) snow depth climatology (Foster and Davy 1988), but Brown and Frei (2007) show that values of spatially averaged SWE over the NH are likely underestimated by about 50% (or 30 mm) in the spring as a result of shortcomings in the USAF dataset and the specification of snow densities that were too low.

The SWE climatology used in this study was derived from the daily global snow depth analysis generated by the Canadian Meteorological Centre (CMC) since March 1998 and described in Brasnett (1999). The daily snow depth analyses were converted to SWE by applying a snow density lookup table developed from analysis of Canadian historical snow course data (Meteorological Service of Canada 2000) for the six main snow climate classes defined by Sturm et al. (1995; Table 3). All six classes are found in Canada, allowing mean density information to be generalized to the rest of the NH using the 0.5° × 0.5° gridded version of the Sturm et al. (1995) classification (Liston and Sturm 1998). The use of a snow density climatology to estimate SWE is considered valid, because snow depth variability is the dominant factor driving spatial and temporal variability in SWE (Pomeroy and Gray 1995) and because snow density tends to follow the same seasonal evolution with relatively low interannual variability (Brown 2000). Mean annual maximum monthly SWE values during the 2001–06 period were interpolated to a 190.5-km NH PS grid for comparison with the climate models. This relatively short period was the longest period of overlap that could be obtained based on the availability of the observed and simulated data.

3. Methodology

a. Sensitivity analysis

One of the main motivations for this paper was to examine the role of snow cover as a potential climate indicator. The question of which snow properties (e.g., snow cover duration, peak SWE, melt onset date, and others) are most appropriate for monitoring for climate change detection was first posed by Goodison and Walker (1993). Barry (1984) proposed that candidate climate monitoring indices should do the following:

- show high sensitivity to meteorological variables responding to greenhouse gas (GHG) forcing;
- have a short response time;
- have a high signal-to-noise ratio; and
- be consistently measured over an extended period of time.

The first three criteria were tested by carrying out a sensitivity analysis of snow cover simulated by the temperature–index snowpack model described in Brown et al. (2003). The model incorporates most of the temperature-dependent processes included in detailed physical models (e.g., partitioning of precipitation into solid and liquid fractions, melt from rain-on-snow events, specification of new snowfall density, snow aging, and snowmelt). Snowmelt is parameterized as a function of snow density following Kuusisto (1980) to simulate the effect of decreasing albedo on spring snowmelt rates. As a minimum, the model requires daily values of maximum

TABLE 3. Mean monthly snow density (kg m^{-3}) lookup table used to estimate SWE from snow depth based on Canadian snow course observations. The snow classes are those defined by Sturm et al. (1995).

Month	Tundra	Taiga	Maritime	Ephemeral	Prairie	Alpine
October	200.0	160.0	160.0	250.0	140.0	160.0
November	210.7	176.9	183.5	300.0	161.6	172.0
December	218.1	179.8	197.7	335.1	185.1	181.6
January	230.3	193.1	216.5	316.8	213.7	207.2
February	242.7	205.9	248.5	337.3	241.6	241.5
March	254.4	221.8	283.3	364.3	261.0	263.5
April	273.6	263.2	332.0	404.6	308.0	312.0
May	311.7	319.0	396.3	458.6	398.1	399.6
June	369.3	393.4	501.0	509.8	464.5	488.9

(T_{max}) and minimum (T_{min}) temperature and daily total precipitation. For this study, the model was run at a 6-hourly time step to capture the diurnal cycle, with 6-hourly temperatures estimated from weighted values of T_{max} and T_{min} following Anderson (1973). The model assumed a temperature threshold of 0°C for rain/snow separation and a melt threshold temperature (T_{melt}) of −1°C based on an evaluation of the model at several sites across Canada by Brown et al. (2003). A melt threshold less than 0°C is physically realistic, since radiation melt can take place when air temperatures are below freezing. Kuusisto (1984) obtained T_{melt} values of −1.3° and −1.2°C for open and forested sites, respectively, in Finland.

The choice of snow cover variables was based on key properties of a snowpack related to the timing of snow cover onset and melt and the amount of accumulated snow. The most commonly used snow cover properties in the literature are the first and last date of snow on the ground (e.g., Ye 2001); the first and last dates of continuous snow cover, which requires definition of depth and duration thresholds; the duration of snow on the ground, which may involve the setting of depth thresholds for in situ data (e.g., Brown 2000); and variables describing the amount of accumulated snow, such as mean, median, and maximum snow depth or SWE. For this study, the following snow cover parameters were selected and calculated for a snow year defined from 1 August to 31 July:

- first (JDfirst) and last (JDlast) dates of any snow on the ground;
- the start (JDstart) and end (JDend) dates of continuous snow cover defined as the first date with seven continuous days of SWE values ≥4 mm, and the first date with seven continuous days of SWE values <4 mm;
- the snow cover duration (days with SWE ≥2 mm) in the fall (SCD1) and spring (SCD2) halves of the snow cover season and during the entire snow year (SCDann);
- annual maximum SWE (SWEmax); and
- date of annual maximum SWE (JDmax).

Snow cover duration variables were set to missing once the snowpack persisted beyond the end of a snow year. Comparison of the coefficient of variation (CV), or relative dispersion, defined as the ratio of the standard deviation to the mean, for observed and simulated annual snow cover statistics during the 1961–90 period (Fig. 2) at sites from three different snow-climate regions in Canada revealed the snowpack model correctly captured the differences in CV between snow cover variables with the exception of JDmax, where the model simulated a lower interannual variability. This is likely linked to the model not having precise information on the solid/liquid fraction of precipitation. Realistic simulation of the CV is important for assessing changes computed with Eq. (1) (see below).

The snow cover sensitivity analysis was carried out over a temperature (T) and precipitation (P) change range of +4°C and +30%, which approximates the upper range of projected future changes over NH mid-to-high latitudes reported in the recent IPCC fourth assessment report (Christensen et al. 2007). Simulations were run using a 30-yr block of observed daily climate data (1961–90) to evaluate the significance of simulated changes following Eq. (1) below. Simulations were also carried out over 10 250-m elevation increments above the station elevation, assuming a mean lapse rate of −6.5°C km^{-1} to investigate the effect of local changes in elevation on snow trends in the climate zones represented by each of the stations. Precipitation amounts were not adjusted for elevation.

The sensitivity of snow cover variables to changes in temperature and/or precipitation was assessed by computing the Student's t statistic for a change in mean following Fyfe and Flato (1999):

$$(S_{fut} - S_{ref})/\sigma, \quad (1)$$

where S_{ref} is the 30-yr snow cover variable mean for 1961–90, S_{fut} is the 30-yr mean for changed T and P, and σ is the pooled standard deviation for the two 30-yr periods.

These sensitivity experiments or *incremental analogs* (Mearns et al. 2001) are easy to apply but contain some inherent weaknesses that must be acknowledged. First, they ignore local feedbacks in the warming response—for example, the snow–albedo feedback, which is stronger in the spring than in the fall (Groisman et al. 1994; Déry and Brown 2007). Second, they assume that the variance in temperature and precipitation does not change in

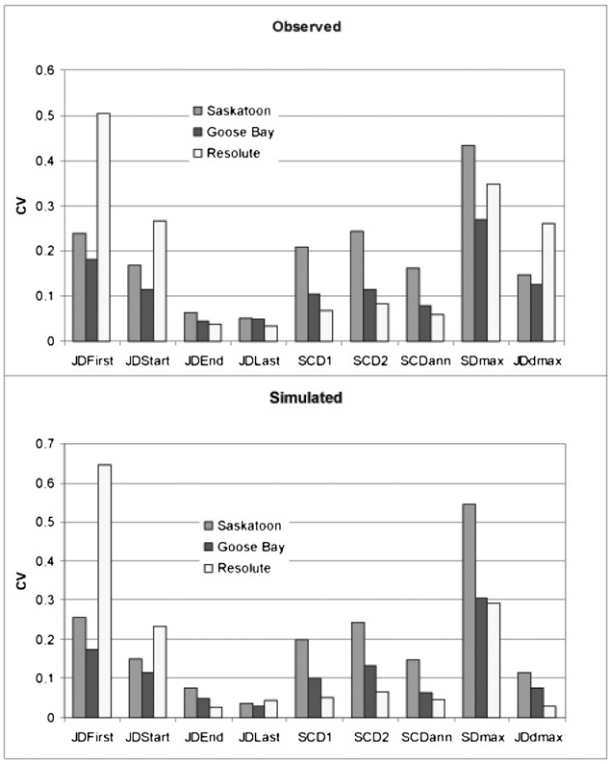


FIG. 2. The CV of annual snow cover variables computed from (a) observed and (b) simulated daily snow depth during the 1961–90 period at three of the sites used in the sensitivity analysis. There were insufficient snow depth observations at TL to compute snow cover statistics. Note that SDmax is used to define maximum accumulation in this comparison.

response to warming. Third, they assume that the seasonality of the precipitation regime remains unchanged and that all regions experience the same warming and increase in precipitation. However, it is argued that this approach is still useful, as it provides information on the *relative* sensitivity of the various snow cover variables in different climate regimes.

b. Trend analysis of observed snow cover

The NOAA dataset was used to document observed trends in snow cover over the Northern Hemisphere land areas from 1966 to 2007. This period was characterized by statistically significant (0.05 level) warming in all months over NH midlatitudes based on analysis of the CRUtem3v dataset (Brohan et al. 2006) with fall and spring season warming of ~1.5°C, which accounts for a large fraction of the warming experienced since 1900. Trends in annual (SCDann) and seasonal (SCD1 and SCD2) snow cover duration during 1966–2007 were computed using the nonparametric Kendall's rank correlation (Sen 1968) and taking serial correlation into account following Zhang et al. (2000). The Kendall estimate is used instead of least squares, as it is less sensitive to nonnormally distributed variables and is less affected by extreme values or outliers in the series. A Monte Carlo method was used to estimate the confidence interval for the estimate slope and the 5% level of significance used to define statistically significant trends. Grid points were excluded from the analysis, where more than half the years had either zero snow cover or were completely snow covered (i.e., points where snow cover duration is strongly skewed) and the trend Z statistic was plotted to give an indication of the relative strength of the changes in snow cover. It was not possible to make inferences about the statistical significance of the trends as a result of the likely presence of technological bias, which requires further work to document and correct.

c. Climate model analysis

The purpose of the climate model analysis was to determine whether they showed any evidence of changes in snow cover during twentieth-century climate simulations and to identify the locations and snow variables where future changes may be first identified. The snow cover variables assessed were seasonal snow cover duration (SCD1, SCD2, and SCDann) and SWEmax. It was not possible to evaluate other snow variables (e.g., onset dates), as daily model output are not readily available for snow-related variables.

Snow cover changes were assessed over 30-yr averages using Eq. (1) to compute the t statistics between a 30-yr period mean and the reference climate mean. The period 1970–99 was defined as the "current climate" reference for future climate changes, which were based on the A2 emission scenario. The 30-yr periods used for assessing future changes were 2010–39, 2040–69 and 2070–99. Twentieth-century changes were assessed with model output from 20C3M runs, with change computed between 1900–29 and 1970–99. The change statistics from the various models were interpolated to a common 190.5-km PS grid (the NOAA grid) using the nearest-neighbor method to summarize the results. The percent of models showing statistically significant changes of the same sign was also plotted on the common grid to obtain an indication of model consensus.

A brief evaluation was made of the models' ability to simulate the observed mean annual snow cover duration and annual maximum monthly SWE climatologies over the Northern Hemisphere, as this was not included in the IPCC model evaluation (Randall et al. 2007), and recent evaluations such as Roesch (2006) only looked at the mean seasonal cycle. Frei and Gong (2005) provided a preliminary evaluation of the CMIP3 models over North America and found significant between-model

variability, with most models underestimating mean continental snow-covered area.

Figure 3 compares the mean model annual SCD climatology for 1970–99 with the corresponding annual SCD climatology from the NOAA dataset with the fall and spring season difference fields presented in Fig. 4. The models do a reasonable job of simulating the mean spatial pattern of annual SCD in agreement with previous evaluations (Foster et al. 1996; Frei and Robinson 1998; Frei et al. 2003, 2005), with spatial correlation coefficients (not shown) ranging from 0.78 to 0.89 across the eight models. The mean model difference in annual SCD averaged over the NH was −5 days. The one area where the models performed poorly was the Tibetan Plateau, where annual SCD was overestimated by an average of more than 100 days. This problem was previously identified by Frei et al. (2003), who attribute this to systematic cold and wet biases in the models over the Tibetan Plateau that still exist in the CMIP3 models (Randall et al. 2007). The underestimation of snow cover over mountain regions in Fig. 4 (e.g., western cordillera of North America) is mainly related to the smoothed topography in most of the models. The spring and fall SCD differences are similar except for a tendency for the models to underestimate SCD in the spring over high latitudes. However, recent evidence from northern Canada indicates the NOAA dataset has a delayed spring melt response (Wang et al. 2005; Brown et al. 2007), which could account for some of this difference.

The SWE climatology comparison was carried out during the 2001–06 period, which was the longest period of overlap between the GCMs and the CMC-estimated SWE values (Fig. 5). The Goddard Institute for Space Studies (GISS) model was excluded from the model mean, as twenty-first-century data only started in 2004. Analysis of individual model spatial correlations and differences (not shown) yielded lower correlations than annual SCD, with little difference between the 13 models (r = 0.60–0.67). The difference field results show that GCM-simulated SWEmax values are typically 40–80 mm higher over mid-to-high latitudes than those estimated from the CMC daily snow depth analyses, although this may be exaggerated, as the CMC SWE climatology appears to underestimate high-latitude SWE when compared to recent field observations (C. Derksen 2007, personal communication). This underestimate is likely related to the fact that the snow depth observations used in the CMC analysis tend to be taken in open locations that are more susceptible to wind scour and earlier snowmelt.

Apart from Frei et al. (2005), who reported positive SWE biases over northern Canada for AMIP2 AGCMs, previous evaluations of snow mass in GCMs (e.g., Roesch 2006) have tended to focus on continental-scale-averaged SWE, where spatial averaging reduces the importance of high-latitude positive SWE biases. For example, in Fig. 5, the mean SWE bias is only +15 mm when averaged over the NH snow-covered area. The high-latitude positive SWE bias is also linked in part to a tendency for the models' simulated precipitation to be about 15–20 cm yr^{-1} more than satellite-derived observations over land areas north of ~40°N (see Fig. S8.10 in Randall et al. 2007), together with a slight cold bias (Randall et al. 2007). Räisänen (2007) found good agreement between climate model simulations of annual maximum monthly SWE and historical snow course data from the former Soviet Union (Fig. 1d of Räisänen 2007). However, his comparison was mainly confined to the region south of 60°N, and there were few observations over northeastern Russia, where Fig. 5 suggests the models overestimate annual maximum SWE.

4. Results

a. Sensitivity analysis

Before proceeding with the presentation of the sensitivity results from the snowpack model, it is important to stress that the results should be interpreted in a relative sense because of the numerous site-related processes affecting snow cover that are not incorporated in the model (e.g., blowing snow, vegetation interactions, among others). This is also the case for the elevation change results that depend on local topography and local lapse rates, which can differ markedly from the assumed mean value of −6.5°C km^{-1} used here. For example, Marshall et al. (2007) report mean daily lapse rates of −4.1°C km^{-1} from Ellesmere Island in the Canadian Arctic. It should also be noted that investigation of the influence of elevation is carried out with respect to the local elevation of the station and is written as ΔZ; thus, ΔZ = 2000 m refers to the estimated snow climate 2000 m above the surface where the data were collected. The purpose of this analysis is to determine how snow cover may vary with elevation in the four snow-climate regions represented by each station, some of which include terrain at altitudes 1000–2000 m higher than the station.

The t statistics for simulated snow cover change in response to joint linear increases in air temperature and precipitation of +4°C and +30%, respectively, by small proportional increments, are plotted in Fig. 6 for $\Delta Z = 0$. The results show that snow cover responds very rapidly to warming, with locally significant reductions in snow cover duration emerging at TL, GB, and RB for an annual warming of 0.6°–0.8°C. In general, snow cover

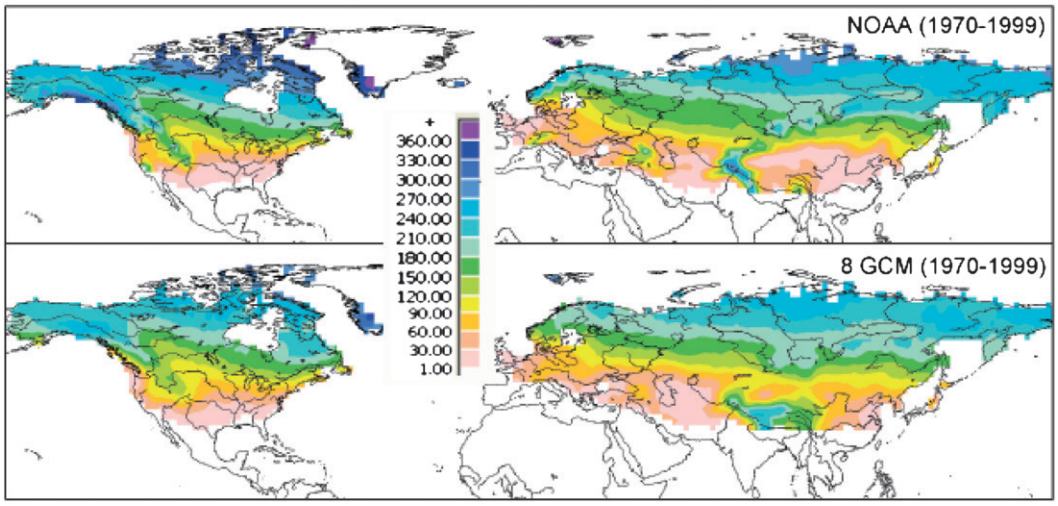


FIG. 3. Comparison of mean annual SCD (days) from the (top) NOAA weekly dataset and (bottom) eight GCM average.

duration variables, such as annual snow cover duration, showed greater sensitivity to warming than variables related to depth, such as SWEmax or JDMax. Averaged across all four sites, SCDann gave the largest change signal followed by JDLAST, SCD1, and SCD2. One of the reasons for this is that these snow cover variables tend to be less noisy than SWEmax, as evident from the CV for observed annual snow cover variables (Fig. 2).

TL exhibited the strongest temperature sensitivity of the four sites, with all nine snow cover variables displaying significant changes for a warming of only 1.5°C. The stronger sensitivity of snow cover and SWE to warming is related to the relatively warm, wet winter climate at TL (see Table 1), which means that small changes in temperature generate large changes in the fraction of precipitation falling as snow and the number of snowfall days. Indeed, for similar climates of western North America south of TL, Knowles et al. (2006) found large changes in the fraction of precipitation falling as snow, and Mote (2006) found a strong statistical connection between temperature variability and springtime SWE variability. SK displayed the lowest temperature sensitivity of the four sites, which seems counterintuitive, since one would expect shallow prairie snowpack to be sensitive to warming at the start and end of the snow season. However, this site tends to have a noisier

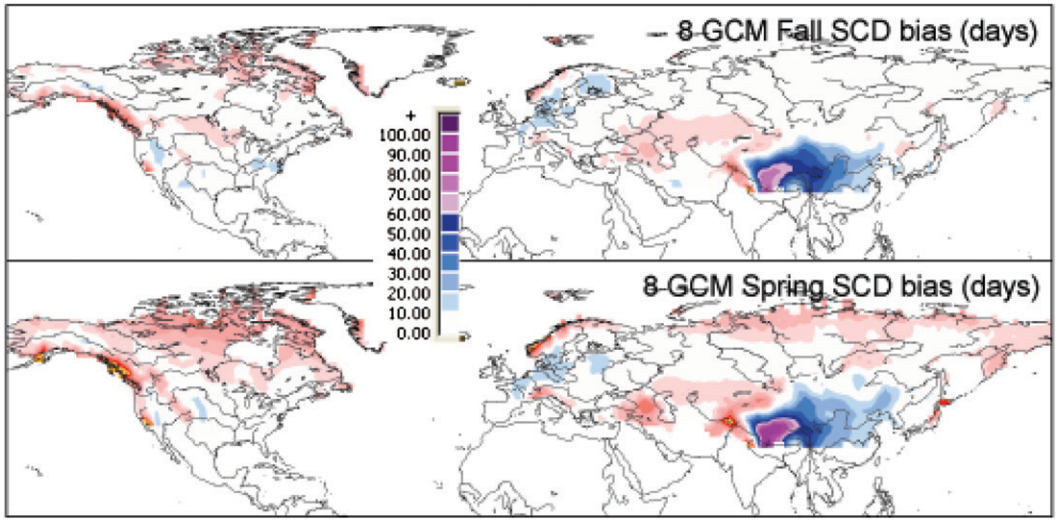


FIG. 4. Mean GCM bias in (top) fall and (bottom) spring SCD (days) compared to the NOAA weekly dataset, 1970–1999.

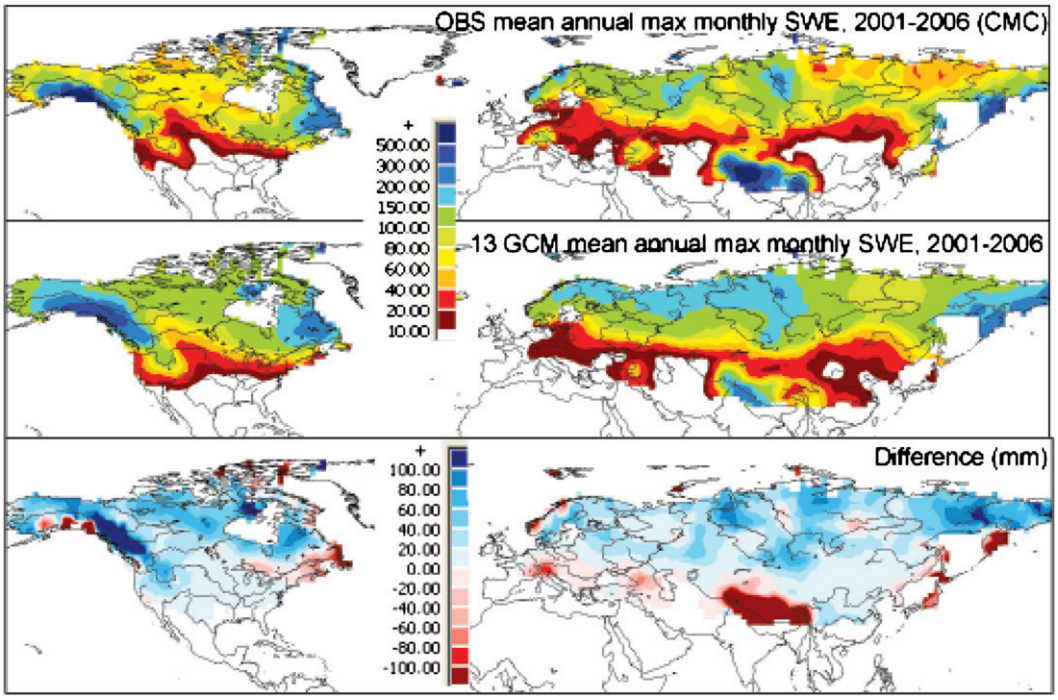


FIG. 5. Comparison of (middle) 13 GCM mean SWEmax (mm), with (top) estimated SWEmax from CMC daily snow depth analyses for 2001–06 and (bottom) mean model bias (mm).

snow cover climate than the other sites (see Fig. 2) and the temperature change signal is small, because the transition period of near-freezing temperature is half as long as TL and winter precipitation is an order of magnitude smaller (Table 1).

Comparison of sensitivity results with P varied and T held constant (not shown) indicated that increasing precipitation has only a small effect in delaying the snow cover change signal at all four stations, but a strong influence in delaying the onset of significant reductions in SWEmax, particularly at the two lower precipitation sites (RB and SK). The stronger sensitivity of SWEmax to precipitation changes in cold, dry environments may complicate the interpretation of snow trends in these climate regions. This appears to be the case in western China, where there is a wide range in reported snow cover trends (e.g., Qin et al. 2006).

An interesting result of the sensitivity analysis was the identification of significant shifts to earlier dates of maximum SWE (JDmax) *before* significant reductions in SWEmax emerged at three of the four sites. This result is partly due to the underestimation of interannual variability in JDmax by the snow model (Fig. 2). However, the observations also indicate there is less noise in the date of maximum accumulation than annual maximum depth, which suggests it may be a useful variable for climate monitoring.

The elevation response of snow cover to the warmer, wetter scenario evaluated in this study is shown in Fig. 7. Increases in SWEmax are only simulated over higher elevations where the warming influence is offset (Fig. 7a) with the height of the zero-change line ranging from $\Delta Z = 750$ m at SK to $\Delta Z = 1100$ m at TL. Increasing SWE at higher elevations has been documented over the western United States (Mote et al. 2005; Regonda et al. 2005) and the mountains of Slovakia (Vojtek et al. 2003). The elevation response of SWEmax involves nonlinear interactions between changing snow season length and changing snowfall amount and frequency. This nonlinear response can be expected to contribute to regional-scale variability in the elevation response of snow cover to climate change, which will be modified by local factors, such as lapse rate, aspect, and topography, among others. For example, Vojtek et al. (2003) estimated a zero-change elevation of 1800 m on north-facing slopes and 2300 m on south-facing slopes for snow cover change in Slovakian mountains during the 1920–2000 period.

TL exhibited the strongest elevation dependence and the strongest temperature sensitivity, which reached maximum values for SWEmax of $-150\,\text{mm}\,^{\circ}\text{C}^{-1}$ at $\Delta Z = 500$ m and SCDann of -25 days $^{\circ}\text{C}^{-1}$ at $\Delta Z = 250$ m (Fig. 7b). This is an order of magnitude higher than GB and almost two orders of magnitude higher than SK or RB, but it is

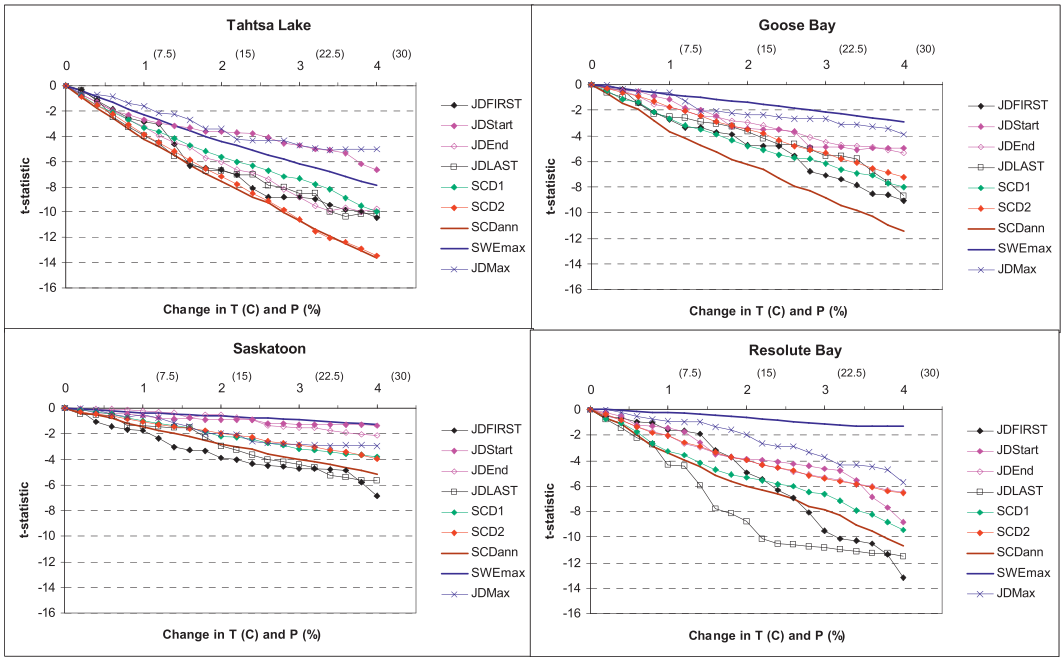


FIG. 6. Results of simulated snow cover response (t statistic) to concurrent linear increases in temperature (+4°C) and precipitation (+30%): (top left) Tahsta Lake and (top right) Goose Bay. (bottom left) Saskatoon and (bottom right) Resolute Bay.

comparable to the maximum temperature sensitivity of 1 April SWE (-200 mm °C^{-1}) observed over lower elevations of the southern Sierra Nevada by Howat and Tulaczyk (2005). The zone of elevated SWE temperature sensitivity is caused by the combined effects of changes in snowfall amount and changes in the length of the snow cover season. At low elevations, the maximum SWE is limited by the shorter snow season, while at higher elevations the snow cover is continuous, so SWEmax can only respond to changes in snowfall amount. In the intermediate zone, SWEmax is influenced by both changing season length and the changes in snowfall amount. These results are consistent with snow cover duration temperature sensitivity results from the European Alps (Hantel et al. 2000; Hantel and Hirtl-Wielke 2007) and Scotland (Mandar et al. 2007), which show a zone of maximum SCD temperature sensitivity of approximately -30 days °C^{-1} in a midelevation band between ~500 and 900 m. Figure 7 also suggests that warming and increasing precipitation may be associated with trends of opposite sign in SCD and SWEmax at some mid-elevation zone $\sim\Delta Z$ = 1000–1500 m, which may complicate the interpretation of the response of snow cover to warming in mountain regions. Analysis of published snow trends with respect to temperature as a proxy for elevation (Fig. 8) confirms the sensitivity study findings of stronger decreasing trends at lower elevations and in regions with more temperate winter climates.

In summary, the sensitivity analysis has provided the following guidance on possible twentieth-century changes in snow cover in response to a warming and a concomitant increase in precipitation:

- The snow variables likely to provide the largest climate change response across a range of snow-climate regimes are snow cover duration (SCDann, SCD1, and SCD2) and the first (JDfirst) and last (JDlast) dates of snow on the ground.
- The first evidence of a snow warming signal is most likely to be found over lower elevations of maritime and mountain snow-climate regions, that is, regions of high SWE with mild winter temperatures and in the *tundra* snow-climate region. In the latter region, the strongest signal is associated with the last date of snow on the ground, while in the former, annual snow cover duration has the strongest signal. These changes will be reinforced by albedo feedbacks in the spring period.
- The elevation response of snow cover and SWE varies with climate region, involves nonlinear interactions between snow cover duration and accumulated snowfall, and also depends on local factors, such as lapse rates, topography, and vegetation cover; these influences will complicate the interpretation of snow cover changes, particularly in mountain regions. The strongest elevation sensitivity of snow cover to

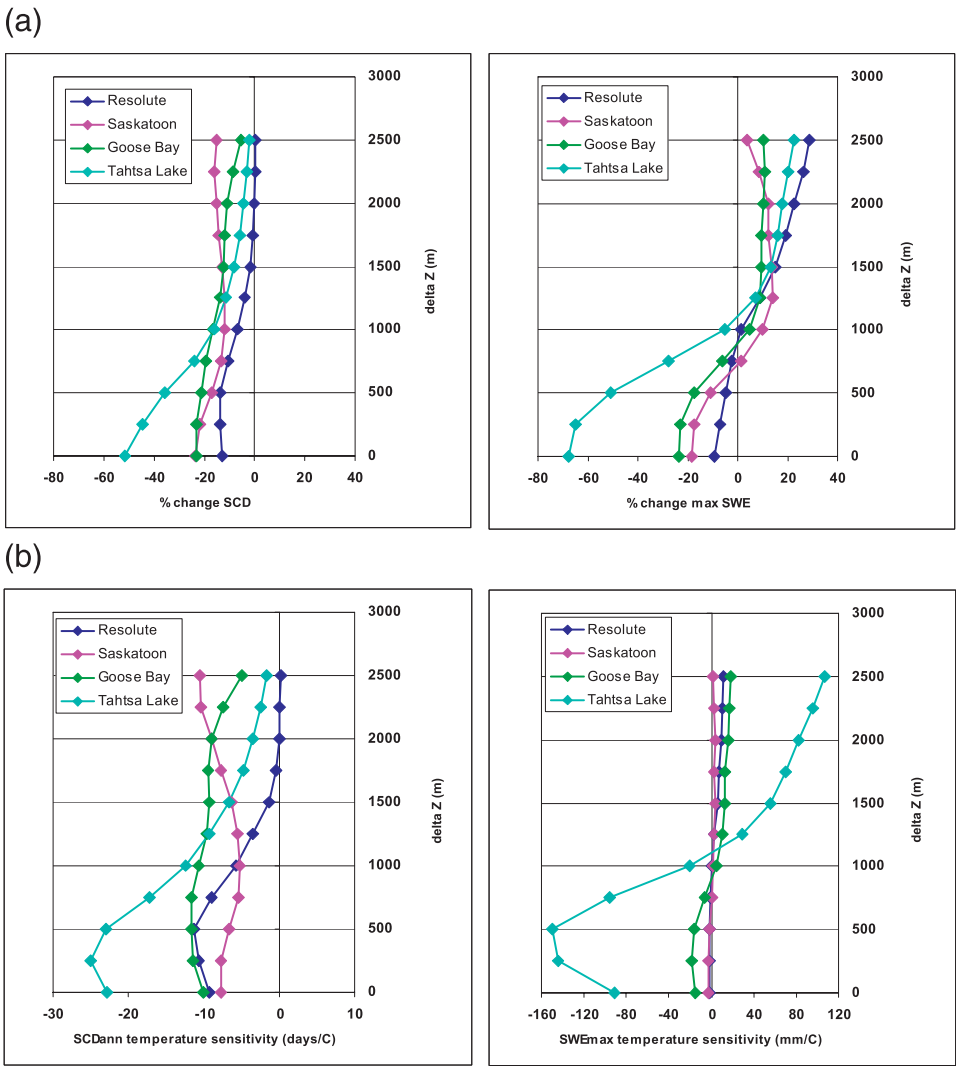


FIG. 7. (top left) Simulated change in SCDann and (top right) SWEmax at the four sites in response to a warming of 4°C and an increase in precipitation of 30%. (bottom left) SCDann and (bottom right) SWEmax temperature sensitivity.

climate change is most likely to be found in maritime climate regions.

- Significant reductions in SWEmax are most likely to be observed in maritime snow-climate regions and at lower elevations; significant increases in SWEmax are only likely to be observed at higher elevations where increases in SWEmax are not offset by decreases in snow season length.
- In mountainous regions, the response of SWEmax and SCD to warming temperature and increasing precipitation may fall into one of three regimes—(i) −SWEmax, −SCD; (ii) +SWEmax, −SCD; or (iii) +SWEmax, +SCD—depending on elevation, which can complicate the interpretation of snow cover trends.
- Increasing precipitation has a relatively small effect on the warming response of snow cover duration variables, but it does play an important role in offsetting reductions in SWEmax.
- JDmax may be a useful variable for climate monitoring, as it has less noise than SWEmax.

b. Observed snow cover response to NH warming in the NOAA dataset

The results of trend analysis for observed fall, spring, and annual snow cover duration during the 1966–2007 period are presented in Fig. 9. The most obvious feature is the pronounced seasonal contrast with fall SCD, characterized by increases over much of Eurasia and the

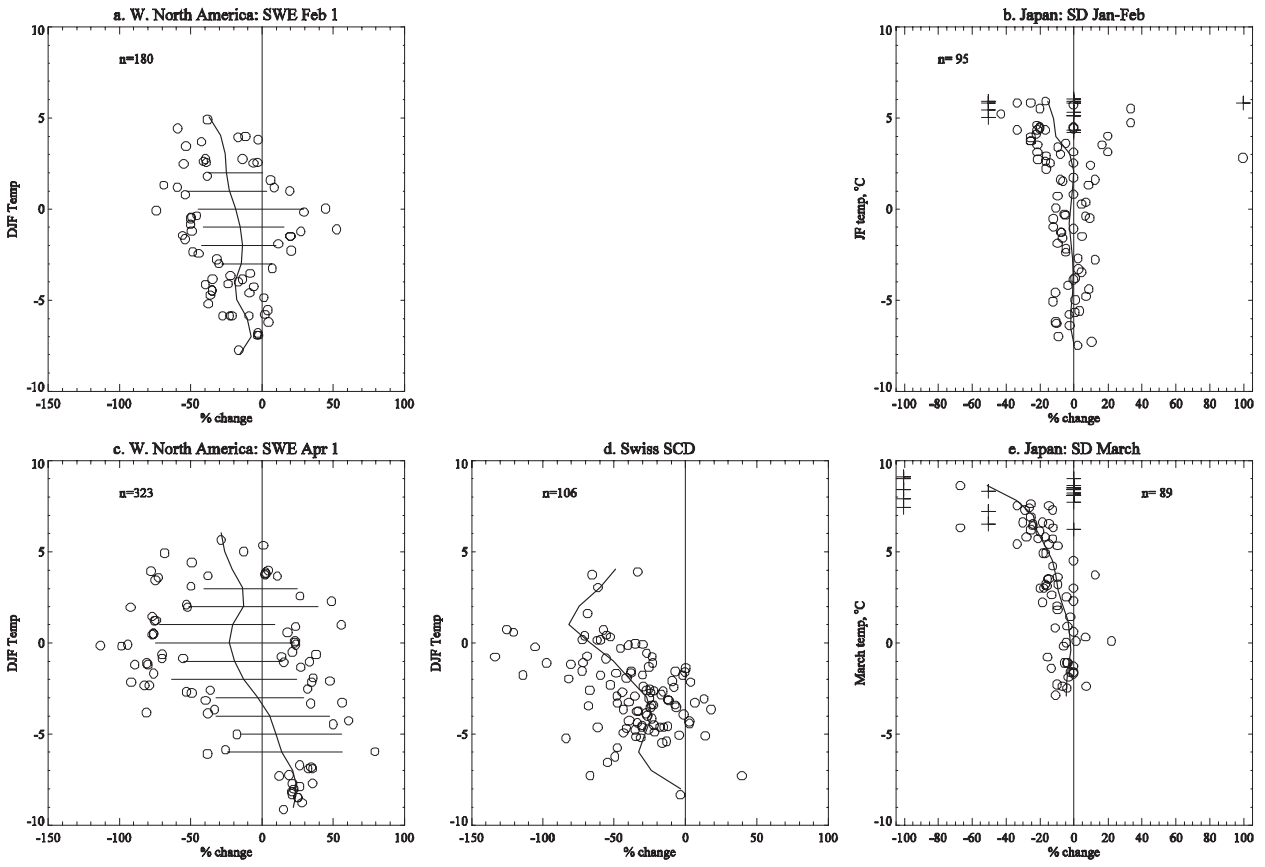


FIG. 8. Dependence of trends in measures of snow on site temperature. The number of data points is given in each panel. (a) Trends in 1 Feb SWE at mountain sites in western NA for 1950–2000. Smooth curve denotes the mean for each 1°C bin, and horizontal lines indicate the 5%–95% range for each bin. Circles show individual snow courses with trends outside that range. (b) Differences in average January–February SD at weather stations in Japan, 1960s minus 1990s. Sites with mean SD more than 2 cm are shown as circles and are used in computing the smooth curves shown; other sites are shown with pluses. (c) Same as (a), but for April. (d) Trends in snow cover duration (December–February) in Switzerland during the period of record 1958–99. (e) Same as (b), but for March. (a),(b) Show quantities and times of year that are less sensitive to temperature than those in (c)–(e).

interior of NA, while spring SCD is characterized by decreases over most NH land areas. Analysis of temperature trends in the CRUtem3v dataset and the National Centers for Environmental Prediction (NCEP) reanalysis (Kalnay et al. 1996) during the same period (not shown) indicate this different response is related to a slight cooling of temperatures in October–November over Alaska and central Canada, eastern Europe, and northern Siberia. In contrast, spring is characterized by a much stronger and widespread warming. The snow model sensitivity analysis for temperature decreases (not shown) indicated that a cooling on the order of 0.5°–1.0°C was sufficient to generate significant earlier onset of continuous snow cover (JDstart) at three of the climate stations (TL, RB, and GB). This effect is enhanced under increasing precipitation—for example, a cooling of only −0.3°C is required to generate a significant earlier onset to the continuous snow cover season at TL, with an increase in precipitation of 10%. We argue, therefore, that increasing precipitation (Zhang et al. 2007; Min et al. 2008) and the different temperature trends between early winter and spring periods are responsible for the asymmetric snow cover response observed in Fig. 9. This is consistent with observed significant increases in early winter SCE and SWE over the midlatitudes of North America during much of the twentieth century (Brown 2000).

The strongest decreases in spring snow cover extent are observed around the coastal margins of NA, Scandinavia, northern Russia, and over the Himalayas. This follows the findings of the sensitivity analysis results, which indicated that snow cover in areas with larger precipitation amounts were likely to be more sensitive to warming. To investigate this further, an analysis was made of ''at risk'' snow cover following Nolin and Daly (2006) by constructing a map of snow cover temperature

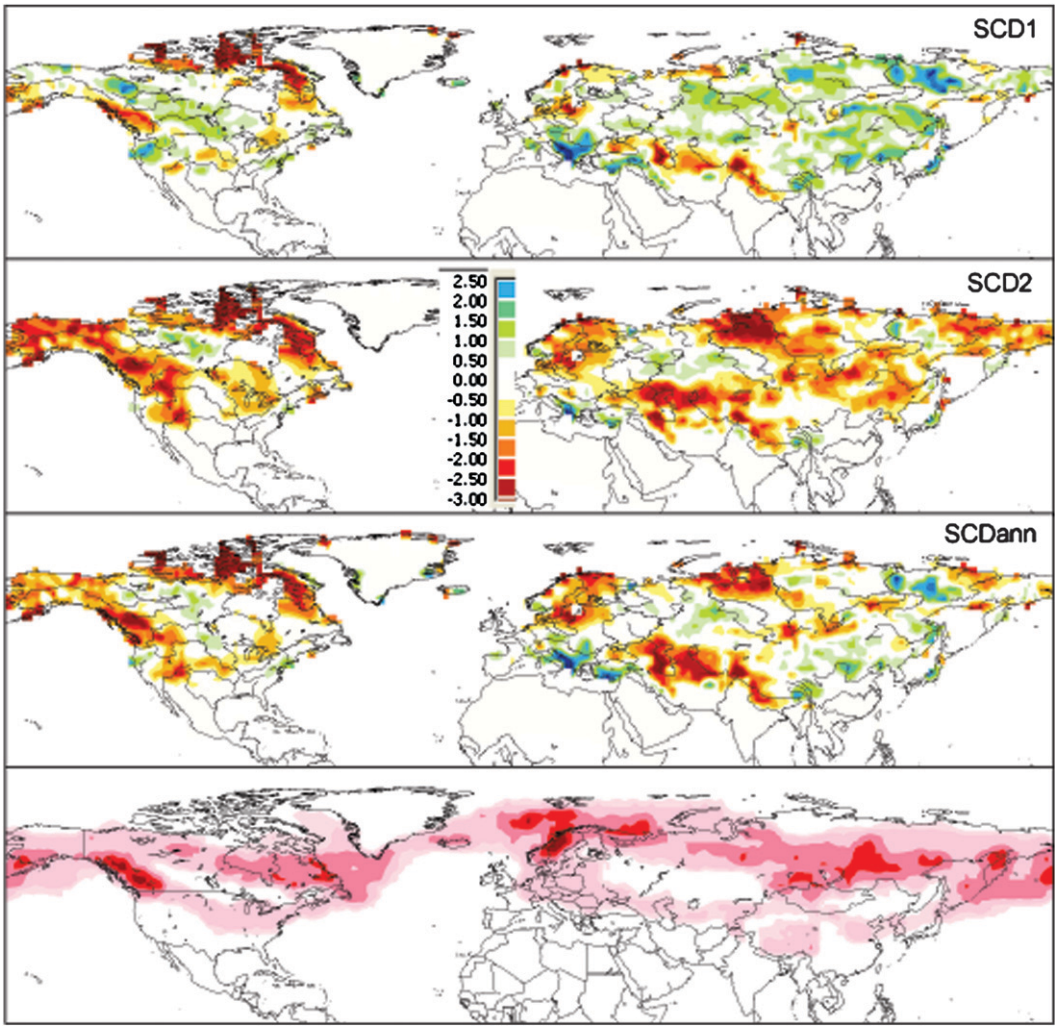


FIG. 9. (top three panels) Local Z statistic for trend in snow cover duration during 1966–2007 from the NOAA weekly snow cover dataset for the fall, spring, and snow year. (bottom) Location of the zone where snow cover should be most sensitive to temperature based on temperature and precipitation climatology from the NCEP reanalysis (the darker the red, the longer the period of snow sensitivity).

change potential from mean monthly surface air temperature and precipitation rate in the NCEP reanalysis for the 1961–90 period. The classification scheme assigned a 0–3 sensitivity ranking based on an air temperature range of +5° to −5°C and mean precipitation amounts: 0—seasonal mean air temperatures warmer than +5° or colder than −5°C; 1—seasonal mean air temperatures in the range +5° to −5°C and the mean precipitation rate of <1 mm day^{-1}; 2—seasonal mean temperatures in the range +5° to −5°C and mean precipitation rate of 1–2 mm day^{-1}; and 3—seasonal mean temperature in the range +5°C to −5°C and mean precipitation rates >2 mm day^{-1}. The temperature range was selected based on analysis of observed snow cover changes in the NOAA dataset and corresponding mean temperatures from the NCEP reanalysis, while the precipitation thresholds are based on the observed range of mean precipitation values in the NCEP reanalysis.

The classification was carried out separately for the fall [September–November (SON)], winter [December–February (DJF)] and spring [April–June (AMJ)] periods, and the seasonal ranks added to provide an index of the amount of time a potential snow cover was in the temperature sensitive regime (Fig. 9, bottom). This temperature- and precipitation-based classification shows good qualitative agreement with the spatial pattern of observed change and correctly replicates the areas showing little change over the interior of both continents. This is consistent with the sensitivity analysis,

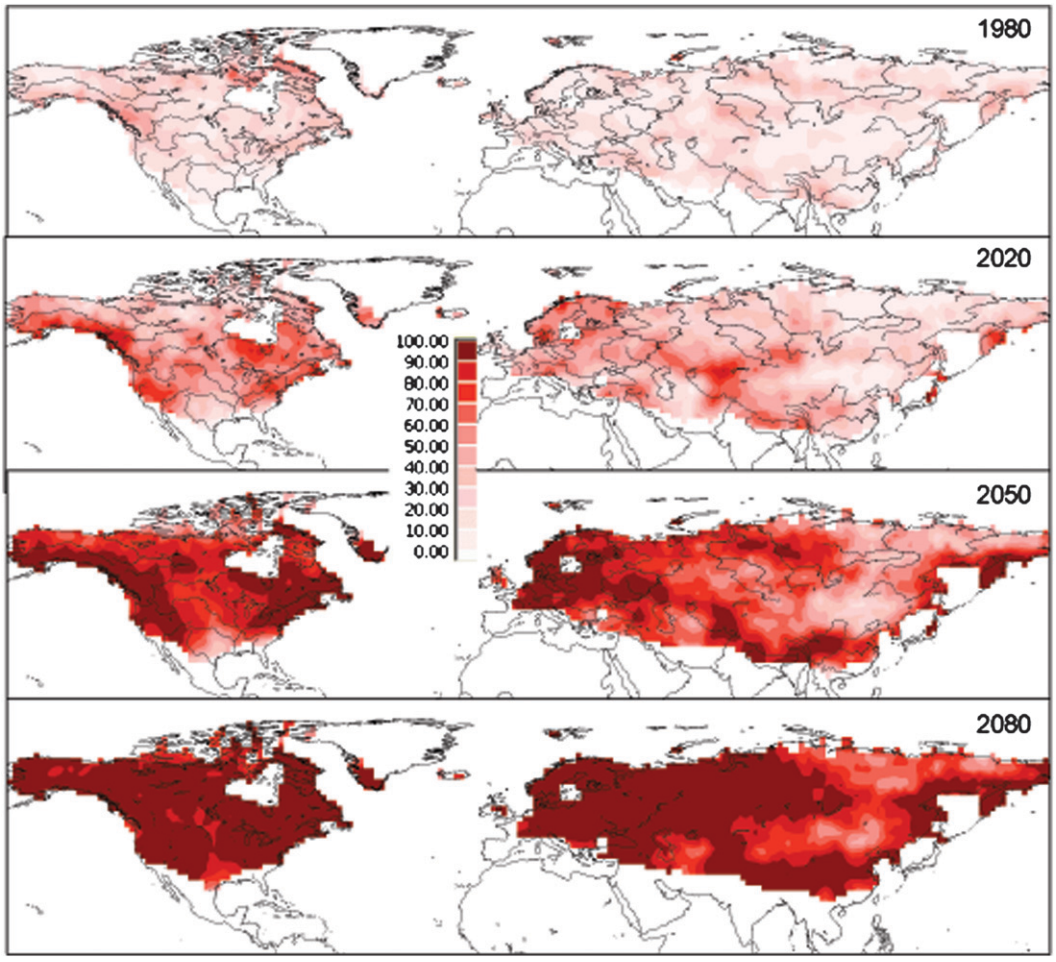


FIG. 10. Evolution of the climate warming signal in annual SCD from eight GCMs (% of models showing significant decreases). (top) Results are for 1970–99 vs 1900–29 from 20C3M simulations. (lower three panels) Results are computed from the A2 scenario with respect to a 1970–99 reference: 2020–2080. There is no model consensus for significant increases in SCDann.

which suggested the prairie snow cover climate regime has a much lower signal-to-noise ratio than other snow climate regions.

c. *Simulated snow cover response to climate change in climate models*

The time sequence of changes in SCDann and SWEmax simulated by the CMIP3 models is shown in Figs. 10–12 as the percentage of models showing a statistically significant change. For SCDann, model consensus results are only presented for *decreasing* snow cover duration, as there was no model consensus for *increasing* snow cover. Hardly any models show significant decreases in SCDann (Fig. 10) during the twentieth century, except over the eastern Canadian Arctic and in British Columbia, but by 2020 a majority of models show significant decreases in snow cover over NA in two broad zones located over the western cordillera and east of ~90°W. The 2020 model consensus is more regional in character over Eurasia, with the main regions of consensus located over Scandinavia and Kazakhstan. By 2050 all models are showing significant decreases in SCD over the west and east coast regions of NA, while in Eurasia all models are indicating significant decreases over Europe and around the southern and eastern edges of the continental snow cover. The other feature that emerges in 2050 is the large zone of lower model consensus for significant SCD decreases over much of eastern Eurasia (east of ~90°E), which persists into 2080. In this region, the models show either no consensus (particular to grid points located over the eastern Tibetan Plateau) or a consensus for no significant change in SCD. The former may reflect difficulties representing complex terrain in GCMs (Cui et al. 2007),

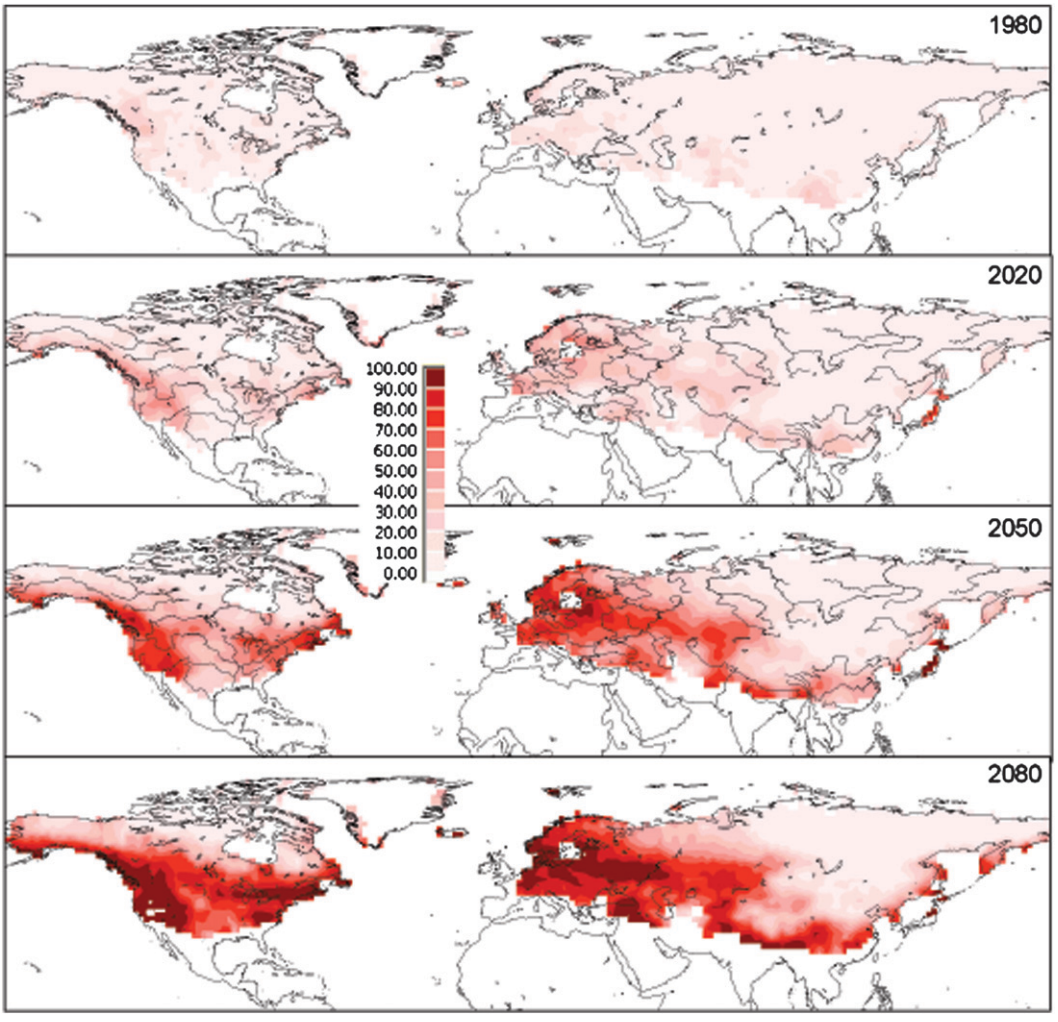


FIG. 11. Model consensus for significant decreases in SWEmax (% of 14 models). See Fig. 10 for definition of panels.

while the latter is likely linked to a slower snow cover change signal for dry continental locations, as seen in Saskatoon in Fig. 6.

The results for significant reductions in SWEmax (Fig. 11) do not show a clear model consensus until 2050, and the spatial domain of areas experiencing significant decreases in SWEmax is more constrained than SCD, with decreases mainly confined to midlatitudinal coastal regions of NA and western Europe. Over northern Canada, eastern Eurasia, and northern Russia, there is no model consensus for decreasing SWE. as these are regions where the model consensus is for no significant change or significant increases in SWE (Fig. 12). The areas of model consensus for significant decreases and increases in SWE correspond approximately to the 0° and −20°C isotherms for November–May mean air temperature (Fig. 13) following Räisänen (2007). Between these two isotherms are large regions (e.g., northern China and the boreal forest zone over NA) where the models show a consensus for no significant changes in SWE. This is consistent with the incremental analog results in Fig. 6, which showed that SWEmax did not exhibit significant change at two of the sites (SK and RB) under a scenario of T +4°C and P +30%.

There is some empirical evidence supporting increasing SWE over northern Eurasia—for example, Hyvärinen (2003) found increases of snow depth in eastern and northern Finland but decreases in the south and west. Ye et al. (1998) reported increased snow accumulation over Siberia in the same region where the CMIP3 models project increases in SWE. This was confirmed more recently by Kitaev et al. (2005), who document a general increase in snow depth and snow cover from daily snow depth observations over northern Eurasia from 1936–2000, although their snow

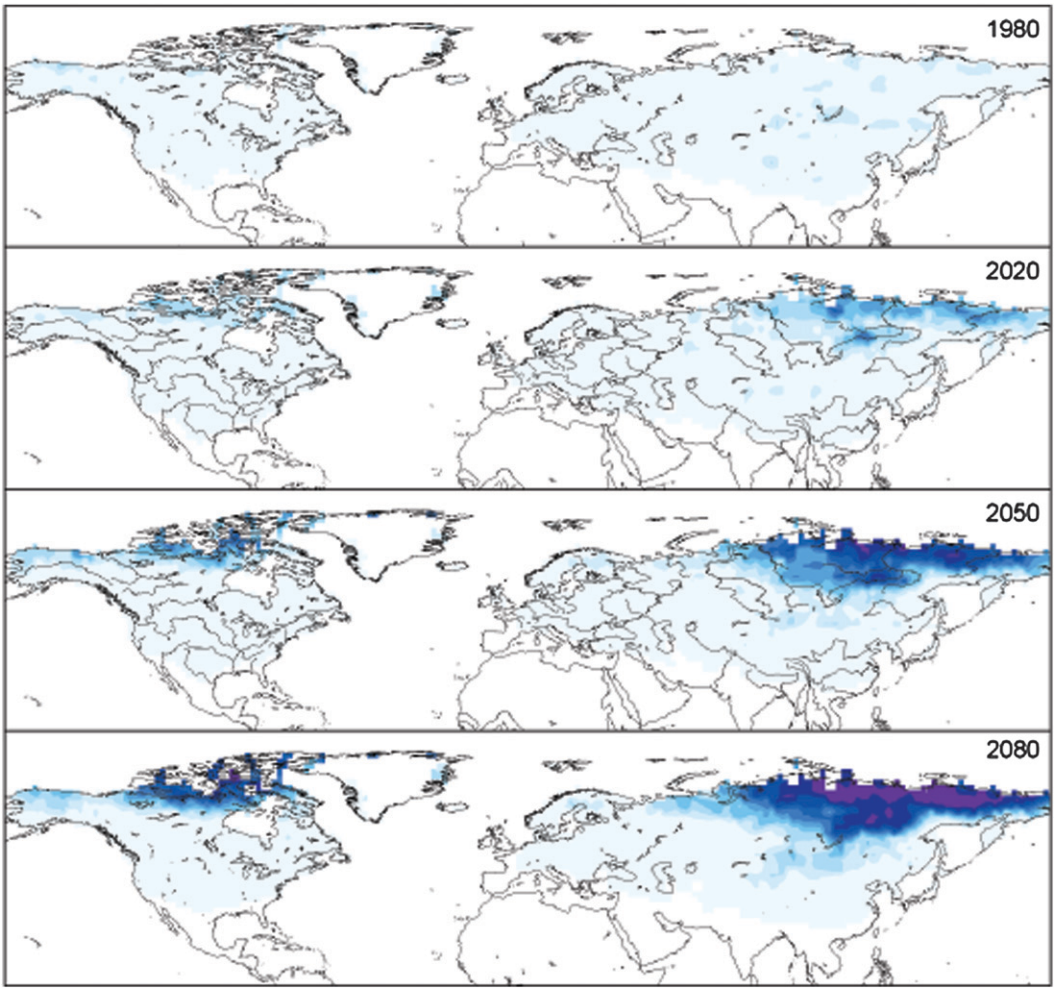


FIG. 12. Same as Fig. 11, but for significant increases in SWEmax (percent of 14 models showing significant increases).

cover duration time series suggest values peaked around the end of the 1970s then started a downward trend similar to twentieth-century trends in NA winter SCE documented by Brown (2000). This is consistent with the results of the sensitivity analysis, which showed that increasing SWE and decreasing SCD can exist for certain combinations of increasing temperature and precipitation (Fig. 7). The CMIP3 GCMs do not have sufficient resolution to capture the increase in SWE at higher elevations suggested by the sensitivity analysis, but this is apparent in dynamically downscaled SWE change scenarios (Fig. 14), which show increased SWE at higher elevations in the Rocky Mountains.

Overall, the model results are consistent with several aspects of the previous analyses. That is, they show an enhanced early response of snow cover to warming over the western cordillera of NA and maritime regions, such as eastern NA, Scandinavia, and the Pacific coast of Russia, and they indicate that snow cover changes are slower over continental interiors. The models are also consistent with the sensitivity analysis, in that they show an earlier consensus for significant reductions in snow cover before SWE (cf. Figs. 10 and 11). The areas where the models results are not consistent with observed trends and the sensitivity analysis are the lack of a consistent warming response over eastern Eurasia; the lack of clear evidence of earlier reductions in spring SCD over high latitudes, suggested from observed trends in SCD from the NOAA dataset and emerging evidence of accelerating reductions in snow cover over the Arctic spring–summer period (Déry and Brown 2007); and a circumpolar increase in maximum snow accumulation, which is not consistent with widespread reductions in winter snow depths over Canada since ~1950 (Brown and Braaten 1998; Kitaev et al. 2005; Derksen et al. 2007a). The first problem may be linked to difficulties simulating the climate interactions of the Tibetan Plateau in GCMs, and the second problem may be linked to

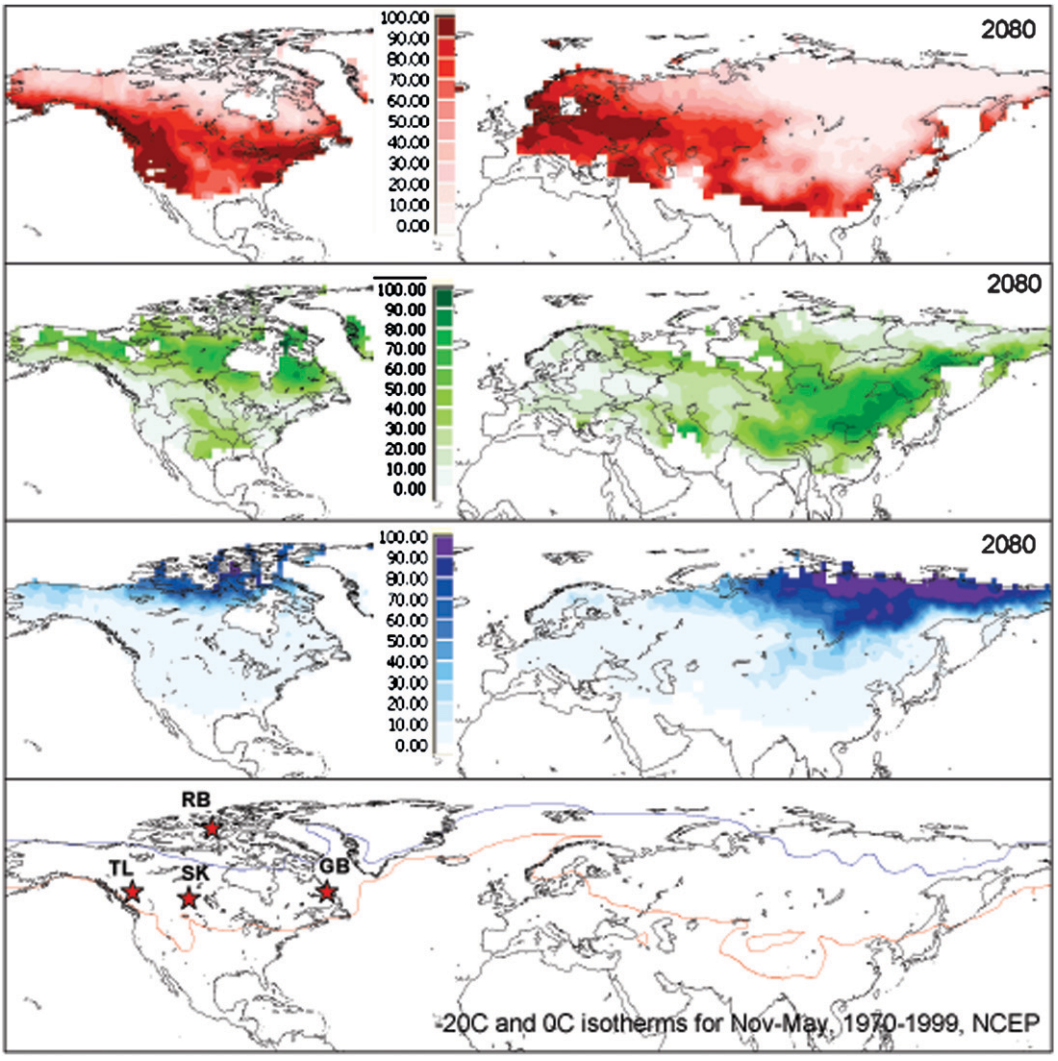


FIG. 13. Percent of 14 models showing (top to next to bottom panel) significant decreases, no increases, and significant increases in SWEmax between 2070–99 and 1970–99 compared to (bottom) the mean location of the November–May −20° (blue) and 0°C (red) isotherms; stars indicate the approximate locations of the four sites used in the sensitivity analysis.

inadequate treatment of snow processes. For example, Qu and Hall (2007) reported a large intermodel spread in snow–albedo feedback related to the treatment of vegetation.

5. Conclusions

The results of the analyses carried out in this paper suggest a complex response of NH snow cover to a climate change of increasing T and P, with the strength (and signal-to-noise ratio) varying with climate regime, elevation, and snow cover variable. Snow cover duration was shown to have the strongest sensitivity to warming as measured by signal-to-noise ratio, with the largest relative changes in SCD and SWE over lower elevations of regions with a maritime winter climate—that is, moist climates with snow season temperatures in the range of −5° to +5°C. This corresponds to a midlatitudinal zone extending around the coastal margins of the continents, with snow cover in the drier, colder continental interior regions showing the lowest temperature sensitivity. Positive albedo feedbacks in the spring period generate a stronger signal in spring snow cover duration, with feedback potential increasing with latitude (Déry and Brown 2007).

The snow model sensitivity analysis suggested a potentially complex elevation response of SCD and SWEmax to increasing T and P in mountain regions as a result of nonlinear interactions between the duration of the snow season and snow accumulation rates. Joint

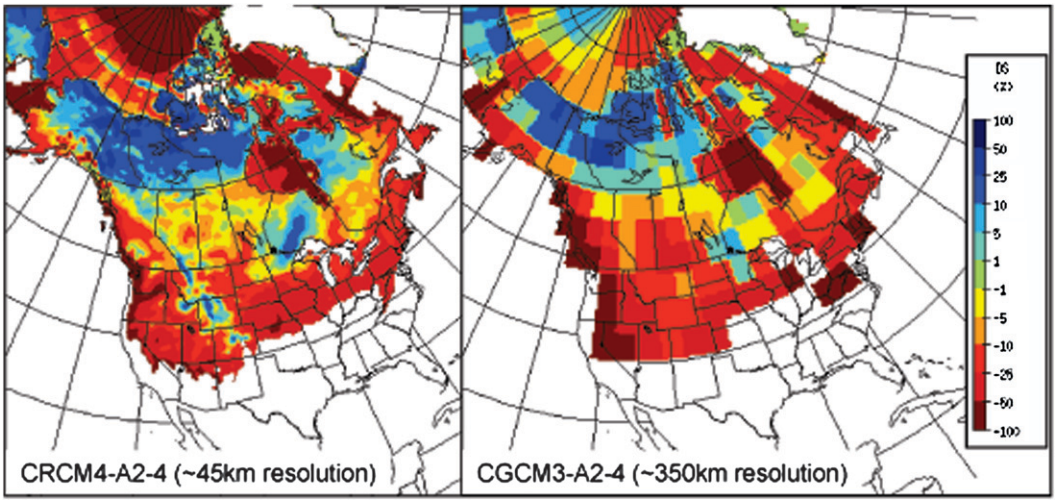


FIG. 14. Percent change in mean winter SWE between 1961–90 and 2041–70 for the Special Report on Emission Scenarios (SRES) A2 scenario. (left) Results from version 4.1.1 of the Canadian Regional Climate Model simulation (Music and Caya 2007) driven at its lateral boundaries by CGCM3 SRES A2 scenario, run4. (right) SWE change simulated by CGCM3 (source: Ouranos Climate Simulation Team).

response surface results of relative changes in SCD and SWEmax indicate three possible snow cover responses depending on the amount of warming and precipitation increase, the elevation, and the climate regime. Drier climate regimes were more likely to have a mixed response of increasing SWEmax and decreasing SCD, which may explain some of the variability in snow trends reported from western China. Local factors—such as lapse rate, aspect, topography, vegetation, and blowing snow transport—will further complicate the snow cover climate response in mountain regions.

Trends in snow cover duration observed from space during the 1966–2007 period are consistent with the spatial pattern of significant snow cover reduction simulated by the CMIP3 climate models, which show an early warming response of snow cover over the western cordillera of NA and maritime regions, such as eastern NA, Scandinavia, and the Pacific coast of Russia. The observed trends and the climate models also agree that significant snow cover changes are slower over continental interiors (in agreement with the sensitivity analysis). The climate models are also consistent with the sensitivity analysis and observed trends in that they show significant widespread reductions in snow cover occurring *before* significant reductions in SWEmax.

Observations of increasing snow depths over northern Eurasia are consistent with climate model projections of increased SWEmax over northern high latitudes. However, there is little observational evidence of increasing SWE from northern Canada. The exception is northern Quebec, where there is some proxy evidence of increasing snowfall from reconstructed lake levels (Bégin 2000) and analysis of black spruce growth forms (Lavoie and Payette 1992). According to Payette et al. (2004), increased snowfall since 1957 was the main climatic driver for accelerated permafrost thawing in northern Quebec between 1957 and 2003. Further work is needed to document snow accumulation trends over northern latitudes, as the surface observing networks are inadequate for monitoring regionally averaged SWE. It may be possible to develop regional SWE estimates from passive microwave data using new algorithms being developed specifically for the tundra region (Derksen et al. 2005; Rees et al. 2006; Derksen et al. 2007b). Further work is also needed to evaluate and improve the representation of snow cover in global climate models—in particular, the apparent overestimation of snow accumulation over northern high latitudes and the muted model response of snow cover duration to climate forcing over high latitudes and eastern Eurasia.

Acknowledgments. The idea for this paper was developed while the coauthor was on sabbatical at the Ouranos climate change consortium in Montréal during July 2007, and he thanks Ouranos for its partial support for the sabbatical. Anne Frigon from the Ouranos Climate Simulation Team is acknowledged for providing the regional climate model snow simulation shown in Fig. 14. We also thank S. Scherrer, M. Ishizaki, and N. Petkova for sharing the data used to construct Fig. 1, and X. Zhang

and the three anonymous referees for their helpful comments on the draft manuscript. We acknowledge the modeling groups, the Program for Climate Model Diagnosis and Intercomparison (PCMDI), and the WCRP's Working Group on Coupled Modelling (WGCM) for their roles in making available the WCRP CMIP3 multimodel dataset. Support of this dataset is provided by the Office of Science, U.S. Department of Energy. This publication is partially funded by the Joint Institute for the Study of the Atmosphere and Ocean (JISAO) under NOAA Cooperative Agreement NA17RJ1232.

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From: Holt, Bryon
To: Grizzle, Betty
Subject: Re: Help with locating references
Date: Monday, February 5, 2018 12:35:19 PM
Attachments: Cayan 1996.pdf
Hamlet et al 2005.pdf
Mote 2006.pdf

2nd email.

On Mon, Feb 5, 2018 at 11:33 AM, Holt, Bryon <bryon_holt@fws.gov> wrote:

> Hi Betty,
>
> I've got all but one document. The one document I don't yet have is by Moran-Tejeda, but I have requested it from our Conservation Library and should have it soon to send to you. I will send the ones I have now in a couple of emails to minimize size issues. Note: Brown and Mote is 2009 not 2008 - could not find 2008. May be a typo?
>
> Bryon
>
> On Wed, Jan 31, 2018 at 10:47 AM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:
>
> > Thank you! Will save me time pulling this final draft together.
> >
> > On Wed, Jan 31, 2018 at 10:44 AM, Holt, Bryon <bryon_holt@fws.gov> wrote:
> >
> > > I can help you with this.
> > >
> > > On Wed, Jan 31, 2018 at 10:26 AM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:
> > >
> > > > Hi Bryon - You offered to help with finalizing the SSA Report, so am writing to ask if you can locate and download the references highlighted in yellow on the first page of the attached publication. There are 7 total and full citations are at the end.
> > > >
> > > > Please let me know whether or not you have time to help. If not, I can ask others. Thanks.
> > > >
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> > > >
> > > > Betty J. Grizzle, D.Env., MPH
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Interannual Climate Variability and Snowpack in the Western United States

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(Manuscript received 23 January 1994, in final form 19 June 1995)

ABSTRACT

An important part of the water supply in the western United States is derived from runoff fed by mountain snowmelt. Snow accumulation responds to both precipitation and temperature variations, and forms an interesting climatic index, since it integrates these influences over the entire late fall–spring period. Here, effects of cool season climate variability upon snow water equivalent (SWE) over the western part of the conterminous United States are examined. The focus is on measurements on/around 1 April, when snow accumulation is typically greatest. The primary data, from a network of mountainous snow courses, provides a good description of interannual fluctuations in snow accumulations, since many snow courses have records of five decades or more. For any given year, the spring SWE anomaly at a particular snow course is likely to be 25%–60% of its long-term average. Five separate regions of anomalous SWE variability are distinguished, using a rotated principal components analysis. Although effects vary with region and with elevation, in general, the anomalous winter precipitation has the strongest influence on spring SWE fluctuations. Anomalous temperature has a weaker effect overall, but it has great influence in lower elevations such as in the coastal Northwest, and during spring in higher elevations. The regional snow anomaly patterns are associated with precipitation and temperature anomalies in winter and early spring. Patterns of the precipitation, temperature, and snow anomalies extend over broad regional areas, much larger than individual watersheds. These surface anomalies are organized by the atmospheric circulation, with primary anomaly centers over the North Pacific Ocean as well as over western North America. For most of the regions, anomalously low SWE is associated with a winter circulation resembling the PNA pattern. With a strong low in the central North Pacific and high pressure over the Pacific Northwest, this pattern diverts North Pacific storms northward, away from the region. Both warm and cool phases of El Niño–Southern Oscillation tend to produce regional patterns with out-of-phase SWE anomalies in the Northwest and the Southwest.

1. Introduction

Much of the water supply in the western United States is derived from snowmelt runoff. For example, it is estimated that 75% of the annual discharge of most of the major streams in the western United States is from melting mountain snowpack (Palmer 1988). Topography plays a strong role, so that the heaviest snow accumulations are at middle to high elevation mountain sites usually with exposures to the west or southwest. In the Sierra Nevada, which is the primary water-bearing region for California, the snowmelt is derived mostly from snowpack accumulated during winter storms at elevations above 1300 m. Precipitation in the region is very seasonal; most locations west of the Rocky Mountains have a winter maximum and a summer minimum. Combined with the annual cycle of temperature, this seasonality produces a spring maximum in snow accumulation, so that most snow courses attain their climatological maximum snow water equivalent (SWE) at about the beginning of April. Snow course observations are generally collected on or about the first of the month during the winter and spring. The greatest frequency of sampling over the historical period is for 1 April.

How much variability occurs in the snow accumulation? A survey of historical snow course records over the western states indicates that the coefficient of variation (CV) of 1 April SWE ranges from 20% to well over 100%. For the 200 snow courses used in this study, 45 had a CV $\geq$ 50%, whereas only 23 had a CV $\leq$ 25%. This level of variability is comparable to that of seasonal precipitation over this region (Granger 1977; Changnon et al. 1991). Snowpack accumulations are coherent over scales of at least a few hundred kilometers; in California, Aguado (1990) showed that for 28 snow courses across the Sierra Nevada Mountains, over 80% of the variance of SWE was accounted for by a single principal component.

Although mountain snow accumulations have been routinely monitored for several years to gauge water supplies, these measurements have not been fully exploited as a climate dataset. A comprehensive study of short period climate variability of snowpack in the West has not been undertaken, but there are noteworthy regional examples. In examining snow course records in the Rocky Mountains, Changnon et al. (1991) dem-

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onstrated that spring snow accumulation serves as a regional climatic indicator, and Changnon et al. (1993) derived effects of synoptic atmospheric circulation patterns and a tendency for out-of-phase multiyear shifts in snow accumulation and associated circulation patterns over the central part of the northern and southern Rocky Mountains. Concerning El Niño, Redmond and Koch (1991) demonstrated that snowfall in Oregon is significantly biased in years with a low Southern Oscillation Index (SOI) relative to those with a high SOI, and Cayan and Webb (1992) derived the pattern of SWE over the western United States in conjunction with the SOI.

It is clear that both precipitation and temperature affect the snowpack, but their aggregate impacts are not entirely clear. There does not appear to be a strong temperature–snow relation over the entire region (Walsh et al. 1982; Leathers and Robinson 1993; Karl et al. 1993). However, anomalous temperature impacts are evident in spring runoff variations in the Sierra Nevada (Aguado et al. 1992; Cayan et al. 1993). Also, winter temperature trends appear to be involved in a decades-long change in the fraction of runoff occuring in late spring and summer runoff found in the Sierra Nevada and many other snowmelt-driven streams over the western United States (Roos 1987, 1991; Wahl 1992; Aguado et al. 1992; Dettinger and Cayan 1995). Its dependence on temperature makes snow a key diagnostic in climate change scenarios (e.g., Gleick 1987; Roos 1989; Lettenmaier and Gan 1990; Jeton et al. 1995). Temperature and precipitation are not well correlated during the cool season in the West (Zhao and Khalil 1993; Cayan and Peterson 1993), so their individual influences on snowpack should be separable.

In view of the above, the purpose of this study is to examine a network of mountain snow courses over the western United States to examine the following issues associated with the snow accumulation:

1) *How does it vary spatially?*
2) *How is it linked to fluctuations in surface temperature, precipitation, and runoff?*
3) *How is it related to atmospheric circulation?*

2. Data

For several decades, the United States Department of Agriculture (USDA) Soil Conservation Service (SCS) has archived snow observations (depth and water content) at several hundred mountain snow courses in the western United States (Stafford 1959; Work and Beaumont 1959). Throughout the West, many snow courses were established in the mid-to-late 1930s. These programs were instituted in response to the early 1930s western drought, after which congress authorized the USDA to develop snow sampling equipment and conduct snow surveys for water supply forecasting (Palmer 1988). In California, a similar set of snow course samples was commenced in 1929 in major water supply regions by the California Department of Water Resources, following over two decades of pioneering efforts by James E. Church of the Cooperative Snow Survey Program (Stafford 1959). Since many rivers in the West are dominated by snowmelt runoff, the snow surveys programs collect snow data for spring and summer water supply forecasts. Snow courses are commonly sited in sheltered alpine meadows, so that drifts, blowing snow, and excessive sublimation are minimized. Snow water equivalent is measured with a Mt. Rose snow sampler, averaged over about 10 samples at each snow course (Clyde and Houston 1951). More recently, there are many automated snow sensor observations available over the western region; these records are limited to the last 10–15 years and may have errors in measuring seasonal accumulation of snow, so they are not employed in the present study. Most snow courses are monitored at the beginning (on or about the first of the month) and sometimes the middle of each of the substantial snow-covered months, usually from January or February through May or June. Here, we use the first of the month snow course samples because the midmonth samples are generally not available over the complete record. Snow courses cover a large range of elevation, from less than 1000 m to over 3000 m. Most of the lowest snow courses are in the coastal ranges of Oregon and Washington, while most of the highest ones are in the southern Sierra Nevada in California or the southern Rocky Mountains.

The 200 snow courses employed, shown by dots on Fig. 1, are from a network of more than 2000 snow courses in the 11 western states of the conterminous United States (Palmer 1988; State of California 1991). Data from Alaska was not employed in the present study, because it has relatively short records and because British Columbia sites were not available. Many snow courses in this network are sited in very close proximity, so anomalous variability at a particular site is likely redundant with that of neighboring stations (Aguado 1990). The 200 snow courses employed were selected to provide a relatively complete record (most of them span 1930–1989, and none begin later than 1940). The stations were culled to provide a relatively uniform spatial sample; the following mix of sites were included: 5 from Washington, 29 from Oregon, 21 from California, 3 from Arizona, 9 from Nevada, 25 from Utah, 32 from Idaho, 27 from Montana, 19 from Wyoming, 26 from Colorado, and 4 from New Mexico. Other studies (Aguado 1990; Changnon et al. 1991) have demonstrated that there is much regional coherence of the spring snow accumulation within networks of the snow courses in California and the Rocky Mountain states, respectively, so the results of this study are not likely to be very sensitive to this choice of stations.

Most snow courses attain their climatological maximum snow water equivalent (SWE) at about the beginning of April. Snow course observations are gen-

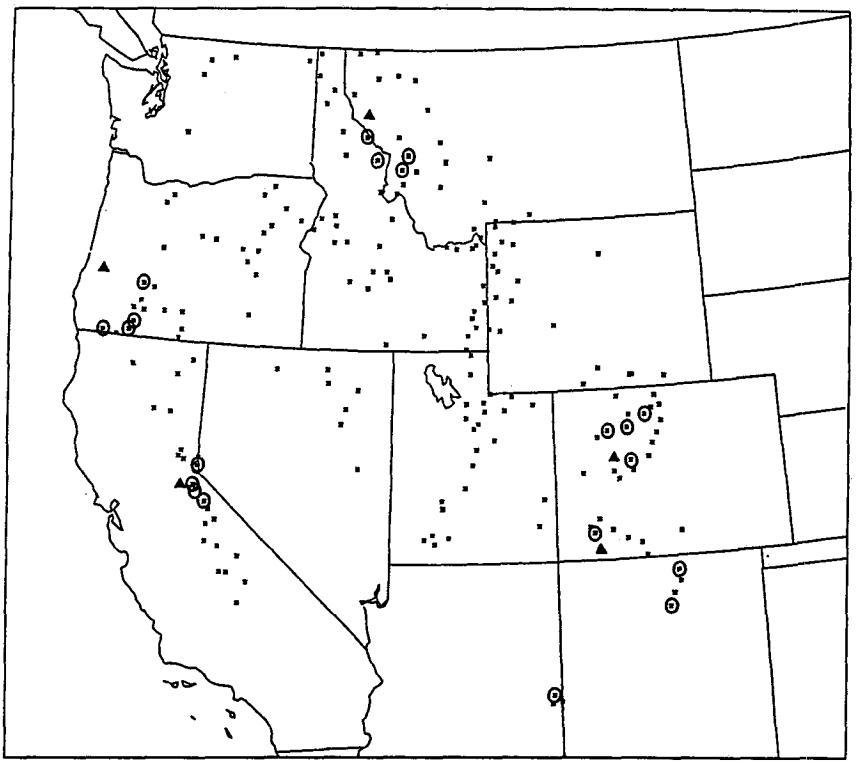


FIG. 1. Snow courses (dots and dots enclosed by circles) and stream gauges (solid triangles) employed in this study. Snow courses enclosed by circles designate those used to calculate regional averages (Table 1). Stream gauges are identified in Table 1.

erally collected on or about the first of the month during the winter and spring; the 1 April sampling tends to be the period with the greatest frequency of sampling over the historical period. However, some low elevation sites and some of the southernmost sites experience maxima as early as 1 February, whereas some of the high elevation sites have maxima delayed until May. Of the 200 snow courses employed in this study, 149 had climatological maximum first-of-the-month values on 1 April, 29 had maxima on 1 March, 19 had maxima on 1 May, and 3 had maxima on 1 February.

Observations from 1 April are the primary data used in the present study, but observations from the other months are also employed to track the seasonal evolution of the snow. Here, 1 April is the most commonly sampled period over the duration of record probably because maximum snow accumulation at most stations is reached in early April. The seasonal progression of snow accumulation at representative locations is shown later in the regional SWE composites. Snow depth is generally measured concurrently with SWE at each snow course. At the 200 snow courses included here, 1 April snow depth and SWE are highly correlated: nearly all (199) have correlation coefficients ≥ 0.80, many (160) have correlations ≥ 0.90, and several (84) have correlations ≥ 0.95.

Atmospheric circulation is represented by Northern Hemisphere 700 millibar (mb) height, which was obtained from the NOAA Climate Analysis Center. The 700-mb height data employed is monthly averages over a 10° long. × 5° lat. "diamond grid" since December 1946. The 700-mb height, approximately 3 km above sea level, avoids most topographic influences, although it is exceeded by some peaks in the southern Sierra and in the central Rockies. This pressure surface is representative of monthly seasonal tropospheric circulation governing North American weather variations (e.g., Namias 1978).

Monthly "divisional" data from the National Climatic Data Center (NCDC) are used to represent temperature and precipitation variability over the conterminous United States for the period of the early 1900s through 1989. These are area averages computed from data for several (usually 20 or more) stations over quasi-uniform climate districts from each state. Weather stations employed in the divisional averages tend to be at elevations lower than the major snow-laden regions studied here, so some mountain climate

features may not be well represented. However, temperature and precipitation anomalies at individual high elevation stations correlate well with those of the divisional data (correlations between anomalies of the middle elevation cooperative stations and those of the divisional data were mostly ≥ 0.8), so we believe the divisional data is adequate for studying the sought after climate anomalies.

3. Spatial variability of SWE

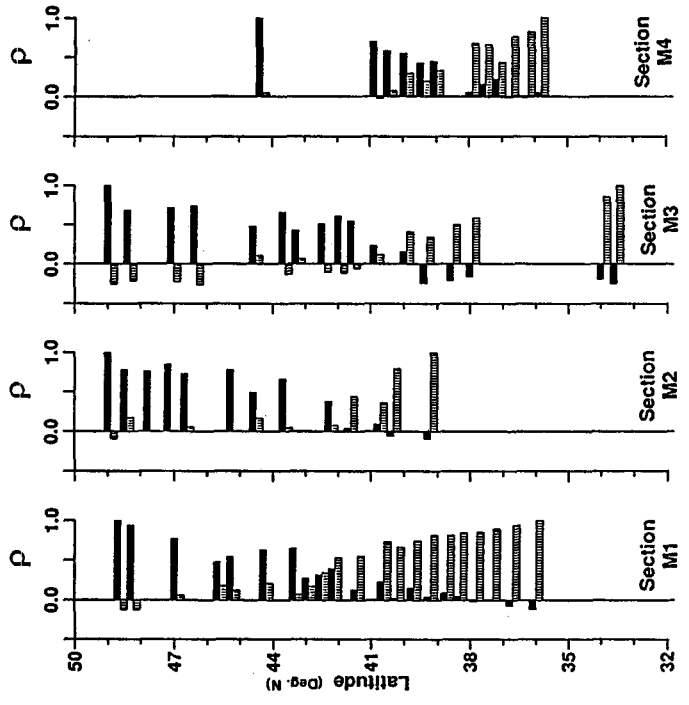
As a first look at the spatial variability of the 1 April SWE anomalies, Fig. 2 gives correlations computed between the time series at selected snow courses along several transects: four meridional transects (M1–M4) and four zonal transects (Z1–Z4). The geographic distribution of the snow courses resulted in a set of stations that were not uniformly dispersed along the transect. A few close neighboring snow courses were included to test the stability of the correlations. For each transect, two sets of correlations are shown: correlations between the snow course at each end of a transect and all of the snow courses along the transect. Because the correlations could be affected by peculiarities of the stations chosen as the end points, a separate analysis was also performed, using snow courses in the interior of the transect as the primary sites. Since the results reinforced those obtained using the stations at the end points, these plots are not shown.

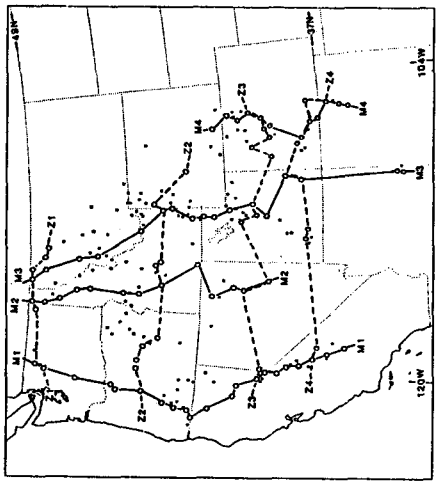
The correlations along the transects show that the SWE anomalies are coherent over distances of a few to several hundred km. The westernmost meridional transect, M1, contains relatively high correlations (≥ 0.5) from the end snow course in northern Washington as far south as a snow courses in south central Oregon, and from the end snow course in the southern Sierra Nevada of California to as far north as a snow course in southern Oregon. Considering all of the meridional transects, the snow courses are quite strongly correlated throughout the Northwest, but are not well correlated with those south of about 42°N, the northern border of California, Nevada, and Utah. The meridional SWE scales along the interior transects tend to be shorter than those along the coastal transect. There is a tendency for northern and southern snow courses to be out-of-phase, as evidenced by the negative correlations linking them (see plots of correlations along M1 and M3). Breaks in the correlations evidently delineate physiographic provinces—there is a discontinuity along the coastal domain between the Cascades and the Sierra Nevada, and there is a break in the interior ranges at the Snake River Plain between the Northern and Southern Rocky Mountains. Each of the zonal transects has significant correlations that extend well into the interior. Correlations exceed 0.5 along the west–east transects from western Washington through Montana, from western Oregon through eastern Idaho, from northern California through western Colorado, and from central California to western Utah. Except for the farthest north zonal transect, correlations beginning with the eastern end snow courses do not sustain high levels westward all the way to the West Coast. In northern Colorado, high correlations along the zonal and meridional transects are confined to a few hundred kilometer distances.

Next, to establish the full geographic patterns of SWE spatial variability, a rotated principal component analysis (RPC) was performed. As explained by Horel (1981) and Richman (1986), rotation of the principal components (PCs) often produces more physically meaningful patterns and prevents artificial mixing of modes whose eigenvalues are not well separated. Following Horel (1981) and Barnston and Livezey (1987), a varimax orthogonal rotation was employed. This rotation is a linear combination that preserves the temporal orthogonality of components, but maximizes the variance of the squared correlation between the RPCs and the original time series. This maximization tends to produce spatial loading vector patterns that are more regionally clustered than those of the original PCs. Also, the spatial loading patterns are no longer constrained to be orthogonal.

Before entry in the PC analysis, the individual 1 April SWE anomaly time series were normalized by their standard deviations. In the present case, 8 PCs were computed, accounting for 78% of the total 1 April SWE variance. Upon rotation, 5 RPCs emerged with spatially coherent, regional-scale patterns (Fig. 3). Two tests were performed to judge the sensitivity of the RPC results: 1) RPCs computed from sets of 5, 6, 10, and 12 PCs yielded little change, namely five dominant RPCs were present with similar spatial patterns to those from rotating 8 PCs; 2) when the dataset was split into two equal segments (1930–1959 and 1960–1990), five dominant RPCs again resulted, which compared closely to those from the original set. Thus, we retained our original choice of 5 RPCs derived from 8 PCs computed from the 1930–1989 dataset.

The 5 RPCs, in order of explained variance, have centers in the following regions: 1) interior Northwest (Idaho and Montana, Wyoming, northern Utah, and northern Colorado); 2) California (the Sierra Nevada and mountainous regions of southern Oregon, Idaho, and Utah; 3) the Southwest (Arizona, New Mexico, southern Colorado, and southern Utah); 4) Oregon, parts of Washington, Idaho, northern California, and northern Nevada; and 5) Colorado, Wyoming, and north central Utah. These 5 RPCs collectively explain about 70% of the variance of the normalized 1 April SWE anomaly field, and individually account for 28% to 8% of the variance. Hereafter, for brevity, these regions are called Idaho, California, New Mexico, Oregon, and Colorado. In Fig. 3, the spatial loadings of the RPCs are expressed as correlation coefficients, where the correlations relate the particular RPC time coefficients and the original SWE data. The RPCs ac-





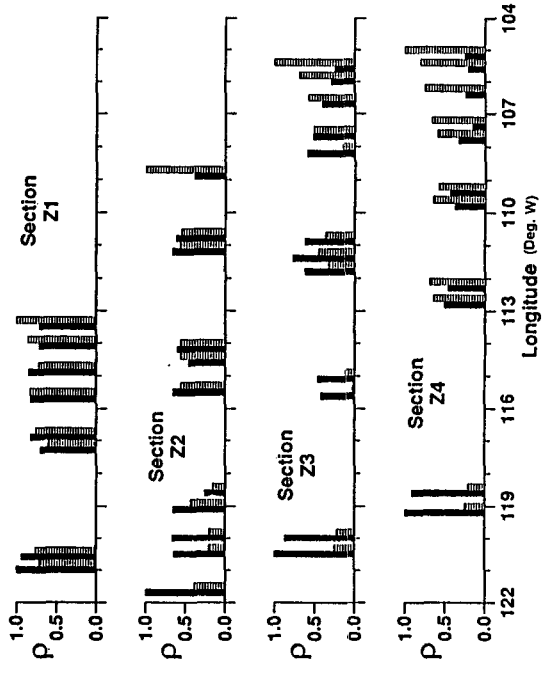


FIG. 2. Correlations among stations along meridional and zonal transects. Inset shows snow course sections (circles denote snow courses along each transect) used for spatial correlation analysis. Meridional transects denoted M1–M4, zonal transects denoted Z1–Z4. Bar graphs show correlations relating snow courses at each end of transect with all others along transect. Solid bars for correlations relating far north/west snow course with other snow courses along meridional/zonal transects. Hatched bars for correlations relating far south/east snow course with other snow courses along meridional/zonal transects.

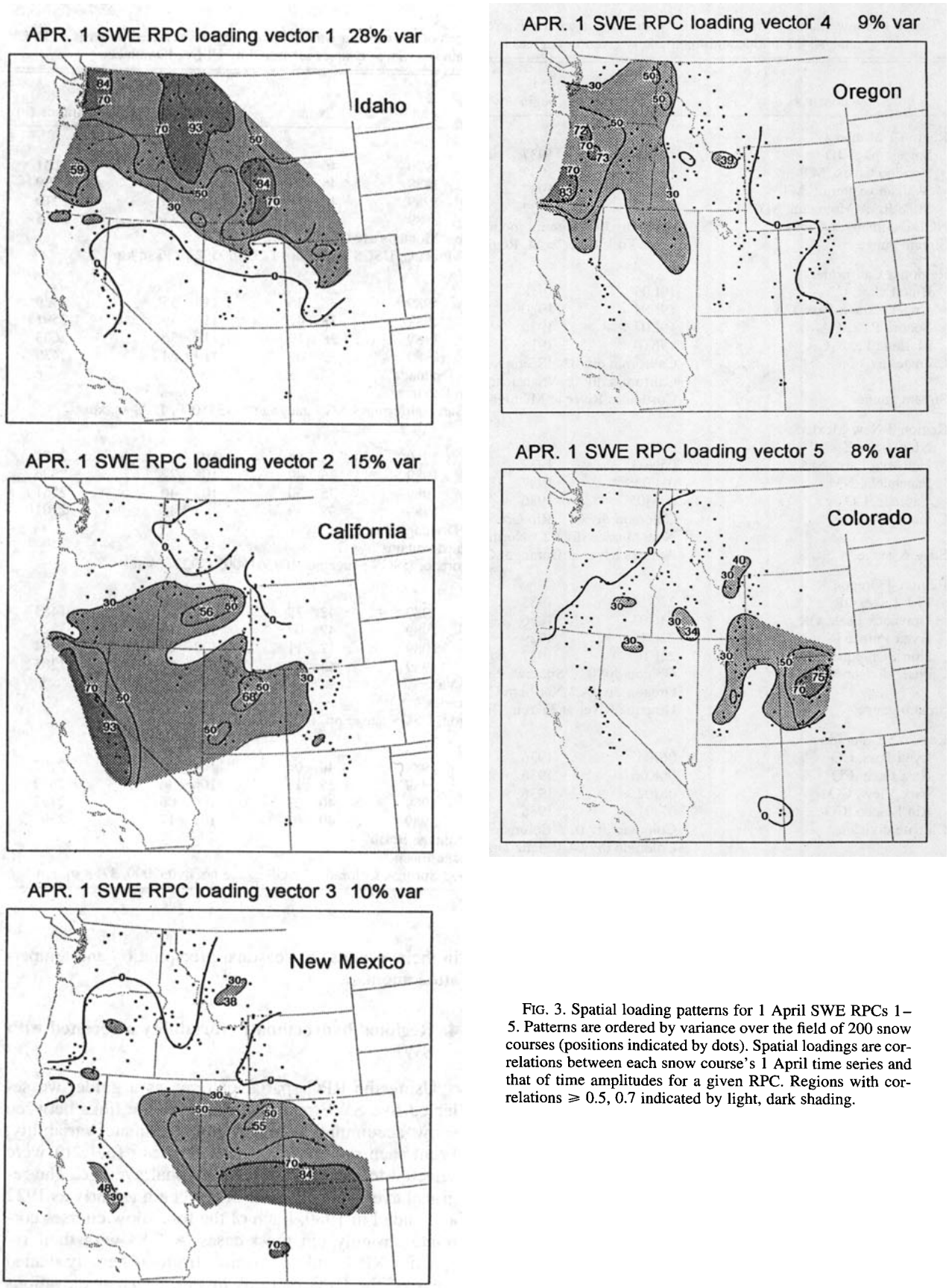


FIG. 3. Spatial loading patterns for 1 April SWE RPCs 1–5. Patterns are ordered by variance over the field of 200 snow courses (positions indicated by dots). Spatial loadings are correlations between each snow course's 1 April time series and that of time amplitudes for a given RPC. Regions with correlations ≥ 0.5, 0.7 indicated by light, dark shading.

TABLE 1. Snow courses, climate divisions, stream gauges for five regions of the western United States (regions selected on basis of RPC analysis). Latitude and longitude is given in degrees and minutes. Snow courses identified by USDA ID number.

Snow course	ID	yr begin	yr end	N lat	W lon	Altitude (m)
Region 1 Montana						
Savage pass, ID	14C04	1937	1990	46 28	114 38	1881
Hoodoo Creek, MT	15C01	1937	1989	46 59	115 01	1799
Skalkaho Summit, MT	13C03	1937	1989	46 15	113 46	2210
Slide Rock Mountain, MT	13C02	1937	1989	46 35	113 34	2165
NCDC climate divisions	Idaho div 04 "Central mountains Montana div 01 "Western"					
Stream gauge	Clark Fork River at St. Regis, Montana, USGS gauge no. 12354500, 27 736 sq km.					
Region 2 California						
Blue Lakes, CA	19L05	1918	1989	38 36	119 55	2439
Upper Carson Pass, CA	19L04	1930	1989	38 42	119 59	2591
Sonora Pass, CA	19L07	1930	1989	38 19	119 36	2683
Marlette Lake, CA	19K04	1915	1989	39 09	119 54	2439
Climate divisions	California div 02 "Sacramento Drainage"					
	California div 05 "San Joaquin Drainage"					
Stream gauge	Cosumnes River at Michigan Bar, California, USGS gauge no. 11335000, 1388 sq. km.					
Region 3 New Mexico						
Nutrioso, AZ	09S04	1938	1989	33 54	109 09	2591
Hematite Park, NM	05N03	1937	1989	36 40	105 22	2896
Panchuela, NM	05P02S	1937	1989	35 50	105 40	2561
Cascade, CO	07M05	1936	1989	37 39	107 48	2701
Climate divisions	Colorado div 05 "Rio Grande Drainage Basin"					
	New Mexico div 02 "Northern Mountains"					
Stream gauge	Animas River at Durango, Colorado, USGS gauge no. 109361500, 1792 sq. km.					
Region 4 Oregon						
Fish Lake, OR	22G14	1933	1989	42 23	122 21	1483
Grayback Peak, OR	23G03	1936	1989	42 07	123 18	1829
Hyatt Prarrie Reservoir, OR	22G16	1929	1989	42 11	122 28	1494
North Umpqua, OR	22F16	1937	1989	43 17	122 09	1285
Climate divisons	Oregon div 03 "Southwestern Valleys"					
	Oregon div 04 "Northern Cascades"					
Stream gauge	Umpqua River at Elkton, Oregon, USGS gauge no. 14321000, 9539 sq. km.					
Region 5 Colorado						
Lynx Pass, CO	06J06	1936	1989	40 05	106 40	2707
Nast Lake, CO	06K06	1936	1989	39 18	106 36	2652
Park View, CO	06J02	1936	1989	40 22	106 06	2793
Rio Blanco, CO	07J01	1936	1989	40 02	107 17	2591
Climate divisions:	Colorado div 02 "Colorado Drainage Basin"					
	Colorado div 04 "Platte Drainage Basin"					
Stream gauge	Roaring Fork River at Glenwood Springs, Colorado, USGS gauge no. 09085000, 3758 sq. km.					

count for considerable regional variance, with correlation maxima ranging from 0.7 to 0.9.

The spatial loadings of the RPCs (Fig. 3) are similar to the transect correlations (Fig. 2) in producing temporally independent modes of variability in close proximity. For example, the RPCs have adjacent centers in the Colorado and New Mexico regions and also in the California and Oregon regions. In the latter example, there is an abrupt delineation between the California RPC and the Oregon RPC at about the California–Oregon border. At first glance, this seemed surprising, since the winter precipitation in Oregon and central California emanates from similar winter atmospheric circulation patterns (Klein and Bloom 1987; Cayan and Peterson 1989). As will be revealed, however, the snow courses in these two regions are quite different in their response to seasonal precipitation and temperature anomalies.

4. Regional hydroclimatic variability associated with SWE

Using the RPC spatial patterns as a guide, we selected five SWE regions to analyze the links between snow accumulation and regional climate variability. From each region, four snow courses (Table 1) were selected to provide a single regional average. The regional average SWE time series begin as early as 1922 and ended in 1989. Each of the four snow courses correlate strongly (in most cases ≥ 0.8) with their respective RPC, and were drawn from the heavily shaded areas of the RPC patterns in Fig. 3. Their elevations

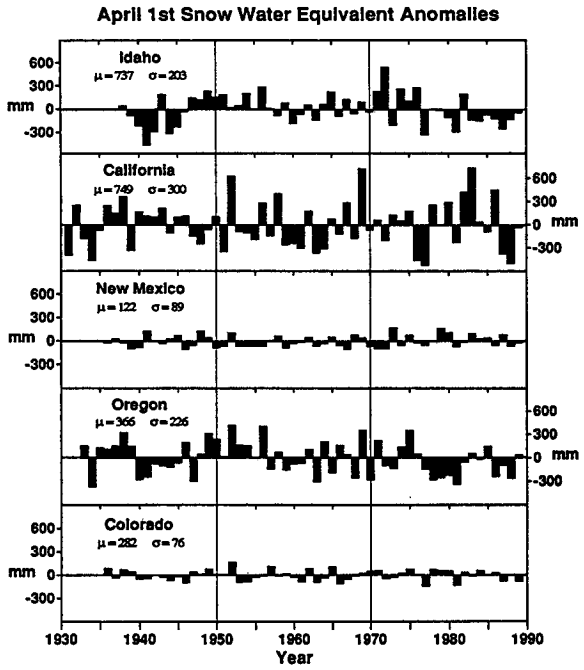


FIG. 4. Time series of April 1 SWE anomalies (mm) for each of the five regions.

(Table 1) range from 1285 m in Oregon to 2876 m in the southern Rocky Mountains in New Mexico. Year-to-year variability of the regional 1 April SWE is illustrated by the anomaly time series plotted in Fig. 4. In the following sections, several analyses are performed based upon the two sets of years having high and low SWE.

To diagnose the climate influences on snow accumulation, the SWE time series for each region were related to surface climate variables throughout the water year, beginning in the fall preceding the 1 April measurement. (The water year is October preceding a given calendar year through September of that calendar year.) Surface variables considered are precipitation, temperature, SWE, and streamflow. The precipitation and temperature are averages of two NCDC climate divisions within or nearby each region (Table 1). Streamflow is from stream gauges above which there is little or no impact of water management, and which have records beginning no later than 1930 (Cayan and Webb 1992). Streamflow records, all from long-standing U.S. Geological Survey gauges (Fig. 1 and Table 1) include: 1) Clark Fork River in Montana, 2) Cosumnes River in California, 3) Animas River in southern Colorado, 4) Umpqua River in Oregon, and 5) Roaring Fork River in Colorado.

Figure 4 indicates considerable year-to-year variability; overall, there is relatively low year-to-year correlation of 1 April SWE (Table 2). However, inspection of the regional time series indicates some interesting multiyear variability. Noteworthy years of anomalous snow accumulation over the much of the West include very high snow years 1938, 1952, 1969, and 1983; and very low snow years in 1934, 1963, 1977, and 1981. There are also some marked "runs" of positive and negative SWE anomalies. In particular, the SWE variability contains evidence of the "shift" in climate in the Pacific North American sector in 1977 that has been reported by several previous authors (Ebbesmeyer et al. 1991; Trenberth 1990; Changnon et al. 1993; Miller et al. 1994; Graham 1994). The shift commenced in the winter of 1977 with a series of winters having a deepened Aleutian low and a strengthened ridge over western Canada; this deflected North Pacific storms north of the western United States. Trenberth (1990) indicated that this amplified Pacific/North American (PNA) circulation was mainly a November–March cool season phenomenon; several authors (e.g., Ebbesmeyer et al. 1991; Miller et al. 1994) demonstrated that it produced changes in physical properties and biological components of several North Pacific and West Coast systems. The SWE anomalies in Fig. 4 show a decrease in 1977 toward negative anomalies in the Oregon and the Idaho time series, and a coincident increase toward positive anomalies in the New Mexico and Colorado time series. This shift is consistent with the change in the Rocky Mountain snow accumulation noted by Changnon et al. (1993), who noticed a pivotal zone in the central Rockies with decreasing snowpack in the northern Rockies (Idaho and Montana) and increasing snowpack in the southern Rockies (Utah and

TABLE 2. Regional SWC autocorrelation (1931–1988).

	Month-to-Month										Year-to-year lag (yrs)				
	Correlation coefficient					Number of years									
	1 Jan	1 Feb	1 Mar	1 Apr	1 May	1 Jan	1 Feb	1 Mar	1 Apr	1 May	1	2	3	4	5
Idaho	0.58	0.80	0.91	1.00	0.82	54	54	54	54	53	0.13	0.20	0.25	0.26	−0.04
California	0.55	0.72	0.91	1.00	0.91	30	60	61	71	65	−0.10	−0.01	−0.06	0.09	−0.04
New Mexico	0.22	0.71	0.80	1.00	0.72	21	54	54	59	54	−0.09	−0.39	0.00	0.37	−0.04
Oregon	0.66	0.70	0.79	1.00	0.62	54	57	56	57	42	−0.01	0.25	−0.02	0.04	−0.13
Colorado	0.56	0.83	0.90	1.00	0.78	28	54	54	54	54	−0.17	−0.13	−0.07	−0.04	0.09

southern Colorado). The present results indicate that in the Northwest the diminished snowpack extended to the Pacific Coast, and that in the Southwest the increased snowpack extends southward into Arizona and New Mexico. Also, from the California time series, note that the Sierra Nevada lies within the zone between these negative and positive shifts, and actually exhibits some elements of both. During 1977, the West Coast ridge was very extensive, so that storms were blocked from southern Canada, the northern and central Rockies southward through the Sierra Nevada. From 1978 through the mid-1980s, however, frequent winter storms deposited ample snowpack in the Sierra Nevada across to the southern Rockies.

Associated with these changes, Changnon et al. (1993) detected changes in daily synoptic circulation patterns: an increase in cutoff and split flow circulations, which favor heavier precipitation in the Southwest; and a decrease in the meridional northwest and west northwest circulations, which favor precipitation in the Northwest. In section 5d it is shown that these changes are consistent with two facets of the circulation changes after 1977: 1) deeper central North Pacific lows in the winters; and 2) increased occurrences of the warm phase of ENSO. These circulation changes are mirrored by composites of winter and March circulation, as shown in association with anomalous SWE in section 5b.

In examining the regional surface variables, we considered sequences of monthly composite anomalies corresponding to high and low 1 April SWE cases. Anomaly composites of monthly SWE, streamflow, precipitation, and temperature are based upon the regional average 1 April SWE from the four selected snow courses. Tests showed that the sets of years identified (Table 3) are not very sensitive to the particular choice of the four snow courses. Alternate sets of four snow courses for each of the regions yielded nearly the same set of years. The composites are monthly values averaged over the 20 years with the greatest positive or negative regional average 1 April SWE anomalies. (Twenty years divides the data approximately into terciles.) The months depicted by the composites include September of the previous year through September following the 1 April SWE criterion. Significance of the positive and negative anomalies, relative to the overall sample long-term mean, was computed using the 90% significance threshold of a two-tailed t test. However, the significance of the results may be stronger than the two-tailed t test indicates, because in all of the composites with significant anomalies, the relations that emerge are in the expected sense of positive precipitation (wet) and negative temperature (cool) anomalies accompanying positive SWE anomalies. A multiple regression analysis, reported in Table 4, reinforces these results.

Composite monthly anomaly sequences, based upon the high and low 1 April SWE cases, are shown in Figs. 5a–e. The Idaho high and low SWE extreme composites (Fig. 5a) exhibit significant differences in precipitation, temperature, SWE, and streamflow. The Idaho composites are typical of the other regions in that an anomalous SWE begins to build up early in winter, with significant departures appearing in 1 February and prevailing through 1 May. The persistence of the Idaho snowpack anomalies is indicated by the month-to-month correlations linking the SWE anomalies—these are 0.58, 0.80, 0.91, and 0.82 for 1 April versus 1 January, 1 February, 1 March, and 1 May, respectively. SWE autocorrelations are similar for the other four regions (Table 2). Most significant precipitation anomalies occur from November through January, with differences exceeding 25% of monthly long-term means. An apparent temperature signal appears in February through May, with composite differences exceeding 1°C. The temperature difference is particularly large in March, at −2.1°C (high-minus-low SWE composites). While most temperatures during the following months

TABLE 3. Years with high and low 1 April SWC for five western United States regions (20 years each 1931–1988).

Idaho																				
high	1938	1943	1946	1947	1949	1950	1951	1952	1954	1956	1964	1965	1967	1969	1971	1972	1974	1975	1976	1982
low	1936	1937	1940	1941	1942	1944	1945	1960	1963	1966	1968	1973	1977	1979	1980	1981	1984	1986	1987	1988
California																				
high	1922	1927	1932	1936	1938	1940	1941	1943	1952	1956	1958	1962	1965	1967	1969	1978	1980	1982	1983	1986
low	1924	1936	1929	1931	1933	1934	1939	1948	1951	1959	1960	1961	1963	1964	1968	1976	1977	1981	1987	1988
New Mexico																				
high	1937	1941	1945	1948	1949	1952	1957	1962	1965	1968	1969	1973	1975	1978	1979	1980	1982	1983	1984	1987
low	1931	1933	1934	1935	1939	1940	1946	1947	1953	1954	1955	1959	1963	1966	1967	1971	1972	1977	1981	1988
Oregon																				
high	1933	1935	1936	1937	1938	1946	1949	1950	1952	1956	1958	1962	1964	1969	1971	1975	1983	1984	1985	1989
low	1934	1940	1941	1943	1944	1947	1959	1960	1963	1965	1968	1970	1972	1977	1978	1979	1981	1986	1987	1988
Colorado																				
high	1936	1949	1952	1956	1957	1958	1962	1965	1969	1970	1971	1974	1975	1978	1979	1980	1982	1983	1984	1986
low	1937	1939	1940	1941	1943	1944	1945	1946	1950	1953	1954	1955	1961	1963	1964	1966	1972	1977	1981	1987

(April–August) do not reach the 90% significance threshold, it is intriguing that those associated with high SWE composite anomalies are consistently cooler than average, and those for low SWE are consistently warmer than average. A similar temperature lag also occurs in New Mexico, Colorado, and to some extent California. The March temperature signal suggests that delayed or accelerated snowmelt influences the 1 April SWE anomalies in Idaho, in addition to effects of excess or deficit precipitation. Clark Fork River runoff lags the snowpack anomalies by 2–3 months, with marked streamflow anomalies first appearing in April and persisting through August. Over the water year, high SWE streamflow averages 123% of its long-term average compared to that for low SWE at 77%.

For the California region, composite SWE anomaly conditions (Fig. 5b) are similar to those for Idaho, but with a stronger relation to precipitation and a weaker one to temperature. Significant precipitation anomalies occur in all of the major precipitation months, first appearing in November and continuing through April. The water year mean precipitation is 130% and 73%, respectively, of long-term average. Both high and low SWE cases contain significant anomalies, although not in all of the months. The evidence for a temperature influence in the California composites is subtle, although previous studies (Aguado et al. 1992; Cayan et al. 1993; Cayan and Peterson 1993; Dettinger and Cayan 1995) detect winter and spring temperature effects on spring-summer runoff in the Sierra Nevada, particularly for middle elevation (1800–2500 m) watersheds. Interestingly, the multiple regression analysis presented below picks up a winter temperature influence on spring SWE in California. While they do not reach the significance threshold, the temperature anomalies are consistent with those of the snowpack, in that winter and spring months are cooler for high SWE and warmer for low SWE. Interestingly, the greatest difference occurs during April, the period *after* the 1 April SWE criteria upon which the composites are based. The California SWE anomalies build steadily from 1 January, and are greatest for the 1 May samples, which reach 156% and 39%, respectively, of the May long-term average. Composite streamflow in Cosumnes River is remarkably different between the two cases, more than any of the other four regions. The Cosumnes streamflow annual hydrograph resembles that of the entire Sacramento–San Joaquin River outflow into the San Francisco Bay estuary (Peterson et al. 1989), having a broad peak from February through April. The broad shape of the Cosumnes River response is similar to that of the Umpqua River in Oregon (Fig. 5d), although the Cosumnes peak is delayed by 1–2 months. On the other hand, the Cosumnes flow response is 1–2 months earlier than that of the rivers in the Idaho, New Mexico, and Colorado regions (Figs. 5a, 5c, 5e), which have sharp peaks in May and June. Significant streamflow differences begin in December and continue through July. Composite Cosumnes River flow mirrors the great difference between high and low SWE cases, with water year discharge at 179% and 43% of its long-term average.

For New Mexico, SWE fluctuations respond to both precipitation and temperature (Fig. 5c). Although they are not as pronounced as in California, precipitation differences of the high versus low SWE composites are significant for October through April. Over the water year, composite precipitation for the high and low SWE was 108% and 90% of long-term average. The precipitation climatology in New Mexico and in the Colorado region differs from the other three regions in not having a decided winter maximum. Next to those in Oregon, the SWE anomalies in the New Mexico region display the greatest temperature signals. Temperature differences are negative (cooler for high SWE cases) in November, December, March, and April, with differences exceeding 1°C. Significant SWE differences first appear in February and continue through 1 May. The 1 April SWE anomalies are 120% and 78% of long-term average. In response to anomalous SWE, streamflow of the Animas River has greatest differences in May–July, the period of peak flow. The Animas River annual discharge for the two composite cases is 126% and 70% of its long-term mean.

Oregon is unique among the other four regions in that SWE exhibits a stronger association with temperature than with precipitation (Fig. 5d). High-minus-low April SWE composite temperature differences are about 2°C in February and March. This sensitivity is probably because elevations are relatively low (generally less than 2000 m), so winter temperature determines whether precipitation occurs as rain or snow, and whether the snow melts or accumulates (elevations of the snow courses are shown in Table 1). The Oregon precipitation signal is not convincing, with the largest positive difference occurring in November, a month when little snowpack is established. None of the core winter months have precipitation anomalies exceeding the t test threshold. The insensitivity to precipitation is not because Oregon precipitation has low anomaly variability—the coefficient of variation of monthly precipitation in Oregon during the cool season ranges from about 0.3 to 0.6, which is comparable to that of the other regions. Also, to test the stability of the Oregon results, another set of temperature and precipitation composites was computed using high and low years from an alternative set of snow courses in Washington and northern Oregon. These results were nearly identical to those in Fig. 5d, so it appears that these are representative of the lower elevation mountainous region of the Pacific Northwest. In Oregon, the SWE anomalies begin to accumulate by 1 January, reach their greatest excursion on 1 April, but fall off rapidly on 1 May. For the low SWE case, SWE is nearly zero on 1 May. Although Umpqua River has less contribution by snowmelt runoff than the other rivers, it still

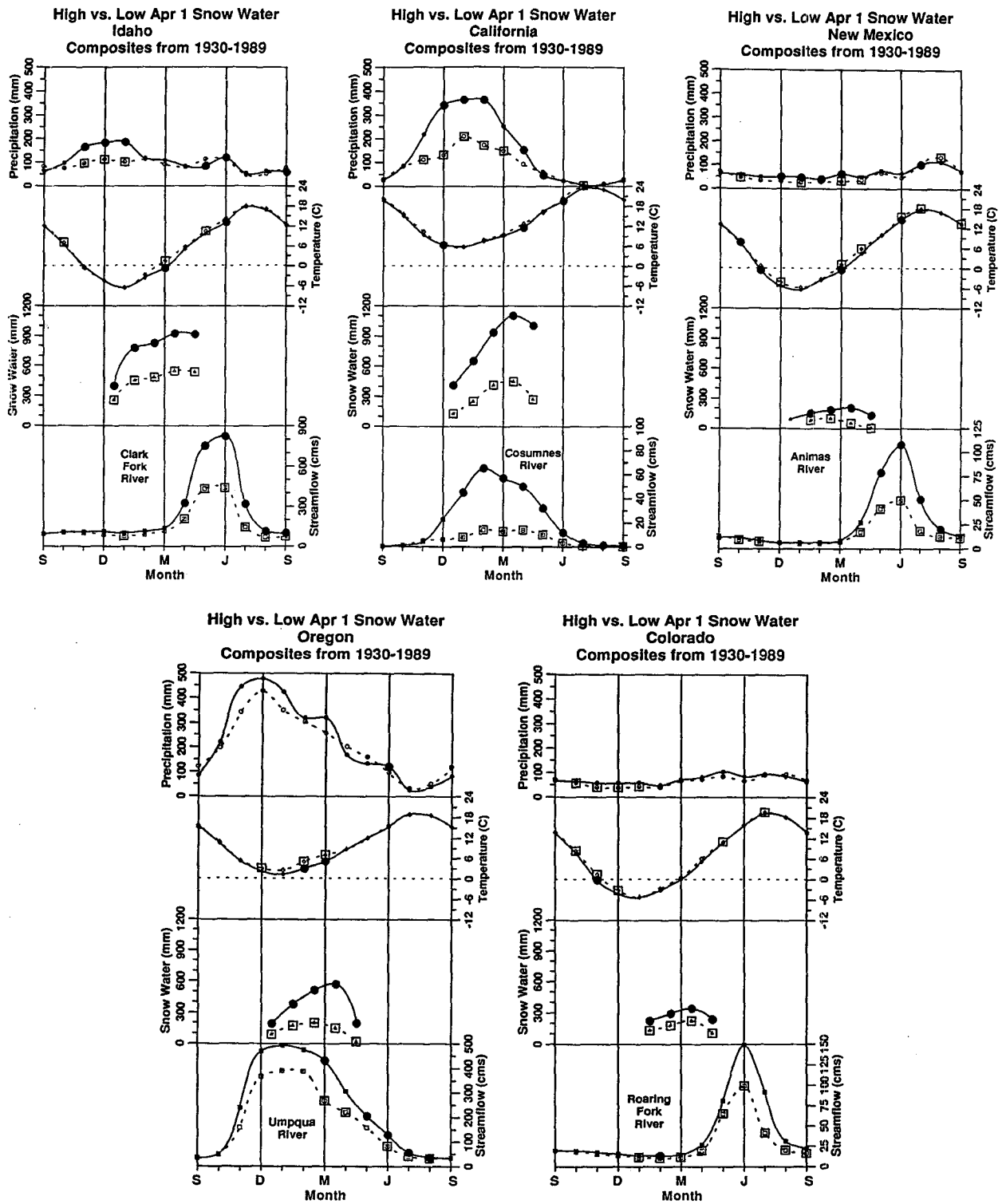


FIG. 5. Composite of monthly precipitation (mm), temperature (°C), SWE (mm), and streamflow ($m^3 s^{-1}$) for years with 20 lowest (dashed lines and open squares) and 20 highest (solid lines and solid dots) 1 April SWE. Composite values whose anomalies have magnitude exceeding 90% significance threshold of two-tailed t test are shown with large symbols. Shown for each of five regions described in text.

TABLE 4. Multiple regression: sensitivity of SWE to precipitation and temperature 1 April SWE is predicted from winter and March precipitation and temperature.

Region	Variance accounted for	Predictors* Winter precipitation	March precipitation	Winter temperature	March temperature
Idaho	77%	.75	.18		−.27
California	84%	.80	.32	−.15	
New Mexico	41%	.36	.17		−.35
Oregon	58%	.21	.14	−.41	−.44
Colorado	34%	.60			

* Values given under predictors are normalized regression coefficients.

displays a significantly greater discharge under high SWE conditions than low. The average Umpqua River water year discharge for high and low SWE cases is 115% and 86%, respectively, of its long-term mean.

For the Colorado region, associations between SWE and the other climatic variables are weaker than those for the other regions in this study. Precipitation differences are positive (high SWE has more precipitation) in November–January. The deficit precipitation associated with low SWE is more unusual than its high precipitation counterpart. Concerning temperature, high-minus-low SWE temperature differences are negative, but it is mostly the warmer temperatures associated with low SWE that stand out. The temperature anomalies are most important in fall and early winter (October–December). The reason for the fall temperature response may be that 1) fall temperatures are closer to the rain versus snow boundary (winter is always cool enough for snow in the higher elevations) and 2) fall precipitation tends to be greater than winter, so temperature anomalies then would have added importance. Interestingly, low SWE cases are preceded by anomalously warm temperatures in summer, particularly July and August. SWE differences (high minus low) are positive for all 5 months, building from 1 January to a maximum on 1 April. High and low 1 April SWE composite values are 110% and 78%, respectively, of its long-term mean. Roaring Fork River, which has a marked June peak streamflow, has its greatest response to the SWE extremes in June and July. Over the water year, high and low SWE discharge was 108% and 74% of its long-term average. Like the precipitation and temperature, the streamflow anomalies display greater significance for the low SWE composite than for those of the high SWE composite.

To verify the relationship of local precipitation and temperature to SWE, we also conducted a multiple linear regression analysis (Table 4). For each region, 1 April SWE was specified by four predictors: winter precipitation, March precipitation, winter temperature, and March temperature. The entire period of record (beginning no later than 1937) was employed. All of the predictors presented in Table 4 were retained in the multiple regression at above the 97% level of confidence under an F test. The results are in accord with the composite monthly precipitation and temperature described above. For Idaho, California, and Colorado, precipitation is the dominant predictor, especially during winter, and to a lesser extent in March. Temperature was selected as a significant predictor in all regions except Colorado; in three of these cases, March temperature is the dominant temperature predictor. While precipitation always enters as a positive coefficient (higher precipitation associates with higher SWE), temperature always enters as a negative coefficient (warmer temperatures associate with lower SWE). The regression models account for more than 70% of the SWE variance in California and Idaho, and for less than 50% of the SWE variance in New Mexico and Colorado.

5. Large-scale spatial structure

a. Winter circulation patterns, high versus low SWE

If anomalous SWE behaves similarly to anomalous precipitation and temperature, it should be strongly organized by the atmospheric circulation. To shed further light on local and regional SWE variations, composites of the anomalous 700-mb height field were constructed, analogous to composites of surface variables presented above. The Northern Hemisphere 700-mb monthly means begin in December 1946, so these composites encompass fewer cases (about 14 in each) than those from the precipitation, temperature, SWE, and streamflow (20 in each). The 700-mb height composites are shown for the winter (December–January–February) average and for March. In addition to the circulation, maps of divisional surface precipitation and temperature anomalies over the United States are included and are discussed in section 5c.

Considering the Idaho low SWE cases (Fig. 6a, upper), the winter circulation pattern appears to be the deep Aleutian Low phase of the Pacific/North American (PNA) pattern (Barnston and Livezey 1987). This pattern has a bold North Pacific signature, with a neg-

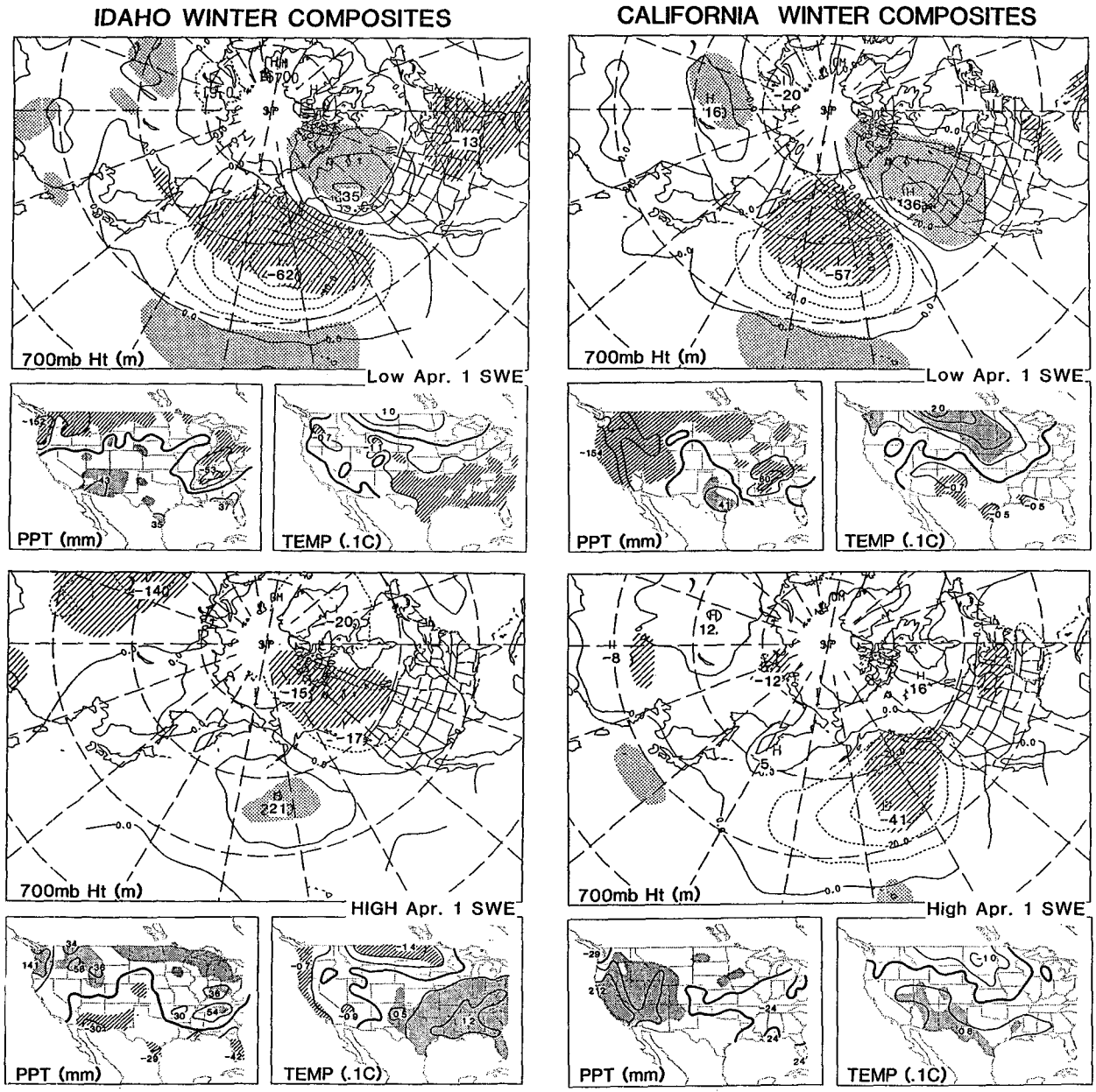


FIG. 6. (a) Map of composites of winter Northern Hemisphere 700-mb height (m), United States precipitation (mm), and United States temperature (°C) for groups of years with low SWE (top frame) and high SWE (bottom frame) for Idaho region. Stipling (hatching) indicates regions where composite values with positive (negative) anomalies have magnitude exceeding 90% significance threshold of two-tailed t test. (b) Same as Fig. 6a but for winter composites for California region.

ative anomaly center south of the Aleutian Islands. A positive anomaly center is located over western Canada and shields much of the western United States from storms, shunting cold arctic air farther east. Accordingly, the Northwest is dry, consistent with the low SWE conditions in Idaho. Downstream, weather over the eastern United States is dry and cold, a well-known symptom of positive PNA episodes. For high SWE in Idaho (Fig. 6a, lower), positive 700-mb height anomalies appear in the central North Pacific south of the Aleutians, and negative anomalies occur over western Canada. This pattern produces two elements conducive to heavy snow: ample precipitation and cooler than normal air throughout the Northwest. Note that the Idaho low SWE pattern has larger anomalies than the high SWE composite case, presumably indicating a more consistent circulation for the dry cases. Interestingly, both high and low SWE precipitation maps exhibit an

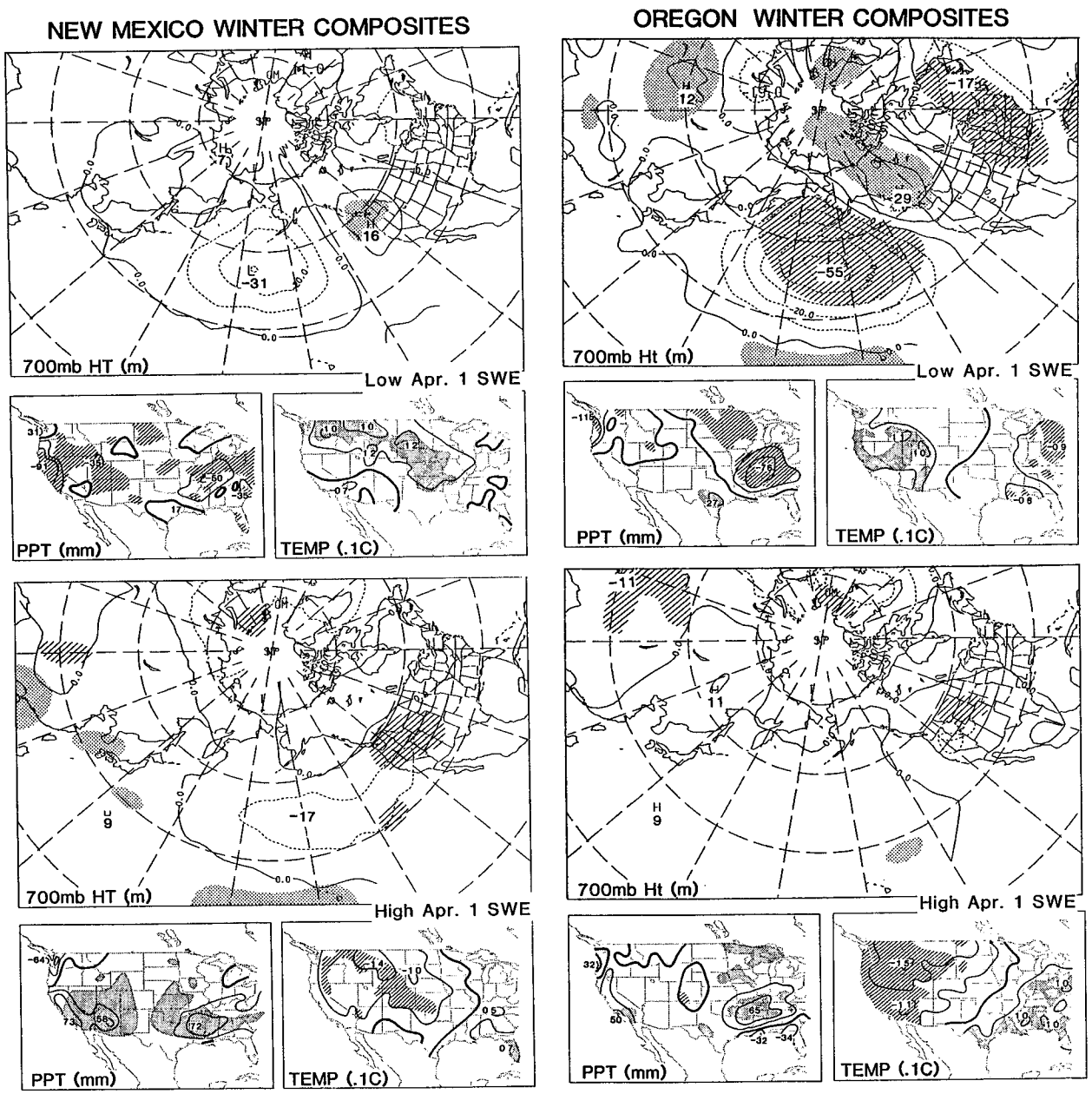


FIG. 6. (*Continued*) (c) Same as Fig. 6a but for winter composites for New Mexico region.
(d) Same as Fig. 6a but for winter composites for Oregon region.

out-of-phase relation of precipitation anomalies over the Northwest with those in the Southwest. The high and low SWE winter circulation patterns for Oregon are quite similar to those for Idaho.

The California low SWE winter circulation pattern (Fig. 6b, upper) resembles drought-producing circulations in the state (Namias 1978), with negative 700-mb height anomalies in the central North Pacific and strong positive anomalies seated over the West Coast centered over British Columbia. Although there are frequent North Pacific storms with this pattern, they are carried northward toward Alaska, leaving dry conditions over California and much of the western states. Winter circulation leading to high SWE winter in California (Fig. 6b, lower) is quite clearly the same pattern that produces heavy precipitation and high winter–spring streamflow, as noted in previous studies (e.g., Namias 1979; Klein and Bloom 1987; Cayan and Peterson 1989). Wet winters in California have considerable variability in their mean circulation patterns, but

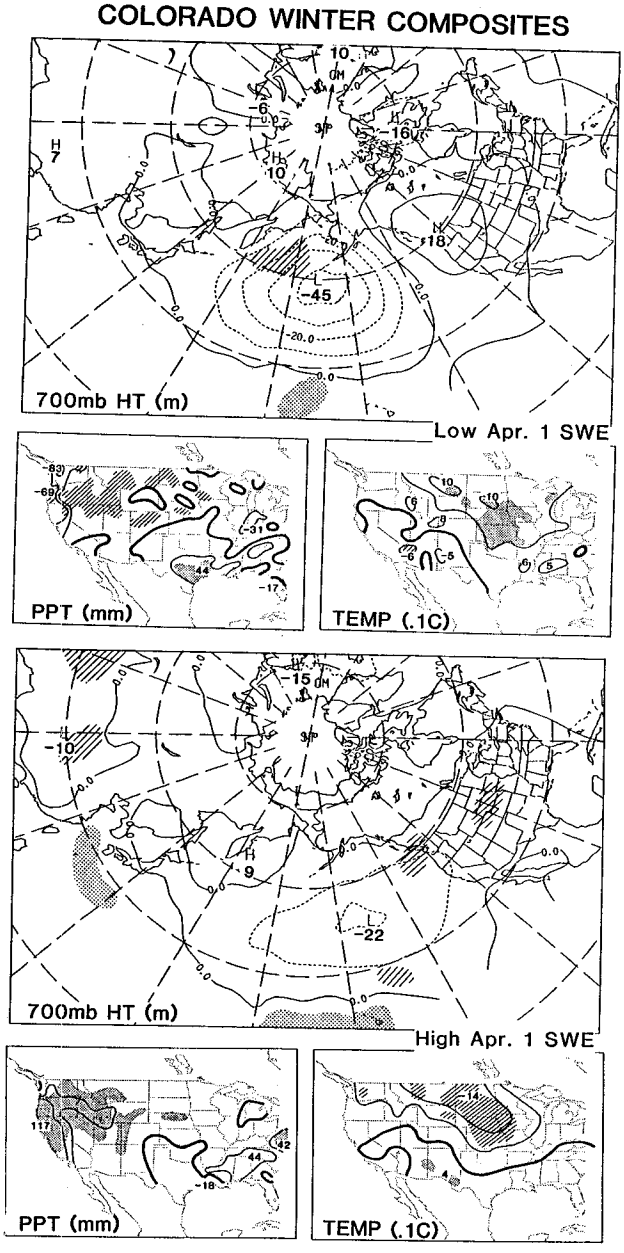


FIG. 6. (*Continued*) (e) Same as Fig. 6a but for winter composites for Colorado region.

they usually involve negative height anomalies just offshore, centered at about 40°N, 130°–140°W. This feature is clearly seen in the high SWE composite 700-mb height anomalies of Fig. 6b, lower. This low pressure anomaly is a symptom of frequent storms and strong westerly-to-southwesterly winds over the western states, producers of heavy precipitation and snow at higher elevations (Fig. 6b, lower panel).

There is quite a contrast in the character of circulation patterns associated with the two extremes of SWE. Much similarity is found for the low SWE; notice how the 700-mb height anomaly pattern for each of the five regions (Figs. 6a–e, upper panels) contains negative anomalies in the central North Pacific and positive anomalies over the Pacific Northwest. This resembles the PNA pattern (Barnston and Livezey 1987), and tends to block severe storms and inhibits cold outbreaks over the West. Section 5d shows that low winter pressures over the central North Pacific is often sufficient to reduce the 1 April snowpack over the Northern Rockies, Cascades, and northern Sierra Nevada. In comparison to the low SWE cases, there is more diversity in the circulation patterns associated with high regional SWE. While high SWE in each region (Figs. 6a–e, lower panels) is associated with negative anomalies (anomalously low pressure) in its vicinity, these centers are not as strong and are not very similar from one region to the next. For high SWE in the New Mexico and Colorado regions, the circulation composite is not as strong as that for low SWE cases. This helps to explain the relatively low correlations plotted on meridional section M3 (Fig. 2) between SWE anomalies at the Arizona snow courses and those to the north in southern Utah. Sellers (1968) showed similar boundaries between monthly precipitation anomalies in the southern Rocky Mountains using an empirical orthogonal function analysis of precipitation over the western United States. Changnon et al. (1993) demonstrated that several synoptic circulation patterns produce heavy precipitation in different parts of the southern and northern Rockies. Their results and inspection of individual seasons indicates that heavier than normal snows are produced by a variety of different circulation patterns, implying that there will not be a strong concensus when all of the cases are aggregated.

b. March circulation patterns, high versus low SWE

Because the regional SWE is also affected by early spring climate variations (section 4), a second set of diagnostics was produced for March. As for winter, precipitation and temperature anomalies over the western United States are shown in companion; recall that early spring temperature and precipitation both contribute to 1 April SWE anomalies (section 4). As for winter, the criteria for the two groups of years were high and low extreme 1 April SWE; again, there are only about 14 members in each 700-mb height composite, compared to 20 members for the temperature and precipitation composites. Because several of the composite maps are quite similar to those for winter, only two March composite maps are displayed, although the results for each of the regions are discussed.

Considering March 700-mb anomalies for Idaho, the low SWE pattern (Fig. 7a, upper) is quite similar to that for winter. This pattern, which resembles the PNA, has anomalously high pressure in most of the subtropical North Pacific, anomalously low pressure centered

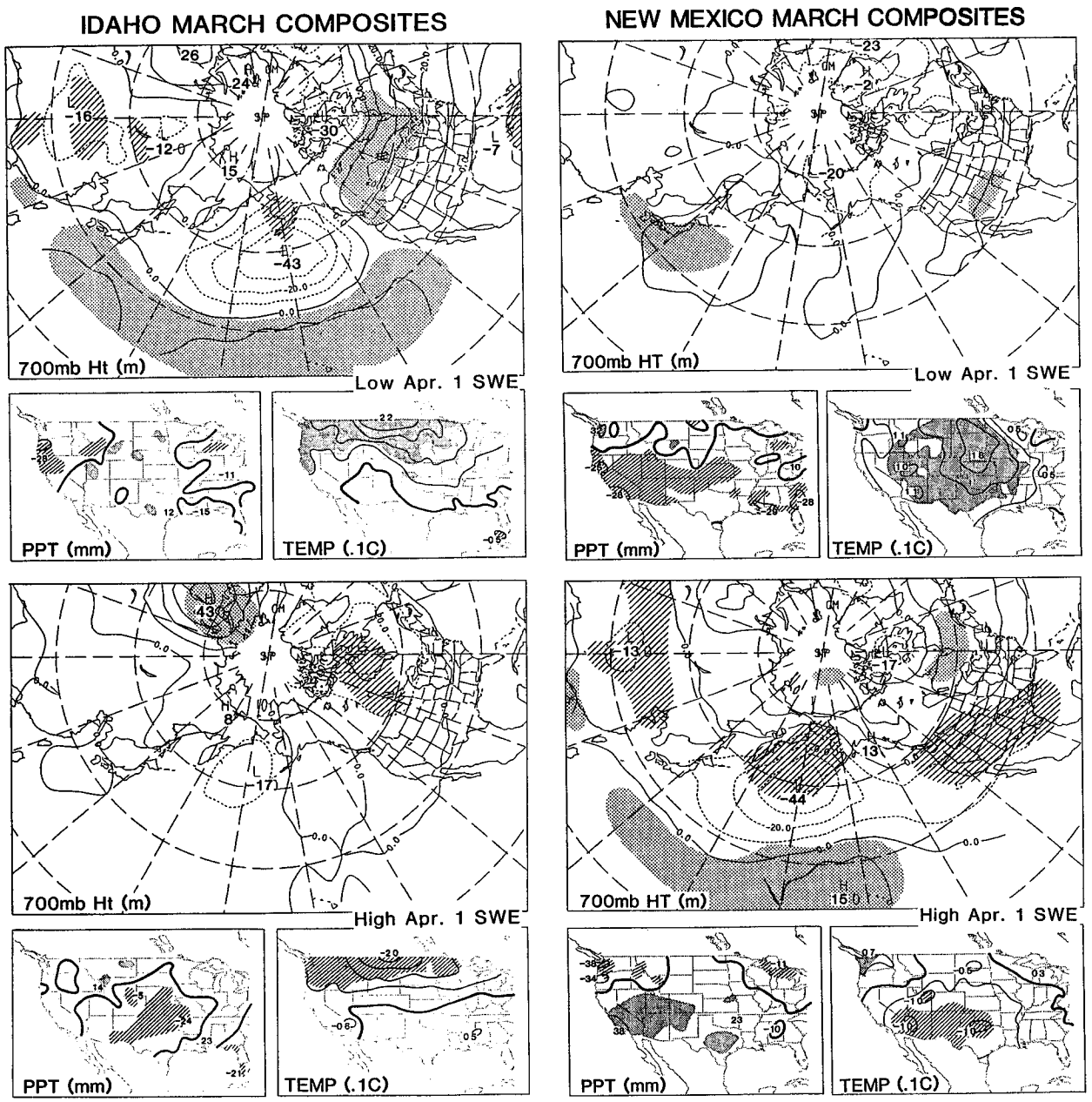


FIG. 7. (a) Same as Fig. 6a but for March composites for Idaho region. (b) Same as Fig. 6a but for March composites for New Mexico region.

south of the Aleutian Islands, and anomalously high pressure over western central Canada. The effect is to have fewer storms and warmer temperatures than average. The high SWE pattern (Fig. 7a, lower), is less distinct over the North Pacific, but does have significant negative 700-mb height anomalies stationed over central Canada. As described in section 5c, this circulation allows cold air and storminess to penetrate southward into the Northwest.

For California, the March low SWE pattern (not shown) contains a positive anomaly center well off the Pacific Coast, with negative anomalies farther north in the Gulf of Alaska. This circulation would inhibit storms from approaching the California coast. The March circulation associated with high 1 April SWE is quite similar to those for winter, with a negative height anomaly in the eastern North Pacific just offshore from central California. This is a signature of increased

storminess and anomalously high precipitation across the Southwest.

For New Mexico, the March 700-mb height anomaly pattern for the low SWE cases (Fig. 7b, upper) is not particularly strong, but does contain significant positive anomalies over the Southwest. This feature is associated with light precipitation and warmer than average temperatures. The March circulation pattern associated with New Mexico high SWE (Fig. 7b, lower) is stronger, with negative height anomalies over much of the North Pacific, extending from south of the Aleutian Islands through the eastern North Pacific and into the southwestern United States. This pattern produces heavier than normal precipitation in the Southwest and cooler than normal temperatures from California through the southern Great Plains.

For Oregon, the March 700-mb height anomaly pattern for the low SWE cases (not shown) is quite strong, and very similar to the Oregon low SWE pattern for winter, in that the strong Aleutian Low phase of the PNA pattern produces deficit snowpack throughout the Northwest, as noted above and further discussed in section 5d. For the high SWE cases, the March 700-mb height anomaly pattern is quite weak, but does contain negative anomalies across the North Pacific from the Aleutians to the West Coast. This pattern would encourage more storms, heavier precipitation, and cooler temperatures in the Oregon region.

For Colorado, the March circulation patterns (not shown) are not very strong, although they exhibit the general tendency for positive 700-mb height anomalies associated with low SWE, and negative height anomalies with high SWE.

c. *Regional precipitation and temperature anomaly patterns*

The spatial patterns of precipitation and temperature are worthy of discussion, since they affect the anomalous snow patterns and help to explain how regional snow anomalies relate to the atmospheric circulation. Composite precipitation and temperature anomalies, aggregated from the 20 years of the low and high 1 April SWE categories, are presented at the bottom of each of the circulation maps in Figs. 6a–e (winter) and Figs. 7a–b (March). Shading indicates anomalies whose magnitudes exceed the 90% confidence level of rejecting the null hypothesis, using a two-tailed t test.

First consider the winter precipitation and temperature composites in Figs. 6a–e. In most cases, where the low and high cases have significant anomalies, they are expressions of opposite, symmetric spatial patterns. For Idaho (Fig. 6a), the winter precipitation has a bold pattern, reinforcing the monthly anomaly traces in Fig. 5a. Both low and high SWE maps contain a zone of anomalous precipitation (low and high, respectively) across the northern United States that extends eastward from the Northwest into the Great Lakes. Interestingly, both maps contain opposite phase precipitation in the southwestern United States, a tendency that was noted in the meridional section correlations in Fig. 2. The winter temperature associated with extreme 1 April SWE in Idaho does have broadscale anomalies, especially over the Southeast. However, the temperature anomalies are not significant in the Northwest, indicating that the downstream temperature is a teleconnected feature of the circulation but not a direct influence on Idaho SWE.

As it was for Idaho, winter precipitation in California is the major influence in both the low and high SWE cases (Fig. 6b). Negative (positive) precipitation anomalies blanket the western United States during low (high) SWE conditions. Largest magnitudes ($\geq$100 mm) occur in California, but have significant values that reach the interior Northwest. These patterns are consistent with the extensive positive (negative) anomaly circulation patterns (Fig. 6b) that occur during the low (high) SWE cases in the Sierra Nevada.

The New Mexico extreme SWE winters also display extensive precipitation anomalies, even though their winter circulation anomaly composites are weak; this is another symptom suggesting that several circulation patterns produce anomalous precipitation in the Rocky Mountains. The precipitation anomalies extend from California to New Mexico, with same-sign anomalies also occuring downstream over the southern Great Plains through the Southeast. Winter temperature anomalies also tend to accompany the low and high 1 April SWE anomalies, although the pattern is spotty.

Unlike the other regions, the Oregon extreme 1 April SWE is not associated with strong spatially extensive precipitation patterns, but does exhibit an impressive temperature signal. Consistent with the composite monthly temperature sequences in Fig. 5d, anomalous 1 April SWE is driven by anomalous temperature (in the expected sense of low temperatures producing higher snow accumulation). These anomalies typically approach 1°C, and cover a broad swath of the western United States.

For 1 April SWE in Colorado, winter precipitation forcing appears to be most important, with significant anomalies occuring from the West Coast through the Rocky Mountains. Like the Idaho and California composites, the temperature anomalies are not significant in the region.

Some interesting changes appear in the March maps in comparison to those for winter. For Idaho (Fig. 7a), the March precipitation pattern is small in scale, but the temperature composites contain strong, zonally extensive anomalies from the Pacific Northwest across the northern states to Minnesota. Anomalies have magnitudes from 1° to 2°C. For California (not shown), precipitation is again a major feature, but the negative precipitation anomalies associated with the low 1 April SWE are more extensive than the positive anomalies associated with the high SWE. For New Mexico (Fig. 8b), both precipitation and temperature contain strong

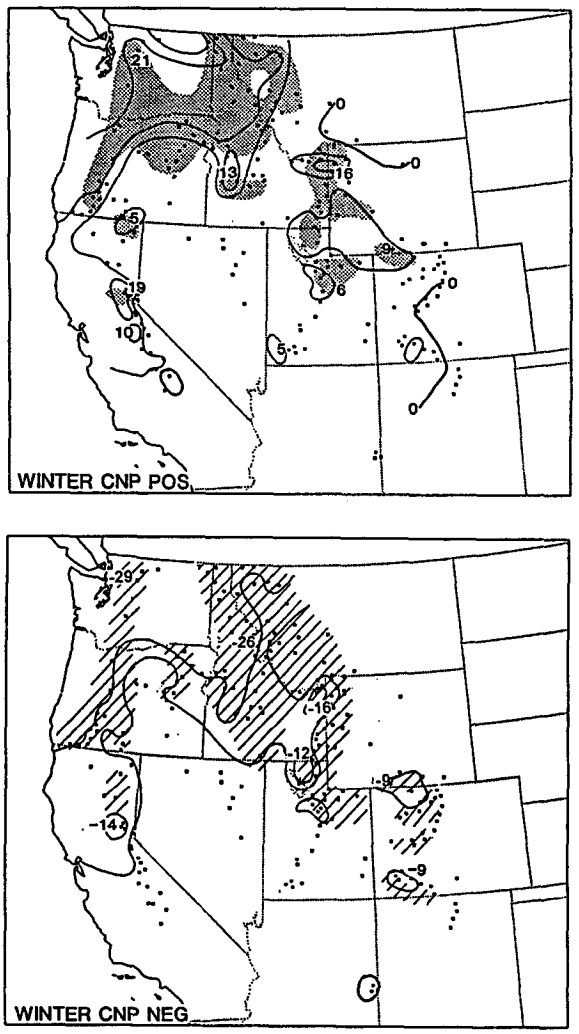


FIG. 8. Composite 1 April SWE (cm) following winters with positive and negative CNP. Contours at 0, 10, and 20 cm. Stipling (hatching) indicates regions where composite values with positive (negative) anomalies have magnitude exceeding 90% significance threshold of two-tailed t test.

spatially extensive anomaly patterns. For Oregon (not shown), like the winter composites, there is a strong temperature pattern, with widespread positive (negative) anomalies over most of the West associated with low (high) 1 April SWE. Precipitation is weaker with spotty anomalies over the Oregon region, but does have consistent anomalies (negative anomalies associated with low SWE and vice versa). Concerning the Colorado March composites (not shown), there are few significant regions of precipitation or temperature signals.

d. Influence of the Central North Pacific Low and ENSO

The regional composite circulation suggests that anomalous winter central North Pacific Lows provoke a broadscale regional response by the snowpack over the West. For an overall view of the effect of anomalous circulation from the Aleutian and western Gulf of Alaska low regions, Fig. 8 is a composite of the anomalous SWE over the western region associated with the central North Pacific sea level pressure index (CNP), an average of sea level pressure anomalies over (35°–55°N, 170°E–150°W). CNP is strongly correlated ($\rho = -0.9$) to the PNA index in winter (Cayan and Peterson 1989), a symptom of the strong teleconnectivity (Wallace and Gutzler 1981) of the central North Pacific center in the winter PNA circulation. The deep Aleutian low phase of the CNP (or PNA) is predominantly dry over much of the West and inhibits cold air masses from settling into the region. As indicated by the composite at the bottom of Fig. 8, there is deficit SWE over much of the western United States, including the northern Sierra, the Cascades, and most of the Rocky Mountains. Conversely, during winters having the weak Aleutian low phase of the CNP (or PNA), more winter storms penetrate the western states, and nearly the reverse anomaly pattern prevails, although the spatial pattern is quite weak (Fig. 8, top).

Since the El Niño/Southern Oscillation (ENSO) is known to affect both precipitation and temperature in the West (Redmond and Koch 1991; Cayan and Webb 1992), it should certainly affect the snowpack. Previous studies (Redmond and Koch 1991; Cayan and Webb 1992) have indicated that the warm tropical Pacific phase of ENSO during Northern Hemisphere winter is associated with diminished snowpack and streamflow in the Northwest and enhanced snowpack and streamflow in the Southwest. Using the winter Southern Oscillation Index (SOI) as a measure of ENSO, composites (Fig. 9) of the SWE anomalies over the West were formed, analogous to those for the CNP in Fig. 8. The choice of warm-phase and cool-phase years of ENSO are given in Cayan and Webb (1992). The April SWE observations that are tested in the composites are during the boreal spring following the strongest development of warm or cool tropical conditions. The negative SOI (warm Tropics) composite yields a SWE pattern similar to, but not as strong as that for the deep central North Pacific low phase of the CNP/PNA. This pattern (Fig. 9, bottom) exhibits greatest negative SWE anomalies in the northern Rocky Mountains, and a hint of positive SWE anomalies farther south in Utah and southern Colorado. The opposite behavior, but stronger, was found during the analogous cool phase of ENSO (Fig. 9, top). This result is consistent with the link between the central North Pacific Low (CNP or PNA) and snowpack and streamflow. It is well known that there is a tendency for the deep Aleutian low phase of the PNA to accompany the warm phase of ENSO (Dickson and Livezey 1984; Namias 1985). In contrast to the response to CNP (or PNA), however, the ENSO patterns contain SWE anomalies in the Southwest that are out of phase with the primary snow

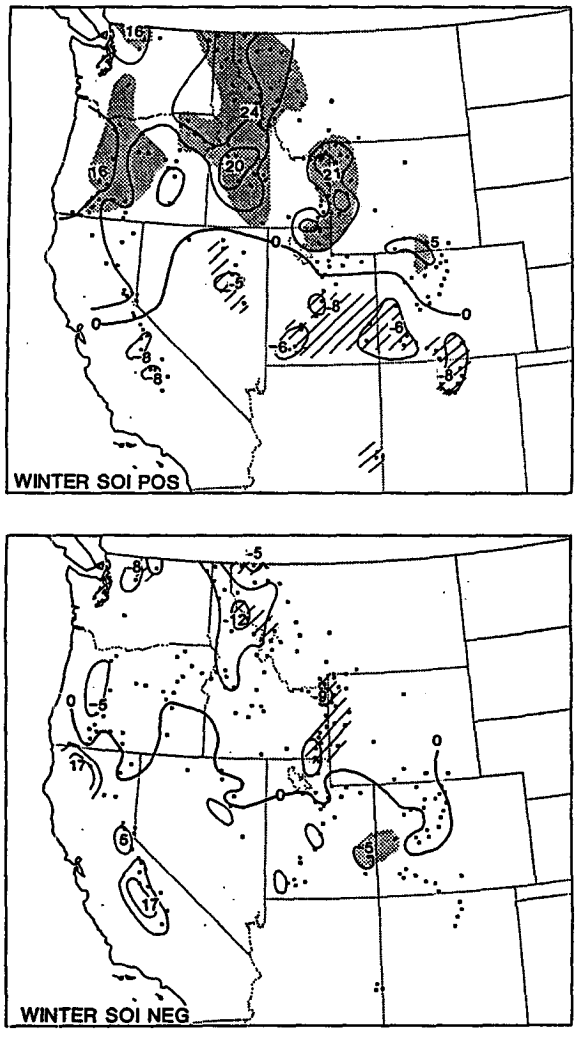


FIG. 9. Composite 1 April SWE (cm) following years with negative and positive SOI. Contours and shading as in Fig. 8.

anomaly signal in the Northwest, while the CNP patterns do not. This is consistent with evidence that the ENSO linkage is more complex than a single PNA response; Livezey and Mo (1987) and Cayan and Redmond (1994) have illustrated other ENSO-related circulation modes with a strong signal over the Southwest.

Is snow, being a hybrid of different climate effects, more or less sensitive to the circulation than either precipitation or temperature? This question was examined using a multiple regression scheme that predicted 1) winter precipitation, 2) winter temperature, and 3) 1 April SWE. Predictors were seasonal means (a) CNP and (b) SOI; fall, winter, and spring were provided as for each of the two predictor variables. Thus, separate regressions were developed that paired each of the three predictands with each of the two predictor variables. We selected the Idaho region for this test, since it contains at least a modest response to the Aleutian low and the ENSO phenomenon. Predictors chosen were admitted only if they contributed variance at the 95% confidence level, via an F test. While the results (explained variance is given in Table 5) are not particularly strong, they do imply that the snow accumulation is possibly more sensitive to large-scale fluctuations in atmospheric circulation than is precipitation and temperature. For the SOI, this seems quite conclusive, but for the CNP the winter precipitation appears to be almost as well predicted by CNP as is the spring SWE. Of course, this is only one region and the data may be somewhat suspect in that it is biased toward lower elevation sites, so this is only suggestive.

6. Summary and conclusions

Six decades of snow course records over the 11 western states of the conterminous United States provide a good description of interannual variability of mountain snow accumulation.

The snow course records reveal that 1 April snow water content (SWE) variations have regional spatial coherence. SWE anomaly patterns are produced by winter and spring precipitation and temperature anomalies, which are often considerably larger than individual mountain ranges or even statewide areas. A rotated principal components (RPC) analysis of the SWE data partitions 70% of the SWE variability into five regional patterns with centers in Idaho, California, New Mexico, Oregon, and Colorado. Among the most remarkable features of the SWE anomaly patterns is that the anomalies in the Northwest tend to lose their coherence at about the latitude of the Oregon–California border. Discontinuities appear between the Cascades and the Sierra Nevada, and also between the Northern and Southern Rocky Mountains. There is also a tendency for SWE anomalies in the Northwest to be out-of-phase with those in the mountains of the Southwest. West-to-east tracking of storms is evidenced by relatively high correlations (≥ 0.5) of SWE anomalies from the Pacific Coast across to the Rockies. Regions in the interior Rocky Mountain States (namely Wyoming, Colorado, and Utah) are affected by several different winter storm patterns; consequently, they exhibit rather weak connections to the seasonal mean atmospheric circulation and quite provincial spatial SWE correlations.

TABLE 5. Sensitivity of precipitation, temperature, and SWE to CNP and SOI circulation indices, Idaho region (via multiple linear regression, 53 cases, 1937–1989).

	Variance explained	
Predictand	CNP (F, W, Sp)	SOI (F, W, Sp)
winter temperature	31%	0%
winter precipitation	35%	6%
1 April SWE	45%	23%

The spatial coherence of the snow anomalies and other associated surface variables is driven by the atmospheric circulation, which encompasses regional to hemispheric anomaly scales over the North Pacific/North American sector. Perhaps the most remarkable circulation pattern associated with anomalous snow in the West is a quite repeatable configuration associated with low SWE at several of the different regions. This pattern resembles the well-known PNA pattern, and appears in both winter March circulation composites. The deep Aleutian low phase of the PNA circulation mode is predominantly dry over much of the West and inhibits cold air masses from settling into the region. The response of the snowpack is deficit SWE over much of the western United States, including the northern Sierra, the Cascades, and most of the Rocky Mountains. Another pattern occurs in association with the ENSO phenomenon. During springs following the mature warm tropical Pacific phase of ENSO, there are likely to be negative SWE anomalies in the northern Rocky Mountains, and possibly positive SWE anomalies farther south in Utah and southern Colorado, Arizona, and New Mexico. The opposite behavior, but stronger, was found during the analogous cool phase of ENSO. Thus, anomalies of SWE are quite sensitive to the large-scale atmospheric circulation; in fact, in the Pacific Northwest, snow appears to be more sensitive than precipitation and temperature to fluctuations in the PNA and the SOI.

For most regions, anomalous snow accumulation is affected most by fluctuations in precipitation-winter and early spring precipitation anomalies extend over broad regional scales. In composite, these patterns tend to be symmetric for positive and negative anomalies at the target snow catchments. For all regions, a strong response in spring or summer runoff was evident, with a delay of 1–2 months from the SWE anomalies.

How much does anomalous temperature control the variability of the snowpack? Previous studies (Walsh et al. 1982; Leathers and Robinson 1993) have suggested that the influence of temperature upon snow cover in the western United States is weak, but the snow course data suggest that this relationship may be weak or strong, depending on topography and season. In the cases examined here, the temperature responses detected were as expected, in the sense that cooler (warmer) temperatures produce positive (negative) SWE anomalies. Presumably, the temperature effect operates both during and after the snow is deposited: it determines the partition of snow versus rain that falls over the landscape during a given storm, and it determines the amount of snow that is melted after it is deposited. In regions with relatively high elevations, the temperature signals are modest, and are strongest in spring when mean temperatures are warmer. However, near the range of lower elevations close to the snow line, temperature fluctuations can have a profound effect upon snow accumulation—this is particularly evident in the coastal Pacific Northwest. In the Cascades of Oregon and Washington, where snow course elevations are generally lower than 2200 m, the amount of 1 April snowpack is strongly influenced by winter and spring temperatures, but is not so dependent upon precipitation.

Finally, regional composites of monthly temperature associated with anomalies of SWE suggest that the amount of snow accumulated in spring foreshadows the temperature departures in the next few months; for example, high 1 April SWE promotes lower temperatures in the region during April through June. This evidence is subtle, but it occurs in most of the five regions. It is not clear whether this is simply a seasonal persistence in the climate variability, or whether it represents a physical influence of snow on the regional temperature.

Acknowledgments. This work was supported by the University of California Water Resources Center, project UCAL-W-798, by the NOAA Climate Program Office through the Scripps Experimental Climate Prediction Center, and by the NOAA Climate and Global Change Program, Grant NA-16RC 0123-01. David Garen of the Soil Conservation Service provided the SCS snow course data over the western United States, and Frank Gehrke of the Cooperative Snow Surveys of the California Department of Water Resources provided snow course data from California snow courses. Thanks to Larry Riddle and Roger Shih for computations and graphics, and to Marguerette Shultz, Jean Seifert, and Nicki Pyles for illustrations and word processing. Mike Dettinger, Dave Peterson, Larry Riddle, Greg McCabe, Ed Aguado, and Roger Barry offered advice and three anonymous reviewers provided useful comments that improved the manuscript substantially.

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Effects of Temperature and Precipitation Variability on Snowpack Trends in the Western United States*

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(Manuscript received 4 March 2005, in final form 10 May 2005)

ABSTRACT

Recent studies have shown substantial declines in snow water equivalent (SWE) over much of the western United States in the last half century, as well as trends toward earlier spring snowmelt and peak spring streamflows. These trends are influenced both by interannual and decadal-scale climate variability, and also by temperature trends at longer time scales that are generally consistent with observations of global warming over the twentieth century. In this study, the linear trends in 1 April SWE over the western United States are examined, as simulated by the Variable Infiltration Capacity hydrologic model implemented at 1/8° latitude–longitude spatial resolution, and driven by a carefully quality controlled gridded daily precipitation and temperature dataset for the period 1915–2003. The long simulations of snowpack are used as surrogates for observations and are the basis for an analysis of regional trends in snowpack over the western United States and southern British Columbia, Canada. By isolating the trends due to temperature and precipitation in separate simulations, the influence of temperature and precipitation variability on the overall trends in SWE is evaluated. Downward trends in 1 April SWE over the western United States from 1916 to 2003 and 1947 to 2003, and for a time series constructed using two warm Pacific decadal oscillation (PDO) epochs concatenated together, are shown to be primarily due to widespread warming. These temperature-related trends are not well explained by decadal climate variability associated with the PDO. Trends in SWE associated with precipitation trends, however, are very different in different time periods and are apparently largely controlled by decadal variability rather than longer-term trends in climate.

1. Introduction and background

Snowpack is crucial to the water resources of the western United States (Serreze et al. 1999). A dominant fraction (by some accounts, more than 70%) of streamflow in the western United States originates as melting montane snowpack. Over most of the western United States, precipitation in the mountains is strongly winter dominant and snow accumulation (along with storage in the soil and groundwater) is the primary physical mechanism by which winter precipitation is stored and transferred to the relatively dry summers. For river basins with substantial snow accumulation, scenarios of increasing winter and spring temperatures in climate change assessments typically result in increased winter runoff, reduced peak water equivalent stored as snow, earlier peak streamflows, and reduced summer streamflow volumes (e.g., Cayan 1996; Dettinger et al. 2004;

* Joint Institute for the Study of the Atmosphere and Ocean Contribution Number 1066.

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Gleick and Chalecki 1999; Gleick 2000; Hamlet and Lettenmaier 1999b; Lettenmaier and Gan 1990; Leung et al. 2004; Knowles and Cayan 2002; Lettenmaier et al. 1999; McCabe and Wolock 1999; Mote et al. 2003; Service 2004).

Observational studies have also shown increasing trends in both temperature and precipitation over the western United States in the twentieth century (McCabe and Wolock 2002; Mote et al. 2003, Mote 2003a; Sheppard et al. 2002) and widespread trends toward earlier runoff timing in the western United States associated with earlier spring snowmelt (Dettinger and Cayan 1995; Stewart et al. 2005; Regonda et al. 2005). Mote (2003b) examined trends in 1 April snow water equivalent (SWE) for the U.S. Pacific Northwest (PNW), and Mote et al. (2005) have extended the earlier study to examine snow course records over the entire western United States and southern British Columbia (BC), Canada. These studies both show strong downward trends in 1 April SWE over most of the domain from about 1950 onward. Mote et al. (2005) also corroborated observed SWE trends by comparing simulated and observed SWE trends from 1950 to 1997 and used correlations between 1 April SWE and winter temperatures to show that temperature trends were a major driver of these observed trends, particularly at moderate elevations with relatively warm winter temperatures. The role of temperature and precipitation in determining snowpack variations was also examined by Serreze et al. (1999).

Observed data provide direct empirical evidence of trends in SWE over the western United States, but model simulations are needed to corroborate and extend these observations in order to avoid problems with limited spatial coverage, coarse temporal resolution, and longevity of observed data sources (see also Mote et al. 2005). Observations from snow course measurements (i.e., direct measurements of average SWE over a small transect), for example, have had a reasonably consistent level of coverage over the western United States for the period from about 1950 to present, but the number of observations declines rapidly prior to 1950, and coverage is very uneven before about 1940 (Mote et al. 2005). These changes in station density prevent a consistent analysis for a period of record longer than about a half century. This is important because the period of record, 1950–present, has been shown to contain substantial (and regionally specific) trends in precipitation and temperature that are strongly influenced by climatic variability associated with the El Niño–Southern Oscillation (ENSO) and the Pacific decadal oscillation (PDO; Dettinger et al. 1998; Gershunov and Barnett 1998; Hamlet and Lettenmaier 1999a; Hidalgo and Dracup 2003; Mantua et al. 1997; Mote et al. 2003; Sheppard et al. 2002). In particular, the precipitation trends that accompany the well-known shift in the PDO from cool phase to warm phase in 1976–77 (Mantua et al. 1997) are a confounding element in the attempt to attribute observed losses of SWE from 1950 to present to long-term changes in climate.

Another important issue is that observations of SWE are representative of only a small subset of the spatial domain, and relatively few observations are available at very high elevations or in the transient snow zone where winter snowpacks are ephemeral. These areas that are not well observed are a dominant fraction of the land area and of the SWE contributing to streamflow in the mountainous western United States and, in the case of the transient snow zone, they represent one of the most sensitive areas to warming. Thus a comprehensive understanding of long-term changes in western U.S. snowpacks and associated water resources impacts requires attention to not only a period of record longer than the last 50 yr, but to parts of the domain not covered by the snow course observations, the locations of which were dictated by short-term water supply forecasting needs, rather than detection of the impacts of climate variations.

In addition, snow course observations lack temporal resolution (they are typically reported only on the first of the month) and are unavailable before midwinter at most stations. Automated observations of SWE via the Natural Resources Conservation Service Snowpack Telemetry (SNOTEL) network are available at daily time steps throughout the snow season, but the longest records available for these observations go back only slightly more than 20 yr in all but a few isolated cases. Thus observations alone cannot provide useful information about long-term trends in patterns of snow accumulation and melt (such as date of peak accumulation, date of 90% melt, etc.).

While the use of model simulations introduces uncertainties into the analysis that are not present in observational studies, these uncertainties can be minimized if the model forcing data are carefully quality controlled, and the use of model simulations avoids a number of important limitations in the observed datasets discussed above. Here we examine simulations of spring SWE on 1 March, 1 April, and 1 May, with particular attention to the critical 1 April date, which corresponds roughly to the date of maximum snow accumulation over much of the mountainous western United States. By isolating the effects of temperature and precipitation trends in the model's forcing dataset in separate model runs, we also use the model to ex-

amine explicitly the effects of temperature and precipitation variability on SWE trends. The role of decadal-scale variability on SWE trends is also examined by selecting periods of record for analysis that reflect known shifts in decadal variability associated with the PDO (Mantua et al. 1997).

It is worth noting here that the trends in precipitation in the PNW appear to be consistently related to the PDO time series (i.e., dry to wet from 1925 to 1976 associated with warm to cool PDO; and wet to dry from 1947 to 2003 associated with cool to warm PDO; Hamlet and Lettenmaier 1999a; Mote et al. 2003; Mote 2003a), whereas a consistent relationship between precipitation trends and the PDO over time is not necessarily apparent in other parts of the domain, perhaps most notably in the Colorado River basin (Hidalgo and Dracup 2003).

The questions that motivate this paper are 1) To what extent are observed trends in western U.S. snowpack attributable to precipitation and temperature trends, and how do these effects vary with region, topography, climate, and the time period examined? 2) What areas of the western United States show the greatest trends in simulated snowpack, and what are the specific climatic regimes associated with these areas? 3) What is the role of decadal climate variability in determining the observed snowpack trends? 4) What changes in the timing of snow accumulation and melt are apparent and how do they vary with region, topography, and climate?

2. Hydrologic model and meteorological driving data

a. VIC hydrologic model

We use the Variable Infiltration Capacity (VIC) hydrologic simulation model (Liang et al. 1994; Cherkauer and Lettenmaier 2003) implemented at 1/8° (latitude–longitude) spatial resolution over the western US and southern BC. The 1/8° grid cells are roughly 12km by 10km, varying in size to some degree with latitude over the domain. Daily SWE for each grid cell (which represents the areal average SWE for the mosaic of elevation bands and vegetation types within the cell) is the primary output used in this study. SWE simulations from each elevation band were also used for comparison with point data from the snow course records, although these results are only briefly discussed here. We do not examine other variables from the simulations in this study, but the model also simulates a detailed water balance in each cell. Simulated daily runoff from the model produced by the runs in this study will be used in subsequent studies to examine changes in the water balance associated with changes in snow accumulation and melt. The VIC model has been used in numerous climate studies of large river basins around the world (e.g., Nijssen et al. 2001), and has been well validated with streamflow observations, particularly in the mountainous western U.S. (Christensen et al. 2004; Hamlet and Lettenmaier 2005, 1999b; Maurer et al. 2002; VanRheenen et al. 2004). The model has also been used extensively for streamflow forecasting applications (Hamlet and Lettenmaier 1999a; Hamlet et al. 2002; Wood et al. 2002) and for climate change assessments (Christensen et al. 2004; Hamlet and Lettenmaier 1999b; Payne et al. 2004; Snover et al. 2003; VanRheenen et al. 2004; Lettenmaier et al. 1999). For this study we primarily make use of the detailed energy balance snow model incorporated in the VIC model (Cherkauer and Lettenmaier 2003). The snow model is well suited for simulating mountain snowpack and includes the effects of forest canopy on snow interception and the attenuation of wind and solar radiation, which are fundamental drivers of snow accumulation, sublimation, and melt processes (Storck 2000; Storck et al. 2002).

b. Meteorological driving data

For this study, a new climate dataset was developed following methods outlined by Hamlet and Lettenmaier (2005). This dataset was intended specifically for application to problems where careful quality control to avoid spurious trends in long records of daily temperature and precipitation is necessary. Daily values of maximum temperature, minimum temperature, and total precipitation were obtained for 1915–2003 at the 1/8° latitude–longitude spatial resolution by interpolating daily values from National Weather Service Cooperative Observer stations using the Symap algorithm (Shepard 1984) following methods described in Maurer et al. (2002). Temperatures were lapsed by 6.1°C km^{-1} (i.e., the theoretical moist adiabatic lapse rate) during the interpolation process, and precipitation was scaled for topographic effects using monthly precipitation maps produced by the Precipitation Regression on Independent Slopes Method (PRISM) algorithm (Daly et al. 1994). Maurer et al. (2002) describe these data-processing steps in more detail. Temporal inhomogeneities (e.g., inconsistencies in time due to changing station groups, station moves, changes in instrumentation, urban heat island effects, etc.) in the regridded time series were adjusted for each calendar month using regridded and temporally smoothed data from the U.S. Historical Climatology Network and the monthly Historical Canadian Climate Database. Additional technical details are available in Hamlet and Lettenmaier (2005).

Despite the fact that the model forcing data are carefully quality controlled, there are inevitably some limitations associated with the approach. One is that the gridded (precipitation and temperature) forcing data are derived primarily from low or moderate elevation stations. Because the trends in the temperature and precipitation data are ultimately derived from the time series of these low elevation stations, there is no explicit information contained in the driving dataset regarding potentially different trends at very high elevations. In some parts of the domain, the primary driving data are also quite sparse, which may result in gridded data that may artificially suppress spatial variability, especially at short (e.g., daily) time steps. Topographic adjustments to the precipitation and temperature data (discussed above) likewise tend to suppress spatial variability in the gridded product. These limitations in the forcing data, while important for high-resolution modeling applications, are generally of secondary importance here. One major reason is that snow accumulation and ablation in mountainous regions is dominated by the characteristics of accumulated precipitation in snowpacks, and issues associated with suppression of spatial variability tend to be filtered out. In any event, the model is able to reproduce observed trends in spring snowpack quite well at a macroscale over the western United States (as discussed below); hence it meets the requirements for a study such as ours that focuses on macroscale aspects of the snow accumulation and ablation process.

c. *Role of temperature and precipitation driving data in the snow simulations*

The relationship between hourly snowfall and precipitation estimates in the model forcing data is fairly straightforward. The phase of the precipitation is determined by a simple partitioning scheme based on estimated hourly time step temperatures (derived from Tmax and Tmin) during the simulation. Temperatures below -0.5°C are assumed to result in precipitation that is 100% snow, those above 0.5°C are assumed to result in 100% rain, and a linear relationship is assumed between these two extremes.

Temperature driving data play a much more complex role in the simulations. Aside from the direct effects of temperature on convective heat transfer from the air to the snowpack (e.g., Storck 2000), temperature also affects snow accumulation and melt in the model due to meteorological variables that are derived from the temperature data. The difference between Tmax and Tmin determines the attenuation of solar radiation and variations in long-wave radiation due to cloudiness, for example, and Tmin is used in calculating the dewpoint and vapor pressure deficit which influence snow sublimation (Thornton and Running 1999). Thus trends in temperature are also indirectly related to trends in long- and shortwave radiation, humidity, and vapor pressure contributing to sublimation and snowmelt in the model. When we refer to "temperature-related trends" in later sections of the paper, we are referring to trends associated with these indirect effects as well.

d. *Evaluation of simulated SWE trends*

In corroborating trends in observations with the model results from this study, Mote et al. (2005) show remarkable broad-based agreement over the western United States between observed trends in 1 April SWE from 1950 to 1997 and trends derived from VIC simulations of SWE for the same period. Although some differences in the absolute value of relative trends are apparent, it is clear that the model successfully captures the large-scale characteristics of the spatiotemporal trends of snow accumulation and melt over the western United States. It is largely on the basis of this comparison and the longer streamflow comparisons discussed by Hamlet and Lettenmaier (2005) that we argue that the model results provide a reasonable surrogate for observations from 1916 to 2003 and are suitable for a trend analysis at these spatial scales.

It should be noted however that on a point-by-point basis, there are often substantial differences between the SWE observations analyzed by Mote et al. (2005) and the VIC simulations, both in absolute value and in relative trends. These differences between simulations and observations are primarily due to the fundamental differences in spatial scale between observations and simulations, and to frequent discrepancies between the actual precipitation and temperature at snow course sites (a point within the VIC cell) and the gridded meteorological driving data used in the VIC simulations. Snow courses are also generally located in open areas, which is inconsistent with the VIC simulations in parts of the domain with substantial forest canopy.

The issues associated with the model forcing data are to be distinguished from validation of the VIC snow model itself. When the model is driven by accurate temperature, precipitation, and vegetation characteristics (i.e., in locations where there is a nearby meteorological station and detailed information about the vegetation), the model very closely reproduces daily snow accumulation measurements recorded at these same locations (see, e.g., Storck 2000). Furthermore, additional analysis (not shown) found good general agreement between monthly VIC snow accumulation and melt statistics (e.g., date of peak snow accumulation, volume of accumulation and melt in each month, etc.) and those

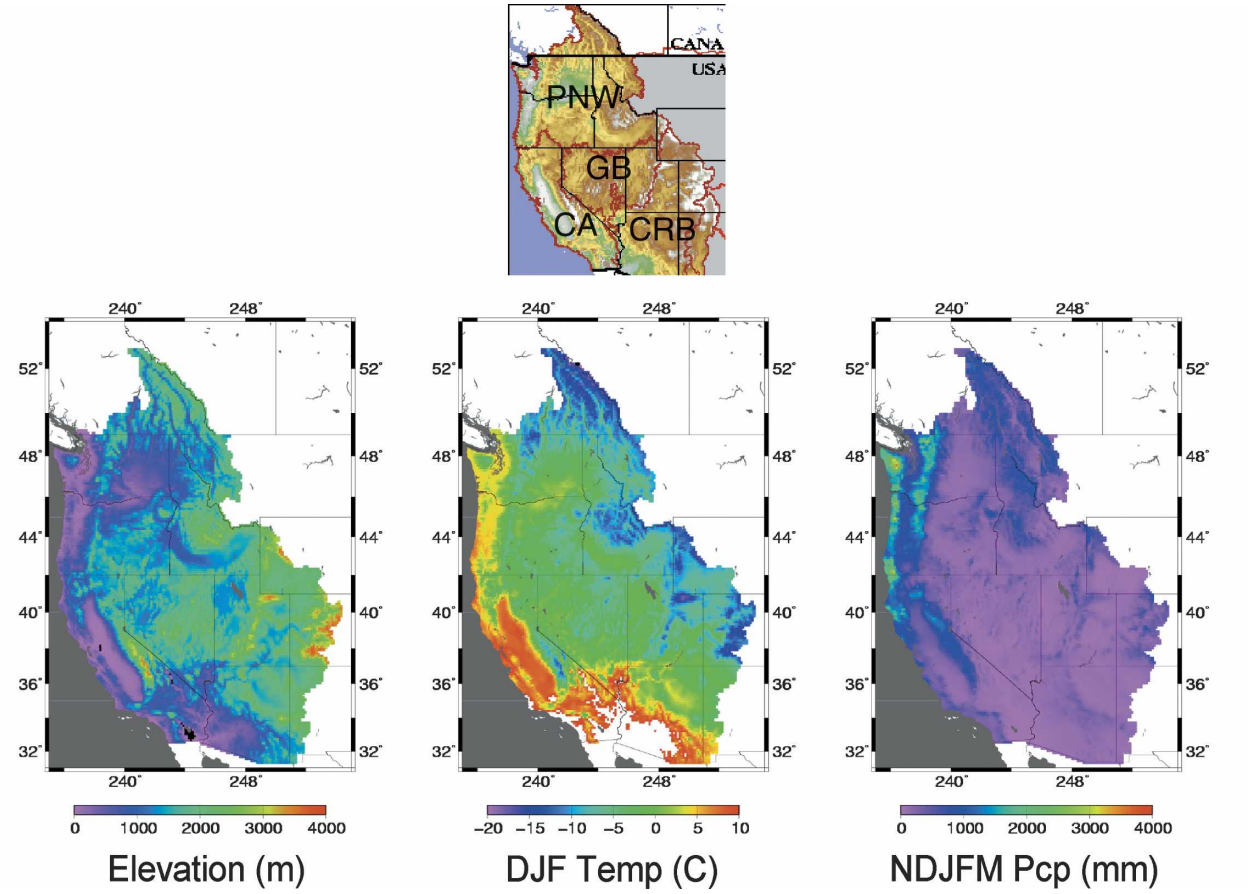


FIG. 1. Topographic and climatic characteristics of the 1/8° VIC hydrologic model simulation domain.

extracted from daily time step observations from SNOTEL stations. These comparisons with the SNOTEL observations help to confirm that the VIC model captures the primary physical mechanisms associated with winter climate and topography that ultimately determine trends in mountain snowpack.

Mote et al. (2005) also examined the model's ability to capture the relationship between winter temperature regimes and trends in 1 April SWE in the observations from 1950 to 2003. The model successfully reproduces this relationship over most of the western United States; however, in some specific parts of the domain the model does display some bias. There are, for example, some apparent systematic errors in the trends in high-elevation precipitation in the Sierra Nevada Mountains in the model forcing data that result in a general bias toward smaller simulated upward trends in the high-elevation areas in the southern part of the domain, and there are larger downward trends in simulated SWE at moderate elevations in California (CA) from 1950 to 1997 than are apparent in the observations. In the case of the lower-elevation sites in California, this problem might be partly attributable to sampling bias in the observations, or to inaccurate temperature lapse rates in spring when the weather is clear (i.e., the model forcing temperatures are probably biased toward warmer temperatures on clear days at high elevations because of the assumption of a fixed pseudoadiabatic lapse rate; see Hamlet and Lettenmaier 2005).

3. Methods and experimental design

The VIC model (version 4.0.5) was run from 1915 to 2003 at 1/8° resolution over the 16 526 grid cells that make up the continental United States west of the Continental Divide, plus the portion of the Columbia River basin that lies in southern British Columbia (see Fig. 1). In the following, we will refer to the PNW (containing the Columbia River basin and coastal drainages), California (CA), Colorado River basin (CRB), and Great Basin (GB) as subregions of this domain (see Fig. 1). The snow model time step was 1 h, which was required to capture the effects of the diurnal cycle of temperature and solar radiation on snow accumulation and

melt. In the absence of shorter time step observations, daily total precipitation was equally distributed throughout the 24 h of the day. To represent subgrid topographic variability, the model uses up to five equal-area elevation bands with a vertical interval of approximately 500 m. Simulated SWE values are reported as the area-weighted average SWE over all the elevation bands in each cell, each of which includes a vegetation mosaic comprised of up to 10 vegetation types. The vegetation characteristics of each cell were assumed to be stationary with time in the simulations. Note that the SWE values reported from the model represent the average snow water content over the entire grid cell, which can be quite different from the value that would be simulated for any particular point location within the cell. Elevation bands in each grid cell with July mean temperatures below 10°C were assumed to be above the tree line (Körner 1998), and vegetation classes with an overstory (if any) were removed from these bands during the simulation. The first nine months of the simulation (1 January 1915–30 September 1915) were used as model spinup and were discarded. This spinup is adequate for snow simulations, for which there is essentially no year-to-year carryover effect. Figure 1 shows the 1/8° digital elevation model for the simulation domain, average December–February (DJF) temperatures for each grid cell, and average November–March precipitation summarizing the model forcing data.

Three different model runs were made:

1) Base run: Unperturbed precipitation and temperature data from the meteorological input files (daily total precipitation and minimum and maximum temperature and wind speed) were used to drive the model. Trends in SWE are the result of both temperature and precipitation variations.
2) Fixed precipitation (fixed *P*) run: The precipitation forcing data were fixed (for each calendar month) at the climatological value for each grid cell, but daily temperature was allowed to vary as in the original time series. Trends in SWE are the result of trends in temperature and other temperature-related meteorological variables alone.
3) Fixed temperature (fixed *T*) run: The temperature forcing data were fixed (for each calendar month) at the climatological value for each grid cell, but daily precipitation varied as in the original time series. Trends in SWE are the result of precipitation trends alone.

For the fixed *P* and fixed *T* runs, the forcing data were perturbed as follows: For each grid cell, a monthly climatological value for either precipitation or temperature was calculated for each calendar month (i.e., a separate climatology for January, February, etc.). Then the daily time series of the variable to be held constant was forced to reproduce this climatological value in each month of the simulation—that is, the daily values were allowed to vary within the month as in the original time series, but the monthly totals (precipitation) or averages (maximum and minimum temperature) in each calendar month are identical in each year of the simulation. This method preserves the daily covariance between temperature, precipitation, solar radiation, etc., while removing the trends and monthly variation in the fixed variable from the simulation. Note that an average seasonal cycle remains in the fixed variable after this adjustment. Wind data were allowed to vary as in the original time series in each case. If precipitation and minimum and maximum temperature were all held fixed as described above, the mean of the resulting simulated SWE would be comparable (but not equal) to the long-term mean for each calendar date in the base simulation, and any trends in the simulated SWE would be small. It should be noted that the *timing* of precipitation within the month is unaffected by these adjustments. This has some important implications when evaluating the fixed *P* trends in the calendar date of 10% snow accumulation discussed in later sections, for example.

The 1 March, 1 April, and 1 May SWE values were extracted from the model output files for four time periods based on warm and cool PDO epochs defined by Mantua et al. (1997) as follows:

A) 1916–2003 (full period)
B) 1925–76 (warm PDO to cool PDO)
C) 1947–2003 (cool PDO to warm PDO)
D) 1925–46 concatenated with 1977–2003 (warm PDO to warm PDO)

Linear trends were calculated for the time series for each grid cell and were normalized by the long-term mean of the simulated 1 April SWE for each cell over the time period of analysis (i.e., the trends for scenario C were normalized by the mean SWE calculated from the 1947–2003 data). Cells included in the plots were required, on average, to have at least 50 mm of SWE on 1 April in order to avoid including spurious trends in the analysis.

For each grid cell, a time series of the date of peak SWE [day 1 =1 September, day 365 (or 366 in leap years) = 31 August] was extracted from the daily time series of simulated SWE, as well as the day associated with 10% of peak accumulation and 90% melt. Linear trends were calculated for each of these variables from 1916 to 2003 for the base, fixed *P*, and fixed *T* simulations.

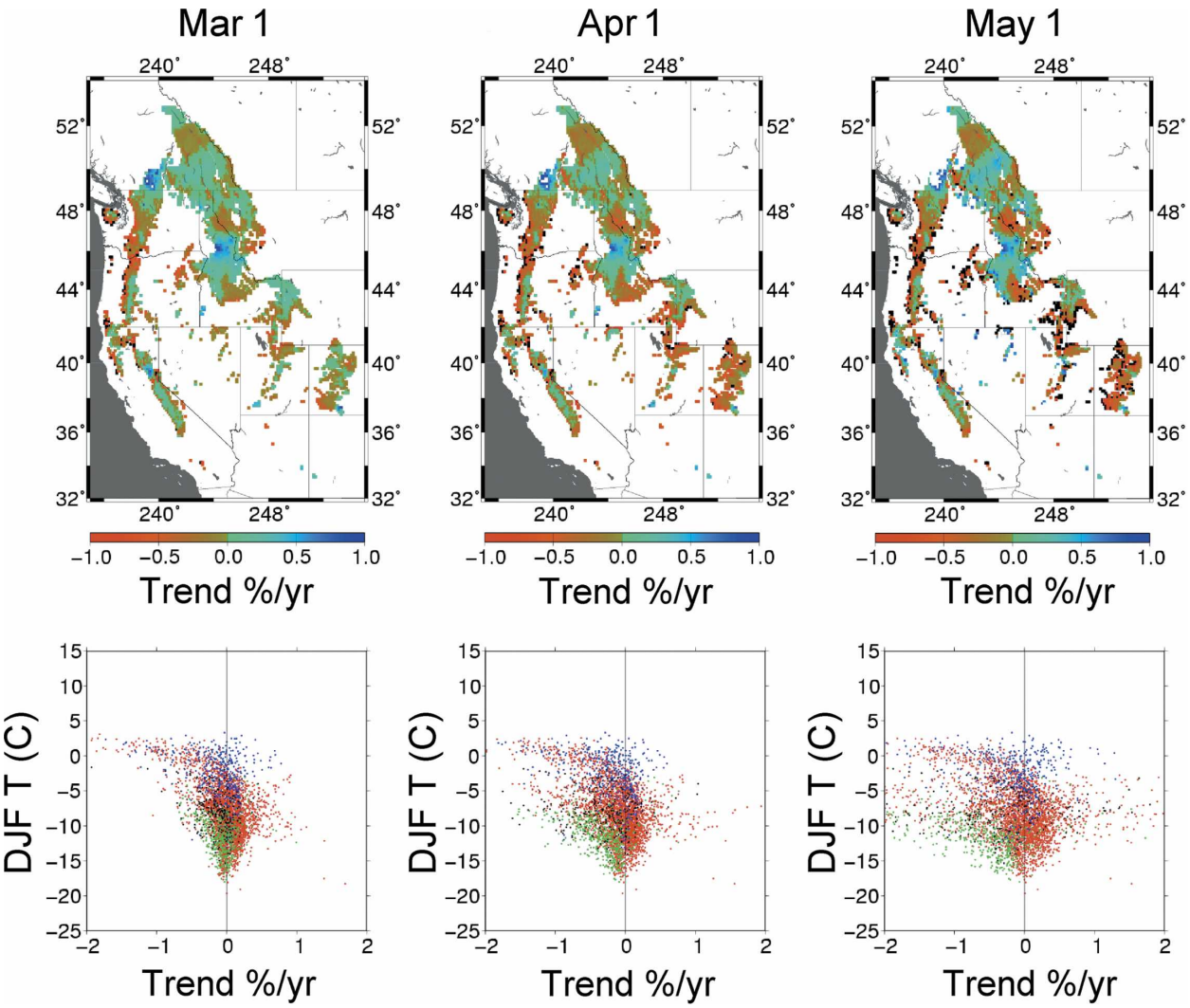


FIG. 2. Relative trends (% yr^{-1}) in simulated SWE for three calendar dates for the period from 1916 to 2003. (Black pixels in the spatial plots are off scale on the low side. Scatterplots are color coded: PNW is red, CA is blue, CRB is green, and GB is black.)

4. Results and discussion

a. Notes on presentation of results and interpretation of figures

The results in this section will be summarized primarily using spatial plots of the trends in SWE in each grid cell and scatterplots showing the relationship between average midwinter (DJF) temperatures in each cell and the trends in SWE. The dots in the scatterplots are color coded so that the different regions in the domain can be distinguished. By comparing the patterns in these figures for the base, fixed *P*, and fixed *T* runs, qualitative conclusions about the effect of temperature and precipitation trends on the overall trends can be drawn. DJF average temperatures for each grid cell used in the scatterplots of trend versus winter temperature were calculated from the VIC driving data from 1916 to 2003 and are independent of the period of analysis used.

b. SWE trends for 1 March, 1 April, and 1 May for 1916–2003

To begin with, we show overall trends in simulated SWE for 1 March, 1 April, and 1 May for the base run from 1916 to 2003 (Fig. 2). As noted above, only grid cells with average SWE greater than 50 mm on 1 April are shown. The results show downward trends over many grid cells in the domain, and a relationship between midwinter temperatures and the relative trends in SWE characterized by an inverted "J" shape in the scatterplots. As we shall see in subsequent sections, this relationship is characteristic of downward trends in

snowpack associated with warming. Downward trends in SWE on 1 April and 1 May are more widespread than for 1 March. A substantial part of the domain shows upward trends in SWE, particularly over much of the Columbia River basin. As we shall see in subsequent sections, these upward trends in SWE are primarily due to upward trends in precipitation. The increased scatter for 1 May that is evident in Fig. 2 is probably due to the fact that many grid cells have relatively small amounts of snow remaining on 1 May.

c. Results for 1916–2003

This period is characterized by widespread, modestly upward trends in precipitation that result in upward trends in SWE in the fixed *T* simulations (Fig. 3c) and strong downward trends in SWE due to upward temperature trends over essentially the entire domain in the fixed *P* simulations (Fig. 3b). Thus the majority of the downward trends in SWE from 1916 to 2003 are attributable to large-scale warming, which overwhelms the effects of widespread increases in winter precipitation. Note, for example, the predominantly upward trends in grid cells in CA associated with precipitation (Fig. 3c), but predominantly downward trends for CA in warmer areas with DJF temperatures above about −2.5°C (Fig. 3a). Several distinct climatic regimes are also apparent in these results, which will be discussed separately below.

d. Results for 1924–76

Strong upward trends in precipitation throughout the region dominate the SWE trends in this period (Fig. 4c), and upward trends in SWE follow these precipitation trends over most of the domain (Fig. 4a). The overall trends in precipitation appear to reflect a shift from widespread drought from 1925 to 1946 to wetter conditions overall in the second half of the period. The fixed *P* analysis (Fig. 4b) shows that there is not a consistent trend in SWE due to temperature alone, and this is the only period examined for which there were substantial upward trends in SWE associated with downward trends in temperature (and solar radiation) alone.

e. Results for 1947–2003

The 1947–2003 period of analysis corresponds most closely with the time periods covered by the snow course and streamflow records that were discussed in the introduction. This period is characterized by different precipitation trends in different regions (Fig. 5c) and by strong downward trends in SWE associated with temperature trends (Fig. 5b). The trends in precipitation in each region are broadly consistent with patterns of observed climate variability associated with the PDO since 1947 that were discussed in the introduction. The fixed *P* analysis for this period (Fig. 5b) shows that without precipitation trends, essentially the entire domain would have experienced strong downward trends in 1 April SWE. Because the temperature-related trends are almost always downward for this period, areas with downward precipitation trends (e.g., most of the PNW cells with DJF temperatures above −5°C) exhibit very strong downward trends in SWE. Some areas with upward precipitation trends (e.g., about half the cells in CA) show downward SWE trends due to temperature effects despite increases in precipitation. This effect is less evident in the colder areas of the domain (e.g., the CRB) but can still be seen in the base run (Fig. 5a) as a shift toward stronger downward trends in SWE when the effects of temperature trends are combined with the effects of precipitation trends. Trends in the base run analysis (Fig. 5a) are less than −0.25% per year (strong downward trends) over about 70% of the cells shown in the figures. This is a considerably larger fraction of the total area with downward SWE trends than for the base run for 1916–2003, in part because of widespread downward precipitation trends in the colder areas of the PNW from 1947 to 2003.

f. Results for 1924–46 concatenated with 1977–2003

The intent here is to minimize the effects of decadal variability associated with the PDO by combining two climatologically similar epochs. The fixed *P* analysis for this period (Fig. 6b) shows somewhat more scatter than the fixed *P* analysis for 1947–2003, but again the trends associated with temperature alone are overwhelmingly downward. The fixed *T* analysis (Fig. 6c) shows little consistent trend in SWE associated with precipitation trends in comparison with the 1916–2003 or 1925–76 periods. As in the 1916–2003 and 1947–2003 periods, a major driver of the SWE trends in this period is a large-scale warming that affects most of the domain, and the analysis suggests that this warming cannot be readily explained by decadal-scale variability associated with the PDO. In particular, the earlier warm phase PDO epoch was clearly much cooler overall than the most recent warm phase epoch. Further evidence that decadal variability is a poor explanation for the temperature-related trends can be seen by comparing Figs. 3b, 5b, and 6b. It is apparent that any period paired with the 1977–2003 period shows dramatic downward trends in SWE due to temperature trends alone.

g. Effects of climatic regimes

Three distinct climatic regimes are apparent in the areas that experienced downward trends in 1 April

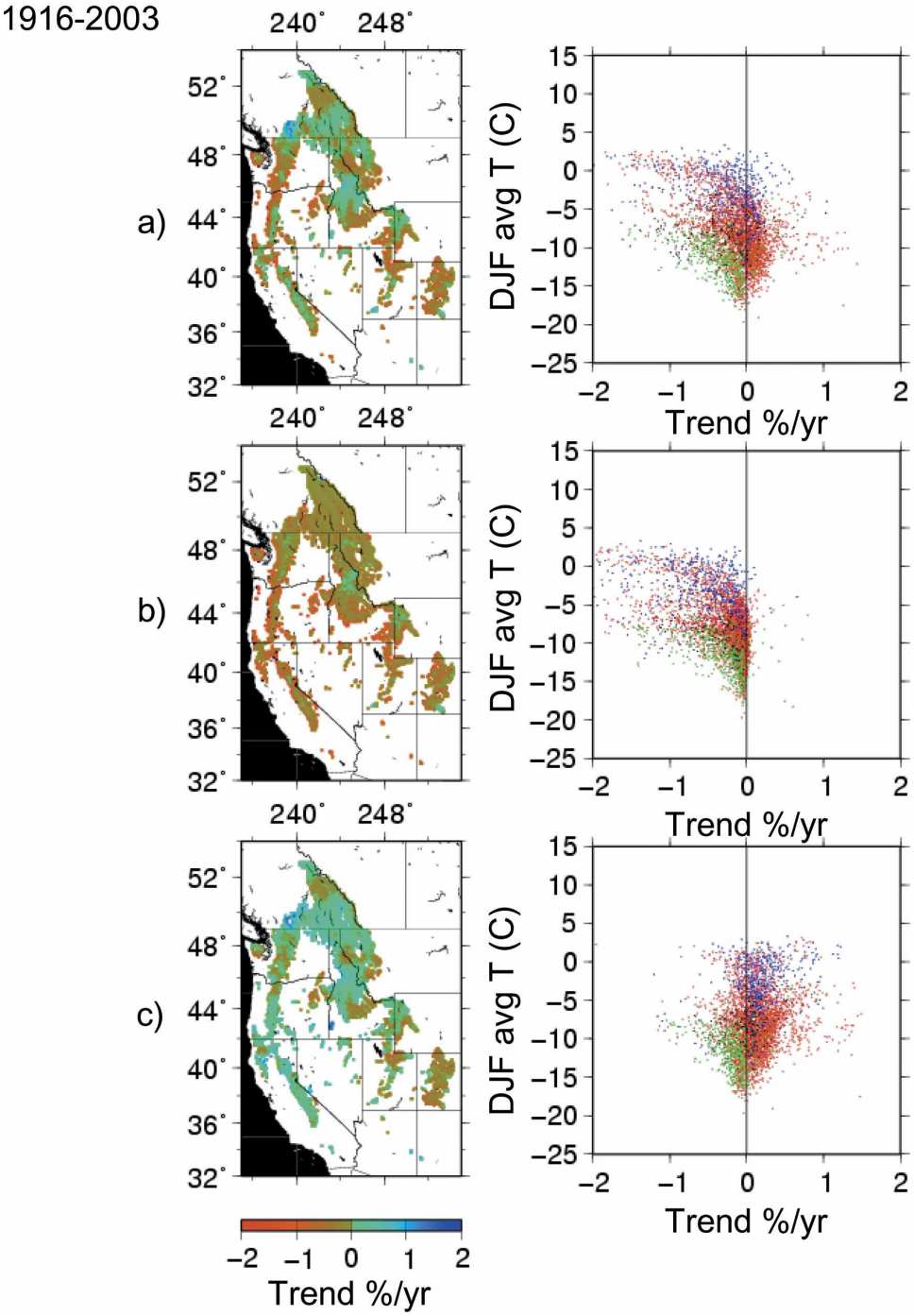


FIG. 3. Same as in Fig. 2, but in simulated 1 Apr SWE for 1916–2003: (a) combined effects of temperature and precipitation trends, (b) effects of temperature trends alone, and (c) effects of precipitation trends alone.

SWE from 1916 to 2003 (Fig. 7). Areas with large downward trends in 1 April SWE and DJF temperatures between −1.0° and +2.5°C are shown to be coastal areas in the PNW and northern CA (Fig. 7, top right). Areas with large downward trends in 1 April SWE and DJF temperatures less than −1.0°C are shown to be inland areas with a more continental climate (Fig. 7, middle right and lower right). These differences have to do with specific interactions between precipitation and temperature regimes that determine

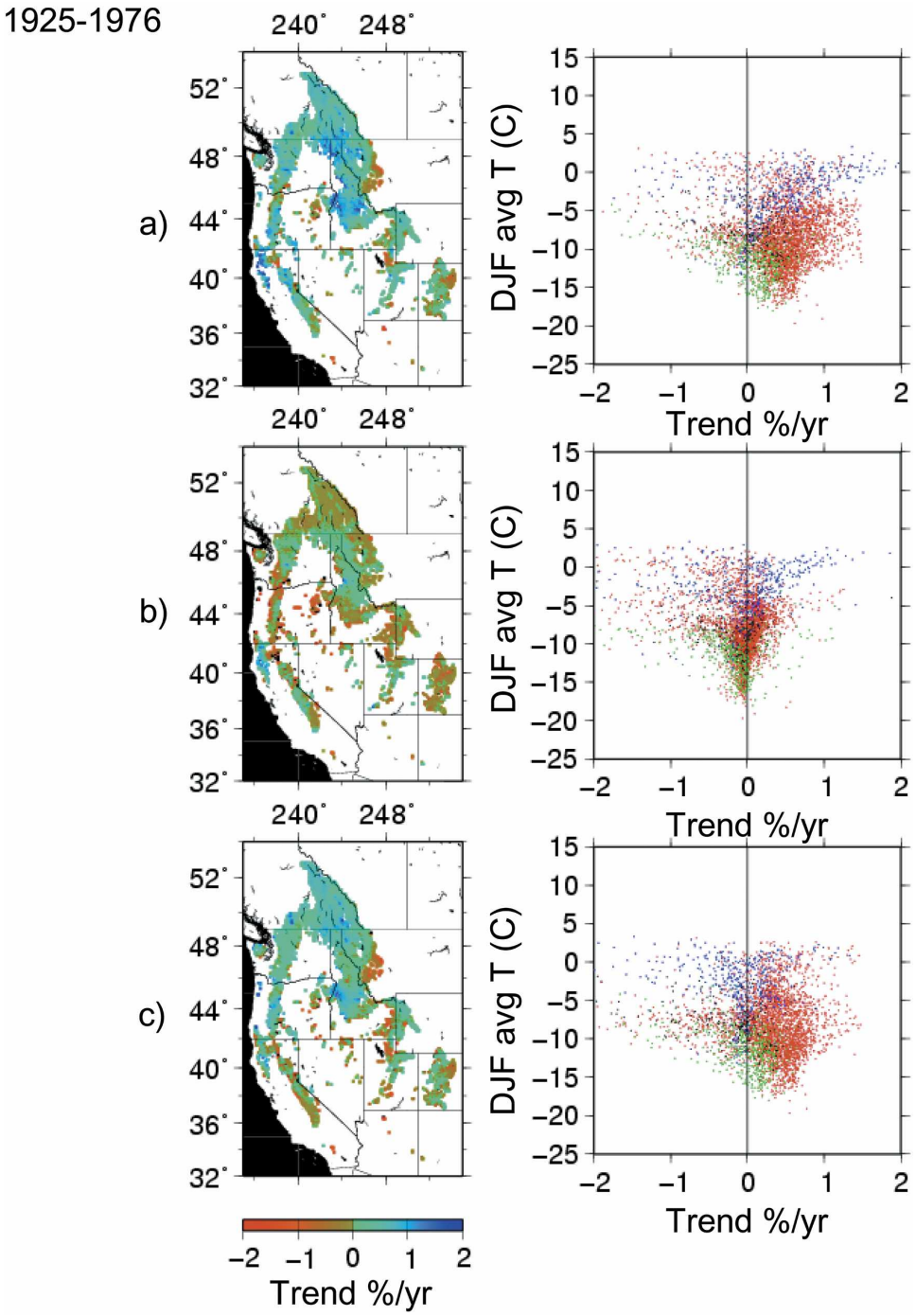


FIG. 4. Same as in Fig. 3, but for 1925–76.

the time scales of snow accumulation and melt. Coastal mountain ranges are warmer but are able to produce very large snowpacks in midwinter that can persist until 1 April because of high mean precipitation and cool, cloudy spring conditions. These areas are affected by temperature increases from midwinter through spring (which also explains why they are most sensitive to warming). Continental areas, by contrast, are typically dryer and colder, and snow accumulation that ultimately persists until 1 April takes place over a number of months from late fall to early spring. Analysis of winter precipitation and the seasonal cycle of snow ac-

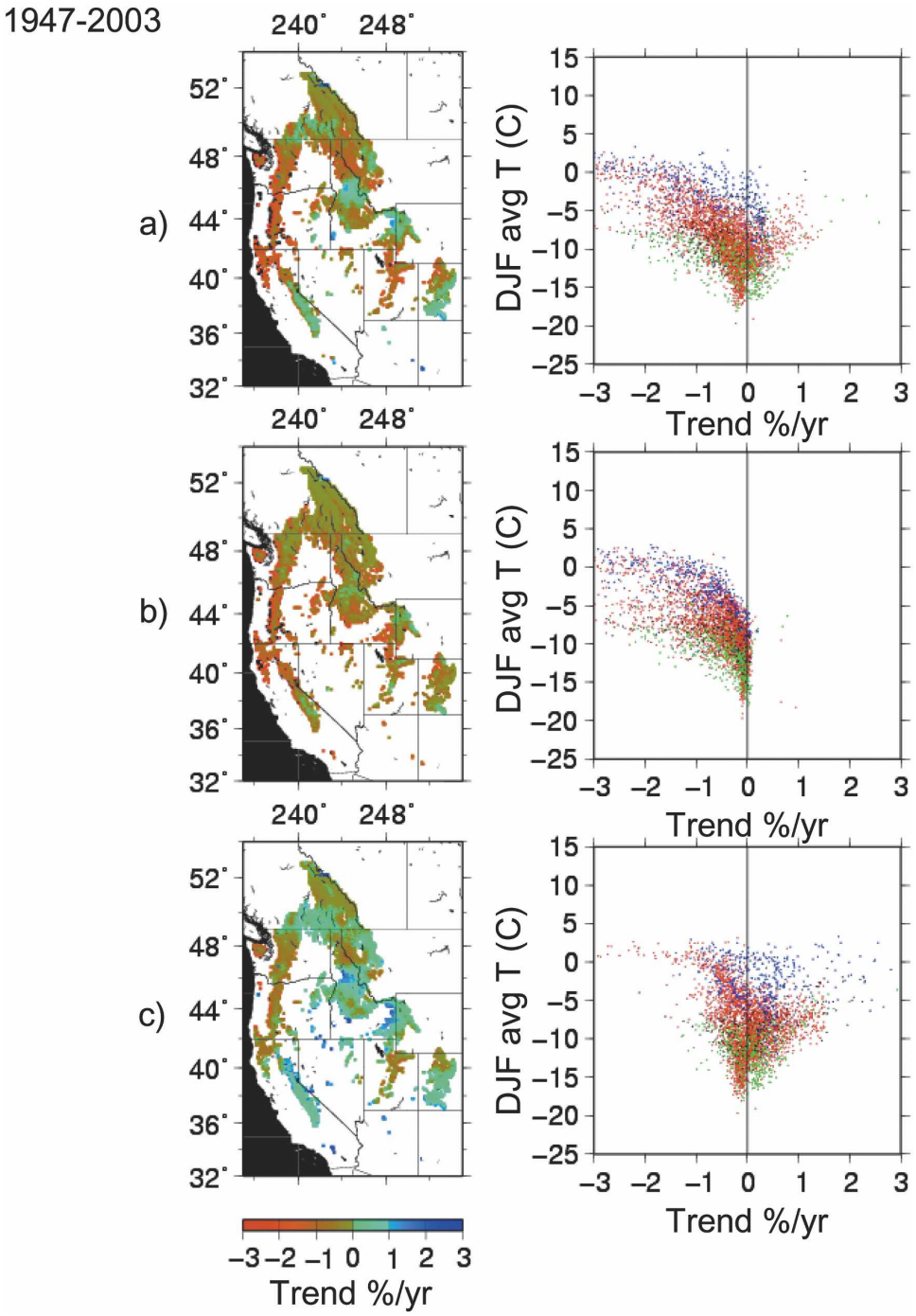


FIG. 5. Same as in Fig. 3, but for 1947–2003.

cumulation in these areas (not presented here) showed that precipitation trends over the entire winter and warmer temperatures during the late spring were the predominant drivers of the overall trends in SWE. For areas with cold midwinter temperatures (Fig. 7 lower right), the temperature sensitivity becomes relatively small and the large downward trends in SWE are associated primarily with strong downward trends in winter precipitation (cf. Fig. 7 with Fig. 3c). These climatic regimes are consistent with climatic characteristics observed by Serreze et al. (1999) in SNOTEL observations.

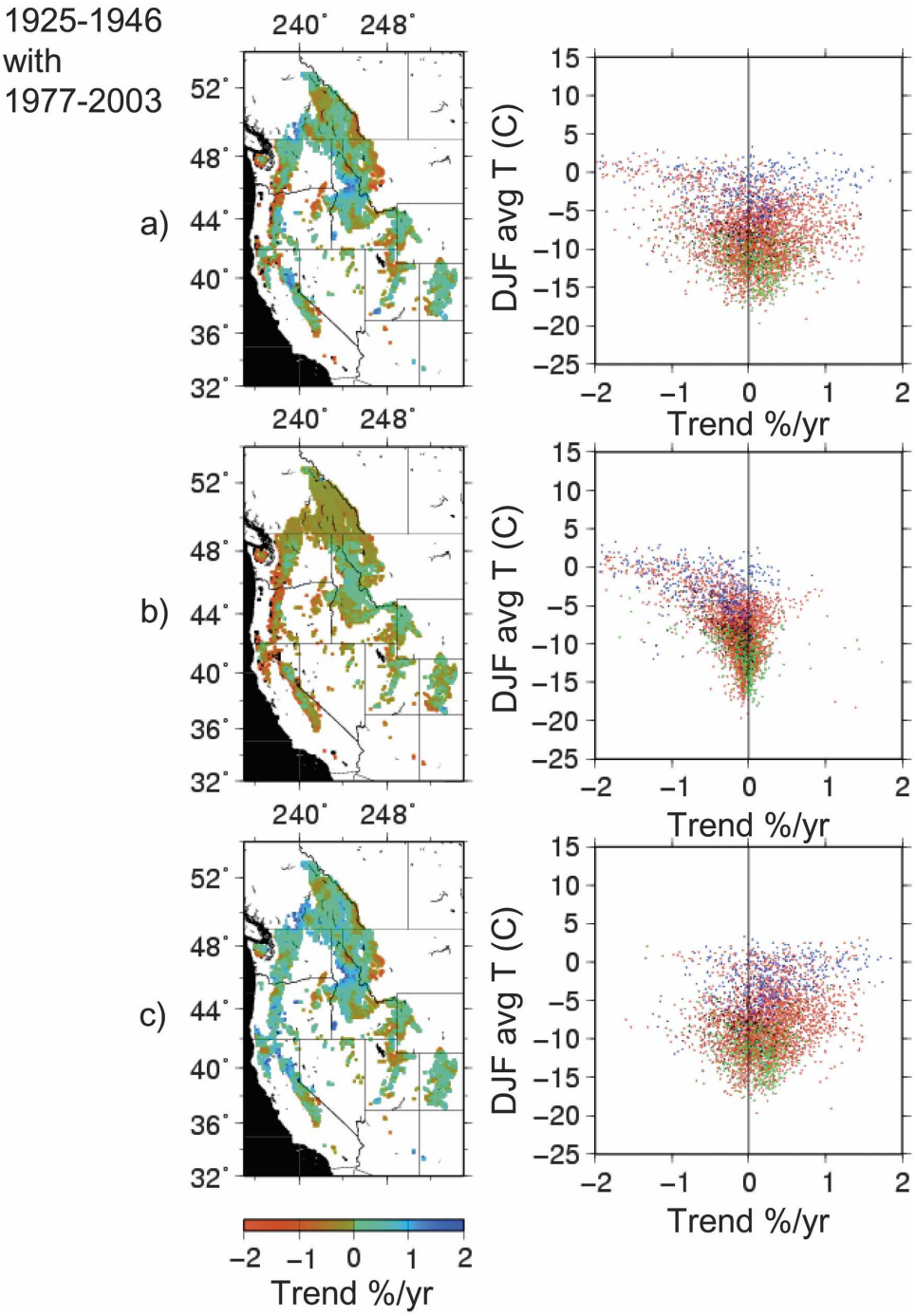


FIG. 6. Same as in Fig. 3, but for 1925–46 combined with 1977–2003.

h. Trends in the timing of snow accumulation and melt

Figure 8 shows the trends in the date of simulated snow accumulation and melt. Trends in the date of simulated peak snow accumulation and the date of 90% melt are strongly related to DJF average temperatures. Almost all the cells with winter temperatures warmer than −5°C, for example, show earlier dates of peak accumulation and 90% melt. In sensitive areas (such as the coastal areas in the PNW and CA shown in Fig. 7), peak accumulation occurs between 15 and 45 days earlier, and 90% melt occurs 15–40 days earlier. Comparison of the base run scatterplots with those from the

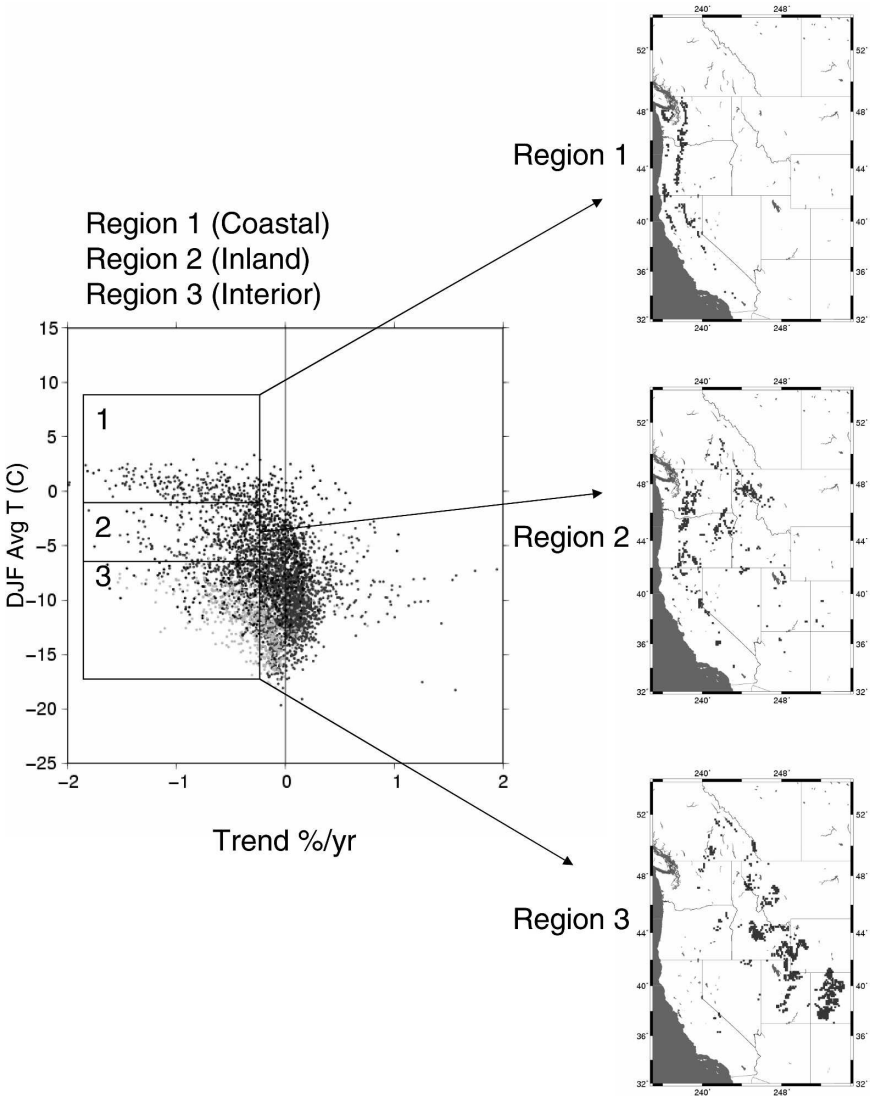


FIG. 7. Location of winter climatic regimes associated with strong downward SWE trends (<0.25% yr^{-1}) from 1916 to 2003. Midwinter (DJF) temperatures in region 1 are above −1°C, in region 2 are between −1° and −7°C, and in region 3 are below −7°C.

fixed *P* and fixed *T* runs (Fig. 8) shows that the changes in the timing of peak snow accumulation and 90% melt are a complex function of precipitation and temperature changes, but that the dominant effect is due to temperature trends. These changes in snow accumulation and melt are consistent with observed changes in streamflow timing discussed in the introduction. A similar examination of the scatterplots shows that trends in the date of 10% accumulation are predominantly due to upward trends in precipitation in the beginning of the snow accumulation season. Interestingly, the time between 10% accumulation and 90% melt has not changed very much overall. This somewhat counterintuitive result occurs because the 10% accumulation dates are most affected by precipitation trends (cf. base and fixed *T* runs in Fig. 8), whereas the peak SWE and 90% melt dates are most affected by temperature trends (cf. base and fixed *P* runs in Fig. 8). Despite different mechanisms at different times of the year, the overall trends in the date of snow accumulation and melt are characterized by shifts in the entire snow accumulation season earlier in time, but with a smaller amount of peak SWE.

The results for the four time periods are summarized for different portions of the domain and the different model runs in Tables 1 and 2. Table 1 shows the areal average of the relative trends calculated for each cell (i.e., the spatial average of the trends shown in Figs.

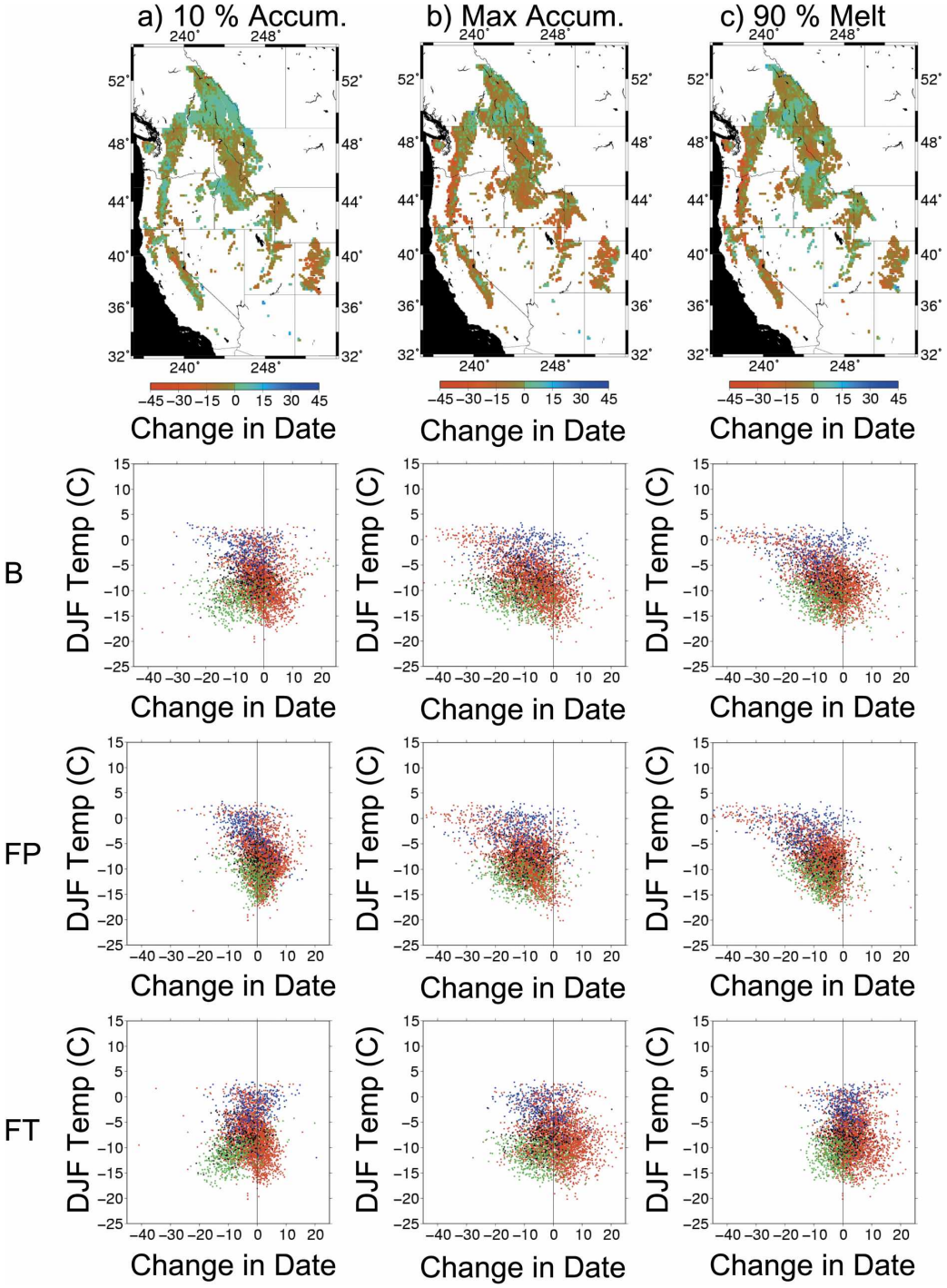


FIG. 8. Cumulative changes (trend × number of years) in calendar date of (a) 10% SWE accumulation, (b) maximum SWE accumulation, and (c) 90% melt for 1916–2003 for three model runs. (top) Spatial plot for the base run, and (bottom three) scatterplots for the base (B), fixed *P* (FP), and fixed *T* (FT) runs, respectively.

3–6). Table 2 shows the trend in the areal average SWE (i.e., a time series of SWE averaged over each domain is extracted first, and the trend in this aggregate time series is reported).

The results for 1925–76 shown in Table 2 are somewhat confusing and require some additional explanation. For CA, for example, the fixed *P* and fixed *T* runs for this period show small downward trends in areal

TABLE 1. Areal averages of the relative trends for each grid cell calculated for each portion of the domain (% yr^{-1}).

	PNW	CA	GB	CRB
1916–2003				
Base	−0.14	−0.21	−0.40	−0.69
Fixed *P*	−0.60	−0.91	−0.95	−0.71
Fixed *T*	0.40	0.44	0.40	−0.22
1925–76				
Base	1.52	1.05	1.47	0.74
Fixed *P*	0.73	0.29	1.33	0.68
Fixed *T*	0.67	−0.05	0.37	0.36
1947–2003				
Base	−0.39	−0.84	0.46	0.39
Fixed *P*	−0.94	−1.61	−0.72	−0.35
Fixed *T*	0.45	0.42	0.63	0.26
1925–46 with 1977–2003				
Base	0.53	−0.35	0.09	−0.22
Fixed *P*	−0.36	−1.75	−0.62	−0.80
Fixed *T*	0.98	0.54	0.75	0.43

average SWE, yet the trend in the base run has a strong upward trend. These counterintuitive effects are associated with the spatial distribution of the upward and downward trends associated with temperature and precipitation trends. For example, if specific areas in the domain that are getting wetter are also getting colder, then the overall trend may be strongly upward (as for CA for this period), whereas if areas that are getting wetter are also getting warmer there may be little overall trend. These complex effects are not seen in the other time periods because the temperature-related trends (and in many cases, the precipitation related trends as well) are much more spatially homogeneous.

5. Summary and conclusions

Widespread warming has occurred in the western United States from 1916 to 2003, resulting in downward trends in 1 April SWE over large areas of the domain. High-elevation areas with upward precipitation trends, however, are shown to produce upward trends in SWE over the same time period. The results show that almost all the upward trends in SWE from 1916 to 2003 are due to modest upward precipitation trends and that many of the downward trends in SWE are caused by widespread warming. In areas with relatively warm winter temperatures, such as coastal areas of the PNW and CA, the effects of warming frequently overwhelm the effects of increasing precipitation. Colder areas are predominantly driven by precipitation changes, and temperature effects on SWE trends, while apparent in the combined trends, are relatively small.

TABLE 2. Relative trend in the areal average SWE for each portion of the domain (% yr^{-1}).

	PNW	CA	GB	CRB
1916–2003				
Base	−0.06	−0.13	−0.25	−0.26
Fixed *P*	−0.18	−0.39	−0.40	−0.22
Fixed *T*	0.10	0.14	0.04	−0.14
1925–76				
Base	0.48	0.40	0.12	0.12
Fixed *P*	−0.06	−0.04	−0.13	−0.12
Fixed *T*	0.38	−0.03	−0.14	0.03
1947–2003				
Base	−0.40	−0.51	−0.51	−0.19
Fixed *P*	−0.34	−0.73	−0.60	−0.30
Fixed *T*	−0.03	0.21	0.34	0.13
1925–46 with 1977–2003				
Base	0.07	−0.04	−0.28	−0.29
Fixed *P*	−0.26	−0.57	−0.55	−0.36
Fixed *T*	0.27	0.24	0.03	−0.10

The period from 1925 to 1976 was characterized by widespread increases in precipitation, which dominated temperature trends to produce widespread upward trends in SWE. The period from 1925 to 1976 was also the only period in which widespread upward trends in SWE are observed due to temperature trends alone. From 1947 to 2003, regionally specific precipitation trends associated with decadal variability combine with strong downward trends in SWE due to widespread warming. The effects of warming are frequently large enough to overwhelm upward trends in SWE associated with precipitation trends alone in warmer areas of the domain.

The results for the back-to-back warm PDO epochs combined with the 1916–2003 and 1947–2003 analysis described above strongly suggest that widespread regional warming trends are not well explained by decadal-scale variability, but that the decadal variability probably does account for the trends in winter precipitation that have been apparent over shorter periods of the record. Although the precipitation trends from 1916 to 2003 are broadly consistent with many global-warming scenarios, it is not clear whether the modestly increasing trends in precipitation that have been observed over the western United States for this period are primarily an artifact of decadal variability and the time period examined, or are due to longer-term effects such as global warming.

Several distinct climatic regimes exist in areas that have experienced downward trends in 1 April SWE in the simulations. Coastal areas are strongly affected by warming throughout the winter and spring, whereas areas with a more continental climate are more sensitive to precipitation trends during the winter and to warming in late spring. Many high-elevation areas in the Rockies and southern Sierra Nevada are relatively insensitive to temperature trends, and downward trends in SWE are primarily due to downward trends in precipitation.

The dates of peak snow accumulation and 90% (of peak) melt have generally been occurring earlier in the year, and these trends are clearly sensitive to winter temperature regimes, with the greatest changes apparent in areas with warmer winter temperatures (e.g., near-coastal mountains in the PNW and CA). These effects are consistent with the observed trends toward earlier peak snowmelt runoff. The sensitivity analysis shows that the changes in the timing of peak accumulation and 90% melt are primarily a temperature-related effect. The date of 10% (of peak) accumulation has also trended earlier in the year but is shown to be related primarily to trends in fall precipitation rather than to trends in temperature.

The results of this study strongly suggest that the effects of global warming are apparent in the surrogate observations examined in this study. Furthermore, because the trends in SWE are most strongly affected by trends in temperature, we should expect, based on projections of continued warming, that these trends will continue. Because precipitation variability seems most strongly associated with decadal variability rather than long-term trends, the use of observed precipitation variability in conjunction with scenarios of warmer temperatures may currently be the best approach for understanding the overall effects of global warming on the hydrologic variability of the western United States. Such an approach implies, as well, that both "warm and wet" and "warm and dry" periods are likely to occur in the future at different times, and that water resources planning should consider both scenarios in testing alternative management plans.

Acknowledgments. This publication was funded by the Joint Institute for the Study of the Atmosphere and Ocean (JISAO) under NOAA Cooperative Agreement NA17RJ1232.

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Climate-Driven Variability and Trends in Mountain Snowpack in Western North America*

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(Manuscript received 21 November 2005, in final form 17 February 2006)

ABSTRACT

Records of 1 April snow water equivalent (SWE) are examined here using multiple linear regression against reference time series of temperature and precipitation. This method permits 1) an examination of the separate roles of temperature and precipitation in determining the trends in SWE; 2) an estimation of the sensitivity of SWE to warming trends, and its distribution across western North America and as a function of elevation; and 3) inferences about responses of SWE to future warming. These results emphasize the sensitivity to warming of the mountains of northern California and the Cascades of Oregon and Washington. In addition, the contribution of modes of Pacific climate variability is examined and found to be responsible for about 10%–60% of the trends in SWE, depending on the period of record and climate index.

1. Introduction

In most river basins of the West, especially in Washington, Oregon, and California, snow (rather than man-made reservoirs) is the largest component of water storage; hence, the West is (to varying degrees) vulnerable to climatic variations and changes that influence spring snowpack.

Since the midtwentieth century, as numerous studies have documented, important changes have occurred in hydrological properties in the West. In most of the West, these changes include trends toward earlier snowmelt-driven streamflow (Cayan et al. 2001; Stewart et al. 2005) that are more pronounced at lower elevation (Regonda et al. 2005), reduced flow in June and increased flow in March (Stewart et al. 2005), reduced spring snow water equivalent (SWE) (Mote 2003a; Mote et al. 2005), and reduced frequency of low-elevation snowfall (Groisman et al. 2004; Scott and Kaiser 2004; Knowles et al. 2006). Simulating the West's hydrology on a daily time scale from 1915 to 2003, Hamlet et al. (2005) showed that most of the large-scale declines in snowpack and shifts in snowmelt runoff are related to temperature trends.

The present paper uses climate data from nearby stations to evaluate the sensitivity to temperature and precipitation of 1 April SWE at nearly 1000 locations in the West. Multiple linear regression is used to reconstruct SWE using only the climate variables, permitting a separation of the influences of temperature and precipitation on the observed trends. The role of Pacific climate patterns in the observed trends in SWE is also examined.

2. Data

As in Mote et al. (2005), snow course data through 2002 were obtained from the Water and Climate Center of the U.S. Department of Agriculture's Natural Resources Conservation Service (NRCS; www.wcc.nrcs.usda.gov/snow/snowhist.html) for most states in the United States, from the California Department of Water Resources for California (http://cdec.water.ca.gov), and from the Ministry of Water, Land, Air, and Parks for British Columbia (http://www.env.gov.bc.ca./rfc/). In some instances, manual snow courses have been replaced by automated snow telemetry (SNOTEL) instruments; where overlap was sufficient to estimate the relationship between the snow course and SNOTEL estimates, NRCS provides two time series with back-

* Joint Institute for the Study of Atmosphere and Ocean Contribution Number 1134.

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FIG. 1. Map of study area including major features and locations of certain snow courses mentioned in the text.

and forward-estimated values,[1] and we use both. In total 995 snow records span the time period 1960–2002 and are used here. Results for other periods of record (e.g., 1935–2002) are also shown, and results for 1 March and 1 May are examined but not shown.

For each snow course location, reference time series of temperature and precipitation are composed by combining observations at the five nearest climate stations for the months November through March, which roughly correspond to the snow accumulation season. These climate observations are drawn from the U.S. Historical Climate Network (USHCN; Karl et al. 1990) and from the Historical Canadian Climate Database (HCCD; Vincent and Gullett 1999), and the process of selecting and combining station records into reference time series is as described by Mote (2003a). The terrain and other geographic details are shown in Fig. 1.

3. Regression analysis

The 1 April SWE represents a cumulative, simplified summary of the previous several months' weather: deposition of snow, melting or ablation of snow, and rain events that may either partially melt the snow or be absorbed in the snowpack, increasing SWE. This simplified summary of weather will be represented in different fashions by 1 April SWE and by the mean temperature T and precipitation P during the snow accumulation season, taken here as the months November through March. We explore here how successfully the seasonal means of T and P at the nearest set of five climate stations can represent the observed interannual variability SWE(t) of mountain SWE for each year t, using the regression equation

$$\mathrm{SWE}(t) = a_P P(t) + a_T T(t) + \varepsilon(t), \quad (1)$$

where a_p and a_T are the regression coefficients for precipitation $P(t)$ and temperature $T(t)$, respectively, and ε is the residual at year t (see, e.g., von Storch and Zwiers 1999, p. 160ff, for a description of regression analysis.) After determining the regression coefficients, we derive $S(t) \equiv \max[a_P P(t) + a_T T(t), 0]$. In words, $S(t)$ represents the variability associated (via regression) with the precipitation and temperature at the nearest climate stations but is not permitted to be negative. Three examples are given in Fig. 2. Emerald Lake has the highest correlation (0.95) between SWE(t) and $S(t)$ of any snow course in the West: its 1 April SWE is determined almost entirely by total seasonal precipitation, and it is uncorrelated with temperature owing to its exceptionally high elevation. The simulated trend over the period of record (+7 cm or +8%) is fairly close to the observed trend (−3 cm or −3%) and both are statistically indistinguishable from zero.

Fish Lake, Oregon, has one of the highest correlations with temperature (−0.66) and has correspondingly shown a very large decline. On several occasions the observed SWE there was zero, as happens at many locations in the Cascades; but only twice (1934 and 1977, both exceptionally warm dry winters) does $S(t)$ reach zero. Because the simulated variance is always less than the observed variance, $S(t)$ tends to be too high at low values and too low at high values (see Fig. 2, right) and the simulated trends are usually smaller than the observed trends, as illustrated at Fish Lake.

At some sites, this regression approach does poorly at characterizing the interannual variations. For the West as a whole, a total of 61 (6%) snow course/SNOTEL locations have a correlation[2] between SWE(t) and $S(t)$ that is less than 0.4 (Fig. 3). Big South, Colorado, is an example of such a location—only a handful of sites

[1] That is, estimated values for the discontinued snow course have been derived for the years since it ended, and estimated values for the SNOTEL measurements have been derived for the period before it began. In most cases the mean values are different, but trends are similar. This affects about 10% of the data points.

[2] For $n = 41$ the critical value of the correlation coefficient is 0.254 at $\alpha = 0.05$.

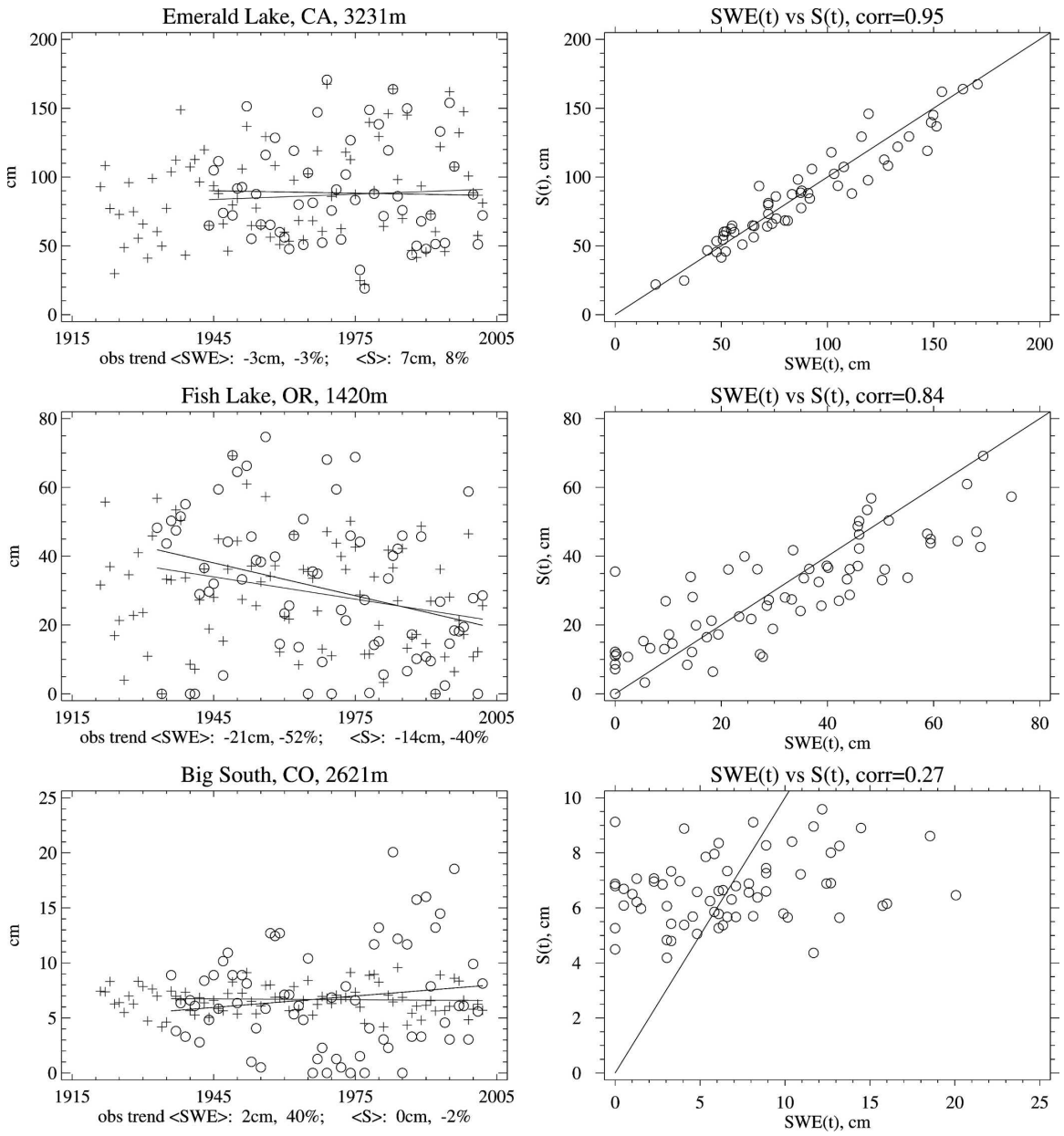


FIG. 2. (left) 1 Apr SWE as observed (circles) and as represented using regression on Nov–Mar temperature and precipitation at nearby USHCN (climate) stations (+), for three locations. Each time series also includes a linear fit over the period of record of the snow data. (right) Scatterplots of the observed SWE and $S(t)$ [see Eq. (1)], along with the 1:1 line. Emerald Lake is insensitive to temperature and has shown very little decline; Fish Lake has a correlation with temperature of -0.66 and shows a large decline related to the warming, in common with other temperature-sensitive areas. Big South has one of the lowest correlations with $S(t)$.

have correlations as low as 0.3. The variance explained is quite small, so the trends in climate-derived SWE are necessarily small and unrepresentative of observed trends in such cases, in this case of a substantial positive (40.4%) trend owing to increases in precipitation there (note also the low mean SWE at Big South.)

To visualize the correlations between SWE and the reference time series of temperature and precipitation at each location, Fig. 3 plots the correlations as a scatterplot. The mean correlation with precipitation is 0.65, and half the values fall between 0.53 and 0.81. The mean correlation with temperature is -0.22, and half the values fall between -0.10 and -0.36. Many of the locations exhibiting very high correlations with precipitation (like Emerald Lake; Fig. 2) are found in the highest, coldest locations in the Sierra Nevada and the

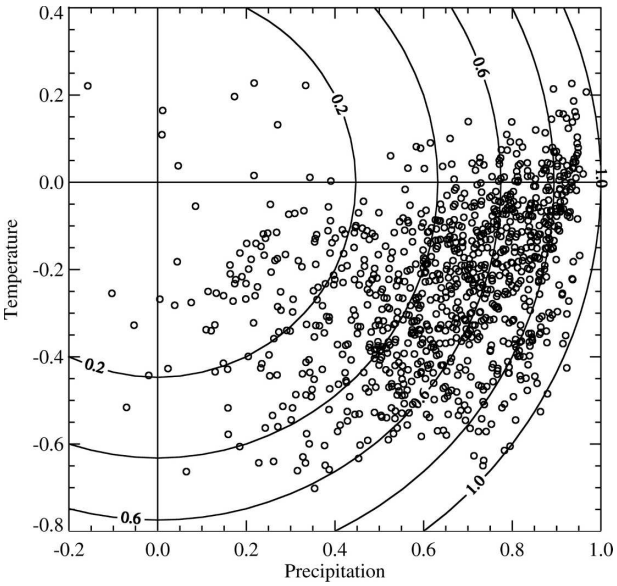


FIG. 3. Each small circle marks the correlations between 1 Apr SWE at one of the 995 snow course locations and the reference time series of Nov–Mar precipitation (x axis) and temperature (y axis). Contours indicate the quantity ($r_T^2 + r_P^2$), an approximation of the variance explained.

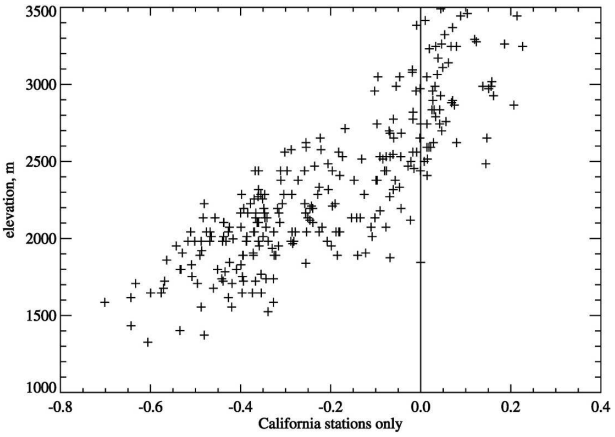


FIG. 4. Correlation between 1 Apr SWE and Nov–Mar temperature at each snow course in California, plotted as a function of snow course elevation.

Rockies. The ones with large (negative) correlation with temperature are generally found in the Cascades and the mountains of northern California and also in the Southwest (see Fig. 4 of Mote et al. 2005). A handful of sites have negative correlations with precipitation; two are in northeastern Oregon and three are in southern British Columbia. The reference time series of precipitation at these sites may not be a good representation of actual precipitation owing to topography and distance. A larger number of sites have positive correlation with temperature, which is more physically plausible, especially when we examine where these sites are found.

Temperature sensitivity—or correlation with temperature—is clearly dependent on elevation, especially for California (Fig. 4). The average correlation is near zero above about 2500 m and grows increasingly negative at lower elevations, exceeding −0.5 below about 1800 m. As we will see shortly, most of the low-elevation and most temperature-sensitive snow courses in California lie in the northern half of the state.

Using the regression coefficients a_p and a_T, we can explore the roles of temperature and precipitation trends in producing observed trends in SWE. That is, evaluating the long-term trends of each term in Eq. (1) permits a comparison of the separate components $a_p\langle P\rangle$ and $a_T\langle T\rangle$, where $\langle Y\rangle$ indicates the linear trend in $Y(t)$. Maps of $\langle$SWE$\rangle$, $a_p\langle P\rangle$, $a_T\langle T\rangle$, and $\langle S\rangle = a_p\langle P\rangle + a_T\langle T\rangle$ are shown in Fig. 5 for the period 1960–2002. As noted previously for the periods of record 1950–97 (Mote et al. 2005) and 1916–2003 (Hamlet et al. 2005), observed trends in 1 April SWE over the period of record 1960–2002 are also predominantly negative: in fact, the fraction of sites having negative trends (76%) is approximately the same regardless of period of record and is roughly the same for the gridded output of the Variable Infiltration Capacity (VIC) hydrologic model and for the unevenly distributed snow course sites.

Comparing observed trends (Fig. 5a) with trends in $\langle S\rangle$ (Fig. 5b) provides an indication of the degree to which climate variability can explain the observed trends in SWE. Broadly, the sign and spatial distribution of trends agree, with correlations between $\langle$SWE$\rangle$ and $\langle S\rangle$ of 0.71. As observed, the largest absolute trends are found in the Cascades of Oregon and the mountains of northern California, as well as parts of the Rockies; positive trends prevail in the southern Sierra Nevada mountains.

The largest trends occurred in the Cascades and the mountains of northern California and were evidently a result of both temperature and precipitation trends over this interval. Modeling work (Hamlet et al. 2005) suggests that temperature slightly predominates over this interval but that the declines over the longer 1916–2003 interval were almost entirely due to temperature trends.

The quantity $a_T\langle T\rangle$ (Fig. 5c) is almost everywhere negative, reflecting the predominance of negative values of a_T (except at a few very high-elevation locations; see Fig. 4) and positive values of $\langle T\rangle$, and values of $a_T\langle T\rangle$ are particularly large in the Cascades and the mountains of northern California. The field $a_p\langle P\rangle$

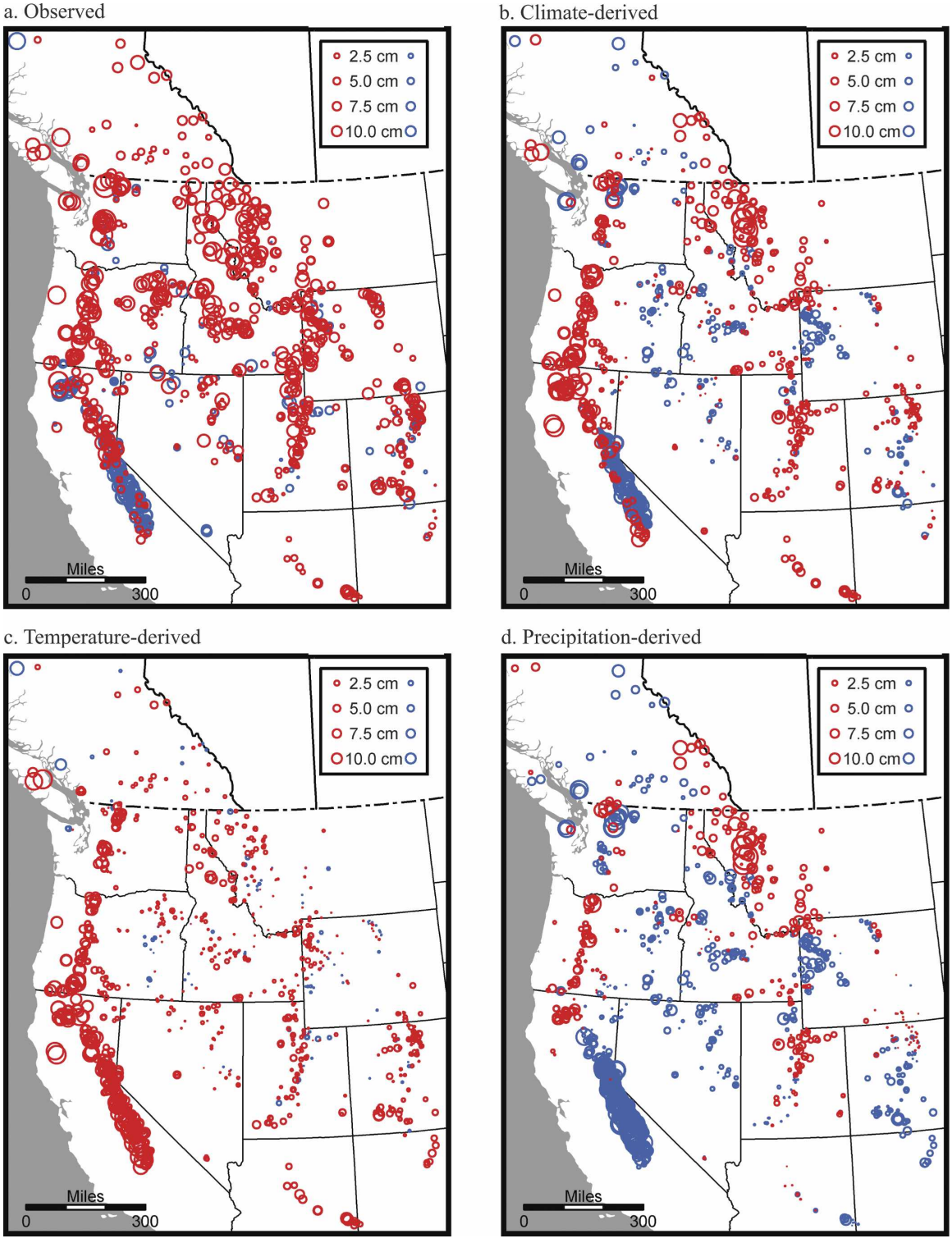


FIG. 5. Trends in 1 Apr SWE over the 1960–2002 period of record: (a) directly from snow course observations; (b) $\langle S \rangle = a_P \langle P \rangle + a_T \langle T \rangle$; (c) $a_T \langle T \rangle$; and (d) $a_P \langle P \rangle$. Positive trends are shown in blue and negative in red, by the scale indicated in the legend.

(Fig. 5d) is much more heterogeneous, with roughly equal numbers of positive and negative values; the typical scale of a patch of the same sign is on the order of 500 km.

Discrepancies in sign exhibited in Fig. 5b have a positive bias and tend to be clustered. Overall, the fraction of positive trends in S is much higher (42%) than for observations (24%). Discrepancies in sign occur predominantly in western Wyoming, eastern Oregon, southwestern Idaho, and Nevada, but there are none in Arizona and almost none in Washington, New Mexico, western Oregon, or Montana.

There are several possible causes of differences in sign between <SWE> and $<S>$:

1) changes in land cover at a snow course (e.g., growth of forest canopy) that would introduce a nonclimate trend in <SWE>;
2) trends in precipitation or temperature at the snow course location that do not match those at the lower-elevation stations used to form the reference time series;
3) changes in site characteristics at the reference climate stations such that the trends $<T>$ and $<P>$ are not representative of the true climate trends; and
4) other factors, like sensitivity to wind direction, that may have a time-dependent influence on the regression coefficients a_T and a_P.

Anecdotal evidence from snow surveyors suggests that possible cause 1 is a factor at some locations, though not generally the dominant factor. No systematic documentation of changes in site characteristics has been compiled. Tests with the VIC hydrology model and paired-catchment comparisons suggest that even drastic changes in land cover (clearcut logging) have an effect on hydrology that is secondary to that of climate (Bowling et al. 2000).

The second factor has been the subject of much recent research, at least for temperature (e.g., Diaz and Bradley 1997; Pepin and Seidel 2005), but most find little evidence that trends in temperature in high-elevation areas are significantly different from those in low elevations. Algorithms that relate temperature and precipitation to elevation for producing gridded climate data (Daly et al. 1994; Hamlet and Lettenmaier 2005) assume constant lapse rates in most conditions, ensuring that trends in temperature and precipitation at high elevation will be locked to those at low elevation, and hence are not suitable for evaluating this question. In mountainous areas, surface trends have been larger than free-air trends except at sites well exposed to the free atmosphere [which Pepin and Seidel (2005) called mountaintop sites]. Knowles et al. (2006) found a slight increase of temperature trend with elevation for the western United States, but these were in cooperative weather data, which were uncorrected for station moves, etc., as in the USHCN. Far more difficult than determining how temperature trends depend on elevation is determining how precipitation trends may depend on elevation: obviously the absolute trends will usually be larger at higher, wetter locations, but a variety of local and synoptic-scale factors could produce different relative trends at different altitudes, and in fact even at similar elevations the trends in precipitation at nearby locations can be quite different (Mote 2003b). Hence we cannot evaluate to what extent the discrepancies between <SWE> and $<S>$ may result from different climate trends at lower and higher elevations.

The third factor should be minimal, as the USHCN and HCCD datasets have been compiled and adjusted with precisely these factors in mind. In addition, the averaging of five climate stations to form a reference time series should further reduce the role of any uncorrected site changes.

Despite the concerns about these and other discrepancies, this simple regression approach clearly has some merits in diagnosing the changes in SWE that have been observed. In particular, it corroborates the findings of Hamlet et al. (2005) that temperature and precipitation both played a role in the recent decline of snowpack in the Cascades.

4. Latitude–altitude transect

The largest absolute trends, largest relative trends, and greatest sensitivity to temperature generally are found on the coastal cordillera, the string of mountains from coastal British Columbia through the Cascades and Olympics, the coastal mountains of Oregon and northern California, and the northern Sierra Nevada. In the southern Sierra Nevada mountains, mean elevations are so much higher that despite their low latitude they are cold enough to accumulate substantial snowpack. To visualize the behavior of SWE in these critically important mountains, Fig. 6 shows a transect of snow course locations in California and in Washington, Oregon, and British Columbia west of 120° longitude (the meridian of the California–Nevada border north of Lake Tahoe).

Several observations emerge when looking at the snow course data from this perspective. First, the elevation of snow courses declines with latitude; it turns out that the slope of the distribution of snow courses roughly follows an isotherm, approximately 1 km in altitude per 7.3° latitude. [The isotherms were deter-

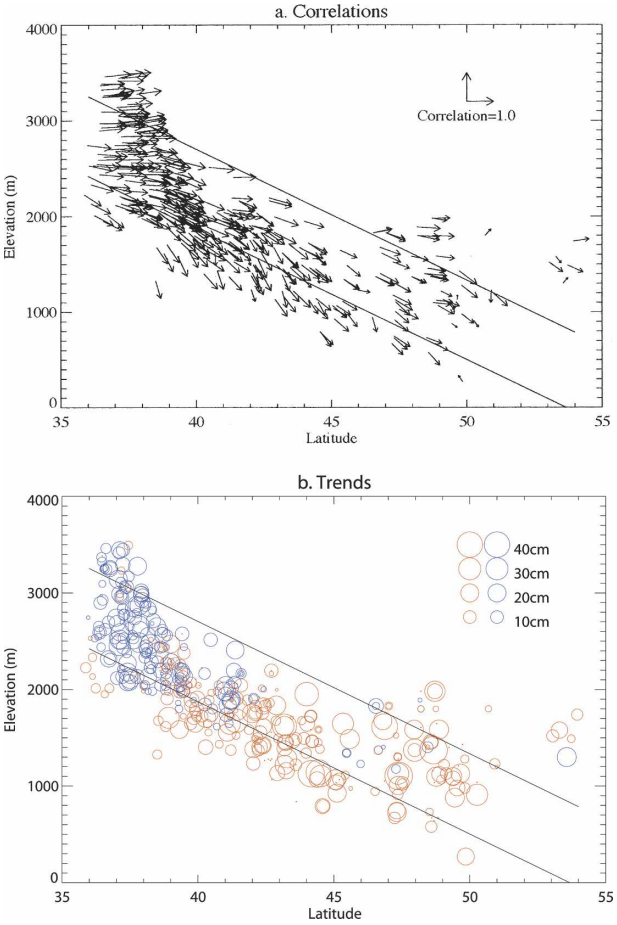


FIG. 6. Transect along the western cordillera of (a) correlations between 1 Apr SWE and Nov–Mar precipitation (x direction) and temperature (y direction), with unit correlations shown in the legend. Sloping lines are the 0° and −5°C isotherms. (b) Linear trends in 1 Apr SWE over the 1960–2002 period of record.

mined by multiple linear regression on latitude and altitude, using December through February mean temperatures at the snow course locations from the Precipitation-Elevation Regression on Independent Slopes Model (PRISM) dataset (Daly et al. 1994; see Mote et al. 2005).] The exception to this distribution is in the southern Sierra Nevada, where a large area of the mountain range and a large number of snow courses are higher than 3000 m, well above the −5°C isotherm that defines the highest elevation of snow courses farther north.

The correlations with precipitation and temperature are clearly dependent on elevation (Fig. 6a). At most latitudes, low-elevation locations (those at or below the 0°C isotherm) have high negative correlation with temperature and lower correlation with precipitation; traveling higher in elevation reduces the magnitude of the correlation with temperature and increases the correlation with precipitation. At Antelope Springs, California (latitude 38.5°N, elevation 1326 m) 1 April SWE is determined almost entirely by temperature. In the southern Sierra Nevada mountains, as one travels up past the −5°C isotherm the correlations with temperature go from negative to weakly positive, indicating that at these rare, very cold locations the warmest winters merely increase the moisture content of the air without increasing the risk that precipitation falls as rain (see also Fig. 4). Some of the locations in British Columbia (latitude >49°N) have very small correlation, which may reflect the sparsity of climate stations there.

Trends in 1 April SWE are also easier to interpret in this latitude–altitude transect (Fig. 6b). North of 40°N, negative trends prevail and positive trends tend to be found only at higher elevation. South of 40°N, positive trends prevail, but there are also negative trends mostly at lower elevation. Some of the largest absolute losses in SWE are found at lower elevations, even though mean SWE there is typically less than at higher elevations. That is, relative losses have been quite large—in excess of 80%—at low elevations, especially in Oregon (see Figs. 1 and 2 of Mote et al. 2005).

5. Connections to climate variability over the Pacific Ocean

Patterns of variability over the Pacific Ocean have known influences on winter and spring climate of the West. Statistical connections between snowpack and any index of climate variability $X(t)$ can be established based on interannual fluctuations and used to deduce the role of X in longer-term trends, using the relationship

$$\mathrm{SWE}(t) = a_X X(t) + \varepsilon(t) \quad (2)$$

[where, as in Eq. (1), a_X is the regression coefficient of SWE at each snow course location, $X(t)$ is an annually varying index, and $\varepsilon(t)$ is a residual time series] from which

$$\langle \mathrm{SWE} \rangle_X = a_X \langle X \rangle, \quad (3)$$

that is, the portion of trend in SWE attributable to X is simply the product of the trend in X and the local regression coefficient. Note that this does not presume that the trend in residuals is zero, and note also that if $\langle X \rangle$, the trend in X, is zero over a particular interval, then by definition $\langle \mathrm{SWE} \rangle_X = 0$. This is not the case for the trends examined here. In this portion of the study, each set of statistics was calculated for starting years at 5-yr intervals from 1935 to 1965 and for months March, April, and May, to examine the effects of period of record and of seasonality on the results.

For climate indices, several candidates are available that could influence SWE on decadal time scales. Pa-

cific decadal oscillation (PDO; e.g., Mantua et al. 1997) and El Niño–Southern Oscillation (ENSO) are both indices of variability in sea surface temperature in different regions. ENSO has well-documented influences on 1 April SWE in much of the West, with warm phase associated with lower snowpack in the Northwest and higher snowpack in the Southwest (Clark et al. 2001; their analysis of ENSO influence will not be repeated here).

Accounting for these patterns is important in understanding trends in SWE, especially since the period of most abundant snow course observations since 1945 corresponds approximately with a shift from predominantly cool phase PDO (relatively high snowpack and streamflow in the Northwest, opposite in the Southwest) during 1945–76 to predominantly warm phase PDO (low snowpack and streamflow) from 1977 to perhaps 1998.

Another measure of Pacific climate variability, one based not on properties of the ocean but of the atmosphere, is the North Pacific Index (NPI; Trenberth and Hurrell 1994), defined as the mean atmospheric sea level pressure over the region 30°–60°N, 160°E–140°W. PDO and NPI are weakly correlated (r = 0.30) during the 1900–2002 period of record and share similar spectral characteristics with statistically significant (α = 0.05) interdecadal variability; PDO has higher lag-1 autocorrelation (0.40 versus 0.22). Both PDO and NPI have nonzero trends over the period of record of the snow courses, roughly since 1930 (Fig. 7); while the magnitude of the trends depends on the starting year, the sign is consistent for all starting years. Both indices reflect the late-twentieth-century trend toward a deeper Aleutian low. Though the PDO had predominantly positive values in 1925–45 and 1977–95, and predominantly negative values in 1945–76, since 1995 the index has changed sign frequently; in fact, the pattern of North Pacific variability has changed from predominantly first empirical orthogonal function (EOF1) (PDO) to predominantly EOF2 in the early 1990s (Bond et al. 2003).

The two indices, PDO and NPI, influence SWE over different spatial domains (Fig. 8). The general pattern of correlations is for SWE in the Northwest to be positively correlated with NPI and negatively correlated with PDO, and the opposite in the Southwest, but the dividing line between positive and negative correlations is quite different for the two indices. For PDO, it runs diagonally from northern California through central Idaho to the Montana–Wyoming border. For NPI, it runs more east–west from central California. Interesting details about these differences will emerge when we examine the correlations again as a latitude–altitude

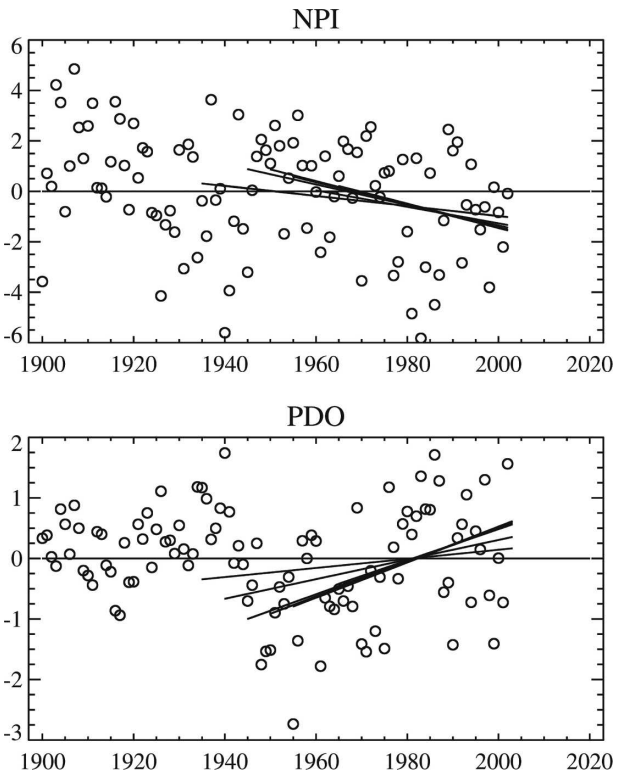


FIG. 7. Winter (Oct–Mar) average values of (top) the North Pacific Index and (bottom) the Pacific Decadal Oscillation index, along with linear fits for data spanning 1935–2002, 1940–2002, . . . 1965–2002.

transect. Correlations between SWE and NPI are generally larger than between SWE and PDO, and far more are positive. The starting year has very little influence on the correlation; for instance, the mean absolute difference between correlations of PDO and 1 April SWE for 1935–2002 and 1965–2002 (for stations with data 1935–2002) is only 0.07.

The quantities $<\text{SWE}>_{\text{PDO}}$ and $<\text{SWE}>_{\text{NPI}}$ for 1960–2002 are plotted in Figs. 8c,d, using the same scale as Fig. 5. These large-scale climate patterns clearly help explain some portion of the negative trends in the northern snow courses and also some portion of the positive trends in the southern snow courses. NPI is particularly helpful at explaining the downward trends in the northwestern snow courses, and PDO is helpful at explaining the upward trends in the central and southern Sierras.

From the maps it is difficult to evaluate the fraction of trend explained by PDO or NPI, so in Fig. 9 we show a scatterplot of the observed trend (Fig. 5a) against the quantity shown in Fig. 8d, $<\text{SWE}>_{\text{NPI}}$, along with the $y = x$ line and two least squares linear fits passing through the origin, one for snow courses in the Northwest (latitude >42°, longitude >120°, points repre-

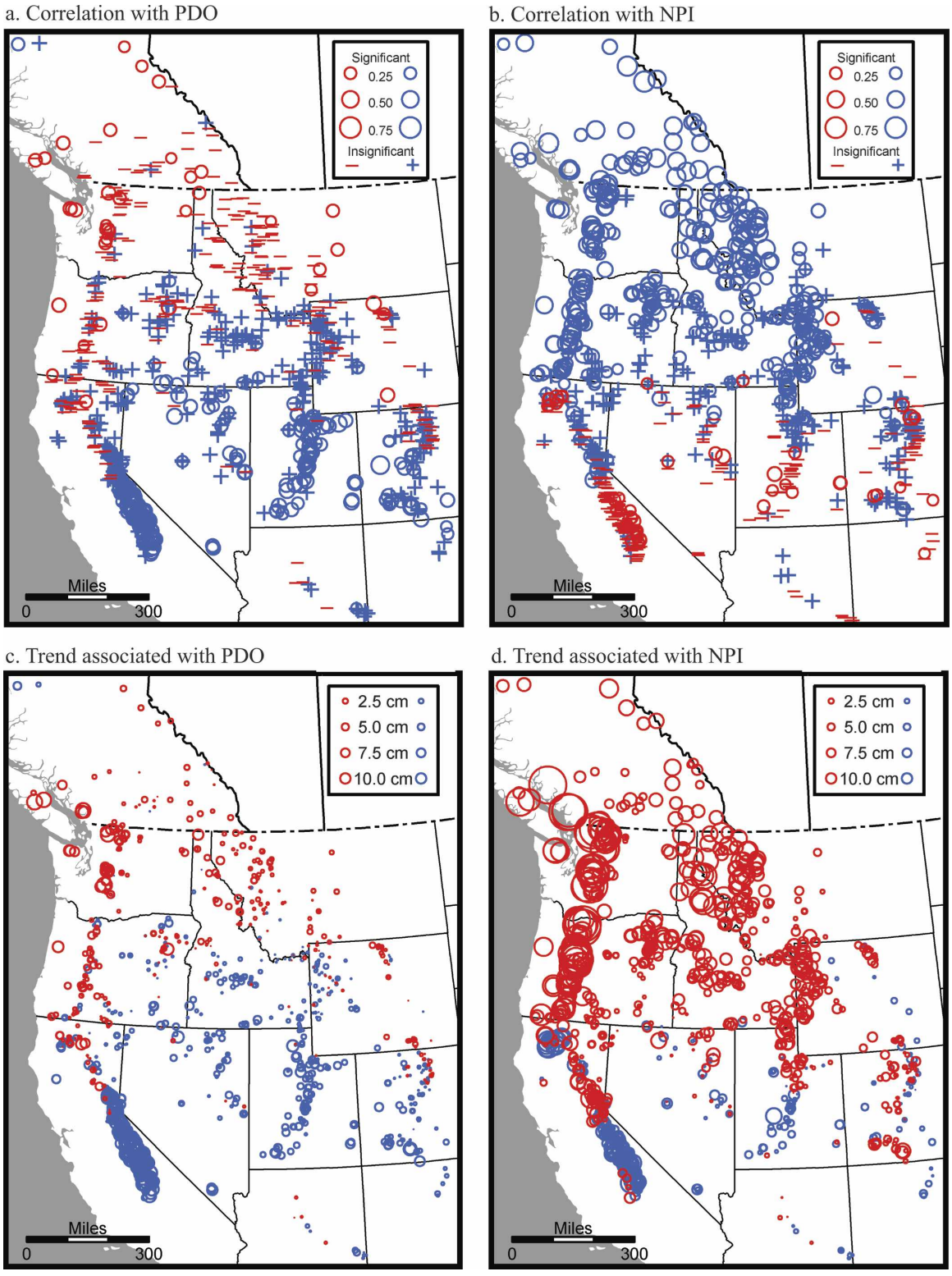


FIG. 8. Relationships between two climate indices, NPI and PDO, and 1 Apr SWE, over the 1960–2002 period of record. (a), (b) Correlations are shown as red for negative and blue for positive; circles indicate statistically significant trends, and + or − indicates insignificant trends. (c), (d) The trend explained by regression with the index, $<\text{SWE}>_X$ [see Eq. (3)], in units of cm as in Fig. 5.

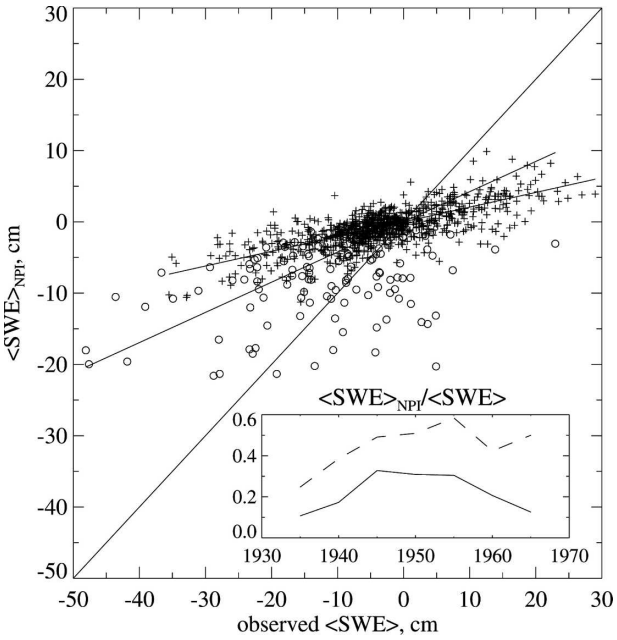


FIG. 9. Scatterplot between absolute trends and trends explained by NPI, that is, between the points on the map in Fig. 5a and the points in Fig. 8d. One line is fitted to PNW points (circles) and one is fitted to non-PNW points (+ symbols). The inset shows how the slopes of the two lines depend on the starting year of the analysis, with the dashed line representing the PNW points.

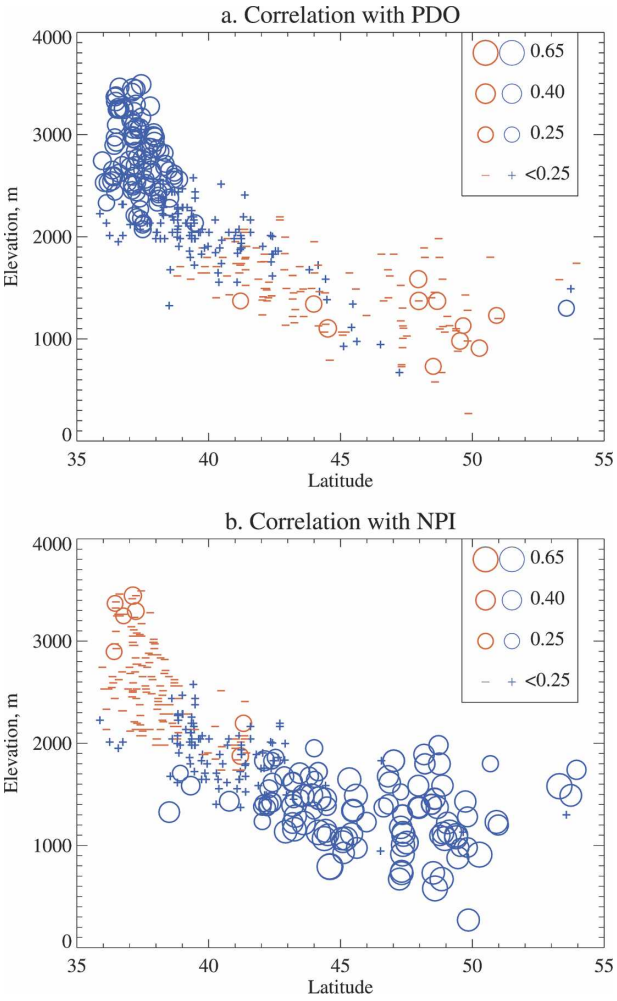


FIG. 10. Transect (as in Fig. 6) of the correlations shown in Figs. 8a,b.

sented by circles) and one for all the rest of the snow courses. The slope of the line represents the mean fraction of trend attributable to NPI, in this case 0.42 for points in the Pacific Northwest (PNW) and 0.21 for non-PNW points. A few of the snow courses in the PNW with trends between 10 and −30 cm lie near the $y = x$ line, indicating that those trends can be almost completely explained by the NPI, but the largest negative trends are more typical, with about 20%–40% of the trends explained by NPI. The mean trend for PNW snow courses over 1960–2002 is −15 cm or −34%, and for the rest of the West it is −9.9 cm or −22%. The inset shows how the slopes of the two fitted lines depend on the starting year, from 1935 when very little of the overall trend could be explained by NPI to 1945–55 when about half of the trend in the PNW and a third of the trend in the rest of the West are explained by NPI. Results are roughly similar for PDO (not shown), except that the slopes of the lines are typically less for PNW than for non-PNW snow courses owing to the smaller correlation with PDO in the PNW.

Plotting the correlations in Fig. 8 as a latitude–altitude transect (Fig. 10) reveals a striking fact about the relationship between these climate indices and SWE: the correlations depend strongly on elevation in California, as well as on latitude from north to south. In northern California and southern Oregon, higher snow courses are positively (but not significantly) correlated with PDO, while lower snow courses are negatively correlated with PDO. In central and southern California, the strength of the correlation generally increases with altitude. For NPI, the situation is somewhat reversed: in northern California and southern Oregon, most of the lower-elevation snow courses are significantly positively correlated with NPI, and at higher elevations some significant negative correlations emerge. In southern California (south of 38°N), a few of the lowest-elevation snow courses are positively (though weakly) correlated with NPI and most of the highest-elevation snow courses are significantly negatively correlated with NPI. These differences can be understood as an interaction between the separate influences of the climate pattern (PDO or NPI) on temperature and on precipitation: at low elevations, temperature matters

most, and temperature is negatively correlated with NPI throughout the transect (see www.cdc.noaa.gov/USclimate/Correlation). At higher elevations, precipitation matters more, and precipitation is (weakly) negatively correlated with NPI in California and positively in the PNW. Thus, the elevationally determined competition between the effects of temperature and the effects of precipitation are convolved with the spatially varying response to NPI (or PDO) to produce this complex pattern of correlations with NPI and PDO.

6. Conclusions

The modest success of this simple regression approach suggests that for most snow course locations in the West, the long-term variations in spring SWE are reasonably well explained by summaries of seasonal climate at nearby locations. This implies that day-to-day details of snow accumulation, ablation, and melt are generally of secondary importance, except where correlations between observed SWE and climate-derived SWE are low.

On longer time scales, another implication of the general agreement between observed and climate-derived SWE is that other sources of trends are also generally of secondary importance. Where discrepancies exist, they could be caused by, for example, changes in forest canopy or other aspects of site characteristics that influence SWE, or substantially different climate trends between the relatively low-elevation climate stations and the snow course locations.

During the second half of the twentieth century, and likely even since 1916 (Hamlet et al. 2005), winter and spring warming in the West have reduced spring snowpack at most locations. Increases in precipitation appear to have offset this loss in some places since midcentury, notably in the southern Sierra Nevada mountains, where large increases have occurred. Some of the interannual variability and long-term trends can be explained as a response to variability and change in North Pacific climate, especially as represented by the North Pacific Index (NPI), which responds to the oceanic variations ENSO and PDO. However, NPI can only account for about half of the trends in the PNW since midcentury (and less elsewhere or from earlier starting points), in rough agreement with the modeling results of Hamlet et al. (2005). The remaining portion clearly includes the influence of the monotonic warming (e.g., Fig. 5c) observed throughout the West, which is largely unrelated to Pacific climate variability and may well represent human influence on climate (Karoly and Wu 2005). That is, even after accounting for the role of known patterns of climate variability, there is a substantial downward trend in overall snowpack in the West that is consistent with the observed warming.

Even a conservative estimate (0.3°C/decade^{-1}) of the likely warming rate for western mountains in winter (Leung et al. 2004) would, by 2100, move the 0° isotherm where the 3°C isotherm now lies: most of the westernmost mountains would be in the transient snow zone, in which snow accumulates and melts repeatedly during the snow season. One implication of Fig. 6 is that spring SWE will grow more and more sensitive to temperature. In other words, the more it warms, the more the warming will affect snowpack even at higher elevations. These results underscore those from modeling efforts (e.g., Hamlet and Lettenmaier 1999; Knowles and Cayan 2004), which have illustrated the likely declines in snowpack from a warming climate, but with the added detail that the intraseasonal behavior of snowpack will change from one of steady accumulation to alternating accumulation and loss. Simple regression-based methods for forecasting seasonal volumetric streamflow will have to be revised or replaced by more sophisticated methods that can account for the changing role of temperature both in determining the quantity of spring snowpack (the subject of this paper) and the rate at which it melts (Hamlet et al. 2005).

Acknowledgments. I am grateful to Rob Norheim, who expertly constructed the maps. These analyses would not have been possible without the efforts of the dedicated snow surveyors who laboriously collected the data. I thank Alan Hamlet, Dennis Lettenmaier, Mike Wallace, Martyn Clark, Todd Mitchell, and Dan Cayan for helpful discussions and suggestions on an earlier version of this manuscript. This publication is funded by the Joint Institute for the Study of the Atmosphere and Ocean (JISAO) under NOAA Cooperative Agreement No. NA17RJ1232.

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From: Holt, Bryon
To: Grizzle, Betty
Subject: Re: Help with locating references
Date: Monday, February 5, 2018 12:38:33 PM
Attachments: Moran-Tejeda 2012.pdf

Here is Moran-Tejeda.

On Mon, Feb 5, 2018 at 11:35 AM, Holt, Bryon <bryon_holt@fws.gov> wrote:

> 2nd email.
>
> On Mon, Feb 5, 2018 at 11:33 AM, Holt, Bryon <bryon_holt@fws.gov> wrote:
>
> > Hi Betty,
> >
> > I've got all but one document. The one document I don't yet have is by Moran-Tejeda, but I have requested it from our Conservation Library and should have it soon to send to you. I will send the ones I have now in a couple of emails to minimize size issues. Note: Brown and Mote is 2009 not 2008 - could not find 2008. May be a typo?
> >
> > Bryon
> >
> > On Wed, Jan 31, 2018 at 10:47 AM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:
> >
> > > Thank you! Will save me time pulling this final draft together.
> > >
> > > On Wed, Jan 31, 2018 at 10:44 AM, Holt, Bryon <bryon_holt@fws.gov> wrote:
> > >
> > > > I can help you with this.
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> > > > On Wed, Jan 31, 2018 at 10:26 AM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:
> > > >
> > > > > Hi Bryon - You offered to help with finalizing the SSA Report, so am writing to ask if you can locate and download the references highlighted in yellow on the first page of the attached publication. There are 7 total and full citations are at the end.
> > > > >
> > > > > Please let me know whether or not you have time to help. If not, I can ask others. Thanks.
> > > > >
> > > > > --
> > > > >
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Reg Environ Change (2013) 13:873–885
DOI 10.1007/s10113-012-0380-8

ORIGINAL ARTICLE

Evolution and frequency (1970–2007) of combined temperature–precipitation modes in the Spanish mountains and sensitivity of snow cover

Enrique Morán-Tejeda · Sixto Herrera · J. Ignacio López-Moreno · Jesús Revuelto · Anthony Lehmann · Martin Beniston

Received: 13 July 2012 / Accepted: 18 November 2012 / Published online: 29 November 2012
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Abstract Snow cover in Spanish mountains is crucial for ensuring water availability in spring and summer months, for the success of winter tourism or for the maintenance of biodiversity in mountain ecosystems. A changing climate may affect the volume of snow cover even in high mountains, where weather conditions are usually favorable for snow accumulation. In this paper, we aim to investigate the evolution (1970–2007) of combined precipitation–temperature modes in the Spanish mountains, and the sensitivity of the snowpack to their occurrence. The climatic database "Spain02" and snow thickness data for Spanish mountains were used for this purpose. Results showed that the frequency of dry-warm and wet-warm days has increased over time in all mountain ranges, while the frequency of the "cold" modes has decreased. The thickness of the snowpack in the Pyrenees has also decreased and its evolution is negatively correlated with the frequency of dry-warm days, and positively correlated with the frequency of dry-cold and wet-cold days. This work constitutes the first approach that relates the evolution of climatic conditions favorable or unfavorable for snow accumulation and the evolution of the snowpack in Spanish mountains.

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Keywords Precipitation · Temperature · Joint-quantile indices · Spain02 · Snow · Mountains · Iberian Peninsula

Introduction

Snow cover in mountains is a natural resource that is of high ecological, socio-economic and scientific value. The mountain snowpack constitutes the primary source of water in many areas, ensuring water supply to the lowlands even during the dry season (Barnett et al. 2005; Knowles et al. 2006). But snow cover in mountains has many other implications, including the functioning of alpine and subalpine ecosystems (Pauli et al. 2003; Keller et al. 2005) or the financial success of ski resorts (Elsasser and Bürki 2002). In the mountains of the Iberian Peninsula, under the predominant influence of Mediterranean climates, the environmental and economic role of snow can in some years be critical. This is due to the highly variable thickness of the snowpack related to the characteristic interannual variability of precipitation (López-Moreno 2005).

The accumulation of snow cover is a process that is highly dependent on particular weather conditions. These clearly include the combination of precipitation and below-zero temperatures for triggering the snowfall and the persistence of low temperatures during a reasonably long period that enables a sustained snowpack (Beniston et al. 2011; López-Moreno et al. 2011). These conditions are characteristic of mountains. Their topography favors the gradients of temperature such that the zero-degree isotherm can be reached at certain elevations, as well as enhanced precipitation due to the uplift of moist air (Barry 2005). Knowledge of the type of meteorological conditions that prevail in mountains during the snow accumulation and melting seasons is therefore essential to predict the

accumulation of the snowpack as well as the subsequent availability of water resources. However, snow-prone weather conditions might be threatened in mountain regions as a result of greenhouse gas induced warming (Beniston 2003). There is wide agreement that mean temperatures have increased globally during the last 40 years, in relation to the observed increases in radiatively active greenhouse gases (IPCC 2007; Solomon et al. 2007). Global warming is responsible for the retreat of glaciers in different parts of world, as well as for the reduction in the snowpack in mountains (see examples in Beniston 2003). In Europe, the Alps have been extensively studied in terms of the effects of climate variability on snow cover (Beniston et al. 2003, 2011; Uhlmann et al. 2009), and most studies highlight a reduction in the snowpack due to warmer conditions, both in the past and projected for coming decades. Similar conclusions were drawn for the mountains of the western US and Canada, where snow also constitutes a primary source of water resources (Hamlet et al. 2005; Knowles et al. 2006).

There is a clear need for research in other mountains of Europe, due to their role as "climatic islands," such as those in the Mediterranean basin (López-Moreno et al. 2011). In the Iberian Peninsula, various studies highlighted changes in the seasonality of mountain river flows, which have been attributed to reduced snowpack (López-Moreno et al. 2008; Morán-Tejeda et al. 2011). Contrary to temperature, precipitation has not exhibited a homogeneous trend signal around the world. Some regions, including the Mediterranean, are expected to experience drier conditions on the future decades (Milly et al. 2005; IPCC 2007). To date, however, there has been no clear trend toward more or less precipitation in Mediterranean countries (Bladé and Castro-Díez 2010).

Given the specific meteorological conditions for snowfall, that are seasonally dependent in mid-latitudes, studies pertaining to the behavior of snow can be placed within the study of climatic extremes, that is, the values of the climatic variables located within the tails of the frequency distributions. Much effort has been made in recent years to characterize the behavior of extremes, as exemplified by the number of indices recommended by the World Meteorological Organization Commission for Climatology (CCl) jointly with the World Climate Research Programme (WCRP) project on Climate Variability and Predictability (CLIVAR) (see examples in Alexander et al. 2006). These indices are derived from series of daily precipitation and temperatures, and in general, they characterize one single variable (e.g., warm days, heavy precipitation days, etc.), but their statistical calculation also enables different variables to be combined. Beniston (2009) demonstrated that the use of joint-quantile indices can provide new insights for the understanding of climate variability, due to the mutual feedbacks that both temperature and precipitation may exert upon each other. This approach was shown to be useful as a proxy for predicting the evolution of snowpack in the Alps (Beniston et al. 2011), because particular combinations of temperature and precipitation are closely related to the synoptic circulations favorable (or not) for snowy winters. We believe that this method can also be applied to the mountains of southern Europe and thus its performance when applied to Spanish mountains will be examined in this work.

Based on the above considerations, the objective of this paper is to study the evolution and trends of combined precipitation and temperature quantile-based indices in the mountain ranges of the continental Spain during the last three decades. The underlying hypothesis is that long-term changes of moisture and temperature in mountains may trigger significant changes in thickness of the snowpack. We thus also seek to examine the relation of the observed evolution in climate indices with that of the snowpack during winter and spring seasons. This work constitutes the first assessment of climate evolution in Spanish mountains in relation to snow accumulation. The results will then form the basis for understanding the future evolution of the snowpack in the Iberian mountains in a changing climate.

Study area

The areas of research are the main mountain chains in continental Spain. The most important mountain ranges, with elevations generally over 2,000 m, are the Pyrenees, the Central System, the Cantabrian Range, the Iberian System and the Baetic System (Fig. 1a). The latter range is the one with the highest altitude of the Iberian Peninsula, namely the Mulhacen Peak (3,478 m). The geographical disposition of the Iberian Peninsula, located between the Atlantic and the Mediterranean, together with the complex topography, induces a notable variety of climates throughout the territory. A northwest–southeast gradient of precipitation and temperature is observed, with Atlantic climate conditions (precipitation over 1,000 mm/year and cool temperatures) in the northwestern quadrant, semi-arid conditions (precipitation less than 300 mm/year and warm temperatures) in the southwest and variants of Mediterranean climates in between. Precipitation is concentrated mainly in winter (DJF with about 40 %), when minimum temperatures usually drop to 0 °C or below, depending on the elevation. Thus, snowfalls are frequent especially in the mountainous areas where a deep snowpack usually accumulates. Warmer conditions in spring trigger the melting of the snowpack, with a lag in time for different mountain systems, ranging from late March to early June depending on the elevation and latitude.

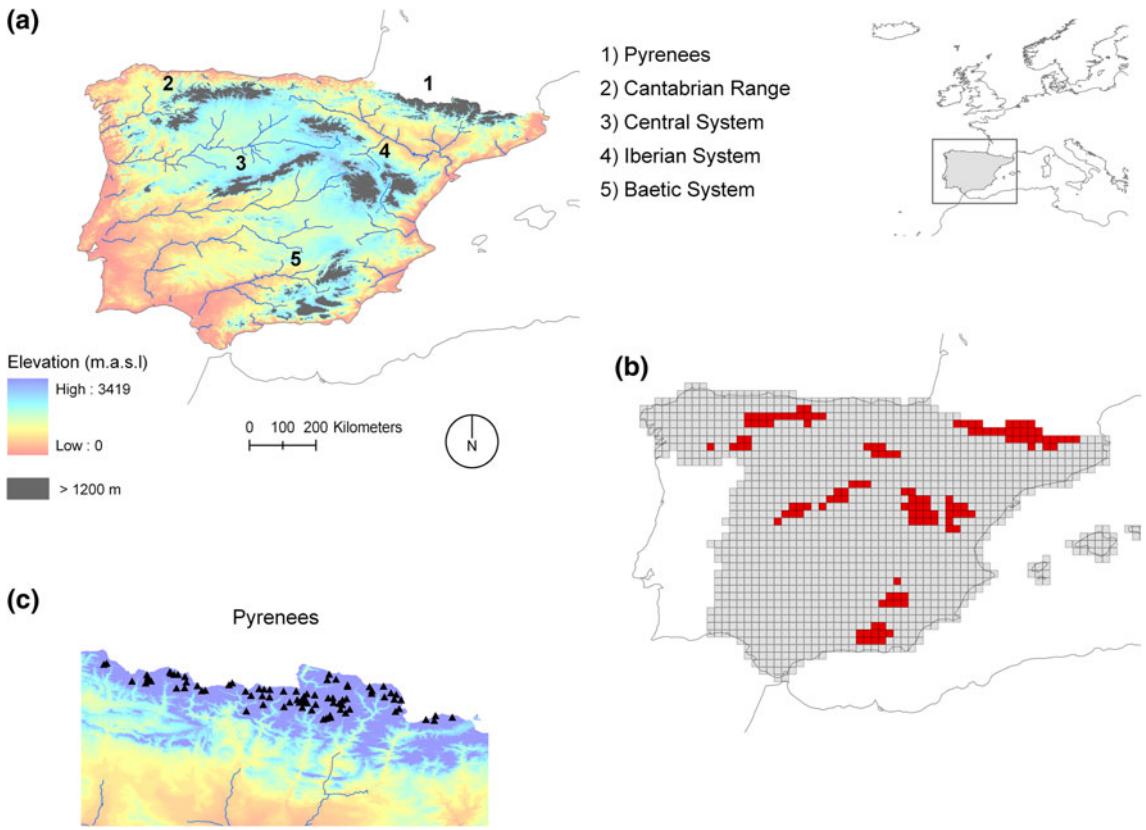


Fig. 1 **a** The Iberian Peninsula and the location of the mountain ranges considered in the study. **b** Grid resolution of the Spain02 database showing pixels with average altitude >1.200 m (*red*). **c** Location of snow poles in the Pyrenees for the study of snow-depth evolution

Data and methods

The "Spain02" database

The climatic data—precipitation and temperature—used in this study was extracted from Spain02 (Herrera et al. 2012). This is a regular 0.2° latitude/longitude (approximately 20 km) daily gridded data set containing precipitation, maximum and minimum temperatures, obtained from quality-controlled surface stations from the *Agencia Estatal de Meteorología* (AEMET, Spanish Meteorological Agency), covering continental Spain and the Balearic Islands during the period 1950–2007. A particular elevation threshold needs to be defined in order to restrict the analyses only to mountain areas. For this, we had to ensure that the entire area of the mountain chains was analyzed, thus avoiding taking into account too many isolated pixels; but also it was necessary that the capability of the database to reflect specific climate characteristic of mountains was guaranteed. The threshold of 1,200 m for this study has been considered to be an optimal level, and as a consequence all pixels with an average elevation higher than 1,200 m were considered for the subsequent analyses (Fig. 1b). As stated by the authors of Spain02 (Herrera et al. 2012), the number of stations used in the construction of the database varies on time and this could lead to erroneous results when using the database for regional trend analyses. To overcome this problem, we only considered the period 1970–2007, for which meteorological stations used to construct the database were evenly distributed in space.

Validation against observations

In order to analyze the capability of Spain02 to reproduce weather conditions in the main mountain ranges of the Iberian Peninsula, a validation of the grid data against observations was performed. For this, we considered 91 precipitation stations and 52 temperature stations from AEMET with heights ranging from 1,210 to 2,507 m. Then, we compared the observations with the nearest grid box of Spain02 whose average height is located above the selected threshold of 1,200 m.

Figure 2 shows the validation results in terms of the bias and Pearson correlation (the same analysis was developed using the Spearman correlation with similar results). Figure 2a shows the height of the AEMET stations used for validation and the corresponding grid points of Spain02.

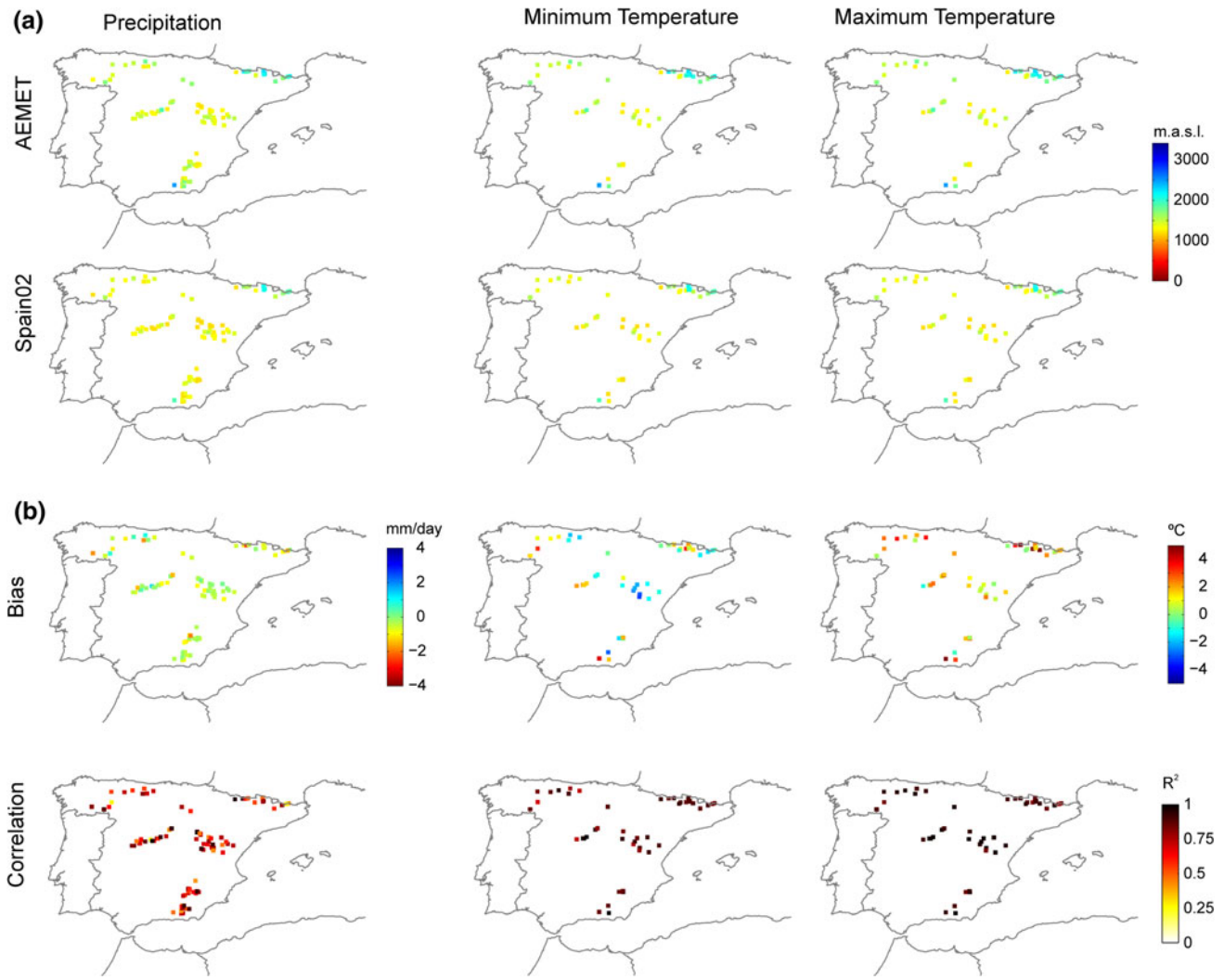


Fig. 2 Validation of Spain02 against observations. **a** Elevation of AEMET stations and corresponding grid points of Spain02. **b** Bias and Pearson correlation between observed and gridded precipitation and temperatures

There is clearly a negative bias for altitude, that is, Spain02 grid points are generally lower than the selected AEMET stations. The consequence is a general positive bias in the temperatures (Fig. 1b). Despite this bias, both temperature and precipitation displayed correlations (Fig. 1b) $R^2 > 0.6$ in all points, indicating that Spain02 reproduces reasonably well the daily variability of the three variables considered.

Moreover, we have compared the temporal evolution of the indices considered in this study (defined in Table 1) for the observations (blue line) and for the Spain02 data set (red line) for the period 1970–2007 (Fig. 3). High correlations are observed in all cases ($R^2 > 0.75$) between observed and estimated series. The variability is therefore well reproduced by Spain02, although it tends to systematically overestimate the number of dry-cold days in winter and dry-warm days in the spring. This systematic bias does not affect, however, the robustness of the trend analyses to be undertaken.

Precipitation–temperature joint quantiles and statistical analyses

The daily resolution of Spain02 database enables the behavior of far-from-normal events to be investigated. In this study, we have computed a series of indices that combine moisture and heat conditions based on the tails of the frequency distributions of daily precipitation and mean temperature series. This approach, proposed earlier by Beniston and Goyette (2007), provides a deeper understanding of the behavior of climate, due to the mutual feedbacks of precipitation and temperature that cannot be revealed when these variables are studied independently from one another. In this case, the resulting indices are representative of the weather conditions that favor or hamper snow accumulation and melting.

The different nature of the frequency distributions of daily precipitation (skewed) and temperature (normal)

Table 1 Definition of the indices considered in the study

Thresholds	Description	Units
$p0$	$P < 0.1$ mm	mm
$p25$	Value of the 25th percentile of the P-$p0$ series	mm
$t25$	Value of the 25th percentile of the T_{mean} series	°C
$t75$	Value of the 75th percentile of the T_{mean} series	°C
Exceedance series		
DW (dry-warm)	Days with $p0$ and $T > t75$	Number of days year^{-1}
DC (dry-cold)	Days with $p0$ and $T < t25$	Number of days year^{-1}
WW (wet-warm)	Days with $P > p25$ and $T > t75$	Number of days year^{-1}
WC (wet-cold)	Days with $P > p25$ and $T < t25$	Number of days year^{-1}
Spells		
A_DW	Average duration of dry-warm spells	Number of consecutive days year^{-1}
A_DC	Average duration of dry-cold spells	Number of consecutive days year^{-1}
A_WW	Average duration of wet-warm spells	Number of consecutive days year^{-1}
A_WC	Average duration of wet-cold spells	Number of consecutive days year^{-1}
M_DW	Maximum duration of dry-warm spells	Number of consecutive days year^{-1}
M_DC	Maximum duration of dry-cold spells	Number of consecutive days year^{-1}
M_WW	Maximum duration of wet-warm spells	Number of consecutive days year^{-1}
M_WC	Maximum duration of wet-cold spells	Number of consecutive days year^{-1}

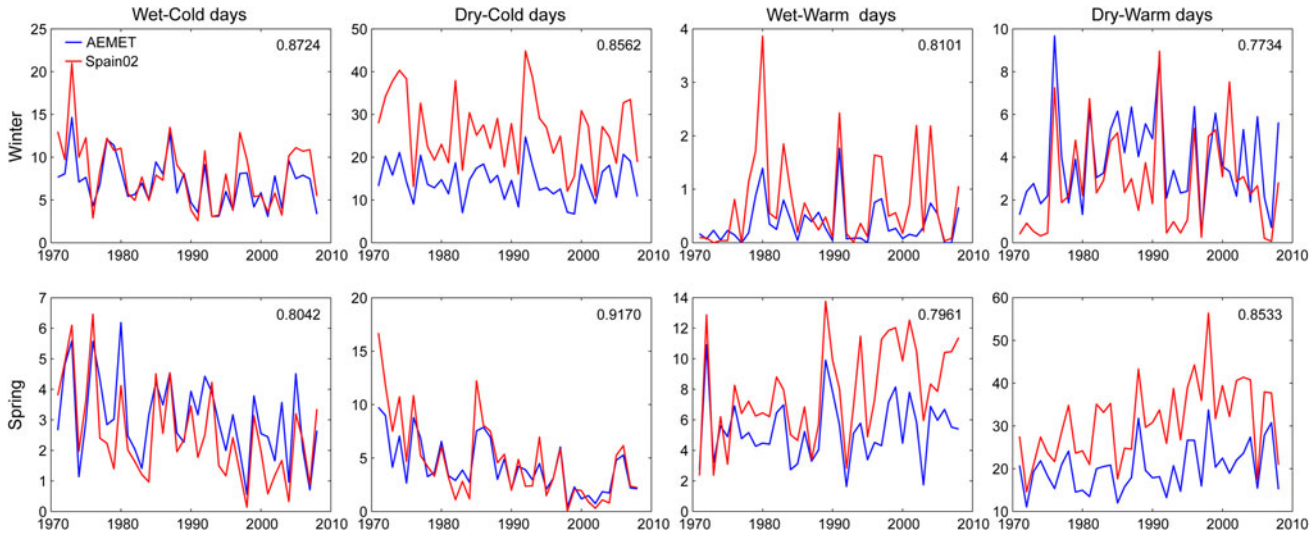


Fig. 3 Comparisons of the evolution of average (by mountain range) precipitation–temperature indices (Table 1) for observations and grid points

required the use of different thresholds for defining the indices to be considered. Calculation of the indices, for 1970–2007 and December-to-May period, involved:

- Separation of dry days ($P < 0.1$ mm) from the precipitation series and computation of the 25th percentile ($p25$) of the $P > 0.1$ mm series. Days with $P < 0.1$ mm ($p0$) were considered "dry days" and days with $P > p25$ mm were considered "wet days"
- Computation of the 25th ($t25$) and 75th ($t75$) percentiles of temperature series. Days with temperature $<t25$ °C are considered "cold days" and days with temperature $>t75$ °C are considered "warm days."
- Combination of thresholds to obtain the precipitation–temperature indices: DW (dry-warm), DC (dry-cold), WW (wet-warm) and WC (wet-cold).

Yearly series of the exceedance of these joint thresholds were calculated separately for winter (DJF) and spring (MAM), resulting in the number of days per year and season belonging to any of the combined classes. Moreover, spells of precipitation–temperature modes, that is, the

number of consecutive days of each class, were computed. For the purposes of trend analyses, we calculated the average duration of spells per year and the duration of the longest spell per year on a seasonal basis. Table 1 summarizes the indices of exceedance series and duration of spells, and Table 2 shows the minimum, mean, median and maximum values of the thresholds of precipitation ($p25$) and temperature ($t25$ and $t75$), for each mountain range considered in this study.

The exceedance series and the series of duration of spells per year and season were then checked against monotonic trends. On the one hand, linear regressions (least squared method) were fitted against time to obtain the values of the linear trend and to calculate the change between the end and beginning of the studied period. Furthermore, the Mann–Kendall non-parametric test was used for detecting the statistical significance of trends. (Kendall 1975; Berryman et al. 1988; Yue et al. 2002). A significance P level <0.05 (95 % of confidence) was set to reject the null hypothesis of the test, that is, no trend in data.

Snowpack thickness data

Snow thickness measures were provided by the ERHIN program (*Evaluación de los Recursos Hídricos procedentes de la Innivación* – assessment of water resources from snow) of the Spanish Ministry of Environment. The ERHIN program has been taking snow measurements in the Spanish mountains since mid-1980 with fixed snow poles; three measurements were carried out for every year: the first in late January or early February, the second in March and the third in mid-April or early May. Quality criteria, including a maximum number of 3 data gaps were set up to obtain reliable snow data series. The data period was set from 1986 (first year of snow sampling) to 2007 (to make it coincide with the last year of the climatic series). Unfortunately, snow data for the Cantabrian, Central, Baetic and Iberic systems exhibited many gaps and did not allow obtaining entire series for the defined period; as a result, only snow measurements for the Pyrenees were considered for the study. Some of the selected series still had few data gaps, and only those with less than 15 % of missing data were filled using the results of linear regressions with the best correlated series (i.e., those where $R^2 > 0.7$). A total of 84 series of snow depth in April–May, covering most of the Pyrenees area, were finally selected (location of snow stakes in Fig. 1c) and related to the annual frequency of the different temperature and precipitation modes.

Results and discussion

Evolution in the frequency of days and duration of spells with combined precipitation–temperature conditions

Figure 4a shows the Mann–Kendall coefficients for the exceedance series of the different combined indices for winter and spring seasons in each pixel of elevation >1.200 m, and Fig. 4b summarizes the results by mountain range. Dry and warm (DW) days showed an inhomogeneous signal in winter among mountain ranges. Negative coefficients (although mainly not significant) predominated in the Iberian, Cantabrian, Pyrenees and northern Baetic systems, and positive and significant trends in Central System and southern Baetic. During spring, on the contrary, the signal is strong and homogeneous with positive

Table 2 Minimum, median, mean and maximum values (per mountain range) of the thresholds ($p25$, $t25$ and $t75$) set to obtain the combined indices

Threshold	Pyrenees	Cantabrian range	Central system	Iberic system	Baetic system
$p25$ (mm)					
Min	0.73	1.39	0.75	0.89	0.95
Median	1.40	2.24	1.83	1.63	2.12
Mean	1.73	2.29	1.94	1.73	2.05
Max	4.69	3.57	3.85	3.05	2.80
$t25$ (°C)					
Min	−2.48	0.24	0.30	1.07	2.45
Median	0.20	1.68	2.29	2.13	4.19
Mean	0.00	1.80	2.15	2.13	4.47
Max	2.12	3.18	3.17	3.72	6.31
$t75$ (°C)					
Min	3.96	5.42	6.70	6.74	8.28
Median	6.13	7.16	8.19	8.02	10.41
Mean	6.20	7.24	8.04	7.91	10.18
Max	8.79	8.99	8.87	9.18	11.85

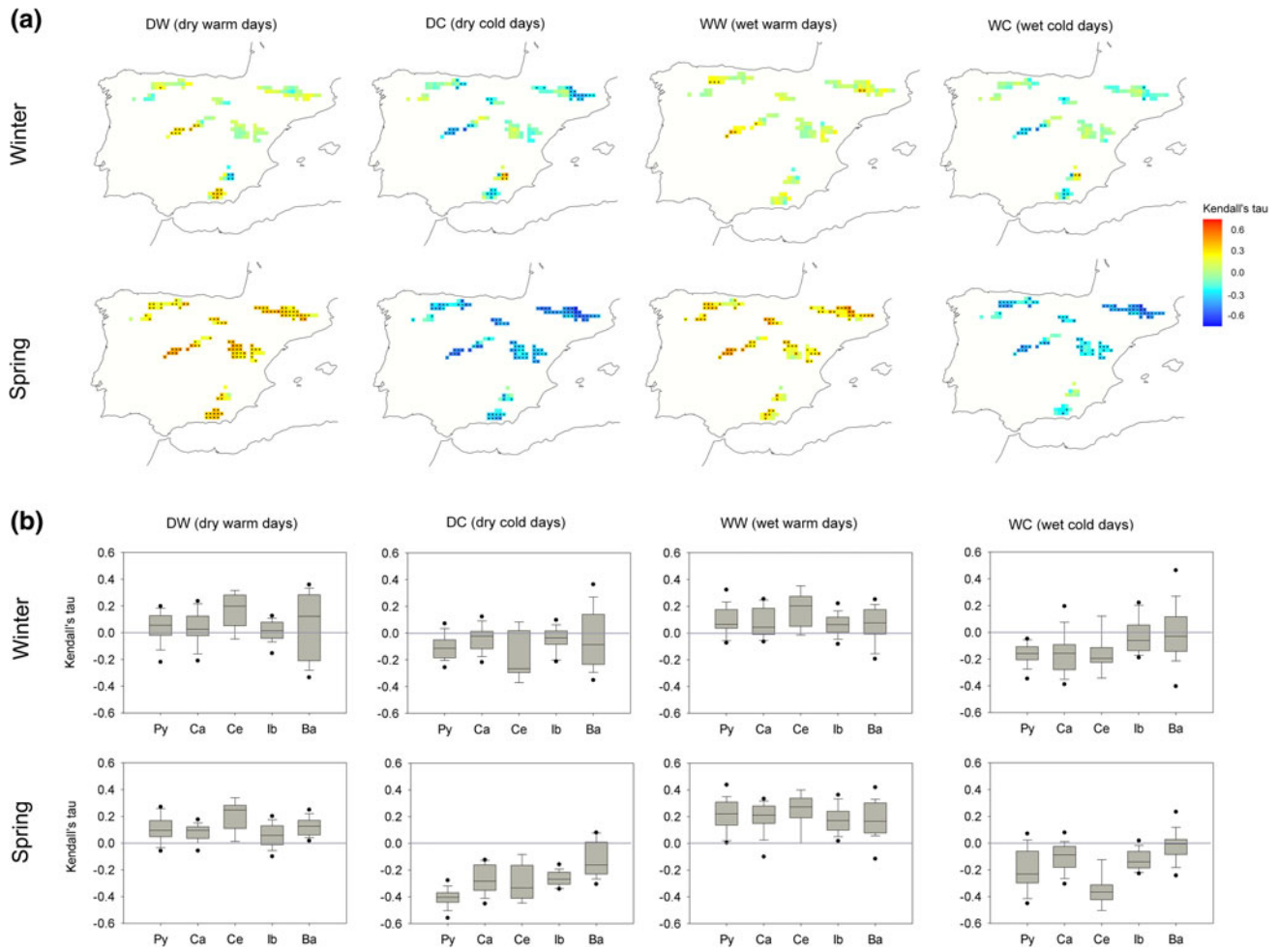


Fig. 4 Trends (Kendall's coefficient) in the joint-quantile indices in winter and spring. **a** Warm (cool) *colors* correspond to positive (negative) trends. *Black dots* indicate significant trend at the 95 % confidence level

significant trends in all mountain systems, indicating a sharp increase in the number of days with combined dry and warm conditions. Dry and cold (DC) days showed a similar behavior, but with trends in the opposite direction. The behavior of the wet modes is pretty similar to that of the dry modes, especially for the wet and warm (WW) days, with weak and inhomogeneous signals in winter and strong positive trends in spring. In the case of wet and cold (WC) days, a negative signal is observed for most mountains, although trends are more pronounced and significant in spring than in winter. Spatially, the Pyrenees, Central System and western Cantabrian range recorded the most pronounced decrease in wet and cold conditions. The Iberian and Baetic systems show, in contrast, more attenuated signals of change.

We calculated as well the amount of change in the indices derived from the linear trend. Two coefficients were extracted from this analysis: the difference in the number of days of each index between the end and beginning of the period (n) and the fold change (f), which is the ratio between the values at the end and at the beginning of the period, and provides insight about the relative magnitude of the change. In Table 3, a summary with the averages n and f for the pixels corresponding to each mountain system is shown. The greatest change is found in the Pyrenees for the cold modes, especially in spring, with a decrease in nearly 10 days in DC and 4.6 days in WC, and a 91-fold (this excessive amount of change is derived by dividing a value of 9.3 days at the beginning of the period by 0.1 days at the end of the period; these data are not shown) and 16-fold change, respectively. The decrease in DC days in spring is the greatest among the indices in all the mountain systems, with a 7-fold decrease in the Cantabrian range or an 8.6-fold decrease in the Iberian System. Wet-cold days have also experienced substantial decreases especially in spring, with less than half (fold change <-2.0) days with WC conditions at the end of the period with respect to the beginning in all mountain systems except in Baetic system (-1.7-fold change).

The homogeneous signal of increase in the two "warm" modes and decrease in the two "cold" modes a priori

Table 3 Change in the average (by mountain range) exceedance series of the combined precipitation–temperature indices

Index	Pyrenees			Cantabrian range			Central system			Iberian system			Baetic system		
	n	*f*	Sig	*n*	*f*	Sig	*n*	*f*	Sig	*n*	*f*	Sig	*n*	*f*	Sig
DW															
Winter	−0.2	−1.1		−0.1	−1.0		3.2	3.5		0.0	1.0		0.8	1.2	
Spring	15.3	1.7	*	9.9	1.4	*	17.8	1.8	*	12.6	1.5	*	18.2	1.7	*
DC															
Winter	−10.2	−1.5	*	−2.8	−1.1		−11.7	−1.6	*	−5.2	−1.2		−3.9	−1.1	
Spring	−9.2	−91.6	*	−5.6	−7.0	*	−7.5	−6.2	*	−7.8	−8.6	*	−6.5	−4.3	*
WW															
Winter	0.2	2.0		0.4	1.3		0.5	2.5		0.0	−1.1		0.1	1.6	
Spring	6.0	1.9	*	5.4	1.9	*	6.2	2.5	*	4.5	1.8	*	3.0	2.5	*
WC															
Winter	−1.6	−1.2		−0.1	−1.0		−3.1	−1.5		−1.0	−1.1		−1.8	−1.2	
Spring	−4.6	−16.2	*	−2.2	−3.6	*	−2.0	−2.5	*	−2.8	−3.2	*	−1.3	−1.7	

n = number of days (absolute); *f* = fold change (relative)

* indicates significant with 95 % of confidence

indicates that the temperatures are the driving factor of change in climatic conditions. Brunet et al. (2007) found that the increases in temperature in Spain from 1973 onwards was more pronounced in spring than in winter which is consistent with our findings. A number of studies have previously explored the trends in extreme quantiles of the temperatures and precipitation for different time periods during the twentieth century for Europe. Although most studies differ in the period or weather stations analyzed, most of them report significant and spatially consistent trends in the extreme temperatures, while for precipitation more uncertainty is observed (e.g., Klein and Können 2003; Moberg and Jones 2005; Alexander et al. 2006). A recent study by Beniston (2009), in which the "joint-quantile" technique was applied for various European sites for the twentieth century, showed a similar signal to that observed in this work, with increasing (decreasing) trends in the warm (cold) modes of the studied indices. The author, however, analyzed in depth the internal behavior of temperatures and precipitation when the different modes were exceeded and concluded that the observed trends were not only explained by the evolution of temperatures. He demonstrated, in fact, that the average amount of precipitation had decreased during the wet-cold modes and increased during the wet-warm modes, while overall the mean annual amount of precipitation remained stationary. The extended geographical scope of the cited work, covering the Iberian Peninsula, suggests that this could be the most likely behavior for our selected areas. Further research should, however, verify this hypothesis.

The averaged series of the different indices for every mountain chain were plotted to visualize the inner characteristics of the detected trends (Fig. 5). It is observed that the evolution of the indices exhibited similar dynamics in most mountain ranges. Some common features that can be observed include: the sustained increase in spring DW that peaked in the late 90s and the later decrease until the end of the series; or the change points in the middle of the 1980s and the middle of the 1990s for DC days. In general, there is an appreciable change of tendency during the last years of the record, namely a decrease in warm modes (while the general trend is to increase) and an increase in the cold modes (while the general trend is to decrease). These similarities suggest that the specific weather conditions that lead to each particular mode are similar over most of the continental Spain. This is somehow controversial: it is well known that climate conditions in the Iberian Peninsula are generally governed by the position and strength of the Azores High, allowing/impeding the penetration of disturbances associated with Atlantic fronts, which are closely related to the behavior of the North Atlantic Oscillation (NAO) (Martín-Vide and Fernández 2001; Hurrell et al. 2003; Trigo et al. 2004); however, recent research has highlighted that especially for the eastern half of the Iberian Peninsula, other atmospheric patterns such as the Western Mediterranean Oscillation (Martin-Vide and López-Bustins 2006) also contribute significantly to regional weather patterns. Recently López-Moreno et al. (2011) concluded that the NAO is a sound indicator of the variability of winter types favorable/unfavorable for snow accumulation in the west-Mediterranean mountains. In this work, spatial homogeneity is seen, but it is evident that the Pyrenees and the western sector of the Central System show the strongest signal of change. These spatial differences leave some open questions that merit to be further explored, such as the weather/synoptic conditions

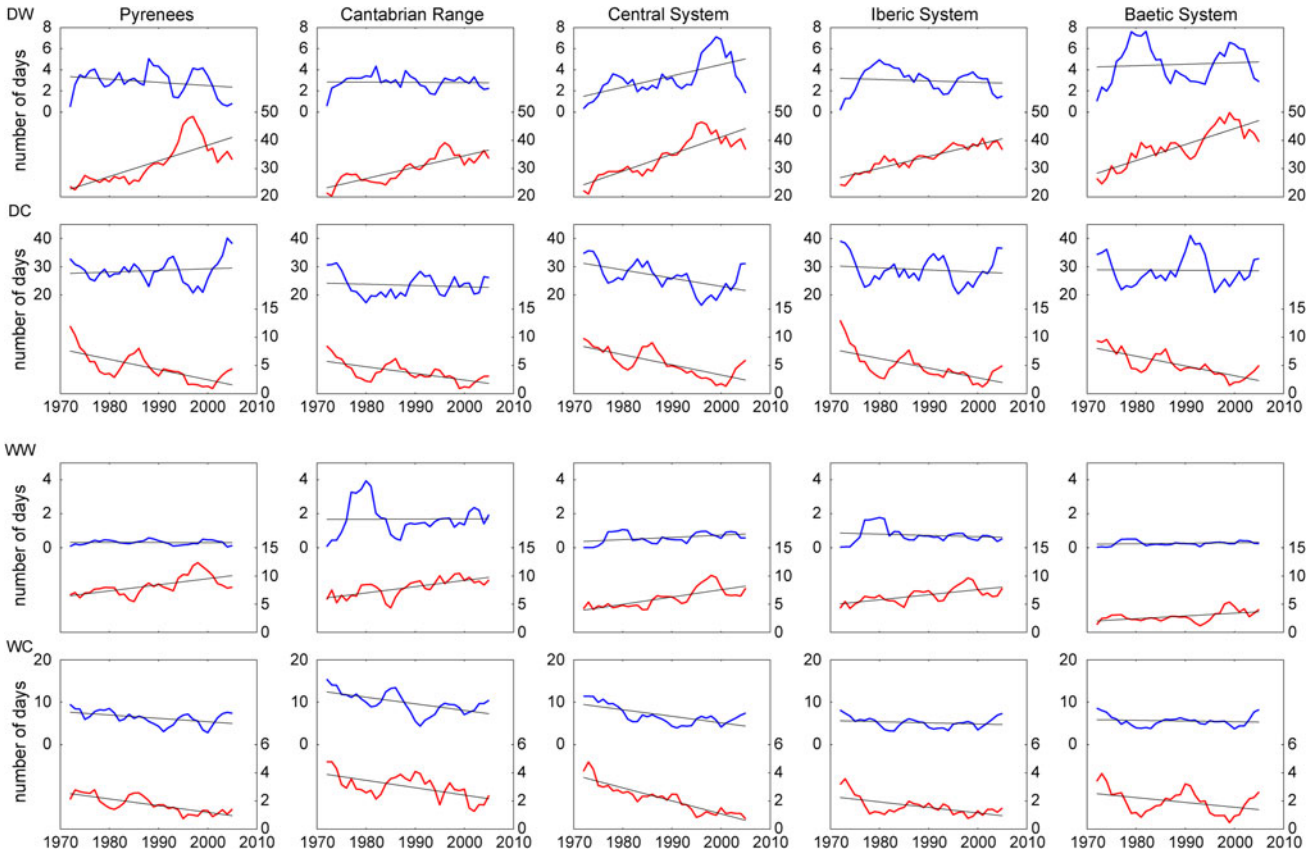


Fig. 5 Plots of the average evolution of the combined indices in winter (*blue line*) and spring (*red line*), for each mountain chain. The time series were smoothed with a 5-year moving average in order to reduce noise and enhance possible cycles

associated with the occurrence of the different precipitation–temperature modes studied here, or the relationship of the studied indices with the various teleconnections affecting the Iberian Peninsula, not only with the North Atlantic Oscillation (NAO), but also the Eastern-Atlantic Pattern (EA), the Scandinavian Pattern (SCA) or the Western Mediterranean Oscillation (WEMO).

Trends in the evolution of the average duration of spells and of the spells of maximum duration per year were as well calculated (Fig. 6). In general, the pattern of increase (decrease) in warm (cold) modes is also observed for the duration of spells, although in this case trends are not as robust as for the case of the evolution in the number of days. The most significant and generalized trends are observed for DC (negative) and WW (positive) spells in spring. The Pyrenees experienced once again the greatest trends, especially in DC. The general negative evolution in the duration of spells with WW conditions is also noticeable.

Influence of climatic evolution on snow accumulation

The second objective of the study was to assess the evolution of the snowpack in the studied mountains and its sensitivity to the evolution of the climatic indices. Various drawbacks related with data quality made this objective difficult to achieve, although the results next presented enabled conclusions to be drawn.

Spring snow depth in the Pyrenees has decreased between the time in which measurements began and recent years (Fig. 7a). The average series constructed with data from the 84 snow poles shows a variable evolution in snow depth, but a negative linear trend is superimposed to this variability. The negative evolution seems more pronounced in the first decade, whereas from 1997 onwards, there is no clear tendency. This is a first sign of coincidence with the evolution of the climatic indices, which in general presented a change of tendency in the last decade of the studied period. In spite of this general evolution, large variability among cases exists, as expressed by the amplitude of the inter-quartile range throughout the period. A principal component analysis with the 84 snow-depth series as input variables allowed us to obtain the most common patterns of evolution of snow depth, therefore reducing the initial inter-cases variability (Fig. 7b). We thus observed the first two components (which together explain about 60 % of the original variance) with a clear decrease in snow depth, although with different dynamics in the

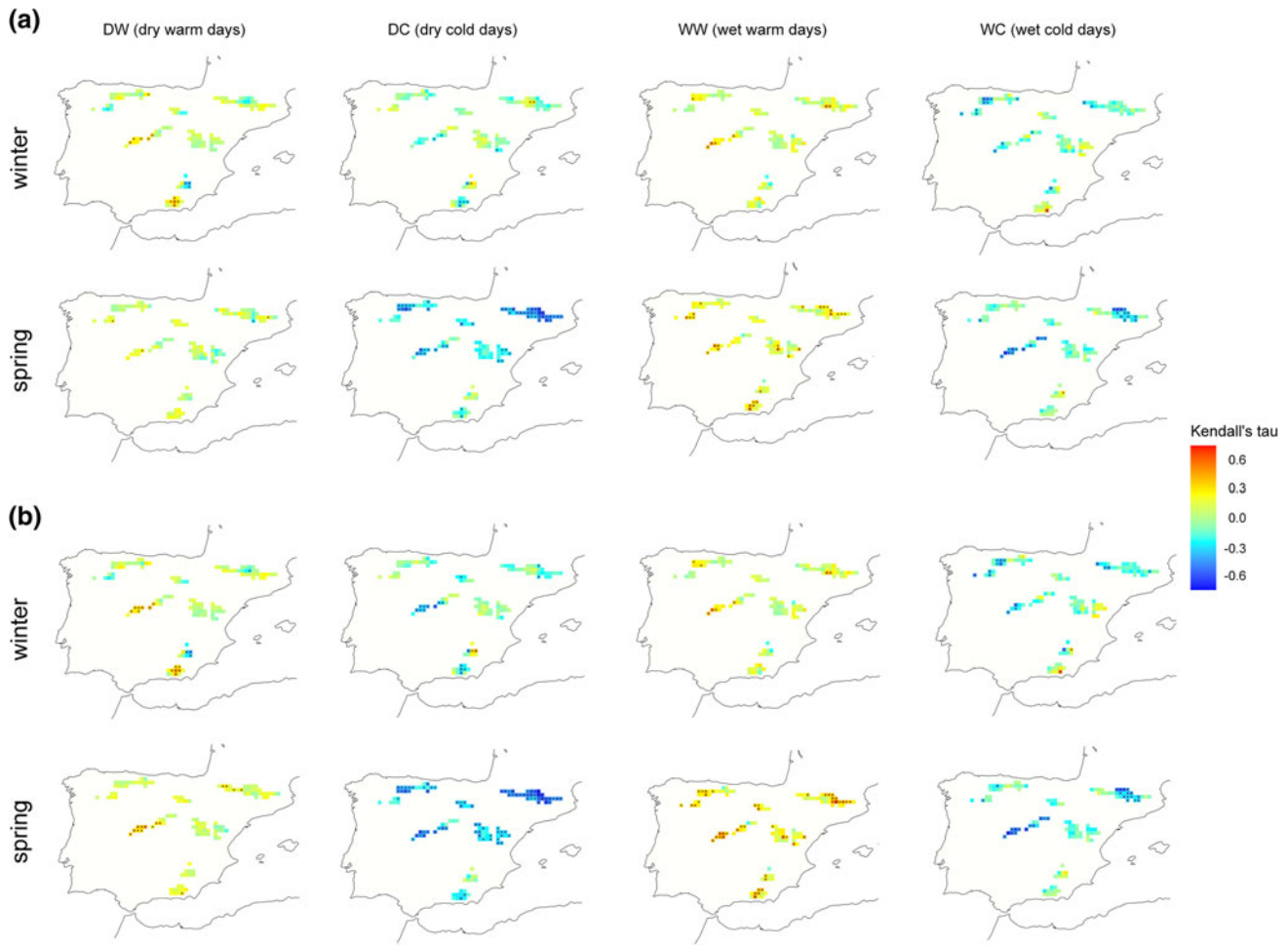


Fig. 6 Trends (Kendall's coefficient) in the mean duration of spells **a** and in the duration of the longest spells **b** with combined climatic conditions. Warm (cool) *colors* correspond to positive (negative) trends. *Black dots* indicate significant trend at 95 % of confidence

evolution of snow. In contrast, the last two components (with quite a smaller proportion of the explained variance) showed no clear signal of trends.

In order to asses to what extent the snow-depth evolution is sensitive to the evolution of the studied climatic indices, we have performed Pearson's correlations between each index in each of the 0.2° resolution pixels of the Pyrenees ($n = 31$), and the standardized (factorial scores) series of snow depth resulting from each principal component, for the common period 1986–2007. Results (Fig. 8a) show that the first principal component (PC1) is mostly anti-correlated with the evolution of DW days in spring and, to a lesser extent (and mostly not significantly), correlated with the evolution of DC days in spring. Otherwise, indices were seen to lie within the non-significance intervals. The evolution of snow depth for PC2 shows a significant and generalized correlation with DC in spring and also with the evolution of WC days in spring. The third and fourth principal components (which represent a minority of snow poles) do not show, in general, any significant correlations with the evolution of the climatic indices. Only the PC4 shows a pattern of negative correlations with the "warm modes" and positive correlations with the "cold modes," but in the majority of cases correlations are not statistically significant. Local particularities of each sampled site (e.g., topography, solar radiation, wind, exposure) may determine these differences in the sensitivity of the snowpack to climate conditions. Figure 8b graphically shows that abundant (sparse) snow springs coincide with the years of low (high) frequency of DW days for PC1 and with years of high (low) frequency of DC days for PC2, although the latter are less evident.

These results are in agreement with those by Beniston et al. (2011) in the Alps, where the snow depth correlated negatively with the evolution of dry-warm days, that is, with the conditions that favor the disappearance of snow in spring. It must be highlighted that the coefficients of correlation obtained, when significant, were not of great magnitude in the majority of cases. We believe that the explanation for the low correlation coefficients has to do

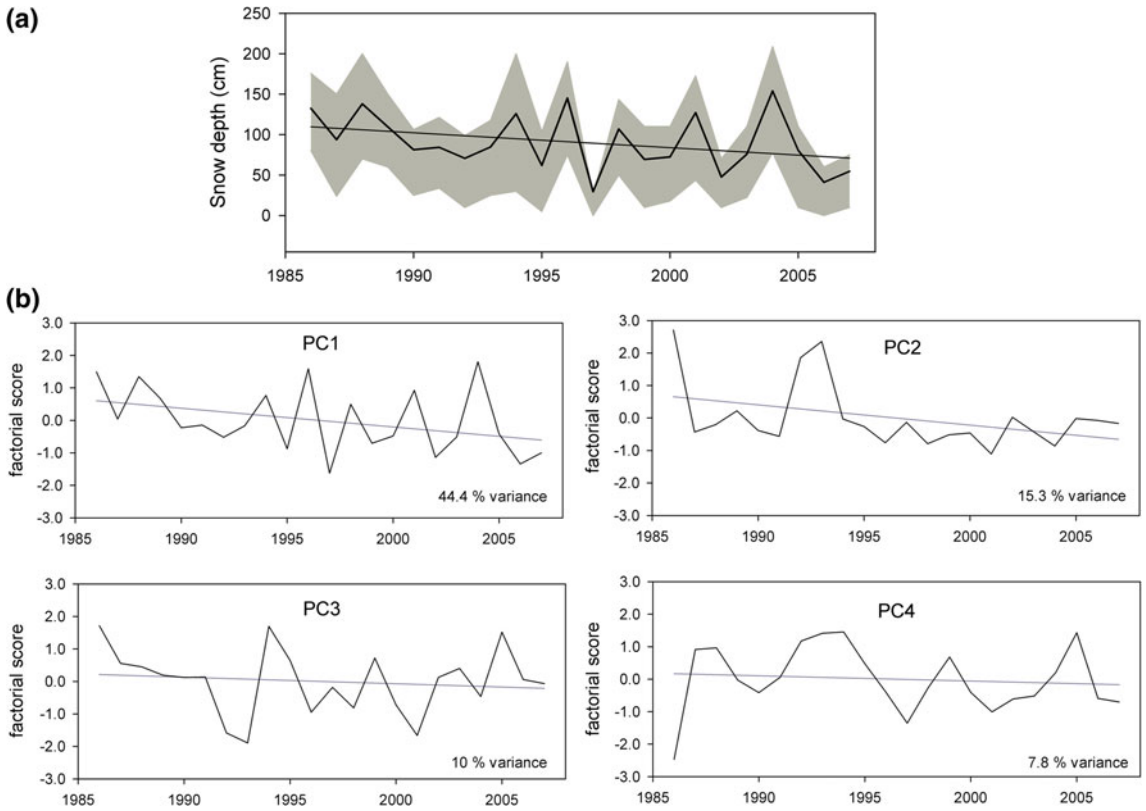


Fig. 7 Evolution of the snow depth in the Spanish Pyrenees (1986–2007). **a** Regional series of snow depth showing the average (*black line*) and the inter-quartile range (*gray shade*). **b** Series of the four first principal components

with the characteristics of the snow-depth data. As explained before, the series of snow depth used in the study correspond with a measurement per year, taken in late April/early May. This value of snow depth is representative of the accumulated snow along the winter/spring, but may be as well masking a history of accumulation/melting related to the specific weather conditions during the year at the sampled site. This might be the explanation, as well, for the fact that snow depth is basically controlled by temperature conditions in spring, whereas the precipitation appears not to be a limiting factor. Daily snow series, such as those used by Beniston et al. (2011) for the Alps, would probably provide better results, as they allow computing statistics such as the mean or standard deviation of snow depth, as well as the duration of the snowpack. Having into account these limitations, we find that the patterns of correlation found reproduce a signal of sensitivity of the snowpack to the evolution of the studied indices. Although snow data in the other studied mountains did not enable a similar analysis, it is expected a similar sensitivity of the snowpack to that observed in the Pyrenees.

A reduction in the snowpack in a warmer climate is likely to take place, given the displacement in altitude and latitude of the 0 °C isotherm. This is, however, difficult to demonstrate directly, due to the lack of continuous monitoring systems of climatic variables and snow depth in mountainous areas. To overcome this, the use of climate indices as proxies has been proven to yield reliable results in several parts of the world. In the western US and Canada, Hamlet et al. (2005) observed downward trends in the snow water equivalent (a measure of the quantity of water contained in a given volume of snow) during the twentieth century, which were in the long term explained by warming temperatures and, on a decadal time scale, by the variability of precipitation. In a study of snow evolution in the Pyrenees, López-Moreno et al. (2005) demonstrated that precipitation of previous months explained the evolution of the snowpack in winter and spring, whereas temperatures partly explained the evolution of snowpack only in spring. They also found that temperatures correlated better with snow depth at low elevations (1,700–2,200 m) than at higher elevations, which implies that there is a certain threshold of elevation for which the observed warming does not yet influence the evolution of the snowpack. Moreover, Pons et al. (2009) demonstrated that the decrease in the occurrence of "snow days" in northern Spain during 1950–1999 was related to increasing temperatures, although precipitation played a role as well, especially at high elevation sites.

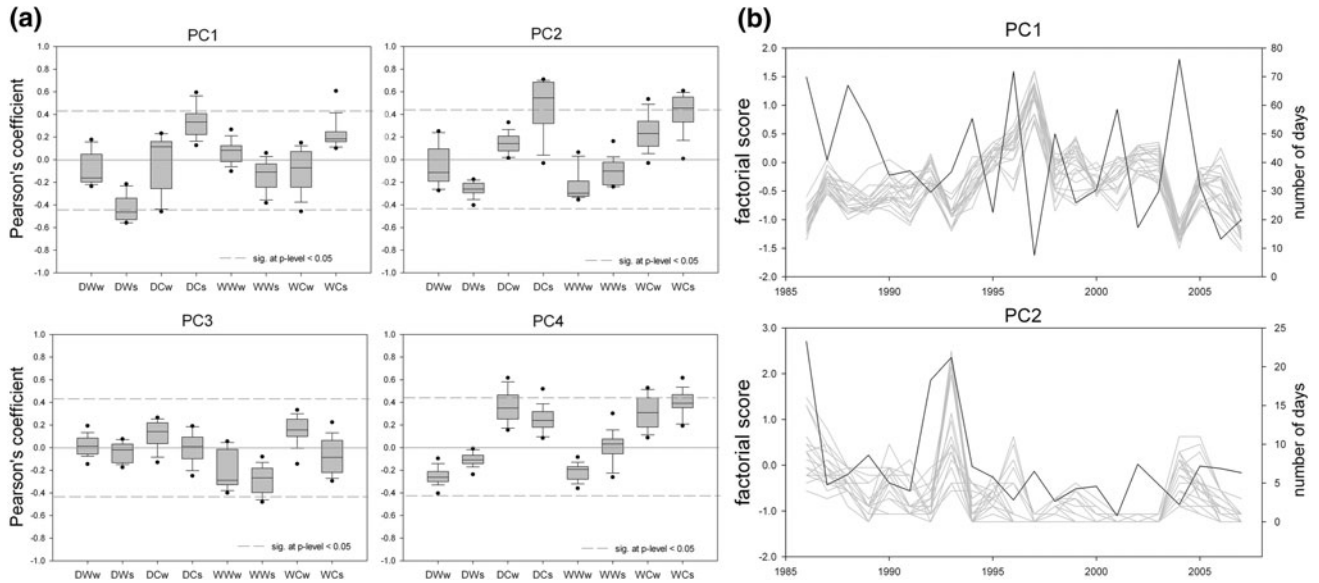


Fig. 8 Correlation between climatic indices and snow depth (1986–2007). **a** Boxplots showing the variability in the Pearson coefficients between the indices in each of the pixels of the Pyrenees and the series of each of the four principal components displayed in Fig. 7b). Series of the two first principal components of snow evolution (*black line*) and the index in the Pyrenees pixels that best correlates with snow evolution (*gray lines*), that is, DW days in spring for the PC1 and DC days in spring for PC2

These observations highlight the need for an in-depth understanding of the relationships between snowpack and climatic variables, in order to better simulate future projections of snow evolution in a "greenhouse climate." For the Swiss Alps, Uhlmann et al. (2009) and Beniston et al. (2011) projected a reduction in the snowpack and a much lower frequency of snow-abundant winters in the course of the twenty-first century in an enhanced greenhouse climate; López-Moreno et al. (2011) predicted a dominance of "warm" winter modes, with a rare occurrence of "cold" winters, for a moderate warming scenario (A1B1, IPCC 2007), in Mediterranean mountains between 2010 and 2100. Based on these precedents, it is reasonable to believe that a reduction in the snow cover is likely to take place in the Iberian mountains under conditions of climatic warming over coming decades.

Conclusions

A noteworthy change in climatic conditions was detected throughout the different mountain systems during the 1970–2007 period. This involved a significant increase in the number of dry-warm and wet-warm days, and also in the duration of spells with identical conditions, particularly in spring, and the respective decrease in the cold modes. The increase in temperatures is therefore the main driver of change in these climatic conditions. The signal of change was homogeneous in the territory, but the magnitude of trends was greater in the Pyrenees and smaller in the Baetic System.

Snowpack in the Pyrenees showed a reduction in depth during the last two decades and in the majority of sampled sites the snowpack evolution was negatively correlated with the frequency of dry-warm days, that is, with the conditions favorable to snow melting. There was in addition, for other cases, a significant correlation with the frequency of wet-cold and dry-cold days, that is, conditions that favor the snowfall and snow persistence. These results confirmed the hypothesis of the sensitivity of the snowpack to changes in the joint precipitation–temperature modes.

This work constitutes the first approach that relates the evolution of climatic conditions favorable or unfavorable for snow accumulation, and the evolution of the snowpack in Iberian mountains. Although results provide evidence of a change in climatic conditions and a certain sensitivity of the snowpack to these, a series of questions have arisen out of the present research that include: the uncertain role of precipitation in the change of joint-quantile modes; the synoptic conditions leading to each particular mode, and the relation with large-scale teleconnection patterns that affect the Iberian Peninsula; or the future evolution of the studied indices and the snowpack during the course of the twenty-first century, in an enhanced greenhouse climate. Such topics would justify future work to attempt to clarify many of these issues.

Acknowledgments This work has been possible thanks to the financial support of the Spanish Government (Ministry of Education)

through the postdoctoral program "Ayudas de movilidad postdoctoral en centros extranjeros (Orden EDU/2728/2011, de 29 de septiembre)." The authors would like to give special thanks to Fernando Pastor Argüello, from the ERHIN program (Ministry of Environment of Spain), for providing the snow thickness data.

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From: Evans Mack,Diane
To: Grizzle, Betty (betty_grizzle@fws.gov)
Subject: FW: Recent wolverine trapping
Date: Monday, February 5, 2018 3:13:31 PM
Attachments: image002.png
SPTR-ENF-KM18020514550.pdf

Hi Betty –
Attached is a summary of the incident, with reference to the trapper's name blacked out. This document (with black-outs) came from our Wildlife Bureau in IDFG Headquarters in Boise. Hope this covers the info you're looking for.

Diane

Diane Evans Mack
Regional Wildlife Diversity Biologist
Idaho Department of Fish and Game
McCall Subregion
555 Deinhard Lane
McCall, ID 83638
208-634-8137
diane.evansmack@idfg.idaho.gov

From: Grizzle, Betty [mailto:betty_grizzle@fws.gov]
Sent: Thursday, February 01, 2018 8:32 AM
To: Evans Mack,Diane
Subject: Recent wolverine trapping

Hi Diane - Can you provide information on the female wolverine that was reportedly trapped in GMU21A in December 2017 (which resulted in mortality)? We received a copy of the letter sent to your agency from the Center of Biological Diversity (dated January 16, 2018), but was hoping you could provide additional information (e.g., more specific location).
Thanks,
Betty

--

Betty J. Grizzle, D.Env., MPH
Fish and Wildlife Biologist
U.S. Fish and Wildlife Service
Carlsbad Fish and Wildlife Office
2177 Salk Ave, Suite 250
Carlsbad, CA 92008
760-431-9440, ext. 215
760-431-5901 fax


IDAHO
FISH & GAME

MEMORANDUM

Date: December 12, 2017

To: Sharon Kiefer, Scott Reinecker, Mike Keckler, Kathleen Trever

Cc: Tom Curet, David Silcock, Greg Painter

From: Beth Waterbury, Salmon Region Wildlife Diversity Biologist

Subj: Incidental capture of wolverine in GMU 21A, Fourth of July Creek drainage, Lemhi County

This memo summarizes knowledge to date on the incidental capture of an adult wolverine in GMU 21A on approximately Friday, December 8, 2017. The wolverine was captured in the Fourth of July Creek drainage about 18 miles N-NE of Salmon in Lemhi County, ID (Fig. 1). The trapper reported the nontarget capture to IDFG Salmon Region's voicemail on Sunday, December 10, 2017. The trapper is [redacted]

Nontarget Capture Incident

In [redacted] voicemail message, he stated he made a nontarget catch of a wolverine on Friday, December 8, 2017 in Fourth of July Creek. He said he would store the wolverine in his shop freezer in case IDFG personnel wanted to drop by to retrieve it. He left his phone number for follow up. He reported the capture within the 72-hour mandatory reporting period for dead nontargets.

I called [redacted] Monday morning at about 8:30 and spoke with [redacted]. [redacted] said [redacted] wasn't available and reiterated the information from [redacted] voicemail message. [redacted] said [redacted] was coming into Salmon later in the day and could drop off the wolverine carcass. I told [redacted] that would be fine and that I would like to further discuss the incident with [redacted]

[redacted] stopped by the Salmon Region office at around 1:30 pm on Monday, December 11, 2017 to transfer the wolverine and complete a BGMR (C127058). [redacted] provided the following additional information about the incident:

- The wolverine was captured in a snare targeting "mountain" fox. Snare height was 10 inches and snare loop was 8 inches. Bait of beef scraps was placed 50 yards from the snare set. There was no stop on the snare loop. Snare was set at approximately 6,500 ft. elevation.
- Location coordinates are approximately: 45.426625° N / -113.806603° W
- [redacted] last checked the snare 3 days (72-hours) prior to Friday, Dec. 8. Given the semi-frozen condition of the wolverine when he found it on Dec. 8, he estimated its date of death was the previous Wednesday night or Thursday morning.
- [redacted] said he has snared several fox at this site since deploying the set in mid-November.
- A few weeks ago, he had a wolverine consume an American marten at one of his marten sets deployed at higher elevation on the divide between Fourth of July and the Sheep Creek drainage to the north. I have high confidence in [redacted] ability to identify wolverine tracks and sign.
- [redacted] said he pulled all his traps in the Fourth of July Creek drainage as a precaution to avoid additional nontarget captures.

I told [redacted] he is eligible to receive a $10 reimbursement for turning in his nontarget captured wolverine. I did not pay him the $10 at the time of his visit, as I wasn't sure how these funds are

disbursed (IDFG check or out of the Salmon Region cash register). I told [redacted] I would find out and follow up with [redacted]

Wolverine Info

The wolverine carcass was received intact and appears to be an animal in good condition and in full winter coat (Figs. 2-4). It is an adult female weighing 18.25 pounds, which is on the low end of the range for females. Body length = 29.5 inches; tail length = 7.25 inches (body length ranges from 26-41 inches; tail length ranges from 5-10 inches).

There is a pronounced groove around the animal's neck where the snare constricted. The teeth appear worn, suggesting a mature or old animal. The upper left canine is sheared off at the gums and 2 upper left incisors are missing; loss of these teeth is likely an older injury/trauma (Fig. 5). Each of the other 3 canines is broken at the tip; these breaks may have occurred when the animal became ensnared, as breaks have sharp edges and are unstained. Reproductive status was undetermined, but pregnancy and past births could be determined through necropsy.

The markings on this female are very unique and striking (Fig. 6). The coat is marked with large patches of white fur, particularly on the right front quarter. Toes on all 4 feet are white. Based on its unique markings, it is highly probable this is the same individual captured on camera (Fig. 7) and video near Ajax Peak, on the Montana side of the Beaverhead Mountains over 4th of July weekend in 2017. Ajax Peak is about 8 air miles N-NW of the site the wolverine was snared. In May 2016, we documented another wolverine in the vicinity (Sheep Creek drainage) by trail camera with similar "white-legged" markings, though on its left leg (Fig. 7). We suspect these 2 animals are siblings or otherwise members of a family group carrying this genetic trait.

The wolverine was snared in an area of known and relatively frequent wolverine occurrence. In 2008, we documented a wolverine maternal den about 2 miles from the snare site. We have collected and confirmed wolverine DNA from tracks in this and adjacent drainages over the past 8 years. We recognize the west flank of the Beaverhead Mountains as an important travel corridor for wolverines, linking higher elevation breeding habitats with mid and lower elevation foraging habitats (i.e., deer/elk/sheep/ pronghorn carrion on winter ranges). The crest and upper flanks of the Beaverhead Mountains, including the site the wolverine was snared (~6,500 ft. elevation), are within high suitability modeled habitat for wolverines.

Salmon Region plans to have this wolverine taxidermy mounted for public display. Tissue samples will be shipped to Rocky Mountain Research Station to contribute to the Northern Rocky Mountains wolverine DNA database. The carcass will be delivered to the IDFG Wildlife Health Lab for necropsy. Please let us know if you have any further instructions for the disposition of this animal. You can reach me at 208-940-2754 or beth.waterbury@idfg.idaho.gov.

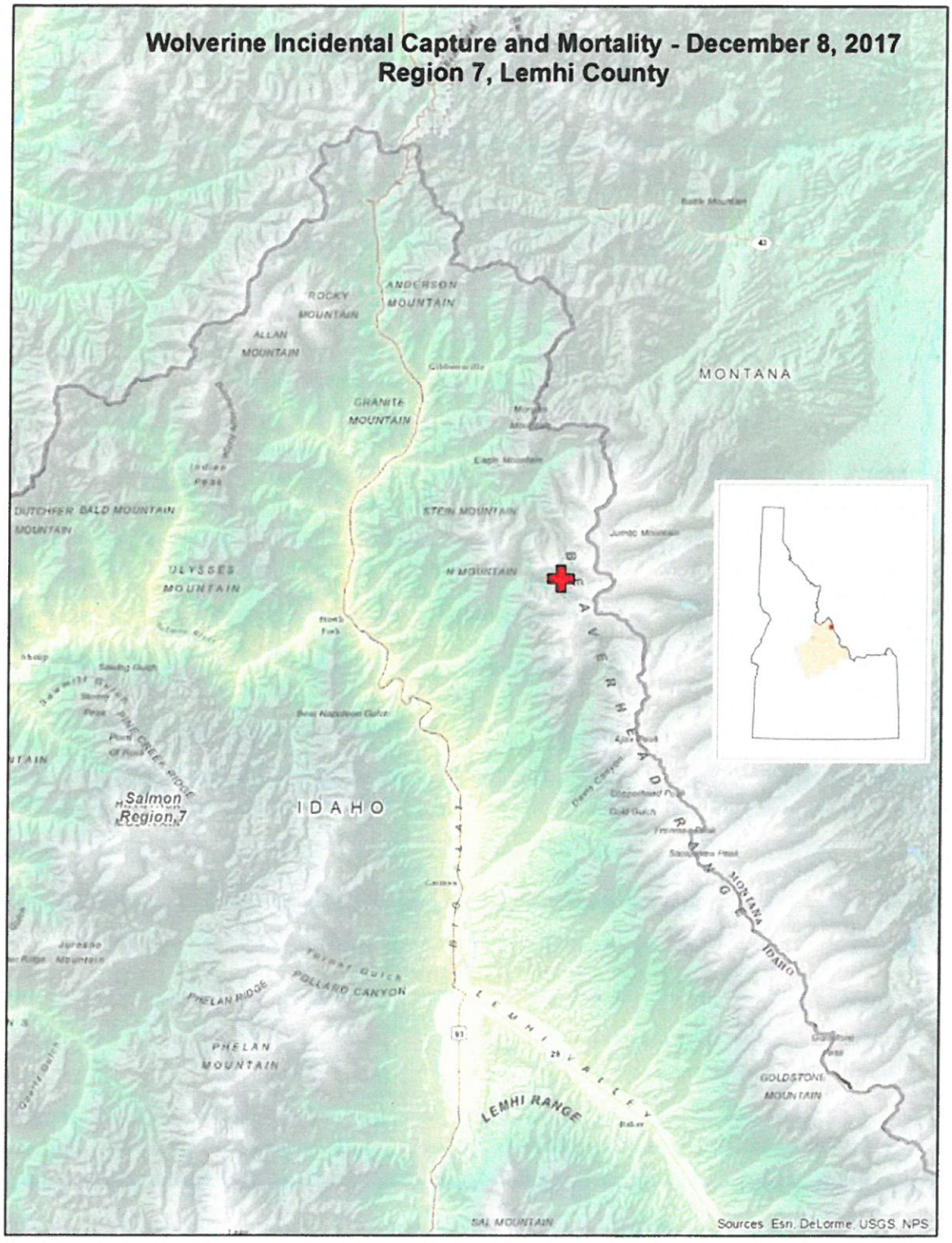


Figure 1. Location of wolverine incidental capture/mortality in Salmon Region, upper Fourth of July Creek drainage, Lemhi County, Idaho.



Figure 2



Figure 3



Figure 4



Figure 5



Figure 6

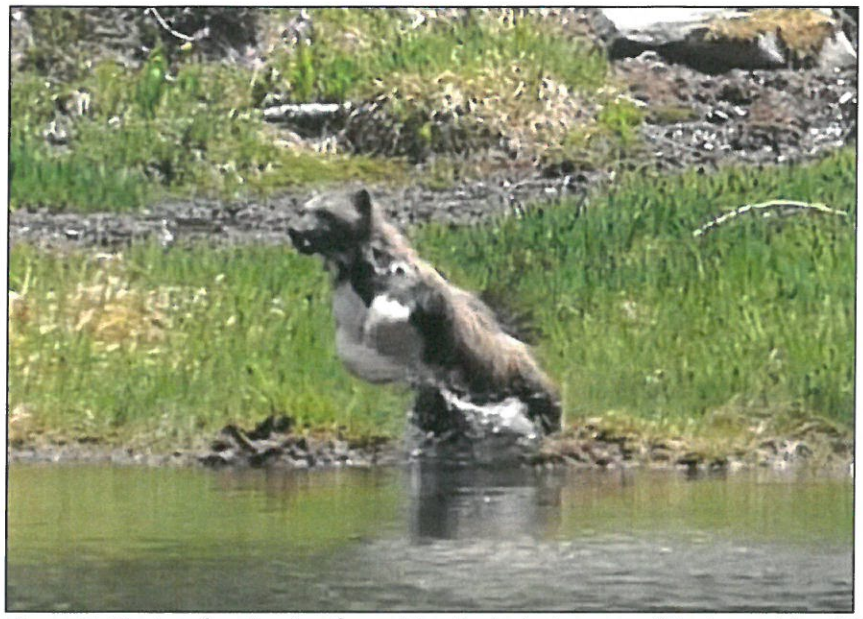


Figure 7. Photo of wolverine from Ajax Peak drainage on Montana side of Continental Divide, 4th of July weekend, 2017 (still from video). Coat pattern is entirely consistent with patterns in Figs. 2-6.



Figure 8. Photo of wolverine taken from bear bait station trail camera in Sheep Creek drainage on May 11, 2016, Beaverhead Mountains, Lemhi County. Note "white sock" on left foot.

From: Grizzle, Betty
To: Evans Mack,Diane
Subject: Re: FW: Recent wolverine trapping
Date: Monday, February 5, 2018 3:31:12 PM
Attachments: image002.png

Thanks, Diane.

On Mon, Feb 5, 2018 at 2:12 PM, Evans Mack,Diane <diane.evansmack@idfg.idaho.gov> wrote:

> Hi Betty –
>
> Attached is a summary of the incident, with reference to the trapper's name blacked out. This document (with black-outs) came from our Wildlife Bureau in IDFG Headquarters in Boise. Hope this covers the info you're looking for.
>
> Diane
>
> Diane Evans Mack
> Regional Wildlife Diversity Biologist
>
> Idaho Department of Fish and Game
>
> McCall Subregion
> 555 Deinhard Lane
> McCall, ID 83638
> 208-634-8137
>
> diane.evansmack@idfg.idaho.gov
>
> 
>
>
> _____
>
> **From:** Grizzle, Betty [mailto:betty_grizzle@fws.gov]
> **Sent:** Thursday, February 01, 2018 8:32 AM
> **To:** Evans Mack,Diane
> **Subject:** Recent wolverine trapping
>
> Hi Diane - Can you provide information on the female wolverine that was reportedly

trapped in GMU21A in December 2017 (which resulted in mortality)? We received a copy of the letter sent to your agency from the Center of Biological Diversity (dated January 16, 2018), but was hoping you could provide additional information (e.g., more specific location).

Thanks,
Betty

--

Betty J. Grizzle, D.Env., MPH

Fish and Wildlife Biologist

U.S. Fish and Wildlife Service

Carlsbad Fish and Wildlife Office

2177 Salk Ave, Suite 250

Carlsbad, CA 92008

760-431-9440, ext. 215

760-431-5901 fax

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Betty J. Grizzle, D.Env., MPH
Fish and Wildlife Biologist
U.S. Fish and Wildlife Service
Carlsbad Fish and Wildlife Office
2177 Salk Ave, Suite 250
Carlsbad, CA 92008
760-431-9440, ext. 215
760-431-5901 fax

From: Shoemaker, Justin
To: Kit Hershey
Subject: Fwd: Wolverine Draft SSA for Tribal Review
Date: Tuesday, February 6, 2018 7:12:58 AM

Example for tribes.

Justin Shoemaker
Classification and Recovery Biologist
U.S. Fish and Wildlife Service, Region 6
Phone: 309-757-5800 x214
Email: justin_shoemaker@fws.gov

---------- Forwarded message ----------
From: **Bush, Jodi** <jodi_bush@fws.gov>
Date: Thu, Oct 26, 2017 at 9:43 AM
Subject: Wolverine Draft SSA for Tribal Review
To: Anna Munoz <anna_munoz@fws.gov>
Cc: Justin Shoemaker <justin_shoemaker@fws.gov>, Betty Grizzle <betty_grizzle@fws.gov>

Anna. Here is the email to share with our tribal partners. Attached is the SSA document for review. Please have the tribal liasons cc Betty Grizzle on their emails so we can maintain the administrative record. Thanks. JB

Dear Tribal Partners:

Attached please find the U.S. Fish and Wildlife Service's DRAFT Species Status Assessment (SSA) for the Wolverine - Contiguous United States Distinct Population Segment (DPS). The SSA is intended to provide the biological and scientific underpinnings for all decisions the Service must make in accordance with the Endangered Species Act (Act). The draft report will undergo concurrent peer review and review by State Fish and Wildlife Agencies and by Federal land management agencies (BLM, NPS, and USFS) within the DPS range.

The Service jointly respects and values the significant role of Indian Tribes in past and ongoing species conservation. We also respect the sovereignty of Tribal governments and our collective Trust responsibility to Tribes. Continuing this effective relationship with interested Tribes and others is essential to achieving conservation of species including wolverine. Therefore, we are providing this draft for review by members of your organization with expert knowledge of the species and its habitat. That review will help us ensure that we have appropriately considered the best scientific and commercial information when evaluating the current status and future viability of the wolverine DPS.

We request your organization's independent scientific perspectives on the comprehensiveness and logic of the document, as well as how well the technical conclusions are supported by the data and analyses. We ask that your comments on the draft SSA report focus specifically on whether the best available information was used, the quality of the scientific information, and our interpretation and analyses of the data with regard to the species' viability in the contiguous United States. We request that you direct your review to the scientific issues and assumptions related to your expertise.

We welcome consolidated comments from your organization on the Draft SSA by November 30, 2017. Please send comments by that date to betty_grizzle@fws.gov. Thank you for your interest and assistance.

The Draft SSA is not intended to solicit public comment and will be revised after this scientific review. The Draft SSA does not predetermine any future agency decision under the Endangered Species Act.

General Information about SSAs:

The Species Status Assessment framework is a new tool the U.S. Fish and Wildlife Service is using to improve transparency

while conducting listing determinations and other actions in accordance with the Act, and peer review of our analyses of the viability of species is part of that new process. The attached draft SSA report is a rough draft; we are seeking comments at this stage to ensure that we have time to incorporate any substantial comments as we finalize the report.

In reviewing the document, please note that this draft SSA report does not result in or predetermine a decision by the Service on whether the Canada lynx warrants protections of the Act. This document is strictly a characterization of the species' viability in the contiguous United States. As a reminder, all reviews and comments submitted to the Service will become public documents and part of our administrative record for this document.

Thank you.

Jodi L. Bush
Office Supervisor
Montana State Ecological Services Office
585 Shepard Way, Suite 1
Helena, MT 59601
(406) 449-5225, ext.205

From: Shoemaker, Justin
To: Caitlin Snyder
Subject: Wolverine revised timeline
Date: Wednesday, February 7, 2018 3:06:50 PM
Attachments: Wolverine Detailed Timeline_post RTM.docx

Caitlin,

Can you take a look at this? Marj talked to Q about wolverine FR date. I revised the timeline given our recommendation and aiming for end of Aug to FR. Highlighted steps I'm not sure how long should be allocated given current direction from HQ. I'm allowing 3 months from HQ submission of the package to FR, not sure if that is appropriate these days or not.

Justin Shoemaker
Classification and Recovery Biologist
U.S. Fish and Wildlife Service, Region 6
Phone: 309-757-5800 x214
Email: justin_shoemaker@fws.gov

Wolverine Listing Determination Timeline
2/7/18 version

Task	Responsible Parties	Dates	Length of time
Listing Decision Analysis Phase			
SSA report to recommendation team	Justin Shoemaker, Jodi Bush	Jan 10, 2018	At least 2 weeks prior to recommendation team meeting
Decision meeting	RDs or delegates, ARDs, other management, SSA core team	Jan 25, 2018	1 days
Draft decision summary for the record or certify decision meeting notes	R6 RD or delegate	Mid Feb 2018	
Briefing for Acting Director	Greg Sheehan, Gary Frazer, Bridget Fahey, Noreen Walsh, Mike Thabault, Marj Nelson, Justin Shoemaker, Jodi Bush, Betty Grizzle, Paul Souza, Sarah Quamme	Feb 20	30 min
Process for final withdrawal of proposed listing			
Draft final withdrawal (not-warranted) FR notice	Justin Shoemaker and Betty Grizzle	Feb 20 – April 6, 2018	
Core team reviews FR notice, make revisions, Prelim SOL review	SSA core team, Justin Shoemaker	April 6 – April 20, 2018	2 weeks
SOL surname and revisions	DOI SOL	April 20 – May 4, 2018	2 weeks
Regional Office Surnames and concurrence and revisions	Marjorie Nelson, Mike Thabault, Matt Hogan, Noreen Walsh, and concurring regional RDs/ARDs or delegates	May 4 – May 18, 2018	2 weeks
PPM review and revisions	PPM	May 18 – June 1, 2018	2 weeks
HQ review	Sarah Quamme, Bridget Fahey	June 1, 2018	submit 3 months prior to FR submittal date
Asst. Director for ES Surname	Asst. Director for ES	April 4, 218	5 business days
FWS Director Surname	Director of FWS	April 11, 2018	5 business days
Fish, Wildlife, and Parks Surname	FWP	April 18, 2018	10 business days
Executive Secretary Surname	Executive Secretary's Office	May 2, 2018	3 business days

*Includes States, Tribes, Federal Agencies

Deliver to FR	HQ	Aug 31, 2018	3 months after submittal of package to HQ
Publication of withdrawal	Federal Register	Sept 7, 2018	

*Includes States, Tribes, Federal Agencies

From:	trina_vigil@fws.gov
To:	Marjorie Nelson; Michael Thabault; Justin Shoemaker; Jodi Bush; Betty Grizzle; John Guinotte; Stephen Torbit
Subject:	Wolverine Dry Run
Start:	Wednesday, February 14, 2018 10:00:00 AM
End:	Wednesday, February 14, 2018 11:00:00 AM
Location:	RD Conference Room

From: Trina Vigil
To: Justin Shoemaker
Subject: RE: Wolverine Dry Run
Date: Thursday, February 8, 2018 9:52:56 AM

Okay, I have it written down for her to add. She will get the information you to you guys. Sorry I don't have the info to send to you.

Trina Vigil
Administrative Support Assistant
134 Union Blvd
Lakewood, CO 80228
(303) 236-4253
Telework Friday's

From: Shoemaker, Justin [mailto:justin_shoemaker@fws.gov]
Sent: Thursday, February 08, 2018 9:51 AM
To: Trina Vigil
Subject: Re: Wolverine Dry Run

I think Stephanie likes to use their vidyo link, so its fine for now. Thanks.

Justin Shoemaker
Classification and Recovery Biologist
U.S. Fish and Wildlife Service, Region 6
Phone: 309-757-5800 x214
Email: justin_shoemaker@fws.gov

On Thu, Feb 8, 2018 at 10:28 AM, Trina Vigil <trina_vigil@fws.gov> wrote:

Justin-

I was in the process of adding the conference line info and it has now been added. I have left a note with Stephanie letting her know that we will need to add webex or vidyo for this meeting. Or is this something we provide? Sorry, I am just covering for the RD assistant today not sure of the process.

Trina Vigil
Administrative Support Assistant
134 Union Blvd
Lakewood, CO 80228
(303) 236-4253
Telework Friday's

From: Shoemaker, Justin [mailto:justin_shoemaker@fws.gov]
Sent: Thursday, February 08, 2018 9:23 AM
To: Vigil, Trina
Subject: Re: Wolverine Dry Run

Thanks Trina. We will need webex or vidyo for this meeting. And a conference line.

Justin Shoemaker
Classification and Recovery Biologist
U.S. Fish and Wildlife Service, Region 6

Phone: 309-757-5800 x214
Email: justin_shoemaker@fws.gov

2018-02-08 10:14 GMT-06:00 <trina_vigil@fws.gov>:

From:	Shoemaker, Justin
To:	Grizzle, Betty
Subject:	Fwd: Briefing Requirements for ESA Decisions
Date:	Thursday, February 8, 2018 10:03:05 AM
Attachments:	066863 Briefing Up on ESA Packages.pdf

FYI

Justin Shoemaker
Classification and Recovery Biologist
U.S. Fish and Wildlife Service, Region 6
Phone: 309-757-5800 x214
Email: justin_shoemaker@fws.gov

---------- Forwarded message ----------
From: **Quamme, Sarah** <sarah_quamme@fws.gov>
Date: Wed, Dec 6, 2017 at 2:59 PM
Subject: Fwd: Briefing Requirements for ESA Decisions
To: FWS ULT <fws_ult@fws.gov>

Hello Team - Please see the attached memo and guidelines regarding formal briefings for our decisions.

---------- Forwarded message ----------
From: **Fahey, Bridget** <bridget_fahey@fws.gov>
Date: Wed, Dec 6, 2017 at 3:17 PM
Subject: Briefing Requirements for ESA Decisions
To: FWS ES Regional ARDs <fws_es_regional_ards@fws.gov>, FWS ES Deputy ARDs <fws_es_deputy_ards@fws.gov>, Sarah Quamme <Sarah_Quamme@fws.gov>, Susan Jacobsen <Susan_Jacobsen@fws.gov>, Alisa Shull <alisa_shull@fws.gov>, Drew Crane <drew_crane@fws.gov>, Gina Shultz <Gina_Shultz@fws.gov>, Jeff Newman <jeff_newman@fws.gov>, Carey Galst <Carey_Galst@fws.gov>, Don Morgan <Don_Morgan@fws.gov>, Janine Van Norman <janine_vannorman@fws.gov>, "marilet_zablan@fws.gov" <marilet_zablan@fws.gov>, Marjorie Nelson <marjorie_nelson@fws.gov>, Martin Miller <martin_miller@fws.gov>, "michael_long@fws.gov" <michael_long@fws.gov>, "Merritt, Timothy" <timothy_merritt@fws.gov>, Aaron Valenta <Aaron_Valenta@fws.gov>

Hey y'all. This memo from Gary went out to DTS on Friday, but since not all of us are regular DTS users, I wanted to make sure that everyone has seen it. The key point is that Regions will need to set up briefings with the Director's office on ESA decisions in advance (shortly after the Regional Director has developed a recommended path forward). Let us know if you have any questions.

Bridget Fahey
Division Chief for Conservation and Classification
U.S. Fish and Wildlife Service
(703) 358-2163

--

Sarah Joan Quamme
Unified Listing Team Manager
U.S. Fish & Wildlife Service
703-358-1796

United States Department of the Interior

FISH AND WILDLIFE SERVICE

In Reply Refer To:
FWS/AES/DCC/ULT/066863

NOV 30 2017

Memorandum

To: Regional Directors, Regions 1-8
(Attention: ARD-ES)

From: Assistant Director for Ecological Services

Subject: Briefing Up on ESA Decision Packages

As we discussed during the Fall 2017 ARD meeting, attached is guidance on how to successfully brief Service leadership on ESA decision packages going forward. This guidance will provide direction on which actions are required to be briefed, the timing and duration of the briefing, the content that should be included, how to set the briefings up, and who should attend. Finally, the attached will provide some tips based on previous successes and lessons learned.

Attachment

Briefing Up on Endangered Species Act Decision Packages

Each Endangered Species Act decision should be briefed up to the Director and Principal Deputy Director early enough in the process so that they can provide feedback before the package is on surname.

Types of actions that should be briefed up: All proposed and final rulemakings to list, reclassify, or delist species under the Endangered Species Act require a briefing. Likewise, all proposed and final critical habitat rulemakings require a briefing. Warranted but precluded 12-month findings should also be briefed. Not warranted 12-month findings and 90-day findings will usually *not* need to be briefed, with some exceptions for determinations with high visibility and/or controversy. For example, actions that will be of great interest or concern to the States or industry should be considered for briefing up, at the discretion of the Regional Director.

Content: The purpose of the briefing is to allow the Director's office to become familiar with the recommended direction on a package well in advance of that package arriving in Headquarters on surname, as well as to allow for an earlier opportunity to provide feedback on the recommendation. To that end, the briefing should include the overall conservation picture, the approach to critical habitat (if applicable), the contents of a 4(d) rule (if applicable), interested stakeholders, and significant or controversial issues. If applicable, the briefing should reference any relevant Candidate Conservation Agreements (CCAs), Candidate Conservation Agreements with Assurances (CCAAs), Safe Harbor Agreements (SHAs), Habitat Conservation Plans (HCPs), or other conservation commitments. The briefing should also include information on past conservation efforts and expected concerns of state fish and wildlife agencies. Finally, the briefing should include a discussion about the quality of the scientific information informing the decision.

Timing of Briefing: The Director and Principal Deputy should be briefed after the Regional Director (or Regional Directors, for decisions that cross multiple regions)) has provided direction to his/her staff on a recommended course of action. The briefing should occur within 30 days of the recommendation team meeting, and before the appropriate Federal Register document has been put on surname in the Region(s).

Setting up Meetings: The Regions will have the lead in working with the Director's scheduler to find a time that works for the RD, Director, Principal Deputy Director, and Assistant Director for Ecological Services. For listing packages, the Chief of the Division of Conservation and Classification, the Species Assessment Team (SAT) Manager, and the SAT Project Manager should be included; for delisting or downlisting packages, the Chief of the Division of Recovery and Restoration, the Chief of the Branch of Recovery and State Grants, and the HQ lead biologist should be included (although we will be flexible on participation when needed to adhere to tight deadlines). Regional Directors can include subject matter experts; however, we recommend that they keep this to a small number of staff. For court-ordered and/or significant determinations, there may be a narrow window between the recommendation team meeting and the need for RO surnames, so Regional Directors should consider scheduling these briefings in advance.

Length of Briefing: The typical briefing should be 30 minutes.

Briefing Recommendations: We recommend that the lead region prepare a short (<10 slide) powerpoint which is shared with the SAT Manager or HQ recovery lead ahead of time. Presentation materials should include maps and other visual aids needed to fully relay the conservation picture. Webinar technology may not always be available in MIB meeting rooms, so Regions should be prepared to brief from paper copies of the presentation, which should be provided to the Director's administrative staff to print out if need be. The Regional Director for the lead region should lead the briefing, turning over discussion to subject matter

experts as appropriate, while also trying to limit the number of speakers to avoid confusion. The lead region should target about 15 minutes to present with 15 minutes for questions and answers.

From: Grizzle, Betty
To: Scott Sobiech
Subject: Upcoming briefing for North American Wolverine
Date: Thursday, February 8, 2018 10:27:18 AM
Attachments: 066863 Briefing Up on ESA Packages.pdf

Scott - I have been asked to prepare and likely will deliver a webinar briefing in the form of a PowerPoint presentation for the Principle Deputy Director, Greg Sheehan, and the Assistant Director, Ecological Services, Gary Frazer, which will also include the participation of managers/staff in HQ and Region 6. This is scheduled for February 20. Please see the attached memorandum that describes the purpose and timing of these ESA decision briefings.

Region 6 has asked for a run through of this presentation on **February 14 from 9:00 to 10:00 am**.

Please let me know if you would like to participate in, or provide feedback during, the Feb 14 call/meeting.

--

Betty J. Grizzle, D.Env., MPH
Fish and Wildlife Biologist
U.S. Fish and Wildlife Service
Carlsbad Fish and Wildlife Office
2177 Salk Ave, Suite 250
Carlsbad, CA 92008
760-431-9440, ext. 215
760-431-5901 fax

United States Department of the Interior

FISH AND WILDLIFE SERVICE

In Reply Refer To:
FWS/AES/DCC/ULT/066863

NOV 30 2017

Memorandum

To: Regional Directors, Regions 1-8
(Attention: ARD-ES)

From: Assistant Director for Ecological Services

Subject: Briefing Up on ESA Decision Packages

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Length of Briefing: The typical briefing should be 30 minutes.

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experts as appropriate, while also trying to limit the number of speakers to avoid confusion. The lead region should target about 15 minutes to present with 15 minutes for questions and answers.

From: Shoemaker, Justin
To: Grizzle, Betty
Subject: powerpoint
Date: Thursday, February 8, 2018 2:20:58 PM
Attachments: DPS Presentation_wolverine RTM_Jan 2018.pptx

Justin Shoemaker
Classification and Recovery Biologist
U.S. Fish and Wildlife Service, Region 6
Phone: 309-757-5800 x214
Email: justin_shoemaker@fws.gov

Wolverine
Regulatory and Litigation History



Timeline of Federal Actions

1995, 2003: 90-day Not Substantial Findings

- 2003 finding challenged; 12-mo finding ordered by Court
- **2008: Not Warranted Finding (no DPS in contiguous U.S.)**
- Challenged; FWS agreed to issue new finding

2010: 12-month Finding

- **DPS warranted**, but precluded by higher priority listing actions
- MDL agreement – publish either proposed rule or withdraw warranted finding; agreed to publish in FY 2013

2013: Proposed Rule, with DPS (using 2010 analysis), 4(d)

- Based primarily on synergistic effects of climate change, persistent snow
- **2014: Withdrawal of proposed rule**
- Remanded back to FWS

Current Action: Finalize Propose Rule or Withdrawal

- Informed by SSA Report
- Withdrawal or Final Rule to publish FY 2018



2013 p-Listing

- Assumed dependence on deep persistent spring snow, cold temperatures yr-round
- This feature is likely to be reduced by 33% by 2045, 63% by 2085
- Other threats are speculative or minor and absent CC would not be considered threats
  trapping, winter recreation, development, inbreeding



Determination: Threatened -- by 2085 we expect wolverine populations to be smaller and more fragmented as a result of CC to the point of being endangered

2014 Withdrawal

- Peer & partner review, expert meeting highlighted uncertainties.
- Information supporting the proposed listing rule was too uncertain and speculative to be relied upon in the listing decision.
- Habitat changes due to climate change do not constitute a threat to the DPS at this time.



- Other risk factors such as effects of inbreeding due to small and semi-isolated populations, and harvest are not threats to the DPS.

2014 Withdrawal Litigation

- Court Opinion highlighted the Service's determinations regarding:
  - (a) the threat posed to the wolverine by the effects of climate change at the reproductive denning scale,
  - (b) the threat posed to the wolverine by small population size and lack of genetic diversity, and
  - (c) application of the SPR Policy to the wolverine.



History of DPS



- The ESA allows listing of populations as endangered species or threatened species
  - "species" includes any subspecies of fish or wildlife or plants, and **any distinct population segment** of any species of vertebrate fish or wildlife which interbreeds when mature
- 1978 ESA Amendments - DPS restricted to vertebrates
- DPSs are to be used "sparingly and only when the biological evidence indicates that such action is warranted" (Senate Report 151, 96th Congress, 1st Session)

DPS – Defined by Policy*

- Policy identifies principles that guide the Services' listing, delisting, and reclassification of DPSs of vertebrate species

- Three Elements:

  - **Discreteness -** Discreteness of the population segment in relation to the remainder of the species to which it belongs;
  - **Significance -** The significance of the population segment to the species to which it belongs
  - **Conservation Status -** The population segment's conservation status in relation to the ESA standards for listing



*Joint FWS/NMFS policy published February 7, 1996 (61 FR 4722)

Discreteness

- It is markedly separate from other populations of same taxon as a consequence of physical, physiological, ecological, or behavioral factors. Quantitative measures of genetic or morphological discontinuity may provide evidence of this separation.

- It is delimited by international governmental boundaries within which differences in control of exploitation, management of habitat, conservation status, or regulatory mechanisms exist that are significant in light of section 4(a)(1)(D) of the Act.



Red tree vole



Marbled murrelet



Significance



- …its biological and ecological significance will then be considered in light of Congressional guidance (see Senate Report 151, 96th Congress, 1st Session) that the authority to list DPS's be used "***sparingly" while encouraging the conservation of genetic diversity.

- In carrying out this examination, the Services will consider available scientific evidence of the discrete population segment's importance to the taxon to which it belongs.

- This consideration may include, but is not limited to, the following:
  - Persistence of discrete population segment in an ecological setting **unusual or unique** for the taxon,
  - Evidence that loss of the discrete population segment would result in a **significant gap** in the range of a taxon,
  - Evidence that the discrete population segment represents the **only surviving natural occurrence** of a taxon that may be more abundant elsewhere as an introduced population outside its historical range, or
  - Evidence that the discrete population segment **differs markedly** from other populations of the species in its **genetic characteristics**.

Conservation Status

- Does it meet the Act's definitions of "endangered species" or "threatened species”?
- Example: **Black-footed albatross**
  - Hawaiian Islands DPS and Japanese Islands DPS
  - Determined discrete and significant
    - Discrete based on behavior of strong nest fidelity, physical separation, genetics
    - Significant because loss of 1 population would result in significant gap in range, and markedly different genetics
  - After evaluation, neither of these DPSs warranted listing under the ESA



Wolverine DPS History

2008: Lower 48 NOT DPS

2009: Challenged on 2008 finding

2010: Lower 48 IS a DPS

2013: Lower 48 IS a DPS

2014: Lower 48 IS a DPS

2018: For Discussion



2008 (12-mo finding)

Discreteness = NO

1. Markedly Separated = NO

- U.S. population connected to populations in Canada and likely dependent on them to some degree to maintain genetic diversity.

2. International Governmental Boundary per 4(a)(1)D = NO

- No difference in biological conservation status
- Difference in exploitation (e.g. trapping) not significant due to adequate regulation.

Overall

Not discrete, so not a DPS. We did not explore significance.



2008 (12-mo finding)

2009 Litigation

- We were quickly challenged by Defenders of Wildlife, Center for Biological Diversity, and others.
- In 2009, we entered into a settlement agreement to reconsider our 12-mo finding.
- Record reflects a last minute change in 2008 DPS determination.



2010 (warranted but precluded); 2013 (proposed listing rule), 2014 (vacated withdrawal)


Vincent Van Zalinge

All used the following rationale under the DPS policy.

Discreteness = YES

1. Markedly Separated = NO
We provided no analysis

2. International Governmental Boundary per 4(a)(1)D = YES
Population Size:
Canada = large populations;
Contiguous US = population numbers low continued existence at risk
Effective Population Size:
Canada = in good shape with sufficient genetic diversity;
Contiguous US = small with poor genetic diversity
Habitat Connectivity:
Canada = well-distributed and interconnected;
Contiguous US =fragmented habitat.
Exploitation
Did not rely on control of exploitation to establish discreteness

2010 (warranted but precluded); 2013 (proposed listing rule), 2014 (vacated withdrawal)


Vincent Van Zalinge

Significance = YES

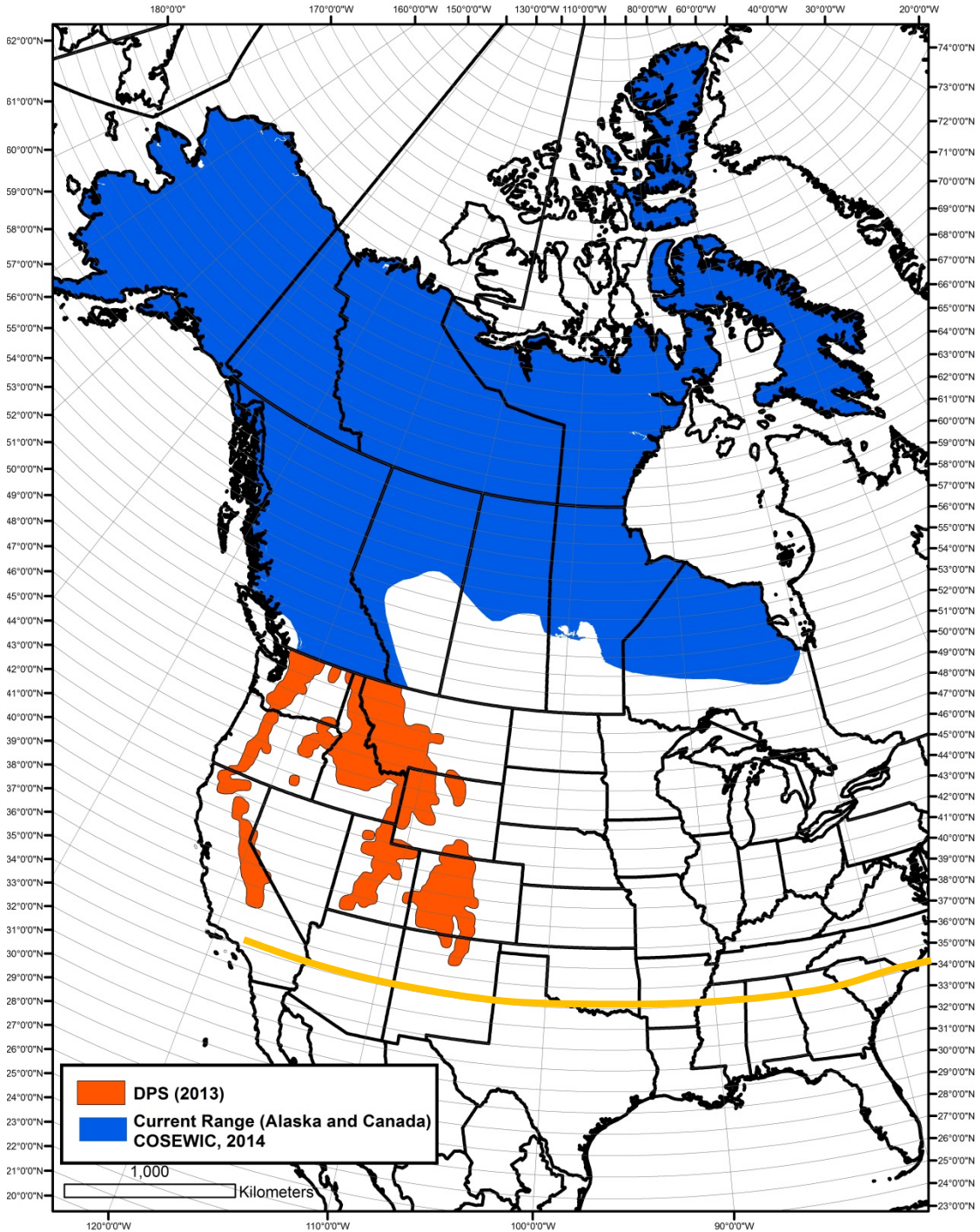
Criteria #2 (Evidence that loss of the discrete population segment would result in a significant gap in the range of the taxon) = YES

Loss would substantially reduce southern range terminus approximately 15 degrees of latitude (40 percent of the latitudinal extent of the wolverine range); and loss of 205,942 km2 (79,515 mi2) of habitat (based on Copeland et al 2010 model)

Criteria #1 (persistence of the discrete population segment in an ecological setting unusual or unique for the taxon) = NO

Criteria #3 (only remaining occurrence of a taxon) = NO

Criteria #4 (…discrete population segment differs markedly from other populations of the species in its genetic characteristics) = NO



12-Month Finding on a Petition To List the North American Wolverine as Endangered or Threatened (2010)

--

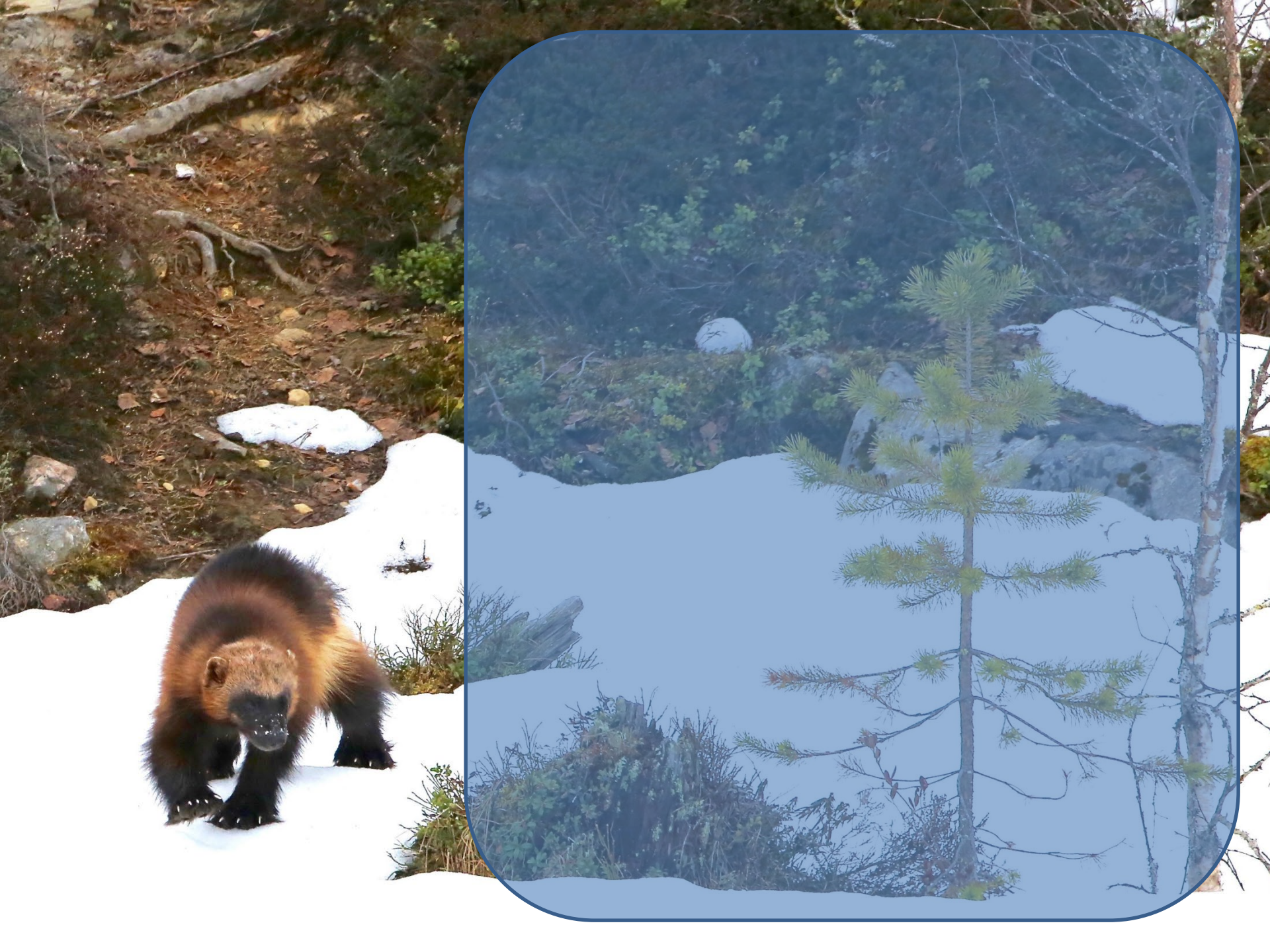
Today, the contiguous United States represents the southernmost reach of the wolverine's range. ***The loss of this population would be significant because it would substantially curtail the range of the wolverine by moving the southern range terminus approximately 15° of latitude to the north (or approximately 40 percent of the latitudinal extent of wolverine range)*** *and eliminate wolverines from the fauna of the contiguous United States. Therefore, the loss of this population would result in a significant gap in the range of the taxon.*

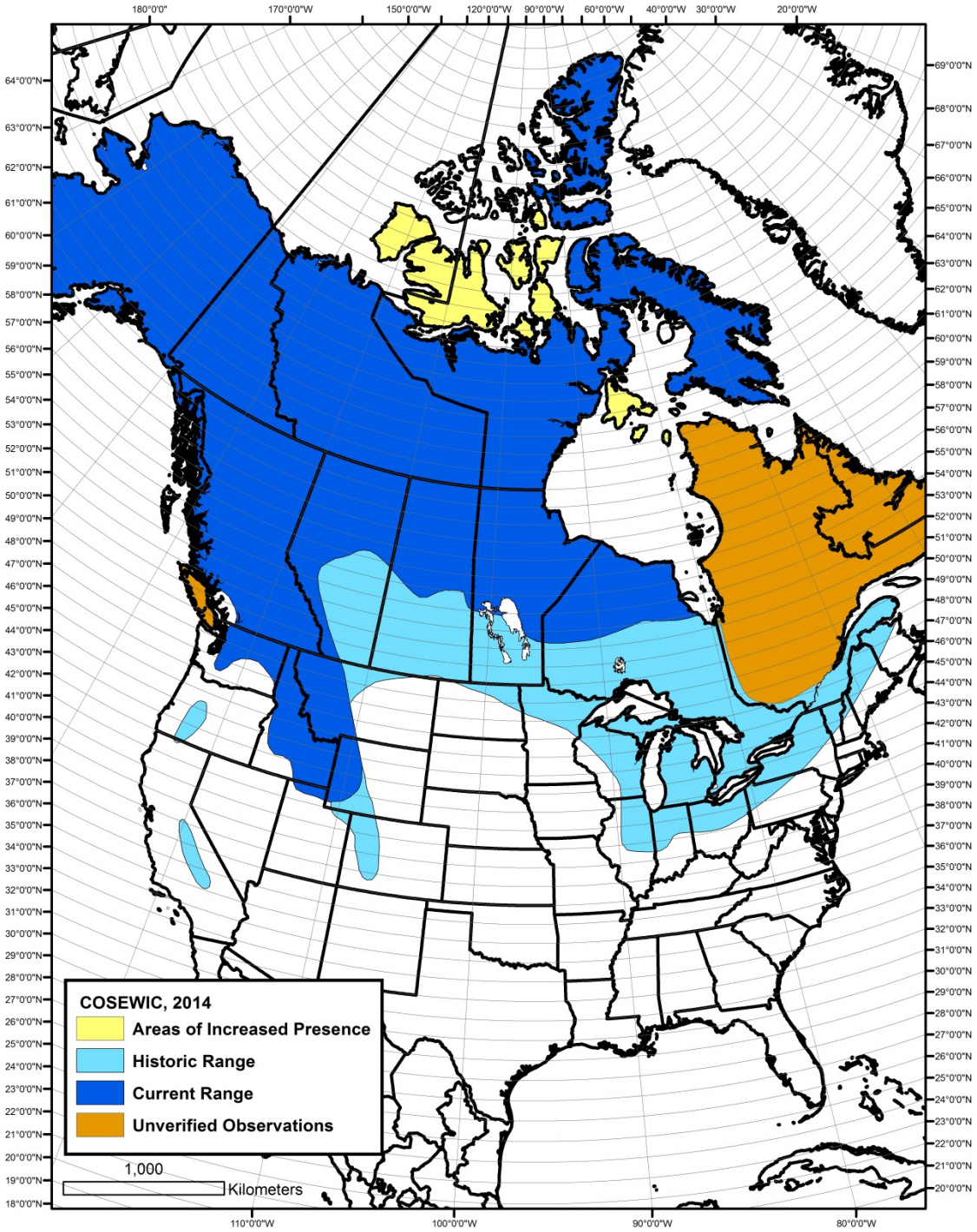
Based upon the 15 degree latitude gap that would result in the range of the wolverine if the U.S. population was lost, we determine that the loss of the contiguous U.S. wolverine population would result in a significant gap in the range of the taxon. *Thus, the DPS meets the definition of significant in our DPS policy.*

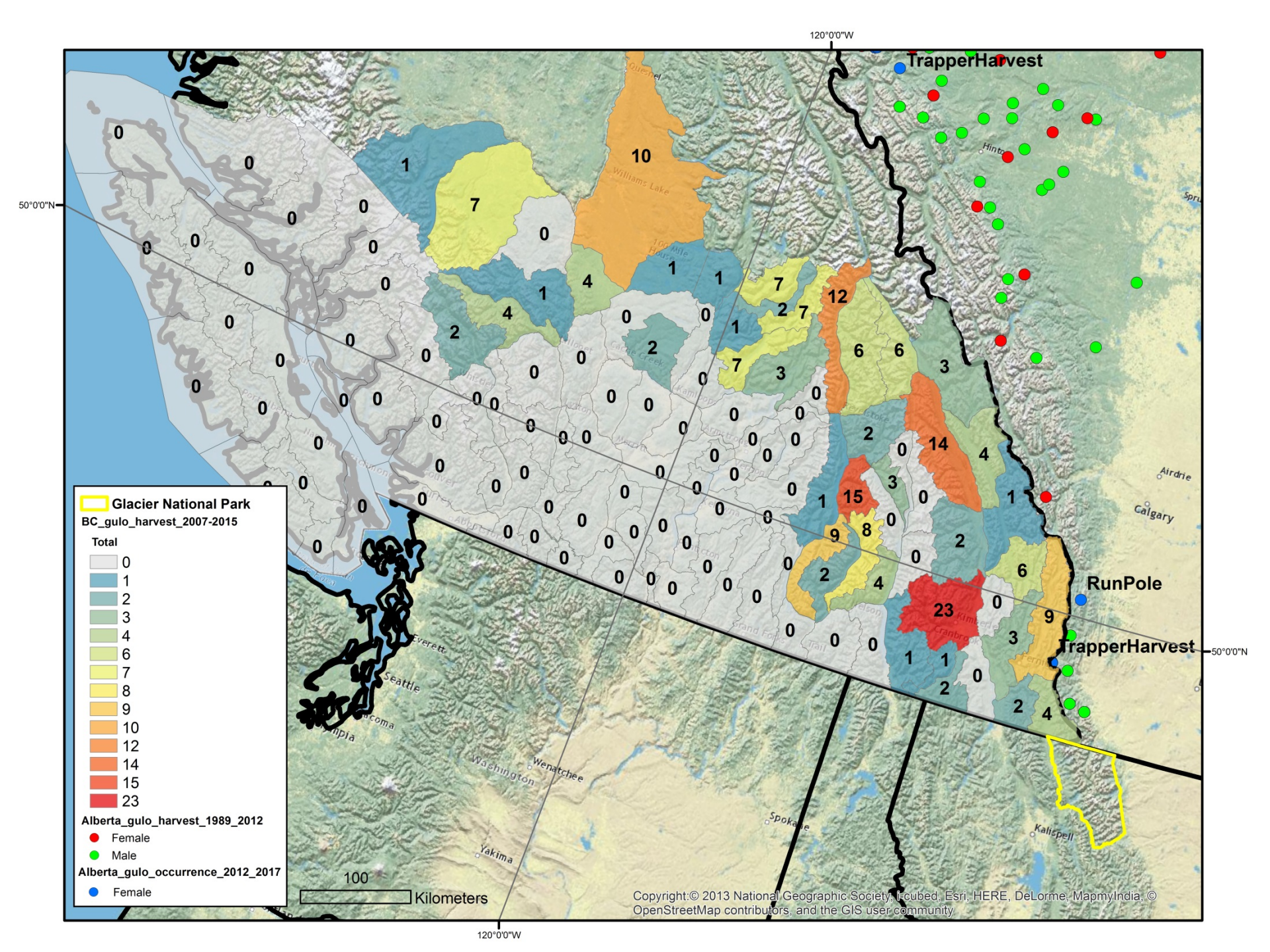

Roy Andersen



Roy Andersen




COSEWIC, 2014
Areas of Increased Presence
Historic Range
Current Range
Unverified Observations
1,000
Kilometers


120°0'0"W
50°0'0"N
TrapperHarvest
RunPole
TrapperHarvest
Glacier National Park
BC_gulo_harvest_2007-2015
Total
0
1
2
3
4
6
7
8
9
10
12
14
15
23
Alberta_gulo_harvest_1989_2012
Female
Male
Alberta_gulo_occurrence_2012_2017
Female
100
Kilometers
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From: Stephen Torbit
To: betty_grizzle@fws.gov; john_guinotte@fws.gov
Subject: Fwd: Individual Identification of Wolverine from Multistate Wolverine Effort
Date: Friday, February 9, 2018 9:32:40 AM
Attachments: image001.png
image003.png
image004.png

Keep this close for now. Had to beat on people yesterday.

Hopefully haplotypes coming soon.

Heading back to Denver soon.

Sent from my iPhone

Begin forwarded message:

> **From:** "Schwartz, Michael K -FS" <michaelkschwartz@fs.fed.us>
> **Date:** February 9, 2018 at 8:47:39 AM MST
> **To:** "Evans Mack,Diane" <diane.evansmack@idfg.idaho.gov>, "Pilgrim, Kristine L -FS" <kpilgrim@fs.fed.us>
> **Cc:** "jgude@mt.gov" <jgude@mt.gov>, "Sallabanks,Rex" <rex.sallabanks@idfg.idaho.gov>, "bob.lanka@wyo.gov" <bob.lanka@wyo.gov>, "Jackson, Scott -FS" <sjackson03@fs.fed.us>, "Stephen Torbit (stephen_torbit@fws.gov)" <stephen_torbit@fws.gov>, "Zack Walker (zack.walker@wyo.gov)" <zack.walker@wyo.gov>, "Lewis, Jeffrey C (DFW)" <Jeffrey.Lewis@dfw.wa.gov>, "Swanson, Jessica - FS" <jessicaswanson@fs.fed.us>, "Bob Inman (bobinman@mt.gov)" <bobinman@mt.gov>
> **Subject: RE: Individual Identification of Wolverine from Multistate Wolverine Effort**
>
> Hi Diane,
> Thanks for asking if these are all new individuals. The short answer is that they are not. However, I want to be careful about sharing data that is not ours to share. If it is OK with everyone on this e-mail I will approach other parties that have detected individuals that you each detected and ask if they are OK sharing. I would like to offer to share that their animal was recaptured by the multi-state effort. Please advise me on how to proceed.
> Thanks
> Mike
>
> ---
>
> **From:** Evans Mack,Diane [mailto:diane.evansmack@idfg.idaho.gov]
> **Sent:** Friday, February 09, 2018 8:17 AM
> **To:** Schwartz, Michael K -FS <michaelkschwartz@fs.fed.us>; Pilgrim, Kristine L -FS <kpilgrim@fs.fed.us>
> **Cc:** jgude@mt.gov; Sallabanks,Rex <rex.sallabanks@idfg.idaho.gov>;

bob.lanka@wyo.gov; Jackson, Scott -FS <sjackson03@fs.fed.us>; Stephen Torbit (stephen_torbit@fws.gov) <stephen_torbit@fws.gov>; Zack Walker (zack.walker@wyo.gov) <zack.walker@wyo.gov>; Lewis, Jeffrey C (DFW) <Jeffrey.Lewis@dfw.wa.gov>; Swanson, Jessica - FS <jessicaswanson@fs.fed.us>; Bob Inman (bobinman@mt.gov) <bobinman@mt.gov>
Subject: RE: Individual Identification of Wolverine from Multistate Wolverine Effort

Hi Mike and Kristy –
Thanks so much for this. Can't wait to map it all out. Questions – are all of these individuals new to the existing individual wolverine database? None was a familiar animal from previous work?

Thanks.
Diane

Diane Evans Mack
Regional Wildlife Diversity Biologist
Idaho Department of Fish and Game
McCall Subregion
555 Deinhard Lane
McCall, ID 83638
208-634-8137
diane.evansmack@idfg.idaho.gov

From: Schwartz, Michael K -FS [mailto:michaelkschwartz@fs.fed.us]
Sent: Thursday, February 08, 2018 7:58 PM
To: Evans Mack,Diane; Zack Walker; Jeff Lewis; Swanson, Jessica - FS; Inman, Bob
Cc: jgude@mt.gov; Sallabanks,Rex; bob.lanka@wyo.gov; Jackson, Scott -FS; Stephen Torbit (stephen_torbit@fws.gov); Pilgrim, Kristine L -FS
Subject: Individual Identification of Wolverine from Multistate Wolverine Effort

Dear Multistate Wolverine Team;
Attached are the individual identification results from the multi-state effort. We are working on the haplotype data and should be finished with that shortly.
Looking forward to the groups interpretation of the results.
Best Regards,
Mike

Michael K. Schwartz, Ph.D.
Program Manager Wildlife and Terrestrial Ecosystem Program &
Director of National Genomics Center for Wildlife and Fish Conservation

Forest Service
Rocky Mountain Research Station
p: 406-542-4161

f: 406-543-2663
mkschwartz@fs.fed.us
https://www.fs.fed.us/research/genomics-center/
https://www.fs.fed.us/rm/wildlife-terrestrial/

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IDAHO
FISH & GAME





From: Grizzle, Betty
To: Stephen Torbit
Cc: John Guinotte
Subject: Re: Individual Identification of Wolverine from Multistate Wolverine Effort
Date: Friday, February 9, 2018 10:55:31 AM
Attachments: image003.png
image001.png
image004.png

Steve - Since the original attachment was not included on your email message, I assume that the Working Group will be releasing this information to the Service. Is that correct?

On Fri, Feb 9, 2018 at 8:32 AM, Stephen Torbit <stephen_torbit@fws.gov> wrote:

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> Hopefully haplotypes coming soon.
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> Heading back to Denver soon.
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> Sent from my iPhone
>
> Begin forwarded message:
>
> **From:** "Schwartz, Michael K -FS" <michaelkschwartz@fs.fed.us>
> **Date:** February 9, 2018 at 8:47:39 AM MST
> **To:** "Evans Mack,Diane" <diane.evansmack@idfg.idaho.gov>, "Pilgrim, Kristine L -FS" <kpilgrim@fs.fed.us>
> **Cc:** "jgude@mt.gov" <jgude@mt.gov>, "Sallabanks,Rex" <rex.sallabanks@idfg.idaho.gov>, "bob.lanka@wyo.gov" <bob.lanka@wyo.gov>, "Jackson, Scott -FS" <sjackson03@fs.fed.us>, "Stephen Torbit (stephen_torbit@fws.gov)" <stephen_torbit@fws.gov>, "Zack Walker (zack.walker@wyo.gov)" <zack.walker@wyo.gov>, "Lewis, Jeffrey C (DFW)" <Jeffrey.Lewis@dfw.wa.gov>, "Swanson, Jessica - FS" <jessicaswanson@fs.fed.us>, "Bob Inman (bobinman@mt.gov)" <bobinman@mt.gov>
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>
> Thanks
>
> Mike

From: Evans Mack,Diane [mailto:diane.evansmack@idfg.idaho.gov]
Sent: Friday, February 09, 2018 8:17 AM
To: Schwartz, Michael K -FS <michaelkschwartz@fs.fed.us>; Pilgrim, Kristine L -FS <kpilgrim@fs.fed.us>
Cc: jgude@mt.gov; Sallabanks,Rex <rex.sallabanks@idfg.idaho.gov>; bob.lanka@wyo.gov; Jackson, Scott -FS <sjackson03@fs.fed.us>; Stephen Torbit (stephen_torbit@fws.gov) <stephen_torbit@fws.gov>; Zack Walker (zack.walker@wyo.gov) <zack.walker@wyo.gov>; Lewis, Jeffrey C (DFW) <Jeffrey.Lewis@dfw.wa.gov>; Swanson, Jessica - FS <jessicaswanson@fs.fed.us>; Bob Inman (bobinman@mt.gov) <bobinman@mt.gov>
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Hi Mike and Kristy –

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Thanks.

Diane

Diane Evans Mack
Regional Wildlife Diversity Biologist

Idaho Department of Fish and Game

McCall Subregion
555 Deinhard Lane
McCall, ID 83638
208-634-8137

diane.evansmack@idfg.idaho.gov

From: Schwartz, Michael K -FS [mailto:michaelkschwartz@fs.fed.us]
Sent: Thursday, February 08, 2018 7:58 PM
To: Evans Mack,Diane; Zack Walker; Jeff Lewis; Swanson, Jessica - FS; Inman, Bob
Cc: jgude@mt.gov; Sallabanks,Rex; bob.lanka@wyo.gov; Jackson, Scott -FS; Stephen Torbit (stephen_torbit@fws.gov); Pilgrim, Kristine L -FS
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Mike

Michael K. Schwartz, Ph.D.

Program Manager Wildlife and Terrestrial Ecosystem Program &

Director of National Genomics Center for Wildlife and Fish Conservation

Forest Service

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https://www.fs.fed.us/rm/wildlife-terrestrial/

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Missoula, MT 59802
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--

Betty J. Grizzle, D.Env., MPH
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U.S. Fish and Wildlife Service
Carlsbad Fish and Wildlife Office
2177 Salk Ave, Suite 250
Carlsbad, CA 92008
760-431-9440, ext. 215
760-431-5901 fax

From: Stephen Torbit
To: Grizzle, Betty
Cc: John Guinotte
Subject: Re: Individual Identification of Wolverine from Multistate Wolverine Effort
Date: Friday, February 9, 2018 11:12:02 AM
Attachments: image001.png
image004.png

It should have been attached. May have to send this afternoon from my computer, not my phone.

Sent from my iPhone

On Feb 9, 2018, at 10:55 AM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:

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Betty J. Grizzle, D.Env., MPH
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U.S. Fish and Wildlife Service
Carlsbad Fish and Wildlife Office
2177 Salk Ave, Suite 250
Carlsbad, CA 92008
760-431-9440, ext. 215
760-431-5901 fax

From: Grizzle, Betty
To: Shoemaker, Justin
Subject: Draft PowerPoint
Date: Friday, February 9, 2018 2:32:50 PM
Attachments: Briefing for Deputy Director_Wolverine 12mo finding.pptx

---

Justin - Here is my first draft for the Deputy Director's briefing. It is 10 slides, if I don't add the DPS topic, but 13 if I do include this.

Feel free to make any changes or if you have other comments/concerns.

--

Betty J. Grizzle, D.Env., MPH
Fish and Wildlife Biologist
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North American Wolverine (*Gulo gulo luscus*): 12-Month Finding



Photo credit: Mark Packila; used with permission.

Mountain Prairie Regional Office

February 20, 2018

Listing History

1995, 2003: 90-day Not Substantial Findings

- 2003 finding challenged; 12-mo finding ordered by Court
- **2008: Not Warranted Finding (no DPS in contiguous U.S.)**
- Challenged; FWS agreed to issue new finding

2010: 12-month Finding

- **DPS warranted**, but precluded by higher priority listing actions
- MDL agreement – publish either proposed rule or withdraw warranted finding; agreed to publish in FY 2013

2013: Proposed Rule, with DPS (using 2010 analysis), 4(d)

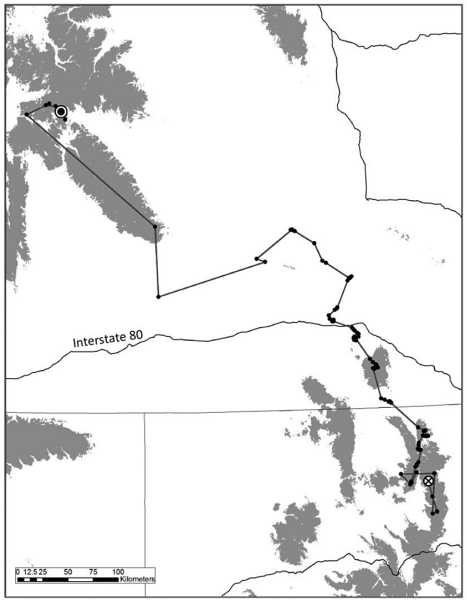
- Based primarily on synergistic effects of climate change, persistent snow
- **2014: Withdrawal of proposed rule**
- Challenged, Court remanded back to FWS

Current Action: Finalize Propose Rule or Withdrawal

- Informed by SSA Report
- Withdrawal or Final Rule to publish FY 2018

Species Description & Needs

- Key physical/ecological needs
  - Large territories, remote areas, high elevations
  - Access to variety of food, seasonal variability
  - Physical/Structural habitat features (e.g., talus slopes)
- Extensive movement and dispersal capabilities
  - Colorado, Utah, California, Michigan (all in 2000s)



Map of VHF locations of M56 - 2008 to 2009



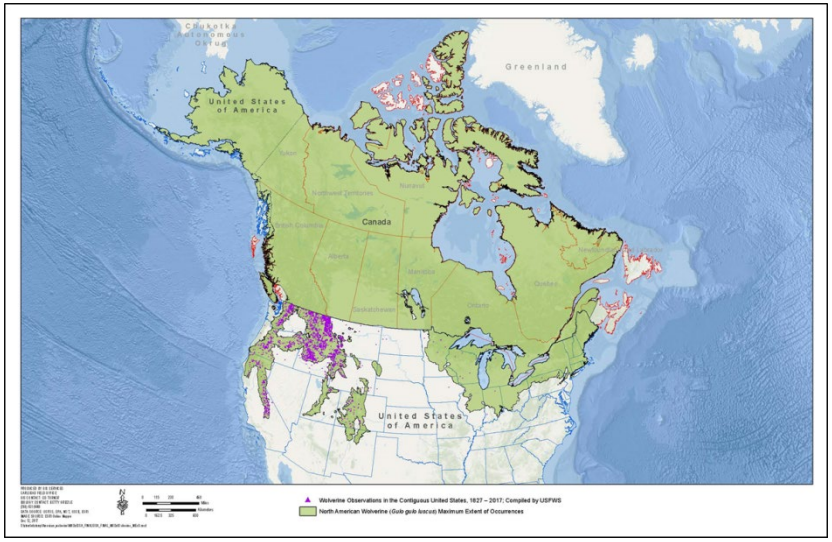
Suspected wolverine M56, from Packila et al. 2017

SSA Report –New information and analysis

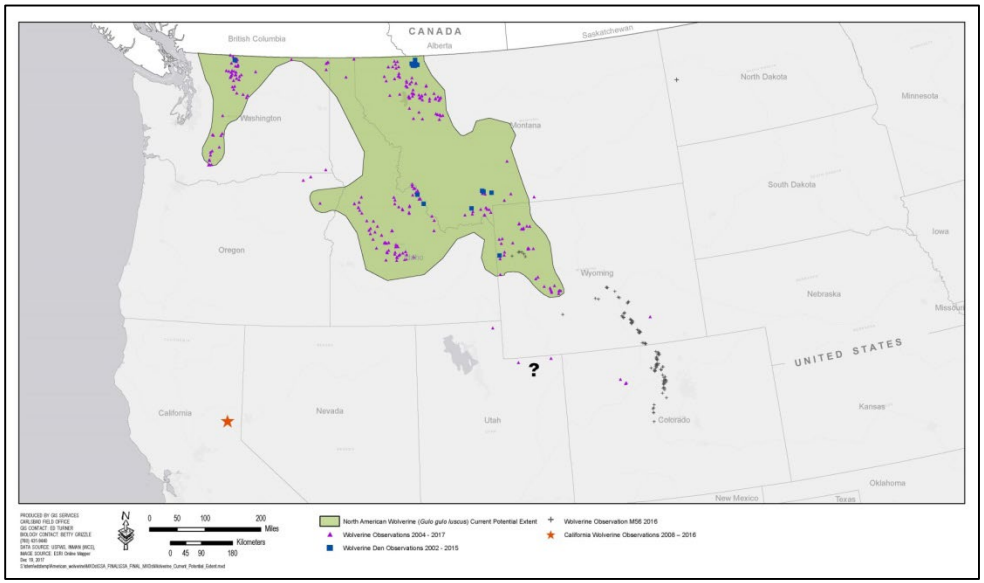
- Recent Scandinavian & Canadian studies providing new insights
- Physiological discussion added; not well-documented in previous rules
  - Important to describe critical temperature ranges and elements of thermoneutrality for evaluating effects of climate change
- New range maps prepared
- Contacted many wolverine experts (both in and outside U.S.), State agencies, and other scientists
- NOAA/CU snow study prepared (funded by Science Applications Program)

North American Wolverine Range Maps

- Maximum Extent of Occurrences (North America)



- Current Potential Extent of Occurrences (Lower 48)

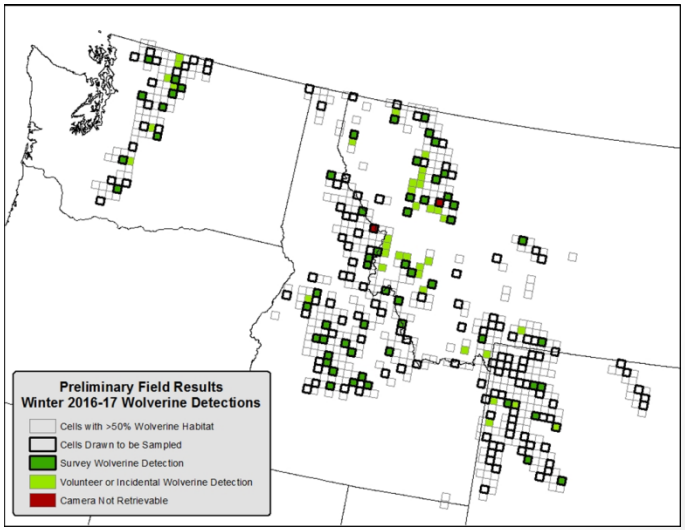


Denning and Snow: only one piece of the puzzle

- New studies – Copeland et al. model does not always predict either wolverine occurrences or den sites
  - At least 80 dens in Scandinavia and several in Alberta, Canada, recorded outside model
- Limited information on relationship of denning habitat with **reproductive success**
- Denning habitat varies across Holarctic range
- Young moved from natal den in late April
- Many factors influence den locations and shifts in den locations (not just temp/snow depth or persistence)
- Our analysis of "melt-out" dates for natal den sites in U.S. – earliest was May 14

Biological Status (current conditions)

- Population size
  - In No America: uncommon and at naturally low densities
  - Contiguous US: 2013 model estimate of 318 individuals (range 249-626)
    - Wolverine working group developing occupancy analysis
  - Canada – over 10,000 adults
    - Stable or increasing in this northern part of No Am range
  - AK uses density estimates (5-10 per 386 mi^2); considered stable
- Movement between AK and Canada & lower 48 and Canada
- Wolverine Occupancy Study
  - 4 Western States
  - 42% probability of occupancy



Potential Stressors Evaluated

- Effects from roads
- Effects from wildland fire
- Disturbance – winter recreational activity, other infrastructure
- Disease and Predation
- Overutilization
  - Research activities
  - Harvesting and incidental trapping
- For Future Conditions
  - Effects of Climate Change
  - Cumulative Effects

Overall Assessment

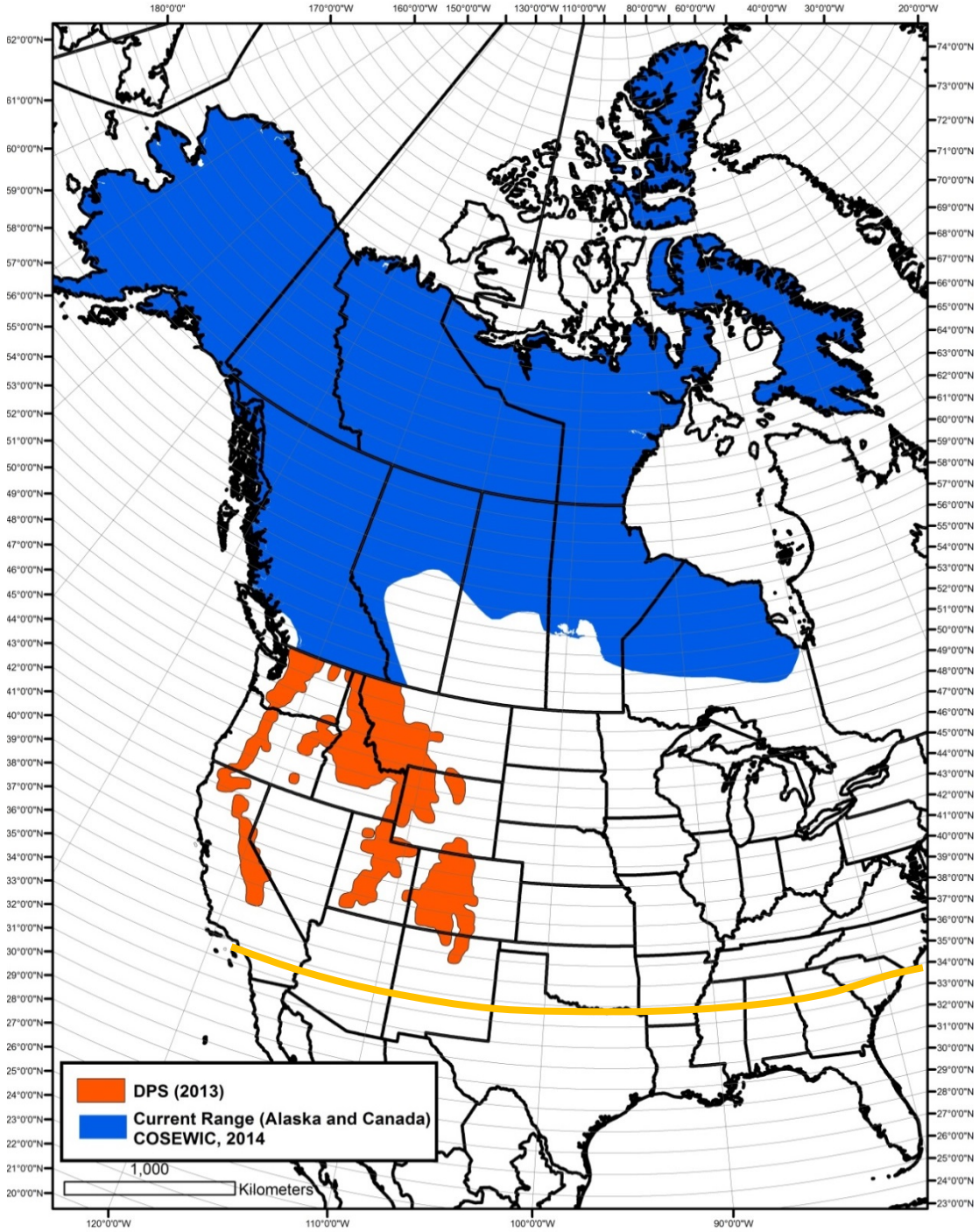
- Known or likely potential stressors represent low risk to the species
- Legal trapping currently prohibited in contiguous U.S.
  - Some incidental trapping still occurring
  - Trapping in Canada & Alaska appears to be sustainable
- Legal protections (T, E, Candidate) in several western states and Canada
  - Conservation plans (States and Tribes) in effect or planned
- No population trend information for contiguous U.S.
  - Wolverines now found in areas previously extirpated; still recovering from unregulated persecution
- Climate change effects: increase temperature and reduced snowpack
  - Model results: higher elevations (where wolverines are found) retain more snow than lower elevations
  - Modeled study areas: large portions have snow > 20” on May 1

Comments Received

- Response to Oct 2016 NOI (proposed rule/status review)
  - Over 200 comments posted (from several States, Canada, U.K.)
  - Most, but not all, general public comments favored listing
    - CC effects (need for deep snow) and rarity/small population size were primary concerns
- SSA Report Review
  - 11 States and Agency (Technical) Comments submitted
  - No Tribal comments submitted
  - Peer Reviewers
    - Complete, comprehensive, well-organized summary; additional references and technical comments provided
    - One reviewer was very negative – claimed “pervasive” bias in use and interpretation of info; CC projections should go beyond 50 years

DPS

- For 2010 (warranted but precluded); 2013 (proposed listing rule), 2014 (vacated withdrawal)
  - Discreteness = YES
    - 1. Markedly Separated = NO (No analysis provided)
    - 2. International Governmental Boundary per 4(a)(1)D = **YES**
      - Population Size: Canada = large populations; Contiguous US = population numbers low continued existence at risk
      - Effective Population Size: Canada = in good shape with sufficient genetic diversity; Contiguous US = small with poor genetic diversity
      - Habitat Connectivity: Canada = well-distributed/interconnected; Contiguous US = fragmented habitat
      - Exploitation: Did not rely on control of exploitation to establish discreteness
  - Significance = YES (see next slide)
    - Criteria #2 (Evidence that loss of the discrete population segment would result in a significant gap in the range of the taxon) = **YES**
      - *Loss would substantially reduce southern range terminus approximately 15 degrees of latitude (40 percent of the latitudinal extent of the wolverine range); and loss of 205,942 km^2 (79,515 mi^2) of habitat (**based on Copeland et al. 2010 model**)*
    - Criteria #1 (persistence of the discrete population segment in an ecological setting unusual or unique for the taxon) = NO
    - Criteria #3 (only remaining occurrence of a taxon) = NO
    - Criteria #4 (…discrete population segment differs markedly from other populations of the species in its genetic characteristics) = NO

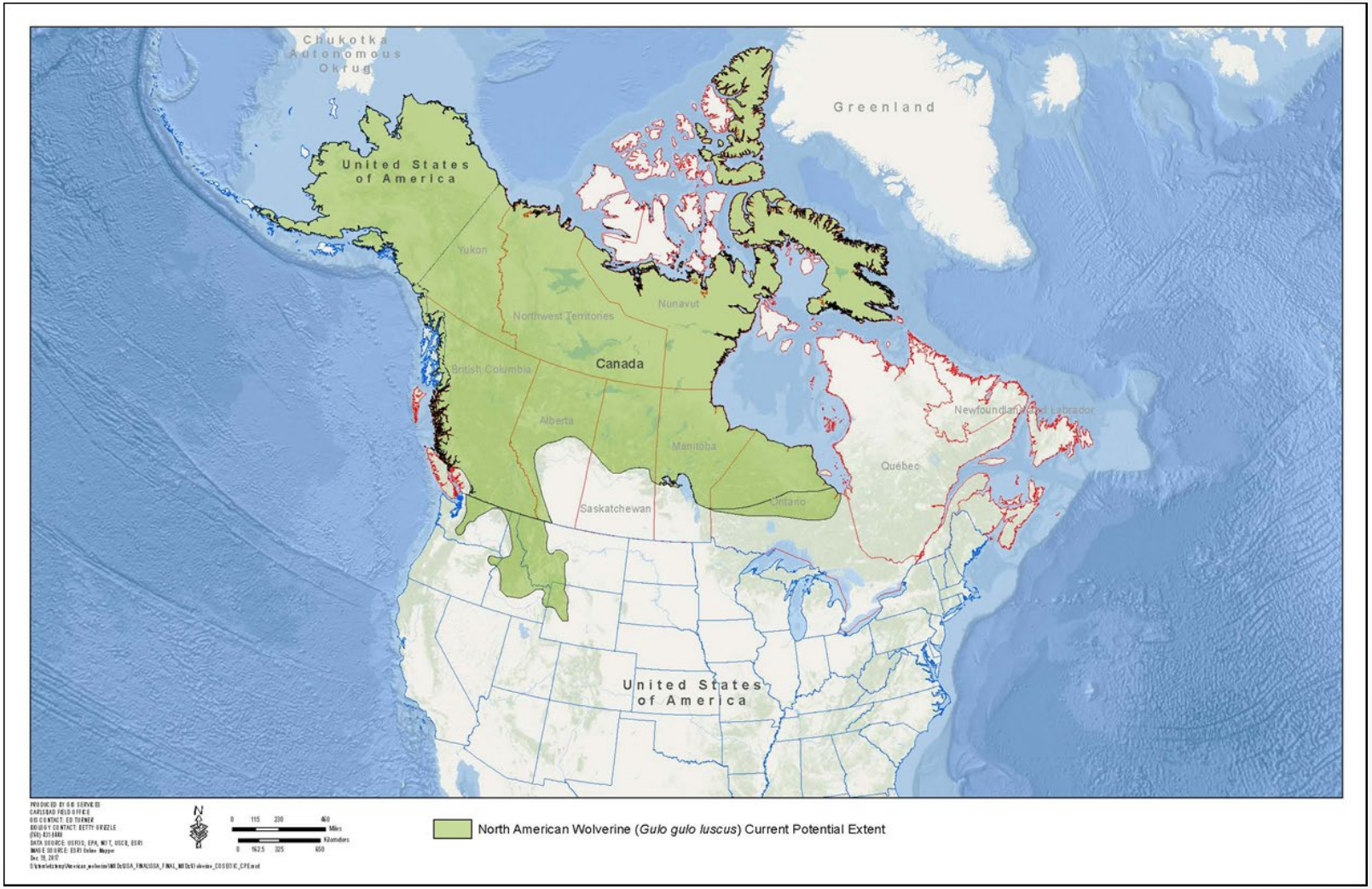


12-Month Finding on a Petition To List the North American Wolverine as Endangered or Threatened (2010)

" Today, the contiguous United States represents the southernmost reach of the wolverine's range. ***The loss of this population would be significant because it would substantially curtail the range of the wolverine by moving the southern range terminus approximately 15° of latitude to the north (or approximately 40 percent of the latitudinal extent of wolverine range)*** *and eliminate wolverines from the fauna of the contiguous United States.* ***Therefore, the loss of this population would result in a significant gap in the range of the taxon****."*

*"****Based upon the 15 degree latitude gap that would result in the range of the wolverine if the U.S. population was lost, we determine that the loss of the contiguous U.S. wolverine population would result in a significant gap in the range of the taxon.*** *Thus, the DPS meets the definition of significant in our DPS policy."*

Current Potential Extent – North America



From:	Stephanie Potter
To:	John Guinotte; Michael Thabault; Stephen Torbit; Marjorie Nelson; Jodi Bush; Justin Shoemaker; Betty Grizzle; Rollie White; Michael Fris
Subject:	Wolverine Dry Run
Start:	Wednesday, February 14, 2018 10:00:00 AM
End:	Wednesday, February 14, 2018 11:00:00 AM
Location:	VIdyo link below or RD conference room

---

https://fwsvideo.fws.gov/join/vRLe1DJE0c

In-person: RD Conference Room 4th Floor

From:	Stephen Torbit
To:	Betty Grizzle; John Guinotte
Subject:	FW: Individual Identification of Wolverine from Multistate Wolverine Effort
Date:	Monday, February 12, 2018 8:06:28 AM
Attachments:	image002.png image003.png Multistate Gulo_Individual ID 2_8_18_from NGC.pdf

Should work this way.

Stephen C. Torbit
Assistant Regional Director
Science Applications
U.S. Fish and Wildlife Service
134 Union Blvd.
Lakewood, Colorado 80228
303-236-4602 – Office
720-626-7504 – Cell

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REPORT

Project Title: DNA-based Individual Identification of Wolverine (*Gulo gulo*) Samples from the Multi-State Wolverine Survey

Date Issued: February 8, 2018

Recipients: **Diane Evans Mack**
Regional Wildlife Diversity Biologist
Idaho Department of Fish and Game
555 Deinhard Lane
McCall, ID 83638

Robert M. Inman, PhD
Carnivore-Furbearer Coordinator
Montana Fish, Wildlife and Parks
1420 East 6th Ave., PO Box 200701,
Helena, MT 59620-0701

Jeffrey C. Lewis, PhD
Mesocarnivore Conservation Biologist
Washington Department of Fish and Wildlife
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Jeffrey.Lewis@dfw.wa.gov

Jessica Swanson
Wildlife Biologist
Flathead NF
Tally Lake Ranger District
650 Wolfpack Way
Kalispell, MT 59901

Zack Walker
Certified Wildlife Biologist
Statewide Nongame Bird and Mammal Program Supervisor
Wyoming Dept. of Game and Fish
260 Buena Vista
Lander, WY 82520

Cc: **Stephen Torbit**
Scott Jackson

REPORT

Prepared By:

Kristine Pilgrim, M.S.
National Genomics Center for Wildlife and Fish Conservation
Missoula, MT 59801, USA
kpilgrim@fs.fed.us

Cory Engkjer
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clengkjer@fs.fed.us

Michael Schwartz, Ph.D.
Director
National Genomics Center for Wildlife and Fish Conservation
Missoula, MT 59801, USA
mkschwartz@fs.fed.us

REPORT

We analyzed 1,400 hair samples submitted under the auspices of the Multi-State Wolverine Survey (Table 1) for species identification. This included 1,342 hair samples submitted in 2017 and 58 hair samples collected in 2016 by Wyoming Game and Fish (WYGF). From these 1400 samples, 208 were identified as being from wolverine. These 208 samples were further analyzed using 15 microsatellite loci used in previous mustelid studies (Kyle and Strobeck 2001, 2002, Cegelski et al. 2003, 2006, Zielinksi et al. 2006, Schwartz et al. 2007): Gg3, Gg4, Gg7, Ma2, Ma8, Tt1, Tt4 (Davis and Strobeck 1998); Ggu101, Ggu216, Ggu234, Ggu238 (Duffy et al. 1998); Mvis020, Mvis72, Mvis075 (Flemming et al. 1999); and Lut604 (Dallas and Piertney 1998) to attempt to provide an individual genotype. One hundred and twenty four of the 208 samples had sufficient DNA to produce an individual genotype (59.6%; Table 2). In addition, five samples had more than two alleles at multiple loci indicating more than one individual wolverine ("mixed individual") and six samples had alleles both from wolverine and marten ("mixed with marten"). A total of 43 unique individual wolverine were identified from the 124 genotypes: 20 males and 23 females.

Table 1. Summary of samples received and wolverine results from the Multi-State Wolverine Survey 2016 & 2017

Survey Conducted By	Total Hair Samples Analyzed	Wolverine Samples	Individuals	Males	Females
IDFG	439	66	11	4	7
FNF	37	7	3	2	1
MTFWP	559	87	21	10	11
WDFW	147	29	4	2	2
WYGF	160	8	1	1	0
WYGF_2016	58	11	3	1	2
Total	**1400**	**208**	**43**	**20**	**23**

Table 2. Wolverine DNA individual and sex identification results for samples submitted from the Multi-State Wolverine Survey 2016 & 2017

Survey Conducted By	Sequential Sample Number	Cell Number	Cell Name	Collection Date	DNA Species ID	Sex ID	Individual ID
MTFWP	362.12.12.2016.W03	362	Blodgett Canyon	12/12/2016	Wolverine	poor DNA	poor DNA
MTFWP	342.14.12.2016.W01	342	Granulated Mountain	12/14/2016	Wolverine	Male	MS-Gulo-M1
MTFWP	342.14.12.2016.W04	342	Granulated Mountain	12/14/2016	Wolverine	poor DNA	poor DNA
MTFWP	342.14.12.2016.W03	342	Granulated Mountain	12/14/2016	Wolverine	Male	MS-Gulo-M1
MTFWP	347.16.12.2016.W03	347	Carpp Ridge	12/16/2016	Wolverine	poor DNA	poor DNA
MTFWP	347.16.12.2016.W04	347	Carpp Ridge	12/16/2016	Wolverine	Female	MS-Gulo-F2
MTFWP	347.16.12.2016.T02	347	Carpp Ridge	12/16/2016	Wolverine	Female	MS-Gulo-F2
MTFWP	347.16.12.2016.T03	347	Carpp Ridge	12/16/2016	Wolverine	Female	MS-Gulo-F2
MTFWP	307.04.01.2017.W01	307	Twin Lakes	1/4/2017	Wolverine	Male	MS-Gulo-M4
MTFWP	307.04.01.2017.W03	307	Twin Lakes	1/4/2017	Wolverine	Male	MS-Gulo-M4
MTFWP	307.04.01.2017.W04	307	Twin Lakes	1/4/2017	Wolverine	Male	MS-Gulo-M4
MTFWP	307.04.01.2017.L01	307	Twin Lakes	1/4/2017	Wolverine	Male	MS-Gulo-M4
MTFWP	307.04.01.2017.L02	307	Twin Lakes	1/4/2017	Wolverine	poor DNA	poor DNA





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MTFWP	307.04.01.2017.L03	307	Twin Lakes	1/4/2017	Wolverine	poor DNA	poor DNA
MTFWP	1	412	Hoodoo	1/6/2016	Wolverine	poor DNA	poor DNA
MTFWP	4	536	DNRC Whitefish	12/28/2016	Wolverine	Female	MS-Gulo-F5
MTFWP	5	536	DNRC Whitefish	12/28/2016	Wolverine	Female	MS-Gulo-F5
MTFWP	6	536	DNRC Whitefish	12/28/2016	Wolverine	poor DNA	poor DNA
MTFWP	8	412	Hoodoo	12/8/2016	Wolverine	poor DNA	poor DNA
MTFWP	9	412	Hoodoo	12/8/2016	Wolverine	Male	MS-Gulo-M6
MTFWP	10	412	Hoodoo	12/8/2016	Wolverine	Male	MS-Gulo-M6
MTFWP	11	412	Hoodoo	12/8/2016	Wolverine	Male	MS-Gulo-M6
MTFWP	12	412	Hoodoo	12/8/2016	Wolverine	poor DNA	poor DNA
FNF	485_02012017_01	485	n/a	2/1/2017	Wolverine	Female	MS-Gulo-F7
FNF	485_02012017_03	485	n/a	2/1/2017	Wolverine	Female	MS-Gulo-F7
FNF	485_02012017_05	485	n/a	2/1/2017	Wolverine	poor DNA	poor DNA
WYGF	2017.02.08.161.v3L1	161	Wyoming-North, Blackwater Cr.	2/8/2017	Wolverine	poor DNA	poor DNA
IDFG	McCall5	293	McCall	12/8/2016	Wolverine	Male	MS-Gulo-M8
IDFG	McCall7	293	McCall	12/8/2016	Wolverine	Male	MS-Gulo-M8
IDFG	McCall8	166	McCall	12/19/2016	Wolverine	poor DNA	poor DNA
IDFG	McCall10	166	McCall	12/19/2016	Wolverine	Female	MS-Gulo-F9
IDFG	McCall26	251	McCall	1/4/2017	Wolverine	Female	MS-Gulo-F10
IDFG	McCall27	251	McCall	1/4/2017	Wolverine	Female	MS-Gulo-F10
IDFG	McCall29	214	McCall	1/5/2017	Wolverine	Male	MS-Gulo-M11
IDFG	McCall30	214	McCall	1/5/2017	Wolverine	Male	MS-Gulo-M11
IDFG	McCall31	214	McCall	1/5/2017	Wolverine	Male	MS-Gulo-M11
IDFG	McCall32	214	McCall	1/5/2017	Wolverine	Male	MS-Gulo-M11
IDFG	McCall33	214	McCall	1/5/2017	Wolverine	Male	MS-Gulo-M11
IDFG	258-V2-1	258	Salmon	1/24/2017	Wolverine	Male	MS-Gulo-M12
IDFG	258-V2-2	258	Salmon	1/24/2017	Wolverine	Male	MS-Gulo-M12
IDFG	258-V2-3	258	Salmon	1/24/2017	Wolverine	Male	MS-Gulo-M12
IDFG	258-V2-4	258	Salmon	1/24/2017	Wolverine	Male	MS-Gulo-M12
IDFG	214-V2-4	214	McCall	2/2/2017	Wolverine	Female	MS-Gulo-F13
IDFG	293-V3-1	293	McCall	2/15/2017	Wolverine	poor DNA	poor DNA
IDFG	293-V3-2	293	McCall	2/15/2017	Wolverine	poor DNA	poor DNA
IDFG	293-V3-3	293	McCall	2/15/2017	Wolverine	Female	MS-Gulo-F14
IDFG	293-V3-4	293	McCall	2/15/2017	Wolverine	Female	MS-Gulo-F14
IDFG	9	78	Stanley	12/13/2016	Wolverine	Female	MS-Gulo-F15
IDFG	12	78	Stanley	12/13/2016	Wolverine	Female	MS-Gulo-F15
IDFG	14	167	Stanley	1/6/2017	Wolverine	Female	MS-Gulo-F9
IDFG	15	167	Stanley	1/6/2017	Wolverine	Female	MS-Gulo-F9
IDFG	16	167	Stanley	1/6/2017	Wolverine	Female	MS-Gulo-F9
IDFG	17	167	Stanley	1/6/2017	Wolverine	poor DNA	poor DNA
IDFG	18	167	Stanley	1/6/2017	Wolverine	Female	MS-Gulo-F9
IDFG	20	167	Stanley	1/6/2017	Wolverine	poor DNA	poor DNA
IDFG	26	91	Stanley	1/14/2017	Wolverine	poor DNA	mixed with marten
IDFG	31	170	Stanley	1/16/2017	Wolverine	Female	MS-Gulo-F16
IDFG	32	170	Stanley	1/16/2017	Wolverine	poor DNA	poor DNA





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IDFG	33	170	Stanley	1/16/2017	Wolverine	Female	MS-Gulo-F16
WYGF	2017.02.14.101.Track1	NA	n/a	2/14/2017	Wolverine	Male	MS-Gulo-M17
WYGF	2017.03.15.160.V4L1	160	n/a	3/15/2017	Wolverine	poor DNA	poor DNA
WYGF	2017.03.15.160.V4W1	160	n/a	3/15/2017	Wolverine	poor DNA	mixed with marten
WYGF	2017.03.15.160.V4W4	160	n/a	3/15/2017	Wolverine	poor DNA	mixed with marten
WYGF	2017.03.17.100.V4W2	100	n/a	3/17/2017	Wolverine	poor DNA	poor DNA
WYGF	2017.03.17.100.V4W3	100	n/a	3/17/2017	Wolverine	poor DNA	poor DNA
MTFWP	520.01.27.2017.W01	520	Two Med Lake	1/27/2017	Wolverine	poor DNA	mixed with marten
MTFWP	520.01.27.2017.L01	520	Two Med Lake	1/27/2017	Wolverine	poor DNA	poor DNA
MTFWP	520.01.27.2017.L02	520	Two Med Lake	1/27/2017	Wolverine	Female	MS-Gulo-F18
MTFWP	536.02.28.2017.W03	536	Whitefish, DNRC	2/28/2017	Wolverine	Female	MS-Gulo-F5
IDFG	McCall34	293	McCall	1/17/2017	Wolverine	Female	MS-Gulo-F14
IDFG	McCall36	293	McCall	1/17/2017	Wolverine	Female	MS-Gulo-F14
IDFG	McCall37	293	McCall	1/17/2017	Wolverine	poor DNA	poor DNA
IDFG	293-V4-1	293	McCall	3/8/2017	Wolverine	Female	MS-Gulo-F14
IDFG	293-V4-2	293	McCall	3/8/2017	Wolverine	poor DNA	poor DNA
IDFG	293-V4-4	293	McCall	3/8/2017	Wolverine	poor DNA	poor DNA
IDFG	41	91	Stanley	2/12/2017	Wolverine	poor DNA	poor DNA
IDFG	43	91	Stanley	2/12/2017	Wolverine	Male	MS-Gulo-M19
IDFG	46	91	Stanley	2/12/2017	Wolverine	Male	MS-Gulo-M19
IDFG	67	127	Stanley	3/6/2017	Wolverine	Male	MS-Gulo-M19
IDFG	237-v3-6	237	Salmon	3/1/2017	Wolverine	Female	MS-Gulo-F20
IDFG	237-v3-7	237	Salmon	3/1/2017	Wolverine	poor DNA	poor DNA
IDFG	237-v3-7	237	Salmon	3/1/2017	Wolverine	poor DNA	poor DNA
MTFWP	287.17.01.2017.L01	287	Snow Lake/Goose	17-Jan-17	Wolverine	poor DNA	poor DNA
MTFWP	259.21.02.2017.W01	259	Taylor Fork	21-Feb-17	Wolverine	Female	MS-Gulo-F21
MTFWP	287.27.02.2017.W02	287	Snow Lake/Goose	27-Feb-16	Wolverine	Female	MS-Gulo-F22
MTFWP	287.27.02.2017.L01	287	Snow Lake/Goose	27-Feb-16	Wolverine	Female	MS-Gulo-F22
FNF	Wolverine Collar 1	498	Don Scharff Volunteer Cell	3/9/17	Wolverine	Male	MS-Gulo-M23
FNF	Wolverine Collar 2	498	Don Scharff Volunteer Cell	3/9/17	Wolverine	poor DNA	poor DNA
FNF	Wolverine Collar 3	498	Don Scharff Volunteer Cell	3/9/17	Wolverine	Male	MS-Gulo-M23
IDFG	258.FV.1	258	258.FV.1	4/4/17	Wolverine	Male	MS-Gulo-M12
IDFG	258.FV.5	258	258.FV.5	4/4/17	Wolverine	Male	MS-Gulo-M12
IDFG	251-Pull-1	251	251-Pull-1	4/5/17	Wolverine	poor DNA	poor DNA
IDFG	251-Pull-3	251	251-Pull-3	4/5/17	Wolverine	poor DNA	poor DNA
IDFG	166-P-1	166	166-P-1	4/6/17	Wolverine	Female	MS-Gulo-F9
IDFG	166-P-2	166	166-P-2	4/6/17	Wolverine	Female	MS-Gulo-F9
IDFG	166-P-4	166	166-P-4	4/6/17	Wolverine	Female	MS-Gulo-F9
IDFG	214-P-1	214	214-P-1	4/6/17	Wolverine	Female	MS-Gulo-F13
IDFG	84	148	84	4/4/17	Wolverine	Female	MS-Gulo-F16
IDFG	90	127	90	4/7/17	Wolverine	Male	MS-Gulo-M19
IDFG	91	127	91	4/7/17	Wolverine	Male	MS-Gulo-M19
IDFG	92	127	92	4/7/17	Wolverine	poor DNA	poor DNA
IDFG	93	127	93	4/7/17	Wolverine	Male	MS-Gulo-M19





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IDFG	94	127	94	4/7/17	Wolverine	Male	MS-Gulo-M19
IDFG	82	170	82	4/3/17	Wolverine	Female	MS-Gulo-F16
MTFWP	287.21.04.2017.W02	287	Snow Lake/Goose	4/21/17	Wolverine	poor DNA	poor DNA
MTFWP	350.29.01.2017.W02	350	Wahoo	1/29/17	Wolverine	Male	MS-Gulo-M25
MTFWP	350.29.01.2017.W04	350	Wahoo	1/29/17	Wolverine	poor DNA	poor DNA
MTFWP	441.01.02.2017.W01	441	Lucifer	2/1/17	Wolverine	Male	MS-Gulo-M26
MTFWP	441.01.02.2017.W02	441	Lucifer	2/1/17	Wolverine	Male	MS-Gulo-M26
MTFWP	441.01.02.2017.W03	441	Lucifer	2/1/17	Wolverine	Male	MS-Gulo-M26
MTFWP	441.01.02.2017.W04	441	Lucifer	2/1/17	Wolverine	Male	MS-Gulo-M26
MTFWP	342.02.02.2017.W01	342	Granulated	2/2/17	Wolverine	poor DNA	poor DNA
MTFWP	342.02.02.2017.W02	342	Granulated	2/2/17	Wolverine	poor DNA	poor DNA
MTFWP	347.08.02.2017.W01	347	Carpp	2/8/17	Wolverine	poor DNA	multiple individuals
MTFWP	347.08.02.2017.W03	347	Carpp	2/8/17	Wolverine	Female	MS-Gulo-F2
MTFWP	362.18.02.2017.L02	362	Blodgett	2/18/17	Wolverine	Male	MS-Gulo-M25
MTFWP	554.25.03.2017.W01	554	Pass Creek	3/25/17	Wolverine	poor DNA	multiple individuals
MTFWP	342.02.03.2017.W01	342	Granulated	3/2/17	Wolverine	poor DNA	poor DNA
MTFWP	342.02.03.2017.W02	342	Granulated	3/2/17	Wolverine	Male	MS-Gulo-M27
MTFWP	342.02.03.2017.W03	342	Granulated	3/2/17	Wolverine	poor DNA	multiple individuals
MTFWP	342.02.03.2017.W04	342	Granulated	3/2/17	Wolverine	Male	MS-Gulo-M28
MTFWP	342.02.03.2017.L01	342	Granulated	3/2/17	Wolverine	poor DNA	poor DNA
MTFWP	342.02.03.2017.L02	342	Granulated	3/2/17	Wolverine	poor DNA	poor DNA
MTFWP	347.03.03.2017.W01	347	Carpp	3/3/17	Wolverine	Female	MS-Gulo-F2
MTFWP	347.03.03.2017.W04	347	Carpp	3/3/17	Wolverine	Female	MS-Gulo-F2
MTFWP	353.04.03.2017.W02	353	Fox	3/4/17	Wolverine	poor DNA	poor DNA
MTFWP	353.04.03.2017.W03	353	Fox	3/4/17	Wolverine	poor DNA	multiple individuals
MTFWP	353.04.03.2017.W04	353	Fox	3/4/17	Wolverine	Female	MS-Gulo-F24
MTFWP	353.04.03.2017.T01	353	Fox	3/4/17	Wolverine	Female	MS-Gulo-F24
MTFWP	441.06.03.2017.W01	441	Lucifer	3/6/17	Wolverine	Female	MS-Gulo-F29
MTFWP	441.06.03.2017.W04	441	Lucifer	3/6/17	Wolverine	Female	MS-Gulo-F29
MTFWP	415.22.03.2017.W01	415	Neihart	3/23/17	Wolverine	Male	MS-Gulo-M31
MTFWP	415.22.03.2017.W03	415	Neihart	3/25/17	Wolverine	Male	MS-Gulo-M31
MTFWP	415.22.03.2017.W04	415	Neihart	3/26/17	Wolverine	poor DNA	poor DNA
MTFWP	415.22.03.2017.L01	415	Neihart	3/27/17	Wolverine	poor DNA	poor DNA
MTFWP	446.2017.03.28.L01	446	Crown	3/28/17	Wolverine	Male	MS-Gulo-M30
MTFWP	415.22.03.2017.L03	415	Neihart	3/29/17	Wolverine	Male	MS-Gulo-M31
MTFWP	536.2017.03.30.W01	536	Whitefish	3/30/17	Wolverine	Female	MS-Gulo-F5
MTFWP	536.2017.03.30.W02	536	Whitefish	3/30/17	Wolverine	Female	MS-Gulo-F5
MTFWP	536.2017.03.30.W04	536	Whitefish	3/30/17	Wolverine	Female	MS-Gulo-F5
MTFWP	427.03.04.2017.W01	427	SF Jocko	4/3/17	Wolverine	Female	MS-Gulo-F32
MTFWP	429.04.04.2017.W01	429	Copper	4/4/17	Wolverine	Female	MS-Gulo-F33
MTFWP	429.04.04.2017.W02	429	Copper	4/4/17	Wolverine	Female	MS-Gulo-F33
MTFWP	429.04.04.2017.W04	429	Copper	4/4/17	Wolverine	Female	MS-Gulo-F33
MTFWP	429.04.04.2017.L01	429	Copper	4/4/17	Wolverine	poor DNA	poor DNA
MTFWP	442.03.04.2017.W01	442	Richmond Ridge	4/3/2017	Wolverine	poor DNA	poor DNA





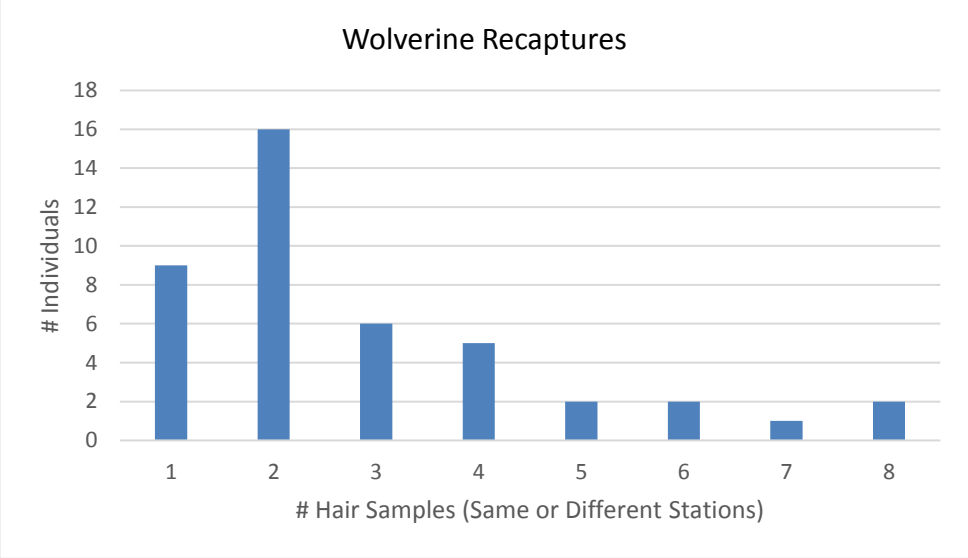
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MTFWP	442.03.04.2017.W03	442	Richmond Ridge	4/3/2017	Wolverine	Female	MS-Gulo-F34
MTFWP	307.02.04.2017.W04	307	Twin	4/2/17	Wolverine	Male	MS-Gulo-M35
MTFWP	307.02.04.2017.L01	307	Twin	4/2/17	Wolverine	Male	MS-Gulo-M35
MTFWP	441.04.04.2017.W01	441	Lucifer	4/4/17	Wolverine	poor DNA	poor DNA
MTFWP	446.2017.04.19.W01	446	Crown	4/19/17	Wolverine	Female	MS-Gulo-F36
MTFWP	446.2017.04.19.W02	446	Crown	4/19/17	Wolverine	Female	MS-Gulo-F36
MTFWP	446.2017.04.19.W03	446	Crown	4/19/17	Wolverine	Female	MS-Gulo-F36
MTFWP	446.2017.04.19.L01	446	Crown	4/19/17	Wolverine	poor DNA	poor DNA
MTFWP	342.21.04.2017.W04	342	Granulated	4/21/17	Wolverine	Male	MS-Gulo-M28
MTFWP	347.03.04.2017.W01	347	Carpp	4/3/2017.	Wolverine	Female	MS-Gulo-F2
MTFWP	347.03.04.2017.W04	347	Carpp	4/3/2017.	Wolverine	Female	MS-Gulo-F2
FNF	9	479	Pentagon	5/16/17	Wolverine	Male	MS-Gulo-M3
MTFWP	LC1-4	302	Lambert Creek	3/1/17	Wolverine	poor DNA	poor DNA
MTFWP	LC1-5	302	Lambert Creek	3/1/17	Wolverine	poor DNA	mixed with marten
WYGF	073.2017.05.30.v1.pc.B	73	WY-N	5/30/2017	Wolverine	poor DNA	poor DNA
WDFW	060--3	419(60)	Placer	2/21/2017	Wolverine	poor DNA	poor DNA
WDFW	060--4	419(60)	Placer	2/21/2017	Wolverine	poor DNA	poor DNA
WDFW	060--5	419(60)	Placer	2/21/2017	Wolverine	poor DNA	poor DNA
WDFW	060--7	419(60)	Placer	2/21/2017	Wolverine	Female	MS-Gulo-F37
WDFW	060--8	419(60)	Placer	2/21/2017	Wolverine	Female	MS-Gulo-F37
WDFW	108--2	381(108)	Alder Creek	1/1/2017	Wolverine	Male	MS-Gulo-M38
WDFW	108--3	381(108)	Alder Creek	1/1/2017	Wolverine	Male	MS-Gulo-M38
WDFW	WA-156-003	558(156)	Mt Shuksan South	6/7/2017	Wolverine	poor DNA	poor DNA
WDFW	ID-96-1	515(96)	Company Creek	8/23/2017	Wolverine	poor DNA	poor DNA
WDFW	ID-96-2	515(96)	Company Creek	8/23/2017	Wolverine	poor DNA	poor DNA
WDFW	id-96-3	515(96)	Company Creek	8/23/2017	Wolverine	poor DNA	poor DNA
WDFW	ID-96-4	515(96)	Company Creek	8/23/2017	Wolverine	poor DNA	poor DNA
WDFW	ID-96-6	515(96)	Company Creek	8/23/2017	Wolverine	poor DNA	poor DNA
WDFW	502 hair1_1-7	502(160)	Perry Creek	1/7/2017	Wolverine	poor DNA	poor DNA
WDFW	Cell 526 sample 11	526(15)	Twisp River	1/27/2017	Wolverine	Female	MS-Gulo-F39
WDFW	Cell 526 sample 29	526(15)	Twisp River	3/2/2017	Wolverine	poor DNA	poor DNA
WDFW	Cell 526 sample 30	526(15)	Twisp River	3/2/2017	Wolverine	Male	MS-Gulo-M40
WDFW	Cell 526 sample 31	526(15)	Twisp River	3/2/2017	Wolverine	poor DNA	poor DNA
WDFW	Cell 526 sample 32	526(15)	Twisp River	3/2/2017	Wolverine	Male	MS-Gulo-M40
WDFW	Cell 526 sample 33	526(15)	Twisp River	3/2/2017	Wolverine	poor DNA	poor DNA
WDFW	Cell 533 Sample 3	533(80)	Bridge Creek	1/6/2017	Wolverine	Male	MS-Gulo-M40
WDFW	Cell 533 Sample 18	533(80)	Bridge Creek	2/3/2017	Wolverine	poor DNA	poor DNA
WDFW	Cell 533 Sample 19	533(80)	Bridge Creek	2/3/2017	Wolverine	poor DNA	poor DNA
WDFW	Cell 533 Sample 20	533(80)	Bridge Creek	2/3/2017	Wolverine	poor DNA	poor DNA
WDFW	Cell 533 Sample 21	533(80)	Bridge Creek	2/3/2017	Wolverine	Female	MS-Gulo-F39
WDFW	Cell 533 Sample 22	533(80)	Bridge Creek	2/3/2017	Wolverine	Female	MS-Gulo-F39
WDFW	Cell 533 Sample 23	533(80)	Bridge Creek	2/3/2017	Wolverine	Female	MS-Gulo-F39
WDFW	Cell 533 Sample 24	533(80)	Bridge Creek	2/3/2017	Wolverine	poor DNA	poor DNA
WDFW	Cell 457 Sample 12	457(64)	Beverly	4/25/2017	Wolverine	poor DNA	mixed with marten
IDFG	95-Pull-2	95	Salmon	8/3/2017	Wolverine	poor DNA	poor DNA

IDFG	95-Pull-4	95	Salmon	8/3/2017	Wolverine	poor DNA	poor DNA
IDFG	95-Pull-5	95	Salmon	8/3/2017	Wolverine	poor DNA	poor DNA
IDFG	109-Pull-1	109	R3-BNF	7/29/2017	Wolverine	poor DNA	poor DNA
IDFG	109-Pull-3	109	R3-BNF	7/29/2017	Wolverine	poor DNA	poor DNA
MTFWP	344_W3	344	n/a	7/15/2017	Wolverine	Male	MS-Gulo-M25
IDFG	79VOL-1	79	Boulder Lake	7/16/2017	Wolverine	poor DNA	multiple individuals
WYGF_2016	27.2016.01.07.V2A	27	Miller Park	1/17/2016	Wolverine	poor DNA	poor DNA
WYGF_2016	27.2016.01.07.V2C	27	Miller Park	1/17/2016	Wolverine	poor DNA	poor DNA
WYGF_2016	016.2016.01.20.v2a	16	Cony Mtn.	1/20/2016	Wolverine	Female	MS-Gulo-F41
WYGF_2016	016.2016.01.20.v2b	16	Cony Mtn.	1/20/2016	Wolverine	Female	MS-Gulo-F41
WYGF_2016	016.2016.01.20.v2c	16	Cony Mtn.	1/20/2016	Wolverine	Female	MS-Gulo-F41
WYGF_2016	016.2016.01.20.v2d	16	Cony Mtn.	1/20/2016	Wolverine	Female	MS-Gulo-F41
WYGF_2016	027.2016.02.05.v3a	27	Miller Park	2/5/2016	Wolverine	Male	MS-Gulo-M42
WYGF_2016	027.2016.02.05.v3d	27	Miller Park	2/5/2016	Wolverine	Male	MS-Gulo-M42
WYGF_2016	027.2016.03.16.v4	27	Miller Park	3/16/2016	Wolverine	Female	MS-Gulo-F43
WYGF_2016	085.2016.03.21.v4	85	Five Pockets	3/21/2016	Wolverine	poor DNA	poor DNA
WYGF_2016	027.2016.04.08.v5	27	Miller Park	4/8/2016	Wolverine	Female	MS-Gulo-F43

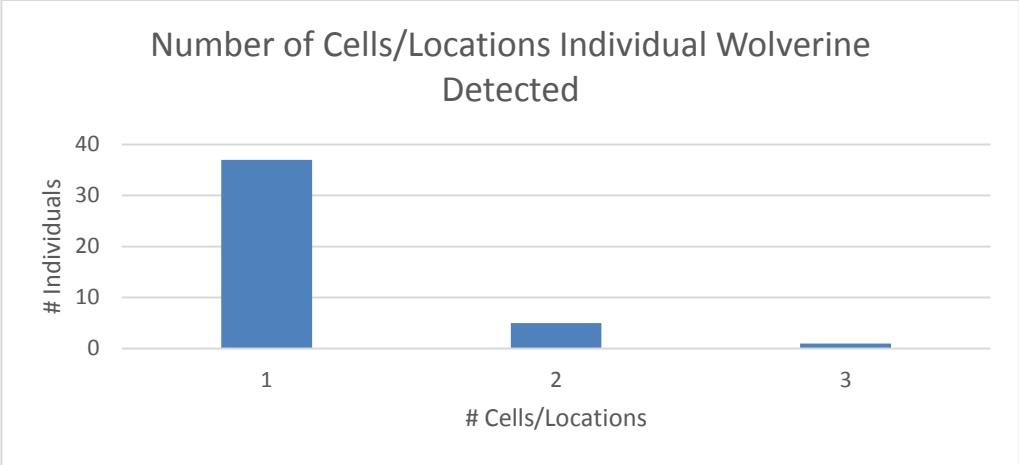
Nine of the 43 individual wolverine were captured from only one hair sample. Thirty-four were captured multiple times. Two individuals were identified from 8 hair samples (Figure 1).

Figure 1. The number of hair samples identifying each individual wolverine



Individual identification was obtained for wolverine collected from 39 cells (and one snow tracking sample submitted by WYGF). The majority of individual wolverine (37) were detected only at one cell/location (Figure 2). Five wolverine were detected at two cells/locations and one wolverine was detected at 3 stations (M25; Figure 2).

Figure 2. The number of cells/locations that wolverine individuals were detected.



We are currently finishing DNA haplotyping work at the mitochondrial DNA control region. These results will be available shortly. Please let us know if you have any questions; we look forward to working with you in the future.

Appendix A.
Data Generation and Quality Control (Genotyping Error Checking)

Genotyping errors, such as allelic dropout or false alleles, are prominent with low quantity DNA samples from hairs, scats, or other forensic or non-invasive genetic sampling. Our National Genomics Center uses at least two independent approaches to catch and remove these errors. First, DNA from hair samples are run and analyzed at least two times at each microsatellite locus (Taberlet et al. 1996). Allele scores are trusted if there is concordance between runs. Samples that failed or are inconsistent are re-analyzed until they either produced a high quality genotype or are culled from the dataset and labeled as "no result - poor quality DNA". Second, the entire database is subject to at least one computer screen such as program DROPOUT (McKelvey and Schwartz 2005), ALLELEMATCH (Galpern 2012), MICROCHECKER (Van Oosterhout et al. 2004), or a GIS based analysis (Marucco et al. 2011). Which screen we use is dependent on the use of the data and the life history of the species under study.

For the data in this study we used the multiple analysis approach (multi-tube approach), the two screens implemented in program DROPOUT (following the protocol outlined in Schwartz et al 2006) and MICROCHECKER.

Computer Screens for Quality Control

The first screen we implemented was the EB test in DROPOUT. After running the first test we found that individuals differed by four or more loci (Figure A1). This suggests that there are no errors in the dataset, as errors often present themselves as a second mode on the left side of the graph (e.g., samples that differ at 1, 2 or 3 loci different). The second test implemented in DROPUT is the DCH test. This test uses a subset of loci that achieve a reasonable probability of identity, then tests what occurs when a new locus is added to this subset. In theory, once the minimum number of loci needed to identify unique individuals is achieved any extra locus should not add a new individual. If a new individual is detected when this new locus is added then it is probably due to genotyping error at that locus. When each locus is used as a "new locus" past the number needed to identify individuals, we can evaluate if that locus may have errors. In this dataset no microsatellite locus was found to have errors falsely increasing the number of individuals in the dataset (Figure A2). For the records, the dataset had adequate power to detect individuals with a subset of loci. We calculated the probability of identity (PI; Paetkau & Strobeck 1994) and the probability that siblings are identical (PIsib; Evett & Weir 1998; Waits et al. 2001) for this data as $3.62E\text{-}^{09}$ and $1.61E\text{-}^{04}$, respectively.

Figure A1. Results of the Examining Bimodality (EB) test to detect genotyping errors using the wolverine dataset from the Multi-State Survey. There is no sign of error in this database

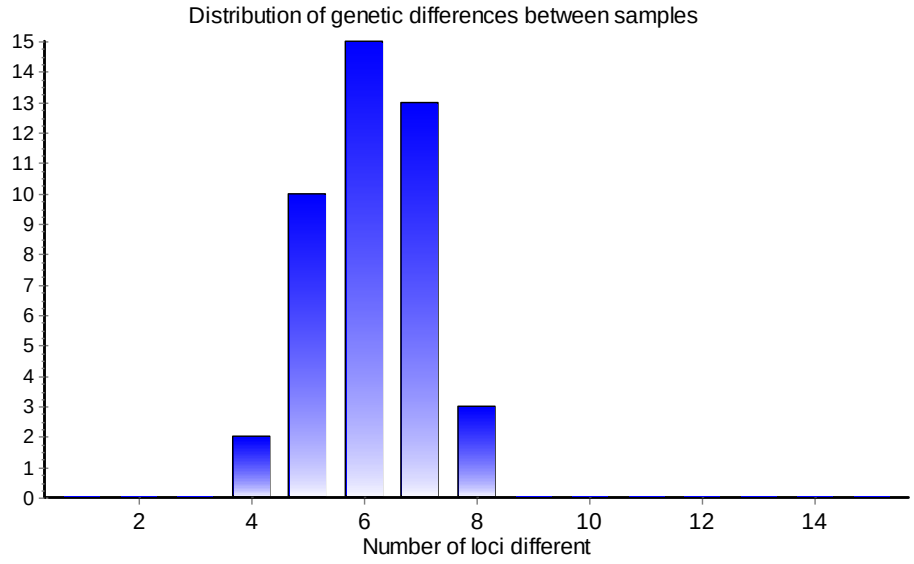
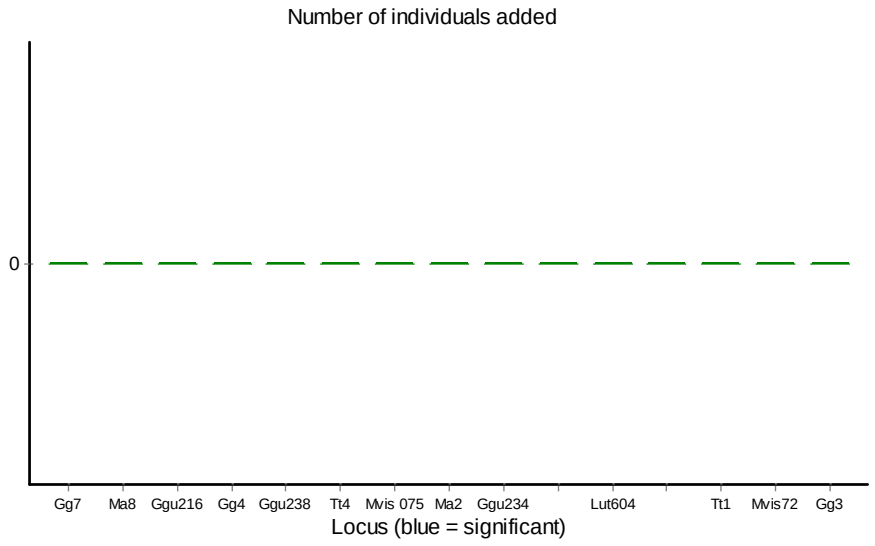


Figure A2. Results of the Difference in Capture History (DCH) test designed to detect loci with errors erroneously contributing to the number of individuals in the data. There is no sign of error in your database.



Data were also evaluated using the program Micorchecker (VanOosterhout et al. 2004) which looks for population genetic signals that may indicate allelic dropout or false alleles, producing homozygote excesses at individual microsatellite loci. We tested the 39 individuals identified from ID, MT, and WY using this analyses (there were too few individuals from WA to run this analysis; and based on previous studies we know that there is substantial substructure between the Rockies and the Cascades; Schwartz et

al. 2007). None of the loci in this study showed evidence of having null alleles (Table A1). A "yes" in the "null present" column indicates that samples at that locus may have allelic dropout. Microchecker also reported: No significant evidence for homozygote excess, "No evidence for scoring error due to stuttering", "No evidence for large allele dropout", and "No evidence for null alleles."

Given all of the genotyping error screens, we feel that errors are minimal or non-existent in this dataset. We encourage you to scrutinize these data with any additional field information that you have available.

Table A1. Summary statistics for null allele estimates for the Multi-State Survey wolverine data

Locus	Null Present?	Oosterhout
Gg7	no	-0.06
Ma8	no	0.02
Ggu216	no	0.06
Gg4	no	-0.08
Ggu238	no	-0.03
Tt4	no	0.01
Mvis 075	no	0.06
Ma2	no	0.11
Ggu234	no	-0.05
Ggu101	no	-0.17
Lut604	no	0.05
Mvis020	no	0.11
Tt1	no	0.11
Mvis72	no	-0.08
Gg3	no	-0.08

From: Grizzle, Betty
To: Guinotte, John
Subject: Re:
Date: Monday, February 12, 2018 4:31:31 PM

Might be something else to add, with revisions. Is the point that the Rhoades et al. study developed projections that are further out in time vs. Ray et al. 2017 report, only? Some of the other models evaluated in the SSA Report, for other regions, do go past 2050s.

On Mon, Feb 12, 2018 at 3:20 PM, Guinotte, John <john_guinotte@fws.gov> wrote:

> Hi Betty, Is this useful to you?
>
> John Guinotte
> Spatial Ecologist
> Ecological Services
> U.S. Fish and Wildlife Service
> Mountain Prairie Region 6
> 134 Union Blvd., Lakewood, CO 80228
> 303-236-4264
> john_guinotte@fws.gov
>
> ---------- Forwarded message ----------
> From: **Andrea Ray - NOAA Federal** <andrea.ray@noaa.gov>
> Date: Mon, Feb 12, 2018 at 4:13 PM
> Subject: Re:
> To: "Guinotte, John" <john_guinotte@fws.gov>, Steve Torbit <Stephen_Torbit@fws.gov>
>
> John -- I'm double checking a couple of numbers, hope to hear back from
>
> here it is tentatively, let me know if this is off base -- I'll send later as a final word doc and with the figures as separate high res.
>
> Although analysis of snowpack at the end of century was beyond the scope of the analysis done for FWS, a recent study analyzes snow to 2100 for several western mountain ranges (Rhoades et al, 2018), which was also used in the National Climate Assessment (Figure 8.2 and Wehner et al 2017). They simulate the changes in several snow-related variables using the NCAR CESM model and RCP8.5 (the higher emissions pathway), which is then run through the VIC hydrology model. Compared to the GCMs analyzed for FWS, the NCAR CESM model these Rhoades et al (2018) used is hot and very wet for GLAC by end of century (JOHN – I'm double checking these numbers +5.0 C and +21.4% precipitation), with a somewhat higher temperature and precipitation increase than the spread of the models we used for 2041-2070. For GLAC, CESM is hot and wet (+5.1 C and +11.1% precipitation), although within the spread of the models we used for 2041-2070.
>
> Rhodes et al (2018) find resiliency in snowpack for interior mountain ranges for the 2075-2010 period compared to the 1980-2005 climatology they use. For this period,

at higher elevations (2000 to >2500m), they find snowfall changes in the Rockies to be to be nominally positive or negative (their figure 10, panels e, j, o). A high snow-rain ratio is also maintained in the higher elevation Rockies (Fig. 8 panels, c and e). They also find that the freezing levels for interior mountain ranges (Rockies, Wasatch and Uintas) does not move as far upslope as in the coastal mountain ranges they also analyze. In the Rockies, the freezing range moves up from 500 m to 1000m.

Rhoades, A. M., P. A. Ullrich, and C. M. Zarzycki, 2017: Projecting 21st century snowpack trends in western USA mountains using variable-resolution CESM. *Climate Dynamics*, **44**, 1-28, doi:10.1007/s00382-017-3606-0
https://link.springer.com/article/10.1007%2Fs00382-017-3606-0

Wehner, M.F., J.R. Arnold, T. Knutson, K.E. Kunkel, and A.N. LeGrande, 2017: Droughts, floods, and wildfires. In: *Climate Science Special Report: Fourth National Climate Assessment, Volume I* [Wuebbles, D.J., D.W. Fahey, K.A. Hibbard, D.J. Dokken, B.C. Stewart, and T.K. Maycock (eds.)]. U.S. Global Change Research Program, Washington, DC, USA, pp. 231-256, doi: 10.7930/J0CJ8BNN.

On Fri, Feb 9, 2018 at 1:28 PM, Guinotte, John <john_guinotte@fws.gov> wrote:

> Hi Andrea, Are you going to send me the paragraph we spoke about? We are getting close to deadline. I made a few edits to your abstract and Steve said he'd look when he gets back to Denver this weekend. Thanks John
>
> John Guinotte
> Spatial Ecologist
> Ecological Services
> U.S. Fish and Wildlife Service
> Mountain Prairie Region 6
> 134 Union Blvd., Lakewood, CO 80228
> 303-236-4264
> john_guinotte@fws.gov

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Andrea J. Ray, Ph.D.
NOAA Earth System Research Lab, Mailcode R/PSD1
325 Broadway, Boulder, CO 80305-3328
(tel) 303-497-6434
(fax) 303-497-6449
andrea.ray@noaa.gov
www.researchgate.net/profile/Andrea_Ray2

--

Betty J. Grizzle, D.Env., MPH
Fish and Wildlife Biologist
U.S. Fish and Wildlife Service
Carlsbad Fish and Wildlife Office
2177 Salk Ave, Suite 250
Carlsbad, CA 92008
760-431-9440, ext. 215
760-431-5901 fax

From: Shoemaker, Justin
To: Marjorie Nelson; Jodi Bush; Grizzle, Betty
Subject: Revised wolverine timeline
Date: Tuesday, February 13, 2018 1:05:54 PM
Attachments: Wolverine Detailed Timeline_post RTM_02132018.docx

---

Let me know if you have any concerns w/ this. I'll get it out to the group soon.

Justin Shoemaker
Classification and Recovery Biologist
U.S. Fish and Wildlife Service, Region 6
Phone: 309-757-5800 x214
Email: justin_shoemaker@fws.gov

Wolverine Listing Determination Timeline
2/13/18 version

Task	Responsible Parties	Dates	Length of time
Listing Decision Analysis Phase			
SSA report to recommendation team	Justin Shoemaker, Jodi Bush	Jan 10, 2018	At least 2 weeks prior to recommendation team meeting
Decision meeting	RDs or delegates, ARDs, other management, SSA core team	Jan 25, 2018	1 days
Draft decision summary for the record or certify decision meeting notes	R6 RD or delegate	Mid Feb 2018	
Briefing for Acting Director	Greg Sheehan, Gary Frazer, Bridget Fahey, Noreen Walsh, Mike Thabault, Marj Nelson, Justin Shoemaker, Jodi Bush, Betty Grizzle, Paul Souza, Sarah Quamme	Feb 20	30 min
Process for final withdrawal of proposed listing			
Draft final withdrawal (not-warranted) FR notice	Justin Shoemaker and Betty Grizzle	Feb 20 – April 23, 2018	
Core team reviews FR notice, make revisions, Prelim SOL review	SSA core team, Justin Shoemaker	April 23 – May 7, 2018	2 weeks
SOL surname and revisions	DOI SOL	May 7 – May 21, 2018	2 weeks
Regional Office Surnames, Regional concurrence and revisions	Marjorie Nelson, Mike Thabault, Matt Hogan, Noreen Walsh, and concurring regional RDs/ARDs or delegates	May 21 – June 4, 2018	2 weeks
PPM review and revisions	PPM	June 4 – June 18, 2018	2 weeks
HQ review	Sarah Quamme, Bridget Fahey	June 18 – July 2, 2018	submit to HQ 2 wks and 2 mo prior to FR submittal date
Asst. Director for ES Surname	Asst. Director for ES	July 2 – July 10, 2018	5 business days
FWS Director Surname	Director of FWS	July 10 – July 17, 2018	5 business days

*Includes States, Tribes, Federal Agencies

Fish, Wildlife, and Parks Surname	FWP	July 17 – August 17, 2018	1 month
Executive Secretary Surname	Executive Secretary's Office	August 17 – August 30, 2018	10 business days
Deliver to FR	HQ	Aug 31, 2018	
Publication of withdrawal	Federal Register	Sept 7, 2018	

*Includes States, Tribes, Federal Agencies

From: Grizzle, Betty
To: Shoemaker, Justin
Subject: Re: Revised wolverine timeline
Date: Tuesday, February 13, 2018 3:15:26 PM

For my tasks/responsibilities, I think this is fine.

On Tue, Feb 13, 2018 at 12:05 PM, Shoemaker, Justin <justin_shoemaker@fws.gov> wrote:

> Let me know if you have any concerns w/ this. I'll get it out to the group soon.
>
> Justin Shoemaker
> Classification and Recovery Biologist
> U.S. Fish and Wildlife Service, Region 6
> Phone: 309-757-5800 x214
> Email: justin_shoemaker@fws.gov

--

Betty J. Grizzle, D.Env., MPH
Fish and Wildlife Biologist
U.S. Fish and Wildlife Service
Carlsbad Fish and Wildlife Office
2177 Salk Ave, Suite 250
Carlsbad, CA 92008
760-431-9440, ext. 215
760-431-5901 fax

From: Nelson, Marjorie
To: Shoemaker, Justin
Subject: Re: Revised wolverine timeline
Date: Tuesday, February 13, 2018 4:55:36 PM

We did! We said we'd change pack tack later.

Marjorie Nelson
Chief, Division of Ecological Services
Mountain-Prairie Region
U.S. Fish and Wildlife Service
303-236-4258 direct
720-582-3524 cell

On Tue, Feb 13, 2018 at 2:53 PM, Shoemaker, Justin <justin_shoemaker@fws.gov> wrote:

> Marj,
>
> Caitlin tells me Q is still expecting wolverine in June. I thought you talked to her about moving it back in the FY. ???
>
> Justin Shoemaker
> Classification and Recovery Biologist
> U.S. Fish and Wildlife Service, Region 6
> Phone: 309-757-5800 x214
> Email: justin_shoemaker@fws.gov
>
> On Tue, Feb 13, 2018 at 2:05 PM, Shoemaker, Justin <justin_shoemaker@fws.gov> wrote:
>
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> >
> > Justin Shoemaker
> > Classification and Recovery Biologist
> > U.S. Fish and Wildlife Service, Region 6
> > Phone: 309-757-5800 x214
> > Email: justin_shoemaker@fws.gov

From: Grizzle, Betty
To: John Guinotte
Subject: Fwd: Draft PowerPoint
Date: Tuesday, February 13, 2018 10:22:44 AM
Attachments: Briefing for Deputy Director_Wolverine 12mo finding.pptx

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---------- Forwarded message ----------
From: **Grizzle, Betty** <betty_grizzle@fws.gov>
Date: Fri, Feb 9, 2018 at 1:32 PM
Subject: Draft PowerPoint
To: "Shoemaker, Justin" <justin_shoemaker@fws.gov>

Justin - Here is my first draft for the Deputy Director's briefing. It is 10 slides, if I don't add the DPS topic, but 13 if I do include this.

Feel free to make any changes or if you have other comments/concerns.

--

Betty J. Grizzle, D.Env., MPH
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North American Wolverine (*Gulo gulo luscus*): 12-Month Finding



Photo credit: Mark Packila; used with permission.

Mountain Prairie Regional Office

February 20, 2018

Listing History

1995, 2003: 90-day Not Substantial Findings

- 2003 finding challenged; 12-mo finding ordered by Court
- **2008: Not Warranted Finding (no DPS in contiguous U.S.)**
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- **DPS warranted**, but precluded by higher priority listing actions
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2013: Proposed Rule, with DPS (using 2010 analysis), 4(d)

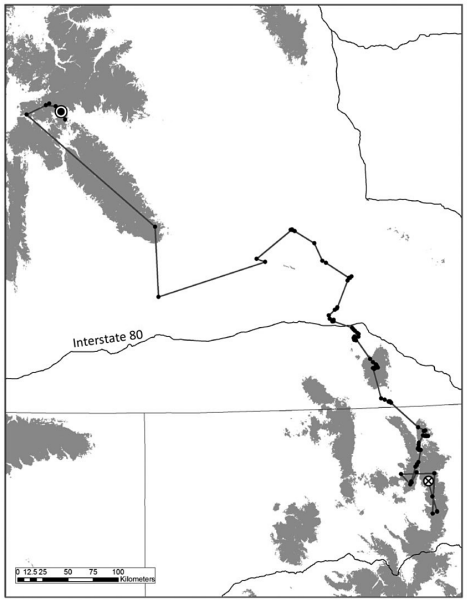
- Based primarily on synergistic effects of climate change, persistent snow
- **2014: Withdrawal of proposed rule**
- Challenged, Court remanded back to FWS

Current Action: Finalize Propose Rule or Withdrawal

- Informed by SSA Report
- Withdrawal or Final Rule to publish FY 2018

Species Description & Needs

- Key physical/ecological needs
  - Large territories, remote areas, high elevations
  - Access to variety of food, seasonal variability
  - Physical/Structural habitat features (e.g., talus slopes)
- Extensive movement and dispersal capabilities
  - Colorado, Utah, California, Michigan (all in 2000s)



Map of VHF locations of M56 - 2008 to 2009



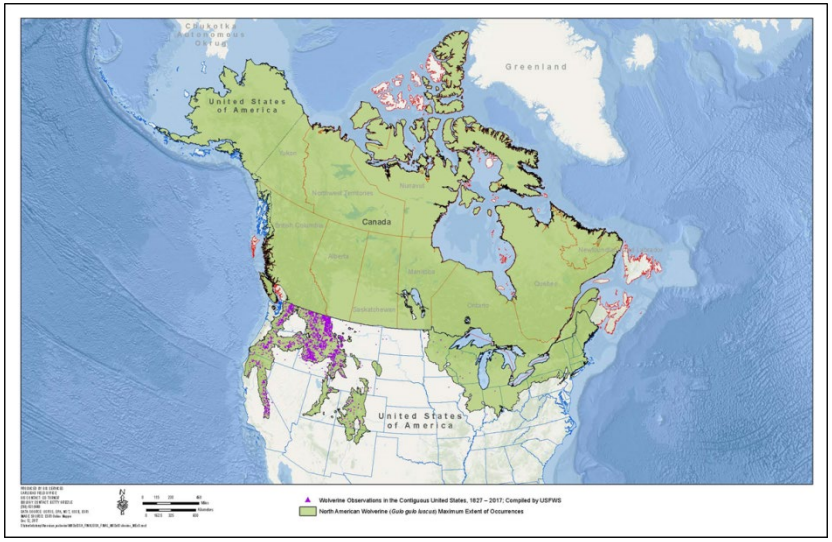
Suspected wolverine M56, from Packila et al. 2017

SSA Report –New information and analysis

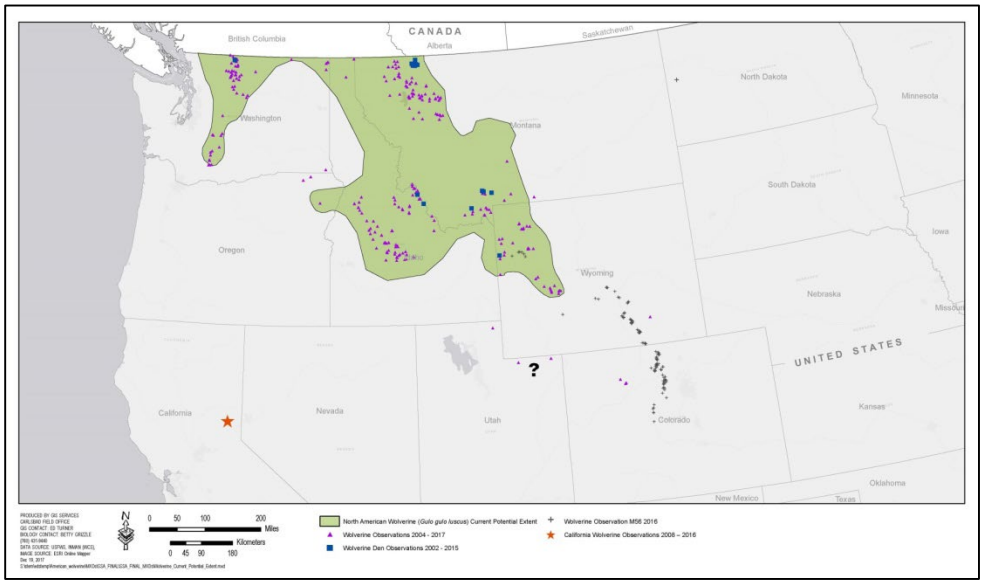
- Recent Scandinavian & Canadian studies providing new insights
- Physiological discussion added; not well-documented in previous rules
  - Important to describe critical temperature ranges and elements of thermoneutrality for evaluating effects of climate change
- New range maps prepared
- Contacted many wolverine experts (both in and outside U.S.), State agencies, and other scientists
- NOAA/CU snow study prepared (funded by Science Applications Program)

North American Wolverine Range Maps

- Maximum Extent of Occurrences (North America)



- Current Potential Extent of Occurrences (Lower 48)

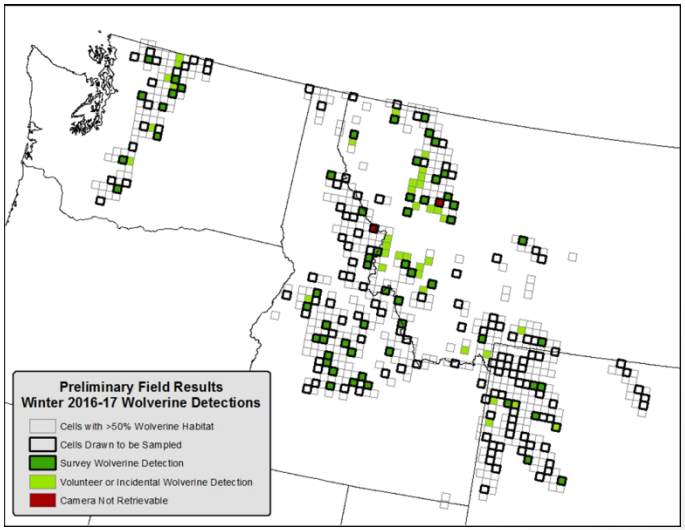


Denning and Snow: only one piece of the puzzle

- New studies – Copeland et al. model does not always predict either wolverine occurrences or den sites
  - At least 80 dens in Scandinavia and several in Alberta, Canada, recorded outside model
- Limited information on relationship of denning habitat with **reproductive success**
- Denning habitat varies across Holarctic range
- Young moved from natal den in late April
- Many factors influence den locations and shifts in den locations (not just temp/snow depth or persistence)
- Our analysis of "melt-out" dates for natal den sites in U.S. – earliest was May 14

Biological Status (current conditions)

- Population size
  - In No America: uncommon and at naturally low densities
  - Contiguous US: 2013 model estimate of 318 individuals (range 249-626)
    - Wolverine working group developing occupancy analysis
  - Canada – over 10,000 adults
    - Stable or increasing in this northern part of No Am range
  - AK uses density estimates (5-10 per 386 mi^2); considered stable
- Movement between AK and Canada & lower 48 and Canada
- Wolverine Occupancy Study
  - 4 Western States
  - 42% probability of occupancy



Potential Stressors Evaluated

- Effects from roads
- Effects from wildland fire
- Disturbance – winter recreational activity, other infrastructure
- Disease and Predation
- Overutilization
  - Research activities
  - Harvesting and incidental trapping
- For Future Conditions
  - Effects of Climate Change
  - Cumulative Effects

Overall Assessment

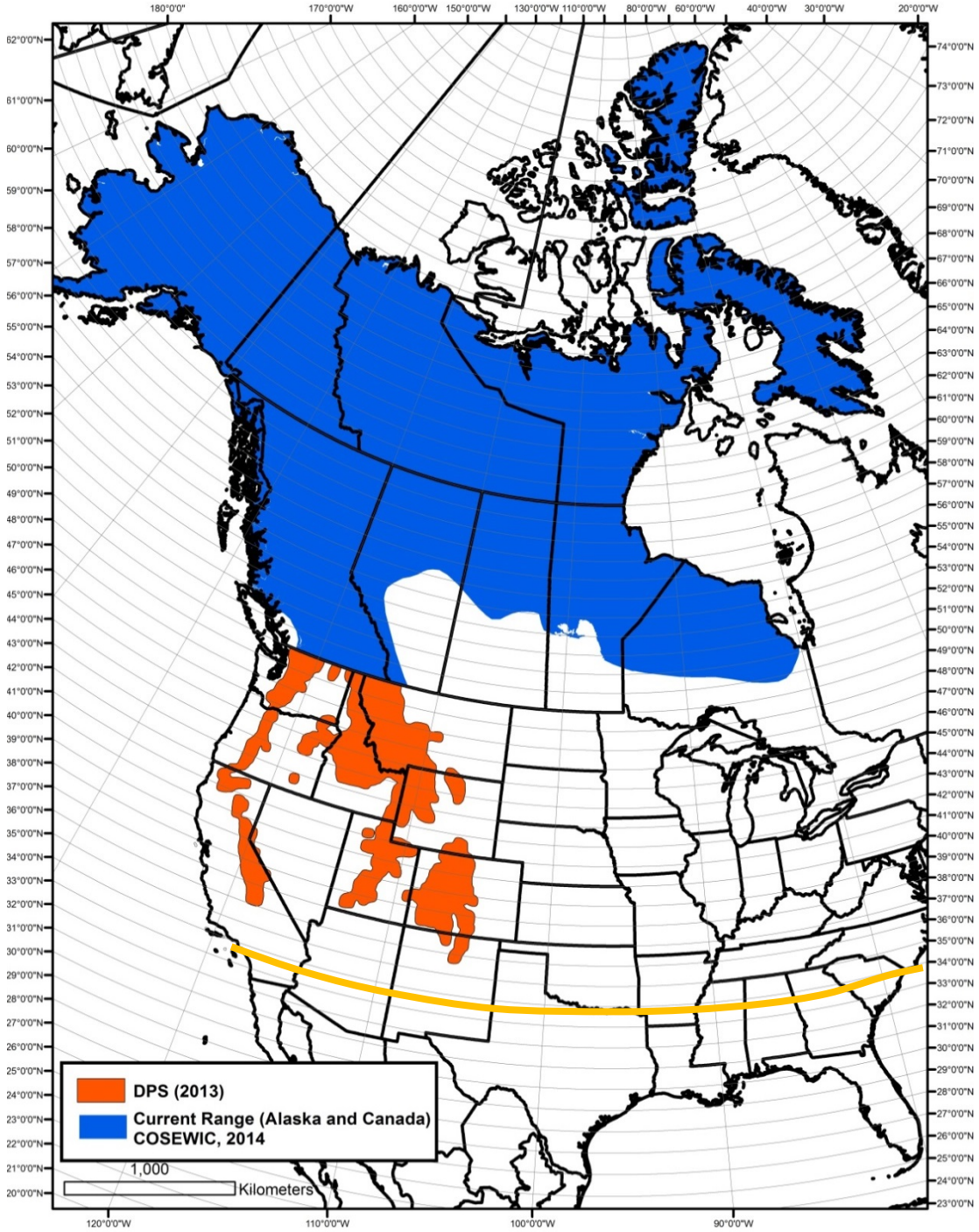
- Known or likely potential stressors represent low risk to the species
- Legal trapping currently prohibited in contiguous U.S.
  - Some incidental trapping still occurring
  - Trapping in Canada & Alaska appears to be sustainable
- Legal protections (T, E, Candidate) in several western states and Canada
  - Conservation plans (States and Tribes) in effect or planned
- No population trend information for contiguous U.S.
  - Wolverines now found in areas previously extirpated; still recovering from unregulated persecution
- Climate change effects: increase temperature and reduced snowpack
  - Model results: higher elevations (where wolverines are found) retain more snow than lower elevations
  - Modeled study areas: large portions have snow > 20” on May 1

Comments Received

- Response to Oct 2016 NOI (proposed rule/status review)
  - Over 200 comments posted (from several States, Canada, U.K.)
  - Most, but not all, general public comments favored listing
    - CC effects (need for deep snow) and rarity/small population size were primary concerns
- SSA Report Review
  - 11 States and Agency (Technical) Comments submitted
  - No Tribal comments submitted
  - Peer Reviewers
    - Complete, comprehensive, well-organized summary; additional references and technical comments provided
    - One reviewer was very negative – claimed “pervasive” bias in use and interpretation of info; CC projections should go beyond 50 years

DPS

- For 2010 (warranted but precluded); 2013 (proposed listing rule), 2014 (vacated withdrawal)
  - Discreteness = YES
    - 1. Markedly Separated = NO (No analysis provided)
    - 2. International Governmental Boundary per 4(a)(1)D = **YES**
      - Population Size: Canada = large populations; Contiguous US = population numbers low continued existence at risk
      - Effective Population Size: Canada = in good shape with sufficient genetic diversity; Contiguous US = small with poor genetic diversity
      - Habitat Connectivity: Canada = well-distributed/interconnected; Contiguous US = fragmented habitat
      - Exploitation: Did not rely on control of exploitation to establish discreteness
  - Significance = YES (see next slide)
    - Criteria #2 (Evidence that loss of the discrete population segment would result in a significant gap in the range of the taxon) = **YES**
      - *Loss would substantially reduce southern range terminus approximately 15 degrees of latitude (40 percent of the latitudinal extent of the wolverine range); and loss of 205,942 km^2 (79,515 mi^2) of habitat (**based on Copeland et al. 2010 model**)*
    - Criteria #1 (persistence of the discrete population segment in an ecological setting unusual or unique for the taxon) = NO
    - Criteria #3 (only remaining occurrence of a taxon) = NO
    - Criteria #4 (…discrete population segment differs markedly from other populations of the species in its genetic characteristics) = NO

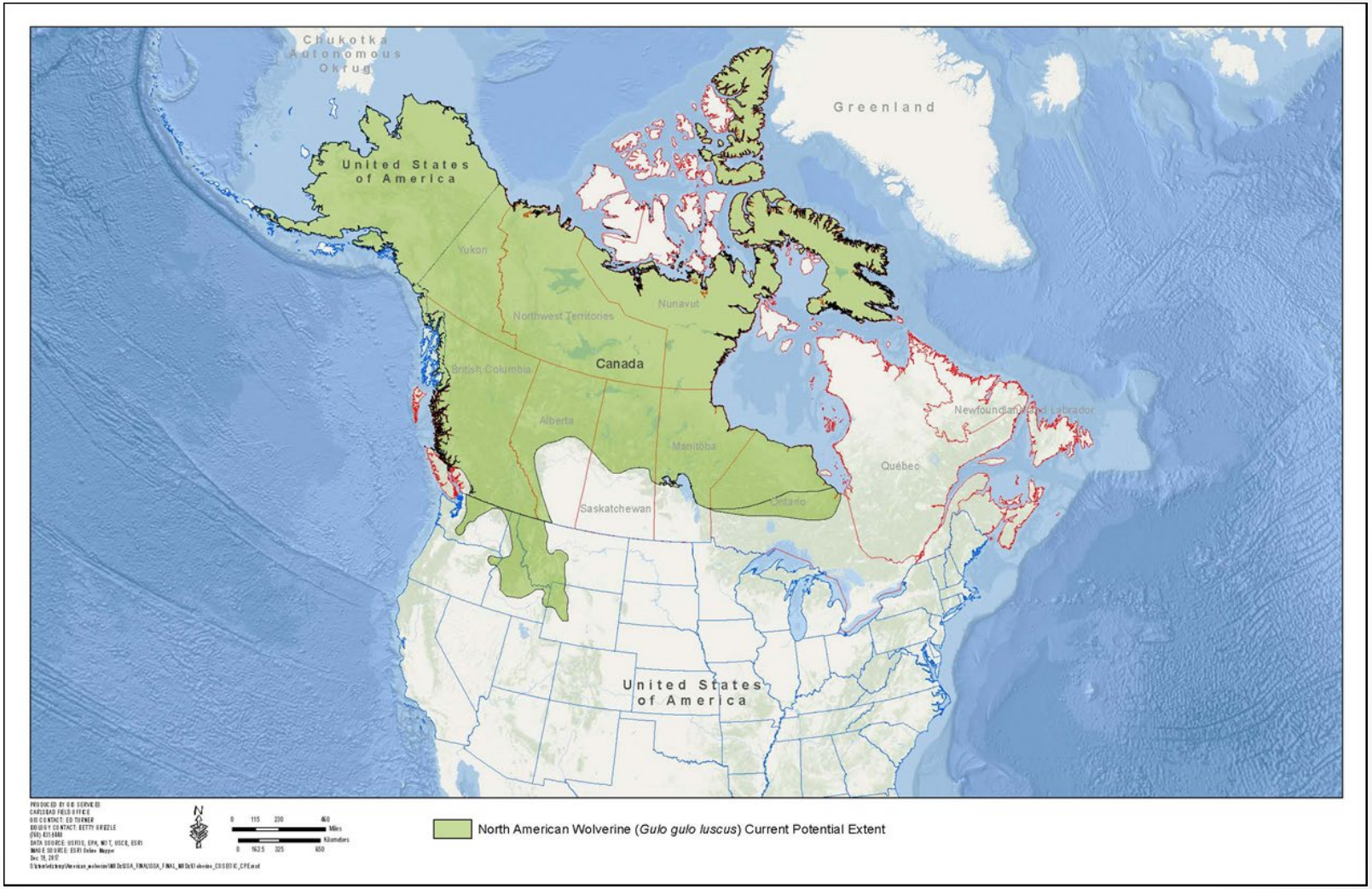


12-Month Finding on a Petition To List the North American Wolverine as Endangered or Threatened (2010)

" Today, the contiguous United States represents the southernmost reach of the wolverine's range. ***The loss of this population would be significant because it would substantially curtail the range of the wolverine by moving the southern range terminus approximately 15° of latitude to the north (or approximately 40 percent of the latitudinal extent of wolverine range)*** *and eliminate wolverines from the fauna of the contiguous United States.* ***Therefore, the loss of this population would result in a significant gap in the range of the taxon****."*

*"****Based upon the 15 degree latitude gap that would result in the range of the wolverine if the U.S. population was lost, we determine that the loss of the contiguous U.S. wolverine population would result in a significant gap in the range of the taxon.*** *Thus, the DPS meets the definition of significant in our DPS policy."*

Current Potential Extent – North America



From: Shoemaker, Justin
To: Grizzle, Betty
Subject: Re: Draft PowerPoint
Date: Tuesday, February 13, 2018 12:27:58 PM
Attachments: Briefing for Deputy Director_Wolverine 12mo finding_JS.pptx

---

Betty,

Looks pretty good. A few comments in the attached. Some of it we can sort out on tomorrow's call.

A little wordy on some slides. And I've been told Greg likes pictures/visuals...

Justin Shoemaker
Classification and Recovery Biologist
U.S. Fish and Wildlife Service, Region 6
Phone: 309-757-5800 x214
Email: justin_shoemaker@fws.gov

On Fri, Feb 9, 2018 at 3:32 PM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:

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>
> --
>
> Betty J. Grizzle, D.Env., MPH
> Fish and Wildlife Biologist
> U.S. Fish and Wildlife Service
> Carlsbad Fish and Wildlife Office
> 2177 Salk Ave, Suite 250
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> 760-431-9440, ext. 215
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North American Wolverine (*Gulo gulo luscus*): 12-Month Finding



Photo credit: Mark Packila; used with permission.

Mountain Prairie Regional Office

February 20, 2018

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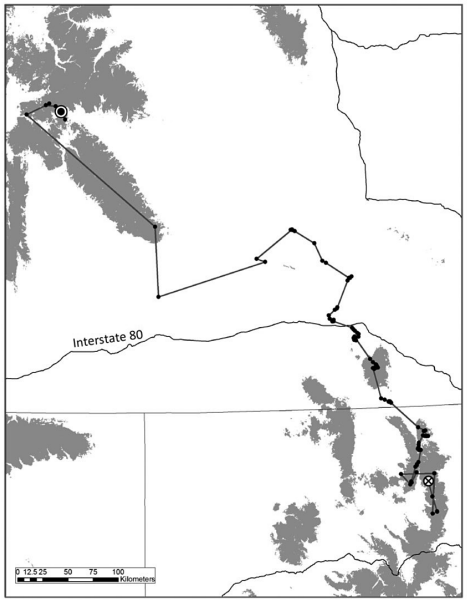
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  - Colorado, Utah, California, Michigan (all in 2000s)



Map of VHF locations of M56 - 2008 to 2009



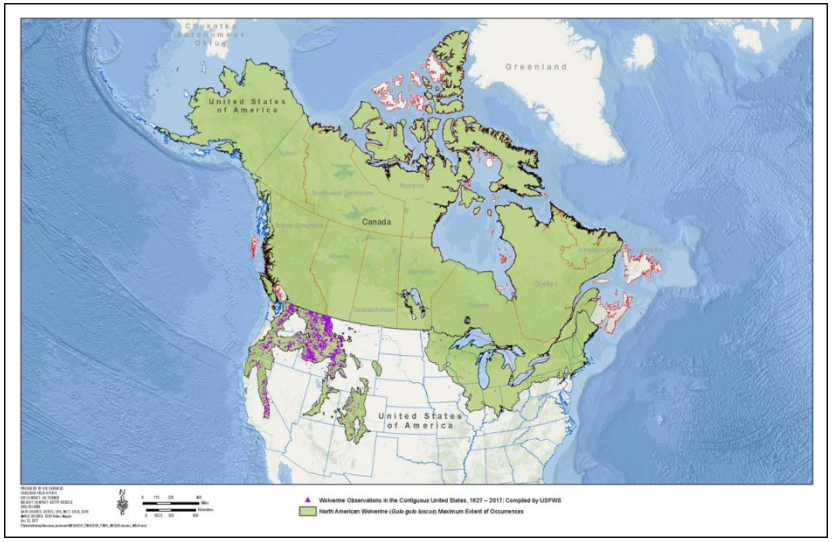
Suspected wolverine M56, from Packila et al. 2017

SSA Report –New information and analysis

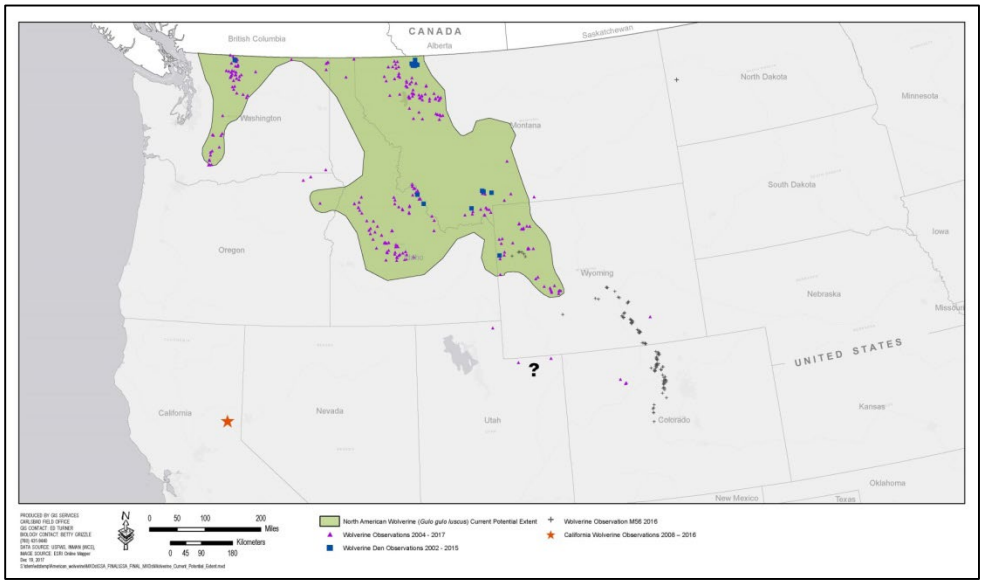
- Recent Scandinavian & Canadian studies providing new insights
- Physiological discussion added; not well-documented in previous rules
  - Important to describe critical temperature ranges and elements of thermoneutrality for evaluating effects of climate change
- New range maps prepared
- Contacted many wolverine experts (both in and outside U.S.), State agencies, and other scientists
- NOAA/CU snow study prepared (funded by Science Applications Program)

North American Wolverine Range Maps

- Maximum Extent of Occurrences (North America)



- Current Potential Extent of Occurrences (Lower 48)

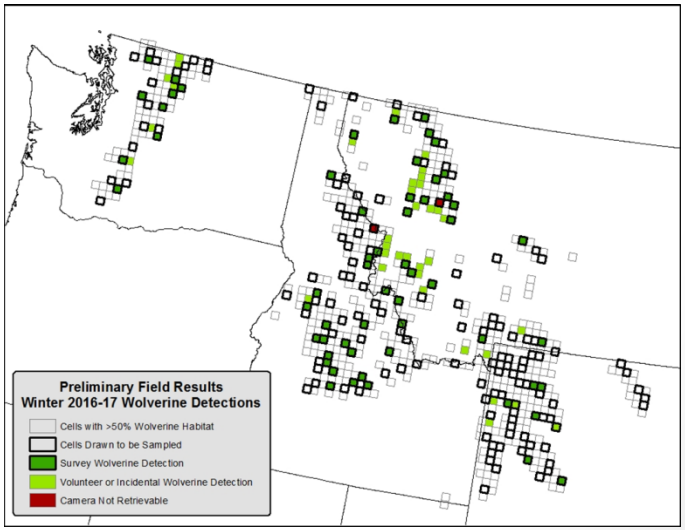


Denning and Snow: only one piece of the puzzle

- New studies – Copeland et al. model does not always predict either wolverine occurrences or den sites
  - At least 80 dens in Scandinavia and several in Alberta, Canada, recorded outside model
- Limited information on relationship of denning habitat with **reproductive success**
- Denning habitat varies across Holarctic range
- Young moved from natal den in late April
- Many factors influence den locations and shifts in den locations (not just temp/snow depth or persistence)
- Our analysis of "melt-out" dates for natal den sites in U.S. – earliest was May 14

Biological Status (current conditions)

- Population size
  - In No America: uncommon and at naturally low densities
  - Contiguous US: 2013 model estimate of 318 individuals (range 249-626)
    - Wolverine working group developing occupancy analysis
  - Canada – over 10,000 adults
    - Stable or increasing in this northern part of No Am range
  - AK uses density estimates (5-10 per 386 mi^2); considered stable
- Movement between AK and Canada & lower 48 and Canada
- Wolverine Occupancy Study
  - 4 Western States
  - 42% probability of occupancy



Potential Stressors Evaluated

- Effects from roads
- Effects from wildland fire
- Disturbance – winter recreational activity, other infrastructure
- Disease and Predation
- Overutilization
  - Research activities
  - Harvesting and incidental trapping
- For Future Conditions
  - Effects of Climate Change
  - Cumulative Effects

Overall Assessment

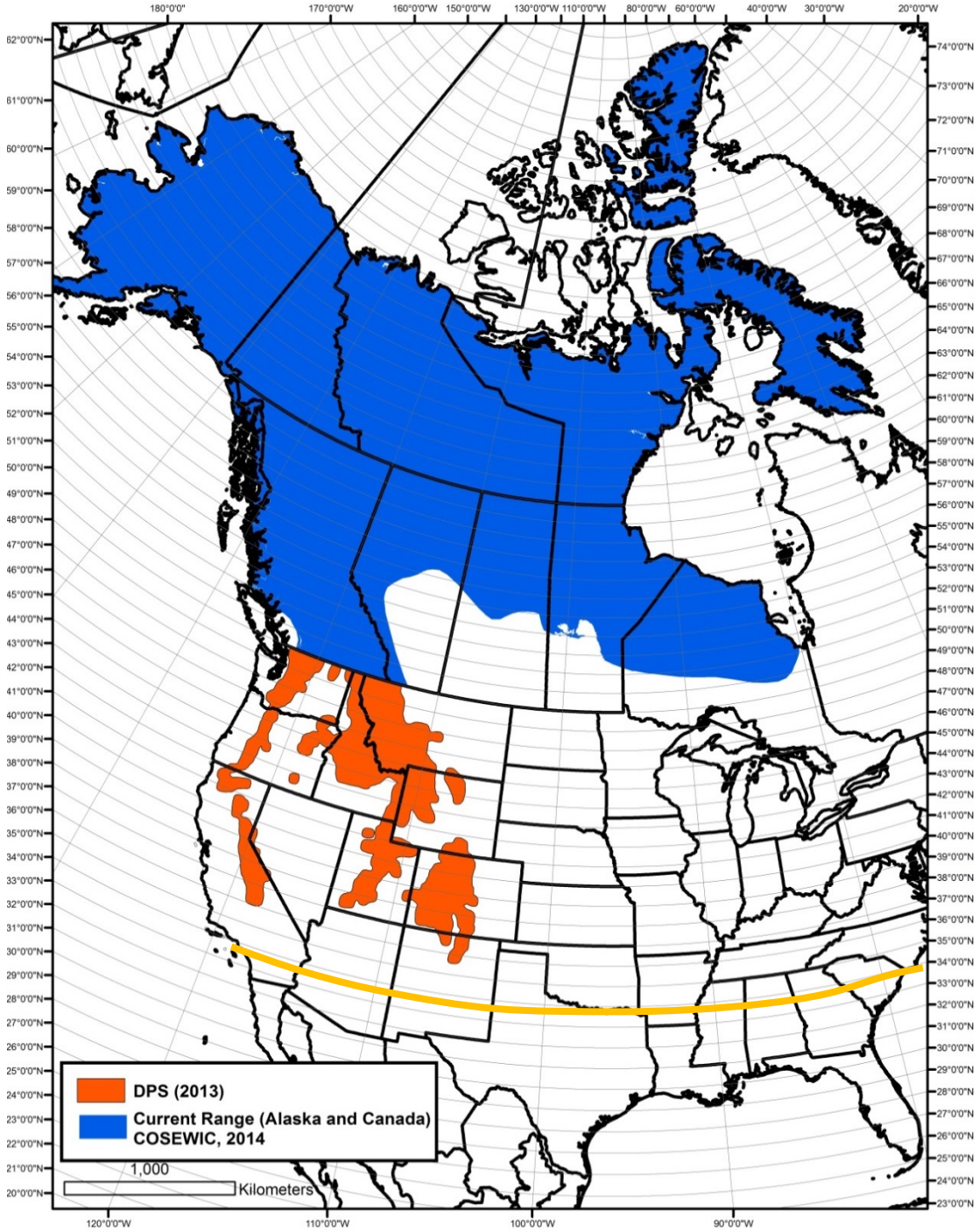
- Known or likely potential stressors represent low risk to the species
- Legal trapping currently prohibited in contiguous U.S.
  - Some incidental trapping still occurring
  - Trapping in Canada & Alaska appears to be sustainable
- Legal protections (T, E, Candidate) in several western states and Canada
  - Conservation plans (States and Tribes) in effect or planned
- No population trend information for contiguous U.S.
  - Wolverines now found in areas previously extirpated; still recovering from unregulated persecution
- Climate change effects: increase temperature and reduced snowpack
  - Model results: higher elevations (where wolverines are found) retain more snow than lower elevations
  - Modeled study areas: large portions have snow > 20” on May 1

Comments Received

- Response to Oct 2016 NOI (proposed rule/status review)
  - Over 200 comments posted (from several States, Canada, U.K.)
  - Most, but not all, general public comments favored listing
    - CC effects (need for deep snow) and rarity/small population size were primary concerns
- SSA Report Review
  - 11 States and Agency (Technical) Comments submitted
  - No Tribal comments submitted
  - Peer Reviewers
    - Complete, comprehensive, well-organized summary; additional references and technical comments provided
    - One reviewer was very negative – claimed “pervasive” bias in use and interpretation of info; CC projections should go beyond 50 years

DPS

- For 2010 (warranted but precluded); 2013 (proposed listing rule), 2014 (vacated withdrawal)
  - Discreteness = YES
    - 1. Markedly Separated = NO (No analysis provided)
    - 2. International Governmental Boundary per 4(a)(1)D = **YES**
      - Population Size: Canada = large populations; Contiguous US = population numbers low continued existence at risk
      - Effective Population Size: Canada = in good shape with sufficient genetic diversity; Contiguous US = small with poor genetic diversity
      - Habitat Connectivity: Canada = well-distributed/interconnected; Contiguous US = fragmented habitat
      - Exploitation: Did not rely on control of exploitation to establish discreteness
  - Significance = YES (see next slide)
    - Criteria #2 (Evidence that loss of the discrete population segment would result in a significant gap in the range of the taxon) = **YES**
      - *Loss would substantially reduce southern range terminus approximately 15 degrees of latitude (40 percent of the latitudinal extent of the wolverine range); and loss of 205,942 km^2 (79,515 mi^2) of habitat (**based on Copeland et al. 2010 model**)*
    - Criteria #1 (persistence of the discrete population segment in an ecological setting unusual or unique for the taxon) = NO
    - Criteria #3 (only remaining occurrence of a taxon) = NO
    - Criteria #4 (…discrete population segment differs markedly from other populations of the species in its genetic characteristics) = NO

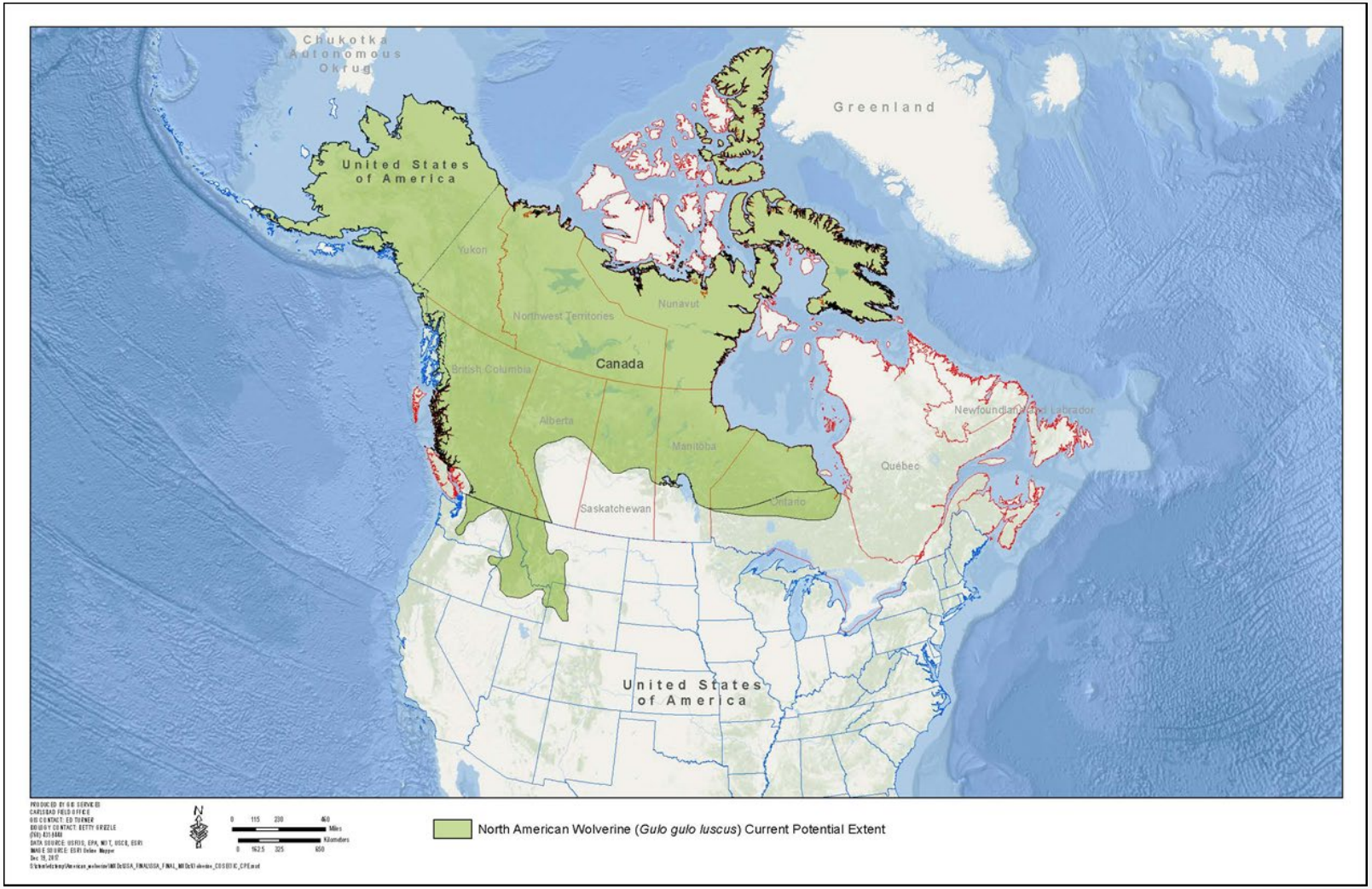


12-Month Finding on a Petition To List the North American Wolverine as Endangered or Threatened (2010)

" Today, the contiguous United States represents the southernmost reach of the wolverine's range. ***The loss of this population would be significant because it would substantially curtail the range of the wolverine by moving the southern range terminus approximately 15° of latitude to the north (or approximately 40 percent of the latitudinal extent of wolverine range)*** *and eliminate wolverines from the fauna of the contiguous United States.* ***Therefore, the loss of this population would result in a significant gap in the range of the taxon****."*

*"****Based upon the 15 degree latitude gap that would result in the range of the wolverine if the U.S. population was lost, we determine that the loss of the contiguous U.S. wolverine population would result in a significant gap in the range of the taxon.*** *Thus, the DPS meets the definition of significant in our DPS policy."*

Current Potential Extent – North America



From: Guinotte, John
To: Betty Grizzle
Subject: Fwd:
Date: Tuesday, February 13, 2018 12:52:12 PM
Attachments: NCA4figure8_2.png
end_of_century_snow_.docx

John Guinotte
Spatial Ecologist
Ecological Services
U.S. Fish and Wildlife Service
Mountain Prairie Region 6
134 Union Blvd., Lakewood, CO 80228
303-236-4264
john_guinotte@fws.gov

---------- Forwarded message ----------
From: **Andrea Ray - NOAA Federal** <andrea.ray@noaa.gov>
Date: Tue, Feb 13, 2018 at 10:52 AM
Subject: Re:
To: "Guinotte, John" <john_guinotte@fws.gov>
Cc: Steve Torbit <Stephen_Torbit@fws.gov>

Take a look at this version...give me a ring if you've got questions, I'm here most of the day, and tomorrow.

if you want to use a figure, I'd suggest the version that the NCA4 CCSA re-plotted (and publically available), I've attached a high-rez image, and here's a suggested caption:

Figure X. Projected changes in winter (DJF) snow water equivalent at the middle and end of this century under the higher scenario (RCP8.5) from a high-resolution version of the Community Atmospheric Model, CAM5. (Figure 8.2 from NCA 4, Wehner et al, 2017, based on Rhoades et al. 2018)

On Tue, Feb 13, 2018 at 8:56 AM, Andrea Ray - NOAA Federal <andrea.ray@noaa.gov> wrote:

> I'll look -- I just got clarifications on the numbers for the GCMs from Imtiaz, so I'm about to get back to this - i have to edit anyway, b/c those numbers are for the change from 1980-2005. freezing line in the cascades went up 1000m (750 to 1750) Wherease the Rockies only went up from 500m to 1000m -- but that's current time. the point for giving freezing line is that, according to their analysis, freezing line stays below the elevations you've told us are the denning habitat areas (see fig 10L)
>
> On one hand, I hesitate to get into comparing to the changes in coastal ranges (like the cascades & sierras) but so many people know about that work, and so that's the argument for doing it.

It's a dense paper.

I'll get the edited text back to you within the hour.

On Tue, Feb 13, 2018 at 7:26 AM, Guinotte, John <john_guinotte@fws.gov> wrote:

> Hi Andrea,
> I'm having trouble following what is below. The last two sentences of the first paragraph seem to contradict each other??
> Best, John
>
> On Mon, Feb 12, 2018 at 4:13 PM, Andrea Ray - NOAA Federal <andrea.ray@noaa.gov> wrote:
>
> > John -- I'm double checking a couple of numbers, hope to hear back from
> >
> > here it is tentatively, let me know if this is off base -- I'll send later as a final word doc and with the figures as separate high res.
> >
> > Although analysis of snowpack at the end of century was beyond the scope of the analysis done for FWS, a recent study analyzes snow to 2100 for several western mountain ranges (Rhoades et al, 2018), which was also used in the National Climate Assessment (Figure 8.2 and Wehner et al 2017). They simulate the changes in several snow-related variables using the NCAR CESM model and RCP8.5 (the higher emissions pathway), which is then run through the VIC hydrology model. Compared to the GCMs analyzed for FWS, the NCAR CESM model these Rhoades et al (2018) used is hot and very wet for GLAC by end of century (JOHN – I'm double checking these numbers +5.0 C and +21.4% precipitation), with a somewhat higher temperature and precipitation increase than the spread of the models we used for 2041-2070. For GLAC, CESM is hot and wet (+5.1 C and +11.1% precipitation), although within the spread of the models we used for 2041-2070.
> >
> > Rhodes et al (2018) find resiliency in snowpack for interior mountain ranges for the 2075-2010 period compared to the 1980-2005 climatology they use. For this period, at higher elevations (2000 to >2500m), they find snowfall changes in the Rockies to be to be nominally positive or negative (their figure 10, panels e, j, o). A high snow-rain ratio is also maintained in the higher elevation Rockies (Fig. 8 panels, c and e). They also find that the freezing levels for interior mountain ranges (Rockies, Wasatch and Uintas) does not move as far upslope as in the coastal mountain ranges they also analyze. In the Rockies, the freezing range moves up from 500 m to 1000m.
> >
> > Rhoades, A. M., P. A. Ullrich, and C. M. Zarzycki, 2017: Projecting 21st century snowpack trends in western USA mountains using variable-resolution CESM. *Climate Dynamics*, **44**, 1-28, doi:10.1007/s00382-017-3606-0
> > https://link.springer.com/article/10.1007%2Fs00382-017-3606-0

Wehner, M.F., J.R. Arnold, T. Knutson, K.E. Kunkel, and A.N. LeGrande, 2017: Droughts, floods, and wildfires. In: *Climate Science Special Report: Fourth National Climate Assessment, Volume I* [Wuebbles, D.J., D.W. Fahey, K.A. Hibbard, D.J. Dokken, B.C. Stewart, and T.K. Maycock (eds.)]. U.S. Global Change Research Program, Washington, DC, USA, pp. 231-256, doi: 10.7930/J0CJ8BNN.

On Fri, Feb 9, 2018 at 1:28 PM, Guinotte, John <john_guinotte@fws.gov> wrote:

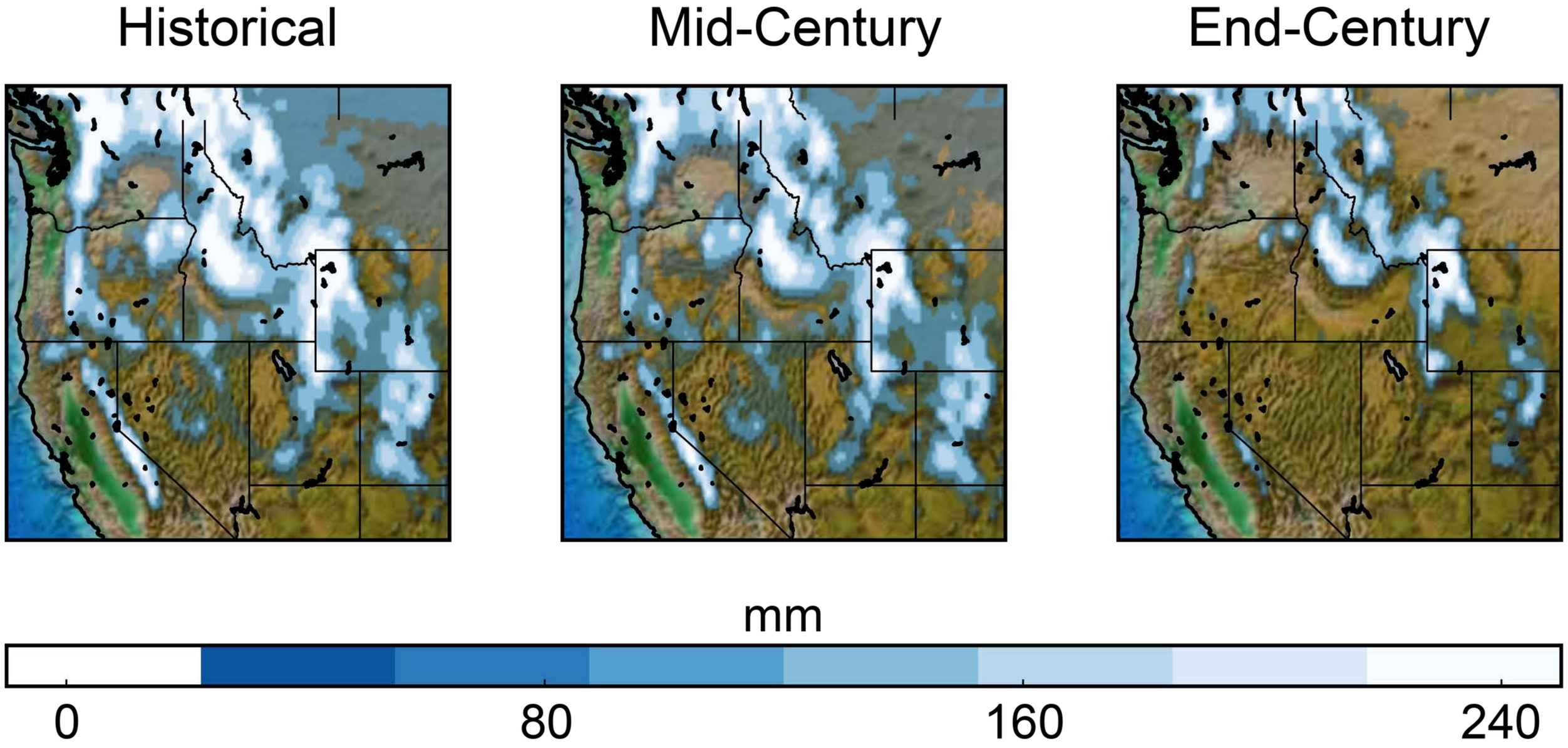
> Hi Andrea, Are you going to send me the paragraph we spoke about? We are getting close to deadline. I made a few edits to your abstract and Steve said he'd look when he gets back to Denver this weekend. Thanks John
>
> John Guinotte
> Spatial Ecologist
> Ecological Services
> U.S. Fish and Wildlife Service
> Mountain Prairie Region 6
> 134 Union Blvd., Lakewood, CO 80228
> 303-236-4264
> john_guinotte@fws.gov

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(fax) 303-497-6449
andrea.ray@noaa.gov
www.researchgate.net/profile/Andrea_Ray2


Historical
Mid-Century
End-Century
mm
0
80
160
240

Although analysis of snowpack at the end of century was beyond the scope of the work done for FWS, a recent study analyzes the impacts of climate change on mountain snowpack to 2100 for several western mountain ranges (Rhoades et al, 2018), which was also used in the National Climate Assessment (Figure 8.2 and Wehner et al 2017). They simulate the changes in several snow-related variables using the NCAR CESM model and RCP8.5 (the higher emissions pathway also used in the NOAA study), which is then run through the VIC hydrology model. Compared to the GCMs analyzed by NOAA, the NCAR CESM model used by Rhoades et al (2018) is hot and very wet for GLAC by end of century (+5.0 C and +21.4% precipitation). This temperature and precipitation increase somewhat higher than the spread of the models in 2041-2070 used in the NOAA study. For GLAC, CESM is hot and wet (+5.1 C and +11.1% precipitation), although within the spread of the models in 2041-2070.

Rhodes et al (2018) analyze simulated snowpack variables for the 2075-2100 end-of-century period compared to a 1980-2005 climatology. For the whole Rockies area, which includes areas well below the current denning zone, they find a slight but not significant decrease in mean December-January-February (DJF) snowfall (-2.29%), with a slight increase in the median (1.4%) (see table 4). Snow water equivalent decreases by about 65% in the mean. However, mean snow cover area decreases by less than the SWE (as found in the NOAA study), by 38.1%.

They provide analysis by elevation, but it is not broken out by mountain range. Across the west, for the end-of-century, snowcover above 2000m has a mean zero area change, with a range of -24% to +15.3% among the ensemble members (their Table 5).

Overall, their findings show that snow can be maintained in the higher elevations of the Rocky Mountains and other interior mountain ranges for the 2075-2010 period. Two other factors that help persistence of snow are that the snow-rain ratio does not decrease as much as in other areas (their figure 8e, j) and an increase in the liquid to snow ratio (Fig. 8 c, e). They also find that the freezing levels for interior mountain ranges (Rockies, Wasatch and Uintas) does not move as far upslope as in the coastal mountain ranges they also analyze, and does not go above freezing until end of century and only below 1000m (Figure 10L).

Rhoades, A. M., P. A. Ullrich, and C. M. Zarzycki, 2017: Projecting 21st century snowpack trends in western USA mountains using variable-resolution CESM. *Climate Dynamics*, **44**, 1-28, doi:10.1007/s00382-017-3606-0
https://link.springer.com/article/10.1007%2Fs00382-017-3606-0

Wehner, M.F., J.R. Arnold, T. Knutson, K.E. Kunkel, and A.N. LeGrande, 2017: Droughts, floods, and wildfires. In: *Climate Science Special Report: Fourth National Climate Assessment, Volume I* [Wuebbles, D.J., D.W. Fahey, K.A. Hibbard, D.J. Dokken, B.C. Stewart, and T.K. Maycock (eds.)]. U.S. Global Change Research Program, Washington, DC, USA, pp. 231-256, doi: 10.7930/J0CJ8BNN.

Suggested figure sent separately

Figure caption. Projected changes in winter (DJF) snow water equivalent at the middle and end of this century under the higher scenario (RCP8.5) from a high-resolution version of the Community Atmospheric Model, CAM5. (Figure 8.2 from NCA 4, Wehner et al, 2017, based on Rhoades et al. 2018)

From: Shoemaker, Justin
To: Grizzle, Betty
Subject: Re: Revised SSA Report
Date: Tuesday, February 13, 2018 3:48:44 PM
Attachments: 20180207 Wolverine SSA Report Version 1.2_JS.docx

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A few edits in the attached.

Justin Shoemaker
Classification and Recovery Biologist
U.S. Fish and Wildlife Service, Region 6
Phone: 309-757-5800 x214
Email: justin_shoemaker@fws.gov

On Wed, Feb 7, 2018 at 3:17 PM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:

> Please find attached for your review an edited (in track changes) Wolverine SSA Report incorporating: 1) new technical material as recommended by the Recommendation Management Team (e.g., snow discussion, Alberta map and new info on dens), 2) new material from a reference recommended by peer reviewer 3 (effective pop size), and 3) new material on climate trends and climate projections (in Cascades and MT) recommended in comments provided by State/USFS. I was not able to incorporate 2 and 3 in the last version. I also did some editing for grammar (extra spaces, spelling, etc.) that may not show up in track changes. Note: There is one new figure (Alberta map), one new appendix, and several new references.
>
> I have not yet updated the figure numbers, table of contents, page numbers, and headers as those changes don't work well while in track changes mode.
>
> Assuming we receive genetic data in the next week or so, I can incorporate that as well.
>
> Please make any recommended edits/comments in track changes mode.
>
> Thanks!
>
> --
>
> Betty J. Grizzle, D.Env., MPH
> Fish and Wildlife Biologist
> U.S. Fish and Wildlife Service
> Carlsbad Fish and Wildlife Office
> 2177 Salk Ave, Suite 250
> Carlsbad, CA 92008
> 760-431-9440, ext. 215
> 760-431-5901 fax

SPECIES STATUS ASSESSMENT
FOR THE
NORTH AMERICAN WOLVERINE
(*Gulo gulo luscus*)



Wolverines in southwestern Montana. *Photo credit: Mark Packila; used with permission.*

U.S. Fish and Wildlife Service

Version 1.~~1~~2
~~January 10~~February 7, 2018

Suggested citation:

U.S. Fish and Wildlife Service. 2018. Species status assessment report for the North American wolverine (*Gulo gulo luscus*). Version 1.2~~1~~. ~~January~~February 2018. U.S. Fish and Wildlife Service, Mountain-Prairie Region, Lakewood, CO.

Executive Summary

The North American wolverine (*Gulo gulo luscus*; wolverine) is a medium-sized carnivore found within the west-northwestern contiguous United States, Alaska, and Canada. Systematic monitoring across the wolverine's North American range has not been conducted given the difficulty in surveying this highly mobile species, and its occupancy within large and relatively inaccessible areas. A multi-state effort to determine wolverine occupancy in Montana, Idaho, and Washington was conducted in winter of 2016–2017 and in Wyoming for the winters of 2015–2016 and 2016–2017. Results from this study are still being analyzed, but photographic detections of wolverines were found across all surveyed States, including areas where wolverines have not recently been observed. In Canada, the population is estimated to exceed 10,000 mature individuals and has been stable over the last two decades. Recent density estimates indicate no declining trend for wolverines in Alaska. Wolverine populations in Alaska are considered to be continuous with populations in the Yukon and British Columbia provinces of Canada. Wolverines that occupy the North Cascades region are known to move from Washington into British Columbia.

Wolverines are capable of moving and dispersing over great distances over short periods of time. Wolverine populations are characterized by naturally low densities in North America. The species is highly territorial, with very little overlap between same-sex adults. Wolverines occupy a variety of habitats, but generally select habitat in locations away from human settlements. Wolverines consume a variety of food resources and seasonal switching of prey is commonly observed. As with other Arctic mammals, wolverines have the ability to adapt to both warm and cold ambient temperatures and solar radiation through both physiological and behavioral responses, such as vasodilation, increase in skin temperature, seasonal adjustments in fur insulation, and micro- and macro-habitat selection.

Wolverine reproduction includes the following characteristics: polygamous behavior (i.e., male mates with more than one female each year), delayed implantation (up to 6 months), a short gestation period (30–40 days), denning behavior, and an extended period of maternal care. The reproductive behavior in wolverines is temporally adapted to take advantage of the availability of food resources, limited interspecific competition, and snow cover in the winter.

Since the publication of the Service's 2013 proposed rule to list the distinct population segment of the North American wolverine in the contiguous United States (78 FR 7864; February 4, 2013), several new wolverine studies have been published. These studies have improved our understanding of wolverine biology while also highlighting new insights into identifying key species' needs and their interactions with both abiotic and biotic factors. In particular, wolverine populations and wolverine dens have been observed outside previously modeled projections of spring snow cover. Our evaluation of snow cover at previously recorded natal den site locations in the western United States indicated that 'melt-out' dates at these locations extend well past the May 15 date used in persistent spring snow cover models.

Overall, the best available information indicates that within the contiguous United States the wolverine's physical and ecological needs include:

(1) large territories in relatively inaccessible landscapes; at high elevation (1,800 to 3,500 meters (5,906 to 11,483 feet));
(2) access to a variety of food resources, that varies with seasons; and
(3) physical/structural features (e.g., talus slopes, rugged terrain) linked to reproductive behavioral patterns.

In this Species Status Assessment (SSA) Report, we provide a discussion of the ecological needs of the wolverine, its current conditions, and projected future conditions. We evaluate potential stressors to the species, with a particular focus on the impacts associated with projected effects of climate change.

In our analysis, we applied the conservation biology principles of redundancy, resiliency, and representation (collectively known as the "3Rs") to evaluate the current and projected future condition of the wolverine and its ability to sustain itself (as one or more populations) in the wild over time (Carroll *et al.* 1996, entire; Wolf *et al.* 2015, entire). This evaluation considers the demographic, distribution, and diversity characteristics unique to the species. After applying the framework of the 3Rs, we determined the following:

(1) Redundancy: The wolverine occurs in the contiguous United States within a metapopulation structure. The best available information indicates that the species continues to expand into historical, previously occupied areas in the contiguous United States following decades of unregulated hunting or trapping and poisoning.
(2) Representation: The wolverine is currently found in the west-northwestern United States, as well as much of Canada, and Alaska. The best available information indicates that the species is found in Arctic, boreal, and subalpine habitats. We estimated a Current Potential Extent (of Occurrence)[1] for the wolverine in the contiguous United States of approximately 280,316 square kilometers (km^2) (69,267,592 acres (ac)), which represents approximately 3.45 percent of the species' total Current Potential extent in North America (see Figures 3 and 4 below).
(3) Resiliency: The wolverine appears resilient within its contiguous United States range. The species exhibits physiological (e.g., seasonal changes in fur) and behavioral plasticity in its life history (e.g., reproduction, feeding, movement and use of habitat). Estimated population size and growth rates across its North American range are uncertain, but the best available information does not suggest that abundance is declining in the contiguous United States, or in North America. The most significant stressor currently and in the future appears to be the effects of climate change, such as warming temperatures and loss of snowpack. However, we have no information to indicate that wolverines are unable to adapt or adjust to potential changing conditions as described in the analysis provided in this SSA Report.

Demographic risks to the species from either known or most likely potential stressors (i.e., effects from roads, disturbance due to winter recreational activities, effects of wildland fire, and

[1] Current Potential Extent of Occurrence represents represents the area defined by circumscribing all localities of a species where we know or expect it to still persist. It is generally created by starting with the full set of localities for the species, then removing the localities where the species is known or expected to have been extirpated (Service 2017).

overutilization) are low based on our evaluation of the best available information as it applies to current and potential future conditions for the wolverine and in the context of the attributes that affect its viability. We analyzed the potential effects of climate change to wolverine habitat, including snow persistence in the Northern and Southern Rocky Mountains. The future timeframe evaluated in this analysis is approximately 38~~40~~ to 50 years. This range represents our best professional judgment of the projected future conditions related to climate change for the western United States, wildland fire conditions, or other potential cumulative impacts. While population information is lacking for this subspecies in some parts of its range, the best available information does not indicate that, winter recreational activities, infrastructure features, mortality from road crossings or trapping (authorized and incidental), currently or in the future will result in a decline in the subspecies across its range. Our evaluation of climate change indicates that snow cover is projected to decline in response to warming temperatures and changing precipitation patterns, but this varies by elevation, topography, and by geographic region. In general, models indicate higher elevations will retain more snow cover than lower elevations, particularly in early spring (April 30/May1). Further, significant snow persistence (greater than 0.5 meters (20 inches)) is projected at high elevations.

Legal protections in the contiguous United States include State listing in California and Oregon (threatened), endangered in Colorado, a candidate species in Washington, non-game species protections in Idaho and Wyoming, and protected from collection, importation, and possession in Utah. In Canada, provincial designations range from endangered to threatened in eastern provinces, and sensitive/special concern to no ranking in other provinces (definitions provided by the Committee on the Status of Endangered Wildlife in Canada, 2014). Legal trapping or hunting of wolverines is currently prohibited in the contiguous United States. Trapping effort along the U.S.–Canada border does not represent a barrier to wolverine movement and dispersal along the international border.

Within the contiguous United States, approximately 72 percent of the Current Potential Extent of the wolverine is located on Federal lands, with 18 percent of this area located in designated wilderness areas. Management actions, including State Wildlife Action Plans, the Idaho Wolverine Conservation Plan, and USDA Forest Service Land and Resource Management Plans, and other Federal and Tribal partners, include winter road closures, fire management, land acquisition or conservation easements. The Western Association of Fish and Wildlife Agencies (WAFWA), in coordination with Tribal partners, have formed a multi-state, multi-agency working group (Western States Wolverine Working Group) to design and implement the Western States Wolverine Conservation Project (WSWCP)–Coordinated Occupancy Survey. In addition, a WAFWA Wildlife Chiefs Wolverine Subcommittee is working collaboratively with each other and with the Service, Tribes, and other partners to conserve wolverines in the western United States. To date, approximately $1.5 million of that funding has been applied towards conservation and management actions for the wolverine.

This SSA Report was submitted for review by State and Federal agencies, Tribal nations, and four independent peer reviewers.

Abbreviations and Acronyms Used

ADF&G = Alaska Department of Fish and Game
BLM = Bureau of Land Management
°C = degrees Celsius
CEC = Commission for Environmental Cooperation
CDFW = California Department of Fish and Wildlife
CMIP = Coupled Model Intercomparison Project
CNDDB = California Natural Diversity Database
COSEWIC = Committee on the Status of Endangered Wildlife in Canada
cm = centimeter
DNA = deoxyribonucleic acid
EIS = Environmental Impact Statement
EPA = U.S. Environmental Protection Agency
°F = degrees Fahrenheit
ft = feet
GCMs = Global Climate Models
GHG = Greenhouse gas
GLAC = Glacier National Park
GPS = Global Positioning System
IDFG = Idaho Department of Fish and Game
in = inch
IPCC = Intergovernmental Panel on Climate Change
IUCN = International Union for Conservation of Nature and Natural Resources
kg = kilogram
km = kilometer
lb = pound
m = meter
mi = mile
MODIS = Moderate-Resolution Imaging Spectroradiometer
Montana FWP = Montana Fish, Wildlife, & Parks
NDSI = Normalized Difference Snow Index
NRC = National Research Council
NRIS = Natural Resource Information System
ODFW = Oregon Department of Fish and Wildlife
RCPs = Representative Concentration Pathways
ROMO = Rocky Mountain National Park
Service = U.S. Fish and Wildlife Service
SSA = Species Status Assessment
SCA = Snowcovered Area
SGCN = Species of Greatest Conservation Need
SWCC = Southwestern Crown of the Continent
SWE = Snow Water Equivalent
UDNR = Utah Division of Natural Resources
WAFWA = Western Association of Fish and Wildlife Agencies
WDFW = Washington Department of Fish and Wildlife

WGFD = Wyoming Game and Fish Department
WRCC = Western Regional Climate Center
WSWCP = Western States Wolverine Conservation Project
YBP = Years Before Present

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Introduction

The wolverine (*Gulo gulo*) is the largest terrestrial member of the Mustelidae family (weasels, fisher, mink, marten, and others) and resembles a small bear with a bushy tail (Hash 1987, p. 575). Wolverines have a Holarctic distribution that includes the northern portions of Europe, Asia, and North America. In North America, they are found in Alaska, much of Canada, and the western-northwestern United States. The wolverine is important to the culture of Native Americans and Aboriginal Peoples in North America, as is its conservation status in aboriginal territory (Cardinal 2004, p. iv; Edmo 2016; pers. comm.; Miles 2017, pers. comm.).

Wolverines possess a number of morphological and physiological adaptations that allow them to travel long distances and they maintain large territories in remote areas (Pasitschniak-Arts and Larivière 1995, p. 6). They have been described as curious, intelligent, and playful, but cautious animals (e.g., Krott 1958, p. 241; Krott 1960, pp. 25–26; Magoun 1985, p. 94; Cardinal 2004, p. 7–8; Woodford 2014, entire), though their social behavior and social organization has not been well-studied.

During the late 1800s and early 1900s, the wolverine population declined or was extirpated in much of the conterminous United States (lower 48 States), which has been attributed to unregulated trapping and habitat degradation (Hash 1987, p. 583). Similar range reductions and extirpations of some wolverine populations were observed in parts of Canada during this time period (van Zyll de Jong 1975, entire; Committee on the Status of Endangered Wildlife in Canada (COSEWIC) 2014, p. iv), attributed largely to human exploitation and availability of food (e.g., decline in caribou (*Rangifer tarandus*)), not climate or habitat changes (van Zyll de Jong 1975, pp. 434, 436). Wolverine numbers have recovered to some extent from this decline (e.g., Aubry *et al*., 2007, p. 2,151; Aubry *et al*., 2012, entire; Aubry *et al*. 2016, pp. 14–15; Magoun *et al*. 2013, p. 27). In the United States, wolverines are currently found in parts of Washington, Oregon, Idaho, Montana, Wyoming, California (single male) and Alaska, and as recently as 2010 in Michigan, 2012 in Colorado, and 2016 in Utah. Known reproducing wolverine populations are found in Washington, Idaho, Montana, and Wyoming.

Species Status Assessment Methodology

In preparing the Species Status Assessment (SSA) Report for the wolverine, we reviewed available reports and peer-reviewed literature, incorporated survey information, and contacted species experts to collect additional unpublished information for the North American subspecies (*Gulo gulo luscus*), including Canada and Alaska. We identified uncertainties and data gaps in our assessment of the current and future status of the species. We also evaluated the appropriate analytical tools to address these gaps and conducted discussions with species experts and prepared updated maps of the known species' range and denning areas in North America. In some instances, we used publications and other reports (primarily from Fenno-Scandinavia) of the Eurasian subspecies (*Gulo gulo gulo*) in completing this assessment.

Importantly, we note here that, since the publication of the 2013 proposed listing rule (78 FR 7864; February 4, 2013), many new wolverine studies have been published, which has added to our understanding of wolverine biology while also highlighting new insights into identifying key

species' needs and their interactions with both abiotic and biotic factors. This is particularly relevant for a difficult to study animal like the wolverine.

Using the species, individual, and population needs identified for the wolverine and location results from surveys and studies, we conducted a geospatial analysis to estimate the Current Potential Extent (of Occurrence) for the North American wolverine in the contiguous United States. Current Potential Extent represents the Maximum Extent of Occurrences of a species minus those areas where we believe the species has been extirpated (see Figures 3 and 4 below). We then evaluated this area and previous estimates of potentially suitable habitat in the west-northwestern United States to assess the species' current conditions within that region. Our future condition analysis includes the potential conditions that the species or its habitat may face, that is, the most probable scenario if those conditions are realized in the future. This most probable scenario includes consideration of the sources that have the potential to most likely impact the species at the population or rangewide scales in the future, including potential cumulative impacts. Potential future impacts associated with climate change (probabilistic estimates for temperature and precipitation) were based on downscaled climate model projections, including a detailed study of two regions in the western United States— Glacier National Park (currently occupied by reproducing wolverines) and Rocky Mountain National Park (not known to be currently occupied).

For the purpose of this assessment, we generally define viability as "consisting of self-sustaining populations that are well distributed throughout the species' range," and where "[s]elf-sustaining populations are those that are sufficiently abundant and have sufficient genetic diversity to display the array of life history strategies and forms that will provide for their persistence and adaptability in the planning area over time" (Committee of Scientists 1999, p. 38). We use a timeframe of approximately 38~~40~~ to 50 years because it is within the range of the available modeling efforts related to climate change; beyond this range, modeling uncertainty increases substantially. We believe this is a reasonable timeframe to consider as it includes the potential for observing these effects over several generations of the wolverine.

Using the SSA framework (Figure 1), we consider what the species needs to maintain viability by characterizing the status of the species in terms of resiliency, redundancy, and representation (Wolf *et al*. 2015, entire).

- **Resiliency** is having sufficiently large populations for the species to withstand stochastic events (arising from random factors). We can measure resiliency based on metrics of population health; for example, birth versus death rates and population size. Resilient populations are better able to withstand disturbances such as random fluctuations in birth rates (demographic stochasticity), variations in rainfall (environmental stochasticity), or the effects of anthropogenic activities.

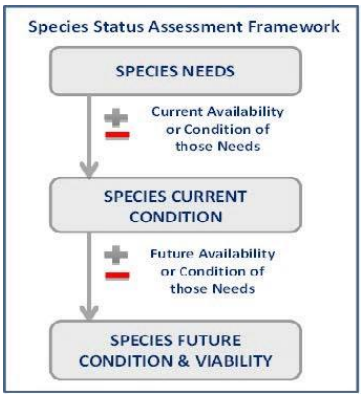


Figure 1. Species Status Assessment Framework.

- **Redundancy** is having a sufficient number of populations for the species to withstand catastrophic events (such as a rare destructive natural event or episode involving many populations). Redundancy is about spreading the risk and can be measured through the duplication and distribution of populations across the range of the species. The greater the number of populations a species has distributed over a larger landscape, the better it can withstand catastrophic events.
- **Representation** is having the breadth of genetic makeup of the species to adapt to changing environmental conditions. Representation can be measured through the genetic diversity within and among populations and the ecological diversity (also called environmental variation or diversity) of populations across the species' range. The more representation, or diversity, a species has, the more it is capable of adapting to changes (natural or human caused) in its environment. In the absence of species-specific genetic and ecological diversity information, we evaluate representation based on the extent and variability of habitat characteristics within the geographical range.

Species Description

Taxonomy

The taxonomic relationship between North American and Eurasian wolverines has been a debated topic (Pasitschniak-Arts and Larivière 1995, p. 1). Most authorities consider all wolverines to belong to a single species, *Gulo gulo* (Rausch 1953, p. 114; Kurten and Rausch 1959, p. 19; Wozencraft 2008 [in *Wilson and Reeder's Mammal Species of the World*, online publication]). Some also further consider the New World and Old World wolverines to be two subspecies, *Gulo gulo luscus* and *G. g. gulo,* respectively, based on morphological measurements. Degerbøl (1935, pp. 35–43) noted slight color differences and very slight, if any, cranium differences, based on 10 North American (Hudson Bay) specimens examined, and regarded the North American and Old World wolverines as conspecific, but identified two subspecies. This reference also cites Coues (1877, p. 43), who, based on observations of a slight similar cranium difference, had posited that the wolverines of the Old World and New World were the same species (Degerbøl 1935, p. 35).

In their *Checklist of Palaearctic and Indian Mammals* (1st and 2nd editions) Ellerman and Morrison-Scott (1951, p. 251; 1966, p. 251) identified one species of wolverine, but listed several subspecies. A comparative analysis of various measurements from 1 wolverine skull collected from the northern Ural Mountains to 41 Alaskan skulls by Rausch (1953, entire) reported "no appreciable differences," noting the highly variable skull characteristics for the Alaskan specimens. Additionally, Krott (1960, p. 20) found no distinct differences between Old World and New World wolverines, and stated that pelt size and quality were not distinguishable. However, using biometric measurements of both more recent (at that time) collections and previously published cranial measurements (e.g., Degerbøl 1935; Rausch 1953), Kurtén and Rausch (1959, p. 19) reported that the North American and European wolverine were significantly different in several quantitative characters related to the size and shape of the skull and teeth. They concluded that the two wolverine populations represented two distinct subspecies, but were the same species, *Gulo gulo*.

The International Union for Conservation of Nature and Natural Resources (IUCN) states that "Most recent accounts [citing Jones *et al*. 1992, Pasitschniak-Arts and Larivière 1995, Wozencraft 2005] treat *luscus* as a subspecies of *Gulo gulo*, following Degerbøl (1935) and Kurtén and Rausch (1959)" (Abramov 2016, p. 1). We reviewed the references cited by IUCN. Jones *et al*. (1992, p. 17) only considers *Gulo gulo*. Pasitschniak-Arts and Larivière (1995, p. 1) state there are differences in the taxonomic treatment, and that, while *Gulo gulo* is now considered by most to be the extant *species*, others (including the above-cited Kurtén and Rausch (1959) and Rausch (1953)) have considered two *subspecies*. The Wozencraft (2005) citation is from Wilson and Reeder's previous 2005 publication, which was updated as of 2008. That account lists several "offspring" of *Gulo gulo*, but does not provide citations for the subspecies identified there, and at least two of those listed are not considered to be subspecific entities (e.g., *G. g. vancouverensis* and *G. g. luteus* (see Banci 1982, p. ii; Banci 1994, p. 104)). Finally, the COSEWIC Assessment and Status Report on the Wolverine (*Gulo gulo*) in Canada indicated that taxonomists recognize only a single subspecies (*Gulo gulo luscus*) in North America *or* consider *G. gulo* as single Holarctic taxon (COSEWIC 2014, p. 4).

Genetic analyses for the North American wolverine populations have primarily focused on genetic structure and variation of wolverine populations or subpopulations (see Kyle and Strobeck 2001; Kyle and Strobeck 2002; Zigouris *et al*. 2012, Zigouris *et al*. 2013). However, Frances' (2008, pp. 20–21) assessment of wolverine spatial genetic structure and demographic history (using mitochondrial DNA) indicated incomplete lineage sorting between North American and Eurasian populations, though comprehensive sampling has not been conducted for some areas (e.g., eastern Asia). A study by Tomasik and Cook (2005, entire) also concluded that reciprocal monophyly (i.e., distinct species) had not been attained between Eurasian and North American wolverine populations. Until additional studies are published, including robust genetic analyses in conjunction with additional sampling, the Service recognizes the North American wolverine as *G. g. luscus*.

Physical Appearance

Detailed descriptions of the wolverine are provided in Novikov (1962, pp. 196–202), Hash (1987, p. 575), Pasitschniak-Arts and Larivière (1995, pp. 1–2), and Wilson (1982, pp. 644–646), among others. Key distinguishing features are summarized here.

Wolverines are a medium-sized (about 1 meter (m) (3.3 feet (ft)) in length) carnivore, with a large head, broad forehead, and short neck (Pasitschniak-Arts and Larivière 1995, p. 1). Males are larger than females (Hall 1981, p. 1,007; Banci 1987, p. 35). Wolverines have heavy musculature and relatively short legs, and large feet with strong, curved claws for digging and climbing (Hash 1987, p. 575). Their feet are adapted for travel through deep snow and, during the winter, dense, stiff, bristle-type hairs are found between the toes and around the foot pad (Grinnell *et al*. 1937, pp. 265–266; Hash 1987, p. 575); this characteristic becomes diminished in the summer (Hash 1987, p. 575).

Adult wolverines are sexually dimorphic, with females weighing from 7 to 13 kilogram (kg) (15.4 to 29 pounds (lbs)) and males weighing between 10 to 18 kg (22 to 40 lbs) (North America) (Rausch and Pearson 1972, p. 264; Magoun 1985, pp. 19–21; Banci 1994, p. 99;

Copeland 1996, p. 20; Cardinal 2004, p. 8; Lofroth 2001, p. 11; Inman 2013, pers. comm.; Magoun 2013, pers. comm.; Aubry *et al*. 2016, pp. 17–18). The skulls of wolverine are large and heavy, and the strong jaw structure allows animals to feed on frozen flesh and crush bone (Haglund 1966, p. 269; Hash 1987, p. 575). Some geographic variation and sexual differences in skull morphology have been reported (Pasitschniak-Arts and Larivière 1995, p. 2). Wolverines have small, wide-set eyes, and are reported to have excellent hearing (Grinnell *et al*. 1937, p. 265; Krott 1960, p. 25; Bevanger 1992, p. 8).

Wolverine fur is short, thick, and uniform in thickness on the head and becomes longer towards the rear of the body (Hash 1987, p. 1). The coat consists of dense, woolly underfur (2-3 centimeters (cm) (0.8-1.2 inches (in) long) and coarse, stiff guard hairs, 6-10 cm (2.4-4 in) in length (Hash 1987, p. 1). The rich glossy coat can vary from medium brown to black (Banci 1994, p. 99; Pasitschniak-Arts and Larivière 1995, p. 1). Seasonal and individual variation in pelt color has been described (Degerbøl 1935, pp. 38–42; Grinnell *et al*. 1937, p. 252). In general, the head, tail, and legs are darker than the face (Pasitschniak-Arts and Larivière 1995, p. 1). An upper body stripe, which varies from creamy, pale buff to light brown or reddish in color (Pasitschniak-Arts and Larivière 1995, p. 1), extends from the nape of the neck, along the sides of the body, to the base of the bushy tail (Banci 1994, p. 99). White or orange patches are commonly found on the throat or chest (Pasitschniak-Arts and Larivière 1995, p. 1; Magoun *et al*. 2008, p. 24; Figure 14). The unique property of wolverine fur to shed frost (Hardy 1948, p. 330; Quick 1952, pp. 492–493), along with its rarity, has made wolverine pelts valuable for trade (Hash 1987, p. 575).

Various accounts state that wolverines have a strong sense of smell (Grinnell *et al*. 1937, p. 265; Bevanger 1992, p. 8) that allows them to locate carrion from great distances (Hornocker and Hash 1981, p. 1,297; *in litt.* Bevanger 1992, p. 8, citing Røskaft 1990; Copeland 1996, p. 100; Cardinal 2004, p. 8); however, experiments with young wolverines indicated a poor sense of smell, and that wolverines may locate food (areas where previously located or cached) based on their memory skills (Magoun 2013, pers. comm.) or learning abilities (e.g., Krott 1958, p. 241).

Scent-marking is used by mammalian carnivores for chemical communication (Hutchings and White 2000, p. 160). For wolverines, this behavior commonly includes urination (e.g., trees, stumps, snow) (Copeland 1996, p. 115; Magoun 1985, p. 105), but also includes scat, and scratches and bites on trees (Haglund 1966, pp. 225, 277; Copeland 1996, p. 115). Scent rubbing (see review by Rieger 1979) of the ventral (abdomen/stomach) area and anal rubbing have also been observed in wolverines (Pulliainen and Oyaskainen 1975, pp. 268–269; Rieger 1979, p. 22, *in litt.* Goethe 1964; Magoun 1985, p. 105). Scent marking by wolverines may also be an important chemical communication signal for potential wolverine prey. Field experiments conducted by Sullivan *et al*. (1985, pp. 928, 930) and Sullivan (1986, p. 388) found that black-tailed deer (*Odocoileus hemionus columbianus*) and snowshoe hares (*Lepus americanus*) avoided feeding on seedlings that were marked with wolverine urine.

Life History and Ecology

In this section we provide a summary of the individual and population needs (collective, species needs), including its life history, physiology and behavior, resource functions necessary for each

life stage (i.e., breeding, feeding, sheltering, dispersal), demographic information (abundance and distribution) and ecological setting.

Overview

Wolverines are active year-round and have been considered as primarily nocturnal (Iversen 1972b, p. 319; Pasitschniak-Arts and Larivière 1995, p. 7, and references cited therein). In his observational studies, Krott (1958, p. 168; 1960, p. 25) described periods of 3-4 hours of activity followed by 3-4 hours of sleep for wolverines in Scandinavia, a pattern also observed in Idaho (Copeland 1996, p. 77).

A study of body temperatures of caged wolverines, along with direct observations of animals obtained from Alaska and Sweden and previous studied animals (Alaska), suggested that wolverines were a day-active species, being very active in the morning, with periods of sleep during the night, a pattern that persisted in both winter and summer (Folk *et al*. 1977, p. 233). However, crepuscular activity (period just after dawn and just before sunset) may be a more accurate description for wolverine behavior (McCue *et al*. 2007, pp. 98–99). Others have remarked that wolverines exhibit a plasticity in their behavior (i.e., different behavior under different conditions) (Krott 1960, p. 26), a result attributed, in part, to their being a scavenging carnivore covering large areas (Stewart *et al*. 2016, pp. 1,495, 1,497). Several aspects of this plasticity are described below.

Wolverines are wide-ranging animals and known for traveling great distances in a short period of time (Krott 1960, p. 21; Gardner *et al*. 1986, p. 603; Woodford 2014, entire) (see Movement section below). This is due, in part, to their unique body structure. As described by Krott (1960, p. 20), they are "lumbrosacrally overbuilt" with heavy musculature and legs that are acutely angled when walking. Wolverine gait is characterized as either a 2X pattern (when patterns of two footprints repeat), used primarily in deep snow, and the more common 3X lopes (patterns of three footprints), for covering long distances over more compacted snow (Halfpenny *et al*. 1995, p. 104). The latter is described as a bouncing gait where all four feet may leave the ground at the same time (Halfpenny *et al*. 1995, p. 104).

As noted in our Species Description section above, in winter, the dense hairs on the foot pad and its body structure supports a low foot load, which has been estimated at 22 gram/cm^2 (Knorre 1959, p. 26) and 27–35 gram/cm^2 (Novikov 1962, pp. 22–23 (citing Dulkeit 1953)). This foot loading is believed to provide an advantage for wolverines preying on ungulates and other large mammals whose movements become restricted in deep snow (Knorre 1959, p. 26; Formozov 1963, pp. 40–41; van Zyll de Jong 1975, p. 435; Banci 1994, p. 113). However, a study of wolverines in boreal forest habitat in Canada present a differing interpretation of the wolverine foot adaptation based on tracking wolverines in snow over three winters (Wright and Ernst 2004a, pp. 58–59), in which they observed wolverines in their study area continuously selected for a path of least snow cover, where practicable, and only traveled in upland areas (Wright and Ernst 2004a, p. 59). They concluded that the low foot load is advantageous when snow crusts form, but, in deep snow, wolverines shift to an inefficient walking gait or lunge, which increases energy demand (Wright and Ernst 2004a, p. 59). They hypothesized that traveling in deep snow

during winter in search of food may increase the risk of starvation due to the greater energy expenditure (Wright and Ernst 2004a, p. 59).

Physiology

The wolverine is a snow-adapted, cold climate animal in its physiology, morphology (Telfer and Kelsall 1984, p. 1,830), behavior, and habits. Wolverines have been classified as a "chionphile" or those animals with adaptations for snow (e.g., increased surface area on feet, pelt characteristics) (see definitions in Pruitt 1959, p. 172; Cathcart 2014, p. 22).

In general, mustelids weighing more than 1 kg (2.2 lbs) have a basal metabolism (defined as the minimum metabolic rate for maintaining a comfortable warm temperature; Irving 1972, p. 121) that is about 20 percent higher than other mammals (Iversen 1972a, p. 343). For the wolverine, Young *et al*. (2012, p. 222) estimated a basal metabolic rate for a 15 kg (33 lbs) adult at 669.4 kcal/day, using Iversen's derived equation [Metabolic rate (M)=84.6*Weight (W, in kg)$^{(0.78)}$ ± 0.15] (Iversen 1972a, p. 343). By comparison, the estimated basal metabolic rate for a 53 g (1.9 ounce (oz)) least weasel (*Mustela nivalis*) is about 40 kcal/day, and approximately 250 kcal/day for a 3.8–5.5 kg (8.4–12 lbs) Arctic fox (*Vulpes lagopus*) (both sampled from Barrow, Alaska) (Irving 1972, p. 115; Figure 9.1).

Experimental studies by Iversen (1972, pp. 320–321; Figure 4) found that during their first 2.5 months, the basal metabolic rate for young wolverines was substantially higher than rates reported for other mammals ($W^{1.41}$ vs. $W^{1.0}$), then declined after 3 months, and declined again after 8 months. Because the early period coincides with weaning, Wilson (1982, p. 646) suggested that the observed peak may be related to changes in food consumed as well as improved thermoregulation since the mother is leaving the young for longer periods of time.

Energy expenditure during pregnancy is relatively low for mustelids (Oftedal and Gittleman 1989, p. 374); however, energy requirements for lactation in mammals can be over 4 to 7 times basal metabolic rates (Allen and Ullrey 2004, p. 478). Wolverines are known to consume a variety of food resources and seasonal switching of prey likely allows for adjustment for nutritional needs depending on life history stage (e.g., lactating females) (Krebs *et al*. 2007, p. 2,187 (Canada); Koskela *et al*. 2013a, pp. 103–104 (Finland); Yates and Copeland 2017, *in prep* (Montana)). Additional details on diet and feeding behavior for wolverines are provided below.

Thermal neutrality (or thermoneutrality) is the temperature range at which resting metabolism is at minimum (Barnett and Mount 1967, p. 468) and animals produce heat at a minimum rate in a thermal neutral environment (Barnett and Mount 1967, p. 413). For a resting mammal at thermal neutrality, body temperature is primarily maintained by "physical thermoregulation," that is, control of circulation in the skin and by sweating (Barnett and Mount 1967, p. 413). The body temperature of wolverine (measured by an implanted temperature transducer) at thermoneutrality has been reported at 38°C (100.4°F) (Folk *et al*; 1977, p. 231; Casey *et al*. 1979, pp. 332–333). The lower and upper bounds of the thermoneutral zone are established by the lower and upper critical temperatures (Monahan 2009, p. 5), which are related to physiological and behavioral adaptive responses. Because the critical temperature is the point at which the metabolic rate starts to rise, animals with lower critical temperatures are able to better conserve their energy

expenditure (Barnett and Mount 1967, p. 413). Arctic mammals, including wolverine, Arctic fox, and wolf (*Canis lupus*), have a (lower) threshold of thermoneutrality of between −30°C to −40°C (–22°F to –40°F) (Iversen 1972b, p. 322; citing studies by Scholander *et al*. (1950b) and Hart (1956)).

With increasing ambient temperatures, mammals, including Arctic mammals, have the ability to dissipate heat through vasodilation and rise of skin temperature (Scholander *et al*. 1950a, p. 251). Casey *et al*. (1979, p. 340) estimated a critical (upper) temperature for wolverine (14 kg (31 lb)) in summer pelage of 5°C (41°F) based on an observed increase in oxygen uptake at air temperatures below this temperature. For comparison, measurements of metabolic rates for the red fox (*Vulpes vulpes alascensis*) (Alaska) observed critical temperatures of 8°C (46°F) in summer (Irving *et al*. 1955, p. 184).

Arctic mammals, particularly small mammals, also adapt behaviorally to cold temperatures by creating burrows and building nest sites under the snow. Wolverines are known to dig holes in snow for shelter (Pruitt 2005, p. 120), and wolverine reproductive den sites located under deep snow may provide a thermoneutrality advantage for newborn wolverines (cubs or kits) (Magoun and Copeland 1998, p. 1,313). This topic is discussed in more detail below under Use of Dens and Denning Behavior.

Wolverines can also adapt to changes in ambient temperature and solar radiation by movement and micro- and macro-habitat selection (e.g., movemement to higher elevations in summer, shaded areas, and water sources). For example, wolverines are good swimmers, easily crossing lakes and rivers, and have been observed near and in lakes and other water bodies (Seton 1909, p. 950; Krott 1960, p. 23; Magoun 2017a, pers. comm.). They likely use these areas more frequently during warmer months both for cooling and hydration, and possibly for hygienic reasons (Krott 1960, p. 23).

Seasonal adjustments of fur insulation also provide an additional mechanism for mammals to overcome large seasonal changes in temperature (Casey *et al*. 1979, p. 340) and have been described for wolverine and other mammals in Alaska (Henshaw 1970, p. 522). An evaluation of the seasonal change in the insulation of fur of wolverine (pelts from Canada) found a 41.2 percent change in mean insulation values (measured as °C/cal/m^2/hr) from winter to summer (Hart 1956, p. 56). Relatedly, a single annual molting (between August and December) was noted in Grinnell *et al*. (1937, p. 251) (California), but twice yearly was described by Novikov (1962, p. 201) (Russia). The reported seasonal decrease in wolverine fur thickness also correlates with experimental results of Casey *et al*. (1979, p. 337) who indicated that a seasonal shift in oxygen consumption below critical temperature was likely due to an increased rate of heat loss in summer.

Physiological responses (e.g., changes in respiration rate, heart rate, skin temperature, or endocrine (hormone) functions) to changes in ambient temperature or solar radiation are also observed in mammalian species (e.g., Norris and Kunz 2012, p. 207). In a study to evaluate metabolic and respiratory responses of eight terrestrial Arctic mammals to ambient temperature during summer months, Casey *et al*. (1979, pp. 335, 338; Figure 7) found that, for wolverines, the frequency of respiration was generally constant (15-20 per minute) for air temperatures

ranging from −20°C to +20°C (−4 to 68°F). However, with decreasing ambient temperature, the tidal volume (air moved per breath) of wolverines exhibited a constant increase (Casey *et al*. 1979, p. 335). The researchers inferred that the increased ventilation of wolverines at low ambient temperatures was the result of an increased energy metabolism (Casey *et al*. 1979, p. 336). One veterinarian reported an enlarged thyroid in a wolverine during a necropsy procedure (Copeland 2017, pers. comm.), which is suggestive of a high metabolism.

Beyond these cited observational and laboratory/field studies, the best available information indicates that behavioral, morphological, or physiological responses of wolverines relative to increases in either colder or warmer temperatures have not been evaluated (i.e., thermal imaging studies combined with biophysical models (e.g., McCafferty *et al*. 2011, entire) or physiological niche models). These types of studies could provide useful information for estimating metabolic heat loss and relative energy costs for wolverines for certain activities, and for measuring seasonal changes in physiological states.

Range and Habitat Use

Historical Range and Distribution

Phylogeography/Phylogenetics

Results from a molecular study of phylogenetic relationships of the Mustelidae family suggest at least six radiation episodes within this family since the Early Eocene Epoch (approximately 50 million years before present (YBP)) (Marmi *et al*. 2004, pp. 488, 492). The split of the marten (*Martes*, *Gulo*) and weasel (*Mustela*) lineages occurred in the Early Middle Miocene Epoch (14 to 11 million YBP), with the separation of Old World and New World lineages (*Martes*, *Gulo*) occurring in the Late Miocene Epoch (8.6 to 5.8 million YBP) (Marmi *et al*. 2004, p. 488). The *Gulo* genus appears in the fossil record in the mid-Pleistocene in both Europe and North America (Bryant 1987, p. 659).

The dispersal of *Gulo* across Beringia (land mass that extended from Siberia into interior Alaska during the Pleistocene) is believed to have produced contemporaneous records for the species in Europe and North America (Bryant 1987, p. 659). Genomic data was examined using a molecular dating technique to estimate an approximate age of the *G. gulo* ancestor (Malyarchuk *et al*. 2015, entire). The researchers estimated a relatively recent origin of the species *Gulo gulo* at about 181,000 to 234,000 YBP (Malyarchuk *et al*. 2015, pp. 1,115–1,116). They note that this latter time period corresponds to the Riss glaciation period (187,000 to 230,000 YBP), a time of genetic divergence of amphi-Beringian (both sides of Beringia) species and speciation events (Hope *et al*. 2013, p. 426). Their results, along with fossil information, also indicate the divergence of the *Gulo* branch and the other *Martes* taxa occurred during the Late Miocene-Early Pliocene (5.6 million YBP), and lends support for strong evolutionary processes in the northern Siberian ecosystems in the Pliocene and Pleistocene Epochs (Malyarchuk *et al*. 2015, pp. 1,116–1,117).

An evolutionary trend was described in which *Gulo* increased in size from the mid- to late-Pleistocene, with a subsequent reduction in size post glaciation, as well as small changes in

selected teeth, and a possible shift to colder habitats (Bryant 1987, p. 660). The Late Pleistocene and the Pleistocene-Holocene transition represent the end of prolonged period that was characterized by climate fluctuations followed by rapid warming (Post 2013, p. 28). This analysis also indicated that both the mid-Pleistocene European *Gulo schlosseri* and the early North American *Gulo* appear to be adapted to a warmer climatic environment, but are likely to have also occupied colder climates (Bryant 1987, p. 660). Other factors such as competition (Guilday 1971, p. 237), predator avoidance, and prey abundance may also have been important in creating significant shifts in geographic ranges for certain species during glacial cycles.

Wolverines are believed to have migrated to North America during the late Pleistocene, although fossil evidence from the Pleistocene Epoch for wolverine is limited (Anderson 1977, p. 15; Bryant 1987, p. 660), and most fossil material is either cranial or dental fragments (Bryant 1987, p. 660). A summary of records for both Pleistocene and extant *Gulo* (Bryant 1987, p. 659; Table 3) includes findings in the United States from Colorado, Idaho (e.g., White *et al*. 1984, p. 248 (lava tubes)), Alaska, Maryland, and Pennsylvania, and Canada (primarily the Yukon region) ranging from the Irvingtonian Age (1.8–2.4 million YBP) to Late Wisconsinan-Holocene (15,000 YBP to present day).

Genetic studies can provide an understanding of the postglacial recolonization of wolverines following the Last Glacial Maximum, a period of rapid cooling, and movement patterns due to changed climatic conditions (Frances 2008; Zigouris *et al*. 2013; McKelvey *et al*. 2014). During the late Wisconsin period (10,000 to 25,000 YBP), approximately 60 percent of North America was covered by glacial ice (Rogers *et al*. 1991, p. 624). However, several ice-free refugia existed at that time including the Beringian refugium, which included eastern Siberia, most of Alaska, areas of northwestern Canada, and areas of the Bering Sea shelf that were exposed by lower sea levels; this refugium harbored a number of mammalian species including wolverine (Rogers *et al*. 1991, pp. 624, 626). Analyses by Frances (2008, entire) and Zigouris *et al*. (2013, entire) supported a wolverine colonization of North America in which individuals "followed retreating glaciers" (Zigouris *et al*. 2013, pp. 10–11), beginning about 21,000 YBP, following the Last Glacial Maximum, when a period of rapid warming resulted in additional extinction events, particularly large mammalian megafauna (Post 2013, p. 29)

A phylogeographic analysis presented by McKelvey *et al*. (2014, p. 331) proposed that a unique haplotype (Cali 1) observed in historical wolverine samples from California was reflective of an independent evolutionary history resulting from isolation (i.e., southern ice-free refugium) of wolverines during glacial retreat. However, Zigouris *et al*. (2013, p. 10, Supplemental Table S5) found the Cali 1 haplotype described by Schwartz *et al*. (2007, p. 2,173; Tables 2 and 4) (relabeled as Haplotype 21) also occurred in historical wolverine samples from the eastern region of Canada (Quebec-Labrador). In addition, as noted by Zigouris (2014, pp. 232–233) the historical samples analyzed by McKelvey *et al*. (2014, p. 327; Table 1) were primarily those from locations at the southwestern edge of the wolverine's North American range (e.g., California, Colorado, Idaho, Montana, Wyoming, Utah, Washington). Without additional sampling, it is unclear if this particular haplotype distribution from two of the most peripheral North American wolverine populations is a reflection of a skewed dispersal after post-glacial colonization, or was a more widely distributed haplotype that declined or was lost due to hunting

and unregulated trapping pressures (beginning in 18th century) or fragmentation (late-20th century) (Zigouris *et al*. 2013, p. 10).

Additional discussion of our current understanding of wolverine genetic structure and diversity is provided in the *Population Structure* section below.

Historical Range

In North America, wolverines were historically distributed in much of the northern portion of the continent, extending southward to the northernmost region of the United States (Maine to Washington) or approximately north of the 38th parallel (Hash 1987, p. 576; Banci 1994, p. 102). An estimate of wolverine observations and distribution in the contiguous United States was prepared by Aubry *et al*. (2007, entire) by compiling 901 verifiable or documented records of wolverine occurrence dating from 1801 to 2005 from 24 states in the contiguous United States. This included a total of 809 verifiable or documented records for the Rocky Mountain and Pacific Coast mountains (west-northwestern United States) for this time period (Aubry *et al*. 2007, p. 2,151).

The historical population size of wolverines in Canada is not known (Fortin 2005, p. 4). Its historical distribution, as depicted by Seton (1909, p. 947; Map 51) and also later by van Zyll de Jong (1975, p. 435; Figure 9) shows a broad range across much of Canada. Examples of early descriptive accounts include de Puyjalon (1900, pp. 126–144), who described wolverines as inhabiting Labrador, Canada (de Puyjalon, p. 101), and extending in range to the 66th parallel and perhaps further (de Puyjalon 1900, p. 144); reports of both trapped and live wolverines in Labrador in the late 1700s (Townsend (ed.), 1911, pp. 73, 93, 228, 255); and reports of wolverines as "common" in Canada's Nunavut Territory (Hudson Bay region) during a 1920s Danish excursion (the Fifth Thule Expedition) to Arctic North America (Freuchen 1935, p. 101). The 2014 COSEWIC report presents a historic range distribution for Canada based on personal accounts and interpretation of the fur trade (COSEWIC 2014, pp. 12–13; Figure 3).

Maximum Extent of Occurrences

We created a map to depict the Maximum Extent of Occurrences of the North American wolverine (Figure 2). This map represents the area defined by circumscribing all known localities of a species, or the perimeter of the outermost geographic limits based on occurrence records from survey reports, literature, natural heritage databases, and citizen sightings (Service 2017), and is similar to the map presented in Seton (1909, p. 947; Map 51). For defining this area in Canada and Alaska, we used the map (in digital format) presented in COSEWIC (2014, p. 12; Figure 3). For the western United States, we used relevant forested areas identified in the U. S. Environmental Protection Agency's (EPA) ecological regions map (Level 2) (Commission for Environmental Cooperation (CEC) 1997, EPA 2010; see **Appendix A** for complete North American map). The Maximum Extent of Occurrences area shown in Figure 2 encompasses about 12,966,221 km^2 (3,204,022,986 ac) (using North American Lambert projection). Of this total, about 88 percent (11,363,639 km^2 (2,808,016,350 ac)) is found north of the U.S.–Canada border.

We also requested all available wolverine observations for the west-northwestern United States from State agencies (e.g., wildlife agencies, natural heritage programs) and the Forest Service Natural Resource Information System (NRIS) Wildlife Database. The NRIS Wildlife Database records represent data that have been quality checked ("passed") by the Forest Service. For the Midwest and eastern United States, we incorporated observations from Aubrey *et al*. (2007; pp. 2,151–2,152; Figure 1). We also located additional historical records in the western United States that do not appear in Aubry *et al*. (2007, p. 2,151); for example, Nead *et al*. (1985, entire) identified several positive and probable reports of wolverines in Colorado in the late 1970s and a wolverine was reported from the Squaw Valley region of California in the summer of 1953 (Ruth 1954, pp. 594–595). For the contiguous United States, these wolverine observations are included in Figure 2, and illustrate the close correlation with the ecological regions boundaries.

Current Potential Extent (of Occurrence)

Using the best available information, we created maps to describe an area of Current Potential Extent for the western-northwestern contiguous United States (Figure 3). Current Potential Extent represents the Maximum Extent of Occurrences minus those areas where we believe the species has been extirpated. For the wolverine, this map was developed beginning with ~~from~~ the current distribution of breeding populations defined by Inman *et al*. (2013, p. 282; Figure 3). We then incorporated records from Forest Service (NRIS Wildlife Database) from 2004 to early 2017, based on an average male life expectancy of about 13 years (Jung and Kukka 2013, pp. 8, 12). We also incorporated more recent observations (e.g., telemetry, camera traps, and/or mortality reports) reported from California, Washington, Colorado, Wyoming, Utah (2014, 2016), Oregon (2016, 2017), Michigan (2004-2010) and North Dakota (2016). Sources include California Department of Fish and Wildlife; Idaho Department of Fish and Game; Michigan Department of Natural Resources; Utah Division of Wildlife Resources; Wyoming Game and Fish Department, Magoun *et al*. 2013, p. 27; Magoun 2017b, pers. comm.), and additional observations from 2016-2017 winter surveys (i.e., WAFWA's Coordinated Occupancy Survey) (Montana Fish Wildlife & Parks (FWP) 2017, pers. comm.)). This Current Potential Extent map is illustrated in Figure 3.

Our Current Potential Extent map also identifies individual wolverine observations outside these boundaries, including a current resident wolverine in California, the tracking locations of a dispersing male (M56) that traveled from northwest Wyoming into Colorado and then to North Dakota where it was legally shot in 2016 (Packila *et al*. 2017, entire), and three observations from experienced observers in Colorado (2015) and one from Wyoming Game and Fish Department (January 2016). We placed a question mark on this map for the Uinta Moutains region of Utah since the last detection was a female wolverine found killed along a highway in July 2016 (Rich County) (Hersey 2017, pers. comm.).

The Current Potential Extent area identified in Figure 3 encompasses approximately 280,316 km^2 (69,267,592 ac) (North American Lambert projection). We also prepared a Current Potential Extent map for all of North America, including Canada and Alaska, shown in Figure 4, using the map presented in the 2014 COSEWIC Status Assessment and Status Report on the wolverine in Canada (COSEWIC 2014, p. 12; Figure 3), for a total estimated Current Potential Extent of 8,114,878 km^2 (2,005,230,024 ac) (North American Lambert projection). The Current Potential

Extent area in the contiguous United States represents approximately 3.45 percent of the total Current Potential Extent of wolverines in North America.

We recognize that this depiction does not necessarily represent current areas where reproducing populations of wolverines are found, nor does it capture unverified accounts from New Mexico, described in Frey (2006, pp. 20–21) for the Sangre de Cristo Range, and visual observations reported by two individuals (2005 and 2016) in response to our *Federal Register* notice (81 FR71670; October 18, 2016) requesting information for our status review.

Habitat loss (historical vs. current range) for the North American wolverine (i.e., Canada and United States) has been estimated at 37 percent (Laliberte and Ripple 2004, p. 126), using "historic" and "current" range maps, an estimation of "human footprint," North American biomes, and a digital elevation model (Laliberte and Ripple 2004, p. 126). Similarly, comparing our estimated North American Maximum Extent of Occurrences map to our estimated Current Potential Extent map for the wolverine also indicates a 37 percent change. However, as noted in Laliberte and Ripple (2004, pp. 125–126), we recommend caution in interpreting this value given the coarse scale used to define dynamic range boundaries.

We provide a discussion of wolverine population abundance and distribution in more detail in the *Biological Status–Current Condition* section below.

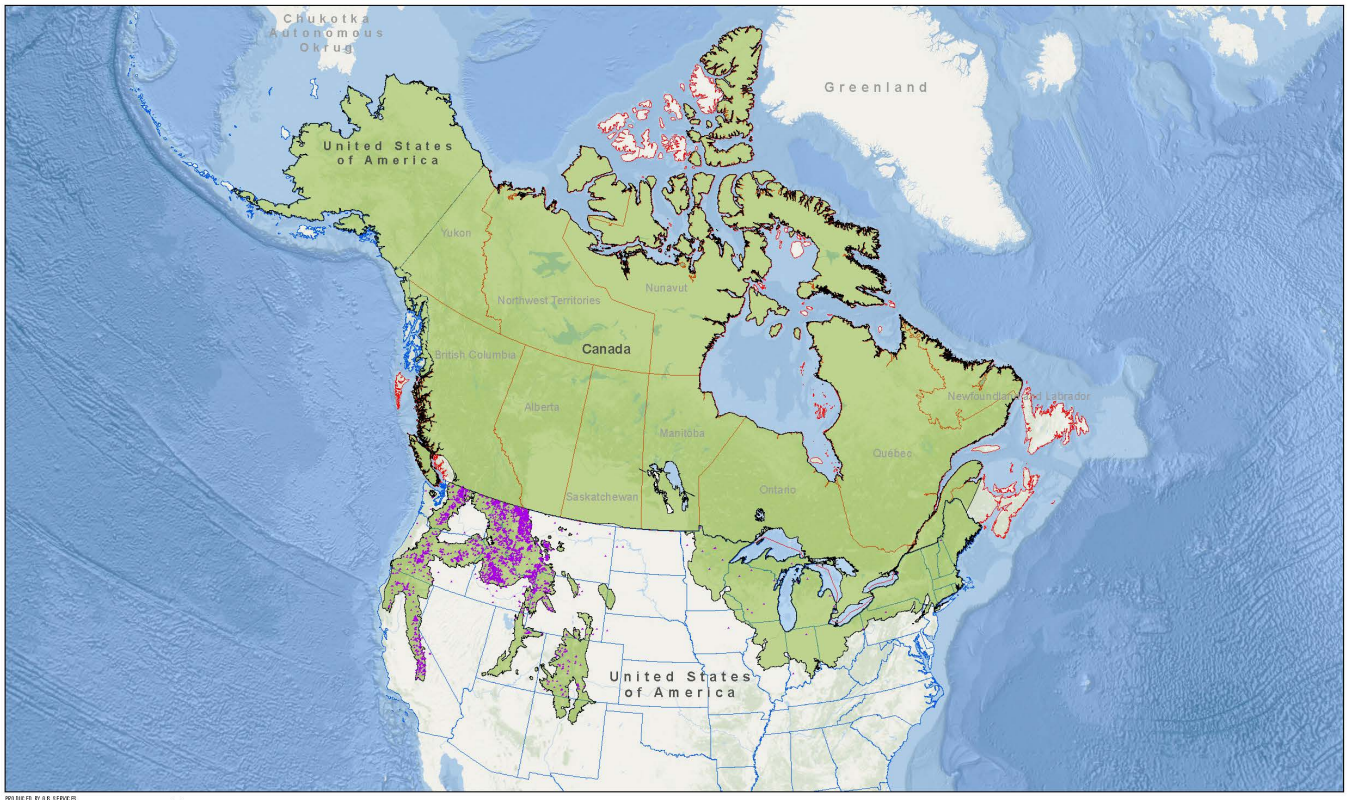


Figure 2. Maximum Extent of Occurences for the North America wolverine shown with observations (1800s to 2017) within the contiguous United States. Boundaries based on COSEWIC (2014) (for Canada and Alaska) and North American Ecological Regions (CEC 1997, EPA 2010) (for the contiguous United States).

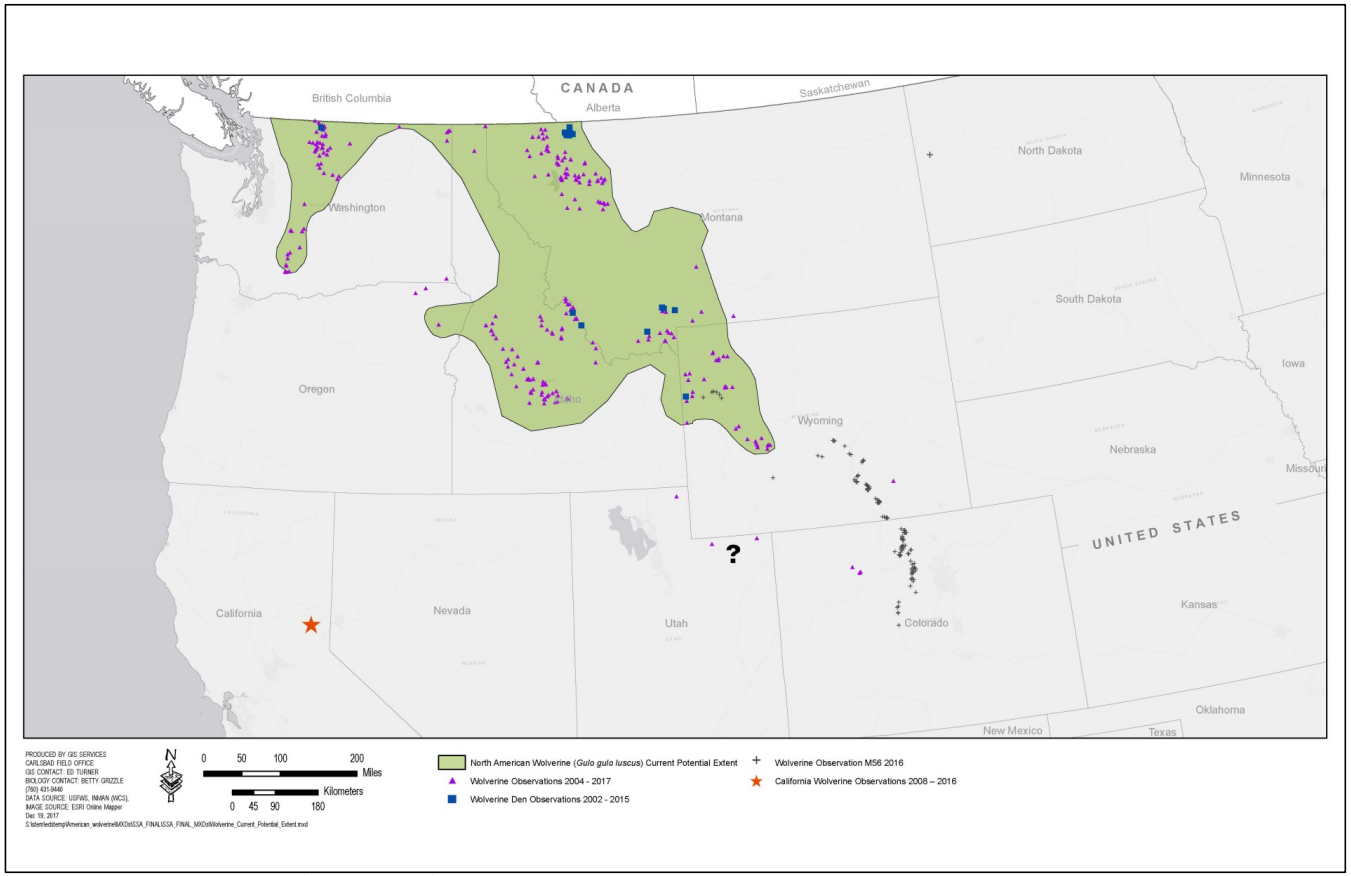


Figure 3. Current Potential Extent of Occurrence for~~of~~ the North American wolverine, western United States. Adapted from Inman *et al*. (2013) and records from California Department of Fish and Wildlife; Forest Service NRIS; Idaho Department of Fish and Game; Magoun *et al*. (2013); Magoun 2017b pers. comm.; Michigan Department of Natural Resources; Montana Fish Wildlife & Parks; Utah Division of Wildlife Resources; Wyoming Game and Fish Department; and den records provided by Copeland, Heinemeyer, and Inman.

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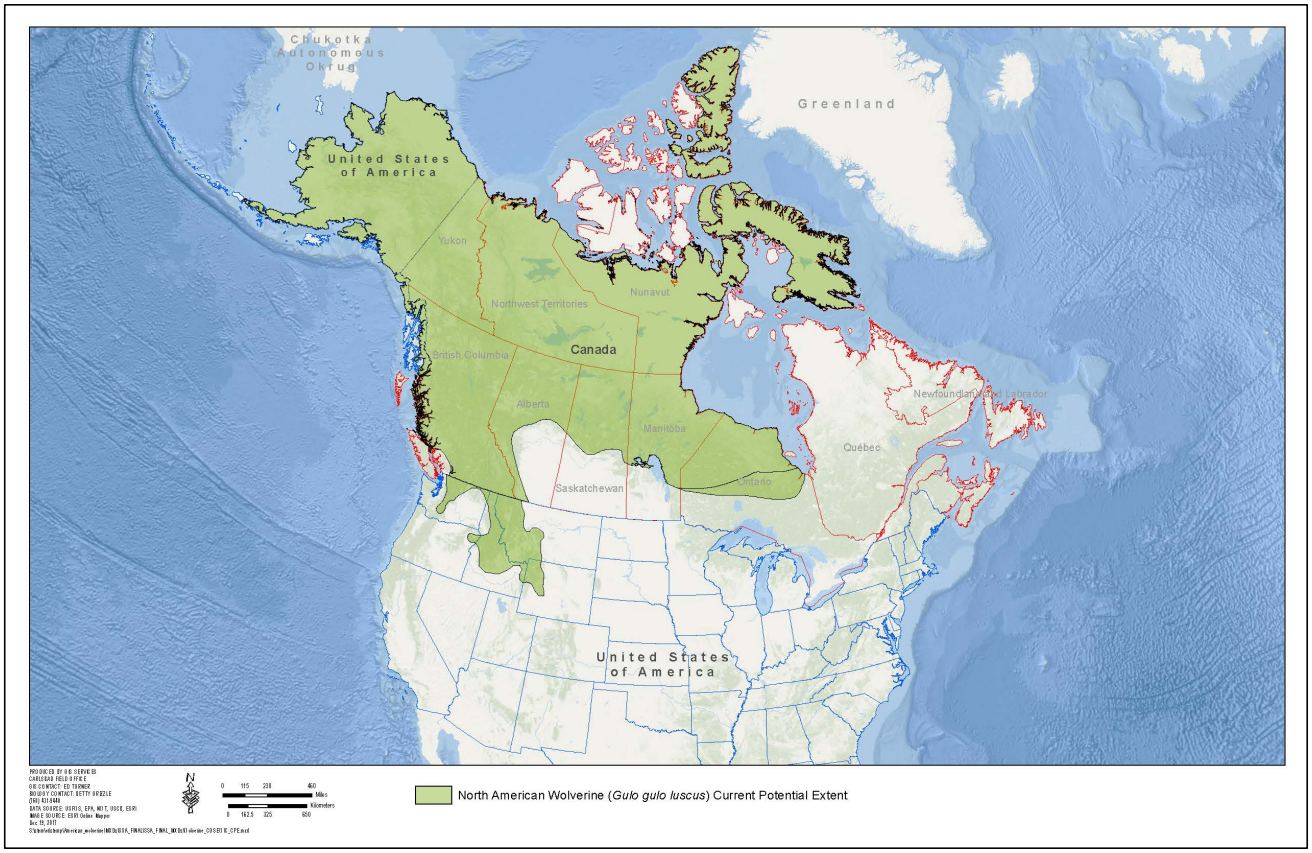


Figure 4. Current Potential Extent of Occurrence for~~of~~ the North American wolverine, North America. Adapted from COSEWIC (2014), and records from Forest Service NRIS; Idaho Department of Fish and Game; Magoun *et al.* (2013); Magoun 2017b pers. comm.; Montana Fish Wildlife & Parks; Utah Division of Wildlife Resources; Wyoming Game and Fish Department; and den records provided by Copeland, Heinemeyer, and Inman.

Habitat Use

Wolverines occupy a variety of habitats within North America, including Arctic tundra, subarctic-alpine tundra, boreal forest, mixed forest, redwood forest, and coniferous forest (Banci 1994, p. 114). However, these broad, landscape-scale vegetation associations can obscure other habitat variables important for wolverines, including features found within peripherally occupied areas or areas of high elevation (Banci 1994, p. 114). In Canada, wolverines use a wide variety of forested and tundra vegetation, at all elevations (COSEWIC 2014, p. 18).

When viewed by ecological region (see EPA 2010), in general, wolverine observations found within our Current Potential Extent of wolverines in the contiguous United States are most commonly found in the Northwestern Forested Mountains ecological region. A complete North American ecological region map is presented in **Appendix A**.

Studies of wolverines in central Idaho found that montane coniferous forests comprised two-thirds of available habitat (Copeland 1996, p. 120). Wolverines in this region also exhibited a seasonal preference, with subalpine rock habitats used in summer and montane coniferous forests used most often in winter (Copeland 1996, p. 120). In addition, individuals within this study population commonly crossed natural openings and those areas with little cover, including burn areas, meadows, or open mountain-top areas (Copeland 1996, p. 124).

Observations of summer movements of wolverines in northwestern Montana indicated that both males and females moved to higher, cooler elevations and remained there throughout the summer (Hornocker and Hash 1981, p. 1,299). In the Greater Yellowstone Ecosystem, wolverines selected areas that contained steep terrain with tree cover, high elevation meadows, boulder or talus fields, and avalanche chutes (Inman *et al*. 2012a, p. 785). In this region, wolverines selected elevations at and above the treeline during summer, moved slightly lower during winter, but avoided low-elevation winter ranges occupied by potential prey (e.g., elk) or areas with little human activity (Inman *et al*. 2012a, p. 785). The avoidance of these areas may be the result of lack of tree or talus field cover at these low elevations, in combination with presence of potential predators (e.g., wolf, mountain lion (*Puma concolor*) or competitors (e.g., coyote (*Canis latrans*), bobcat (*Lynx rufus*) (Inman *et al*. 2012a, p. 785).

Several habitat association-type models have been developed for both North American and European wolverines. In the northern Rockies (including Canada and the United States), Carroll *et al*. (2001, p. 975) found that elevation and north-facing cirque habitat variables (i.e., alpine areas), when incorporated into empirical habitat models, were significantly correlated with wolverine occurrence (Carroll *et al*. 2001, pp. 971, 973–974). Copeland *et al*. (2007, entire) also evaluated habitat associations for wolverines in central Idaho. Wolverines were found to be associated with high elevations (2,200 to 2,600 m (7,218 to 8,530 ft)) with a slight downward shift in summer (Copeland *et al*. 2007, p. 2,207). These movements correspond with a shift in cover types, from high-elevation whitebark pine (*Pinus albicaulis*) communities in summer to mid-elevation Douglas fir (*Pseudotsuga menziezii*) and lodgepole pine (*Pinus contorta*) in winter (Copeland *et al*. 2007, pp. 2,207–2,208). Results from a study of wolverines in Scandinavia suggested that topography may be important in providing refugia from predators and may

therefore facilitate the co-existence of wolverines with larger carnivores such as wolves (Khalil *et al*. 2014, p. 636).

In interior Alaska, wolverines were also found to be positively associated with high elevations (Gardner *et al*. 2010, p. 1,901). The authors concluded that wolverines avoided human influences based on observations that both elevation and human influence were found to have significant effects on wolverine occurrence probabilities, but were not correlated (Gardner *et al*. 2010, p. 1,901). However, the authors indicate that their sampling design was not able to determine which human activities influenced wolverine behaviors; a combination of intensity of development and harvest activities was suggested (Gardner *et al*. 2010, p. 1,901). Current studies are underway in the North Slope region of Alaska to evaluate fine-scale habitat selection of wolverines related to denning, caching, day bed use, and snow holes (Dorendorf 2016, p. 6). Day beds were also described by Haglund (1966, p. 268) for wolverines studied in Sweden.

A Canadian study also found that habitat associations, at least for females, are complex, and include combinations of several modeled variables that supported hypotheses related to food (prey distribution), predation risk (based on a ruggedness index), or human disturbance (winter recreation activity, roads, and forest harvesting) for both summer and winter in two study areas located in northcentral and southeast British Columbia (Krebs *et al*. 2007, pp. 2,186–2,187). Wolverines in the Rocky Mountains of Alberta, Canada, were found to more likely occupy areas with increasingly rugged terrain (Fisher *et al*. 2013, pp. 710–712).

Camera trapping was used to study wolverine behavior in varying habitat in the Rocky Mountains of Alberta, Canada (Stewart *et al*. 2016, entire). That study found that wolverine behavior differed in landscapes that had been significantly modified by human activities as compared to those with light modifications or in protected areas (Stewart *et al*. 2016, p. 1,499). They concluded that wolverine occurrence in their study areas varied more strongly with linear features (seismic lines created from oil and gas exploration, pipelines, transmission lines, roads, and rail lines) than with the degree of snowpack, supporting the idea of human "footprint" as a driver of habitat suitability for wolverines; that is, if snowpack conditions was the only driver for suitable habitat, then a behavioral signal (e.g., latency time for detection or tree climbing at bait station) would not be expected (Stewart *et al*. 2016, p. 1,501).

Aboriginal knowledge holders (Aboriginal Peoples that have accumulated knowledge about wildlife species and their environment) in Canada have reported that while wolverines appear to avoid human habitation and developed areas, some wolverine will visit these areas if they do not appear to be threatened or if development activities cease (Cardinal 2004, p. 22). Wolverines were described as occupying deserted snow huts (Nunavut Territory, Canada) during winter months (Freuchen 1935, p. 98).

A negative association with roads and wolverine (and caribou) occurrence in boreal forest habitat was reported in northwestern Ontario, Canada, and wolverines in that study area avoided deciduous forests (Bowman *et al*. 2010, p. 464). However, a study of wolverines in upland boreal forests of Canada found that wolverines followed open linear corridors that offered compact snow conditions, including winter roads, recent seismic lines, snowmobile trails, and all-terrain vehicle tire tracks for travel of distances up to 3 kilometers (km) (1.86 miles (mi)) (Wright and

Ernst 2004b, p. 59). In central Idaho, wolverines were reported using snowmobile winter access (unmaintained) roads for travel (Copeland *et al*. 2007, p. 2,210).

A study of wolverine selection patterns in boreal forests in northwestern Alberta using resource selection function (RSF) modeling techniques[2] and data from telemetered wolverines found that, for the winter season, both male and female wolverines selected for streams, forested areas (broadleaf, coniferous, and mixed) and bogs or fens, while avoiding active oil and gas well sites and low-traffic winter roads (Scrafford *et al*. 2017, pp. 31, 32). That study also found that wolverines did not avoid older seismic lines, likely due to the intermediate stage of regeneration found in their study area and the availability of small prey in conjunction with minimal risk of human or wolf presence (Scrafford *et al*. 2017, p. 34).

RSF-based modeling was used to quantify the relationship between the observed distribution of the wolverine and variables representative of habitats and human disturbance in the taiga and tundra ecoregions (see ecological regions map in **Appendix A**) of the Canadian central Arctic (Nunavut and Northwest Territories) (Johnson *et al*. 2005, p. 10). Using a range defined by previous studies of collared wolverines, researchers identified two seasons for wolverines, based on presence or absence of barren-ground caribou (*Rangifer tarandus groenlandicus*) (Johnson *et al*. 2005, p. 8). They found that, in winter, the occurrence of wolverines was correlated with patches of heath rock and rock association, and areas dominated by sedge (Johnson *et al*. 2005, pp. 23–25). Results for models for summer season were less clear, but models that included grizzly bear (*Ursus arctos*), caribou, and wolf were found to be positively associated with wolverine, likely due to the scavenging opportunities and hunting of caribou provided by these other carnivores (Johnson *et al*. 2005, p. 24). In Finland, the presence of wolves was found to be one of the most important variables influencing habitat selection of wolverines (Koskela 2013, p. 35) likely due to the increased scavenging opportunities provided by wolf kills (Koskela 2013, p. 36).

A RSF model was also used to develop a predictive map of wolverine habitat for the western United States (Inman *et al*. (2013, p. 281). Their best fit model found that, in general, wolverine were most likely to be distributed at high elevations, with steeper terrain, more snow, fewer roads, and reduced human activity, but also in proximity to high elevation talus, tree cover, and areas that had snow cover on April 1 (Inman *et al*. 2013, pp. 280–281). Primary habitat for the wolverine in the western United States was estimated at 164,125 km^2 (63,369 mi^2) (Inman *et al*. 2013, p. 281). Additional information related to the results of this modeling effort is discussed in the *Population Distribution and Abundance* section below.

[2] RSF is any mathematical function that is proportional to the probability of use of a resource unit (Manly *et al*. 2002, p. 15). A RSF contains several coefficients that quantify the selection for or avoidance of an environmental feature, and the sign/strength of those coefficients represents a differential variation in the distribution of each environmental feature measured at a sample of locations to a comparable set of random sites. Thus, when an animal's observed use of a resource is greater than those random sites, selection of that feature is inferred (Johnson *et al*. 2005, p. 10).

Movement

Wolverine movements are related to both territoriality (within home ranges) and dispersal (adults and young). Movement within home ranges by adult male and female wolverines is extensive. For example, wolverines monitored in the Greater Yellowstone Ecosystem traveled a distance that was equivalent to their average home range diameter in less than 2 days, which is also about the size of their home range circumference in less than 1 week (Inman *et al*. 2012a, pp. 782–783). This study also found that, for a 24-hour period, the average minimum distance traveled was 15.5 (km) (9.63 (mi) for males and 7.5 km (4.66 mi) for females (Inman *et al*. 2012a, p. 783). Telemetry studies of wolverines in south-central Alaska indicate an average distance traveled per day of approximately 12 km (7.46 mi) for females and 8–21 km (4.97–13 mi) for males (Woodford 2014, no page number). Observations from snow tracking studies have found instances where two individual wolverines traveled together (Wright and Ernst 2004b, p. 63).

A study of female Fennoscandinavian wolverines found that most (86 percent) females remained stationary in their established territories, with 8 percent vacating and 6 percent expanding their territory (Aronsson 2017, p. 40). In addition, this study of 42 female wolverines in 122 territories reported that females with established territories only moved to available territories that were higher than average in quality (Aronsson 2017, p. 41). In central Norway, a study of spatial and temporal patterns in wolverines using noninvasive genetic sampling methods also found that individuals tended to stay in the same general area from one year to the next (Bischof *et al*.'s 2016, p. 1,533).

A number of factors can affect wolverine movements within territories, such as availability of food, temperature, and breeding activity. Seasonal shifts in elevation have also been observed for wolverines in the contiguous United States. An ecological study of wolverines in southcentral Alaska found significant movement up in elevation during late winter and early spring as well as significant movement down in elevation during the late fall and winter (Gardner 1985, p. 21). Wolverines were also observed moving to and occupying higher and presumably cooler elevations in summer months in northwestern Montana (Hornocker and Hash 1981, p. 1,299). In Central Idaho, wolverines exhibited a preference for higher elevation areas containing rock and talus cover in summer months, but moved to lower elevations in winter, which was attributed to the result of an increase in availability of carrion related to the fall hunting season (Copeland 1996, p. iv). Two aboriginal knowledge holders in the Kivalliq region (Nunavut, Canada) reported that wolverines will move closer to communities during caribou migration in the fall, likely attracted by the large number of caribou carcasses left by hunters (Cardinal 2004, p. 22).

A study of wolverine movement in boreal forest habitat in Canada (northwestern Alberta and northeastern British Columbia) during winter months found that wolverines chose the most direct travel route with the least snow cover (Wright and Ernst 2004a, pp. 58–59). Woodford's (2014, no page number) account of wolverine observations from studies in Alaska indicated that, when pursued, wolverines will run uphill, which may represent a predator-avoidance adaptive behavior.

As discussed in more detail below (*Diet and Feeding*), several studies have shown that wolverines exhibit a seasonal shift in diet, and Hornocker and Hash (1981, p. 1298) concluded

that food availability was the primary factor determining both movements and home ranges for wolverines studied in northwestern Montana. Movement patterns of adult males during the summer months are also likely influenced by breeding activity (Magoun 1985, p. 66).

Males and females maintain large territories with very little overlap between same-sex adults (Magoun 1985, p. 38; Banci 1994, p. 118; Inman *et al*. 2012a, p. 783; Bischof *et al*. 2016, pp. 1,532–1,533; Regehr and Lacroix 2016, p. 249), but breeding pairs have overlapping territories (Copeland 1996, pp. 55–61; Hedmark *et al*. 2007, p. 19; Dawson *et al*. 2010, p. 413; Persson 2010, p. 52; Inman *et al*. 2012a, p. 787). However, ranges of young males, that have not yet dispersed, can overlap with resident adult male home ranges (Alaska) (Magoun 1985, p. 64). Studies of wolverines in the Greater Yellowstone Ecosystem found a mean percent overlap of 12.7 percent for same sex, adult–subadult pairs and about 24 percent for opposite sex, adult–sub-adult pairs (Inman *et al*. 2012a, p. 787). In addition, Inman *et al*. (2012a, p. 783) found that when a resident adult wolverine died, same-sex adults (not known to be located within the dead wolverine's home range) would begin using (within 3–7 weeks) areas of the unoccupied home range, or same-sex subadults would expand into and then occupy most or all of the dead wolverine's former home range. A study of territoriality of wolverines in central Norway (using scat analysis) indicated that within their study population, wolverines were also more likely to choose a home range area that was previously used by a neighboring same sex individual after that individual's death (Bischof *et al*. 2016, p. 1,533).

Table 1 presents a summary of annual home ranges of resident wolverines.

Table 1. Annual Home Range Size for Adult, Resident Wolverines.

Region	Female, km^2 (mi^2)	Male, km^2 (mi^2)	Reference
Central Idaho	384 (148)	1,582 (610)	Copeland 1996
Central Idaho / Yellowstone Region	357 (138)	1,138 (439)	Heinemeyer and Squires 2015
Greater Yellowstone Ecosystem	303 (117)	797 (308)	Inman *et al*. 2012a
Glacier National Park (MT)	139 (54)	521 (201)	Copeland and Yates 2008
Alaska (Northwestern)	53-232 (20-89.6)	488-917 (188-354)	Magoun 1985
Canada Northwest Ontario	50-400 (19-154) 423 (163)	230-1,580 (89-610) 2,563 (990)	COSEWIC 2014 Dawson *et al*. 2010
Central Norway	331 (128)	757 (292)	Bischof *et al*. 2016
Southern Norway	274 (106)	663 (256)	Landa *et al*. 1998
Northern Sweden	170 (66)	669 (258)	Persson *et al*. 2010

Home range use is smaller for female wolverines during the reproductive period. For a parturient (about to bear young) female, estimates of home range size in the Greater Yellowstone region were significantly smaller, with a minimum of 100–150 km^2 (39–58 mi^2) (i.e., during year raising young) (Inman *et al*. 2012a, p. 782). The average home range size for lactating females

rearing young was estimated at 70 km^2 (27 mi^2) from March through August (Alaska) (Magoun 1985, p. 36). In northwestern Ontario, researchers reported a home range of 262 km^2 (101 mi^2) for a lactating female (Dawson *et al*. 2010, pp. 141–142). In general, the distance traveled by female wolverines depends on the location of the reproductive den site within the home range, the areas used for locating food/prey, and the territory border (Myhr 2017, no page number).

In summary, habitat diversity, food availability, and competition for resources can collectively or individually influence home range sizes of wolverines (Magoun 1985, p. 63; Inman *et al*. 2012a, p. 785), which affects wolverine densities and population structure. Home range sizes of male wolverines are likely influenced by the density and reproductive condition of female wolverines (Magoun 1985, p. 63).

Based on telemetry studies, wolverines have been observed to disperse over very long distances. Both male and females can move long distances (Flagstad *et al*. 2004, pp. 684–686), but young (yearling) females tend to establish home ranges closer to their natal ranges than do young males (COSEWIC 2014, p. 24), which supports a male-biased dispersal pattern (from natal range) for wolverine populations. Vangen *et al*. (2001, p. 1,647) indicated that dispersal patterns of females were likely determined by competition for resources (that is, high quality territories) while male dispersal patterns were likely determined by competition for mates.

As noted above, wolverines readily cross water bodies such as rivers, and can cross rugged terrain (COSEWIC 2014, p. 24; Woodford 2014, entire). Dispersing wolverines in Idaho traveled over 200 km (124 mi) following routes across isolated subalpine habitat (Copeland 1996, p. 130). Inman *et al*. (2012a, p. 784) recorded dispersal-related movements of wolverines in the Greater Yellowstone Ecosystem and found that the maximum dispersal distance of subadults from the home range of their mothers was 170 km (106 mi) for males and 173 km (108 mi) for females, with an average maximum distance per dispersal movement of 102 km (63 mi) for males and 57 km (35 mi) for females (Inman *et al*. 2012a, p. 784). In the Ontario, Canada, region a juvenile male reportedly dispersed 100 km (62 mi) (COSEWIC 2014, p. 24, citing unpublished data from Dawson *et al*. 2013).

Several examples illustrate the extensive dispersal capability of wolverines. A male wolverine apparently dispersed (2008 or earlier) from the western edge of the Rocky Mountain region to the Sierra Nevada region of California (Moriarty *et al*. 2009, p. 160). Another male wolverine (designated as M56), whose natal area was the Greater Yellowstone Ecosystem (northwest Wyoming), moved south to Colorado (about 500 miles), where it remained for about 3 years (2009–2012), when its tracking signal was lost (Packila *et al*. 2017, p. 402). In April 2016, M56 was legally shot and killed by a rancher in western North Dakota, or about 1,288 km (800 mi) from its last known location (Packila *et al*. 2017, p. 404). In addition, a female wolverine was observed over several years (2004–2010) in the lower peninsula of Michigan, and genetic testing after her death in 2010 suggested she was more closely associated with eastern Canada wolverine populations (i.e., Manitoba and Ontario) (*in litt* Zigouris 2013, pers. comm.). It's unclear how this individual came to occupy this region, but given the long distance movements reported for this species, dispersal from Canada is plausible. Wilson (1982, p. 650) reported that wolverines on occasion may enter Minnesota from Canada. Jackson (1961, pp. 359–360) also reported several authentic records of wolverine in Wisconsin and in areas in Minnesota, along the

Wisconsin-Minnesota border. However, the wolverine was likely never abundant in Wisconsin, even before widespread and unregulated (Anglo) hunting in the late 19th and early 20th centuries (Jackson 1961, p. 359).

Additional discussion of population distribution and density estimates is provided below (see Biological Status–Current Conditions).

Reproduction and Growth

Wolverine reproduction includes the following characteristics: polygamous behavior (i.e., a male mates with more than one female each year), delayed implantation (up to 6 months), short gestation period (30–40 days), denning behavior, and an extended period of maternal care (Rausch and Pearson 1972, pp. 255–256; Pasitschniak-Arts and Larivière 1995, p. 5; Magoun and Copeland 1998, pp. 1,315–1,316; Hedmark *et al*. 2007, p. 19; Persson *et al*. 2017, *in prep*).

Table 2 below presents a summary of wolverine reproductive chronology (extent and peak of reproductive events) based on a review of the literature and personal knowledge from field studies (Inman *et al*. 2012b, entire), and studies from Scandinavia (Aronsson 2017; Persson *et al*. 2017, *in prep*).

Table 2. Chronology of wolverine reproductive events (adapted from Inman *et al*. 2012b).

Reproductive Biology Event	Time Interval
Mating Season	May – August; *peak in June*
Nidation (implantation of embryo)	November – March; *peak in late December–early February*
Gestation	November – April; *peak in January–mid-March*
Parturition (birth of young)	late January – mid-April; *peak in February–mid-March* (Sweden: *peak in mid-February*, range from end of January to early March)[a]
Reproductive Den Use	late January – end of June; most commonly, *early February–mid-May*
Lactation	About 10 weeks; generally February–June
Weaning	April – June; most commonly, *late April–May*
Rendezvous Sites[b]	April – June; *peak in early May*
Independence	August – January; *peak in September–December*
Dispersal	Peak period at 10–15 months of age; February–mid-April

[a] Persson *et al*. (2017, *in prep*); [b] Described below.

Wolverine mating is generally assumed to occur in May, June, and July (Pulliainen 1968, p. 341; Rausch and Pearson 1972, p. 249). A review of both the literature and personal observations by Inman *et al*. (2012, p. 636) indicated that June represented the peak in a wolverine mating season, but began in at least May and extended into early August. Female wolverines have been reported as not breeding in their first summer (under 1 year of age) based on examination of reproductive tracts from wolverine carcasses obtained from trappers (Yukon) (Banci and Harestad 1988, p. 268) and ages of pregnant female wolverines were estimated at 1 to 11-plus years of age (Banci and Harestad 1988, p. 266). In another study of wolverine carcasses (also in Yukon), some female wolverines were said to be mature at about 1 year (about 15 months), but first litters were not produced until 2 years of age (Rausch and Pearson 1972, p. 253). In

Scandinavia, the mean age of first reproduction for female wolverines was 3.4 years, based on monitoring of telemetered animals (Persson *et al*. 2006, p. 76). Breeding ages were reported at 2 to 13 years of age for wolverines in Sweden (mean age of first birth was 3.4, range of 2 to 5 years), based on monitoring/observations of female wolverines (Rauset *et al*. 2015, p. 3,157).

A genetic-based wolverine study in Scandinavia found that "females often reproduced with the same male in subsequent breeding years" (Hedmark *et al*. 2007, p. 18). However, this study also found (with some assumptions regarding sampling and paternity) that 8 of 13 female wolverines bred with different males, and, based on telemetry results, 2 females bred with a new male even though their previous breeding partner was still alive (Hedmark *et al*. 2007, p. 18). This shift in partners may have resulted from a change in the resident male wolverine in the area (Hedmark *et al*. 2007, p. 19).

The reproductive rate of wolverines is relatively low. An early study of 31 wolverine dens in Finland, as reported by hunters, found an average of 2 young per den (range 1–4) (Pulliainen 1968, pp. 338–341). Average litter size for northern Europe (161 litters) was 2.5 (range 1–4) (Pulliainen 1968, p. 343). In Alaska, average litter size was reported as 1.75 young, with a reproductive rate of 0.69 young per adult female per year (Magoun 1985, p. 28). A summary of average litter size for earlier studies of New World and Old World wolverines, based on method of determination, was presented in Magoun (1985, p. 29), indicating a range of 2.2 to 3.5. Anderson and Aune (2008, entire) evaluated pregnancy rates based on presence of corpora lutea (CL) and fetuses in trapper-harvested wolverines from western Montana. That study found median CL counts for pregnant adults ranging from 1.6 to 3.0, depending on the subpopulation (Anderson and Aune 2008, p. 22), with a mean litter size based on number of fetuses for pregnant adult females of 2.6 (Anderson and Aune 2008, p. 23). Studies of telemetered female wolverine in Scandinavia, from 1993 to 2002, reported a mean litter size of 1.88, with a range of 1 to 4 young, with a mean annual birth rate of 0.74 young per female (Persson *et al*. 2006, pp. 76–77). More recently, the average number of young per female per year reported for wolverines in Sweden was 0.84 (range 0–3); however, for those animals with recorded denning behavior, this value increased to 1.38 (range 0–3) (Rauset *et al*. 2015, p. 3,157).

Results from studies of telemetered female wolverines indicate that studies of wolverine reproductive tracts are likely to overestimate wolverine productivity (Persson *et al*. 2006, p. 77). Their findings suggest that young are either lost during pregnancy, early after parturition, or both; the losses likely do not occur before implantation (Persson 2006, p. 77). However, the factors that contribute to the observations that female wolverines do not give birth during some years are not well understood, and could be due to failure to breed, pseudo-pregnancy (as demonstrated by Mead *et al*. 1993, entire), failure of a fetus to implant, absorption of implanted fetus, stillbirth, or mortality before emerging from den (e.g. infanticide, etc.) (Magoun 2013, pers. comm.).

Carnivorous mammals generally have altricial young (poorly developed and dependent young (Derrickson 1992, p. 58)), and prepare shelter in dens where the mother can feed their young and keep them warm (Irving 1972, p. 174). Young wolverines (kits or cubs) weigh about 0.1 kg (3.5 oz) at birth and are blind until about 4 weeks of age (Krott 1960, p. 23). Newborns are covered with whitish to yellow hair (Krott 1958, p. 87; Mehrer 1976, p. 570), 4.5 millimeters (mm) (0.18

in) in length (Shilo and Tamarovskaya 1981, p. 147), with unerupted teeth (Mehrer 1976, p. 570; Pasitschniak-Arts and Larivière 1995, p. 5) and closed ear canals (Shilo and Tamarovskaya 1981, p. 147).

Mustelids, in general, have a short period of growth (Iversen 1972b, p. 317). As noted above, the metabolism of young wolverines is highest during the first 2.5 months, and individuals are almost two-thirds grown by the fall (at about 6 months) (Krott 1960, p. 25). As described by Shilo and Tamarovskaya (1981, p. 146), 45 to 50 day old wolverines (Norway) have woolly coats, are muddy grey in color, with teeth beginning to erupt at this age. At about 150 days, all permanent teeth have been established (Shilo and Tamaovskaya 1981, p. 147). After 2.5 months, young wolverines replace their juvenile coat with the adult summer coat (Shilo and Tamarovskaya 1981, p. 147). With growth ending at about 8 months (Iversen 1972b, p. 320; Magoun 1985, p. 23), wolverines are generally full grown by October or November.

Use of Dens and Denning Behavior

Dens and breeding burrows of animals are, in general, carefully constructed, well-camouflaged, and located in areas not easily accessible (Novikov 1962, p. 25). Wolverines use both natal dens (used for birthing) and maternal dens (used subsequent to natal den and before weaning) for rearing young (Magoun and Copeland 1998, p. 1,314). They are generally not left alone at the den during the first 3 to 4 weeks (Krott 1958, pp. 88, 108). The young remain at the natal den site for 6 to 8 weeks (Krott 1960, p. 24), and are weaned at 9 to 10 weeks (Copeland 1996, p. iv (Central Idaho); Koskela *et al*. 2013a, p. 101 (Finland)) (*cf*. 7 to 8 weeks reported by Myhre and Myrberget, 1975, p. 754 (Norway)). After weaning, the young are dependent on the mother and begin to travel with her by late April (Koskela *et al*. 2013a, p. 101 (Finland)). Observations of wolverines in central Idaho reported that females traveled up to 17.9 km (11 mi) from maternal dens to forage (Copeland 1996, p. 97). A study of telemetered wolverines in Scandinavia found that, on average, a female wolverine spends most of her time within 1,000 m (3,281 ft) of the reproductive den during the denning period (Myhr 2017, no page number). The average relocation distance to maternal den sites for active wolverine den sites studied in Norway was 268 m (879 ft) (95% confidence interval: 40–497 m (131–1,631 ft)) (May *et al*. 2012, p. 199).

The exact timing of when females abandon natal dens and begin using maternal dens is difficult to establish (Inman *et al*. 2012b, p. 638). In general, studies have found that den abandonment (natal) occurs before May (Magoun and Copeland 1998, p. 1,315; Table 1; Inman *et al*. 2012b, p. 637; Figure 2). A study by Aubry *et al*. (2016, p. 24) reported that a female wolverine moved her single young (estimated to be at least 9 weeks old) from a natal den in late April in the North Cascades region of Washington. More recently, a comprehensive study of wolverines in Scandinavia found that females begin to shift den locations more frequently beginning in late April, as young are more mobile and are more reliant on solid food brought to them by the mother (Aronsson 2017, p. 46). Natal den abandonment in Alaska and Idaho reportedly "coincided with a period when maximum daily temperatures rose above freezing for a number of days for the first time since denning commenced" (Magoun and Copeland 1998, p. 1,316). Factors other than temperature can influence shifts in the locations of these dens, including intraspecific predation, parasites, or other disturbances (Inman *et al*. 2012b, p. 638). In central

Idaho, Copeland (1996, p. iv) concluded that human disturbance at maternal den sites resulted in den abandonment, but not abandonment of young.

Rendezvous sites are locations in which the female leaves young while she hunts for food, and from which they will not leave without her (Magoun 1985, pp. 16, 77). These areas provide security to young (Copeland 1996, p. 94) and serve as locations at which females bring food to the young, or from which she will guide them to a food source (Inman *et al*. 2012b, p. 638). Rendezvous sites of wolverines studied in central Idaho consisted of large boulder talus or riparian areas associated with mature overstory and dense timber deadfall (Copeland 1996, p. iv). Magoun (1985, p. 76) reported that rock caves and hilltops containing boulders without large snowdrifts were used as rendezvous sites in Alaska. Females may move their young to new rendezvous sites several times over a two month period (Magoun 1985, p. 73), and distances between consecutives sites have been reported as far away as 8.5 km (5.3 mi) (Magoun 1985, p. 76).

Studies of adult female wolverines in Scandinavia (northern Sweden) have provided additional details regarding the temporal patterns of reproductive behavior and den site use. Aronsson (2017, p. 45) (see also Persson *et al*. 2017, *in prep*) found that, in general, most births occurred in mid-February. Females spend very little time outside the natal den for the first 2 weeks (Aronsson 2017, p. 45). During the first period of den site use, or approximately 2 to 2.5 months from mid-February (when females generally give birth and are lactating), females will move short distances and do not need to bring food to young (Aronsson 2017, p. 46). This time period generally coincides with snow cover and favorable conditions for food caching, and dens offer protection from predators and the environment (Aronsson 2017, p. 46). In addition, during the first 1.5 months of the denning period, females rarely changed den sites, but begin to move outside the den in early March (Aronsson 2017, p. 45). In the later denning period (after April 15), females begin to move more frequently and at greater distances between den sites (Aronsson 2017, p. 45). By late April, the young are more active and also begin to rely more on solid food that is brought back to them by their mother (Aronsson 2017, p. 46). This also corresponds to a time period when prey are more available (e.g., reindeer migration and calving period in Sweden and emergence of hibernating rodents in the spring (Yates and Copeland 2017, *in prep*)) and expected shorter distance movements by the mother back to denning or rendezvous sites (Aronsson 2017, p. 46). These observations are consistent with Inman *et al*.'s (2012b, entire) proposed cold, low productivity niche for wolverines based on studies of wolverines in the Greater Yellowstone Ecosystem. That is, reproductive chronology in wolverines is considered to be adapted to take advantage of the availability of food resources, limited interspecific competition, and snow cover in the winter (Inman *et al*. 2012b, p. 635).

In summary, as described by Inman *et al*. (2012b, entire) and Persson *et al*. (2017, *in prep*), reproductive behavior of wolverines reflect seasonal shifts in resource abundance within the wolverine's range; that is, adaptation that matches the time of birth and development of young to changes in the availability of resources and foraging strategies (Persson *et al*. 2017, *in prep*). We present in Figure 5 a visual summary of wolverine feeding strategies relative to resource availability from time of birth to post-weaning.

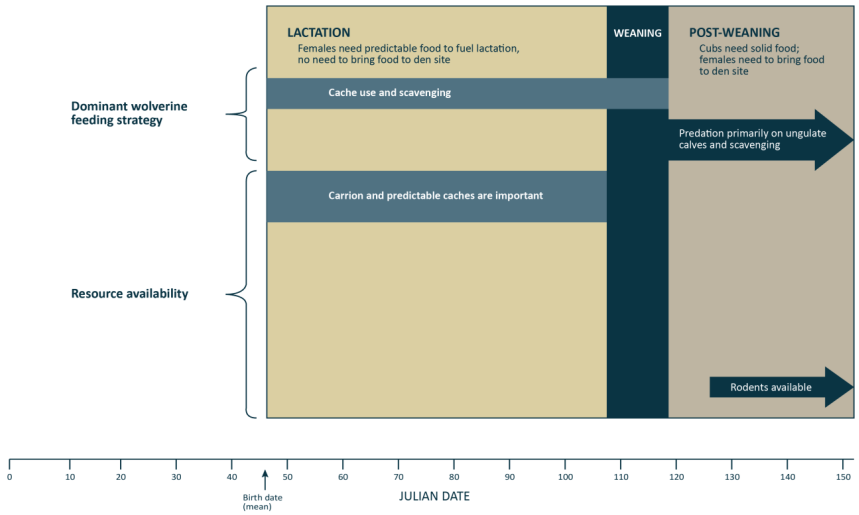


Figure 5. Wolverine feeding strategies relative to resource availability. Adapted from Persson *et al*. 2017, *in prep*.

Denning Habitat

Given the wolverine's observed association with snow, we provide in **Box 1** a summary of the importance of snow, for ecological systems, and as an introduction to this section of the SSA Report. This summary provides a detailed perspective of how various physical properties of snow can influence ecological systems occupied by snow-adapted wildlife, including insulating properties, differences in snow cover in mountainous vs. forested habitat, and changes in snow cover due to wind and slope/aspect.

Wolverine denning habitat varies over its Holarctic range and is dependent on regional environmental conditions. For example, in southcentral Norway, wolverine dens were identified as snow tunnels dug into deep snow at the tree line (elevation 1,100 m (3,609 ft)), but most of the tunnel systems extended down to boulder fields, talus slopes, or rock crevices such that young could crawl around within these structures (May *et al*. 2012, p. 201). Snow tunnels are also reported for wolverine natal dens in Alaska (Magoun 1985, pp. 84, 185, 190). However, reproductive dens are not always excavated in deep snow. In Canada, female wolverines are said to give birth in dens where snow cover persists at least until April, and can den under snow-covered rocks, logs, or within snow tunnels (COSEWIC 2014, p. v). As an example, in northwestern Ontario, den site habitat for a female in lower Boreal Forest habitat (elevation 250 to 500 m (820 to 1,640 ft), 51°N) included large boulders and downed trees, similar to dens described for wolverines in montane ecosystems (Dawson *et al*. 2010, p. 139).

Box 1. Snow Cover in an Ecological Context

Formozov (1961; 1963) prepared comprehensive reviews of the unique properties of snow in the context of its role in the ecology of animals and plants in Russia. In his 1963 review (translated from the 1946 original), he identified two important factors attributed to snow cover — nastization (the thickness of the crust on the surface of mature snow cover) and firnization (process of snow compaction) — relative to its ecological influence (Formozov 1963, p. 8). Snow cover provides not only a substrate that allows some animals to move across the landscape, it also provides a matrix within which other animals can create tunnels and build nests (Formozov 1963, p. 8). Additional fundamental concepts described in this study are provided below:

- Snow has very low thermal conductivity which promotes cooling at the surface while at the same time protects the deeper layers from chilling; but this property varies by region, by depth, by season, and by year (e.g., the more continuous the snow cover during winter, the greater the warming effect). As snow changes to ice (through compaction and melting), the thermal conductivity decreases (Formozov 1963, pp. 7, 8, 108)
- Snow therefore creates a thermo-insulating layer, which allows for a unique temperature regime on the surface and underneath; as an example, soil temperatures measured in January (near Saint Petersburg, Russia) averaged 15°C higher with snow cover than without snow cover, with up to a 32°C difference, depending on the day and depth measured (Formozov 1963, p. 109)
    - At 20–25 cm snow cover, the insulative effect is particularly important for winter nesting of small mammals (Formozov 1963, p. 110)
- Snow cover in mountains:
    - Depth of snow cover and its duration increases with elevation; even minor elevation differences are noticeable (Formozov 1963, p. 123)
    - This spotty distribution is also affected by unequal distribution of snow precipitation on slopes with different exposures, transport of snow by wind, melting of snow on sun-exposed slopes, avalanche or rolling down of snow from steeper areas, and vegetation (Formozov 1963, p. 123)
    - Snow cover areas near Arctic limits and at treeline in mountain regions is more strongly influenced by wind, which compacts and re-works snow cover (Formozov 1963, p. 29)
- Snow cover in forests:
    - The maximum depth, density, duration and date of melting, thickness of snow surface crust are all very different in forested areas as compared to open treeless areas (Formozov 1963, p. 19)
    - Snow accumulates slowly under trees and is generally thicker the further away from the forest than within the forest; thus, the compaction and settling of snow under a forest canopy is less than tundra or open fields (with a less icy crust), so for some vertebrates, forested areas can provide a more preferable place to winter or migrate (Formozov 1963, pp. 24, 26)
    - Snow cover in forested areas also melts slower than open fields and clearings (Formozov 1963, p. 28)
- Snow cover also plays an important role in the overwintering conditions for insect eggs, caterpillars, pupae, and adult insects in litter and soil (Formozov 1963, p. 121)

Large diameter, downed trees can also provide structure for den sites. In Finland, Pulliainen (1968, p. 340) reported a den site (January) at the base of a tree and not covered in snow, and also described other structural features such as rocks, fallen trees, and deep ravines as denning habitat (likely both natal and maternal dens) (Pullianinen 1968, pp. 338–341). In Russia, where wolverine habitat has been described as located far from human-inhabited areas within boreal forests and, to some extent, tundra, and taiga (Novikov 1962, pp. 199, 200), den locations were described as "clefts in rocks, among stones, and under roots of upturned trees" (Novikov 1962, p. 200). A study from northwestern Ontario noted that, because lowland boreal forest habitat in this region does not support deep, wind-hardened snowdrifts, other structural elements within snow layers such as trees and boulders can be important components of wolverine denning habitat (Dawson *et al*. 2010, p. 142).

Although it has been assumed that wolverines have an obligate relationship with snow for natal denning, including persistent spring snow cover, the key elements or combination of elements that define this relationship have not been empirically analyzed. Limited studies to date have evaluated the importance of denning habitat to reproductive success, or the key physiological and ecological characteristics, including avoidance and/or protection from predators, prey availability, availability of caching habitat, that define denning behavior and den site selection. Population density, trapping pressure, population genetics, and other measures of habitat quality may also influence wolverine fecundity (Anderson and Aune 2008, p. 28). Habitat features that facilitate caching (e.g., restricted access and cold temperatures) are likely to be important for wolverine reproduction (Inman *et al*. 2012b, p. 640). Studies of wolverine denning activity have not reported the condition of the natal or maternal den location following abandonment; that is, what is the persistence and/or depth of snow at the natal den at the end of the denning season, and how does this affect survival of young? In addition, previous accounts of den site locations for North America have been biased to tundra regions where dens are more readily observed and located (Banci 1994, p. 110). In Scandinavia, snow cover also had been found to be a poor technique for tracking female wolverines during the time when they give birth and initiate denning (Aronsson and Persson 2016, p. 266).

As noted above, adult wolverines have a wide range of thermoneutrality. However, newborns, who are born with lighter, less dense fur are likely to have a more limited ability to control their internal temperature, though huddling (a thermotactic behavior) of small mammals in dens can conserve heat (Barnett and Mount 1967, p. 439). Relatedly, basal metabolic production of heat, the source of heat that maintains bodily warmth, is not easily modifiable unlike the flexibility of insulation (Irving 1972, p. 121). Metabolic heat above an animal's basal rate for preservation of warmth is restricted by its limited capacity for metabolic production of heat, but also by food availability and the time and opportunity for nourishment (Irving 1972, p. 121). In general, metabolic production of heat is costly to animals, but variable insulation represents a conservative strategy (Irving 1972, p. 121).

A bioclimatic model was used by Copeland *et al*. (2010, p. 234) to test the following hypothesis: "…wolverine distribution at the broadest spatial scale is constrained within a climatic envelope defined by an obligate association with persistent spring snow cover and by an upper limit of thermoneutrality." This hypothesis was based on the premise "If persistence of wolverine populations is linked to the availability of suitable reproductive den sites ([citing] Banci 1994),

snow cover that persists throughout the denning period may be a critical habitat component that limits the wolverine's geographic distribution" (Copeland *et al*. 2010, p. 234). The authors tested their hypothesis by "comparing and correlating the locations of wolverine reproductive dens from throughout their circumboreal range, and telemetry locations from 10 recent wolverine studies in western North America and Scandinavia, with spatial models representing the distribution of spring snow cover and average maximum August temperatures" (Copeland *et al*. 2010, p. 234).

Bioclimatic models "use associations between aspects of climate and known occurrences of species across landscapes of interest to define sets of conditions under which species are likely to maintain viable populations" (Araújo and Peterson 2012, p. 1,527). They are correlational by nature and are often applied to study a variety of conservation issues, including forecasting potential climate change effects on species' distributions (Araújo and Peterson 2012, p. 1,527). These types of correlational models have received some criticisms and require careful framing to avoid misapplication (Sieck *et al*. 2011, p. 6; review by Araújo and Peterson 2012, entire). They generally represent a first step for evaluating current and future species distributions, and, when coupled with climate change scenarios, results are presented at a coarse scale that may not accurately project shifts in species distribution at a smaller scale (Sieck *et al*. 2011, p. 6). In particular, when used to estimate extinction risk, these types of models provide only an estimate of the empirical relationships between a species' current distribution and climate variables and then use inferred relationships to identify potential areas where the species is distributed under future climate scenarios (Araújo and Peterson 2012, p. 1,553). Extinction risk is not represented in the model's input data and therefore is not the targeted parameter of the model; thus, a bioclimatic model's usefulness may be limited in these types of applications given that it only offers partial explanatory evidence for reasons for potential extinction related to the shifts in climate suitability within the time frame being modeled (Araújo and Peterson 2012, p. 1,533 and citations therein). In addition, climate niche projections generally do not incorporate factors such as competition, dispersal, and evolutionary capacity, which also influence range boundaries (Michalak *et al*. 2017, p. 370). Thus, these types of models are more applicable at broad scales in which the effects of fine-scaled topography and biological interactions play a more limited role (Michalak *et al*. 2017, p. 370). However, for the wolverine, topography and biological interactions, such as predator avoidance, are important at the den-site scale.

As noted above, Copeland *et al*. (2010, entire), used a bioclimatic model to evaluate an assumed association not at the den site scale, but at a broad scale. The results presented in Copeland *et al*. (2010, entire) were based not on the condition of snow cover at a particular den site at the time of denning, but rather their evaluation of snow persistence (April 24 to May 15) was based on satellite images summed over a 7-year period (2000 to 2006) for the den locations. The spatial resolution of the snow measurement used to detect daily snow cover was 500 m by 500 m (1,640 ft by 1,640 ft), using Moderate-Resolution Imaging Spectroradiometer (MODIS). If persistent snow cover was observed in any one year, it was included in the bioclimatic model regardless of whether denning occurred during that particular year.

In addition, although the study found that 69 percent of dens for North American wolverines were located within satellite images (pixels) in areas that had snow cover for 6 or 7 years, out of 7 years, 31 percent of the identified den locations were located in areas that were identified as

having spring snow cover 5 years or less out of 7 years (from April 24 to May 15). However, snow depth was not defined in Copeland *et al*. (2010). Also, the den location attributes (e.g., den structure, how long it was used) were not recorded relative to the observed persistent snow cover and some of the 562 dens (e.g., Norway) were identified by snow tracking rather than direct observation.

In summary the results presented by Copeland *et al*. (2010, entire) provided a fairly accurate, assessment at that time of where wolverine populations are expected to be observed, but did not evaluate (model) snow persistence at the den site scale based on location and denning period.

Since the 2013 (78 FR 7864; February 4, 2013) proposed listing rule and 2014 (79 FR 47522; August 13, 2014) withdrawal of the proposed listing rule for the wolverine, several publications have presented additional study results related to wolverine distribution and snow cover. In Alberta, Canada, Webb *et al*. (2016, entire) found that, based on wolverine harvest data, wolverine occurrence relative to spring snow cover (percent of area covered, with greater than 75 percent snow coverage, on April 1 and 15) varied based on the different regions of Alberta. Reproductive den sites were not evaluated. This study used meterological data from 2009 to 2015, in northern Alberta (north of 54°N) and differentiated between spring snow cover that persisted in ≥ 1 of 7 years and more frequently (≥ 4 of 7 yr) (Webb *et al*. 2016, pp. 1,462–1,464). Although the study found an overall positive trend of more frequent wolverine harvests in those areas expected to have spring snow cover, the study did not find consistent large differences between these areas, and did not typically detect significant relationships with frequent spring snow cover (4–7 years) in all regions (Webb *et al*. 2016, p. 6). Additionally, the Rocky Mountains region was the only region in which wolverines were reported in areas with more frequent spring snow cover (4–7 years) (Webb *et al*. 2016, p. 5). This region, which is located along the western border of Alberta, contains montane, subalpine and alpine habitat, with elevations from 1,000 m (3,281 ft) to 3,700 m (12,139 ft) (Webb *et al*. 2016, p. 9). Conversely, the study found that in the Boreal Forest region of Alberta (i.e., wetland habitat interspersed with coniferous, mixed wood, and deciduous forests, with elevations between 1,500 m (4,921 ft) to 1,100 m (3,609 ft)), a female wolverine denned under large boulders and downed trees (Webb *et al*. 2016, p. 8).

Further, the study reported that all female wolverines that were positively identified from camera trapping (including five lactating females) were found in townships where no spring snow cover was predicted (Webb *et al*. 2016, p. 1,461). These areas are depicted in Figure 6 below. This study also located eight wolverine dens from two different females over two winter-spring seasons (Webb 2018b, pers. comm.). The general locations of these dens are in the vicinity of the lactating females circled on the map below.

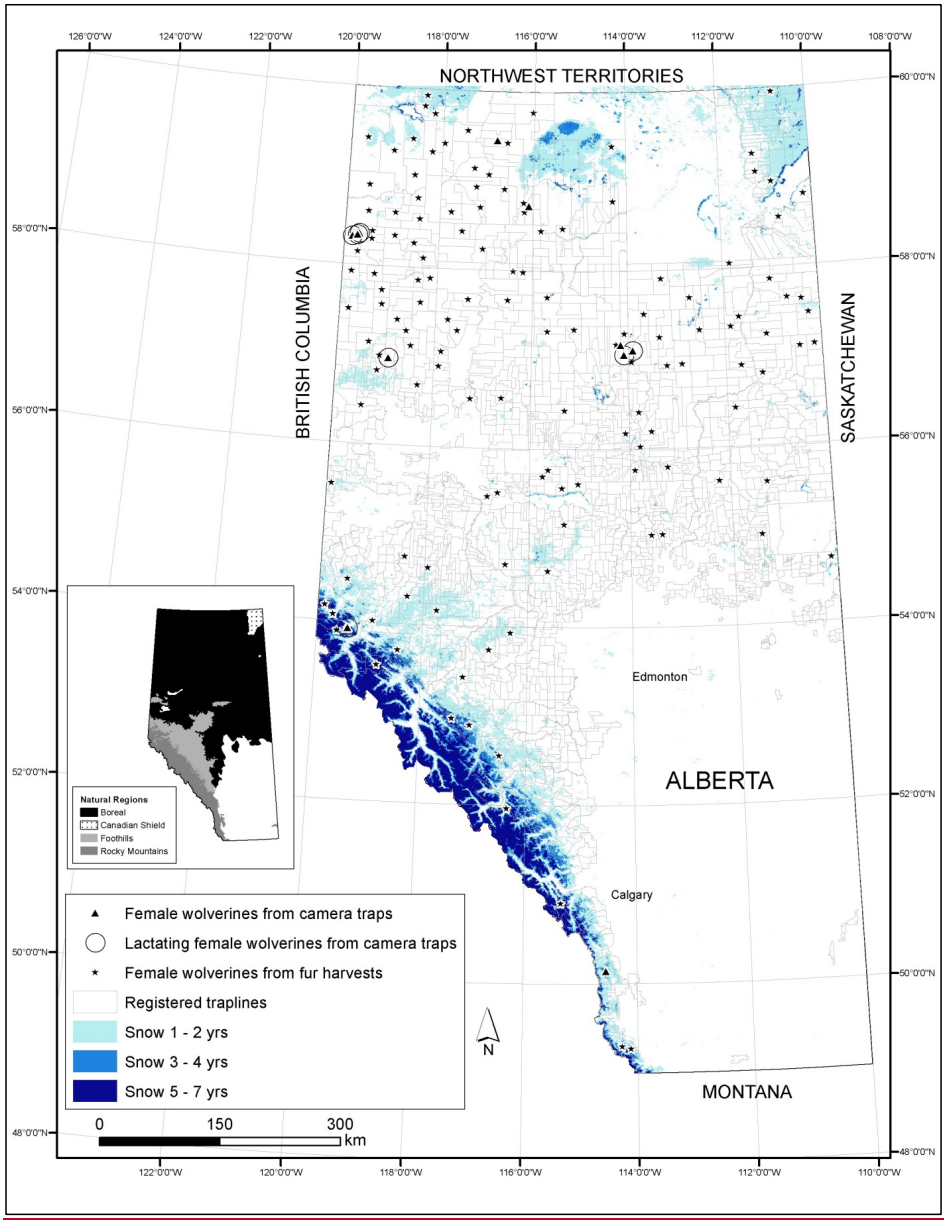


Figure 6. Female and lactacting female wolverine locations in Alberta, Canada, relative to modeled snow cover layer from Copeland *et al*. (2010). Source: Webb 2018b, pers. comm.

Results from another recent Alberta study (Heim *et al*. 2017, entire) evaluated cumulative effects of climate and landscape change, using species distribution models, relative to the spatial distribution of the wolverine. For landcover only models, the study found that six variables (e.g., dense conifer, shrub, herb, snow and ice cover) best explained wolverine frequency (Heim *et al*. 2017, p. 8,908), while for climatic-abiotic models, wolverines selected persistent spring snow cover (measured as years out of 12 that had spring snow between April 14 and May 15, using Copeland et al. 2010 method) (Heim *et al*. 2017, p. 8,908). However, their cumulative effects model better explained wolverine frequency than any single-factor model (Heim *et al*. 2017, p. 8,908; Table 3), which suggests that no single abiotic or biotic variable explains wolverine distribution (Heim *et al*. 2017, p. 8,908).

The study presented in Webb *et al*. noted that wolverine den locations within low elevation, forest habitats have not been well-described (Webb *et al*. 2016, p. 8). In boreal forested habitat of Russia, wolverines were reported denning in rock areas and in tree root structures (Novikov 1962, p. 200). A similar finding was reported in Sweden, where a majority of dens (n=49) were in boulder areas located within mature, mixed coniferous forests (i.e., not alpine or tundra habitat) (Makkonen 2015, p. 14). A recently published study reported two wolverine natal dens in logged areas (cutblocks) in northern Alberta, Canada; specifically, within a slash pile and log deck (Scrafford *et al*. 2017, p. 35).

A study of wolverine populations and distribution in Sweden observed that wolverine populations were found outside areas with persistent spring snow cover (mean snow depth and proportion of years with snow cover on March 15; 1961–1990) and expanding into boreal forest habitat located to the east and south of alpine areas (Aronsson and Persson 2016, p. 266). This southern and eastern expansion (from 1996 to 2014) indicates recolonization of their historical distribution in Sweden, and is thought to be the result of an increase in population, with more dispersers colonizing forest habitat, and an increase in year-round scavenging opportunities due to an increase in Scandinavian wolf packs (Aronsson and Persson 2016, p. 266; Aronsson 2017, pp. 43–44). As of the spring of 2017, over 80 reproductive dens have been observed outside the boundary of the snow model presented in Copeland *et al*. (2010) (Persson 2017, pers. comm.).

Another key element related to den location is the protection that dens provide to a nursing female and her young. Because wolverines are known to den in a variety of structures, it is unclear if the apparent relationship to snow cover is based on selecting den locations in relatively inaccessible, high elevation areas to avoid predators. Bare rock and boulders at den sites can offer dry and secure cavities and enhance the ruggedness of the landscape (May *et al*. 2012, p. 198). "Ruggedness," a measure derived from elevational changes and irregularity of land surface (density of contour lines) traversing a given area (Beasom *et al*. 1983, p. 1,163) has been found to be an important variable (i.e., secure habitat from predation risk) for female wolverines in winter (British Columbia, Canada) (Krebs *et al*. 2007, p. 2,188) and for den site selection at site-specific, home range, and landscape scales (southcentral Norway) (May *et al*. 2012, pp. 200–201).

Snow depth can be affected at a local level by terrain, ruggedness, slope and aspect as well as wind and vegetation cover. For example, slope and aspect together will affect the exposure to snow accumulation (May *et al*. 2012, p. 198). In an effort to document and compare snow persistence at the wolverine den-site scale, Magoun *et al*. (2017, entire) evaluated the use of low-altitude aerial photography during late May 2016 in areas within the Rocky Mountains (Idaho

and Montana) and northwestern Alaska. In Idaho and Montana, flight lines were established along transects through the long axis of previously documented home ranges of denning female wolverines and, in Alaska, known den sites (from 2016) were visited by helicopter and remaining snow was photographed (Magoun *et al*. 2017, p. 383). Transect segments in the Rocky Mountain study areas documented snow on May 31 in all but one segment, with 82 percent classified in low to heavy snow retention categories, and 58 percent considered as moderate to heavy (Magoun *et al*. 2017, p. 383). In the Alaska study area, photographs documented widely scattered patches of snow on May 29, with remnant snowdrifts observed at all four wolverine den sites (Magoun *et al*. 2017, p. 383). The documentation presented in this study of scattered patches of snow in the Rocky Mountains persisting into late May in areas previously detected to be bare of snow on May 29 (MODIS persistent spring snow cover, McKelvey *et al*. 2011, p. 2,889, Figure 4D; Magoun *et al*. 2017, p. 384, Figures 2b and 2d) suggests that persistent spring cover may not always be detectable at the den-site scale using remote sensing methods (Magoun *et al*. 2017, p. 384), such as those used in Copleand *et al*. (2010).

To evaluate, at a finer scale, snow cover at previously recorded den site locations in the western United States, we reviewed natal, maternal, and known den sites relative to derived 'melt-out' dates using MODIS/Terra Snow Cover, 8-day series (Hall and Riggs 2016). Melt-out dates represent the first day of the 8-day composite series when the cell in which the den was located switches from "snow" to "no snow." Snow cover was identified using the Normalized Difference Snow Index (NDSI), on a scale of 0 to 1.0, where NDSI greater than 0.1 is considered "snow." The spatial resolution for these data is 500 m by 500 m (1,640 ft by 1,640 ft). Because MODIS data was only available from the years 2000 to present, we were only able to evaluate 34 of the 47 den sites documented in our records. As shown in Table 3, the earliest melt-out date was May 14 (2006) and the latest was July 12 (2002).

For *natal* den sites only, the range for melt-out dates was May 25 to July 12. All of these sites indicate a melt-out date that is past the May 15 date used for the persistent spring snow cover model presented in Copeland *et al*. (2010). The melt-out dates provided here indicate that snow is persistent at these locations past the time when young begin moving out of natal dens (i.e., late April; see Use of Dens and Denning Behavior section). Because we were interested in analyzing the presence of "significant snow," which we defined as greater than or equal to 0.5 m (20 in) of snow, we also evaluated wolverine denning sites in Glacier National Park relative to snow depth. As shown in the snow depth columns in Table 3, with the exception of one den site that had melted out by May 15, all of the other den sites have snow cover between 0.4 m (15.8 in) and 1.4 m (55 in or 4.6 ft).

Other physical or biotic variables are also likely to be important for wolverine den site locations. Elevation affects snow depth and persistence at the landscape scale (May *et al*. 2012, p. 198). Inman *et al*. (2012a, p. 782) found that wolverines (12 females and 6 males) monitored in the Greater Yellowstone Ecosystem selected, on an annual basis, areas above 2,600 m (8,530 ft) latitude-adjusted elevation. In central Idaho, natal dens were also found in secluded, high elevation (above 2,500 m (8,202 ft)) cirque basins (Copeland 1996, p. 94).

Table 3. Wolverine Den Site Melt-Out Dates, 2002–2015.

Den #	Den Type	Melt-out Date	Elevation, meters (feet)	April 15 snow depth*, meters (feet), (Glacier National Park only)	May 1 snow depth*, meters (feet), (Glacier National Park only)	May 15 snow depth*, meters (feet), (Glacier National Park only)	Structure	State
1	Unknown	7/12/2002	1,814 m (5,951 ft)	n/a	n/a	n/a	None Listed	WA
2	**Natal**	5/25/2003	1,928 m (6,326 ft)	1.32 m (4.33 ft)	1.07 m (3.51 ft)	1.04 m (3.41 ft)	Log Complex	MT
3	Maternal	5/25/2003	1,995 m (6,545 ft)	1.32 m (4.33 ft)	1.07 m (3.51 ft)	1.05 m (3.44 ft)	Log Complex	MT
4	**Natal**	6/4/2004	1,807 m (5,923 ft)	1.96 m (6.43 ft)	1.46 m (4.79 ft)	1.13 m (3.71 ft)	Log Complex	MT
5	**Natal**	6/9/2004	2,399 m (7,871 ft)	n/a	n/a	n/a	None Listed	WY
6	**Natal**	6/17/2004	2,487 m (8,160 ft)	n/a	n/a	n/a	None Listed	MT
7	Maternal	6/29/2004	1,823 m (5,981 ft)	1.0 m (3.28 ft)	0.75 m (2.46 ft)	0.54 m (1.77 ft)	Downed Log	MT
8	Maternal	6/29/2004	1,893 m (6,211 ft)	1.07 m (3.51 ft)	0.83 m (2.72 ft)	0.65 m (2.13 ft)	Log/Boulder	MT
9	Maternal	6/11/2005	1,912 m (6,273 ft)	1.6 m (5.25 ft)	1.11 m (3.64 ft)	0.58 m (1.90 ft)	Spider Tree	MT
10	Maternal	6/11/2005	1,973 m (6,473 ft)	1.0 m (3.28 ft)	0.76 m (2.49 ft)	0.47 m (1.54 ft)	Spider Tree	MT
11	**Natal**	6/11/2005	1,977 m (6,486 ft)	1.6 m (5.25 ft)	1.11 m (3.64 ft)	0.58 m (1.90 ft)	Spider Tree	MT
12	**Natal**	7/12/2005	2,693 m (8,835 ft)	n/a	n/a	n/a	None Listed	MT
13	Unknown	5/14/2006	1,514 m (4,967 ft)	0.68 m (2.23 ft)	0.26 m (0.85 ft)	0	Log Complex	MT
14	Unknown	5/25/2006	2,093 m (6,867 ft)	3.05 m (10.00 ft)	2.56 m (8.40 ft)	2.44 m (8.00 ft)	None Listed	MT
15	Maternal	5/31/2006	1,851 m (6,073 ft)	1.14 m (3.74 ft)	0.79 m (2.59 ft)	0.61 m (2.00 ft)	Log Complex	MT
16	**Natal**	5/31/2006	1,843 m (6,047 ft)	1.14 m (3.74 ft)	0.79 m (2.59 ft)	0.61 m (2.00 ft)	Log Complex	MT
17	Unknown	6/7/2006	2,252 m (7,389 ft)	2.83 m (9.28 ft)	2.4 m (7.87 ft)	2.38 m (7.81 ft)	None Listed	MT
18	**Natal**	6/18/2006	2,695 m (8,842 ft)	n/a	n/a	n/a	None Listed	MT
19	**Natal**	5/25/2007	2,820 m (9,252 ft)	n/a	n/a	n/a	None Listed	MT
20	**Natal**	6/4/2007	1,922 m (6,306 ft)	1.82 m (5.97 ft)	1.28 m (4.20 ft)	0.68 m (2.23 ft)	Log/Boulder	MT
21	Unknown	7/3/2008	2,505 m (8,219 ft)	n/a	n/a	n/a	None Listed	ID
22	**Natal**	6/26/2010	2,276 m (7,466 ft)	n/a	n/a	n/a	Boulder/Downed Log	ID
23	**Natal**	6/18/2010	2,216 m (7,270 ft)	n/a	n/a	n/a	Boulder	ID

24	**Natal**	6/26/2010	2,223 m (7,293 ft)	n/a	n/a	n/a	Boulder/Talus Field	ID
25	Unknown	6/26/2010	2,402 m (7,882 ft)	n/a	n/a	n/a	Boulder	ID
26	**Natal**	7/4/2011	2,167 m (7,110 ft)	n/a	n/a	n/a	Tree Complex	ID
27	Unknown	7/4/2011	2,130 m (6,989 ft)	n/a	n/a	n/a	Tree Complex	ID
28	Unknown	7/4/2011	2,147 m (7,043 ft)	n/a	n/a	n/a	Tree Complex	ID
29	**Natal**	6/9/2012	2,311 m (7,582 ft)	n/a	n/a	n/a	Log Complex	ID
30	Unknown	6/25/2012	2,318 m (7,604 ft)	n/a	n/a	n/a	Downed Log Complex	ID
31	Unknown	6/9/2012	2,350 m (7,711 ft)	n/a	n/a	n/a	Downed Log Complex	ID
32	**Natal**	6/10/2013	2,911 m (9,552 ft)	n/a	n/a	n/a	Boulder Field	ID
33	**Natal**	6/26/2014	2,218 m (7,278 ft)	n/a	n/a	n/a	Boulder Field	ID
34	**Natal**	6/10/2015	2,713 m (8,901 ft)	n/a	n/a	n/a	Boulder Field	MT

*Snow depth was derived from Distributed Hydrology Soil Vegetation Model (DHSVM) snow water equivalent results; see Ray *et al*. (2017, p. 32) for details.

We evaluated 47 den sites in the lower United States using a linear regression model to evaluate whether the elevation of wolverine den sites is related to latitude. We note here that not all of these dens were characterized as to whether they were natal or maternal dens, a few records were not verified through tracking of females or direct observations, and these den locations may not be representative of all current or potential wolverine den sites in the western-northwestern United States. Given these caveats, our examination of these records indicated that, in general, wolverine dens at lower latitudes (36 to 38°N) occur at higher elevations (range: 2,688 to 3,562 m) (8,819 to 11,686 ft) while the converse is seen for those dens at higher latitudes, or approximately 44 to 49°N (range: 1,514 to 2,911 m) (4,967 to 9,552 ft). Given our assumptions (small sample size, test of normality), we used linear regression (R Software; R Development Core Team, 2014) to test this association. We found a moderately significant association with elevation and latitude [adjusted $R^2 = 68.56$, $F = 101.3$, $df = 45$; *p-value* = 4.23×10^{-13}], such that dens found at lower elevations were generally located at more northerly latitudes. These results suggest that other potential explanatory variables (e.g., topographic structure) or interactions between variables (e.g., snow cover and topographic structure) for den site selection should be considered using multiple regression techniques.

The steep slopes found at higher elevations also provide conditions conducive to avalanches, which result in debris and talus/boulder piles that provide structure for dens (Inman 2013, pers. comm.). Steep slopes and the availability of rocks were found to be important to wolverine den site selection for wolverines studied in Norway (May *et al*. 2012, p. 200). These areas also offer either exclusive or higher frequencies of maternal food sources during the high energy demands for reproducing females, such as marmot emerging from hibernation and neonatal or avalanche-killed ungulates (Inman 2013, pers. comm.) (see *Diet and Feeding* discussion below).

In summary, wolverines select den sites for different characteristics depending on location. Dens located under snow cover may be related to wolverine distribution based on other life history traits, including morphological, demographic, and behavioral adaptations that allow them to successfully compete for food resources (Inman 2013, pers. comm.). Structure (e.g., uprooted trees, boulders and talus fields) appears to be important for natal den sites with or without snow cover. Sensitivity to human disturbance and predator avoidance are also likely important factors in selecting both natal and maternal den sites. However, reproductive success of wolverines has not been evaluated relative to the depth and persistence of snow cover, or in combination with these or other important characteristics, including prey availability and predator avoidance. Based on their study of wolverines in Alberta, Canada, Webb *et al*. (2016, p. 1,468) concluded that wolverines are adaptable and do not require large areas of deep spring snowpack for successful reproduction. They suggest that "wolverines may select small areas covered with deep snow at a finer scale than can be detected using satellite imagery, but the amount of snow needed over what portion of the denning duration remains uncertain" (Webb *et al*. 2016, p. 1,468, citations omitted). They add that temperature may play a more important role in determing the distribution of wolverines and restricting the wolverine's niche to cooler environments where wolverines have the competitive edge (citing Aubry *et al*. 2007, Copeland *et al*. 2010, Inman *et al*. 2012) (Webb *et al*. 2016, p. 1,468). Additional studies are needed to further document wolverine den structure, snow conditions at dens, and how long dens are used, particularly for those locations outside of areas expected to have spring snow cover, to better understand the relationship of wolverines and snow cover (Webb *et al*. 2016, p. 8; Magoun *et al*. 2017, pp. 6–7).

Demography

The lifespan of the wolverine is variable. Jackson (1961, p. 361) reported an upper range of 8–10 years in the wild and potentially up to 18 years in captivity. Based on trapper-submitted carcasses from the Yukon, an upper age of 11.9 years for a male wolverine and 12.9 years for (pregnant) female was reported (Jung and Kukka 2013, pp. 8, 12). Inman *et al*. (2012a, p. 781) classified wolverines less than 1 year old as juveniles (or kits/cubs), those 1 to 2 years old as subadults, and those at least 3 years old as adults. Generation time for wolverines has been estimated at 7.5 years (COSEWIC 2014, p. 23).

Survival of adult female wolverines is considered to be an important demographic parameter in the wolverine's life history (Sæther *et al*. 2005, entire). As noted by Aronsson (2017, p. 13), because most polygamous species display a dispersal pattern that is sex-based, their population distribution is generally limited by the dispersal behavior of the sex that is more philopatric (the tendency of a species to remain within or return to its birth area). Thus, the distribution of wolverine populations and colonization is generally limited by dispersal of female wolverines (Aronsson *et al*. 2017, p. 2).

Stochastic factors (both demographic and environmental) also strongly influence the population dynamics of the wolverine (Sæther *et al*. 2005, p. 1,011–1,012). Given the rapid maturity of young wolverines, survival of female wolverines with young is likely dependent on the availability and distribution of food sources during the "snow-free season" (late spring and summer) (Banci 1994, p. 114). For example, a study of wolverines in Norway found that survival of young was primarily influenced by the abundance of small rodents (Landa *et al*. 1997, p. 1,293).

Evaluating how variations in demographic rates are influenced by the interactions between costs of reproduction, individual quality (e.g., breeding status), and environmental factors can provide a better understanding of the dynamics and viability of animal populations (Robert *et al*. 2012; p. entire; Rauset *et al*. 2015, entire). The interactions between individual age, environmental resources, and reproductive costs of wolverines in Sweden were recently examined (Rauset *et al*. 2015, entire). The results of this study provide important details regarding the influences on wolverine reproduction productivity. The study found that age-related variation in reproductive output for female wolverines is driven by the interactions between age, reproductive costs, and availability of resources (Rauset *et al*. 2015, p. 3,160). As an example, female wolverines were found to be more likely to give birth and nurse young in home ranges with greater food resource abundance at the time of fetal development (Rauset *et al*. 2015, p. 3,158). The study also concluded that a favorable reproductive strategy for female wolverines is a conservative one, wherein older female wolverines do not "trade" current reproduction against their own survival (Rauset *et al*. 2015, p. 3,161).

Intraspecific predation of wolverines is another important influence on wolverine population dynamics (Persson *et al*. 2003, p. 26). The altricial life history stage (early May to end of July) is likely a period of high juvenile mortality in solitary carnivores, such as the wolverine, since females are balancing the energetic demands of lactation (Sadleir 1984, pp. 179–180) and providing protection to young (Persson *et al*. 2003, p. 22). Young (juveniles) wolverines are

vulnerable to predation during the time period when left unattended in the natal den (generally March-April) and when they first exit the natal or maternal den and are left at rendezvous sites, or around May-early July (Magoun 1985, pp. 49, 73, 77). An additional vulnerability occurs when juvenile wolverines are required to become nutritionally independent and begin exploratory movements away from their mother's protection, generally August-September (Vangen *et al*. 2001, p. 1,644).

Mortality

There are few natural predators of wolverines. Predation deaths from interactions with American black bears (*Ursus americanus*) have been documented (Inman *et al*. 2007, p. 89). Mountain lions are suspected of killing wolverines (Copeland 1996, p. 46; Krebs *et al*. 2004, p. 497; Aubry *et al*. 2016, pp. 27, 32). Interactions with wolves can lead to severe injury and death (Burkholder 1962, p. 264; Banci 1987, pp. 81, 91; White *et al*. 2002, p. 132). Starvation has also been identified as a cause of mortality in wolverines (Hornocker and Hash 1981, p. 1,296; Banci 1987, pp. 91, 110; Krebs *et al*. 2004, p. 497). Intraspecific predation can also contribute to wolverine deaths. Persson *et al*. (2003, p. 25) found that juvenile survival rates tended to be lower during the altricial period (May–July), and intraspecific predation was the most common cause of mortality, occurring either as infanticide or after independence. Avalanches have also been documented as a cause of wolverine deaths (Inman *et al*. 2007, p. 89).

In North America, anthropogenic causes of mortality for wolverine populations include hunting, trapping, and road kill. Discussion of the effects of hunting, trapping, and other human activities, and infrastructure (roads) is provided below (see Biological Status–Current Condition section).

Diet and Feeding

Wolverines have been described as opportunistic foragers (Inman *et al*. 2012b, p. 639) and as a "seasonal scavenger on the fringe of the food web" (Larsen 1980, p. 399). They are both scavengers and predators, with a diet that varies between seasons and years, and switching between food sources depending on availability (Magoun 1987, p. 396; Cardinal 2004, pp. 19–22; Mattisson *et al*. 2016, p. 9). The term "polyphagous" was used by Landa *et al*. (1997, p. 1,292) to describe the switching of food sources depending on prey availability by wolverines. Regional variations in diet have also been observed for wolverine populations (Nunavut, Canada) (Awan and Szor 2012, p. 9). The availability of ungulate carrion is believed to be more important than a particular habitat type for wolverines (Cardinal 2004, p. 20).

Early studies from northwestern Montana using scat analysis found that carrion (deer or elk) was an important component of wolverine diet (Hornocker and Hash 1981, p. 1,297). During winter, hoary marmots (*Marmota caligata*) were also important food items consumed and, in the spring, Columbian ground squirrels (*Urocitellus columbianus*) were heavily preyed upon (Hornocker and Hash 1981, p. 1,298). As reported by Cardinal (2004, pp. 20–21), wolverines in Canada have a large and varying diet based on reports from aboriginal traditional knowledge holders; in addition to large animals as prey or carrion, wolverine diet includes rabbits and ptarmigans (*Lagopus* sp.), porcupine (*Erethizon dorsatum*), mice, beaver (*Castor canadensis*), fish, ducks, seals, gulls and gull eggs, and lemmings, as well as antlers, bones, and skulls. Native mountain

goats (*Oreamnos americanus*) and bighorn sheep (*Ovis canadensis*), that occupy high elevation winter ranges in portions of North America, have also been suggested as important components of wolverine winter diet, particularly during the reproductive denning period (Buell Environmental 2016, pers. comm.). Snowshoe hares may be an important food item for wolverines in parts of Canada (Jung and Kukka 2013, p. 20).

In northwestern Alaska, analyses of wolverine winter diet using carcasses collected from hunters (1996–2002) within the migratory range of the Western Arctic Caribou Herd found that caribou represented the most common food item, likely through scavenging behavior, followed by moose (*Alces alces*), and to a lesser degree, microtine rodents, Arctic ground squirrels (*Spermophilus parryii*), porcupines, wolverines, red fox (*Vulpes vulpes*), sheep and ptarmigan (Dalerum *et al*. 2009, p. 249). One study found stomach contents contained a large portion of muskoxen (*Ovibos moschatus*) and Dall's sheep (*Ovis dalli*) (Dalerum *et al*. 2009, p. 249). Wolverine diets in winter (scat analysis) and summer (primarily direct observation) were evaluated by Magoun (1987, entire) in northwestern Alaska. Results from that study indicate a large number of Arctic ground squirrels were eaten in summer, while the winter diet consisted primarily of caribou and Arctic ground squirrels (Magoun 1987, p. 393). Scavenging was found to be an important feeding strategy in winter, including remnants of buried caribou carcasses or bone/hide in the tundra (Magoun 1987, p. 396).

Food habits of wolverines from 2002 to 2007 in Glacier National Park were evaluated by Yates and Copeland (2017, *in prep*) by reviewing prey remains and scat samples, or direct observations of feeding behavior. Their scat analysis found that 72 percent of samples contained more than one prey species, and 89 percent contained plant material, primarily conifer needles (Yates and Copeland 2017, *in prep*). The latter may be related to scent-marking behavior of territories, either by defecation after chewing on twigs/shrubs or terpenes released during urination (Wood *et al*. 2009, p. 574), or the result of stomach contents found within their consumed herbivorous prey (Yates and Copeland 2017, *in prep*). Overall, deer and elk represented the most frequent prey item (37 percent), but hibernating rodents were also common in scats (36 percent). Other prey items included mice, voles, lemmings, bovids (e.g., bighorn sheep, mountain goat), birds, and hares (Yates and Copeland 2017, *in prep*). Temporal differences in the occurrence of prey were also observed.

Snow tracking in Montana found that wolverines hunted in brush piles, log jams, and heavy cover, and routinely entered "tree wells," areas immediately under dense, low growing conifers where snow does not accumulate, that provide easy access to small, ground-dwelling mammals (Hornocker and Hash 1981, p. 1298). Wolverines have been described as moving and lifting large stones in order to access human-cached meat (Freuchen 1935, p. 98).

Several foraging strategies have been described for wolverines. Predation behavior on reindeer (Sweden) was detailed by Haglund (1966, p. 275). A study of elk in Siberia, Russia, noted that, in most instances, wolverines will attack young, pregnant females, young of the year, and wounded or sick animals (Knorre 1959, p. 27). Elk were chased, sometimes by two wolverines during periods of heavy snow (Knorre 1959, pp. 10, 27) and wolverines have been observed feeding in groups on large animal carcasses (Cardinal 2004, p.21). However, wolverines have

been described as neither an effective predator of large game animals, nor a serious competitor with other predators (Cardinal 2004, p. 21).

Based on studies of stomach and colon contents of wolverine carcasses in Alaska, Dalerum *et al*. (2009, p. 251) suggested that wolverines occupying this region are large ungulate specialists, but use a generalist feeding strategy by switching between ungulate food sources (e.g., caribou and moose) depending on their availability. Thus, during periods of low caribou abundance, wolverines can switch from caribou (migratory) to moose (non-migratory) while still maintaining their ecological role as a scavenger on ungulate carcasses (Dalerum *et al*. 2009, p. 251).

Wolverines were found to be the main predator of caribou calves (less than 14 days of age) in northern British Columbia, Canada (Gustine *et al*. 2006, pp. 13–14). As noted in the study, the calving areas were located in wolverine denning habitat (Gustine *et al*. 2006, p. 19); thus, this food resource may be especially important to female wolverines and their young in this region.

A study of wolverine diet using scat samples in Finland found that breeding female wolverines opportunistically used carrion and hunted less on small prey as compared to males and non-breeding females (Koskela 2013, p. 35). In addition, in areas with low densities of mid-size ungulates, smaller prey and carcasses may be important in the wolverine diet (Koskela 2013, p. 35). These results supported an optimal foraging theory; that is, wolverines will opportunistically use foods that are the most energy-efficiently available (Koskela 2013, p. 41). In other words, hunting ungulates or smaller prey (rabbits, birds) may incur greater energetic costs than scavenging for food, but searching for wolf- or human-killed carcasses will take more time (Koskela 2013, p. 41).

Finally, diet and feeding strategies of wolverines were evaluated in Scandinavia (Mattisson *et al*. 2016, entire). Wolverine feeding strategies were found to be flexible and temporarily shifted from scavenging to predation and heavily influenced by seasonal dependent responses to availability of prey and the supply of carrion (Mattisson *et al*. 2016, p. 9). Predictable anthropogenic food sources (i.e., remains from hunted ungulates) also influenced wolverine feeding strategies in their study area by increasing scavenging behavior relative to predation (Mattisson *et al*. 2016, p. 10).

Aboriginal traditional knowledge holders in Canada have reported wolverines as being largely dependent on wolves or another large predator to obtain large mammal carrion such as caribou, but also scavenge off polar bear (*Ursus maritimus*) and grizzly bear (summer) kills (Cardinal 2004, p. 20). Wolverines were observed following the tracks of Eurasian lynx (*Lynx lynx*) and then scavenging on prey left behind from lynx kills (Haglund 1966, pp. 272–273). Myhre and Myrberget (1975, p. 756) noted that the hunting abilities of wolverine and Eurasian lynx are not the same and that the two animals use the meat of their prey differently, which, together, may allow the two carnivores to coexist in the same environment.

In Sweden, Mattisson *et al*.'s (2011b, p. 1,326) study of Global Positioning System (GPS)-collared wolverines found that they spent three times longer scavenging ungulate carrion as compared to feeding on wolverine-killed prey, and more than half of the reindeer carcasses

scavenged by wolverines were killed by Eurasian lynx. That study concluded that lynx can increase the availability of food for wolverines and other scavengers and that lynx behavior around kill sites minimizes potential encounter conflicts (Mattisson *et al*. 2011b, p. 1,328). In their study area, Eurasian lynx do not appear to pose a significant threat to wolverines, neither by exclusion in space or time (Mattisson *et al*. 2011a, p. 79) nor from mortality (Persson *et al*. 2009, p. 327). We are not aware of similar evaluations for North American populations of wolverines and Canada lynx. This lack of study on interspecific processes in the more predator-diverse North American landscape is an important gap in our understanding of wolverine distribution (Fisher *et al*. 2013, p. 712).

Large carnivores can act as "sympatric ungulate predators" (Dalerum *et al*. 2009, p. 251), generating carrion at kill sites, particularly during winter months, but also as competitors and potential sources of mortality (White *et al*. 2002, p. 132; Krebs *et al*. 2004, p. 497; Koskela *et al*. 2013b, p. 221). Wolverines apparently balance their exposure to the risk of predation with foraging opportunities (Scrafford *et al.* (2017, p. 32). Thus, even though wolverines may not be dependent on lynx or other sympatric predators for their survival or reproduction, an increase in the availability of carrion and skeletal remains such as bones left by other predators (Yates and Copeland 2017, *in prep*) likely has a positive influence on the reproductive rate (e.g., number of offspring) in wolverine populations (Mattisson *et al*. 2011b, p. 1,328).

Caching of food is an important behavior of wolverines and is a key component of wolverine population dynamics (Hornocker and Hash 1981, p. 1,297; Inman *et al*. 2012b, p. 640). Food is cached in both summer and winter, by both sexes, and allows for food to be available past the peak periods of mortality and predation (Inman *et al*. 2012b, p. 639). Wolverines will typically move between carcasses and cache sites and are able to remove large parts of a carcass in a short time (Mattisson *et al*. 2011b, p. 1,327). Caching behavior in Sweden was reported most commonly in snow, as well as crevices in rock piles, and wolverines carried food to cache sites over long distances (8 and 10 km (5 and 6 mi)) (Haglund 1966, p. 274). As an example, Bjärvall (1982, p. 319) reported a female wolverine carried a reindeer head (with antlers) about 22 km (13.67 mi) back to a den site in Sweden. In northwestern Alaska, wolverines fed on cached ground squirrels during winter (Magoun 1987, p. 395).

A study of wolverine caching behavior in boreal forest habitat in Canada reported that cache sites varied from simple caches, a single feeding site or excavation, to cache complexes, which included feeding stations, latrines, resting sites, and climbing trees dispersed over varying spatial landscapes (Wright and Ernst 2004b, pp. 61–62). All cache sites included bones and hides of moose, which were likely scavenged from wolf kills (Wright and Ernst 2004b, p. 62). Cache sites were often excavated in snow, but also in the ground under boughs of large spruce (*Picea* spp.) trees (Wright and Ernst, 2004b, p. 62) and in peat bogs (Krott 1960, p. 23; Kimmy 2013, no page number). Wolverines also appeared to select cache sites and resting areas that offered good visibility of approaching competitors or predators (Wright and Ernst 2004b, pp. 63–64).

Wolverine energetic demands and food requirements are related to their foraging strategies. Caching provides important energy for female wolverines during the lactation period and helps ensure survival of newborns (Inman *et al*. 2012b, p. 640). Wolverines were found to have high energetic needs compared to other mammalian carnivores (Young *et al*. (2012, p. 2,252), similar

to results previously presented by Iversen (1972a, p. 343), who concluded the basal metabolism of mustelids weighing over 1 kg (2.2 lbs) is approximately 20 percent higher than for other mammals. A study by Andrén *et al*. (2011, p. 36) estimated a 1.2 kg/day (2.65 lbs/day) (range: 1.0–1.4 kg/day (2.2–3 lbs/day)) food requirement for wolverines, while Young *et al*. (2012, p. 223) estimated a male wolverine would require an average of 0.85 kg (1.87 lb) of prey/day in winter and 0.95 kg/day (2.1 lbs/day) in "snow-free" periods (that is, two 182-day seasons). Based on energy equivalent value of various prey sources, Young *et al*. (2012, pp. 223, 225) estimated that a winter diet for a male wolverine would include the equivalent of 1.8 ungulates (mule deer (*Odocoileus hemionus*)), 70.7 sciurids (squirrels), 20.6 lagomorphs (rabbits), and 832.7 small mammals, while in snow free season this would include the equivalent of 0.9 ungulates, 122.9 sciurids, and 3362.1 small mammals.

The study by Young *et al*. (2012, p. 225) concluded that wolverines consume 0.1 kg (3.5 oz) of prey per day more outside winter season, but that prey expected to be consumed in winter had a higher caloric content than other seasons; thus, the mass requirement is lower. As an example, they cite the higher proportion of ungulates consumed in winter, which provide about 1.3 times more energy (kilojoules per kilogram) than squirrels (Young *et al*. 2012, p. 225). Other researchers have also noted that food during the summer is just as important as the availability of cached ungulate food in the winter (e.g., during the energy demanding lactation period) (Inman *et al*. 2012b, pp. 640–642). The post-weaning growth period (May–August) was identified as a high energetic demand for food by a wolverine family group (Inman *et al*. 2012b, p. 640). Taken together with the lactation period, the calories available to wolverines therefore likely reach a maximum from March to August (Inman *et al*. 2012b, p. 640).

Population Structure

As discussed above, wolverines recolonized much of North America after periods of glaciation and then experienced heavy human persecution in much of their historic range. As shown in our Current Potential Extent (of Occurrence) maps (Figures 3 and 4) and described below in our *Population Abundance and Distribution* section, wolverines occur across a broad expanse of North America, where the contiguous United States represents the southern extent of the species' range. However, no comprehensive studies have been conducted across the western-northwestern United States to provide a reliable population estimate of wolverine populations or population structure.

A number of biological factors can affect wolverine populations, including the species' low intrinsic rate of population increase, naturally low densities, and need for large, intra-sexual home ranges (Banci and Proulx 1999, p. 180). Their extensive dispersal abilities make possible the recolonization of individuals into vacant habitats (Vangen *et al*. 2001, p; 1,647; Aronsson 2017, p. 43), though not necessarily establishment of a successful reproducing population. As noted above (*Diet and Feeding*), interactions with sympatric predators and the availability of prey and carrion can also directly and indirectly affect wolverine populations.

Wolverines in the contiguous United States are considered to represent a metapopulation (set of local or subpopulations within a larger area and where migration is possible between patches (Hanski and Simberloff 1997, p. 11)) (Inman *et al*. 2013, p. 277) and occupy habitat in high

alpine patches at low densities, dispersing into suitable areas (Inman *et al*. 2012a, pp. 782–784). Wolverines in Canada are considered to occur as a single large group as they are easily able to move between areas of good habitat and because wolverine habitat is relatively contiguous (Harrower 2017, pers. comm.).Wolverine populations in Alaska are considered to be continuous with populations in the Yukon and British Columbia provinces of Canada based on genetic studies (COSEWIC 2014, p. 37).

Studies of wolverines in the North Cascades region have documented movement of wolverines from Washington into British Columbia (Aubry *et al*. 2016, pp. 16, 20). The 2014 COSEWIC Report indicated that rescue (immigration from another population) of Canadian wolverine populations along the Canada-Alaska international boundary was likely (based on nuclear DNA evidence), but rescue of Canadian wolverine population along the U.S.–Canadian border was negligible (COSEWIC 2014, p. 37; see also Cegelski *et al*. 2006, pp. 205. 208). Based on mitochondrial DNA studies, Tomasik and Cook (2005, p. 390) concluded the gene flow in wolverines in northwestern North America is likely male-mediated, and is primarily due to long distance dispersal between low-density populations. Genetic studies of North American wolverines conducted by Kyle and Strobeck (2002, entire) found high levels of gene flow across northern populations (Canada and Alaska).

Genetics

The geographical genetic structure of wolverines is believed to be largely structured around the strong female philopatry characteristic of this species (Rico *et al*. 2015, p. 2), and, given the species polygamous behavior, wolverine population distributions (at least in Scandinavia) are considered to be primarily limited by dispersal of the more philopatric sex (females) (Aronsson 2017, p. 13). However, the extensive and often asymmetrical movement of male wolverines from core populations to the periphery of their range can result in the addition of nuclear genetic material to these edges (Zigouris *et al*. 2012, p. 1,553). Thus, the dispersal pattern for male wolverines may help explain why allelic richness (i.e., nuclear DNA) can be similar across regions, but haplotype richness (mitochondrial DNA) is lower at the periphery of the species' range (Zigouris *et al*. 2012, p. 1,553). Additionally, the extensive dispersal movements of both male and female wolverines can produce gene flow among diverged populations, making it difficult to distinguish, without additional sampling and analysis, between long-distance dispersal and fragmentation based on the patchy distribution of some haplotypes (Zigouris *et al*. 2013, p.10).

Studies evaluating the genetic structure of wolverines, primarily within its core range in North America, were presented in Chappell *et al*. (2004) and Kyle and Strobeck (2001, 2002). Using microsatellite markers, Kyle and Strobeck (2002) and Zigouris *et al*. (2012) found greater genetic structure of wolverines toward their eastern and southern peripheries of their North American distribution, likely due to a west-to-east recolonization during the Holocene (Zigouris *et al*. 2013, p. 9). Similarly, based on mitochondria DNA, McKelvey *et al*. (2014, p. 330) concluded that modern wolverine populations in the contiguous United States are the result of recolonization (following persecution during a period of unregulated hunting or trapping and poisoning) from the north.

Genetic diversity and population genetic structure of a larger sample size of wolverines were examined by Cegelski *et al*. (2006, entire) for the southern extent of their North American range using both microsatellite markers and mitochondrial DNA. They concluded that the wolverine populations in the contiguous United States were not sources for dispersing individuals into Canada (Cegelski *et al*. 2006, p. 208). They found that there was significant differentiation between most of the populations in Canada and the United States (Cegelski *et al*. 2006, p. 208). However, they cautioned that their statistical analysis may not have been able to detect "effective migrants" and that sample size can affect the detection of dispersers (Cegelski *et al*. 2006, p. 208). They concluded that some migration of wolverines was occurring between the Rocky Mountain Front region (northwestern Montana) and Canada as well as among wolverine populations in the United States, with the exception of Idaho (Cegelski *et al*. 2006, p. 208).

In addition, the authors also conducted model simulations (GENELOSS software) of the number of effective wolverine breeders necessary to maintain genetic variarion (heterozygosity) in their sampled population of the contiguous United States in the absence of gene flow (Cegelksi *et al*. 2006, p. 201). They found that, to maintain at least 95 percent of the genetic variation in the next 100 generations, 200 to 300 wolverine breeding pairs were needed in the Wyoming and Rocky Mountain Front populations, respectively, and 200 breeding pairs were needed in the Gallatin, Crazybelts, and Idaho wolverine populations (Cegelski *et al*., 2006, pp. 208–209). They also indicated that two effective migrants from either Canada or Wyoming into the Rocky Mountain Front population would be needed to maintain the levels of genetic diversity in that population, and one effective migrant was needed to maintain levels of diversity in the Gallatin, Crazybelt, or Idaho populations (Cegelski *et al*. 2006, p. 209). The authors concluded that migration is essential for maintaining diversity in wolverine populations in the contiguous United States since effective population size may never be reached due to the naturally low population densities of wolverines (Cegelski *et al*. 2006, p. 209).

Effective population size (N_e) (see **Box 2**) is defined as "the size of an idealized population that would experience the same amount of genetic drift and inbreeding as the population of interest. In popular terms, N_e is the number of individuals in a population that contribute offspring to the next generation" (Hoffman *et al*. 2017, p. 507). It represents a metric for quantifying rates of inbreeding and genetic drift and is often used in conservation management to set genetic viability targets (Olsson *et al*. 2017, p. 1). It is not the same as the more commonly used metric, census population size (N), but is often assumed to represent the genetically effective population size.

In his review of the minimum viable population size concept, Ewens (1990, entire) emphasized that the term "effective population size" is not a meaningful term unless additional context is provided relative to which the concept of population size is being evaluated (Ewens 1990, p. 309). He introduced the concept of mutation effective population size, defined as the size of population defined by its capacity to maintain genetic variation (Ewens 1990, p. 307). This is essentially the same as the eigenvalue effective population size, but is much different than actual population size (Ewens 1990, p. 309). Demographic factors are needed when interpreting actual population size from an effective population size; thus, there is no justification for a fixed, genetically-derived minimum viable population size value of '500' as each case is unique and is dependent on such factors as sex ratio, subpopulations, dispersal, and immigration (Ewens 1990, pp. 311–313). A review of the minimum viable population concept by Flather *et al*. (2011,

entire) also found that any "rule of thumb" used for minimum viable population will likely be a poor estimate for that population (Flather *et al*. 2011, pp. 311, 313). Minimum viable population estimates vary considerably both within and among species and are sensitive to the time frame in which data are collected (Flather *et al*. 2011, p. 314).

An effective population size analysis for wolverines in the contiguous United States was presented in Schwartz *et al*. (2009, p. 3,225) using wolverine samples from the main part of the Rocky Mountains populations (e.g., central and eastern Idaho, Montana, northwestern Wyoming). Excluded in this analysis, were subpopulations from the Crazy and Belt Mountains in Montana (based on suggestion by Cegelski *et al*. (2003) that they represented separate groups) (Schwartz *et al*. 2009, p. 3,225). Samples were divided into three time frames and the computer program ONeSAMP was used to estimate effective population size in each time frame [sample size appears to be between 142 and 210]. The summed effective population size was estimated at 35, with credible limits from 28–52, and the summed values for the three time frames was reported as follows: $N_{e\ 1989–1994} = 33$, credible limits 27–43; $N_{e\ 1995–2000} = 35$, credible limits 28–57; $N_{e\ 2001–2006} = 38$, credible limits 33–59 (Schwartz *et al*. 2009, p. 3,226).

Box 2. Effective Population Size and Genetic Variation

The concept of effective population size (N_e) (see review by Wang *et al*. 2016) and, relatedly, minimum viable population, has been a topic of debate, particularly the 50/500 rule, which was developed over 30 years ago. As noted by Laikre *et al*. (2016, p. 280), the concept and guidelines for genetically effective population size were developed for single, isolated populations, but it's unclear which of the various N_e metrics was referenced in the original concept proposed by Franklin (1980) (i.e., inbreeding effective size, realized effective size, total inbreeding effective size of a metapopulation, or eigenvalue effective size (Laikre *et al*. 2016, p. 288)). In his review of the minimum viable population size (MVPS) concept, Ewens (1990, p. 311) stated that Franklin's calculation of MVPS were based only on genetic arguments and therefore has no direct connection to addressing demographic questions, without further modification.

There are differing interpretations of the values proposed for effective population size. For example, should the minimum viable effective population size be derived genetically to set a threshold for a minimum viable population? Here, the rule is interpreted as 50 being the short-term number (for inbreeding depression) and 500 as the long-term number (for retention of genetic variation). Or should the N_e value of 500 be interpreted as a long-term goal for maintaining a healthy, genetically robust population, and not a threshold trigger that predicts extinction risk? In addition, some view the 500 value to be a global reference value rather than a local value, and that it may not be necessary to maintain a local N_e of 500 as long as there is some gene flow into a population (Jamieson and Allendorf 2012, p. 580).

Finally, others have recommended changes to the 50/500 rule. Laikre *et al*. (2016, entire) presented an analysis of the metapopulation effective size for the Fennoscandian wolf population and recommended that long-term conservation genetic target for metapopulations (N_{eMeta}) ≥ 500, but also a realized effective size of each subpopulation (N_{eRx}) ≥ 500. Frankham *et al*. (2014, p. 59) have recommended modifying the 50/500 rule to 100/1000.

However, Cegelski *et al*.'s (2006, p. 203) evaluation of nuclear DNA population structure in wolverines in Canada (sample size of 101) and the contiguous United States (sample size of

116), as depicted by a principle component analysis plot and dendrogram, found that all of the Canadian wolverine populations clustered together. In the contiguous United States, the Rocky Mountain Front subpopulation clustered with the Wyoming subpopulation, the Crazy and Belt Mountains area subpopulation clustered with the Gallatin (Montana) population, and the Idaho population was highly differentiated (Cegelski *et al*. 2006, p. 203). However, the sample size used for evaluating the Idaho population was small (n=15) and limited to central Idaho. That study concluded that some exchange of migrants is occurring between the Gallatin and Crazybelt wolverine populations (Cegelski *et al*. 2006, p. 207)

In addition, the map presented in Schwartz *et al*. (2009, p. 3,223) depicting the locations of the wolverine samples used in preparing their effective population size estimate shows significant gaps within the wolverine's range in Idaho and parts of Montana (e.g., interior of the Bob Marshall Wilderness area). Thus, other wolverine subpopulations and/or individuals were likely missed for this analysis.

Wolverines are known to travel (disperse) through areas outside high elevation, forested habitats. For example, tracked dispersal movements of a male wolverine, M56, from Wyoming into Colorado and its subsequent discovery in North Dakota, indicate extensive travel outside of modeled primary wolverine habitat (Inman *et al*. 2013), including through arid grasslands and shrubland habitats of the Wyoming Basin ecoregion (Packila *et al*. 2017, entire). This animal's movement also supports some level of connectivity (and gene flow) between currently occupied habitat (Wyoming) and unoccupied habitat within the wolverine's historic range (Colorado) (Packila *et al*. 2017, p. 404). Similarly, based on genetic analyses, the male wolverine currently occupying an area within the Sierra Nevada Mountains of California also represents evidence of connectivity between wolverine populations of the Rocky and Sierra Nevada Mountain Ranges (Moriarty *et al*. 2009, p. 154).

Studies within the Southwestern Crown of the Continent (SWCC) in northwestern Montana have detected cross-valley movements of wolverines, which researchers believe is an indication of good connectivity in this region (SWCC Wildlife Working Group 2016, pers. comm.). Current efforts to collect additional wolverine hair samples for genetic analyses are underway through a multi-state occupancy survey project (see *Population Abundance and Distribution* section below).

Another evaluation of mitochondrial DNA was conducted by Francis (2008), who found an overall lack of regional (geographic) genetic structure for North American wolverines, but noted that a few populations (the Crazy and Belt Mountains (Montana), Southeast Alaska, Nunavut (Canada), and Kenai Peninsula) appeared to be isolated from the others (Francis 2008, p. 12). However, statistical testing did not identify any genetically defined sampling localities (Francis 2008, p. 13). Minimal differences were found between core and peripheral wolverine populations, as grouped in that analysis (Francis 2008, p. 21; Table 4). Conversely, the study by Zigouris *et al*. (2012, p. 1,554; Table 5) did find support for genetic clusters for wolverine populations in Canada, and Zigouris *et al*. (2013, p. 5; Table 3) identified several worldwide regional genetic groups. In addition, an analysis of estimated population growth found signals of population expansion in several wolverine populations (Francis 2008, p. 13; Table 5) including

Rocky Mountain Front, Wyoming, Central, South, and Northwestern Alaska, British Columbia, Northwest Territories, and Nunavut.

It can be difficult to make inferences about the relationship between population size and point estimates of genetic diversity without continued genetic monitoring and an understanding of the demographic history of a species' population (Hoffman *et al*. 2017, p. 507), including factors that have influenced movement patterns and connectivity. Additionally, the extensive dispersal movements of both male and female wolverines described above (see Movement section) can produce gene flow among diverged populations, making it difficult to distinguish, without additional sampling and analysis, between long-distance dispersal and fragmentation based on the patchy distribution of some haplotypes (Zigouris *et al*. 2013, p.10).

Genetic diversity can be a reflection of favorable adaptations (natural selection) and is necessary for species to locally adapt to environmental stressors or to facilitate range shifts (Zigouris *et al*. 2012, p. 1,544). Genetic distinctiveness in peripheral populations may play a role in both maintaining and generating biological diversity for a species (Zigouris *et al*. 2012, p. 1,544; citing results presented in Channell and Lomolino 2000, p. 84). Genetic variation that is adaptive is a better predictor of the long-term success of populations as compared to overall genetic variation (Hoffman *et al*. 2017, p. 510). The challenge is to be able to determine whether genetic variation is adaptive and is a reflection of remnants of high genetic diversity from ancestral populations, or whether that variation is a reflection of accumulated deleterious, nonadaptive genes due to genetic drift in small populations (Hoffman *et al*. 2017, p. 509).

In summary, the currently known spatial distribution of genetic variability in wolverines in North America appears to be a reflection of a complex history where population abundance has fluctuated since the time of the last glaciation, and insufficient time has passed since human persecution for a full recovery of wolverine densities (Cardinal 2004, pp. 23–24; Zigouris *et al*. 2012, p. 1,554). Zigouris *et al*. (2012, p. 1,545) noted that the genetic diversity reported in Cegelski *et al*. (2006) and Kyle and Strobeck (2001, 2002) for the southwestern edge of the North American range represented only part of the diversity in the northern populations of wolverines. Zigouris *et al*. (2012, p. 1,545) posit that the irregular distribution of wolverines in the southwestern periphery and the genetic diversity observed in those analyses is a result of population bottlenecks that were caused by range contractions from a panmictic (random mating) northern core population approximately 150 years ago coinciding with human persecution. Recent dispersals of wolverines into Colorado, California, and Utah provide evidence for connectivity and the potential for gene flow between Northern Rocky Mountain populations and areas where wolverines were extirpated. Demographic studies as well as additional genetic analyses from contemporaneous wolverines currently occupying the contiguous United States are needed to evaluate the current status of wolverine populations in North America. In addition, ecological, phenotypical, and environmental information should be used to complement genomic data when interpreting the strength of conclusions or inferences of spatial patterns of adaptation or for adaptively divergent populations (Jamieson and Allendorf 2012, p. 492).

Summary

In this SSA report, we have incorporated information from several new studies related to the wolverine published since our 2013 proposed rule (e.g., Magoun *et al*. 2017) and previous studies that were not considered. We have also reviewed new publications and publications in preparation from wolverine researchers in Scandinavia (e.g., Aronsson 2017; Bischof *et al*. 2016; Makkonen 2015; Mattisson *et al*. 2016; Myhr 2015; Persson *et al*. 2017, *in prep*). This information informs our assessment of the most current information regarding the description of the wolverine and its life history and ecology across its North American range. We have included in this SSA Report detailed discussions of wolverine physiology, and spatial and temporal patterns and trends related to reproduction and diet/feeding. We also prepared a revised Current Potential Extent map (see Figure 3) based on information we received from Federal and State agencies, and wolverine researchers.

A species' current and future conditions and overall viability (in terms of resiliency, redundancy, and representation), are largely impacted by the availability of what the species needs at the individual, population, and species level. The needs described below are necessary for wolverines to have resources for the basic requirements of life (breeding, feeding, and sheltering) at all levels. Overall, the best available information indicates that within the contiguous United States the wolverine's physical and ecological needs include:

(1) large territories in relatively inaccessible landscapes; at high elevation (1,800 to 3,500 meters (5,906 to 11,483 feet))
(2) access to a variety of food resources, that varies with seasons; and
(3) physical/structural features (e.g., talus slopes, rugged terrain) linked to reproductive behavioral patterns.

Biological Status – Current Condition

This section provides an overview of the wolverine's current condition, including those stressors that may be impacting the species or its habitat. In this SSA Report, we have identified stressors based on impacts that may negatively affect the physical and ecological needs of the species, including temporary or permanent impacts to habitat features that the species relies on for survival and reproduction.

Population Abundance and Distribution

As shown in Figure 3, wolverines occupy areas in the west-northwestern United States, and have recently dispersed into historically occupied areas, including California, Utah, Colorado, and Oregon; verified reproducing wolverine populations are found in Idaho, Washington (Northern Cascades), Montana, and northwest Wyoming. One individual wolverine (female) was also documented from 2004 until its death in 2010 in Michigan (Michigan Department of Natural Resources 2018). Demographic data collected in the contiguous United States is primarily presence/absence data, which makes it difficult to develop reliable population estimates or population trends. Since our 2013 proposed rule, an updated Canadian status review for the wolverine has been prepared (COSEWIC 2014, entire) and additional studies have also been published related to wolverine populations in British Columbia and Alberta (e.g., Regehr and Lacroix 2016; Stewart *et al*. 2016; Webb *et al*. 2016). As noted above, we developed a Current Potential Extent map for the North American wolverine in the contiguous United States and

North America (see Figures 3 and 4) based on several resources, including the Inman *et al*. (2013), EPA ecological region mapping (2010), Forest Service NRIS data, and information received from State agencies. The 2014 COSEWIC Assessment and Status Report's presents an estimated current range map for Canada and Alaska, and includes the Yukon, Northwest Territories, Nunavut, British Columbia, Alberta, Saskatchewan, Manitoba, Ontario, Québec, Newfoundland, and Labrador (COSEWIC 2014, p. vii).

Contiguous United States

Areas in the western contiguous United States were identified by Inman *et al*. (2013, entire) as suitable for wolverine survival (long-term survival; used by resident adults), or primary habitat (Inman *et al*. 2013, p. 279), reproduction (used by reproductive females), and dispersal (female and male) of wolverines (see methodology in Inman *et al*. 2013, pp. 279–280; Figure 2). From these results, the researchers estimated potential and current distribution and abundance of wolverines in the western contiguous United States (Inman *et al*. 2013, p. 282). They estimated current population size of wolverines to be 318 (range 249–626) located within the Northern Continental Divide (Montana) and areas within the following ecoregions: Salmon-Selway (Idaho, portion of eastern Oregon), Central Linkage (primarily Idaho, Montana), Greater Yellowstone (Montana, Idaho, Wyoming), and Northern Cascades (Washington) (Inman *et al*. 2013, p. 282). Potential wolverine population capacity based on their RSF habitat modeling was estimated to be 644 (range: 506–1881) (Inman *et al*. 2013, p. 282). However, these estimates did not consider spatial characteristics related to behavior, such as territoriality, of wolverine populations.

The discussion below provides a summary of recent studies of wolverine detections and observations in the western United States; however, no comprehensive surveys have been conducted across entire area defined as the species' Maximum Extent of Occurrence (Figure 2) or Current Potential Extent (of Occurrence) (Figure 3) in the contiguous United States.

In the northern Cascades region of Washington and Canada, researchers tracked activity areas for 14 wolverines via satellite telemetry from 2007 through 2015 (Aubry *et al*. 2016, entire). This study demonstrated that the region supports a resident population, with 9 of 11 study animals documented primarily within Washington (Aubry *et al*. 2016, p. 40).

Wolverines have been detected in the Eagle Cap Wilderness Area in the Wallowa Mountains of northeastern Oregon 2011–2012, 2016, and 2017 Magoun *et al*. 2013, p. 17; Magoun 2017b, pers. comm.). Surveys conducted in the Northern Cascades of Oregon in 2012–2014 did not detect any wolverines (McFadden-Hiller and Hiller 2015, p. 107).

In California, camera trap data indicate the continued presence of a single male wolverine in the Truckee area, as of March 2017 (Shufelberger 2017, pers. comm.). The California Department of Fish and Wildlife (CDFW) has received reports of wolverine detections from the public over past several years, particularly the region near Carson Pass, as well as near Meeks Bay, Lake Tahoe (Stermer 2017, pers. comm.). The CDFW has conducted systematic surveys in these areas and has installed baited camera traps in the Carson Pass area, but no wolverines have been detected (CDFW 2017, pers. comm.). CDFW researchers are conducting multi-species predator surveys,

targeting the potential occurrence of Sierra Nevada red fox and wolverine using camera trapping with hair snares in an effort to determine occupancy, detection probability, distribution, and habitat associations (Stermer 2017, pers. comm.).

Since 2010, the Idaho Department of Fish and Game (IDFG) has directed a multi-species forest carnivore monitoring program in the Idaho Panhandle and adjoining areas of Washington, Montana, and British Columbia, Canada (Lucid *et al*. 2016; Lucid *et al*. 2017). From 2010–2014, 497 forest carnivore bait stations were used to survey the study area using a 5 by 5 km (3 by 3 mi) grid (Robinson *et al*. 2016, pp. 6,827–6,828). Wolverines were detected at 13 sites in 13 different cells, and genetic analysis identified 3 individual males (Lucid *et al*. 2016, p. 154). One male was detected on 6 occasions in the Selkirk Mountains from 2010–2013, and was determined to be resident of the area (Lucid *et al*. 2016, p. 161). This male (LS11) was also detected in British Columbia, north of Canadian Highway 3 (Lucid *et al*. 2016, p. 184), which some consider to be a barrier to wildlife passage (IDFG 2017, pers. comm.). LS11 was most recently detected in Idaho, on March 6, 2013 (Lucid *et al*. 2016, p. 175). From 2015–2016, a more focused monitoring effort in the region was conducted and resulted in the detection of two new male wolverines in the Purcell and Saint Joe Mountains, but not the three previously detected males (Lucid *et al*. 2017, p. 12). The 2017 summary report also noted one likely wolverine den in the Saint Joe Mountains (Lucid *et al*. 2017, p. 12).

A pilot study to evaluate wolverine occupancy was conducted in Wyoming from February through June in 2015 (Inman *et al*. 2015, entire). Results from that survey (hair snares and camera traps in 18 stations across 5 mountain ranges) indicated at least three individual wolverines (at five stations) with at least one individual in the Gros Ventre and Wind River mountain ranges, and at least two individuals in the Southern Absaroka mountain range (Inman *et al*. 2015, p. 9). Occupancy modeling estimated a probability of occupancy for sampled sites of 62.9 percent (Inman *et al*. 2015, p. 8).

Building on the results of the Wyoming pilot study, the Western Association of Fish and Wildlife Agencies (WAFWA), in coordination with Tribal partners, formed a multi-state, multi-agency working group (Western States Wolverine Working Group) to design and implement the Western States Wolverine Conservation Project (WSWCP)–Coordinated Occupancy Survey (see Western States Wolverine Working Group 2016 for details of protocol). The primary objectives of the WSWCP include: 1) implement a monitoring program to define a baseline wolverine distribution and genetic characteristics of the metapopulation across Montana, Idaho, Wyoming, and Washington, 2) model and maintain the connectivity of the wolverine metapopulation in western United States, and 3) develop policies to address socio-political needs to assist wolverine population expansion as a conservation tool, including translocation of wolverines (IDFG 2016, pers. comm.; Montana FWP 2016, pers. comm.; WGFD 2016, pers. comm.).

The WGFD began implementation of the survey in Wyoming in the Greater Yellowstone Ecosystem region and the Bighorn Mountains (25 grid cells) in the winter of 2015–2016 (WGFD 2016, pers. comm.). That initial survey detected, based on unique fur markings, at least three unique wolverines in the Wind River and southern Absaroka Mountain Ranges (WGFD 2016, pers. comm.). The WGFD reported 26 independent wolverine visits (defined as a photo of a wolverine greater or equal to 60 minutes from the previous photo of a wolverine), and detections

at least once within their study area during each of the four sampling periods (December 2015 through March 2016) (Bjornlie *et al*. 2017, pp. 4–5). As part of this effort, the WGFD collaborated with the Eastern Shoshone and Arapaho Tribes in Wyoming to conduct surveys on Tribal lands in 2015, and plans to continue this collaboration in the winter of 2017–2018 to monitor wolverines on the Wind River Reservation (WGFD 2017, pers. comm.).

The monitoring effort was expanded in the winter of 2016–2017 to 185 cells (cell area of 225 km^2 (87 mi^2)) in four States (Washington, Idaho, Montana, and Wyoming). **Appendix B** contains a map illustrating these preliminary detections. Wolverines were detected in 59 cells (22,461 detections) and volunteers detected wolverines in another 36 cells for a total of 95 cells (Montana FWP 2017, pers. comm.). Photographic detections of wolverine (both survey and volunteer cells) include 22 from Idaho, 48 in Montana, including detection of wolverines in all 10 cells surveyed in the SWCC region (Davis 2017, pers. comm.), 10 in Washington, including detections south of Interstate 90 (Davis 2017, pers. comm.), and 6 in Wyoming (for 2015–2016 and 2016–2017). Genetic analyses, to date, have reported a total of 157 wolverine samples (Montana FWP 2017, pers. comm.; WGFD 2017, pers. comm.). It has not yet been determined from the camera-trap images and hair samples how many of these detections are the same individual and genetic analyses of collected hair samples, including sex and individual identification are ongoing. A detailed summary of the occupancy survey design and results is presented in **Appendix C** (McDonald (on behalf of WAFWA Wolverine Subcommittee) 2017, pers. comm.).

We also requested and received additional wolverine observations from State and Federal agencies in northwestern Wyoming. A wolverine was detected by camera in northern Grand Teton National Park (Dewey 2017, pers. comm.), and a member of the public reported wolverine tracks in southwestern Grand Teton National Park while skiing (Walker pers. comm. 2017), which was confirmed by a Forest Service biologist (WGFD 2017, pers. comm.). Both of these observations occurred in March 2017. South of this area in the Wyoming Range (about 4 miles east of Alpine, Wyoming), a wolverine was detected by camera in May 2017 (Walker 2017, pers. comm.).

Alaska

The Alaska Department of Fish and Game (ADF&G) Trapper Questionnaire Annual Reports include estimates of relative abundance and trends of wolverines and other furbearers as reported by trappers (Parr 2016, p. 38). Table 4 below provides a summary of abundance and trends from 2010–2016 of those reports by region.

Table 4. Relative Abundance and Trend of Wolverine Populations, Alaska (as reported by trappers), 2010-2016.* For Trend, + indicates increase, − indicates declining/decrease, and n/c indicates no change. N/A (Not Applicable) indicates no value for that year due to changes in how the regions were defined. Sources: ADF&G 2012, 2013a, 2013b; Parr 2016.

Region	Relative Abundance				Trend			
	2010-2011	2011-2012	2012-2013	2015-2016	2010-2011	2011-2012	2012-2013	2015-2016
Region I – Southeast	scarce	scarce	scarce	scarce	n/c	n/c	n/c	−

Alaska								
Region II – Southcentral Alaska	scarce	scarce	scarce	scarce	n/c	n/c	n/c	–
Region III – Interior Alaska	scarce	scarce	scarce	scarce	n/c	n/c	n/c	–
Region IV – Central and Southwest Alaska	N/A	N/A	N/A	scarce	N/A	N/A	N/A	–
Region V - Northwest	N/A	N/A	N/A	scarce	N/A	N/A	N/A	–
Southwest	common	scarce	scarce	N/A	n/c	–	n/c	N/A
Arctic and Western	common	common	scarce	N/A	n/c	n/c	n/c	N/A

*No reports written for years 2009-2010, 2013-2015.

However, relying exclusively on trapping reports is likely to present an incomplete assessment of wolverine populations. The accuracy of information provided in the most recent report is dependent on how many trappers reply to the annual survey; for 2016, the response rate was only 11.7 percent (Parr 2016, p. 3). Trapping effort was reported to have increased by some trappers (45 percent of those reporting) during the 2015–2016 season, and 80 percent of those who increased their efforts reported an increase in their overall catch (Parr 2016, p. 15). However, this assessment does not consider how this increased trapper effort relates to harvest levels for wolverine, nor does it account for an unknown and unreported number of wolverines taken for subsistence purposes (Gardner *et al*. 2010, p. 1,894). Estimates of density, described below, provide a better depiction of wolverine population status in Alaska.

Canada

Similar to Alaska, determining wolverine population abundance and trends in Canada is difficult as numbers are developed from harvest activity (COSEWIC 2014, p. 25). Wolverine harvest trends are also difficult to estimate given the temporal and spatial variability in trapping effort and reporting of harvest, and not all regions use mandatory pelt sealing, compulsory reporting, or fur export permits/fur dealer records (COSEWIC 2014, p. 26).

According to the most recent COSEWIC Assessment and Status Report on the Wolverine, *Gulo gulo* in Canada (COSEWIC 2014, entire), Canada's western subpopulation has been estimated at 15,688 to 23,830 adults, though this value is based on several assumptions (consistent trapping effort and uniform densities across the species' range) (COSEWIC 2014, p. 36). The eastern population (Québec and Labrador) is estimated at less than 100 individuals or may be extirpated (COSEWIC 2014, p. 36). Population trends across all of Canada are not known, but wolverine populations have been stable over areas within the country's northern range for the last three generations (22.5 years) (COSEWIC 2014, p. v).

In northern Manitoba and Ontario, wolverines may be increasing in number as aerial surveys in northern Ontario have indicated an eastward reoccupation of its former range (towards James Bay and Québec) (COSEWIC 2014, p. v). However, although observations of wolverines continue to be reported within Québec and Labrador, there have been no verifiable observations since 1978, and wolverines are likely extirpated from much of southeastern Canada (COSEWIC 2014, p. v). In addition, declines in wolverine populations in the southern regions (within parts of British Columbia and Alberta) are suspected (COSEWIC 2014, p. 36). Table 5 presents a

summary of estimated wolverine populations in Canada; study references for regional estimates are presented in COSEWIC (2014, pp. 26–36).

Table 5. Wolverine Population Estimates for Canadian Territories. Source: COSEWIC, 2014.

Territory	Number of Wolverines
Yukon Territory	3,500–4,500
Northwest Territories	3,430–7,325 (with an additional 220–470 juveniles)
Nunavut	Estimated at 2,000–2,500
British Columbia	2,700–4,760
Alberta	Estimated at 1,500–2,000
Saskatchewan	Less than 1,000
Manitoba	1,100–1,600
Ontario	458–645
Québec	Very rare, at non-detectable level, or extirpated
Labrador (including mainland Newfoundland)	Very rare or extirpated

In addition to the 2014 COSEWIC summary, Cardinal 2004 (entire) prepared a complimentary summary report of wolverine trends in Canada based on Aboriginal traditional knowledge. Trends reported indicate: (1) high, relatively stable levels of wolverines in the Yukon; (2) high levels of wolverines in the North Slave region of the Northwest Territories, with populations estimated to be stable to decreasing; (3) high levels of wolverine along forested areas in the northern portions of the mainland within the Inuvialuit Settlement Region (ISR) (located in the northwest corner of the Northwest Territories) and Kitikmeot region of Nunavut; (4) an increase in wolverines in the Kivalliq region of Nunavut, but at lower levels than populations in the Boreal and North Mountain ecological areas; and (5) least abundant in the northeastern corner of Nunavut and the Arctic Islands (Cardinal 2004, pp. 22–29). In sum, the majority of traditional knowledge holders in Nunavut, Northwest Territory, and Yukon Territory describe wolverine populations as either stable or increasing in northern Canada and wolverines are now found in areas where they occurred in the past, though they are still considered naturally uncommon; only in Yellowknife did people report that wolverines might be decreasing (Cardinal 2004, pp. iii– iv, 10, 23).

Other inventory and occupancy studies include an inventory of wolverines conducted by Regehr and Lacroix (2016, entire) in the winter of 2012 on the east side of the Coast Mountains in British Columbia using a multi-method approach. They identified six individuals using genetic analysis, and one additional individual by photography, which was higher than expected as compared to model predictions of density and habitat quality (Regehr and Lacroix 2016, pp. 248–249). Estimates of wolverine occupancy were also evaluated for the Canadian Crown of the Continent ecosystem (central and southern Canadian Rockies) (Clevenger *et al*. 2016, entire). Occupancy estimates were found to vary from year-to-year and exhibited a north-south gradient, likely due to the differences in habitat quality among areas that were sampled by year (Clevenger *et al*. 2016, p. 4). For 2016, estimated wolverine occupancy probability was 0.40 for their British Columbia Rockies study area, with a declining pattern from north to south (Clevenger *et al*. 2016, p. 4). In general, their research has found that wolverines are more abundant in rugged, remote areas that have minimal human activity and landscape disturbance (Clevenger *et al*. 2016, p. 5). This study projected an expected number of wolverines in their study area of about 28

(Clevenger *et al*. 2017, p. 6). To the south, in the Southwestern Crown of the Continent (SWCC) region, northwestern Montana, wolverine surveys (snow tracking, bait stations/hair snares) have been conducted since 2012 (SWCC Wildlife Working Group 2016, pers. comm.). These survey efforts have detected 22 unique wolverines (11 males, 11 females) within three U.S. Forest Service districts, and they reported an increase in the frequency of detections from 2012 to 2015 (SWCC Wildlife Working Group 2016, pers. comm.).

In the Kootenay Region of southern British Columbia, a non-invasive genetic study of wolverines, including collection of trapper carcasses, was initiated in the southern Columbia Mountains in 2012 (Hausleitner and Kortello 2016, entire). The 2016 report estimated a population size of 9 (7–12) individuals, less than half of the expected population size of 24 (17–36), with a 64 percent estimate of occupancy for the central Purcell Mountains, which was higher than the south Purcell Mountains (38 percent) and south Selkirk Moutains (55 percent), but lower than the Valhalla (70 percent) and central Selkirk Mountains (71 percent) (estimates of detectability were found to be within the range of previous years and other studies) (Hausleitner and Kortello 2016, p. 2; Figures 1 and 2). Although the study results suggest that dispersal events are still occurring among ranges and subranges, they also found three distinct subpopulations in their genetic analyses, which they believe is indicative of limited gene flow (Hausleitner and Kortello 2016, p. 21).

The 2014 COSEWIC report indicates that trends in wolverine populations in the northern range, while uncertain, appear to be stable or increasing, but also notes that there is some concern for populations in the southern areas of British Columbia and parts of the northern United States (COSEWIC 2014, p. 22, citing personal communication from Richard Weir, Carnivore Conservation Specialist, Ministry of the Environment, British Columbia, and Wolverine Science Panel Workshop Report (Wolverine Science Panel 2014)). In British Columbia, researchers are currently conducting a multi-phase project using landscape genetic analyses to identify and delineate functional populations of wolverines and provide an estimate of size and sustainable harvest within each functional population (Weir 2017a, pers. comm.).

Estimates of Density

Wolverine densities vary across North America, and have been described as "naturally low" (van Zyll de Jong 1975, p. 434) and "naturally uncommon" (Cardinal 2004, p. iii) given the species' large home range, wide-ranging movements, and solitary characteristics. The most recent estimates (at that time) of density (number of wolverines per 1,000 km^2 (386 mi^2)) for North America were prepared by Inman *et al*. (2012a, p. 789; Table 5). In the contiguous United States, density estimates ranged from 3.5 for the Greater Yellowstone region (2001–2008) (areas above 2,150 m (7,054 ft) (latitude-adjusted elevation), 4.5 for central Idaho (1992–1995), to 15.4 for northwestern Montana (1972–1977).

In Alaska and Yukon, density estimates presented by Inman *et al*. (2012a, p. 789) range from 3 to about 14 wolverines per 1000 km^2 (386 mi^2), using a number of methods. For example, Royle *et al*. (2011, p. 609) estimated wolverine densities for southeastern Alaska (Tongass National Forest; 2008) from 8.2 to 9.7 per 1000 km^2 (386 mi^2) (using mark-recapture), where the higher estimate incorporates a positive, trap-specific behavioral response. Density of wolverines were

recently reported as an estimated 5–10 wolverines per 1000 km^2 (386 mi^2) (based on snow tracking) for southcentral Alaska, and approximately 10 per 1000 km^2 (386 mi^2) (based on DNA mark-recapture methods) for southeast Alaska (Golden 2017, pers. comm.). A wolverine occupancy study in 2015 within an area of central Alaska reported a density estimate of 9.48 wolverines per 1,000 km^2 (386 mi^2) (ADF&G 2015a, p. 7).

Wolverine density estimates for Canada varies across regions, from 5 to 10 per 1000 km^2 (386 mi^2) in northern mountain and boreal regions to 1 to 4 per 1000 km^2 (386 mi^2) in southern boreal areas (COSEWIC 2014, p. 27). More recently, Clevenger *et al*. (2017, entire) presented a density estimate (using spatial capture/recapture models) for the Kootenay region of British Columbia of 0.78 wolverines per 1000 km^2 (386 mi^2), for 3 study years (2014–2016), which they reported as lower than expected (Clevenger *et al*. 2017, p. 6).

Stressors – Causes and Effects

We reviewed the best available information to identify current conditions and potential stressors that may be affecting wolverine populations or its habitat. These include roads and other infrastructure, recreational activity and other human disturbances, wildland fire, disease or predation, and overutilization for commercial, recreational, scientific, or educational purposes. Because wolverines in North America move between both borders of Canada (i.e., contiguous United States, Alaska), we included in our evaluation stressors identified for wolverines in Canada and Alaska that are also relevant for wolverine populations in the contiguous United States.

As an initial step, we reviewed the land ownership of the area defined as Current Potential Extent in the contiguous United States (Figure 3), and determined that 72.4 percent is located on Federal land (using Albers Projection) (see **Appendix D**). Lands managed by the Forest Service (62 percent) represent the largest portion of Federal lands within this area.

Effects from Roads

Roads and rail lines can be a cause of mortality to wolverines and habitat models have identified road density as an important association (avoidance) for selection of habitat (e.g., Rowland *et al*. 2003; Bowman *et al*. 2010; Inman *et al*. 2013). Road density has been listed as a threat to wolverines occupying the boreal/western mountain regions of Canada (Canadian Boreal Forest Agreement 2014, p. 2). In the wolverine's southern Canadian range, roads may be facilitating direct mortality along transportation corridors and may facilitate indirect mortality through improved motorized access for hunters, trappers and recreational users into remote areas (COSEWIC 2014, p. 21).

In their review of 12 radio-telemetry studies (1972 to 2001) of wolverines in North America, Krebs *et al*. (2004, p. 497) reported 3 mortalities of wolverines from road-rail kills. In addition, a female wolverine was killed on Highway 20 in northwestern Washington in May 1997 (specimen at Burke Museum; Bradley 2017, pers. comm.). More recently, road mortalities have been recorded in Idaho (one confirmed in 2014) (IDFG 2017a); three in Montana (2004, 2007) (Kociolek *et al*. 2016, p. 68; Yates 2017, pers. comm.); one in Utah (2016) (Hersey 2017, pers.

comm.). Two additional wolverine road-rail fatalities in Montana were reported in 2015 (Inman 2017a, pers. comm.).

In Canada, anthropogenic causes of mortality for wolverine populations also include road kill (COSEWSIC 2014, p. v). One road mortality of a male wolverine was reported in a lowland boreal forest region of Ontario, Canada (Dawson *et al*. (2010, p. 142). More recently, Scrafford *et al*. (2017, p. 34) described a report in which nine wolverines were struck and killed by vehicles in the Hay-Zama region of northwestern Alberta, Canada (2013–2015), and one road mortality occurred within the town of Rainbow Lake in Alberta.

Roads also may affect den site selection (May *et al*. 2012, p. 202), particularly areas within their range where they cannot select for high elevation habitat (e.g., central lowland forests of Canada) (Dawson *et al*. 2010, p. 143). In Norway, May *et al*. (2012, p 202) found that wolverine dens were generally located far from infrastructure (public roads and private roads and/or recreational cabins). The authors reported a minimum threshold in den site selection relative to infrastructure of 1.4 km (0.87 mi) from private roads and 7.5 km (4.7 mi) from public roads (May *et al*. 2012, p. 202). However, they found that wolverines in their study area had a wide tolerance range at the home-range scale (1.0–2.75 km (0.62–1.7 mi) for private roads and 6.0–11.0 (3.7–6.8 mi) for public roads) (May *et al*. 2012, p. 201; Figure 4), supporting conclusions from other studies that have found that, once a general area is used, it appears to be re-used in subsequent years including by successive individuals colonizing the sites (May *et al*. 2012, p. 202).

Wolverine road crossings were evaluated in the western Greater Yellowstone region through telemetered animals and visual observations of snow tracks, direct observations of crossings, and road-kill mortality (Packila *et al*. 2007, entire). That study documented 43 crossings of U.S. and State highways by 12 wolverines (Packila *et al*. 2007, p. 105). Within the Big Sky, Montana, area, they documented 67 crossings of MT64/Jack Creek Road by 4 wolverines (Packila *et al*. 2007, p. 105). Most (76 percent) road crossings were made by subadult wolverines, dispersing or otherwise exploring new areas (Packila *et al*. 2007, p. 105). One road-caused mortality was observed and the authors report two others from additional sources during their study period (Inman *et al*. 2007, p. 89). The study results indicate that roads do not act as absolute barriers to movement by wolverines, but they can directly affect individuals (road kill) and may have secondary effects at the population level (Packila *et al*. 2007, p. 105).

In an effort to evaluate the potential impact of major roads to wolverine (individuals and populations), we conducted a spatial analysis of roads[3] found within our Current Potential Extent for the wolverine in the contiguous United States (see Figure 3), as measured by number of kilometers (miles). In our analysis, we identified several road classes: Interstate Highway, U.S. Highway, State Highway, roads identified by Federal agency, County highways, County roads, and local roads. Our analysis found approximately 248,966 km (154,700 mi) of roads with the Current Potential Extent area, with local roads representing 83 percent (209,719 km (128,449 mi)) of all roads within our Current Potential Extent area.

[3] Using U.S. Geological Survey National Transportation Dataset Downloadable Data Collection based on TIGER/Line data provided through U.S. Census Bureau and supplemented with 'HERE' road data to create tile cache base maps.

We then evaluated the type of roads at high elevation within our estimated Current Potential Extent area. Using the 2,300 m (7,546 ft) elevation as a benchmark (based on its use as a predictor variable for wolverine occurrence in Inman *et al*. 2013, and results from predictive models presented in Copeland *et al*. (2007, p. 2,205)), we evaluated the length of roads above and below this elevation, and also the type of roads at or above 2,300 m (7,546 ft). The results are illustrated in **Appendix E**. Overall, we found that approximately 95 percent of all roads were below 2,300 m (7,546 ft). Of the roads located at or above 2,300 m (7,546 ft), 85 percent are local roads.

Using the same dataset, we evaluated road density (km/km^2) based on regional blocks of primary wolverine habitat in the western United States delineated by Inman *et al*. (2013; Figure 3). Those results are shown in Table 6. With the exception of the Southern Rockies (at 0.47 km/km^2), the mean road densities at elevations equal to or greater than 2,300 m (7,546 ft) are very low.

Table 6. Mean Road Density in Wolverine Primary Habitat.

Geographic Region‡	Mean density (km/km^2), all roads	Mean density (km/km^2), all roads ≥ 2,300 m (7,546 ft)
Northern Cascade	0.54	0.00
North Continental Divide	0.54	0.00
Salmon-Selway	0.70	0.03
Central Linkage	0.84	0.06
Greater Yellowstone	0.24	0.06
Southern Rockies	0.55	0.47
Sierra Nevada	0.09	0.03
Uinta	0.15	0.12
Bighorn	0.00	0.00
Great Basin	0.06	0.03
Oregon Cascade	0.72	0.00

‡Regions defined in Inman *et al*. (2013; Figure 3).

We also reviewed den site locations (natal, maternal, or unknown dens) within our database relative to roads (see map in **Appendix E**). Our results indicate that wolverine dens in the contiguous United States are located in areas with minimal roads, including local roads; however, we caution that this analysis is based on a limited den site dataset and should be viewed in the context of other abiotic and biotic variables including landscape features at the den site scale and availability of food. Additionally, most den locations in much of the wolverine's range in the contiguous United States are at high elevations and roads in these areas would likely be impassable or closed entirely to vehicles during the time of denning (January–March).

In summary, wolverines are associated with habitat found in high elevation areas, but are known to disperse over great distances. Major highways can present mortality risks to dispersing individuals and affect immigration to open territories, but roads do not represent absolute barriers to wolverine movements. Wolverines den during winter months in locations that are often inaccessible or restricted to motorized vehicles, though, secondary roads and trails are used for winter recreational activity. Although we recognize there are likely additional events that have not been reported, we estimated the total number of wolverine mortalities due to roads from 1972 to 2016 (44 years) in North America was 20, at least 11 of which are from Canada (see citations

above). As discussed above, we calculated a low proportion of major highways in both modeled primary habitat and a low mean density of roads at high elevations where wolverines have been observed, with the exception of the southern Rocky Mountains. Roads present a low stressor to wolverines at the individual and population level in most of its current contiguous United States range.

Disturbance due to Winter Recreational Activity

Wolverine behavior patterns, such as denning, rearing of young, movement and dispersal, and foraging/scavenging, may be affected by recreational activities (COSEWIC 2014, p. 42), although several wolverines have been captured for research on or near ski areas (e.g., Teton Mountains) (Montana FWP 2017, pers. comm.) As noted above, in Norway, May *et al*. (2012, p. 201) found, at the home range scale, a minimal threshold distance of approximately 1.5 km (0.93 mi) for wolverine den sites from private roads and/or recreational cabins. Krebs *et al*. (2007, entire) evaluated habitat use associations for wolverines in two multiple use areas in British Columbia, Canada. Using logistic regression models, the authors found that in an area of active recreation (Columbia Mountains), female wolverines were negatively associated with helicopter and backcountry skiing in their winter models (Krebs *et al*. 2007, pp. 2,187–2,188). In summer months, Copeland *et al*. (2007, p. 2,210) reported that wolverines in their study area of central Idaho were not uncommonly found near maintained trails and active campgrounds.

The Wolverine–Winter Recreation Study represents an on-going project to evaluate the potential effects of backcountry winter recreation (e.g., backcountry skiers, heli-skiers, cat-skiers, snowmobilers) on wolverines (Heinemeyer 2016, pers. comm.). The multiyear study areas include central Idaho and areas in the western Yellowstone region (Island Park area and Teton Mountains) (Heinemeyer and Squires 2015, p. 3). The study has been monitoring wolverines using GPS collars using movement rates and percent of time active (vs. resting) as indicators of potential responses of wolverines to winter recreation activities (Heinemeyer 2013, pers. comm.). Backcountry winter recreation activities are monitored through GPS units voluntarily carried by recreationists (Heinemeyer 2016, pers. comm.). Early analysis of the data suggest that wolverines demonstrate a behavioral response to recreation activities, such as increased movement rates and a reduction in resting periods in areas of high recreation activity, especially high recreation days (Saturday and Sunday) (Heinemeyer and Squires 2013, pp. 5, 7–8).

However, this research has also found that wolverines maintained their home ranges within areas with relatively high winter recreation activity over several years of monitoring, including some areas found to contain the highest recreational activities (Heimemeyer 2016, pers. comm.). The study has not been able to determine whether these resident wolverines are reproductively successful due to the limited monitoring information available for reproductive females (Heinemeyer 2016, pers. comm.).

A final Winter Recreation Study report was prepared by Heinemeyer *et al*. in December 2017. The study presented results from a RSF model in an effort to characterize habitat selection responses of wolverines to winter recreation patterns. The study found that wolverines maintained multi-year home ranges in areas that support relatively intensive winter recreation, suggesting that wolverines are able to tolerate winter recreation at some scales (Heinemeyer *et*

al. 2017, p. iv). They also reported that wolverines responded negatively to increasing intensity of winter recreation, with off-road and dispersed recreation having a greater effect than recreation that was concentrated on access routes (Heinemeyer *et al*. 2017, p. 34). In addition, wolverine avoidance of roads and groomed areas used by winter recreationists was less than estimated for dispersed recreation, suggesting that wolverines may be less sensitive to predictable winter recreation use patterns (Heinemeyer *et al*. 2017, p. 40). The authors also described habitat selection in females evaluated in the mutli-year study as complex, and likely driven by a combination abiotic (snow, cold) and biotic factors (predator avoidance, food availability) (Heinemeyer *et al*. 2017, p. 36).

Conservation measures currently being implemented that address the effects of roads in the Teton Mountains include winter closures in certain areas (generally from November 1 through May 1), including road closures in the Bridger-Teton and Caribou-Targhee National Forests and in Grand Teton National Park as shown in **Appendix F** (additional details for Grand Teton National Park are described in Superintendent's Compendium (National Park Service (NPS) 2017; pp. 8–9); see also maps at https://jhalliance.org/campaigns/dont-poach-the-powder/ Jackson Hole Conservation Alliance 2017). These closures are being implemented to help minimize disturbance to wildlife (e.g., migration pathways).

State Wildlife Action Plans prepared for individual western States identify recreation management strategies within wolverine habitats. For example, in Oregon, the ODFW Conservation Strategy identifies management of winter recreation use in order to avoid impacts to wolverines (ODFW 2016). In Montana's State Wildlife Action Plan, conservation actions are identified for addressing potential impacts from recreation, such as consideration of seasonal closures during denning season (Montana FWP 2015, p. 63). The IDFG *Management Plan for the Conservation of Wolverines in Idaho* also includes conservation strategies related to winter recreation (e.g., characterizing wolverine response to recreational activities (IDFG 2014, p. 35)), and the State continues to support the Wolverine-Winter Recreation Study. **Appendix G** provides additional details on individual State conservation strategies.

In summary, wolverine behavior (movement) can be affected by winter recreational activity. Results from one long-term study in parts of the wolverine's range in the contiguous United States have found that wolverines can maintain residency in high winter recreational use areas. Wolverines have recently been detected in areas that experience winter recreational activity. Conservation strategies and actions have been identified in several western States' Wildlife Action Plans to address potential impacts of this stressor to wolverines. Based on the best available scientific and commercial information, the effect of winter recreational activity represents a low stressor to wolverines in the contiguous United States at the individual and population level.

Other Human Disturbance

Infrastructure, such as pipelines, active logging or clearcuts, seismic lines, and activities associated with mining (e.g., producing mines, mines under development, mineral exploration areas), may also affect individual wolverine behavior (e.g., avoidance) or loss or modification of wolverine habitat. As discussed above (see Habitat Use section), Johnson *et al*. (2005, entire)

evaluated habitat relationships for the wolverine and other arctic wildlife, including the cumulative effects of human activities and associated infrastructure on the distribution of wolverines in the Canadian central Arctic using RSF modeling. However, because human disturbance factors (i.e., major developments, mineral explorations, seasonal outfitter camps) were mostly absent from the range of monitored wolverines, the researchers were not able to reliably model their effects (Johnson *et al*. 2005, p. 23).

The 2014 COSEWIC status review identified several potential stressors to wolverines and their habitat in Canadian territories. They indicated potential permanent, temporary, and functional losses to wolverine habitat from forestry; oil, gas, and mineral exploration and development; and large hydroelectric reservoirs (COSEWIC 2014, p. 21). As discussed above, Scrafford *et al*. (2017, entire) evaluated habitat selection of wolverines in response to human disturbance in the western Canadian boreal forest in both winter and summer months. Their analysis found that wolverines were attracted to some industrial infrastructure (older seismic lines exhibiting latter stages of regeneration) and disturbance (areas of active logging), likely related to foraging opportunities (e.g., small prey), but avoided interior areas of intermediate-aged cutblocks (areas authorized for logging) (Scrafford *et al*. 2017, pp. 32–34). Their results found evidence of road avoidance, but wolverines were attracted to all-season road sections with borrow pits, which they suggest was related to foraging opportunities at these pits (e.g., presence of beavers in water-filled pits) and less predation risk, since wolves avoid these roads (Scrafford *et al*. 2017, p. 34). In sum, these authors concluded that wolverine selection patterns relative to industrial activity and infrastructure in their study area represent a balance between exposure to predators and foraging opportunities (Scrafford *et al*. 2017, p. 32).

Additional studies of wolverine behaviors related to the effects of disturbance due to infrastructure and other human activities are needed. Based on the best available scientific and commercial information, these effects are small or narrow in scope and scale and appear to represent a trade-off between foraging opportunities in areas that provide minimal risk of predation and avoidance of open areas and/or higher predation risk (Scrafford *et al*. 2017, pp. 33–34).

Effects from Wildland Fire

Wildland fire can produce both direct and indirect effects to wildlife. Direct effects include injury and mortality as well as escape or emigration movement away from fires (Lyon *et al*. 2000, pp. 17–21). Small mammals will generally find refuge underground or within sheltered places within the burning area, while larger mammals will move to safe areas in unburned patches or outside the burn (Lyon *et al*. 2000, p. 18). For animals that emigrate during fire events, the length of time before they return is dependent on the degree to which fire has altered habitat structure and food supply (Lyon *et al*. 2000, p. 20).

We are unaware of any studies evaluating direct effects of wildland fire to wolverines. Wildland fire is likely to temporarily displace wolverines, which could affect home range dynamics. Given that wolverines can travel long distances in a short period of time, individuals would be expected to move away from fire and smoke (Luensmann 2008, p. 14). In addition, because

young are born during winter months, fire risk at that critical life stage is very low (Luensmann 2008, p. 14).

Indirect effects of wildland fire can include habitat-related effects or effects to prey and competitors/predators; however, we are unaware of empirical studies evaluating these potential effects as they relate to wolverines. In a study area within the Yukon (Canada), wolverines were reported occupying regenerating forested habitat that contained remnants of mature timber which had burned 30 years prior (Slough and Mowat 1996, p. 948). Additionally, fire suppression in conjunction with logging activities in boreal forests (northwestern Ontario) can increase the prevalence of deciduous tree habitats, at least at a regional level, which is negatively associated with wolverine occurrence (Bowman *et al*. 2010, p. 464).

A study in northern Idaho of the effects of multiple wildland fires over several years, including very large fires, to forest habitat occupied by another mustelid, the American marten (*Martes americana*) found that fire events had created a mosaic of vegetation that supported a diverse assemblage of cover and food resources that was favorable to this species (Koehler and Hornocker 1977, p. 503). Similar to wolverines, the summer and fall diet of the American marten is represented by diverse prey, and wildland fire events can create and maintain forest openings for ground squirrels and voles (Koehler and Hornocker 1977, p. 504). The development of these types of mosaic forest communities following certain wildland fire events also provides discontinuous fuel loads, which in turn should result in smaller and cooler wildland fires, with less replacement of marten habitat (Koehler and Hornocker 1977, p. 504). However, large, uniform burns would be expected to result in more severe impacts to American marten habitat (Lyon *et al*. 2000, p. 21).

Studies of the effects of wildland fire to a key prey species for the wolverine in parts of its North American range, the caribou, was reviewed by Klein (1982, entire). This review highlighted the importance of separating short-term effects of wildland fire in boreal forests to caribou ecology from long-term effects (Klein 1982, p. 393). Given that long-term benefits to the species' ecology can be disproportionate to the short-term detrimental effects on populations and herds, (including the species' lack of reproductive plasticity), caribou may be more appropriately considered as fire-influenced, rather than fire-adapted (Klein 1982, p. 393). Other ungulate prey species respond more positively to fire. An increase in spring and summer grasses following fall burns can provide forage for elk and deer, and sprouting of deciduous trees, such as aspen, birch and willow, following burns provides forage for moose (Luensmann 2008, p. 18).

Management measures to address this potential stressor are identified in USDA Forest Service National Forest Land Management Plans. Examples of these goals and objectives are described in **Appendix G**. In addition, the Idaho State Wildlife Action Plan includes measures to address fire threats to the wolverine and its habitat, including removal of perceived barriers to allow more prescribed natural fire on State and private forest lands and promoting/facilitating the use of prescribed fire as a habitat restoration tool, on both public and private lands where appropriate, and leaving fire-killed trees standing as wildlife habitat if they pose no safety hazard, all in an effort to restore a more natural fire interval that allows for return to historical forest conditions (IDFG 2017b, pp. 91, 134, 180).

Given the diversity of habitats occupied by wolverines, their occupancy of high elevations, and extensive mobility, wildland fire represents a limited stressor, in scope and scale, to wolverine habitat and its prey in the contiguous United States range.

Disease or Predation

Disease

We are unaware of comprehensive surveys evaluating the prevalence of diseases in wolverines in the contiguous United States. Early accounts of endoparasites species and their prevalence in wolverines include a review by Erickson (1946, p. 503), and a report by Rausch (1959, entire), who documented 7 species of helminth parasites in 86 percent of wolverines examined from trapper-supplied carcasses in Alaska. In 1994, Copeland (1996, p. 26) collected a single specimen of the parasite *Toxascaris* sp. from wolverine scat in Idaho. In Alaska, carcasses sampled (during necropsy or predator control activities) in 2012–2014 to determine the prevalence of *Trichinella* and its genotypes reported one wolverine with *Trichinella* T6 genotype in that single sample (ADF&G 2015b, p. 8). Results from Alaska trapper questionnaires for the prevalence of ectoparasites on wolverines were either scarce or not present across all reporting regions in 2015–2016 (Parr 2016, p. 21).

Rabies is endemic to Alaska in Arctic and red fox along north and west coasts of Alaska (ADF&G 2013c). Under the ADF&G enhanced rabies surveillance program, the agency confirmed rabies in one wolverine (out of 49 sampled) in 2012, a female found dead in the North Slope region (Woodford and Beckman 2012). This was the first confirmed case of rabies in wolverines in North America (Woodford and Beckham 2012).

The 2014 COSEWIC Assessment and Status Report for the wolverine presented a summary of reported parasitic species observed in wolverines in Canada (COSEWIC 2014, p. 25). These observations included: parasitic nematode roundworms (*Trichinella* spp.) in 88 percent of wolverine samples tested from Nunavut and 26 percent from the lower MacKenzie region; helminth parasites (trematodes, cestodes and nematodes) in wolverine digestive tracts from the lower Mackenzie River valley; and, from the Nunavut region, protozoan parasites infections including *Sacrosystis* spp. (80 percent) and *Toxoplasma gondii* (41 percent) (citations omitted). Banci (1987, pp. 81, 110) reported parasitic pneumonia as a cause of mortality in southwest Yukon Territory, a female thought be nutritionally-stressed following the raising of young.

An evaluation of trapper-submitted wolverine carcasses harvested was conducted for the Yukon Territory in the fur trapping seasons 2005–2006 through 2011–2012 (Jung and Kukka 2013, entire). No samples tested positive for rabies (Jung and Kukka 2013, p. 17). Another study of intestinal parasites of wolverine carcasses from both the Yukon and Northwest Territories reported *Trichinella* spp. in 74 percent of carcasses and several intestinal parasites, including cestodes (parasitic flatworms) such as *Taenia* spp. (Luck *et al*. 2016, no page number).

In summary, other than a parasitic pneumonia mortality event and the single rabies case, we are not aware of any other studies documenting impacts of disease to wolverines in North America. At this time, based on the best available scientific and commercial information, we do not find

that disease is a population or species level stressor to the wolverine in the contiguous United States.

Predation

As discussed above (*Diet and Feeding* section), a number of potential natural predators have been identified for wolverines within its North American range, including intraspecific predation. However, we have no information that suggests this predation represents a significant stressor to the wolverine at the population level. At the individual level, we recognize that wolverines likely avoid areas of potential predation risk from wolves (Scrafford *et al*. 2017, p. 34) and other potential predators. Thus, indirect effects of predators may result in predator avoidance behavior of individual wolverines through habitat selection.

In summary, the best scientific and commercial information available indicates that disease or predation is not a stressor for the wolverine. We are unaware of any management or conservation measures currently in place to reduce potential impacts associated with disease or predation.

Overutilization for Commercial, Recreational, Scientific, or Educational Purposes

There is currently no allowable trapping or harvesting of wolverines in the contiguous United States, though incidental trapping mortalities have been documented as we reported in our proposed rule (78 FR 7881; February 4, 2013). Two mortality events from shootings of wolverines were documented in Idaho (2001, 2007) (IDFG 2014, p. 26).

In Montana, wolverines were a legally harvested furbearer in Montana up until 2012. As noted in **Appendix G**, beginning in 2008, the State shifted harvest units and quotas to emphasize proetection in smaller mountain ranges, closing 40 percent of the State to harvest and the trapping season is currently suspended with a zero statewide quota (Montana Natural Heritage Program and Montana FWP 2017). Montana FWP also adopted mandatory trapping education in August 2017, which begins in 2018, which will emphasize avoidance of incidental captures and modern equipment in order to minimize injury (Montana FWP 2017, pers. comm.). There is, however, no evidence to suggest that the harvest of wolverines in Montana at historical rates (about 10 animals per year) was detrimental to wolverine populations, and since 2005, the rates and spatial arrangement of wolverine harvest have been ~~much~~ more conservative than the historical rates (Montana FWP 2017, pers. comm.) (see **Appendix G**).

Unlike populations in Eurasia, wolverines rarely prey on livestock in North America (*cf.* domestic sheep predation in Wyoming reported (Mead 2013, pers. comm.)) and therefore they are not directly targeted for predator control (COSEWIC 2014, p. 41). However, incidental trapping can result in the capture of non-target species such as wolverine. In Idaho, the IDFG has a mandatory furtaker harvest report that requests all live incidental catches be reported by species and any wolverine catch that results in mortality is required to be reported (IDFG 2013, pers. comm.). Since 1965, over a period of just over 50 years, 17 incidentally-trapped wolverines were reported during the State's furbearing seasons, with 6 animals known to be released alive and 6 mortalities (IDFG 2013, pers. comm.; IDFG 2016, pers. comm.; IDFG 2017, pers. comm.). This total includes four wolverines caught during the 2013–2014 furbearer season, with three released

alive and one mortality (IDFG 2014, p. 26), and one animal in 2016, who released itself prior to discovery (identified through DNA testing). Though the fate of that animal is unknown, there was no obvious sign of injury (IDFG 2017, pers. comm.). In December 2017, a trapper reported to IDFG that a female wolverine was a nontarget capture in a snare trap in the Beaverhead Mountain Range Forest near Salmon, Idaho (approximately 1,981 m (6,500 ft) elevation), which resulted in mortality (Mack 2018, pers. comm.).

The State of Utah maintains a highly regulated trapping program, including mandatory trap check intervals (Utah Department of Natural Resources (UDNR) 2017, pers. comm.). Any incidentally trapped wolverines must be released and any mortality must be reported to the Division of Wildlife Resources within 48 hours; there are no records of incidental captures (UDNR 2017, pers. comm.). Within the State of Wyoming, there are two confirmed reports of incidental take, one in 1996 (Mead 2013, pers. comm.) one in 2006. The 2006 animal was released unharmed (Inman 2012, pers. comm.). In Montana, since the closing of the trapping season for wolverine in 2013 (2013 through 2016), four wolverines have been incidentally trapped, one in 2013 and three in 2014 (Inman 2017c, pers. comm.). The 2013 capture was released unharmed and the three caught in 2014 were all mortalities (Inman 2017c, pers. comm.).

Predator control programs targeting wolves, including poison and incidental trapping, can result in incidental losses of wolverines (COSEWIC 2014, p. 41). Specific to wolf control for livestock protection in Idaho, three wolverines have been trapped incidental to authorized wolf control activities since 1995, with two released alive and one animal euthanized (IDFG 2014, p. 26). Preventive measures have been adopted to reduce these incidental captures, including implementation of educational programs to minimize incidental capture of wolverines during trapping seasons (IDFG 2014, p. 27). Licensed wolf trappers are required to complete a Wolf Trapper Education course with specific instruction for reducing incidental trapping of wolverine, Canada lynx, and other non-target species (IDFG 2014, p. 27). In addition, the U.S. Department of Agriculture Wildlife Services (Wildlife Services) agency has also temporarily stopped (as of April 2017) using cyanide predator control devices in the State of Idaho (Moeller 2017). Wildlife Services has also agreed to not use, or fund the use of, M-44 cyanide devices on public lands in Colorado pending issuance of an Environmental Assessment (Civil Action No. 17-cv-00891-WYD) and has agreed to interim measures, including not using EPA-labeled pesticides targeting mammalian species, in the Northern District of California, pending environmental analsysis (Case No. 3:17-cv-3564-WHA) .

In Alaska, wolverine trapping and hunting is controlled by seasons and bag limits, with about 550 animals harvested each year (Alaska Department of Fish and Game (ADF&G) 2017a). For the 2015–2016 reporting period, wolverine harvest, based on furbearer sealing records,[4] totaled 527 animals (Parr 2016, p. 42). This level of harvest has been fairly consistent since 2010, as shown in Table 7 below.

Table 7. Number of wolverines harvested in Alaska, as reported from regulatory year sealing records, 2010–2015. Adapted from Parr (2016, p. 42; Table 10).

[4] Wolverines taken in Alaska are required to be sealed by an authorized department representative before pelts are shipped to an out-of-state buyer or auction house (Parr 2016, p. 44). For those species that require sealing, the number of animals sealed represents the best information regarding the statewide harvest (Parr 2016, p. 41).

Alaska Region	2010	2011	2012	2013	2014	2015
I	25	20	25	31	14	15
II	25	29	50	31	16	37
III	233	235	261	358	268	214
IV	180	160	170	158	99	150
V	140	110	135	133	109	111
Total	603	554	641	711	506	527

Trapping and harvesting of wolverines occurs over much of the range in Canada, as summarized in the 2014 COSEWIC wolverine status review (COSEWIC 2014, pp. 10, 29–35). Specifically, wolverines are harvested in the northern and western territories–Manitoba, Saskatchewan, Alberta, British Columbia, Yukon, Northwest Territories, and Nunavut (COSEWIC 2014, p. 43). Non-aboriginal harvest of wolverines has not been permitted since 2001–2002 in Québec and Labrador (COSEWIC 2014, p. 43). Trapping is closed in Ontario (except through treaty rights), though incidental trapping results in 1 to 4 mortalities per year (Bowman *et al*. 2010, p. 465). Harvest levels in western provinces have remained relatively stable since 1992 (COSEWIC 2014, p. 38; Table 1). An earlier assessment of the sustainability of wolverine harvest in British Columbia was presented in Lofroth and Ott (2007, entire). That study found that the trapping and hunting harvest of British Columbia's wolverine population was likely sustainable at a provincial scale from 1985 to 2004, but wolverine harvest in 15 of 71 population units were reported as likely unsustainable during this period (Lofroth and Ott 2007, p. 2,197).

The management of wolverine harvest in Canada incorporates spatial and temporal elements such as season length, quotas, limited entry, and trapline management by trappers (reviewed by Slough *et al*. 1987). Wolverine harvest levels in Canada are monitored using mandatory pelt sealing, annual harvest reporting, or through monitoring of fur exports (COSEWIC 2014, p. 43). In some northern communities, wolverine pelts are used locally and harvests are monitored through carcass collection programs (COSEWIC 2014, p. 43).

The COSEWIC Assessment and Status Report for the wolverine also noted that range contraction and habitat trends of wolverines in Canada are not solely the result of habitat or trapping pressure (COSEWIC 2014, p. 20). Reductions in ungulate (e.g., caribou) populations, which provide an important winter food resource, were also likely an important factor in range contractions of wolverines in its northern range (COSEWIC 2014, p. 20), and likely continue to influence populations today. Snowmobiles have allowed for better access for hunters and trappers and may be increasing the number of wolverine harvested in its northern North America range; however, the areas of exploitation are still relatively small concentrated areas, and large areas of refugia continue to be found (Cardinal 2004, p. 31).

As described in the 2014 COSEWIC report, trends in wolverine populations in the northern range, while uncertain, appear to be stable or increasing, with some concern for populations in the southern areas of British Columbia and parts of the northern United States (COSEWIC 2014, p. 22, and references cited therein). Similarly, in Alaska, over the past 6 years, on average, 590 wolverines are taken each year (see Table 7). The consistent harvest levels in these regions suggest relatively stable wolverine populations.

We evaluated trapping of wolverines in British Columbia and Alberta regions of Canada in an effort to document potential impacts to dispersing wolverines along the U.S.–Canada border. As described above (*Population Abundance and Distribution*), the population of wolverines in British Columbia is estimated to be 2,700–4,760 and 1,500–2,000 animals in Alberta (COSEWIC 2014, p 36). We obtained 9 years (2007–2015) of harvest data for southern BC wildlife management units from the British Columbia Ministry of Environment, Ecosystems Branch for our analysis (Weir 2017b, pers. comm.). Twenty seven years (1989–2015) of harvest data was obtained for Alberta in addition to locations of wolverines from a 2012–2015 study and other sources (Webb *et al*. 2016, p. 1,465; Webb 2017, pers. comm.).

Figure 6 presents the results from our spatial analysis and indicates a total of 77 wolverines were trapped in British Columbia wildlife management units within 110 km (68.35 mi) of the U.S.–Canada border from 2007–2015 (average of 8.5 animals per year) (Note: the gray areas depicted for British Columbia represent management units with zero harvest). We used this distance since it's similar to both the average maximum distance per dispersal movement of 102 km (63 mi) for male wolverines reported by Inman *et al*. (2012a, p. 784) for the Greater Yellowstone region of Montana, and a reported 100 km (62 mi) dispersal distance for a juvenile male for Ontario, Canada (COSEWIC 2014, p. 24, citing unpublished data from Dawson *et al*. 2013). As shown below, one management area contains nearly one-third (23 individuals) of this total number. The other management units along the international border indicate very few animals harvested over this 8-year period (i.e., areas on map identified as zero). There is no open trapping season or hunting season on wolverines in the management units in the Okanagan (Region 8) (north of Washington State) or South Coast (Region 2) (southwest corner of British Columbia) with a trapping season for wolverines only in the Kootenay (Region 4, the eastern half of the southern part of the province) (Weir 2017c, pers. comm.). In addition, there has not been an open trapping season in Region 2 since at least 1985 and since 1993 in the Okanagan region (Weir 2017d, pers. comm.). For Alberta, we identified a total of 15 wolverines harvested by trappers and data presented in other studies within 110 km (68.35 mi) of the U.S.–Canada border from 1989–2014 (average of less than 1.0 animal per year). As noted above, Regehr and Lacroix's (2016, entire) multi-method inventory of wolverines within an area located in the eastern side of the Coast Mountains of British Columbia (see black star in Figure 6) found unexpectedly high numbers of wolverines, which may have been the result of the rugged landscape features in this mountainous area and abundant food resources (both winter and summer) (Regehr and Lacroix 2016, pp. 249–250). Researchers in Canada are currently conducting a landscape level analysis to estimate the size and sustainable harvest for wolverine populations within British Columbia (Weir 2017a, pers. comm.).

We also requested information related to incidental trapping along the southern Alberta- U.S. border. Only one registered trapline is located 10 km (6.21 mi) north of the border with Montana and no wolverines have been reported in this region, which is primarily grassland habitat (Webb 2018a, pers. comm.). In addition, in Canada, if wolverines are trapped outside of a registered trapline, trappers are required to register that event; however, fur harvest records for southern Alberta do not indicate any trapped wolverines (Webb 2018a, pers. comm.).

Based on this analysis, trapping effort along the U.S.–Canada border does not represent a barrier to wolverine movement and dispersal along the international border.

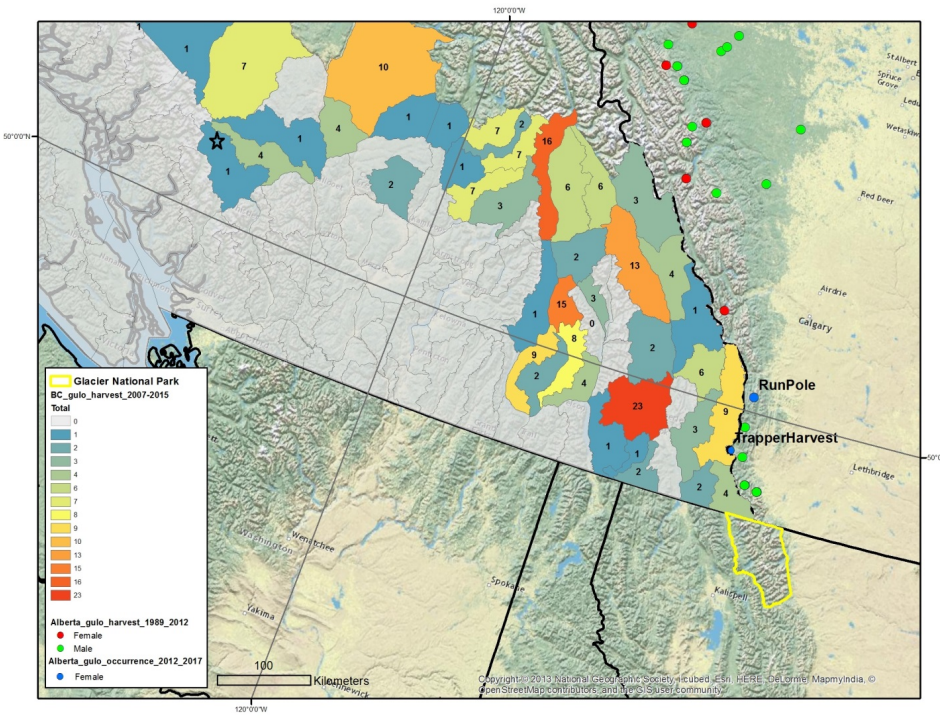


Figure 6. Numbers of wolverines harvested in British Columbia and Alberta, Canada. Sources: Weir 2017a, pers. comm.; Webb *et al*. 2016; Webb 2017, pers. comm.

Legal Status/Protection

In the western United States, the wolverine status is as follows: state-threatened species in Oregon (ODFW 2016) and California (CDFW 2017a); state-endangered species in Colorado (Colorado Parks and Wildlife 2015a); a candidate species in Washington (Washington Department of Fish and Wildlife (WDFW) 2013); protected nongame species and species of greatest conservation need in Idaho (IDFG 2014); protected animal and species of greatest conservation need in Wyoming (WGFD 2017); protected from collection, possession, and importation (Utah Administrative Rule R-657-3-24) and a species of greatest conservation need in Utah (Utah Wildlife Action Plan Joint Team 2015); furbearer and species of concern in Montana (Montana Natural Heritage Program and Montana FWP 2017); and, in Nevada, the Nevada Administrative Code lists wolverines as a protected mammal (NAC 503.030), which provides full legal protection. The State of New Mexico Department of Game and Fish does not recognize the wolverine as a native mammal. Additional discussion regarding State regulatory mechanisms that provide protections for wolverines is provided in **Appendix G**.

The Idaho Department of Fish and Game issues permits allowing live capture, handling, and release of wolverines for scientific studies, which usually involved log box-traps that do not cause physical injury to the captured animals (IDFG 2014, p. 27). The agency also issues scientific collection permits to various agencies and organizations and to IDFG biologists that can include the capture, chemical immobilization, and placement of radio-collars/radio-markers on wolverines (IDFG 2014, p. 27). These permittees (and IDFG staff) are required to comply with animal trapping and handling protocols approved by IDFG's Wildlife Health/Forensic Laboratory and other animal welfare and research institutions. This same process is followed by the State of Montana (Montana FWP 2017, pers. comm.). Over the past 20 years, there have been two documented wolverine deaths due to live capture activities in Idaho (IDFG 2014, p. 27), and one handling-related mortality in Montana (Montana FWP 2017, pers. comm).

In Wyoming, the Wyoming Game and Fish Commission (Commission) Regulation Chapter 52, Nongame Wildlife, authorizes 'take' of wolverine only for scientific or educational purposes as regulated by Commission Regulation Chapter 33 (Regulation Governing Issuance of Scientific, Research, Educational, or Special Purpose Permits). As of November 2017, the WGFD has issued five different Chapter 33 permits for research activities; four of the five are for live capture and one for camera trapping surveys (WGFD 2017, pers. comm.). No incidental take has been reported from permits issued, and no permits allowing for the take of wolverine have been issued in Wyoming (WGFD 2017, pers. comm.).

In California, research permits for State-listed, State-candidate, and fully protected species in California are issued as a Memorandum of Understanding (MOU). Currently, there are no active MOUs for research on wolverine in California (Burkett 2017, pers. comm.).

In Alaska, wolverines are dually classified; that is, depending on the method of take, they can be classified as big game or a furbearer (Parr 2018, pers. comm.). If a wolverine is taken under a hunting license, they are classified as "big game" and a person would have to abide by the hunting season/bag limits; if a wolverine is taken under a trapping license, they are classified as a "furbearer" and a person would have to abide by the trapping season/bag limits (Parr 2018, pers. comm.).

In Canada, provincial designations for the wolverine include Endangered in Labrador, and Threatened in Ontario and Québec ('Threatened' is equivalent to Endangered in Québec), with the remaining provincial designations ranging from no ranking to Sensitive or Special Concern and the Vancouver Island population designated as Imperilled (COSEWIC 2014, p. 44). Recovery planning for the wolverine is focused on the eastern population (Canadian Boreal Forest Agreement Secretariat 2015, p. 3).

In summary, overutilization does not currently represent a stressor to the wolverine in the contiguous United States at the individual, population, or species level. Wolverine populations in the contiguous United States are currently protected under several State laws and regulations. Hunting and trapping activities for wolverines are currently suspended or closed entirely for animals within the contiguous United States, though occasional incidental trapping can occur. Trapping in Montana, Alaska, and Canada has been and appears to be sustainable. Trapping or

harvesting of wolverines along the contiguous U.S.–Canada border does not represent a stressor to wolverines migrating into the contiguous United States at the individual or population level. In addition, wolverine populations along the Alaska–Canada border are continuous with the Yukon region of Canada, which suggests a rescue effect for Canadian populations along this international boundary (COSEWIC 2014, p. 37).

Summary of Current Conditions

Wolverine populations in much of North America are still recovering from large losses of individuals from intensive hunting and unregulated persecution pressures in the late 1880s into the mid-20th century. The distribution of wolverines within suitable habitat provides a better method for estimating population status than using abundance of animal, although there is limited rangewide survey information. Based on the best available information, wolverines continue to be detected within suitable habitat within the west-northwestern contiguous United States including Washington, Oregon, Idaho, Montana, and Wyoming (e.g., Aubry *et al*. 2016, pp. 14–15; Magoun *et al*. 2013, p. 27; Montana FWP 2017, pers. comm.). Studies are currently underway to estimate the species' current distribution and genetic characteristics of these populations. In Canada, the total wolverine population is estimated at over 10,000 adults (COSEWIC 2014, p. 47). In Alaska, estimates of populations are not available and are best evaluated based on density (Parr 2017, pers.comm.), which are naturally low for this species. Recent density estimates range from 5 to 10 wolverines per 1000 km^2 (386 mi^2) for Alaska (Parr 2017, pers. comm). The best available information does not indicate wolverine populations are currently negatively impacted by the lack of genetic diversity and there is no evidence that the population in the contiguous United States is declining.

We prepared a Current Potential Extent (of Occurrence) map to illustrate the species' current distribution in the contiguous United States (Figure 3). We estimated this area represents approximately 3.5 percent of the wolverine's Current Potential Extent in North America.

We determined that 72 percent of our Current Potential Extent of the wolverine in the contiguous United States is found on lands owned or managed by the Federal government (see **Appendix D**) We also evaluated Inman *et al*.'s (2013) previously modeled wolverine primary habitat in the contiguous United States and estimated that 96 percent are owned or managed by Federal agencies and 41 percent of this area is located in designated wilderness areas. **Appendix G**, *Regulatory Mechanisms and Conservation Measures*, provides a more detailed summary of management actions.

We evaluated several potential stressors that may be affecting wolverine populations or its habitat, including effects from roads, disturbance due to winter recreation and other activities, effects from wildland fire, disease and predation, and overutilization for (primarily) commercial purposes. We determined that the effects of roads (evaluated by number of miles, density, and location) and disturbance represent low level stressors to the wolverine in the contiguous United States. Wildland fire was determined to be a short-term stressor to wolverine habitat and its prey. Disease and predation are not considered stressors to the wolverine.

Legal trapping or hunting of wolverines is currently prohibited in the contiguous United States. Incidental trapping of wolverines is infrequent in the contiguous United States and, in Idaho and Montana, education programs are being implemented to reduce this stressor. In Alaska, the level of harvest of wolverines has been fairly consistent since 2010, and, as noted above, density estimates indicate no declining trend in wolverine populations.

Wolverines are harvested in several Canadian provinces with management and monitoring oversight based on spatial and temporal elements. We reviewed trapping information from Canada (within 110 km (68.35 mi) of the contiguous U.S.–Canada border) to assess potential impacts to dispersing wolverines into the United States. We found that, in Alberta, 15 wolverines were harvested over a 25 year period (average of less than 1.0 animal per year), and, for British Columbia, we found an average of 8.5 animals per year, though one management area contained nearly one-third (23 individuals) of this total (Weir 2017b, pers. comm.). Researchers in Canada are currently conducting a landscape level analysis to estimate the size and sustainable harvest for wolverine populations within British Columbia (Weir 2017a, pers. comm.). Based on the best available commercial and scientific information, overutilization does not represent a stressor to the wolverine in the contiguous United States.

Status – Future Conditions

The future timeframe evaluated in our analysis is approximately 38~~40~~ to 50 years, which captures our best professional judgment of the projected future conditions related to trapping/harvesting, climate change, or other potential cumulative impacts. We note here that evaluations of future conditions for species have an inherent level of uncertainty relative to demographic risks, particularly those related to climate change projections.

After considering the current conditions for the wolverine and its habitat, we describe here the most likely future scenario to potentially have an effect on wolverine at the population level in the contiguous United States:

- Climate change effects (i.e., significantly elevated temperatures resulting in decline in snowpack) may modify suitable habitat (including reproductive denning habitat), which could also change the scope of the wildland fire stressor.

Based on our review of the best available information, we determined that there were no other plausible scenarios that were likely to have population level impacts to wolverine in the contiguous United States. We expect that the effects of trapping and roads, human disturbance, effects of wildland fire to continue to be at low levels in the future. We have no information that indicates that mortality from roads or disease would increase within the range of wolverine in the contiguous United States in the future.

Climate Change Effects

In this section, we consider climate changes that may affect environmental conditions that the wolverine relies on. As defined by the Intergovernmental Panel on Climate Change (IPCC), the term “climate” refers to the mean and variability of different types of weather conditions over time, with 30 years being a typical period for such measurements, although shorter or longer

periods also may be used (IPCC 2013a, p. 1,450). The term "climate change" thus refers to a change in the mean or the variability of relevant properties, which persists for an extended period, typically decades or longer, due to natural conditions (e.g., solar cycles) or human-caused changes in the composition of atmosphere or in land use (IPCC 2013a, p. 1,450).

Scientific measurements spanning several decades demonstrate that changes in climate are occurring. In particular, warming of the climate system is unequivocal and many of the observed changes in the last 60 years are unprecedented over decades to millennia (IPCC 2013b, p. 4). The change in temperature reported in the Northern Hemisphere in recent history (past 150 years) at +0.6°C (1.08°F) is twice the change reported for the Southern Hemisphere (+0.3°C (0.54°F)) and there is much year-to-year variation (Post 2013, p. 4). With regard to precipitation over land, there has been a decline in global total annual precipitation, but the variability between years in total precipitation has increased since about the 1970s (Post 2013, p. 9). The Palmer Drought Severity Index (PDSI) compares the actual amount of precipitation received in an area during a certain time period with the normal or average amount expected during that same period (National Weather Service (NWS) 2015) and is generally used as a measure of water stress. Time series analysis of the PDSI indicates worsening persistent drought-like or drought-potential conditions across the globe since 1980, a reflection of the influence of temperature on atmospheric dynamics (Post 2013, pp. 10–11).

Comprehensive assessments of other observed and projected changes in climate and associated effects and risks, and the basis for them, are provided for global and regional scales in recent reports issued by the IPCC (2013c, 2014), and similar types of information for the United States and regions within it can be found in the National Climate Assessment (Melillo *et al*. 2014, entire). Results of scientific analyses presented by the IPCC show that most of the observed increase in global average temperature since the mid-20th century cannot be explained by natural variability in climate and is "extremely likely" (defined by the IPCC as 95 to 100 percent likelihood) due to the observed increase in greenhouse gas (GHG) concentrations in the atmosphere as a result of human activities, particularly carbon dioxide emissions from fossil fuel use (IPCC 2013b, p. 17 and related citations).

Scientists use a variety of climate models, which include consideration of natural processes and variability, as well as various scenarios of potential levels and timing of GHG emissions, to evaluate the causes of changes already observed and to project future changes in temperature and other climate conditions. Model results yield very similar projections of average global warming until about 2030, and thereafter the magnitude and rate of warming vary through the end of the century depending on the assumptions about human population levels, emissions of GHGs, and other factors that influence climate change. Thus, absent extremely rapid stabilization of GHGs at a global level, there is strong scientific support for projections that warming will continue through the 21st century, and that the magnitude and rate of change will be influenced substantially by human actions regarding GHG emissions (IPCC 2013b, 2014; entire).

Global climate projections are informative, and, in some cases, the only or the best scientific information available. However, projected changes in climate and related impacts can vary substantially across and, as noted above, within different regions and hemispheres (e.g., IPCC 2013c, 2014; entire) and within the United States (Melillo *et al*. 2014, entire). Therefore, we use

"downscaled" projections when they are available and have been developed through appropriate scientific procedures, because such projections provide higher resolution information that is more relevant to spatial scales used for analyses of a given species (see Glick *et al*. 2011, pp. 58–61, for a discussion of downscaling).

Further, future projections are generally summarized for a given future scenario (e.g., Representative Concentration Pathways (RCP) 8.5 or RCP 4.5) over a range of future climatological time periods, such as temperature change in 2040–2079 or 2070–2099 relative to 1980–2009 (U.S. Global Change Research Program (USGCRP) 2017, p. 139). This approach has an advantage of developing projections for a specific time horizon, however, the uncertainty in future projections is relatively high; that is, it incorporates both the uncertainty that results from multiple scenarios as well as uncertainty relative to the response of the climate system to human-caused emissions (USGCRP 2017, p. 139). Additionally, as one goes further out in time for these projections, the uncertainties increase (USGCRP 2017, p. 139). Therefore, analyses of projected changes use these transient, scenario-based simulations for a given global mean temperature threshold by extracting a time slice (typically 20 years) that is centered around the point in time at which that change is reached (USGCRP 2017, p. 139; Figure 4.2). A 30-year period is commonly used to better characterize the background state of observed climate around which anomalous conditions and even extremes occur (Arguez *et al*. 2012, p. 1,687).

M~~We note here that m~~ultiple lines of evidence, not just projections derived from quantitative models, should be examined when conducting climate vulnerability assessments (Michalak *et al*. 2017, entire). Thus, we provide below projected effects from climate change in the western United States relative to both abiotic (e.g., temperature, precipitation, snow cover) and biotic (e.g., phenology, behavior) factors.

Abiotic Factors

California

Regional temperature and precipitation observations for assessing climate change are often used as an indicator of how climate is changing. For evaluating climate trends in California, the Western Regional Climate Center (WRCC) has defined 11 climate regions (Abatzoglou *et al*. 2009, p. 1,535). The relevant region for our assessment is the north/north-central Sierra Nevada region (Tahoe National Forest), currently occupied by a male wolverine, or the northeast region as defined in Abatzoglou *et al*. (2009, p. 1,535) .

Two indicators of temperature, the increase in mean temperature and the increase in maximum temperature, are important for evaluating trends in climate change in California. For the climate region that encompasses the Tahoe National Forest region, the 100-year linear trends provided by the WRCC indicate an increase in mean temperatures (Jan–Dec) of approximately 0.92°C/100 yr (± 0.29°C/100 yr) (1.66°F ± 0.53°F/100 yr) since 1895 from present day; 1.55°C/100 yr (± 0.67°C/100 yr) (2.79°F ± 1.21°F/100 yr) since 1949 to present day; and 2.41°C/100 yr (±1.54°C/100 yr) (4.33°F ± 2.78°F/100 yr) since 1975 to present day (WRCC 2017). Thus, the increase in mean temperature has not been constant—the rate of increase over the past 42 years in this region has been 2.6 times higher than the past 122 years. We assume the rate of

temperature increase for this region is higher for the second and third time periods (since 1949 and 1975, respectively) than for the first time period (since 1895) due to the increased use of fossil fuels in the later part of the 20th and early 21st century.

Although these observed trends provide information as to how climate has changed in the past, climate models can be used to simulate and develop future climate projections. Both state-wide and regional probabilistic estimates of temperature and precipitation changes for California (by the 2060s) using downscaled data from 16 global circulation models and 3 nested regional climate models were presented by Pierce *et al*. (2013, entire). The study looked at a historical (1985–1994) and a future (2060–2069) time period using the IPCC Special Report on Emission Scenarios A2 (Pierce *et al*. 2013, p. 841), which is an IPCC-defined scenario used for the IPCCs Third and Fourth Assessment reports, and is based on a global population growth scenario and economic conditions that result in a relatively high level of atmospheric GHGs by 2100 (IPCC 2000, pp. 4–5; see Stocker *et al*. 2013, pp. 60–68, and Walsh *et al*. 2014, pp. 25–28, for discussions and comparisons of the prior and current IPCC approaches and outcomes). Importantly, the projections by Pierce *et al*. (2013, pp. 852–853) include daily distributions and natural internal climate variability.

Simulations using these downscaling methods project an increase in *yearly* temperature for the area that encompasses the Tahoe National Forest (Sierra Nevada) ranging from 2.1°C (3.78°F) to 3.2°C (5.76°F) by the 2060s time period (Pierce *et al*. 2013, p. 844), compared to 1985–1994. The simulations indicated a yearly upper temperature increase of 2.5°C (4.5°F) from 1985–1994 to 2060–2069 (averaged across models) for this area, and an increase of 1.9°C (3.42°F) for the December–February period (Pierce *et al*. 2013, p. 842).

Beginning in 2012 and continuing into 2016, California experienced a severe drought throughout most of the state (Griffin and Anchukaitis 2014, p. 9,020). Although three-year droughts in California are not unusual when evaluated over the past 1000 years, the severity of these drought conditions during this period was demonstrated in the 2014 summer PDSI, which was estimated to be the lowest on record (1901–2014) (Williams *et al*. 2015, p. 6,823). An evaluation of how unusual this drought event was in the context of the last millennium using blue oak (*Quercus douglasii*) tree ring data from four sampling sites (with additional tree sampling following the 2014 growth season) was conducted by Griffin and Anchukaitis (2014, entire). Their paleoclimate drought and precipitation reconstructions for Central and Southern California show that, although the precipitation during this drought has not been anomalously low, it was not outside the range of variability (Griffin and Anchukaitis 2014, p. 9,017). However, the 2014 drought was the worst single drought year of at least the last 1,200 years in California and the 2012–2014 drought was the most severe of three consecutive drought years, based on three events found in the record for the last 1,200 years (Griffin and Anchukaitis 2014, pp. 9,020–9,021). The study concluded that low precipitation combined with high temperatures was responsible for creating this worst short-term drought episode (Griffin and Anchukaitis 2014, pp. 9,021–9,022).

A study by Williams *et al.* (2015, entire) estimated the anthropogenic contribution to California's drought during 2012–2014. They found that the intensifying effect of high potential evapotranspiration on this drought event (measured by summer PDSI) was almost entirely the

result of high temperatures (18–27 percent in 2012–2014; 20–26 percent in 2014) (Williams *et al*. 2015, p. 6,825). Another study evaluating the influence of temperature on the drought in water year 2014 in California found that, although the low level of precipitation was the primary driver for the drought conditions, temperature was an important factor in exacerbating the drought, noting that the water year 2014 was the third year of the multiyear drought event and therefore conditions were drier than normal at the beginning of the water year (Shukla *et al*. 2015, p. 4,392).

In sum, these projections indicate that increased temperatures are likely to occur in the Tahoe National Forest region by the 2060s due to the effects of climate change.

Precipitation patterns can also be used as an indicator of potential climate change. We obtained yearly snowfall data for the Tahoe City station located in the northern Sierra Nevada region from the WRCC (https://wrcc.dri.edu/cgi-bin/cliMAIN.pl?ca8758) since that dataset was the most complete for the area. We then conducted a nonparametric correlation test, the Mann-Kendall statistical test (Hipel and McLeod 1994, pp. 63–64, 856–858), which is commonly used for analyzing climatic time series (e.g., Ahmad *et al*. 2015, entire), to evaluate trends in snowfall over time. This analysis was conducted using the R and R Studio software programs (Version 3.1.2; R Development Core Team, 2014) with the "Kendall" package (Version 2.2) (McLeod 2011). We found that annual snowfall amounts showed no statistically significant trend (increasing or decreasing) from 1909–2017 (tau = –0.0289, two-sided *p*-value of 0.6705) for the Tahoe City station.

State-wide and regional probabilistic estimates of precipitation changes for California were also evaluated by Pierce *et al*. (2013, entire). When averaged across all models and downscaling methods, small annual mean decreases in precipitation were found for the Sierra Nevada region of California (by the 2060s), but that study also found an increase in precipitation for the December through February period (wetter winters) (Pierce *et al*. 2013, pp. 849, 855). There was significant disagreement across the models, with percent changes ranging from a 12 percent decrease to a 9 percent increase (Pierce *et al*. 2013, p. 851), and the projected changes in seasonal-mean precipitation were generally small when compared to natural internal climate variability (Pierce *et al*. 2013, p. 850; Figure 9).

Columbia River Basin Region

Within the Columbia River Basin, Rupp *et al*. (2017, entire) evaluated climate change effects using simulations from 35 Global Climate Models (GCMs). This region covers a large area within eastern and central Washington, east-northeastern Oregon, and most of Idaho, and parts of British Columbia, Canada (Rupp *et al*. 2017; Figure S1), and encompasses portions of the Current Potential Extent area of the wolverine. The simulations provided projections of climate in the Columbia River Basin into the 2080s under two emissions scenarios, Representative Concentration Pathways (RCP) (RCP 4.5, which represents moderate reduction in GHG emissions ("intermediate emissions"), and RCP 8.5, which represents a continued increase in GHG emission "high emission"). The results of their multi-model ensemble for the RCP 4.5 scenario indicate mean annual temperature increases (above Bonneville Dam), above the 1970–1999 baseline average, of 1.3°C (2.34°F) for the 2010–2039 period, 2.3°C (4.14°F) for the 2040–

2069 period, and 2.8°C (5.04°F), for the 2070–2099 future period (Rupp *et al*. 2017, p. 1,788). By season, the winter period (December–February) mean change result indicates an increase of 1.1°C (2.52°F) for 2010–2039, 2.2°C (3.96°F) for 2040-2069, and 2.7°C (4.86°F) for 2070–2099, as compared to the 1970–1999 baseline average (Rupp *et al*. 2017, p. 1,788).

For the RCP 8.5 scenario, the multi-model ensemble projections indicate mean annual temperature increases, above the 1970–1999 baseline average, of 1.4°C (2.34°F) for the 2010–2039 period, 3.1°C (5.58°F) for the 2040–2069 period, and 5.0°C (9.0°F), for the 2070–2099 period (Rupp *et al*. 2017, p. 1,788). For the winter season (December–February) mean change increase of 1.4°C (2.34 °F) for 2010–2039, 2.9°C (5.22°F) for 2040-2069, and 4.7°C (8.46°F) for 2070–2099, as compared to the 1970–1999 baseline average (Rupp *et al*. 2017, p. 1,788). The anthropogenic-forced (human caused or influenced) change for these projections is higher than the annual variability; thus, by the year 2050, it is very unlikely that the temperature for this year or any year following during this century would be as low as the historical average (Rupp *et al*. 2017, p. 1,788).

Precipitation projections were much less robust; the multi-model ensemble mean precipitation projections indicate an increase above baseline of up to 8 percent by 2099 for RCP 8.5 and slightly less for RCP 4.5 (Rupp *et al*. 2017, p. 1,788). When viewed seasonally, for the winter season, the ensemble projections indicate increases in precipitation for all three future time periods for both the RCP 4.5 and RCP 8.5 scenarios (ranging from 3 to 14 percent) as compared to the baseline period (1970–1999) (Rupp *et al*. 2017, p. 1,788). The projections of anthropogenic-forced increases in precipitation are lower than the interannual variability; however, despite these increases, the authors indicate that years of anomalously low precipitation relative to baseline would be expected with high frequency throughout the 21st century (Rupp *et al*. 2017, p. 1,788).

Within three subregions of the Pacific Northwest within the Current Potential Extent of the wolverine, Sheehan *et al*. (2015, p. 20; Table 4) also found that, when compared to a historical baseline (1971–2000), all future climate projections (RCP scenarios 4.5 and 8.5; 2036–2066, 2071–2100) indicate a rise in both minimum and maximum monthly temperatures, and a generally positive change in mean annual precipitation, though the latter results varied across projections.

Northern Cascades (Washington)

The North Cascadia Adaptation Partnership (NCAP) prepared a vulnerability assessment and adaptation planning process for the North Cascade Range (defined here as Mount Rainier north to the Canadian border) (Raymond *et al*. (eds.) 2014). In this region, climate is driven by the regional climate of the Pacific Northwest and is mediated by local effects of mountainous terrain (e.g., orientation and elevational changes) and the region's proximity to the Pacific Ocean (i.e., El Niño-Southern Oscillation and Pacific Decadal Oscillation circulation patterns) (Littell and Raymond 2014, p. 25). Historical records indicate an increase in mean annual temperature of about 0.7° to 0.9°C (1.26°F to 1.62°F) in the Pacific Northwest between 1930 and 1995 (Mote 2003, p. 271). Similarly, using U.S. Historical Climatology Network stations from the Pacific Northwest and Columbia Basin region, Littell *et al*. (2011, p. 10) found increases in both

minimum and maximum temperatures of 0.18°C (0.324°F) per decade for a total of 1.0°C (1.8°F) for the period of 1950 to 2006. For the Pacific Northwest, annual precipitation increased slightly over the 1920–2000 period (Mote 2003, p. 277), but precipitation was found to be more variable relative to the mean than temperature; that is, trends in precipitation are small compared to the interannual variability (Littell and Raymond 2014, p. 27).

For this vulnerability assessment, scientists from the University of Washington, Climate Impacts Group, and partners developed datasets of downscaled climate and hydrologic projections. Details regarding methods and results, as well as archives of the data in grid and summarized forms, are available at http://cses.washington.edu. For projections of most climatic variables, the analysis by Littell *et al*. (2011) was used (Littell and Raymond 2014, p. 30). An ensemble containing a subset of 10 GCMs was used in combination with previously defined IPCC emission scenario A1B (i.e., a medium-high emission scenario that represents an increase in GHGs in the early 21st century followed by substantial reductions in the second half of the 21st century) to evaluate projections of future climate (Littell and Raymond 2014, p. 32). Projections indicate an increase in temperature, on average, of 2.1°C (3.8°F) by the 2040s (average of years in the 30-year window from 2030 to 2059) and 3.5°C (6.3°F) by the 2080s (2070–2099) for the Pacific Northwest/Columbia Basin region (Littell *et al*. 2011, p. 20). Projections of future precipitation for the Pacific Northwest/Columbia Basin region ranged widely across the models, with some models projecting higher annual precipitation and others lower (Littell et al. 2011). For the A1B scenario, the 10-model ensemble mean found no change in annual precipitation for the 2040s and a 2 percent increase for the 2080s and continued high interannual variability (Littell et al. 2011, p. :Table 3.1).

Upper Snake River Basin

The Upper Snake River Tribe Foundation and its Tribal members prepared a climate change vulnerability assessment for the Upper Snake River Watershed (Petersen *et al*. 2017, entire). The assessment covers large areas within our Current Potential Extent for wolverine in the contiguous United States, including southern and central Idaho, and eastern Oregon, western Wyoming, as well as small areas of northern Nevada and northern Utah (Petersen *et al*. 2017, p. 10; Figure 7). Within three geographic/model domains of this larger region, downscaled climate projections were created from 20 GCMs run with two emissions scenarios (RCPs 4.5 and 8.5) and these outputs were then used to calculate potential future changes in temperature and precipitation (Petersen *et al*. 2017, pp. 15–16). The projections were analyzed in reference to a baseline period (1950-2005) for three future time periods—the 2030s (2020–2049), the 2050s (2040–2069), and the 2080s (2070–2099) (Petersen *et al*. 2017, p. 16).

For temperature, their projections indicated an increase in average annual temperatures in both future emission scenarios and across all time periods. Under RCP 8.5 (high emissions scenario), the ensemble mean temperature increase was about 6.11°C (11ºF), and 2.78°C (5ºF) under the RCP 4.5 intermediate emissions scenario across all three geographic/model domains (Petersen *et al*. 2017, Appendix A, p. 2). For the North and East domains (areas with greater topographical variability), there was some indication of a small increase in total annual precipitation by the end

of the century, though there was less agreement among the models (Petersen *et al*. 2017, Appendix A, p. 2).

For all geographic/model domains, the average temperature is projected to increase under both emissions scenarios for all seasons (Petersen *et al*. 2017, Appendix A, p. 2). For the winter months (December, January, February), for RCP 4.5 scenario, the average seasonal temperature is projected to increase by 3.89 to 5°C (7 to 9ºF) by the end of the century, and an increase of approximately 2.22 to 3.33°C (4 to 6ºF) for the other seasons (Petersen *et al*. 2017, Appendix A, pp. 2, 6). The winter season projections for RCP 8.5 add an additional 1.67 to 2.22°C (3 to 4ºF) by the end of the century (Petersen *et al*. 2017, Appendix A, pp. 2, 6).

Rocky Mountain Region (Colorado)

A report by Lukas *et al*. (2014, entire) presented an assessment of observed and future projections of climate change effects for Colorado. They reported that, statewide, annual average temperatures have increased by 1.1°C (2.0°F) over the past 30 years, and 1.4°C (2.5°F) over the past 50 years (Lukas *et al*. 2014, p. 11). These warming trends have been observed in much of the State (Lukas *et al*. 2014, p. 11). They report no significant long-term trends in annual precipitation (30-, 50-, and 100-year trends) through 2012, but they indicate an observed trend towards more severe soil-moisture drought conditions in Colorado, based on the PDSI, over the past 30 years (Lukas *et al*. 2014, pp. 12, 21).

This report also presents results from climate change modeling using an ensemble of Coupled Model Intercomparison Project (CMIP5) model projections, run with RCP 4.5 and 8.5 scenarios (Lukas *et al*. 2014; Section 5). The results indicate future warming in Colorado for all of the climate model projections (Lukas *et al*. 2014, p. 59). By 2050, for the RCP 4.5 (intermediate) emissions scenario, the statewide average annual temperatures are projected to increase by 1.4 to 2.8°C (2.5 to 5°F) (relative to a 1971–2000 baseline), and increase by 1.9 to 3.6°C (3.5 to 6.5°F) under the RCP 8.5 (high) emissions scenario (Lukas *et al*. 2014, p. 59). For precipitation, they report that climate model projections show less agreement regarding future precipitation change for Colorado, but most projections indicate increasing winter precipitation by 2050 (Lukas *et al*. 2014, p. 59).

Northern Rocky Mountain Region (Montana)

An assessment of Montana's climate trends and future projections is presented in the 2017 Montana Climate Assessment (Whitlock *et al.* 2017, entire). Among its findings are observed increases in annual average temperatures (daily minimums, maximums, and averages) from 1950–2015, with a range from 1.1–1.7°C (2.0–3.0°F) across the state, or about 0.3°C (0.5°F) per decade (Whitlock *et al.* 2017, no page number). No historical changes in average annual precipitation was reported for the 1950–2015 time period, but changes in average seasonal precipitation include a decrease in average winter precipitation of 2.3 cm (0.9 in) which was primarily attributed to natural variability and El Niño events (Whitlock 2017, no page number).

Climate projections for Montana reported in this assessment indicate continued warming into the 21st Century. Results from climate change modeling (using an ensemble of 20 individual GCMs

from the fifth iteration of the CMIP5), indicate that by mid-century (2040–2069), annual average temperatures under the RCP 4.5 emissions scenario are projected to increase by about 2.5°C (4.5°F) for most of the state, and by 3.1°C (5.6°F) at end-of-century (2070–2099) (Whitlock *et al*. 2017, Figures 2-9, 2-10). For the RCP 8.5 emissions scenario, average annual temperatures are projected to increase by about 3.3°C (6.0°F) by mid-century and by 5.4°C (9.8°F) at end-of-century (Whitlock *et al*. 2017, Figures 2-9, 2-10).

Average annual precipitation is projected to increase across Montana for both the mid-century and end-of-century projections for both emission scenarios (Whitlock *et al*. 2017, Figures 2-16–2-18), which varies by geographic region and by season/month. As an example, for the Glacier National Park region, under RCP 4.5 scenario, precipitation (average annual) is projected to increase by about 2.3 cm/year (0.9 in/year) for the RCP 4.5 scenario by mid-century (Whitlock *et al*. 2017, Figure 2-17). For end-of-century projections, the projections produced from the GCMs ensemble showed large differences, but indicated a general positive trend for both emission scenarios (Whitlock *et al*. 2017, no page number). An increase in winter month precipitation is projected across the state for both emissions scenarios and for both mid-century and end-of-century time frames (Whitlock *et al*. 2017, Figures 2-20, 2-21).

Summary

Observed trends and future climate model projections indicate warming temperatures for much of the western United States, including areas within the Current Potential Extent of the wolverine. The degree of future warming varies by region and is dependent upon the future emission scenario used during the modeling process. Future precipitation trends are less certain for many regions, in part, due to naturally high, inter-annual variability; some regions are projected to experience greater winter precipitation. Wolverines have been found to have a wide range in critical temperature depending on season, and undergo seasonal changes in fur insulation to adapt to warmer temperatures in summer. Wolverines also exhibit changes in behavior, such as moving to higher elevations in summer months. Wolverines continue to occupy areas that have exhibited increases in temperature (e.g., California, parts of Montana and Washington); however, no empirical studies have evaluated these physiological and behavioral adaptations, including sub-lethal effects, relative to warming temperatures.

Biotic Factors

In addition to evaluating changes in these abiotic factors, biotic interactions should be considered in evaluating species' response to climate change (reviewed by Post 2013). Although abiotic changes drive ecological processes, the alterations in biotic interactions (e.g., competition among conspecifics, interactions with competitors, resources, and predators) represent the ecological responses that result from those changes (Post 2013, p. 1). Changes in certain abiotic factors, such as snow and ice cover, should also be considered in an ecological context since they represent habitat for many species (Post 2013, p. 11).

Ecological studies evaluating the effects of climate change often evaluate phenology, the timing of life history events and how they vary in space and time, generally at the population or site-specific level, though phenological variation at the individual level may also be important (Post

2013, p. 54). Previous meta-analyses of the rate of phenological advancement have suggested advances of between 2–5 days per decade, across taxa, and between low-mid to mid-high latitudes (Post 2013, p. 59). A more recent meta-analysis from Cohen *et al*. (2017, p. 4) found, on average, significant advancement in the phenology of animals since 1950, advancing by about 2.88 days per decade and 3.08 days per degree Celsius.

Within the Pacific Northwest region, Ford *et al*. 2016 (entire) modeled the timing of growth initiation in coast Douglas-fir trees (*Pseudotsuga menziezii* var. *menziezii*) within the species' range in Washington and Oregon to evaluate its ability to track changes in climate with changes in phenology. This study found that, for high latitudes and elevations, growth initiation was predicted to occur earlier in the year, which allows trees to track the beginning of favorable growing conditions, without exposure to frost risk (i.e., adaptive phenological response) (Ford *et al*. 2016, pp. 3718, 3,721). Conversely, their model predicted that at lower latitudes and elevations, growth initiation will lag behind climate change shifts due to reduced chilling with lower productivity, which suggested that coast Douglas-fir has an obligate chilling requirement for height (but not diameter growth initiation) (Ford *et al*. 2016, pp. 3,717–3,719).

Another study reported on the effects of encroachment of woody plants (willows (*Salix* sp.)) in alpine environments to alpine wildflowers and their pollinators due to temporal overlap in flowering phenology, which may result in establishment of plant species with broader environmental tolerance in high alpine ecosystems (Kettenbach *et al*. 2017, p. 6,969). Similarly, in Sweden, Wilson and Nilsson (2009, entire) reported on encroachment of woody vegetation in arctic-mountain habitat in response to observed temperature increase of 2.0°C (3.6°F) over 20 years, though this increase in cover was observed primarily at lower elevations (Wilson and Nilsson 2009, p. 1,682).

A high-latitude, North American study evaluated the effect of weather and broad-scale climate variables and vegetation productivity on the timing of spring and fall migrations of migratory caribou herds in northern Québec and Labrador, Canada (Le Corre *et al*. 2017, entire). That study found that, since 2000, except for the spring arrival, migrations occurred earlier, and were affected by resource availability, likely through intraspecific competition factors (Le Corre *et al*. 2017, p. 266).

In addition to phenological changes related to habitat variables or reproduction patterns, the effects of climate change may affect food resources important to wolverine, either directly (e.g., survival) or indirectly (e.g., effects to their habitat). An early study by Wang *et al*. (2002, p. 217) projected a potential increase in ungulate populations in Rocky Mountain National Park (Colorado) under future climate scenarios due to enhanced survival and recruitment of juvenile animals in response to less severe winters. The authors note that their results should be interpreted qualitatively given the uncertainties in applying climate change scenarios based on global models to ecological systems at the local scale (Wang *et al*. 2002, p. 217). In addition, they report that vegetation response (e.g., succession) to climate change effects may result in changes to ungulate habitat (Wang *et al*. 2002, p. 219). Overall, the study concluded that their results were consistent with those reported in other studies that have evaluated the relationships between the effect of weather and density dependence and ungulate population dynamics (Wang *et al*. 2002, p. 219).

Summary

The results presented above indicate biotic effects resulting from climate change, varying from phenological changes to shifts in vegetation and vegetation succession. We are unaware of studies that have directly evaluated these types of effects to the North American wolverine or its habitat. Given the relatively large area (see Figure 3) and varied habitats occupied by wolverines in the contiguous United States, the shifts in vegetation are likely to be relatively narrow in scope and scale. Furthermore, we have no information to suggest that wolverines selectively use any specific vegetation type and some projected changes in vegetation may be advantageous for wolverine prey.

Climate Change and Potential for Cumulative Effects

Threats can work in concert with one another to cumulatively create conditions that may impact the wolverine or its habitat beyond the scope of each individual threat. Given an expected increase in temperature in the western United States, the best available information indicates that, if there are any cumulative impacts in the future, the most likely to have population level effects on wolverine in the contiguous United States could be: 1) changes in snowpack from the combination of increased temperature and changes in precipitation patterns, or 2) changes in snowpack and increase in wildland fire potential. The best available information does not indicate that the effects of trapping and mortality from roads will act cumulatively with effects of climate change, and those stressors are expected to remain low level impacts into the future.

Snowpack/Snow Cover

The key conditions for the accumulation of snow are sufficiently cold air temperature to support precipitation in the form of snow, and the persistence of temperature below freezing to preserve the snowpack (Sospedra-Alfonso *et al*. 2015, p. 4,429). Several studies have used air temperature and precipitation to explain snowpack in North America and Europe (e.g., Cayan 1996; Hamlet *et al*. 2005; Mote 2006; Brown and Mote 2008; Beniston 2012; Morán-Tejeda *et al*. 2013), in part because air temperature is considered a good proxy indicator of snowmelt (Anderson 2002, p. 7-4) and the general lack of radiation observations, while air temperature and precipitation are more readily available from climatological and hydro-meteorological networks (Sospedra-Alfonso *et al*. 2015, p. 4,429).

Elevational relationships with snowpack with temperature and precipitation have also been examined. In the western United States, little historical change in snowpack (measured as snow water equivalent) was found above approximately 2500 m (8,202 ft) elevation despite observed warming trends (Regonda *et al*. 2005, p. 377). In a more recent study, Sospedra-Alfonso *et al*. (2015, entire) evaluated snowpack variability relative to elevation in the Rocky Mountain region of northern Idaho and northwestern Montana. The authors found that, historically, temperature has been a larger driver of April (aggregated months) snowpack below about 1,560 m (5,118 ft) elevation, with precipitation the main driver of variability above that elevation (Sospedra-Alfonso *et al*. 2015, p. 4,436). Another study, using a single climate change scenario and a high-resolution weather model, found that the threshold elevation below which temperature dominates

snowpack rises by about 250 m (820 ft) in the Colorado (Southern) Rockies, and about 191 m (627 ft) in the Northern Rockies (near Glacier National Park) (Scalzitti *et al*. 2016, p. 5,364). In the Northern Cascades in Washington, the snow water equivalent (SWE) (the water content of snowpack, expressed as depth) for April 1 has declined 15 to 35 percent from the middle of the 20th century to 2006 (Mote *et al*. 2008, p. 193). Larger declines in SWE at low-elevation (below 1,200 m (3,937 ft)) stations have been reported along with smaller declines or increases in SWE at high-elevation (above 1,200 m (3,937 ft)) stations (Mote 2006, p. 6,215; Mote *et al*. 2005, pp. 41–43; Mote *et al*. 2008, pp. 193, 203-204); that is, the warming temperatures are not high enough at higher elevations to cause a change in precipitation from snow to rain (Littell and Raymond 2014, p. 29). However, trends in SWE have also found to be highly variable for both SNOTEL stations (automated snow telemetry locations monitored by the U.S. Department of Agriculture National Resources Conservation Service) and manual snow course stations (Littell and Raymond 2014, p. 29) and many of these stations are at lower elevations.

Topography is also an important factor affecting ~~effecting~~ seasonality, precipitation amount, and potential trends in snow cover (Luce 2017, p. 70). The range of potential changes to climate is complex, particularly for the Northern Rockies, since current climatological settings vary over the landscape at both macroscales and fine scales (Luce 2017, p. 70). In this region, there are broad east to west changes in precipitation seasonality and amount, and thus local differences are found between nearby mountain and valley weather stations; therefore, the trends and drivers for climate variations will differ greatly from east to west. In general, the interior parts of the Northern Rockies (higher elevations) are thought to be cold enough to be relatively insensitive to warming temperatures, but strongly sensitive to variation in precipitation (Luce *et al*. 2017, p. 70). In contrast, the lower elevation mountains found in northern Idaho, which are more heavily influenced by a maritime snow climate (Mock and Birkeland 2000, pp. 2,369–2,370; Figure 2), are sensitive to temperature variability, particularly with respect to snow residence time (Luce *et al*. 2014; Figure 11). Below we provide summaries of projected changes to snowpack/snow cover relevant to the Current Potential Extent of wolverine in the contiguous United States, including an in-depth analysis for the Northern and Southern Rocky Mountains.

Upper Snake River Watershed (Pacific Northwest region)

The Upper Snake River Tribal Foundation assessment (discussed above) included projected changes in snowpack for three locations in the Upper Snake River watershed, including areas located within our estimated Current Potential Extent of the wolverine (from Climate Impacts Group Pacific Northwest (PNW) Hydroclimate Scenarios Project (2860); http://warm.atmos.washington.edu/2860/products/sites/). Model results, based on snow water equivalent (SWE), indicate a projected loss in April 1 snowpack of 36 percent for the 2030–2059 period and 64 percent for the 2070–2099 period for the *Salmon River at White Bird* location (average of percent change across all models relative to the long-term average for 1916–2006 ("historical period")). For the *Snake River at Brownlee Dam* location, the projected loss is 37 percent for the 2030–2059 period and 64 percent for the 2070–2099 period (summary presented in Petersen *et al*. 2017, p. 20). These projected changes were found to be consistent with overall changes projected for the Columbia River Basin snowpack in an earlier study. Hamlet *et al*. (2013, p. 404; Figure 7) found that, relative to the long-term average for 1916 to 2006, the April 1 snowpack in the Columbia River Basin is projected to decline by 29% for the 30-year period

spanning 2030-2059 and decline by 52% for the period spanning 2070-2099 for the A1B emissions scenario. [Note: the A1B emission scenario represents a more balanced energy portfolio than RCP 8.5, with GHG emissions leveling off by the middle of the 21st century].

Northern Cascades (Washington)

As noted above, the North Cascadia Adaptation Partnership (NCAP) prepared a vulnerability assessment and adaptation planning process for the North Cascade Range, which included an assessment of projected snowpack conditions (Raymond *et al*. (eds.) 2014). In the Northern Cascades, snowpack (measured as April 1 SWE) was modeled at an approximate 800 m (0.5 mi) resolution of latitude and longitude (Strauch *et al*. 2014, p. 45). Projections (using A1B emissions scenario and three model configurations) indicate considerable decreases by the 2040s when compared to historical levels (1916–2006) (Strauch *et al*. 2014, p. 59; Figure 4.6). The projected reductions in SWE are higher in warmer areas west of the Cascade crest and at low elevations in eastern Washington; whereas, higher elevations are expected to continue to retain snow cover in early summer (Strauch *et al*. 2014, p. 59). In addition, projections for the date of 90 percent of winter snow melt indicate earlier dates than historically observed, particularly those areas west of the Cascade crest and at the lowest elevations (Strauch *et al*. 2014, p. 59). The observed differences for climate change effects for the east and west slopes of the Cascades are a reflection of the influence of a warmer maritime climate to the west and the more continental climate to the east (Strauch *et al*. 2014, p. 59).

Sierra Nevada

Walton *et al*. (2017, entire) developed snow cover projections for the Sierra Nevada region in California, incorporating snow albedo feedback using a hybrid downscaling approach to develop future climate projections. This feedback loop is known to be important for regional climate change (Thackeray and Fletcher 2016, p. 395) and occurs when warming causes snow pack to shrink at margins and the exposed ground absorbs more sunlight than snow, which enhances the warming, resulting in more melting of snow (Walton *et al*. 2017, p. 1,417). This study (using 3 km (1.86 mi) resolution) found that, by the end of the 21st century (2081–2100), warming and loss of snow cover is expected to occur, though the degree varies depending on the GHG scenario (Walton *et al*. 2017, p. 1,430). Under the RCP 8.5 (high emissions) scenario, the study found that the total area covered by snow during the typical month of April decreases by 48 percent, as compared to historical average (1981–2000) (using ensemble mean) (Walton *et al*. 2017, p. 1,432). Under the RCP 4.5 (moderate emissions) scenario, snow cover losses were projected at about half of those for RCP 8.5 (Walton *et al*. 2017, p. 1,434; Figure 13). The effects of warming were more pronounced at lower elevations and were most severe in May and June (Walton *et al*. 2017, p. 1,431; Figure 12). For the months of March and April, the highest elevations were found to have nearly complete snow cover (measured as snow covered fraction) for all GCM simulations (Walton *et al*. 2017, p. 1,431; Figure 12).

Northern and Southern Rocky Mountains–Glacier and Rocky Mountain National Parks

The effects of climate change on snow persistence has been suggested as an important negative impact on wolverine habitat and populations by the mid-21st century (McKelvey *et al*., 2011,

entire). The Service therefore pursued a refined methodology to provide insights into the potential impacts of climate change on snow persistence.

The Service engaged the National Oceanic and Atmospheric Administration (NOAA) laboratories and University of Colorado in Boulder, Colorado (CU) to evaluate and model fine scale persistence of snow in occupied and potential wolverine habitat in the contiguous United States. Those discussions revealed significant progress in fine scale modeling approaches since the early 2000s. The Service provided funding for an assessment of snow extent and depth to assess the effects of climate change on snow persistence in two areas of the western United States, Rocky Mountain and Glacier National Parks (Ray *et al*. 2017, entire). The primary objective of this study was to refine the spatial and temporal scale of snow modeling efforts and improve the scientific understanding of the extent of spring snow retention currently and into the future under a changing climate (Ray *et al*. 2017, p. 9). The objectives of the study included (Ray *et al*. 2017, p. 10):

- Use of fine-scale models to analyze the topographic effects of snow, including slope and aspect (compass direction that slope faces)
- Use of a range of plausible future climate change scenarios to assess snow persistence
- Analysis of extremes and year-to-year variability by selecting representative wet, dry, and near normal years (using observed conditions) and then modeling changes for those base years under several future climate scenarios
- Assessment of changes in snow persistence by elevation

The study was designed to parallel as much as possible and thereby refine the previous assessment of snow cover persistence in the western United States presented in McKelvey *et al*. (2011). However, an exact replication of the McKelvey *et al* (2011) study was not possible given the time, funding, and computational constraints needed to develop a fine-scale assessment. The current study was limited to two study areas (approximately 1,500 to 3,000 km^2 (579 to 1,158 mi^2) each) in the northern and southern Rocky Mountains (see **Appendix H** for maps). The two study areas were selected because they encompass the latitudinal and elevational range of wolverines within the contiguous United States. Glacier National Park (GLAC) is representative of a high latitude and relatively low elevation area currently occupied by wolverines. The Rocky Mountain National Park region (ROMO) is a lower latitude and higher elevation area within the wolverine's historical range, which was recently occupied by a wolverine from 2009 to at least 2012.

Methods: We provide here a brief summary of the methods used in this study. Additional details are contained in the full report authored by Ray *et al*. (2017). The initial step of the analysis was a review of the observed climate and variability to provide context for trends and year-to-year variability. Next, historical snow cover extent and variability were analyzed using satellite remote sensing (MODIS) data from 2000 to 2016 to calculate a snow disappearance date for each year at each pixel. Summary statistics include total snow covered area (total area covered by snow), representation of snow pack by aspect (percent of land areas covered by snow for each of the 17 years in the historical record by topographic aspect based on compass direction that the slope faces), and elevation dependence for wet, near-normal, and dry years (with median of all years used as reference). Future snow pack projections were then generated using the Distributed

Hydrology Soil Vegetation Model (DHSVM), for the historical period 1998-2013, and then validated against SNOTEL observing stations and MODIS satellite data.

Both Ray *et al*. (2017) and McKelvey *et al*. (2011) used the delta method to estimate future snow persistence. The NOAA-DHSVM delta method uses historical observed weather (1998–2013) as the baseline and applies future changes in temperature and precipitation from the chosen GCMs (approximately Year 2055) to estimate future snow persistence on the landscape. This time period was chosen for concordance with the availability of the MODIS satellite remote sensed snow product and availability of the meteorology dataset used for the study. For GLAC, October–May was typically warmer and drier during 2000–2013 than the long term average, while, for ROMO, the October–May precipitation was typical, though with higher temperatures. Although one might assume that the Spring snowpack would therefore be lower in the baseline period than in prior decades, an assessment of three long-established May 1 snow course sites (i.e., manual measurements) in the GLAC and ROMO regions, indicates that, while this is true, the longer record dating back to 1922 in GLAC and 1936 in ROMO indicates a period of relatively low snowpack prior to the 1950's (see Ray *et al*. 2017, pp. 18–20).

The 30-year averaging period used in Ray *et al*. (2017) centered on 2055, with the projected changes described for 2055 based on projected conditions averaged over the period 2041–2070, which was then compared with the conditions for the historical period of 1971–2000 (unless otherwise noted). As noted above (see *Climate Change Effects* section), this approach is used to better characterize an average state around the anomalies and extremes which are inherent in the climate models. This allows for more clearly distinguishing the anthropogenic future change used to compare to a future time slice that is long enough such that the natural variability is at least partly averaged out (Lukas *et al*. 2014, pp. 60–61).

Five future scenarios (GCMs) were selected from CMIP5 global climate model projections to capture variability in temperature and precipitation, using the RCP 4.5 (moderate) and RCP 8.5 (high) emissions scenarios. Representative wet, near normal, and dry years were analyzed for the historical simulations and evaluated for the five future scenarios. The number of years (out of 16) with snow depth greater than 0.5 m (20 in) was also analyzed as was the change in Snowcovered Area (SCA) (area with depth greater than 0.5 m (20 in)). This snow depth was selected based on an analysis of the depth of snow at documented wolverine den sites in Glacier National Park (Ray *et al*. 2017; Table 5-2). Results were reported for "light snow cover" (snow depth greater than 1.25 cm (0.5 in)) and "significant" snow (snow depth > 0.5 m (20 in)) for April 15, May 1, and May 15 for previously defined representative years. These dates were selected based on studies indicating den site abandonment generally occurs before May 1 (see Use of Dens and Denning Behavior discussion above in *Reproduction and Growth* section). However, the May 15 date does not represent a meaningful biological significance for wolverines relative to the presence or lack of snow.

The term "light snow cover" was incorporated as the most directly comparable parameter to McKelvey *et al*.'s "light" snow cover. The average change in SCA and SWE was analyzed as a function for both study areas of elevation and was overlaid with the elevations of documented wolverine den sites (2003–2007) in GLAC. As noted earlier, snow depth was not defined in Copeland *et al*. (2010).

Comparison with McKelvey *et al*. (2011): Although the methods used in this study have similarities with those presented in McKelvey *et al*. (2011), there are several key differences. Ray *et al*. (2017) used a finer spatial resolution model (DHSVM) than McKelvey *et al*. (2011) (0.0625 km^2 vs. 37 km^2) (see **Appendix I** for comparison figure) that also incorporated slope and aspect. The grid cells represented in McKelvey *et al*. (2011) were assumed to be flat (i.e., north-facing slopes treated as identical to south-facing slopes). McKelvey *et al*. (2011) focused on May 1 snow depth as a proxy for May 15 snow disappearance, while Ray *et al*. (2017) focused directly on May 15 snow disappearance and produced results for the presence or absence of deeper snow (nominally greater than or equal to 0.5 m (20 in) depth) on May 1 and April 15. Ray *et al*. (2017) originally focused on May 15 to compare to the McKelvey *et al*. (2011) study, and June 1 to bracket the snowmelt season. However, the April 15 and April 30 dates were added to the evaluation of snowcovered areas to align with temporal reproductive patterns of the wolverine (see Use of Dens and Denning Behavior discussion in *Reproduction and Growth* section above). Because of the increased resolution of this study, Ray *et al*. (2017) were able to consider whether any areas of snow with depth greater than 0.5 m (20 in) will persist in these areas. Additional comparisons are outlined below in Table 8 and in Ray *et al*. (2017, p. 6).

Table 8. Comparison of Methods, Ray *et al*. (2017) vs. Copeland *et al*. (2010) and McKelvey *et al*. (2011)

Feature	Ray *et al*. (2017)	Copeland *et al*. (2010) and McKelvey *et al*. (2011)
Spatial Resolution	250 m x 250 m = 62,500 m^2 or 0.0625 km^2 (0.24 mi^2)	0.125 degrees (~5 km x 7 km; 37 km^2 (14.29 mi^2))
Geographic Area	Glacier and Rocky Mountain National Parks, 300 m below treeline and above	Western United States, except California and Great Basin
Topography	Slope, aspect, and shading were used	Slope and aspect were not used
Validation	SNOTEL (in-situ observations) and MODIS (satellite remote sensing)	None specific to the snow dataset used
Future Scenario Method	Delta Method, used to project 2000-2013 conditions out to Year 2055 (average of 2041–2070)	Delta Method (Years: 2045 (2030–2059), 2085 (2070-2099)
Future Scenarios (GCMs)	*miroc*, *giss*, *fio, cnrm* (both study areas); *canesm* (Glacier National Park only) *hadgem2* (Rocky Mountain National Park only)	Ensemble mean of 10 GCMs, *pcm1*, and *miroc 3.2*
Time-related Results	Long-term means and year-to-year variability (i.e., wet, near normal, and dry years)	Changes in long-term mean snowpack only
Snow Detection and Measurements	Snow presence: 1.25 cm (0.5 in) snow depth threshold on May 15. "Significant snow": snow depth (0.5 meter (20 in) threshold. Snow depth determined by conversion from Snow Water Equivalent using bulk snow density.	Snow presence (13 cm (5.12 in) snow depth threshold on May 1). Snow depth determined by VIC model.
Number of Years of MODIS Data	17 (2000-2016)	7 (2000-2006)
Snow Model	DHSVM (University of Washington)	VIC (University of Washington)
Snow Cover Dates Analyzed	April 15, May 1, and May 15	May 15 (derived from May 1), May 29 (derived from May 1)

Results: While there are challenges in comparing the results from McKelvey *et al*. (2011) directly to the Ray *et al*. (2017) study due to differences in methodology and focus, the qualitative picture can be summarized as follows: projected warming has a larger effect at lower elevations whereas projected precipitation changes may dominate the springtime snowpack in the high country. We present below a summary of the main results from Ray *et al*. (2017).

MODIS Observed Historical Snowpack Variability Analysis:

- In GLAC, Snowcovered Area (SCA) varies considerably by year, including wet years such as 2011 with very persistent snow, years with strong melt in early May, such as 2012, or in late May (2009, 2001), and dry years (2004, 2005) (Ray *et al*. 2017, Section 4.3).
- Even in dry years, northeast-facing slopes in GLAC tend to hold more snow and melt later in the season.
- More than 80 percent of the GLAC study area above approximately 2,000 m (6,562 ft) elevation on May 1 has snow cover during dry years, and more than 95 percent has snow cover above approximately 1,200 m (3,937 ft) during wet years.
- In ROMO, the SCA also varies considerably by year.
- The northwest-facing slopes in ROMO tend to hold more snow even during dry years. In very dry years, snow cover peaks at intermediate elevations, suggesting that the high-altitude snowpack may be particularly vulnerable in this region under warm/dry conditions.

In addition, results from MODIS-derived summary statistics for snow cover for March 1 to July 1 for all years, 2000–2017, illustrate the importance of aspect on snow retention. Calculations of fraction of total land cover (per octant) are presented in Ray *et al*. (2017; Figure 4.5 (GLAC) and Figure 4.12 (ROMO). These graphical presentations illustrate the asymmetry of snow cover for both regions. For GLAC, the northeast directions retain a much larger fractional SCA than other aspects; even in dry years, more than 60 percent of the northeast facing slopes are snow covered on May 15 (Ray *et al*. 2017; Figures 4.5 and 4.6). Similar results were found for ROMO, except that the northwest facing slopes retian a larger amount of fractional snow cover than other aspects (Ray *et al*. 2017; Figures 4.12 and 4.13). These results highlight the importance of evaluating topographic features when interpreting future snow conditions that are biologically relevant to wolverine habitat.

Future Snowpack Projections: The area-wide SCA results include snow cover changes in both forested and above-treeline (alpine) terrain, which may have different implications for wolverine biology.

Glacier National Park (GLAC):

- Projections for April 15, May 1, and May 15 SCA and area with snow depth greater than 0.5 m (20 in) show declines on average in all scenarios, compared to the 2000–2013 historical average, except for small increases in the Warm/Wet scenario and for almost all years.
    - For April 15, light SCA area is reduced by 3–23 percent and significant snow

cover (greater than 0.5 m (20 in)) declines by 7–44 percent.
    - For May 15, light SCA is reduced by 10–36 percent, and the area with significant snow cover declines by 13–50 percent.
- All projections show declines in the number of years with significant snow (equal to or greater than 0.5 m (20 in)), which varies by scenario (e.g., Figure 5-14 in Ray *et al*. 2017). Areas with frequent availability (at least 14 out of 16 years) of significant snow become concentrated in smaller high elevation areas. Lower elevation areas had the largest decreases in the number of years with significant snow cover.
- Most of the known den sites are located between 1,800 m (5,906 ft) and 2,000 m (6,562 ft) in GLAC. Below that elevation band, large snow losses are predicted (40 to 70 percent decrease for two of the scenarios, 16–20 percent for the other three). Above that elevation band, there is little change in SCA for four of the five scenarios (2–8 percent) except in maximum warming scenario (decline of 40 percent (Ray *et al*. 2017; Figure 5-22). In the 1,800–2,000 m (5,906–6,562 ft) band, the snowpack change is sensitive to elevation and to the future climate scenario used.
- For representative wet years, for May 15, the higher elevations of the study areas experience only 2–7 percent loss of snowpack under the scenarios with "least" change and the "moderate" change, although for the dry years, losses range from 18–57 percent.
    - The implication is that the wet, cold climate of the GLAC study area could act as a "buffer" to change in areas with of 0.5 m (20 in) of (significant) snow on May 1, at least for elevations above 1,800 m (5,906 ft).

Rocky Mountain National Park (ROMO):

- Projections of May 15 SCA in ROMO decline on average in all scenarios, except for small increases in the Warm/Wet scenario, and for almost all years.
    - For April 15, light SCA (depth ≥ 5 mm (0.2 in)) declines by 3–18 percent and significant SCA (depth > 0.5 m (20 in)) changes from −1– +16 percent for the five scenarios considered (compared to the 2000-2013 historical average).
    - For May 15, the area with light snow cover declines 8–35 percent and the area with significant snow cover declines 6–38 percent.
- All projections show declines in the number of years with significant snow (equal to or greater than 0.5 (20 in), which varies by scenario (e.g., Figure 5-21 in Ray *et al*. 2017). The areas with frequent availability (at least 14 out of 16 years) of significant snow become concentrated in smaller high elevation areas. In contrast, lower elevation areas had the largest decreases in the number of years with significant snow cover.
- Although no dens have been documented in ROMO, the elevation band for denning, modeled by regression analysis, is estimated at 2,700 to 3,600 m (8,858 to 11,811 ft). On May 1, modest declines in SWE of about 15 percent and less for areas at 3,400 m (11,155 ft) or above result in losses of only about 10 percent snow cover.
    - The implication is that the wet, cold climate of the higher parts of the ROMO study area could also act as a "buffer" to change in the area of 0.5 m (20 in) deep snow on May 1.

Elevation Dependence of Change: In general, and supported by the literature, the snowpack in the higher elevations of both areas is more responsive to precipitation change, while lower

elevations are more responsive to temperature change. The results from our projections of snow cover at lower elevations (Glacier National Park) versus higher elevations (Rocky Mountain National Park) confirm this elevation-dependent change in snowpack. That is, the qualitative picture indicates that projected warming has a larger effect on spring snowpack? at lower elevations whereas projected precipitation changes are likely to ~~dominate~~have a larger effect on spring snowpack in the high country (Ray *et al*. 2017, p. 45).

For GLAC, most of the observed den sites are located within the zone where temperature dominates the future effects of change. For the elevation of den sites in GLAC (i.e., above 1800 m (5,906 ft)), loss of SCA on May 1 spans the range of 5–40 percent, with a 70 percent decrease for the Hot/Wet (*miroc* GCM) scenario. Above 2,200 m (7,218 ft), the losses are less than 5 percent for all but the Hot/Wet scenario.

Current results may be a reasonable estimate for the high mountain ranges within the Rockies that lie between GLAC and ROMO. However, without further study, we cannot reasonably extend these results to say whether or not snow refugia will persist in the Central Rockies below our study elevations (approximately 1,000 m (3,281 ft)). These lower elevations are where McKelvey *et al*. (2011) predicted the greatest losses in snowpack. The NOAA/CU results also cannot be extrapolated to mountain ranges outside of the Rockies (i.e., the Cascade Range) that have different climates (temperature and precipitation). We note here that we have no documented wolverine den sites in the contiguous United States below 1,500 m (4,921 ft) elevation; that is, no documented den locations in the areas where McKelvey *et al*. (2011) predicted the greatest loss in snowpack.

Interpretation and additional analysis relative to wolverine den site scale: The Service was interested in exploring the question, "If snow cover is required for wolverine denning, will there be a sufficient amount of significant snow cover in the future in areas wolverines have historically used for denning in the contiguous United States?" The Service integrated future DHSVM projections (2000–2013 averages) of snow covered area (greater than 0.5 m (20 in) depth) on May 1 for GLAC and ROMO with new information obtained from a spatial analysis of documented den sites in the contiguous United States. This spatial analysis indicated 31 of 34 documented den sites in the contiguous United States were located in areas with slope less than 25 degrees. Avalanche risk increases significantly in areas with slope greater than 25 degrees (Scott 2017, pers. comm.) and wolverines may avoid these areas for denning due to this risk.

Using the projections prepared by Ray *et al*. (2017), we present in Figures 7–14 the spatial distribution of significant snow covered area with slopes less than 25 degrees and within the elevation bands indicated above for three future scenarios in each study area. The three scenarios for GLAC (*miroc*, *cnrm*, and *giss*) and for ROMO (*hadgem2*, *fio*, and *giss*) were chosen to span the range of GCM uncertainty regarding temperature and precipitation, and by extension significant SCA (see Figure 7 and Figure 8). We found that large portions of the study areas meet all three criteria—greater than 0.5 m (20 in) snow depth on May 1, at elevation 1,514–2,252 m (4,967–7,389 ft) for GLAC or 2,700 to 3,600 m (8,858 to 11,811 ft) for ROMO, and with a slope less than 25 degrees—across both study sites in the future.

The GLAC *miroc* simulation shows the greatest decrease in future snow covered area in the elevation band historically used for denning (orange line in Figure 7). Figure 9 shows the spatial

distribution of significant SCA with slope less than 25 degrees and elevation of 1,514–2,252 m (4,967–7,389 ft) for the *miroc* simulation on May 1 (approximately Year 2055). Approximately 494 km^2 (191 mi^2) of area meet the three criteria with an additional 803 km^2 (310 mi^2) of area retaining significant snow covered area, primarily at higher elevations.

Moreover, we determined that large tracts of significant SCA are projected in close proximity to documented historical den sites across all three scenarios (Figures 9–11). This analysis is limited to Glacier National Park because this is the only area where new snow covered area projections and historical den locations were both available. As shown in Table 9, wolverines would not have to travel far, or at all, relative to either distance or elevation to reach areas with significant snow covered area in the future.

A similar analysis was performed for the ROMO study area and the results indicate that large portions of the study area meet all three criteria identified above. The *hadgem2* (Figure 12) and *cnrm* scenarios were found to have the greatest decrease in significant snow covered area of the five scenarios analyzed. Figure 12 (*hadgem2* simulation) shows the spatial distribution of significant SCA (greater than 0.5 m (20 in) depth), elevation of 2,700–3,600 m (8,858–11,811 ft), and slopes less than 25 degrees where denning would be expected to occur. Total area meeting these three criteria was 339 km^2 (131 mi^2) (dark blue in Figure 12), with an additional 446 km^2 (172 mi^2) with snow depth greater than 0.5 m (20 in) (light blue in Figure 12), mostly at higher elevations. Figures 13 (*fio* scenario) and Figure 14 (*giss* scenario) show a similar distribution, albeit larger areas of significant snow retention in the future (see map legends in Figures 13 and 14 for area estimates).

Table 9. Distance of historical GLAC dens (Years 2003–2007) from projected significant snow covered area in the future (approximately Year 2055) (using 2000–2013 average).

A 0 (zero) value indicates the den site location meets all three criteria in the future (greater than 0.5 m (20 in) snow depth on May 1, at elevation 1,514–2,252 m (4,967–7,389 ft), and with a slope less than 25 degrees).

Den Site[a]	Elevation, m (ft)	Distance from den site to nearest model cell, m (ft)		
		GCM scenario		
		miroc	*cnrm*	*giss*
1	2,252 (7,389 ft)	0	0	0
2	2,093 (6,867 ft)	0	0	0
3	1,995 (6,545 ft)	0	0	0
4	1,977 (6,486 ft)	210 (689 ft)	0	0
5	1,973 (6,473 ft)	208 (682 ft)	0	0
6	1,928 (6,326 ft)	0	0	0
7	1,922 (6,306 ft)	9 (29.5 ft)	8 (26 ft)	8 (26 ft)
8	1,912 (6,273 ft)	170 (558 ft)	0	0
9	1,893 (6,211 ft)	110 (361 ft)	0	0
10	1,851 (6,073 ft)	87 (285 ft)	0	0
11	1,843 (6,047 ft)	74 (243 ft)	0	0
12	1,823 (5,981 ft)	56 (184 ft)	0	0
13	1,807 (5,929 ft)	0	0	0
14	1,514 (4,967 ft)	574 (1,883 ft)	571(1,873 ft)	296 (971 ft)

[a]**Glacier National Park**

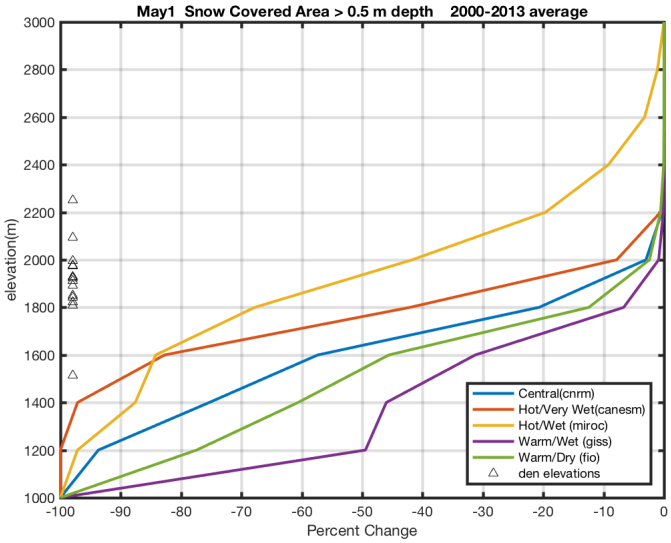


Figure 7. Average Snow Covered Area (depth ≥ 0.5 m (20 in)) percent change at elevation bands for GLAC for five future scenarios on May 1.

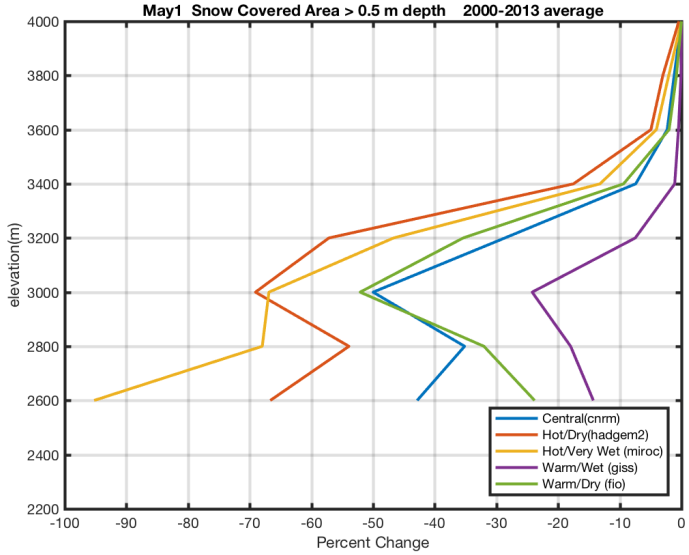


Figure 8. Average Snow Covered Area (depth ≥ 0.5 m (20 in)) percent change at elevation bands for ROMO for five future scenarios on May 1.

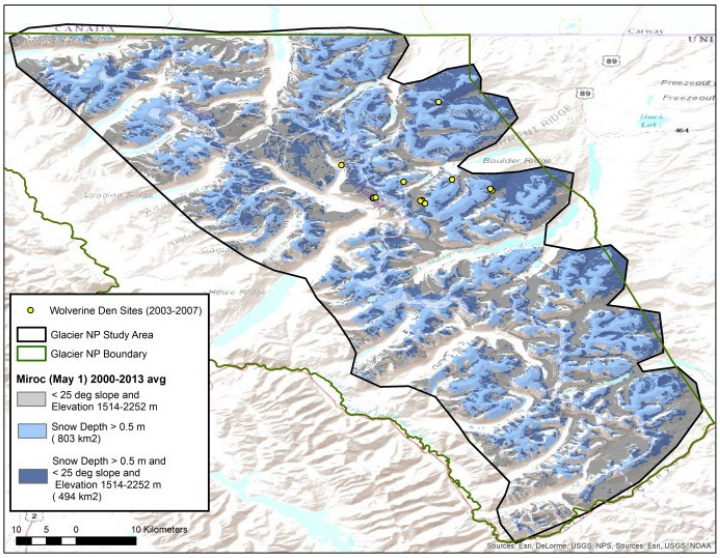


Figure 9. Spatial distribution of averaged (2000-2013) projected snow covered area (depth ≥ 0.5 m (20 in)) for May 1 under the *miroc* (Hot/Wet) scenario in Glacier National Park study area. Map legend shows where slopes are less than 25 degrees and elevations of 1,514–2,252 m (4,968–7,389 ft) (where dens have been documented).

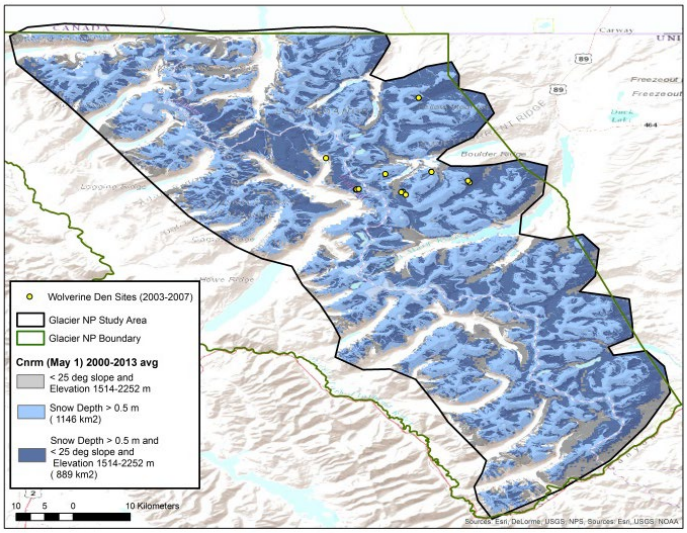


Figure 10. Spatial distribution of averaged (2000-2013) projected snow covered area (depth ≥ 0.5 m (20 in)) for May 1 under the *cnrm* (Central) scenario in Glacier National Park study area. Map legend shows where

slopes are less than 25 degrees and elevations 1514–2252 m (4,968–7,389 ft) (where dens have been documented).

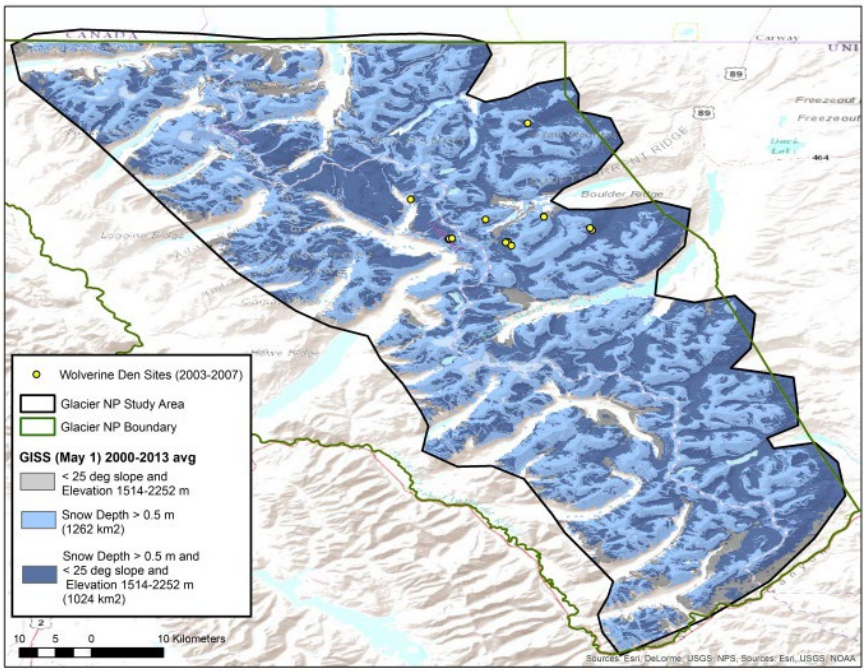


Figure 11. Spatial distribution of averaged (2000-2013) projected snow covered area (depth ≥ 0.5 m (20 in)) for May 1 under the *giss* (Warm/Wet) scenario in Glacier National Park study area. Map legend shows where slopes are less than 25 degrees and elevations 1,514–2,252 m (4,968–7,389 ft) (where dens have been documented).

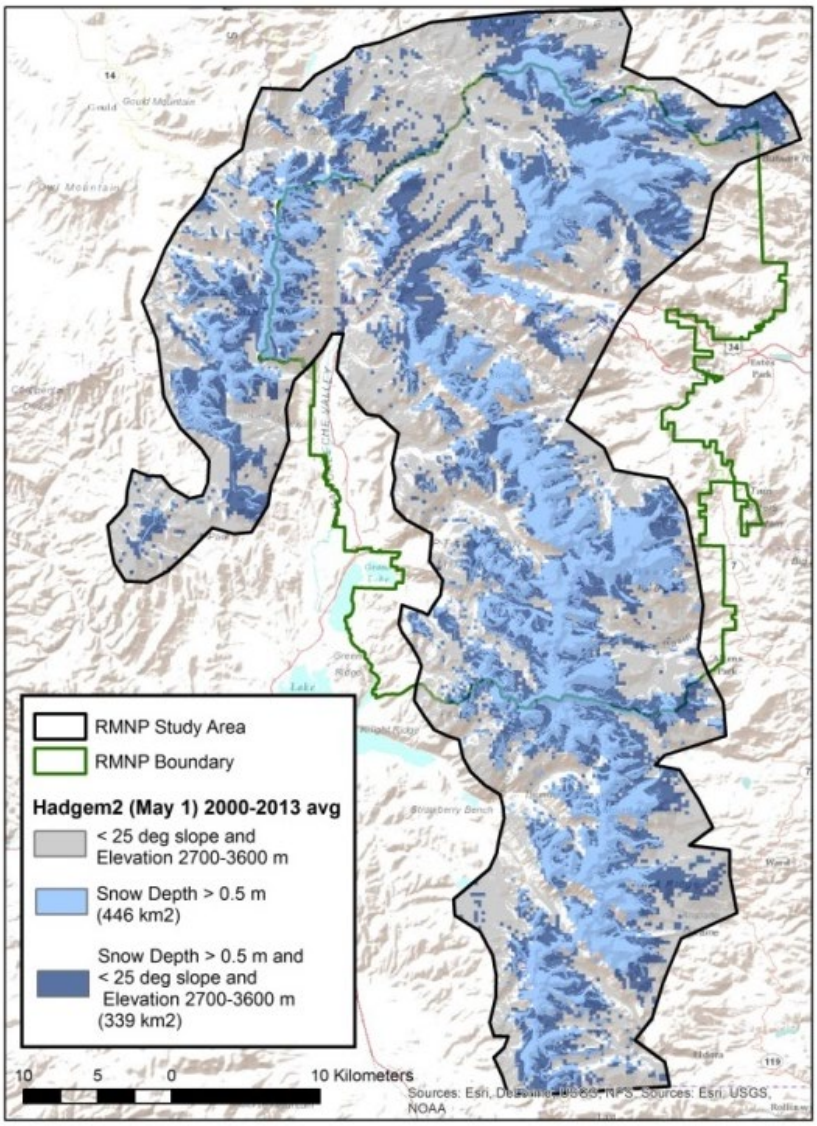


Figure 12. Spatial distribution of averaged (2000-2013) projected snow covered area (depth ≥ 0.5 m (20 in)) for May 1 under the *hadgem2* (Hot/Dry) scenario in Rocky Mountain National Park study area. Map legend shows where slopes are less than 25 degrees and elevations 2,700–3,600 m (8,858–11,811 ft) (inferred elevations where dens would be expected, if occupied).

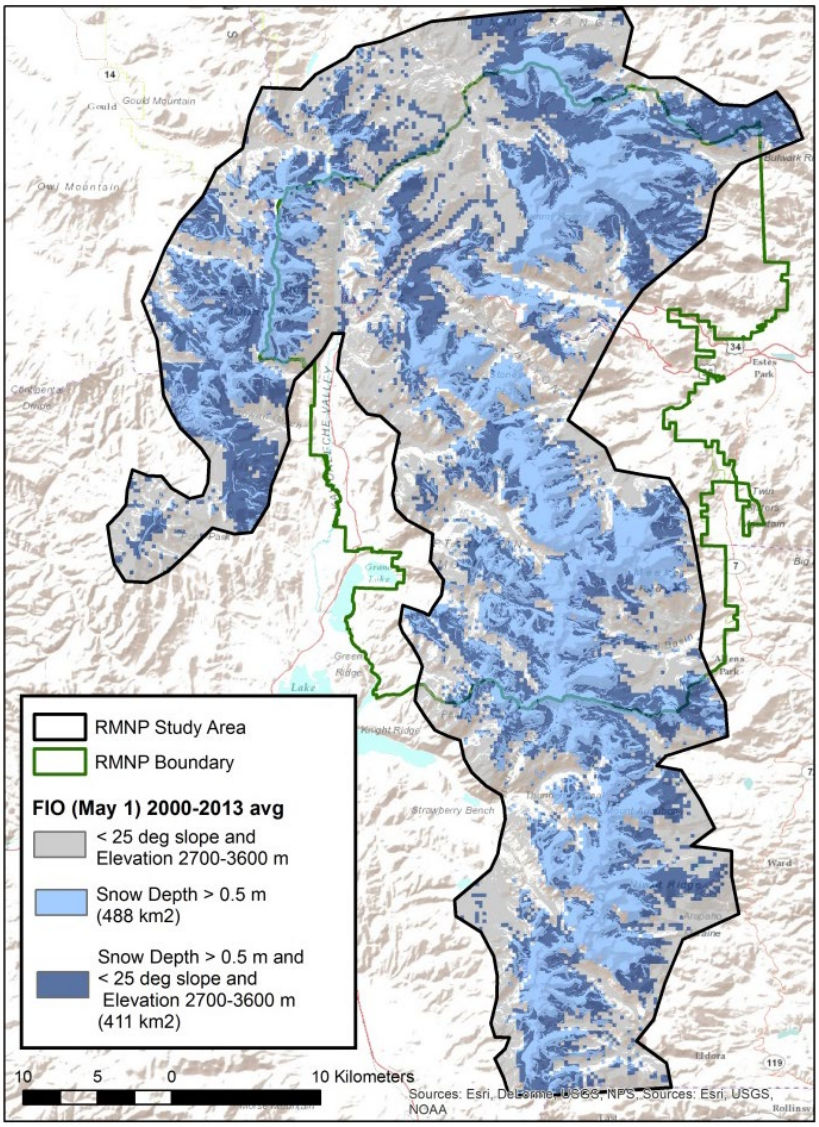


Figure 13. Spatial distribution of averaged (2000-2013) projected snow covered area (depth ≥ 0.5 m (20 in)) for May 1 under the *fio* (Warm/Dry) scenario in Rocky Mountain National Park study area. Map legend shows where slopes are less than 25 degrees and elevations 2,700-3,600 m (8,858–11,811 ft) (inferred elevations where dens would be expected, if occupied).

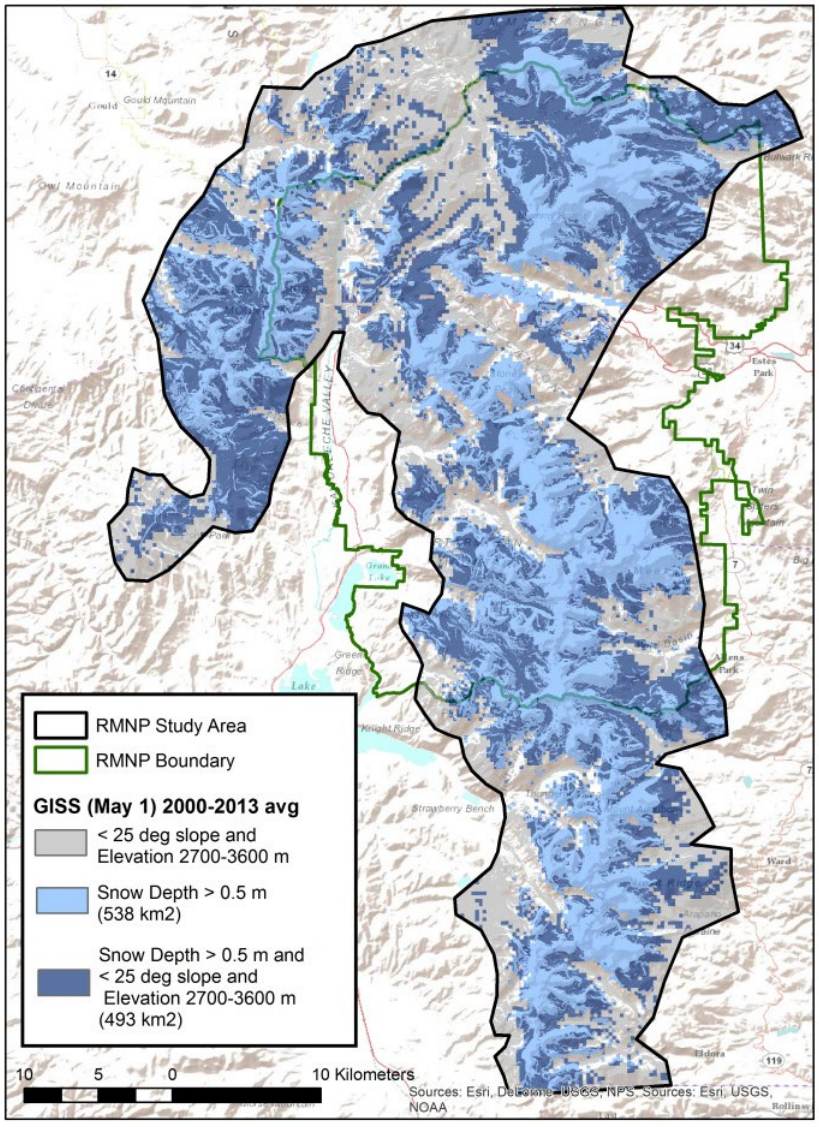


Figure 14. Spatial distribution of averaged (2000-2013) projected snow covered area (depth ≥ 0.5 m (20 in)) for May 1 under the *giss* (Warm/Wet) scenario in Rocky Mountain National Park study area. Map legend shows where slopes are less than 25 degrees and elevations 2,700-3,600 m (8,858–11,811 ft) (inferred elevations where dens would be expected, if occupied).

Montana Climate Assessment

The Montana Climate Assessment also evaluated changes in ~~Another recent assessment of~~ snowpack ~~was conducted for the State of Montana~~ (Whitlock *et al*. 2017, ~~no page number~~entire). The report analyzed recent climate trends in Montana and assessed how climate is projected to change in the future (2040–2069). The study found that snowpack that accumulates at high elevations tends to be more stable and persists longer than at low elevations, due largely to the colder temperatures at high elevations. The largest projected changes in snowpack appear to be in areas located west of the Continental Divide, given their exposure to relatively warm Pacific air masses. Overall, the assessment found that declines in snowpack volume are likely in the future in the basins studied.

Wildland Fire

California

Keeley and Syphard (2016, entire) analyzed fire-climate relationships relative to predicting future fire regimes in California. Their review concluded that: (1) Climate is not a major determinant of fire activity across all landscapes; (2) hotter and drier conditions for areas at lower elevations and lower latitude were found to have little or no increase in fire activity as vegetation types in these regions are ignition limited; (3) increasing annual temperatures by themselves are not good predictors of increased fire activity; seasonality, especially spring and summer temperatures, are more important; and (4) fire-climate models need to be scaled to vegetation types; broad-scale models may produce over-predictions of the total increase in future fire regimes (Keeley and Syphard 2016, pp. 1, 10). Additionally, drought is a key factor in defining fire regimes and annual precipitation is the primary driver of drought variability (Williams *et al*. 2015, p. 6,819), but, at the present time, it is difficult to separate current droughts in California from natural cycles of drought (Keeley and Syphard 2016, p. 6).

Pacific Northwest

Sheehan *et al*. (2015, entire) used downscaled CMIP5 projections to model vegetation and fire changes, with and without fire suppression, within three subregions of the Pacific Northwest. Emission scenarios RCP 4.5 and 8.5 were used for future climate projections (2011–2100). The resulting trends varied by geographic region. In the Western Northwest subregion (from the crest of the Cascade Mountains west), the mean fire interval (MFI) averaged over all climate projections decreased by up to 48 percent, an increase in annual percent area burned (PAB), and the predominant conifer forest is replaced by mixed forest under future climate under both RCP scenarios, with and without fire suppression; thus, climate, rather than fire was found to be the primary influence in this subregion (Sheehan *et al*. 2015, pp. 22–26). In the Eastern Northwest Mountains (ENWM) subregion (mountainous areas east of the Cascade Mountains), the MFI (averaged across all climate projections) decreased by up to 81 percent, there was a projected increase in mean annual PAB, and, while subalpine communities are projected to be lost, conifer forests were projected to continue to dominate this subregion (Sheehan *et al*. 2015, pp. 22–24). When modeled using a without fire suppression regime, the future projections for ENWM indicated a lower MFI and higher mean annual PAB as compared to the with fire suppression

regime (Sheehan *et al*. 2015, p. 22; Table 5). However, the eastern portion of the ENWM subregion was found to show a differing response based on elevation; that is, higher elevations were found to have a higher MFI and a lower mean annual PAB during the 20th century as compared to lower elevations (Sheehan *et al*. 2015, p. 23).

Gergel *et al*. (2017, entire) evaluated the effects of climate change on snowpack, and soil moisture and fuel moisture (fire potential) in the western United States. This study used a statistical downscaling approach, using an ensemble of 10 GCMs across several mountainous regions known to be occupied by wolverines, with a 6.25 km (3.88 mi) spatial resolution hydrologic model. Simulations were run for three future periods: 2020s (2010–2039), 2050s (2040–2069), and 2080s (2070–2099) (Gergel *et al*. 2017, p. 291). Significant declines in snowpack (measured as SWE) were found in all mountain ranges for all future scenarios (using RCPs 4.5 and 8.5) and GCMs (Gergel *et al*. 2017, p. 295). This study found that spring snowpack in mountains along the Pacific Coast is quite sensitive to warmer temperatures, but in the continental mountain ranges (Northern and Southern Rocky Mountains) spring snowpack is more sensitive to changes in precipitation (Gergel *et al*. 2017, p. 295). Differences were observed based on elevation (Gergel *et al*. 2017, p. 292). The study reported on future projected declines of summer soil moisture in forested areas (e.g., Northern Rockies) and the likelihood of increased risk of drought and therefore an increase in wildland fire risk for forested areas (e.g., Northern Rocky Mountains), though they recognize there is significant uncertainty in these future projections in high-elevation areas (Gergel *et al*. 2017, pp. 295–296).

In summary, based on these projections, wildland fire risk is likely to increase in the western United States, but future patterns and trends of wildland fire are dependent on several factors (e.g., degree of warming and drought conditions, fuel and soil moisture, wildland fire management practices, elevation) and geographic region. Based on the best available information, the cumulative effects of wildland fire and climate change (e.g., snowpack) will continue to represent a low impact to the wolverine and its habitat into the mid-21st century, based on climate change projections.

Other Cumulative Effects

Finally, we note here that the effects of climate change on snowpack are projected to negatively affect the season lengths for winter recreational activities, such as skiing and snowmobiling (Wobus *et al*. 2017, entire). Wobus *et al*. (2017) modeled potential changes in snowpack at locations across the contiguous United States using output from five GCMs, two representative pathways (RCPs) that represent a future scenario with continued high emissions growth with limited efforts to reduce GHGs (RCP 8.5) and a future scenario with global GHG mitigation (RCP 4.5), and two future time periods (2050 and 2090) (Wobus *et al*. 2017, pp. 2, 5). Although there was some inter-annual variability in 2050 for some model projections, in general, the Rocky Mountains and Sierra Nevada regions had smaller reductions in season length than other locations due to higher elevation, though for the RCP 8.5 scenario coupled with the 2090 future time period, the smallest projected reduction in season length was 15 percent (Wobus *et al*. 2017, p. 9).

An evaluation of the potential future role of federal lands in the United States (including Alaska and Hawaii) relative to providing outdoor recreation opportunities was prepared in 2014 by the Federal Interagency Council of Outdoor Recreation (White *et al*. 2014, entire). The study used projection models developed for the 2010 Resources Planning Act Assessment (http://www.fs.fed.us/research/rpa/) in conjunction with external projections of relevant factors, including demographic, economic, land use, and climate factors, to simulate future recreation participation (White *et al*. 2014, p. 4). Both motorized (snow mobiling) and non-motorized (developed and undeveloped skiing) winter outdoor activities were evaluated. The study found that for winter motorized activities, the projected trends from 2008 to 2030 indicate per capita participation rates would decline by 10 percent, with days per participant also declining, but at a much lower percentage than the participation rate (White *et al*. 2014, p. 7; Table 3). Snowmobilers still average about 7 days per year on the snow in 2030 (White *et al*. 2014, p. 9).

For non-motorized winter activities, projected trends for developed skiing (downhill skiing and snowboarding) indicated an increase in both per capita participation (6 to 7 percent) and days per participant (3 percent) by 2030 (White *et al*. 2014, p. 9). Undeveloped skiing (cross-country skiing and snowshoeing) has the lowest participation rate for any of the activity groups (approximately 3 percent of the population), and, although little change is expected in per capita participation, the days per participant are projected to increase slightly due to a positive correlation with mean population age (White *et al*. 2014, p. 9).

Summary of Future Conditions

Models represent tools to describe basic physical and biological behaviors using the best available science, and, by presenting a range of plausible future outcomes, they can help generate hypotheses while also identifying knowledge gaps where greater accuracy is needed (Batchelet *et al*. 2016, p. 23). Detecting a species' response to climate change in a single population, and sometimes multiple populations, may not always indicate the response throughout its range given the variation in annual mean surface temperatures over the past century (Post 2013, p. 5).

Climate change model projections for the range of the wolverine within the contiguous United States indicate increases in temperature by the mid-21st century as compared to early to mid-20th century values. Precipitation patterns into the future are less clear as the climate models show significant disagreement in their many regional projections. Although drought conditions in the western United States are not unusual, drought duration and intensity have the potential to be exacerbated by projected temperature increases. Projected temperature and precipitation changes will affect future snow cover and the persistence of snow on the landscape. We note here that a male wolverine occupied (and continues to occupy) the Sierra Nevada Mountain region of California through over 4 years of severe drought with reduced snowpack/snow cover conditions.

Snow cover is projected to decline in response to warming temperatures and changing precipitation patterns, but this varies by elevation, topography, and by geographic region. Simulations of natural snow accumulation at winter recreation locations have found that, overall, higher elevation areas (e.g., Rocky Mountains, Sierra Nevada Mountains) are more resilient to projected changes in temperature and precipitation as compared to lower elevations (Wobus *et*

al. 2017, p. 12). In general, models indicate higher elevations will retain more snow cover than lower elevations, particularly in early spring (April 30/May1).We present results (above) from several recent climate models projecting snowpack declines in the western United States. More specifically, we reviewed a new analysis from NOAA/CU that modeled future snow persistence for Glacier and Rocky Mountain National Parks (areas that encompass the latitudinal and elevational range of the wolverine in the contiguous United States) at high spatial resolution (Ray *et al*. 2017, entire). Their results indicate significant areas (several hundred square kilometers (miles) for each site) of future snow (greater than 0.5 m (20 in) in depth) will persist on May 1 at elevations currently used by wolverines for denning. This is true, on average, across the range of climate models used out to approximately Year 2055.

As described above (see *Life History and Ecology* section), within their North American range, wolverines are found in a variety of habitats within primarily high elevation areas of the western-northwestern United States, and exhibit wide-ranging movements. As discussed above (Denning Habitat), wolverines select den sites for differing characteristics depending on location and natal dens locations are generally associated with snow cover; however, many natal dens have been observed outside of the boundary of the snow model presented in Copeland *et al*. (2010). In addition, reproductive success of wolverines has not been evaluated relative to the depth and persistence of snow cover at the den site scale, or in combination with these or other important key life history characteristics, including avoidance and/or protection from predators, prey availability, availability of food caching habitat.

We also considered temperature and precipitation projections from climate change models in conjunction with wildland fire risk. This risk is likely to increase across the western United States, but patterns and trends are dependent on several factors (e.g., degree of warming and drought conditions, fuel and soil moisture) and geographic region.

Overall Assessment

The wolverine's Current Potential Extent (of Occurrence) includes the west-northwestern United States, large areas of Canada, and Alaska (see Figure 4). In the contiguous United States, potentially suitable habitat (i.e., primary habitat), as determined by the physical and ecological features and the ecological needs of the wolverine, has been estimated at 164,125 km^2 (63,369 mi^2) (Inman *et al*. 2013, p. 281). The species is found in a variety of habitats, but generally occurs in high elevation, relatively inaccessible locations.

In the contiguous United States, the structure of the wolverine population is represented as a metapopulation, although its genetic structure relative to its entire North American range has not been comprehensively evaluated. Wolverine populations in Alaska are considered to be continuous with populations in the Yukon and British Columbia provinces of Canada based on genetic studies (COSEWIC 2014, p. 37). Similarly, studies of wolverines in the North Cascades region have documented recent movement of wolverines from Washington into British Columbia (Aubry *et al*. 2016, pp. 16, 20) and from Idaho (Lucid *et al*. 2016, p. 184) to British Columbia, and earlier from Montana to British Columbia and Alberta (e.g., Newby and Wright 1955, p. 252).

Based on the best available information, wolverines select den sites for different characteristics depending on location. Dens located under snow cover may be related to wolverine distribution based on other life history traits, including morphological, demographic, and behavioral adaptations that allow them to successfully compete for food resources (Inman 2013, pers. comm.). Structure (e.g., uprooted trees, boulders and talus fields) appears to be an important requirement for natal den sites. However, reproductive success of wolverines has not been evaluated relative to the depth and persistence of snow cover, or in combination with these or other important characteristics, including prey availability and predator avoidance. Recent studies of wolverine populations and distribution in Sweden have observed wolverine populations and reproductive den sites outside areas with persistent spring snow cover (Aronsson and Persson 2016; Persson 2017, pers. comm.). The study presented in Webb *et al*. (2016, p. 1,468) (Alberta) concluded that wolverines are adaptable and do not require large areas of deep spring snowpack for successful reproduction, and may select small areas covered with deep snow at a finer scale than can be detected using satellite imagery.

We identified several potential stressors that may be affecting the species' and its habitat currently or in the future, including impacts associated with climate change effects. We recognize there is limited information available for the wolverine, including population estimates and abundance trends. Based on the best available information, demographic risks to the species from either known or most likely potential stressors (i.e., effects from roads, disturbance due to winter recreational activities, effects of wildland fire, and overutilization) are low based on our evaluation of the best available information as it applies to current and potential future conditions for the wolverine and in the context of the attributes that affect the needs of the species.

Climate change model projections for the range of the wolverine within the contiguous United States indicate increases in temperature by the mid-21st century as compared to early to mid-20th century values. Our evaluation of climate change indicates that snow cover is projected to decline in response to warming temperatures and changing precipitation patterns, but this varies by elevation, topography, and by geographic region. In general, models indicate higher elevations will retain more snow cover than lower elevations, particularly in early spring (April 30/May1). Although the persistence of spring snow has not yet been evaluated as critical to wolverine survival in North America, our review of projected snow persistence (to approximately Year 2055) within the Northern and Southern Rocky Mountains, indicates that several hundred kilometers (miles) of deep snow will persist on May 1 at elevations used by the wolverine for denning.

Legal protections include State listing in California and Oregon (as threatened), endangered in Colorado, a candidate species in Washington, protected as a non-game species in Idaho and Wyoming, and protected from collection, importation, and possession in Utah. In Canada, provincial designations range from endangered to threatened in eastern provinces, and sensitive/special concern to no ranking in other provinces. Legal trapping or hunting of wolverines is currently prohibited in the contiguous United States. Trapping effort along the U.S.–Canada border does not represent a barrier to wolverine movement and dispersal along the international border.

Our estimate of Current Potential Extent in the contiguous United States (Figure 3) encompasses approximately 18 percent of designated wilderness areas. Approximately 96 percent of previously modeled wolverine primary habitat (Inman *et al*. 2013) in the contiguous United States is located on Federal lands, with 41 percent of this located in designated wilderness areas. Management actions for conservation of the wolverine and its habitat are included within State Wildlife Action Plans, the *Management Plan for the Conservation of Wolverines in Idaho* (IDFG 2014), and USDA Forest Service Land and Resource Management Plans (see **Appendix G**). Various provisions of these plans include, but are not limited to, winter road closures, fire management, and land acquisition or conservation easements. These management measures, currently and in the future, will alleviate effects associated with potential impacts related to stressors discussed in this report. In addition, the WAFWA Wildlife Chiefs Wolverine Subcommittee is providing a forum for western States to work collaboratively with each other and with the Service and other partners for conserving wolverines found in the west-northwestern United States, and, to date, approximately $1.5 million of funding has been applied towards conservation and management actions for the wolverine (e.g., Western States Wolverine Conservation Project) (McDonald 2017, pers. comm.).

Based on our review of available relevant literature for similar species, we identified the physical and ecological needs of the species as follows: large territories in relatively inaccessible landscapes; at high elevation (1,800 to 3,500 meters (5,906 to 11,483 feet)) within the contiguous United States; access to a variety of food resources, that varies with seasons; and reproductive behavior linked to both temporal and physical features. These needs are currently met for wolverines in the contiguous United States and are expected to be met in the future (i.e., 3840–50 years).

Risk Assessment

In order to characterize a species' viability and demographic risks, we consider the concepts of resilience, representation, and redundancy. We also consider known and potential stressors that may negatively impact the physical and biological features that the species needs for survival and reproduction. Stressors are expressed as risks to its demographic features such as abundance, population and spatial structure, and genetic or ecological diversity. We consider the level of impact a stressor may have on a species along with the consideration of demographic factors (e.g., whether a species has stable, increasing, or decreasing trends in abundance, population growth rates, diversity of populations, and loss or degradation of habitat).

Wolverine populations in much of North America are still recovering from large losses of individuals from unregulated hunting and persecution pressures in the late 1880s into the mid-20th century. Surveys conducted in the winter of 2015–2016, and 2016–2017 continue to document its presence within its "historical" range in the western contiguous United States (representation).

Redundancy, the ability to rebound after stochastic pertrubation, can be characterized by the distribution and connectivity of populations. In considering wolverine in the contiguous United States, individuals are found in alpine, boreal, and subalpine habitats, with breeding populations in four western States. Additionally, wolverines in the contiguous United States appear to be

connected to wolverine populations in Canada along the U.S.–Canadian border, which contributes to current and future redundancy.

Resiliency, the ability to withstand stochastic events, can be characterized by numbers of individuals and abundance trends. As indicated above, population size, growth rate, and current population trends are unknown for the wolverine due to the lack of abundance information. The Current Potential Extent of the wolverine occurs within a large area of northern North America (see Figure 3). The most recent estimate for Canada indicates over 10,000 adult wolverines, as well as expansion of wolverines into historically occupied areas in both Canada and the contiguous United States with movement across both international borders. The 2014 COSEWIC report concluded that a climate-driven decline in wolverine populations in North America is not evident at this time in much of its range (COSEWIC 2014, p. 22). Wolverine populations in Canada are considered stable. Density estimates indicate no declining trend in wolverine populations in Alaska. We recognize that there is limited information on populations (representation) or genetic diversity (resiliency and representation) for the wolverine in the contiguous United States, and no comprehensive studies to indicate what a viable (or minimal) wolverine population size should be across its North American range. However, the best available information does not indicate either increasing or declining numbers of the wolverine in North America, including the contiguous United States. Further, at this time, the best available information does not indicate that the species' abundance is significantly impacted by human-caused stressors and this is unlikely to change in the future, supporting current and future resiliency.

As discussed above (Status–Future Conditions), both direct and cumulative effects of climate change (e.g., higher temperatures, loss of snow cover, wildland fire) may affect the resilience of the wolverine by creating an environment that is less favorable to its physiological and ecological needs. We are unaware of studies of the wolverine that have formally evaluated the species' responses (e.g., reproductive success) to warming temperatures or other climate change effects. Recent studies of behavioral responses to climate change effects of the American pika (*Ochotona princeps*; pika) (Beever *et al*. 2017, entire; Jeffress *et al*. 2017, entire; Stewart *et al*., 2017, entire) highlight the following: 1) the need for monitoring programs for animals like the wolverine that are found in relatively inaccessible locations, in naturally low densities, 2) the importance of evaluating occupancy at multiple scales (Jeffress *et al*. 2017, p. 266), and 3) the potential for geographical variation and habitat structure in adaptation to climate change effects.

As described in this SSA Report, the best available information indicates confirmed observations of wolverines denning in areas with patchy snow cover in Alaska, Canada, and Scandinavia. Given their high rate of movement, large dispersal, and other observed life history traits (e.g., behavioral plasticity) observed in wolverines, we do not predict a significant loss of individual and population resiliency to the species in the future (i.e., ~~40~~38–50 years) within its North America range, including the contiguous United States.

Commented [SJ2]: Do we discuss somewhere that they have been shown to dispers over large un-snowy areas? That snow cover isn't required for movement? I think we need to point that out.

Also, do should we point out that snow cover and denning habitat is not limiting currently or in the future?

Currently, we are unaware of any documented specific risks for the wolverine related to a substantial change or loss of diversity in life history traits, population demographics, morphology, behavior, or genetic characteristics which can be used to characterize species representation (the ability to adapt to change). Rates of dispersal or gene flow are not known to

have changed. Additionally, there is no currently available information to indicate that the current abundance of the wolverine across its Current Potential Extent in the contiguous United States is at a level that is causing inbreeding depression or loss of genetic variation that would affect representation. Nor is there any information to indicate that this species is unable to adapt or adjust to changing conditions (e.g., potential reduction in snow cover). We do not expect a reduction in representation of the wolverines in the contiguous United States in the future.

Acknowledgements

SSA Report review and comments submitted by:

USDA Forest Service (Regions 1 and 6)
Bureau of Land Management
California Deparment of Fish and Wildlife
Colorado Parks and Wildlife
Idaho Department of Fish and Game
Montana Fish Wildlife & Parks
New Mexico Department of Game and Fish
Oregon Department of Fish and Wildlife
Utah Department of Natural Resources, Division of Wildlife Resources
Washington Department of Fish and Wildlife
Wyoming Game and Fish Department
Four independent peer reviewers

Technical and funding support provided by the U.S. Fish and Wildlife Service's Mountain-Prairie Region Science Applications Program

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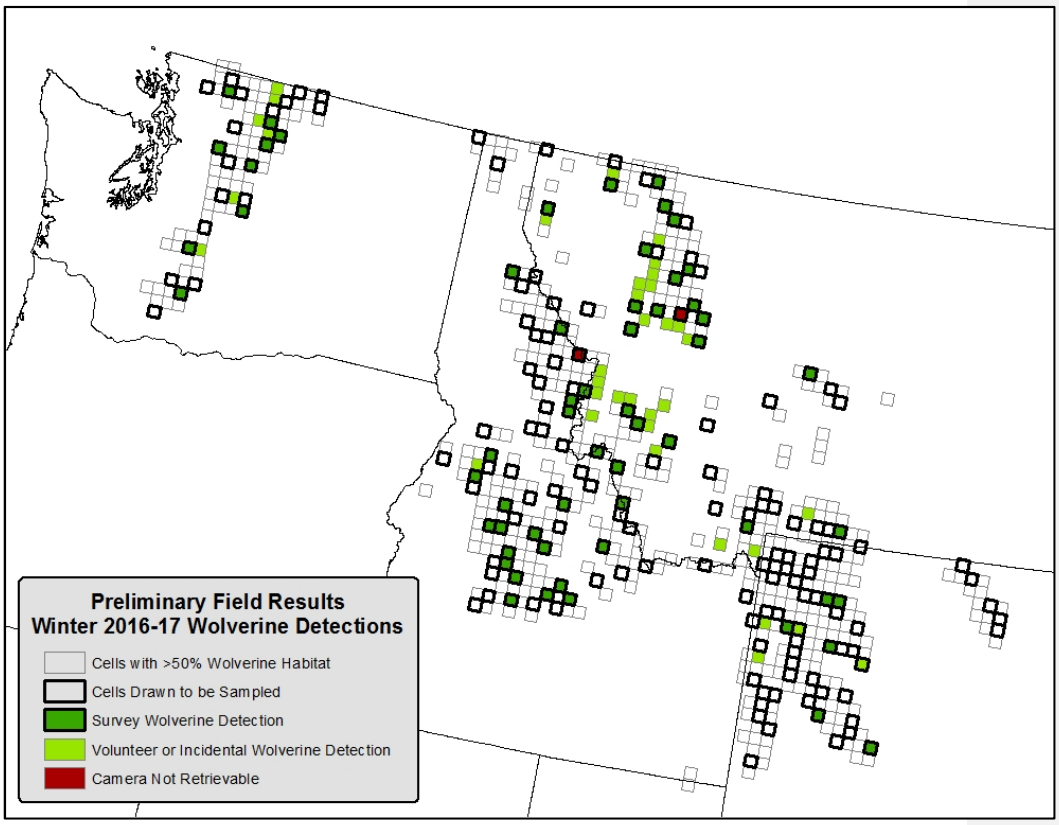
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Appendices

Appendix A – Ecological Regions of North America
(Source: CEC 1997; EPA 2010)



Appendix B – Wolverine Detections, Multi-State Wolverinec Occupancy Survey and Volunteer/Incidental Sightings, Winter 2015–2016 (Wyoming only) and Winter 2016–2017
Source: WAFWA Wolverine Subcommittee 2017.



Appendix C – Summary of Multi-State Wolverine Occupancy Survey
Text provided by Montana FWP 2017, pers. comm.

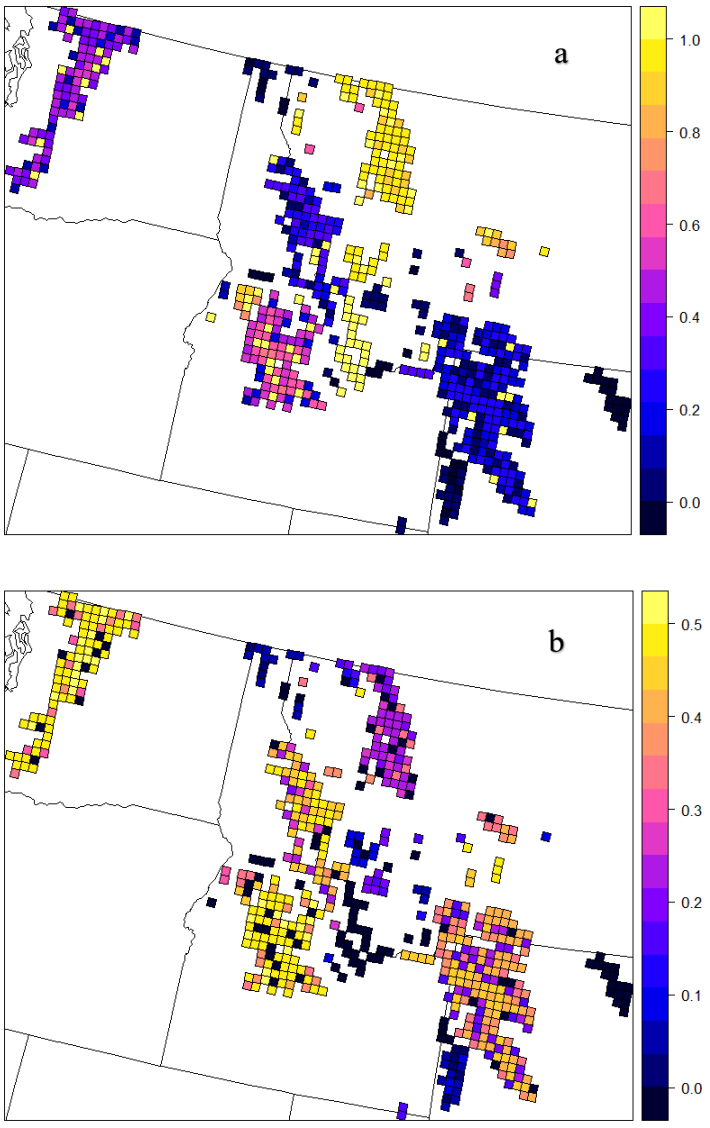
"Wolverines are a naturally uncommon and rarely encountered species that resides in remote, high-elevation areas of the western United States. Proactive conservation measures are needed for this inherently small population that exists as a metapopulation across the region. One key component of a wolverine conservation strategy is development of a population monitoring program – up to now, records of wolverine occurrence have been rare and collected in an uncoordinated and opportunistic fashion, which makes it difficult to assess the species distribution, status and trend. The continued absence of a monitoring program jeopardizes the ability of agencies to allocate resources effectively among wolverines and other conservation priorities. State wildlife agencies in WA, ID, MT, and WY, along with federal, tribal, and NGO partners, developed a collaborative monitoring program that can be implemented in a coordinated fashion across the species range in the western U.S. The foundation for this monitoring program is a coordinated camera survey to obtain baseline information on distribution, genetics, and occupancy of wolverines. When repeated over time, trends in occupancy and distribution can be evaluated.

We developed a sampling frame based on existing models of wolverine habitat. The sampling frame consisted of 633 15x15 km grid cells. We selected 185 cells to sample from the frame using a GRTS sampling method. A single remotely triggered camera was placed in likely wolverine habitat within each selected cell along with a bait or scent attractant. Sampling occurred across a four-state area (ID, MT, WA, WY) during the winters of 2015-16 and 2016-17.

We obtained data from 183 sites and detected wolverines at 59 of the sampled cells. Detection rate was exceptionally high (p=0.92), and did not differ between sites that were baited and revisited multiple times vs. sites that were scented only and visited less frequently. The average estimated probability of wolverine occupancy during our study was 0.42 (95% credible interval 0.29-0.55) suggesting that wolverines used nearly half of all sites during the study period. Wolverine occupancy varied across the region with highest occupancy in the Northern Continental Divide Ecosystem and lower on the southern and eastern periphery of the study area (Figure 1).

The multi-state wolverine survey produced several important outcomes. First, it provided a rigorous and previously unavailable evaluation of wolverine distribution for the first time since the species was extirpated from the contiguous U.S. a century ago. This evaluation confirms the broad distribution of wolverines across the region and that recovery has progressed substantially since historical lows. The effort also demonstrates that occupancy probability varies throughout the region. The variability may be due to differences in habitat quality, differences in wolverine survival, or could be due to the time required to recover from historical absence. Perhaps most importantly, the regional occupancy estimates provide a baseline for future evaluations of change in wolverine distribution through time, including the possibility of detecting any influences due to climate change. In addition, the high detection rate and equal performance of stations visited less frequently indicates the potential for substantial cost savings during future surveys. The effort also demonstrated strong and fruitful collaboration among state, federal and tribal agencies at the scale of a species range, and the results provide management-level guidance on where to direct wolverine conservation in the future. For instance, information on occupancy may be used to inform criteria for translocation, should some states pursue that as a means for enhancing the conservation status of wolverines."

Figure 1. Predicted wolverine occupancy (a) and standard errors (b) for the multistate wolverine survey in Winters 2015–2016 and 2016–2017.

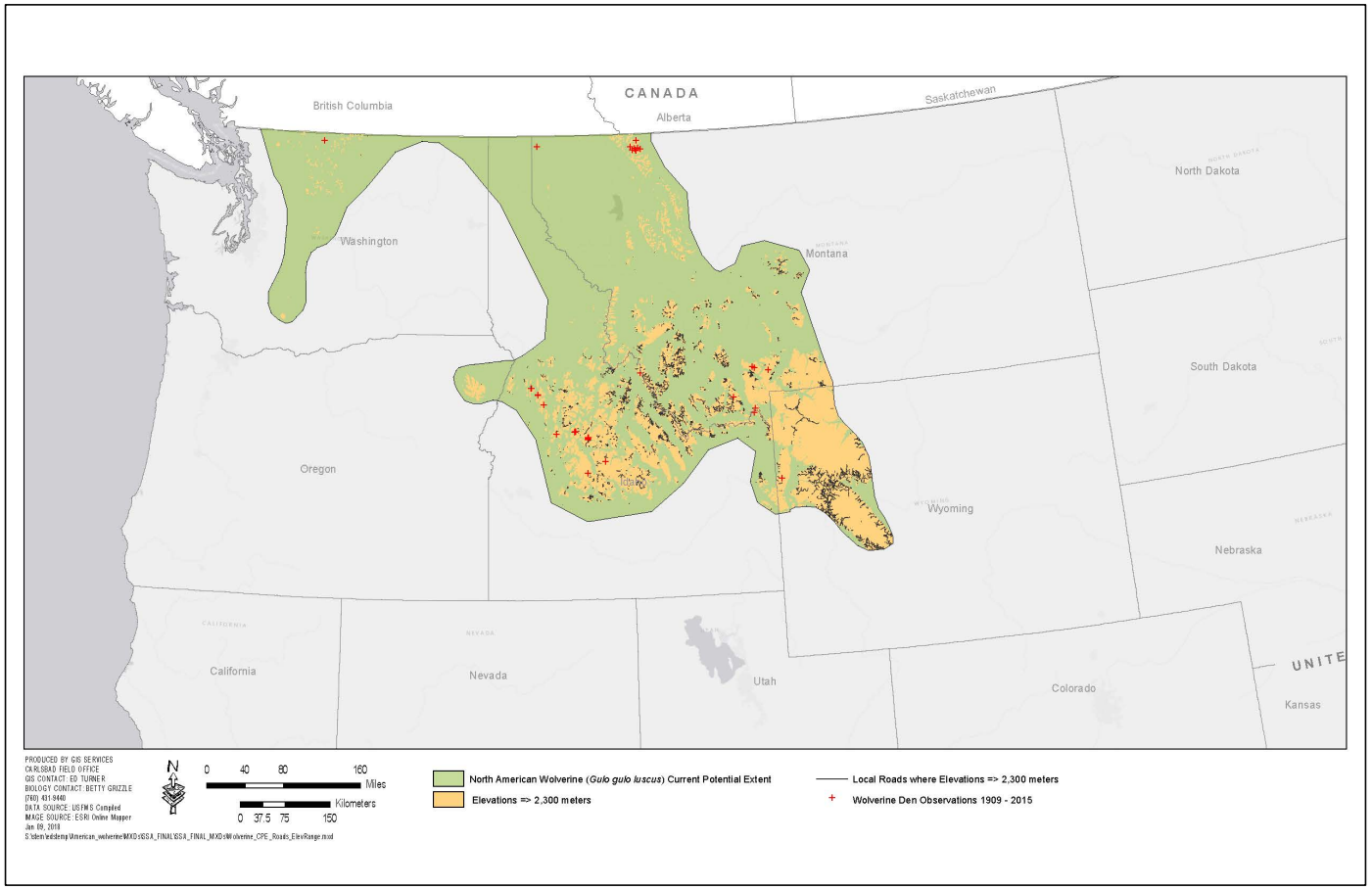


Appendix D – Land Ownership of the Current Potential Extent of Wolverine in the Contiguous United States
(based on Albers Projection; see Figure 3)

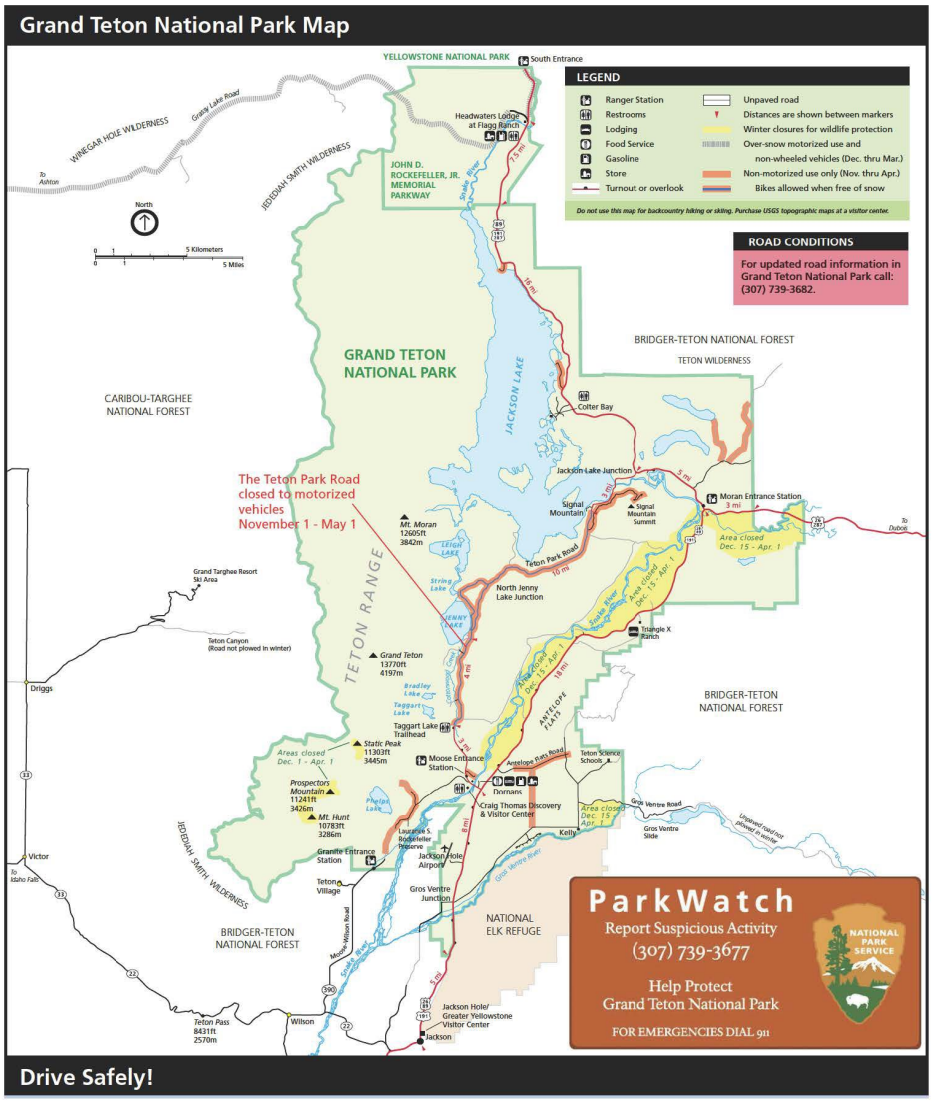
Ownership (% of total)	**Agency or other Entity**	**Total (acres)**	**Total (square kilometers)**
Federal Lands	Bureau of Indian Affairs	2,108,012	8,531
	Bureau of Land Management	4,236,779	17,146
	Bureau of Reclamation	137,567	557
	Forest Service	45,217,891	182,990
	U.S. Fish and Wildlife Service	115,184	466
	National Park Service	4,429,682	17,926
	Other U.S. Department of Agriculture	17,449	71
	Other Federal	76,152	308
Total Federal (72.4%)		**56,338,714**	**227,995**
State Lands (4.3%)	Idaho, Montana, Oregon, Washington, Wyoming	**3,337,474**	**13,506**
Local Government (0.01%)		**9,249**	**37**
Private Lands (22.8%)		**17,771,270**	**71,918**
No Code (“99”) (0.24%)		**187,625**	**759**
Undetermined (0.24%)		**184,767**	**748**
Total (100%)		**77,829,099**	**314,963**

Note: Numbers may not total to 100 percent due to rounding.

Appendix E – Results from Spatial Analysis of Roads within Current Potential Extent of Wolverine



Appendix F – Road Closure Map, Grand Teton National Park
Retrieved from: https://2v9usu38jb9t3l8big1ialsn-wpengine.netdna-ssl.com/wp-content/uploads/2015/11/GTNP-closure-map.pdf



Appendix G – Existing Regulatory Mechanisms and Voluntary Conservation Measures

Federal Mechanisms

Organic Administration Act of 1897 and the Multiple–Use, Sustained–Yield Act of 1960

The USFS Organic Act of 1897 (16 U.S.C. § 475–482) established general guidelines for administration of timber on USFS lands, which was followed by the Multiple–Use, Sustained–Yield Act (MUSY) of 1960 (16 U.S.C. § 528–531), which broadened the management of USFS lands to include outdoor recreation, range, watershed, and wildlife and fish purposes.

National Forest Management Act

The National Forest Management Act (NFMA) (16 U.S.C. § 1600 *et seq*.) requires the Forest Service to develop a planning rule under the principles of the MUSY of 1960 (16 U.S.C. § 528–531). The NFMA outlines the process for the development and revision of the land management plans and their guidelines and standards (16 U.S.C. § 1604(g)).

A new National Forest System (NFS) land management planning rule (Planning Rule) was adopted by the U.S. Department of Agriculture Forest Service (Forest Service) in 2012 (77 FR 21162; April 9, 2012). The new Planning Rule guides the development, amendment, and revision of land management plans for all units of the NFS to maintain and restore NFS land and water ecosystems while providing for ecosystem services and multiple uses. Land management plans (also called Forest Plans) are designed to: (1) Provide for the sustainability of ecosystems and resources; (2) meet the need for forest restoration and conservation, watershed protection, and species diversity and conservation; and (3) assist the Forest Service in providing a sustainable flow of benefits, services, and uses of NFS lands that provide jobs and contribute to the economic and social sustainability of communities (77 FR 21261, April 9, 2012). A land management plan does not authorize projects or activities, but projects and activities must be consistent with the plan (77 FR 21261; April 9, 2012). The plan must provide for the diversity of plant and animal communities including species-specific plan components in which a determination is made as to whether the plan provides the "ecological conditions necessary to…contribute to the recovery of federally listed threatened and endangered species…" (77 FR 21265; April 9, 2012).

The Record of Decision for the final Planning Rule was based on the analyses presented in the *Final Programmatic Environmental Impact Statement, National Forest System Land Management Planning* (77 FR 21162–21276; April 9, 2012), which was prepared in accordance with the requirements of the National Environmental Policy Act (NEPA) (discussed below). In addition, the NFMA requires land management plans to be developed in accordance with the procedural requirements of NEPA, with a similar effect as zoning requirements or regulations as these plans control activities on the national forests and are judicially enforceable until properly revised (Coggins *et al.* 2001, p. 720).

A Species of Conservation Concern (SCC) is defined in the 2012 Planning Rule and in regulation (36 CFR 219.9(c)), as "a species, other than federally recognized threatened,

endangered, proposed, or candidate species, that is known to occur in the plan area and for which the regional forester has determined that the best available scientific information indicates substantial concern about the species' capability to persist over the long-term in the plan area." The 2012 Planning Rule requires Regional Foresters to identify SCC for plan revision, and, when identified for a National Forest, monitoring plans are changed as needed (77 FR 21250, 21267; April 9, 2012). Wolverine is considered a SCC in the Rocky Mountain Region (Region 2). It is considered a Sensitive Species in the Intermountain Region (Region 4) and Northern Region (Region 1).

Within our estimated Current Potential Extent of the wolverine in the contiguous United States (see Figure 3), we identified 22 National Forests. These areas are contained within four Forest Service Regions across the western United States.

National Forest Land Management Plans (Forest Plans)

We reviewed several Forest Plans or related planning documents in an effort to describe how these plans provide conservation management for the wolverine and its habitat, including wildland fire management practices. The sections below are, in most cases, taken directly from relevant documents. However, this discussion is not intended to be inclusive of all NFS management strategies and activities across the entire Current Potential Extent of the wolverine in the contiguous United States.

Sierra Nevada Forest Plan Implementation

The 2004 Sierra Nevada Forest Plan Amendment (referred to as the Sierra Nevada Framework) amended the Land and Resource Management Plans (LRMP) for the eleven National Forests in the Sierra Nevada range to improve protection of old forests, wildlife habitats, watersheds and communities in the Sierra Nevada Mountains and Modoc Plateau. This amendment applies to the Tahoe National Forest, which has been occupied by a single male wolverine since at least 2008 (Moriarty *et al*. 2009, p. 150). The emphasis of the 2004 Sierra Nevada Framework is to adopt an integrated strategy for vegetation management that is aggressive enough to reduce the risk of wildfire to communities in the wildland urban interface, while modifying fire behavior over the broader landscape. Direction is provided as management goals and strategies, desired conditions, management intents and objectives, and management standards and guidelines. The 2004 Framework addressed five problem areas: old forest ecosystems and associated species; aquatic, riparian and meadow ecosystems and associated species; fire and fuels management; noxious weeds; and lower west side hardwood ecosystems (Forest Service 2013, p. 2–3).

Kootenai National Forest

The Kootenai National Forest is located in the northwest corner of Montana along the Canadian border and includes about 2.2 million acres of public land (Forest Service 2015, p. 7). The Forest Service published a Revised Land Management Plan for the Kootenai National Forest in 2015 that identifies forestwide direction, including goals, desired conditions, objectives, standards, and guidelines for physical and biological elements including wildlife such as management activities that promote connectivity and avoiding or minimizing disturbance at known active denning sites

for sensitive, proposed, threatened, or endangered species not covered under other forestwide guidelines. It also outlines objectives and guidelines related to the use of fire to maintain or improve habitat and maintaining unlogged conditions in some portions of areas burned by wildfires for 5 years post-fire (Forest Service 2015, pp. 28–32).

The Kootenai National Forest Land Management Plan also identifies *proposed or possible* actions for wildlife management that includes establishing and maintaining the vegetation diversity necessary to provide food, cover, and security for wildlife species native to the Kootenai National Forest in cooperation with federal, state, and other organizations. For wolverine, those management activities might include maintaining, managing, and protecting lands known or suspected to contribute to landscape linkages for wolverine (and other carnivores) in order to promote genetic dispersal and healthy populations (Forest Service 2015, p. 128).

Beaverhead-Deerlodge National Forest

The Beaverhead-Deerlodge National Forest covers 3.38 million acres in southwest Montana (Forest Service 2009, p. 2). The Beaverhead-Deerlodge National Forest Land and Resource Management Plan identifies goals, objectives, and standards for wildlife management (Forest Service 2009, pp. 45–49). Of relevance to the wolverine, wildlife security management goals include securing areas and connectivity for ungulates and large carnivores and managing the density of open motorized roads and trails by landscape region (Forest Service 2009, p. 45). Objectives include management of habitat conditions for elk security and winter habitat integrity for wolverine and mountain goat relative to changes in abundance of these Management Indicator Species (Forest Service 2009, p. 47). Monitoring elements are defined in the Land and Resource Management Plan that link goals and objectives to elements of the National Monitoring and Evaluation Framework (Forest Service 2009, pp. 273–280). For wildlife security, three performance measures relative to determining whether management activities are effectively protecting high elevation winter habitats for wolverines and mountain goats are defined: (1) presence or absence of wolverines in high elevation habitats, (2) populations of mountain goats (from Montana Fish Wildlife & Parks), and (3) number of snowmobile entries into non-motorized high elevation units protected for wolverines and mountain goats (Forest Service 2009, p. 277). In addition, in order to evaluate objectives related to road and trail densities, a performance measure related to changes in open motorized road and trail density for both seasons by landscape is included (Forest Service 2009, p. 277).

The Forest Service is monitoring the Mount Jefferson Recommended Wilderness boundary for illegal snowmobile intrusions into the wolverine habitat closure; that is, illegal use will be monitored and recorded (number and distance of intrusions) during the period open to snowmobiles December 2 to May 15 and any other time of the year snow conditions make snowmobiling possible (Forest Service 2009, p. 277). A reassessment of the decision to allow snowmobile use will be triggered if: (1) illegal intrusions are documented throughout the closure period; (2) illegal intrusions into the closed area, or (3) illegal intrusions that extend as far as the Bureau of Land Management (BLM) Wilderness Study Area (Forest Service 2009, p. 277).

Flathead National Forest

The Flathead National Forest is located in the northern Rocky Mountains in western Montana and includes approximately 2.4 million acres of public land (Forest Service 2016a, p. 3). This National Forest is surrounded by the Kootenai, Lewis and Clark, and Lolo National Forests, Glacier National Park, and Canada and includes large areas of designated wilderness (e.g., Bob Marshall Wilderness Complex, Mission Mountains Wilderness), Crown of the Continent Ecosystem, and wild and scenic river systems (Forest Service 2016a, pp. 3–4).

A Draft Revised Forest Plan was prepared for the Flathead National Forest in 2016 (Forest Service 2016b, entire). The Draft Revised Forest Plan identifies components to guide future projects and activities and the plan monitoring program, though these components are not commitments or final decisions approving projects or activities (Forest Service 2016b, p. 3). These components include desired conditions, objectives, standards, guidelines, suitability, and monitoring questions and monitoring indicators (Forest Service 2016b, p. 3). [A *desired condition* is a description of specific social, economic, and/or ecological characteristics of the plan area, or a portion of the plan area, toward which management of the land and resources should be directed, while an *objective* a concise, measurable, and time-specific statement of a desired rate of progress toward a desired condition or conditions (Forest Service 2016b, p. 4). A *standard* is a mandatory constraint on project and activity decision making, established to help achieve or maintain the desired condition or conditions, and a *guideline* is a constraint on project and activity decision-making that allows for departure from its terms, and are established to help achieve or maintain a desired condition or conditions, to avoid or mitigate undesirable effects, or to meet applicable legal requirements (Forest Service 2016b, pp. 4–5).]

Relative to wolverine, plan components for the revised forest plan include two guidelines that are protective of wolverine habitat; one that would protect modeled wolverine maternal denning habitat with respect to new projects or activity authorizations involving helicopter use and one that stipulates no net increase in the percentage of modeled wolverine maternal denning habitat where motorized over-snow vehicle use would be suitable on National Forest System lands. Additionally, as described in the Final EIS, management area allocations for Alternatives A, B modified and C include recommended wilderness areas that would add to existing wilderness. Desired conditions related to maintaining connectivity for wolverine and other wildlife are also identified within several geographic areas (Kuennen 2017, pers. comm.).

Federal Land Policy and Management Act (FLPMA) of 1976

FLMPA (43 U.S.C. 1711-1712) represents the BLM's "organic act" for public lands management under the principles of multiple use and sustained yield. Its implementing regulations give BLM regulatory authority over activities for protection of the environment, including mining claims. Under FLPMA and BLM policy, public lands must be managed so as to protect the quality of scientific, scenic, historical, ecological, environmental, air and atmospheric, water resource, and archaeological values (BLM 2005, p. 1).

Land Use and Resource Management Plans

BLM land use planning requirements are established by Sections 201 and 202 of FLMPA and regulations at 43 CFR 1600 (BLM 2005, p. 1). A *Land Use Planning Handbook* (BLM 2005, entire) provides guidance for implementing land use planning requirements established under FLMPA and implementing regulations. Land use plans prepared by BLM include resource management plans (RMPs) and management framework plans (BLM 2005, p. 1). The RMPs establish the basis for actions and approved uses on the public lands and are prepared for areas of public lands, called planning areas (BLM 2005, pp. 1, 14). These plans are periodically evaluated and revised in response to changed conditions and resource demands (BLM 2005, pp. 33–34).

National Environmental Policy Act (NEPA)

All Federal agencies are required to adhere to the NEPA of 1970 (42 U.S.C. 4321 et seq.) for projects they fund, authorize, or carry out. Prior to implementation of such projects with a Federal nexus, NEPA requires the agency to analyze the project for potential impacts to the human environment, including natural resources. The Council on Environmental Quality's regulations for implementing NEPA state that agencies shall include a discussion on the environmental impacts of the various project alternatives (including the proposed action), any adverse environmental effects that cannot be avoided, and any irreversible or irretrievable commitments of resources involved (40 CFR part 1502). The public notice provisions of NEPA provide an opportunity for the Service and other interested parties to review proposed actions and provide recommendations to the implementing agency. NEPA does not impose substantive environmental obligations on Federal agencies—it merely prohibits an uninformed agency action. However, if an Environmental Impact Statement is prepared for an agency action, the agency must take a "hard look" at the consequences of this action and must consider all potentially significant environmental impacts. Federal agencies may include mitigation measures in the final Environmental Impact Statement as a result of the NEPA process that may help to conserve the wolverine and its habitat.

Although NEPA requires full evaluation and disclosure of information regarding the effects of contemplated Federal actions on sensitive species and their habitats, it does not by itself regulate activities that might affect the wolverine; that is, effects to the subspecies and its habitat would receive the same scrutiny as other plant and wildlife resources during the NEPA process and associated analyses of a project's potential impacts to the human environment. The Service receives notification letters for Draft and Final Environmental Impact Statements prepared by the Forest Service, BLM and other Federal agencies pursuant to NEPA for specific proposed projects including those within National Forests or National Parks, and preparation of Forest Service Land and Resource Management Plans, as discussed above.

Wilderness Act

The Wilderness Act of 1964 (16 U.S.C. 1131–1136) provides protection of habitat from most forms of development, though no single agency is responsible for administration of lands provided this designation, which are designated (or modified) by Congress. The Wilderness Act prohibits commercial enterprises and permanent roads within wilderness area and restricts temporary roads, motorized and mechanical transport, and structures, but does not prohibit all commercial uses (e.g., grazing). Within the portion of our estimated Current Potential Extent of

the wolverine in the contiguous United States and Alaska, approximately 18 percent is designated as wilderness areas under the Wilderness Act. We also evaluated wilderness contained within modeled wolverine primary habitat from Inman *et al*. (2013). We found 41 percent of this suitable habitat was designated as wilderness areas.

State Mechanisms

California

As noted above, the wolverine is a threatened species under the California Endangered Species Act or CESA, which prohibits the take of any species of wildlife designated by the California Fish and Game Commission as endangered, threatened, or candidate species (CDFW 2017b). CDFW may authorize the take of any such species if certain conditions are met through the issuance of permits (e.g., Incidental Take Permits) (CDFW 2017b). The wolverine is also a fully protected mammal in California, a designation that provides it with the most restrictive protection regarding take (i.e., may not be taken or possessed at any time) (CDFG Game Code, Division 4, Part 3 (Mammals), Chapter 8 (Fully Protected Mammals) § 4700(a)(1)).

The wolverine is also a Species of Greatest Conservation Need (SGCN) in the State's Wildlife Action Plan[5] and is a focal species of conservation strategies for conservation targets in the Southern Cascades and Sierra Nevada Ecoregions, and in the Mono Ecoregion of the Deserts Province section (Big Sagebrush Scrub (CDFW 2015, pp. 5.2-16, 5.4-23, 5.6-19).

In 2011, the CDFW (formerly California Department of Fish and Game) prepared an assessment/briefing document, *California Wolverine Population Augmentation Considerations*, in response to a *Feasibility Assessment and Implementation Plan for Population Augmentation of Wolverines in California* (November 2010) submitted to the Department by the Institute for Wildlife Studies (California Department of Fish and Game (CDFG) 2011). As of August 2017, no action has been taken by CDFW toward implementation of augmentation of wolverines in California.

Oregon

The wolverine has been listed as threatened species in Oregon since 1975, under the Oregon Endangered Species Act, and is fully protected under management authority of the ODFW (Anglin 2013, pers. comm.).

A Conservation Strategy for conserving the State's fish and wildlife has been prepared by the ODFW. The Conservation Strategy identifies 294 Strategy Species, which are Oregon's SGCN, (including wolverine) and are defined as those species having small or declining populations, are

[5] The U.S. Congress created the State Wildlife Grant (SWG) funding program in 2000 (Title IX, Public Law 106-553 and Title 1, Public Law 107-63). SWG funds are to be used "…for the planning and implementation of [States and territories] wildlife conservation and restoration program and wildlife conservation strategy, including wildlife conservation, wildlife conservation education, and wildlife-associated recreation projects." Congress stipulated that each State or territory applying for this funding program must develop a wildlife conservation strategy (State Wildlife Action Plan (SWAP)) by October 1, 2005. All 56 states and territories submitted SWAPs by 2005 and made commitments to review and/or revise their SWAP at least every 10 years.

at-risk, and/or are of management concern (ODFW 2016). For each of the Strategy Species, the Conservation Strategy identifies information on the special needs, limiting factors, data gaps, and conservation actions. For wolverine, conservation actions include management of recreational use to avoid impacts to the species (ODFW 2016). Other Strategy Species identified in the State's Conservation Strategy are prey species important to wolverine, including the Rocky Mountain bighorn sheep and Columbian white-tailed deer (ODFW 2016).

Washington

The wolverine is a candidate species for listing in the State of Washington and, since 2006, the Washington Department of Fish and Wildlife (WDFW) has been collaborating with wolverine researchers in the Cascades of northern Washington and southern British Columbia to better understand the status, distribution, and general ecology of wolverines in this region (WDFW 2013). It is also considered a SGCN, and is identified as a species whose population is in critical condition (WDFW 2013, pp. 3-7).

Washington's State Wildlife Action Plan (updated in 2015) identifies several major conservation strategies to address the conservation of fish and wildlife habitat and biodiversity in Washington, on both public and private lands (WDFW 2015, pp. 2-12–2-28). The wolverine is included in several identified ecological systems of concern such as alpine scrub, forb meadow, and grassland vegetation, cliff, scree and rock vegetation, and temperate forests (WDFW 2015, pp. 4-19, 4-27, 4-98). The State's *Wildlife Action Plan* identifies major stressors and key actions needed to maintain habitat quality for each of these ecological systems.

Of relevance to wolverine, the WDFW and its partners have been targeting land acquisition and conservation easements with high habitat or biodiversity values such as mixed-conifer forests as well as areas that support winter range and connectivity for wolverine and other carnivores (e.g., Methow River and Okanogan River Watersheds projects) (WDFW 2015, pp. 2-15–2-17). Other landscape conservation efforts highlighted in the State's Wildlife Action Plan include a Federal-State partnership with Washington's Department of Transportation to implement the Interstate-90 Snoqualmie Pass East Project to enhance wildlife connectivity that includes wildlife underpasses under the highway along creeks and rivers and two 150-foot wide wildlife bridges over the highway (WDFW 2015, p. 2-26).

Idaho

In Idaho, the wolverine is a protected nongame species and SGCN in Idaho (IDFG 2014). The *Idaho State Wildlife Action Plan, 2015* is a statewide plan for conserving and managing Idaho's fish and wildlife and their habitats, and provides a framework for conserving Idaho's 205 SGCN and their habitats, which includes the wolverine (IDFG 2017b, pp. xv–xviii). The wolverine is identified as a Tier 1 SGCN, which indicates it represents a species of most critical conservation need (IDFG 2017b, p. xvi). The statewide plan presents a species assessment for each SGCN and ecological section plans. Each of the ecological section plans presents a conservation target (e.g., habitat, species assemblage) that summarizes its viability as well as prioritized threats and strategies (IDFG 2017b, p. xv). A section outlining species designation, planning, and monitoring is also provided. The wolverine is included in three of the defined conservation

targets—forested lowlands, subalpine-high montane conifer forest, and low density forest carnivores (IDFG 2017b, p. 76). Along with objectives and strategies, these summaries identify actions for the SGCNs included in the defined conservation targets. Examples include: develop and implement a long-term multi-taxa monitoring program; determine high risk areas for wildlife crossings; construct highway over- and underpasses; promote and/or facilitate the use of prescribed fire as a habitat restoration tool, on both public and private lands where appropriate; determine best management practices to maintain cool microsites and benefit cool air associated species; and implement strategies to minimize disturbance from winter recreation activities as outlined in the *Management Plan for the Conservation of Wolverines in Idaho, 2014–2019* (IDFG 2017b, pp. 79, 80, 91, 94, 110).

The *Management Plan for the Conservation of Wolverines in Idaho, 2014–2019* (Management Plan) (IDFG 2014, entire) represents a framework for proactive efforts to ensure the long-term persistence and viability of wolverine populations in Idaho (IDFG 2016, pers. comm.). The Management Plan is described as a voluntary guidance document to lead conservation efforts at the State and local level, as well as to facilitate communication and collaboration efforts among wildlife and land managers (IDFG 2014, p. v).

Conservation issues and management actions are described in the Management Plan and the appropriate section plans of the *Idaho State Wildlife Action Plan*. The recommended strategies include development of finer-scale climate projections, research regarding wolverine-snow relationships, characterizing wolverine response to recreational activities, developing predictions of the potential overlap of wolverine and high levels of snow-sports recreation, and educating trappers to minimize incidental trapping of nontarget species, including the wolverine (IDFG 2014, pp. 32–39; IDFG 2017b, p. 1,058). Seven conservation and management objectives are outlined in the Management Plan (IDFG 2014, pp. 32–39) and, as outlined in a November 2016 response letter, there has been progress on all of these objectives (IDFG 2016, pers. comm.). As an example, the agency (under the Multi-species Baseline Initiative) has developed and implemented a baseline micro-climate monitoring protocol for collecting environmental parameters in an effort to identify areas that serve as cool-air refugia (IDFG 2016, pers. comm.). As described above (*Overutilization for Commercial, Recreational, Scientific, or Educational Purposes*), the IDFG has prepared educational materials to promote best management practices for minimizing non-target wolverine captures and continues to educate trappers under a legislative mandate passed in 2016 (State of Idaho House Bill 378) (IDFG 2016, pers. comm.).

In addition, management of prey species important to the wolverine diet is outlined in the Idaho Elk Management Plan 2014-2024 (IDFG 2014a), the Mule Deer Management Plan 2008-2017 (2008) and the Bighorn Sheep Management Plan (2010).

Montana

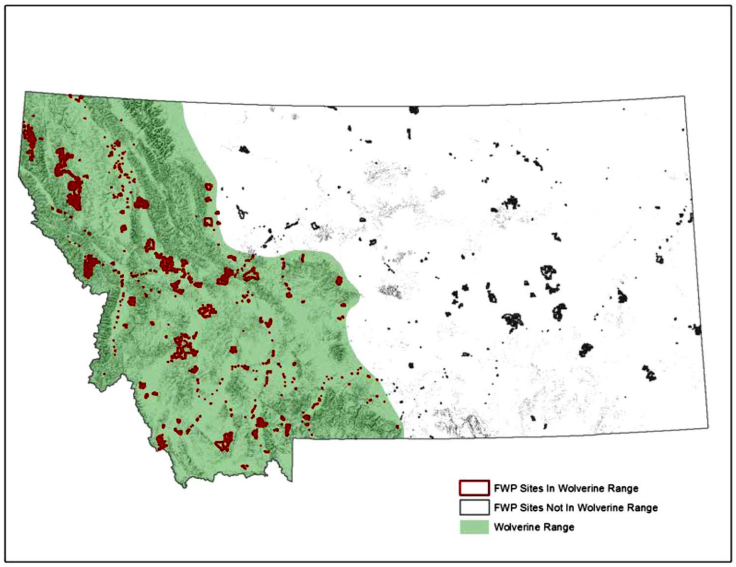
In the State of Montana, the wolverine is classified as a furbearer and species of concern. Since 2013, there has been a zero quota for trapping or harvest of wolverine and trappers that capture a wolverine must notify a designated Montana FWP employee within the relevant trapping district within 24 hours for collection if the animal cannot be released uninjured (Montana FWP 2016, pers. comm.).

There are two broad-scale wildlife conservation efforts that provide conservation benefits to the wolverine. *Montana's State Wildlife Action Plan* (updated and revised in 2015) identifies the wolverine as one of 128 SGCN (Montana FWP 2015, Appendix N). The State's Wildlife Action Plan identifies priority community types, focal areas, and species to help informing Montana FWP's priorities and decisions and to assist other agencies and organizations in making decisions as to where to focus their conservation efforts (Montana FWP 2015, p. 2). Community types and focal areas are designed to identify and direct attention to specific geographical areas in the State that have the greatest conservation need (Montana FWP 2015, p. 5). For the wolverine, *Montana's State Wildlife Action Plan* identifies wolverine habitats in seven community types, all designate Tier I (or those with greatest conservation need), and in all focal areas (also Tier I) within those community types (Montana FWP 2016, pers. comm.). For each community type, impacts, threats, and corresponding conservation actions are identified, as well as specific impacts and threats such as habitat fragmentation (e.g., prioritize land acquisition, provide wildlife under- and overpasses), land management (e.g., management to address altered fire regimes), recreation (e.g., consider seasonal closures during breeding season), and climate change (e.g., collection of baseline data to document shifting range limits of SGCN and Community Types of Greatest Conservation Need) (Montana FWP 2015, pp. 59–63).

The second conservation effort in the State of Montana is a Crucial Area Assessment to identify crucial areas and fish and wildlife corridors, and development of a Crucial Areas Planning System (URL: http://fwp.mt.gov/fishAndWildlife/conservationInAction/crucialAreas.html). This is a Montana FWP mapping application and planning tool designed to assist in future planning of development and conservation (Montana FWP 2016, pers. comm.).

The State of Montana is also conserving wildlife habitat through land acquisition and conservation easements (Montana FWP 2016, pers. comm.) using a landscape scale collaborative approach with other agencies (Montana FWP 2016, pers. comm.; Montana FWP 2017, pers. comm.). In western Montana, including areas known to be occupied by the wolverine, 425 properties for a total 310,523 ha (767,320 ac) have been either acquired (e.g., State Parks, Wildlife Management Areas) or protected by conservation easements, as of November 2016, as shown in the figure below (Montana FWP 2016, pers. comm.). In addition, beginning in 2008, the agency shifted harvest units and quotoas to emphasize protection in smaller mountain ranges (closing 40 percent of the State to harvest), and to provide protection in places where reproduction is occurring in-between large ecosystems, to ensure dispersal and genetic exchange (Montana FWP 2017, pers. comm.).

Other regulatory mechanisms implemented by the State include rules and regulations promulgated by the Montana Fish and Wildlife Commission and codified in State Statute (MCA 87-1-201), and have been in place since 1921. All furbearer regulations are reviewed on an annual basis.



Wyoming

The wolverine is a protected animal and SGCN in Wyoming (WGFD 2017). The *Wyoming Game and Fish Department State Wildlife Action Plan* directs the activities of the WGFD and serves to guide in conserving Wyoming's SGCN through the combined efforts of government agencies, conservation organizations, academia, tribes, and others (WGFD 2017, p. I–1-1). As noted above, the wolverine is identified as a SGCN, a designation intended to identify species whose conservation status warrants increased management attention and funding, and consideration in conservation, land use, and development planning in the State (WGFD 2017, p. IV– i-1). The *State Wildlife Action Plan* incorporates the wolverine as a SGCN in several terrestrial habitat types or ecological systems, including cliffs, canyons, and rock outcrops, montane and subalpine forests, and mountain grasslands and alpine tundra (WGFD 2017, pp. III–2-5, III–5-7, III–6-5).

In 2015, Wyoming funded a pilot project (through The Wolverine Initiative) to evaluate wolverine detection and monitoring of the species in the State and is a contributing collaborator in the Multistate Wolverine Working Group implementing a monitoring strategy (the WSWCP) in the winter of 2016–2017 across four western states (WGFD 2017, p. IV–5-357). Results of those studies (e.g., Inman *et al*. 2015) are summarized above (*Population Abundance and Distribution*). The WSWCP is also updating and refining connectivity models for the wolverine in an effort to focus and prioritize habitat conservation and management (WGFD 2016, pers. comm.).

Utah

The wolverine is protected from collection, importation, and possession in Utah. It is identified as a SGCN in the *Utah Wildlife Action Plan* (UDNR 2015). The plan's threat assessment lists the wolverine as a Level 3 threat based on natural rarity, with a high threat impact (severity x scope) (UDNR 2015, p. 149). As a SGCN, wolverine conservation and management projects are eligible for State Wildlife Grant funding, including landscape-level conservation efforts (UDNR 2017, pers. comm.). The wolverine is also identified in the *Utah Wildlife Action Plan* as a species with a data gap relative to an inadequate understanding of distribution or range (UDNR 2015, p. 174).

Although the State of Utah is not a formal partner in the WAFWA Western States Wolverine Conservation Project (WSWCP)–Coordinated Occupancy Survey, the Division of Wildlife Resources selected grid cells for monitoring wolverines with remote cameras in accordance with the WAFWA protocol. Cameras were placed in potential wolverine habiat in the Unita and Bear River mountain areas from 2014–2017, and the State intends to continue camera monitoring for wolverines (UDNR 2017, pers. comm.).

Colorado

The wolverine is a state-endangered species in Colorado (Colorado Parks and Wildlife 2015a); however, there is no known current resident or reproducing wolverine population.

The *Colorado State Action Plan* (Colorado Parks and Wildlife 2015b) provides a blueprint for a collaborative effort to conserve Colorado's at-risk wildlife and their habitats, with a primary goal for securing wildlife populations in order to avoid protections implemented so that they do not require protection via federal or state listing regulations (Colorado Parks and Wildlife 2015b, p. 1). The wolverine is designated as a Tier 1 (highest conservation priority; up from Tier 2) SGCN (Colorado Parks and Wildlife 2015b, p. 19). The primary conservation action for wolverine described in the 2015 State Action Plan is to continue discussions among wildlife managers, conservation partners and stakeholders of the social and political aspects regarding reintroduction of wolverine populations into the southern Rocky Mountains (Colorado Parks and Wildlife 2015b, p. 186). The State has not yet prepared a potential restoration program for the species (Broscheid 2016, pers. comm.).

Western Association of Fish and Wildlife Agencies (WAFWA) – Western States Wolverine Working Group

The WAFWA, in coordination with Tribal partners, have formed a multi-state, multi-agency working group (Western States Wolverine Working Group) to design and implement the Western States Wolverine Conservation Project (WSWCP)–Coordinated Occupancy Survey. In addition, a WAFWA Wildlife Chiefs Wolverine Subcommittee provides a forum for western States to work collaboratively with each other and with the Service, Tribes, and other partners, for conserving wolverines across the western United States. To date, approximately $1.5 million of that funding has been applied towards conservation and management actions, including the WSWCP (McDonald 2017, pers. comm.).

Other Conservation Mechanisms

Tribes

Nez Perce Tribe

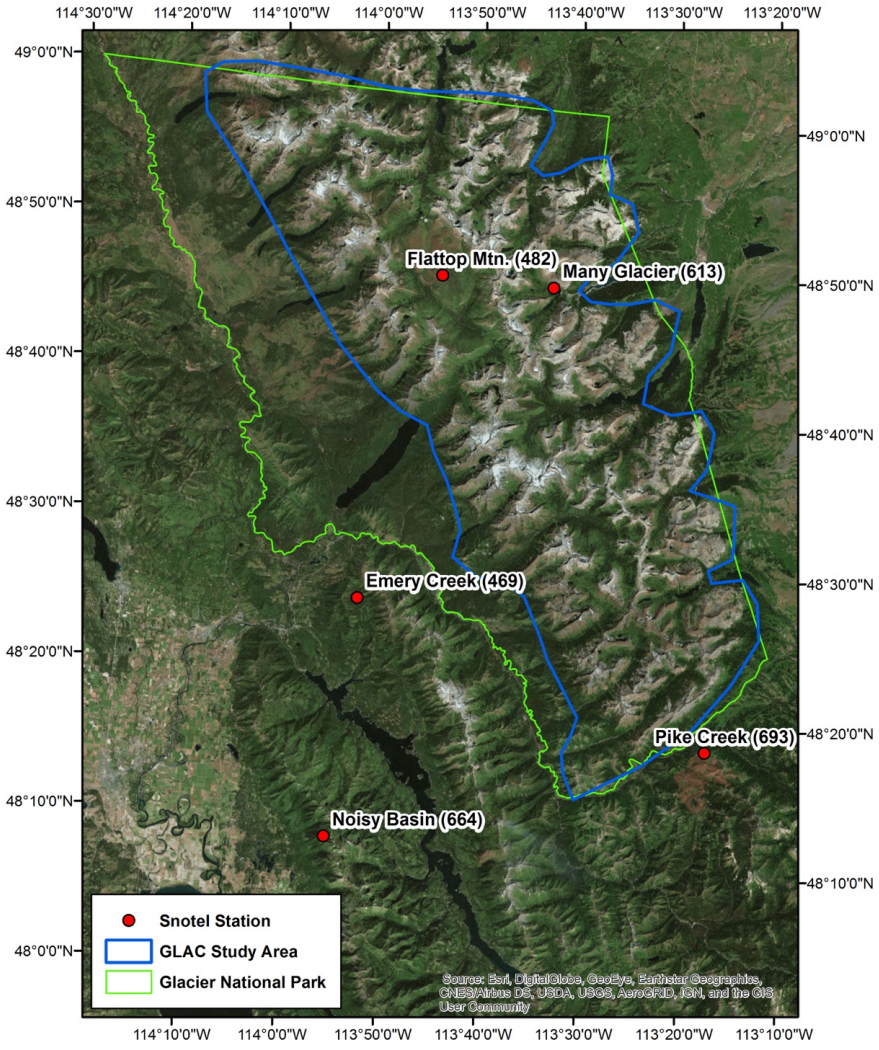
Wolverines are found within the aboriginal territory of the Nez Perce Tribe in north-central Idaho, and conservation and restoration of the species within the Nez Perce homeland is important to the Nez Perce Tribe (Miles 2017, pers. comm.). The Nez Perce Tribe is currently preparing an Integrated Resource Management Plan (IRMP), a Plant and Wildlife Conservation Strategy, and a Forest Management plan with the wolverine defined as a species of conservation concern in all three draft plans (Miles 2017, pers. comm.). The planning area for the IRMP, which is being prepared in partnership with the Bureau of Indian Affairs, incorporates the approximately 311,608 ha (770,000 ac) Nez Perce Reservation, located within portions of Nez Perce, Lewis, Clearwater, Latah, and Idaho Counties in north-central Idaho (http://www.nezperce.org/irmp/; accessed August 24, 2017). The preparation of the IRMP is currently at the scoping stage in the NEPA process for development of a Programmatic Environmental Impact Statement (http://www.nezperce.org/irmp/; accessed August 24, 2017).

The Shoshone-Bannock Tribes

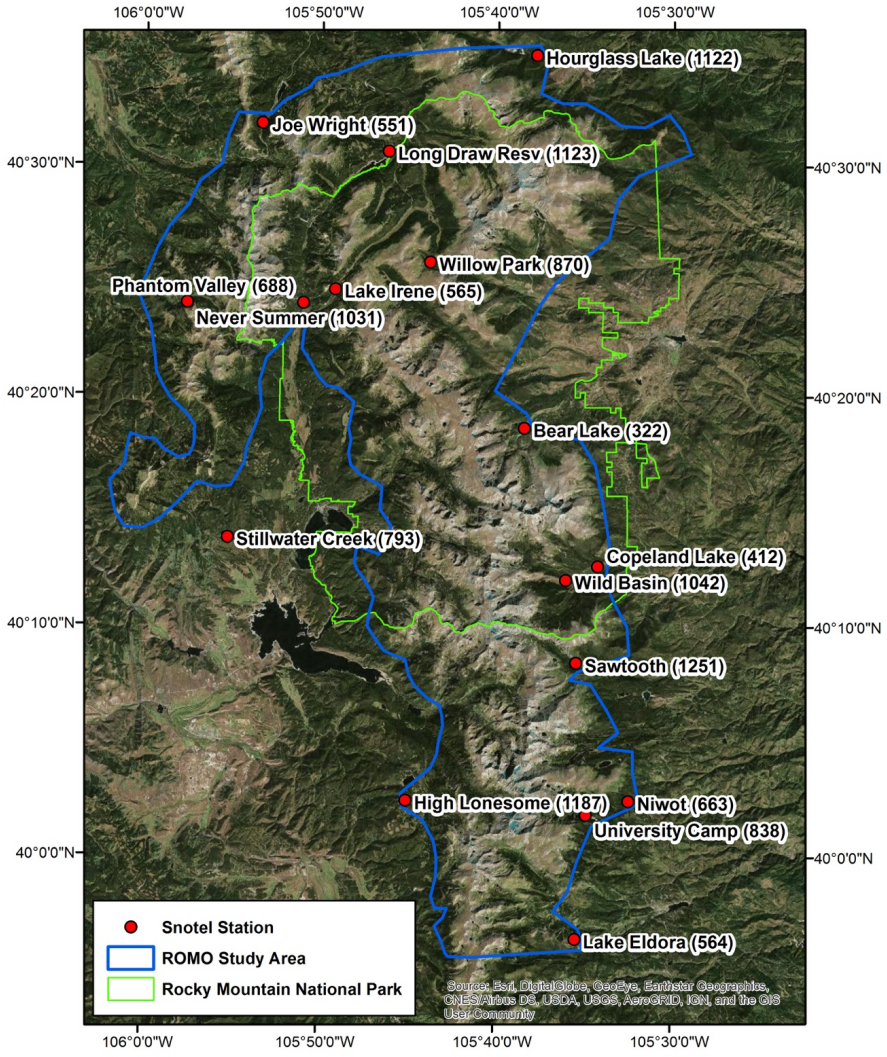
The Shoshone-Bannock Tribes are currently conducting climate change modeling for the Northern Rocky Mountains as part of its preparation of a Climate Change Adaptation Plan (Edmo 2016, pers. comm.). The Upper Snake River Tribes Foundation (USRT), which is comprised of four member tribes—the Burns Paiute Tribe, Fort McDermitt Paiute-Shoshone Tribe, Shoshone-Bannock Tribes of the Fort Hall Reservation, and Shoshone-Paiute Tribes of the Duck Valley Reservation—within the Upper Snake River Watershed region, prepared a *Climate Change Vulnerability Assessment* in February 2017 (Petersen *et al*. 2017, entire). The assessment is the first of three steps the USRT and its member tribes plan activities over the next several years as part of a comprehensive climate change effort, and will include an Adaptation Plan (expected to be completed in 2017–2018), and, depending on future funding, a process for development of Implementing Adaptation Actions and Monitoring (Petersen *et al*. 2017, p. 7).

Appendix H–NOAA/CU Study Areas Used to Evaluate Future Snow Persistence (from Ray et al., 2017)

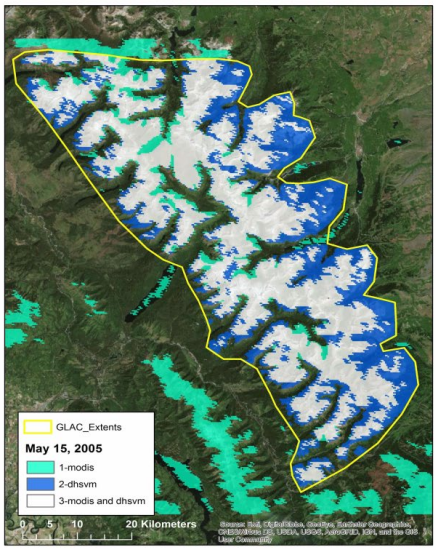
Glacier National Park Study Area

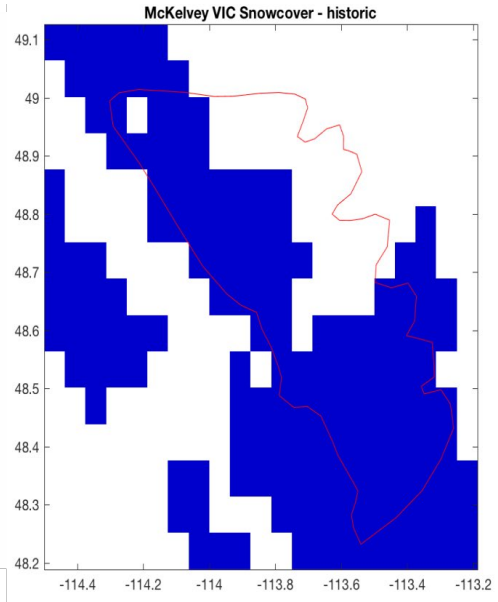


Rocky Mountain National Park Study Area



Appendix I–Visual comparsion of spatial resolution of Ray et al. (2017) with McKelvey et al. (2011)
(from Ray et al., 2017; Figure 2-4)





From: Shoemaker, Justin
To: Nelson, Marjorie
Subject: Re: Invitation: Wolverine Director presentation revisions @ Wed Feb 14, 2018 2:30pm - 3:30pm (MST) (marjorie_nelson@fws.gov)
Date: Wednesday, February 14, 2018 1:25:12 PM

I was thinking hangout would be easy and allow us to see revisions in real-time.

Justin Shoemaker
Classification and Recovery Biologist
U.S. Fish and Wildlife Service, Region 6
Phone: 309-757-5800 x214
Email: justin_shoemaker@fws.gov

On Wed, Feb 14, 2018 at 2:03 PM, Nelson, Marjorie <marjorie_nelson@fws.gov> wrote:

> Are we doing vidyo (I know you hate that) or google hangout or webex for 2:30?
>
> Marjorie Nelson
> Chief, Division of Ecological Services
> Mountain-Prairie Region
> U.S. Fish and Wildlife Service
> 303-236-4258 direct
> 720-582-3524 cell
>
> On Wed, Feb 14, 2018 at 11:09 AM, Justin Shoemaker <justin_shoemaker@fws.gov> wrote:
>
> **Wolverine Director presentation revisions** **more details »**
>
> Wolverine Director briefing dry run debrief and presentation revisions.
>
> When Wed Feb 14, 2018 2:30pm – 3:30pm Mountain Time
>
> Video call https://plus.google.com/hangouts/_/doi.gov/justin-shoemake
>
> Calendar marjorie_nelson@fws.gov
>
> Who
> - justin_shoemaker@fws.gov - creator
> - john_guinotte@fws.gov
> - marjorie_nelson@fws.gov
> - betty_grizzle@fws.gov
> - jodi_bush@fws.gov
>
> Going? **Yes** - **Maybe** - **No** more options »
>
> Invitation from Google Calendar
>
> You are receiving this email at the account marjorie_nelson@fws.gov because you are subscribed for invitations on calendar marjorie_nelson@fws.gov.
>
> To stop receiving these emails, please log in to https://www.google.com/calendar/ and change your notification settings for this calendar.
>
> Forwarding this invitation could allow any recipient to modify your RSVP response. Learn More.

From: Nelson, Marjorie
To: Shoemaker, Justin
Subject: Re: Invitation: Wolverine Director presentation revisions @ Wed Feb 14, 2018 2:30pm - 3:30pm (MST) (marjorie_nelson@fws.gov)
Date: Wednesday, February 14, 2018 1:27:04 PM

---

great!

Marjorie Nelson
Chief, Division of Ecological Services
Mountain-Prairie Region
U.S. Fish and Wildlife Service
303-236-4258 direct
720-582-3524 cell

On Wed, Feb 14, 2018 at 1:25 PM, Shoemaker, Justin <justin_shoemaker@fws.gov> wrote:

> I was thinking hangout would be easy and allow us to see revisions in real-time.
>
> Justin Shoemaker
> Classification and Recovery Biologist
> U.S. Fish and Wildlife Service, Region 6
> Phone: 309-757-5800 x214
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> >
> > > **Wolverine Director presentation revisions** **more details »**
> > >
> > > Wolverine Director briefing dry run debrief and presentation revisions.
> > >
> > > When Wed Feb 14, 2018 2:30pm – 3:30pm Mountain Time
> > > Video call https://plus.google.com/hangouts/_/doi.gov/justin-shoemake
> > > Calendar marjorie_nelson@fws.gov
> > > Who
> > > - justin_shoemaker@fws.gov - creator
> > > - john_guinotte@fws.gov
> > > - marjorie_nelson@fws.gov
> > > - betty_grizzle@fws.gov
> > > - jodi_bush@fws.gov
> > >
> > > Going? **Yes** - **Maybe** - **No** more options »
> > >
> > > Invitation from Google Calendar
> > >
> > > You are receiving this email at the account marjorie_nelson@fws.gov because you are subscribed for invitations on calendar marjorie_nelson@fws.gov.
> > >
> > > To stop receiving these emails, please log in to https://www.google.com/calendar/ and change your notification

settings for this calendar.

Forwarding this invitation could allow any recipient to modify your RSVP response. Learn More.

From: Shoemaker, Justin
To: Grizzle, Betty; Guinotte, John; Marjorie Nelson; Jodi Bush
Subject: webex for our wolverine call now
Date: Wednesday, February 14, 2018 2:19:32 PM

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http://www.mymeetings.com/nc/join.php?i=742652114&p=&t=c

Justin Shoemaker
Classification and Recovery Biologist
U.S. Fish and Wildlife Service, Region 6
Phone: 309-757-5800 x214
Email: justin_shoemaker@fws.gov

From: Inman, Bob
To: Grizzle, Betty
Subject: map
Date: Wednesday, February 14, 2018 8:28:47 AM

Hey Betty, I received your message. Happy to make a map for you. Will check with others first to do so.

-Bob Inman

Robert M. Inman, PhD
Carnivore-Furbearer Coordinator
Montana Fish, Wildlife and Parks
1420 East 6th Ave., PO Box 200701,
Helena, MT 59620-0701
406-444-0042 (o)
406-570-5326 (c)
bobinman@mt.gov

From: Grizzle, Betty
To: 222wsheridan@gmail.com
Subject: AK wolverine presentation
Date: Wednesday, February 14, 2018 9:39:52 AM
Attachments: M. Harrington Wolverines.jpg

Just an FYI - a colleague forwarded the attached announcement to me.

--

Betty J. Grizzle, D.Env., MPH
Fish and Wildlife Biologist
U.S. Fish and Wildlife Service
Carlsbad Fish and Wildlife Office
2177 Salk Ave, Suite 250
Carlsbad, CA 92008
760-431-9440, ext. 215
760-431-5901 fax

Wildlife Wednesdays

2017-2018 Free Wildlife Speaker Series At The Alaska Zoo







Photos courtesy of Mike Harrington (Top and middle) and John Gomes.

February 14 at 7pm

Tracking the Wolverine

Follow the tracks of wolverines as Mike Harrington, with the Alaska Department of Fish and Game, shares insights from a wolverine research project. The project, which used tracking collars and trail cameras, helped determine a more accurate population estimate, and to better understand wolverine movements, home ranges, and relationships with other wolverines.

Presentations will be held on the second Wednesday, October through April.
All talks will begin at 7pm in the Alaska Zoo's Gateway Education Building.
The doors will open at 6pm, coffee and tea will be available during lectures.

All Wildlife Wednesday talks are ASL interpreted.
Door prizes are donated by Alaska Geographic

Great door prizes!

Alaska Geographic · U.S. Fish & Wildlife Service · the Alaska Zoo · Alaska Department of Fish and Game · USGS science for a changing world

The Alaska Zoo 4731 O'Malley Road Anchorage, AK 99507 (907)341-6463

From: Grizzle, Betty
To: John Guinotte
Subject: Revised powerpoint
Date: Wednesday, February 14, 2018 11:23:52 AM
Attachments: Briefing for Deputy Director_Wolverine 12mo finding.pptx

---

Here is the version we walked through this morning. Haven't made any changes yet.
--

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North American Wolverine (*Gulo gulo luscus*): 12-Month Finding



Photo credit: Mark Packila; used with permission.

Mountain Prairie Regional Office

February 20, 2018

Listing History

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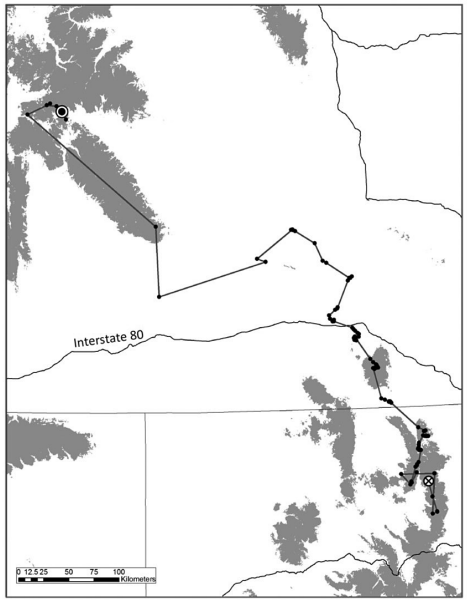
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Map of VHF locations of M56 - 2008 to 2009



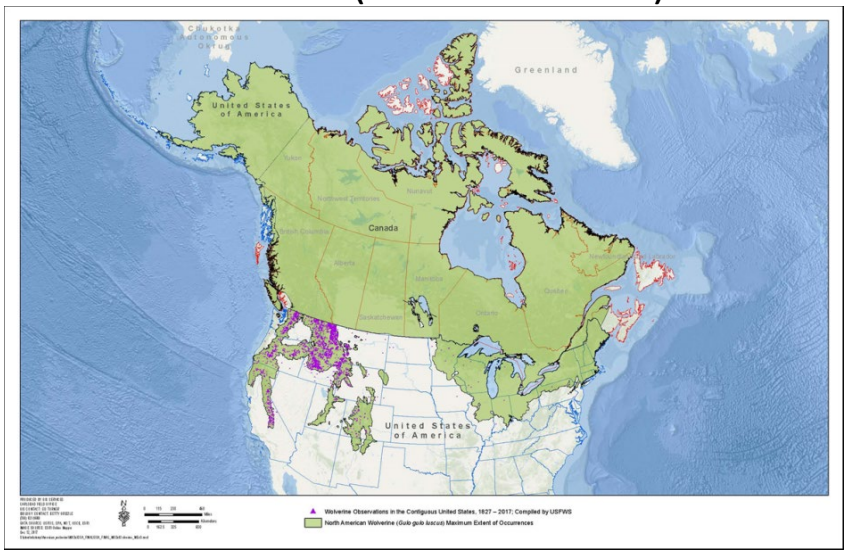
Suspected wolverine M56, from Packila et al. 2017

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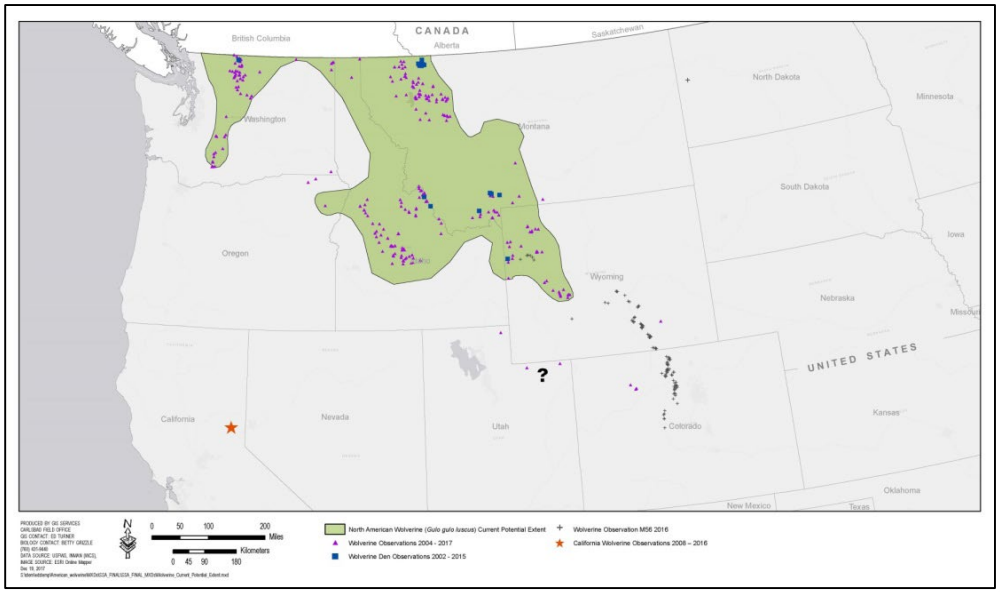
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North American Wolverine Range Maps

- Maximum Extent of Occurrences (North America)



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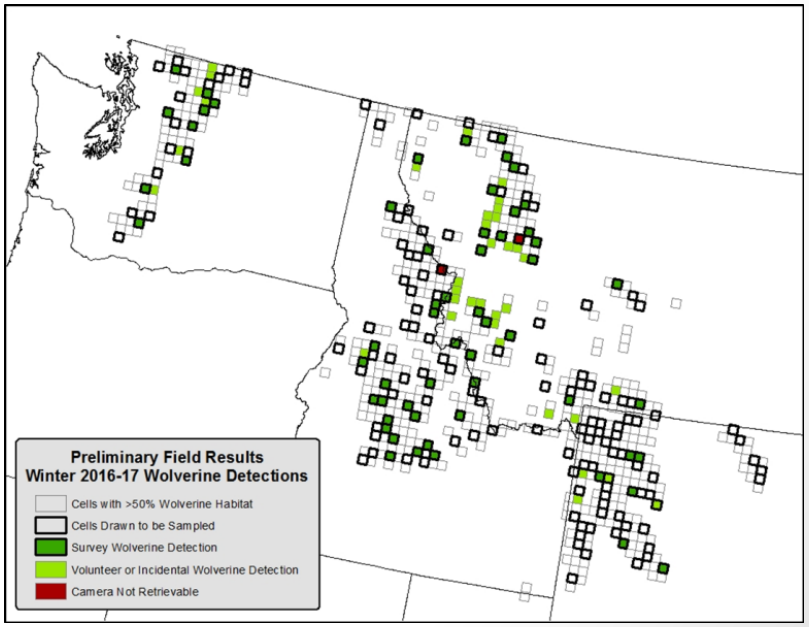
Denning "habitat" in Glacier NP; from Copeland and Yates 2008



March 2, 2017 den in Sweden (lower left, under boulder)

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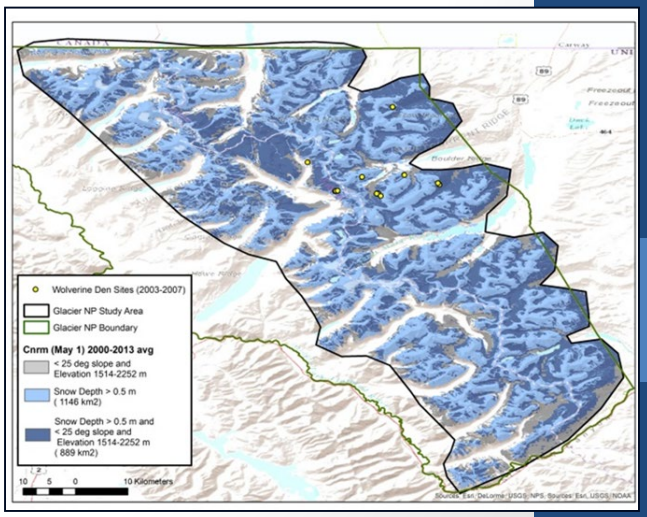


Potential Stressors Evaluated

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- Effects from wildland fire
- Disturbance – winter recreational activity, other infrastructure
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- Known or likely potential stressors represent low risk to the species
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Recommendation Management Team

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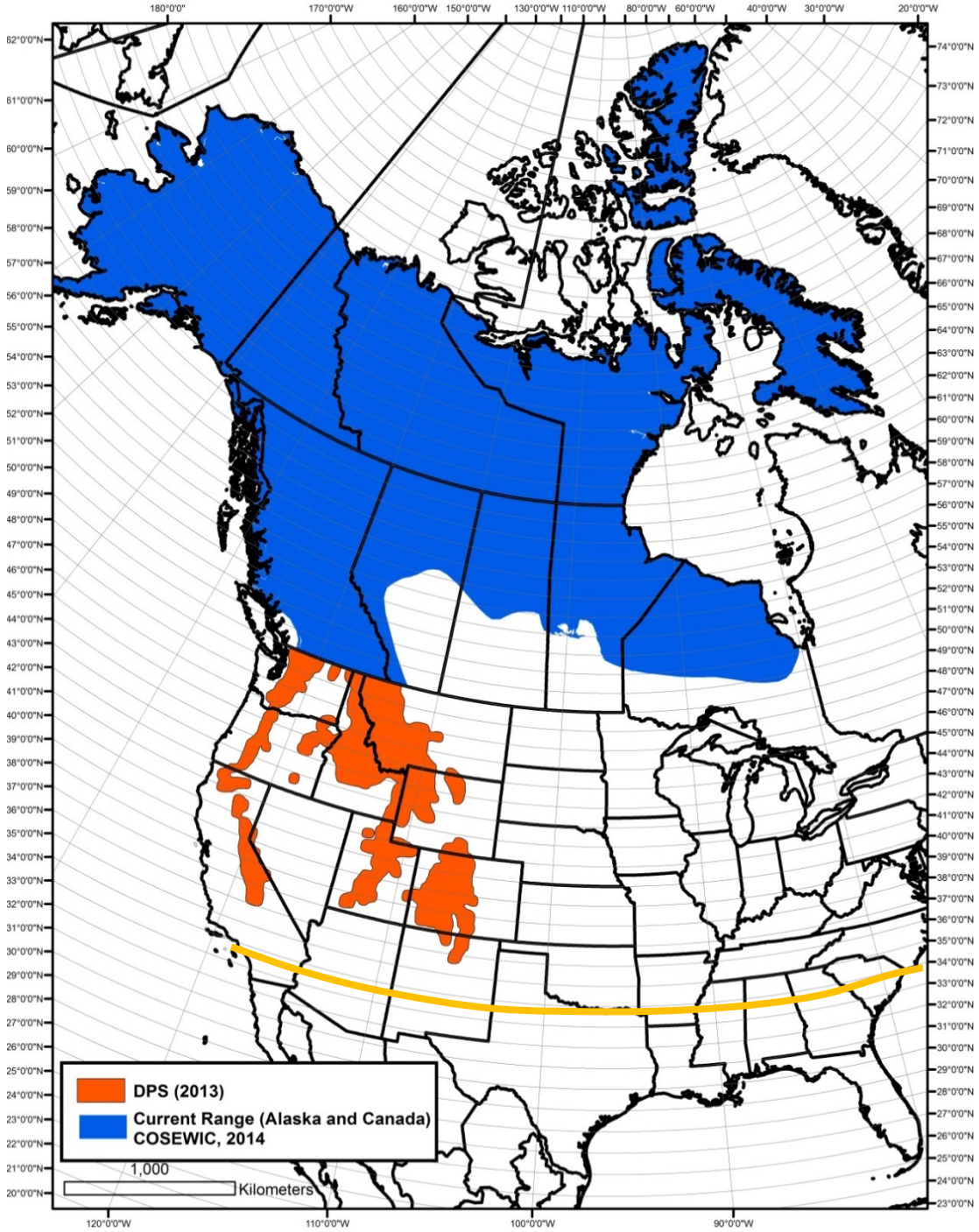
"Buddy" in California, May 2016; Photo credit: Chris Stermer, CDFW

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      - Exploitation: Did not rely on control of exploitation to establish discreteness
  - Significance = YES (see next slide)
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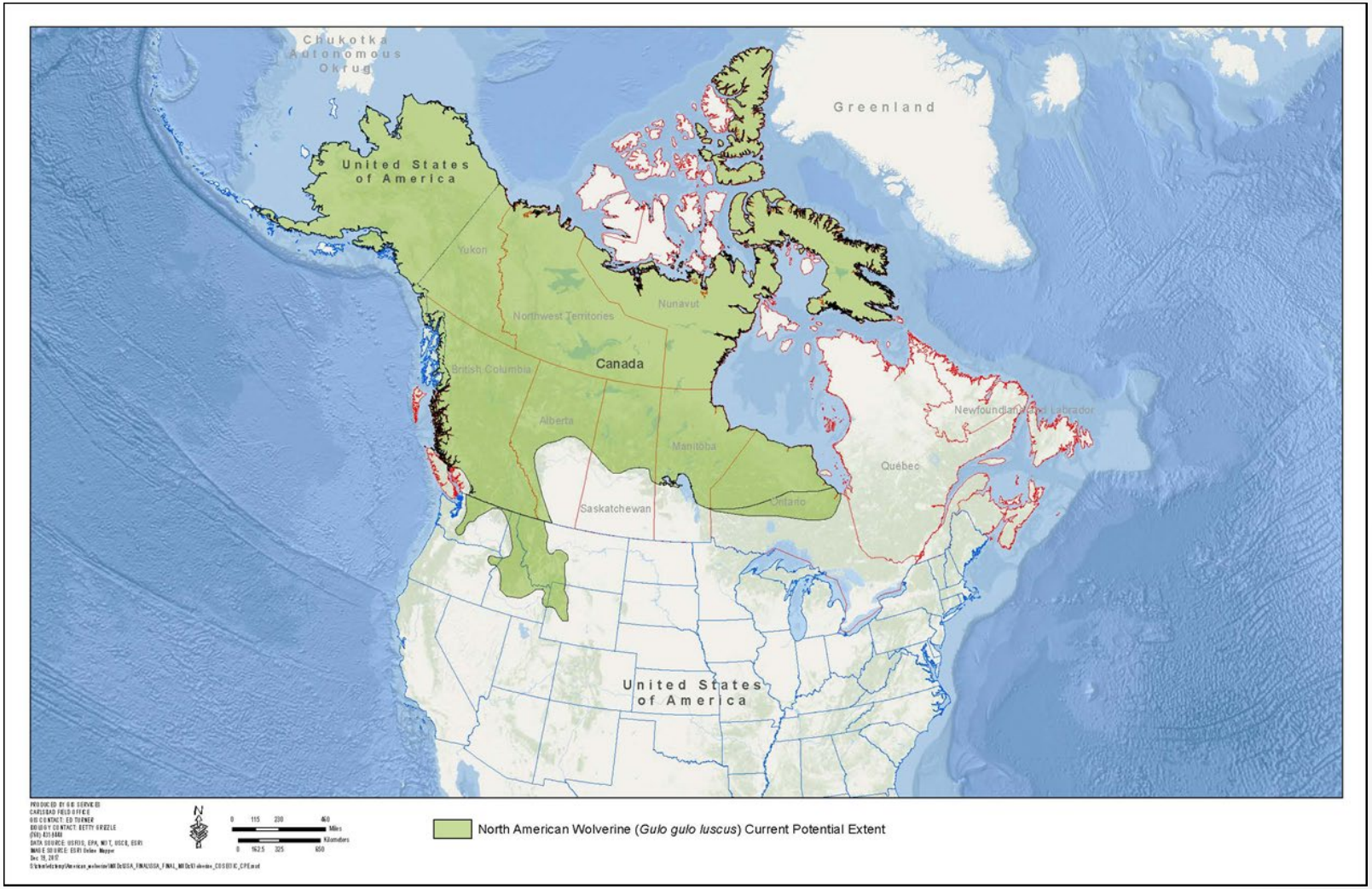


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Current Potential Extent – North America



From: Guinotte, John
To: Betty Grizzle
Subject: added slides
Date: Wednesday, February 14, 2018 1:49:17 PM
Attachments: Briefing for Deputy Director_Wolverine 12mo finding (1)_jmg.pptx

---

John Guinotte
Spatial Ecologist
Ecological Services
U.S. Fish and Wildlife Service
Mountain Prairie Region 6
134 Union Blvd., Lakewood, CO 80228
303-236-4264
john_guinotte@fws.gov

North American Wolverine (*Gulo gulo luscus*): 12-Month Finding



Photo credit: Mark Packila; used with permission.

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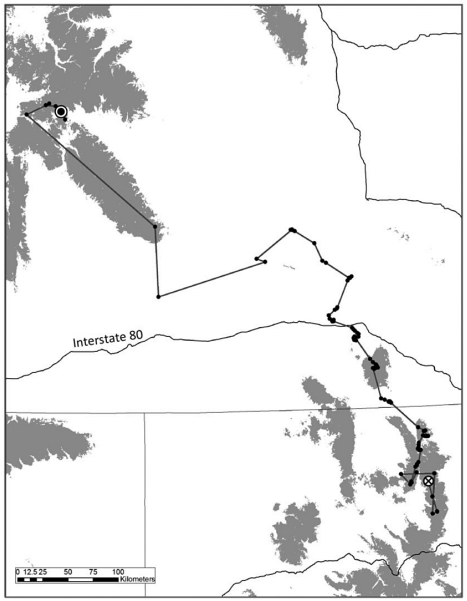
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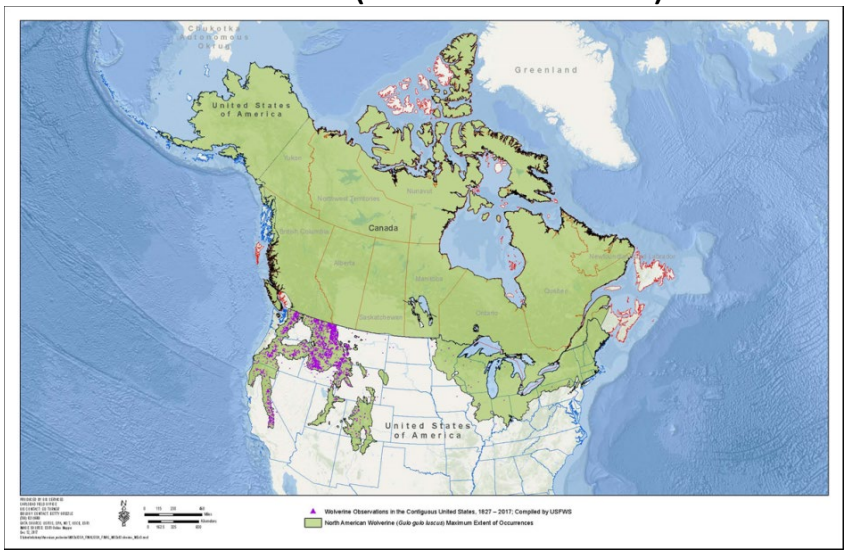
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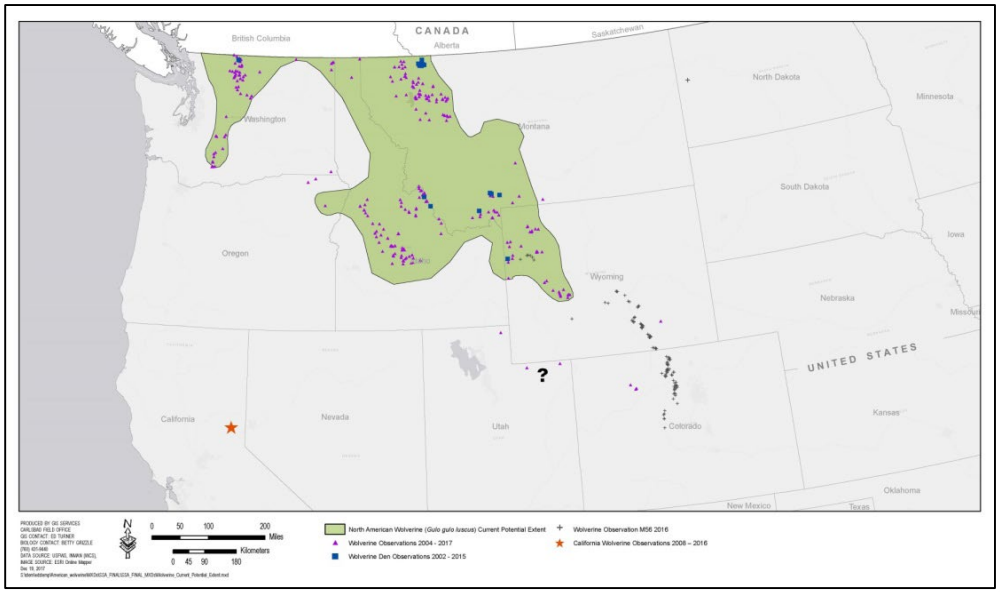
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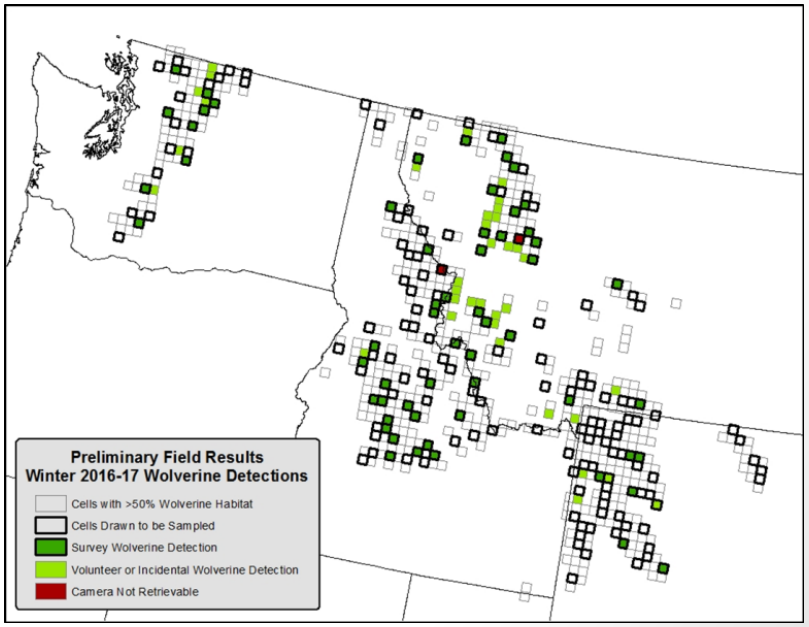
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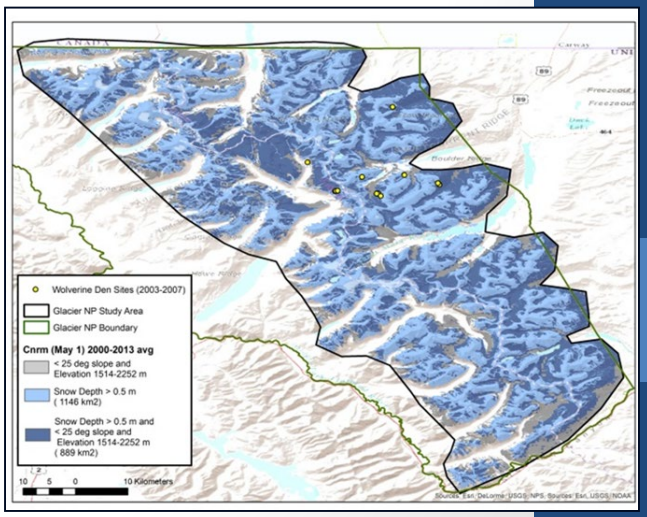


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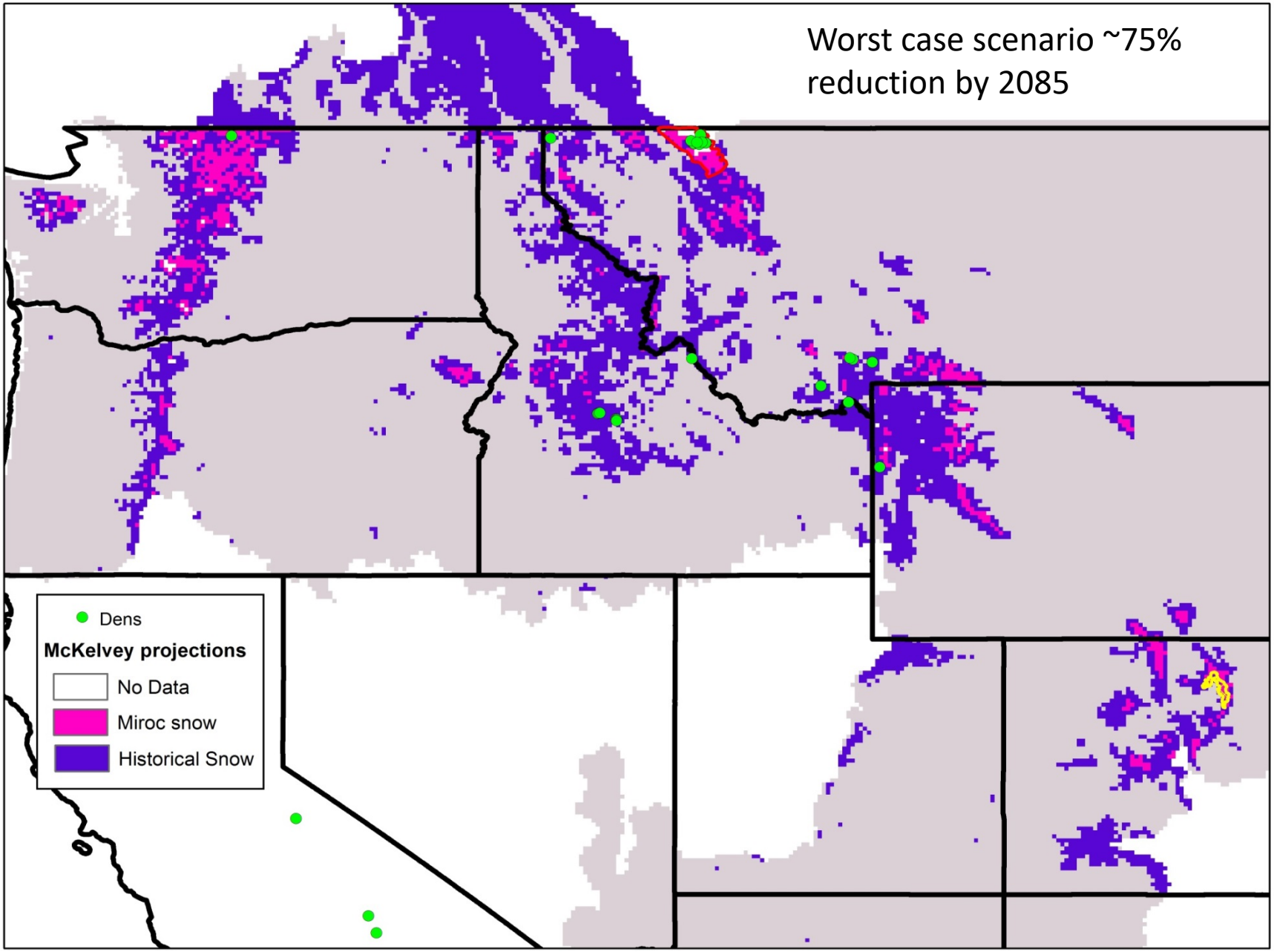
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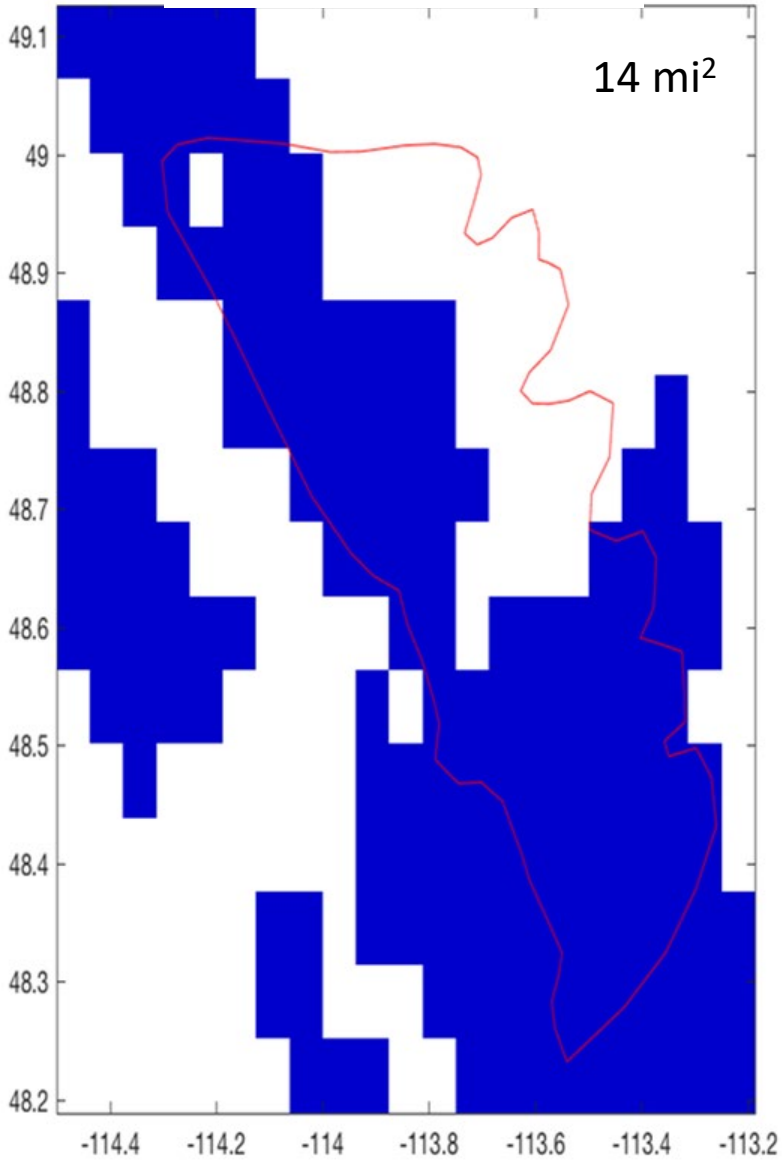


McKelvey et al (2011) snow projections

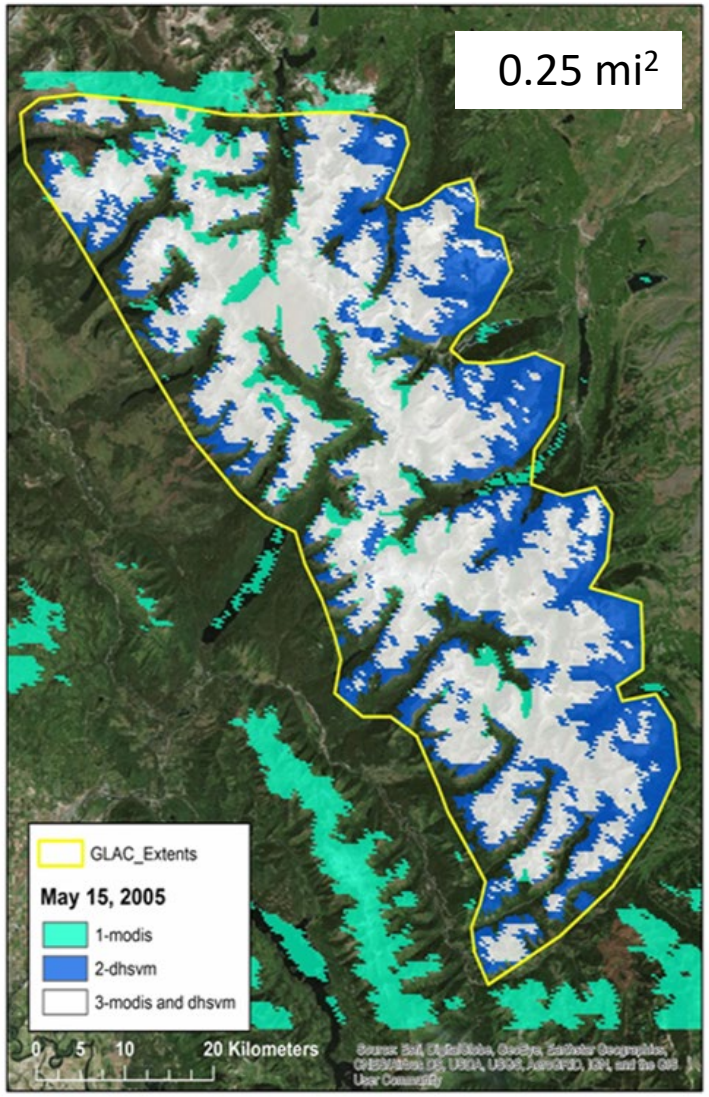


Improved Snow Projections (2017)

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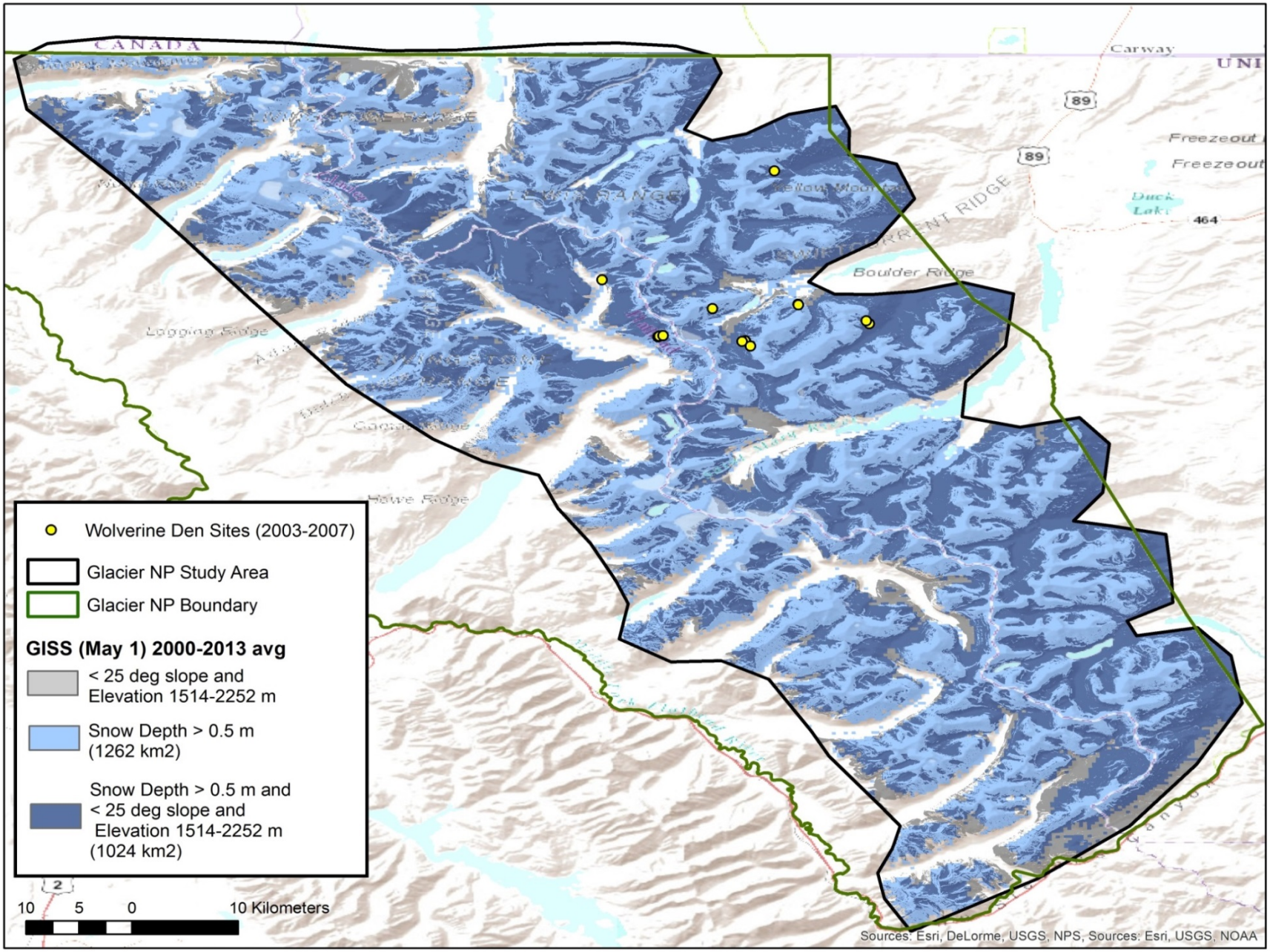


NOAA / CU 2017

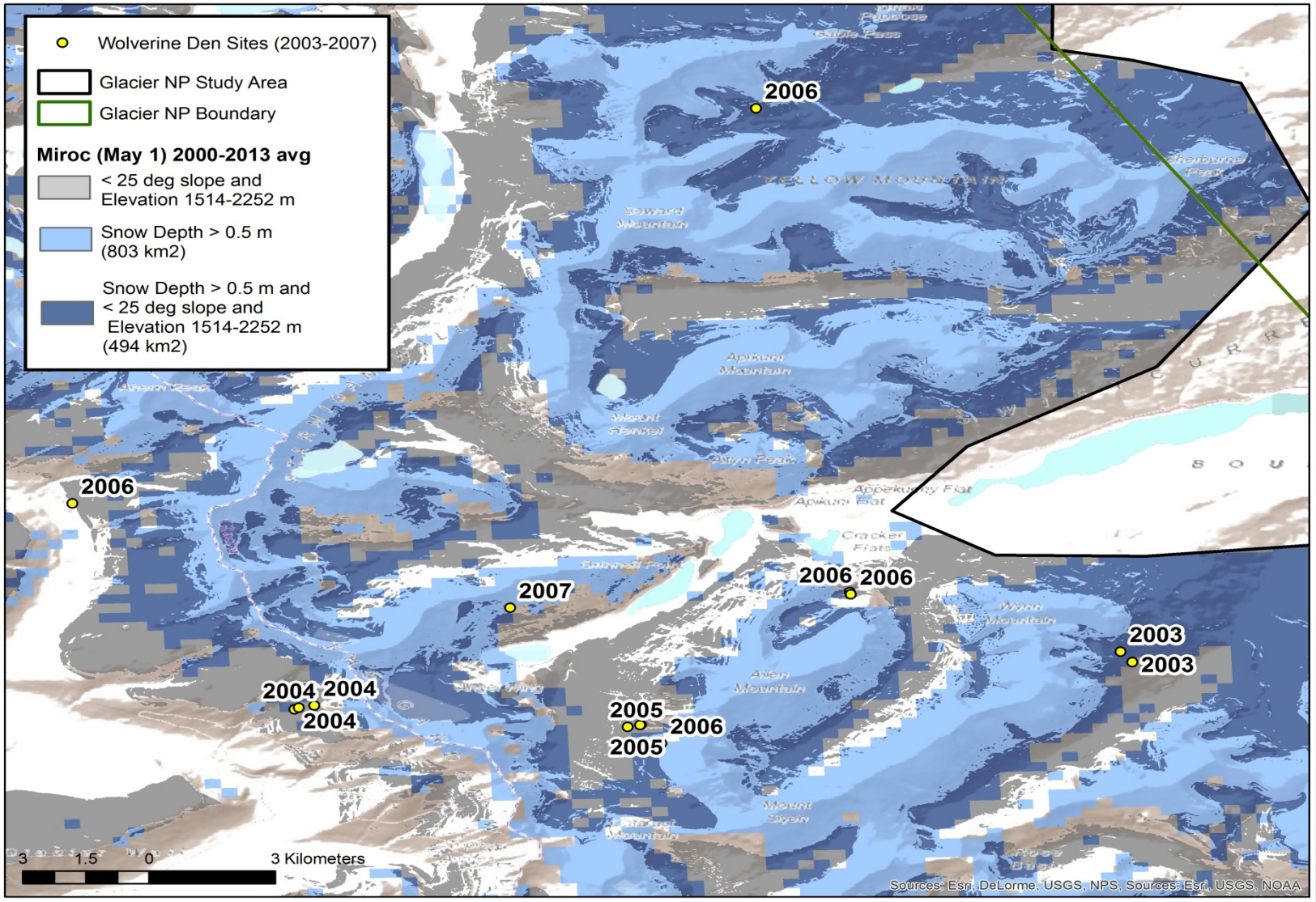


(John Guinotte, FWS)

Glacier NP Future Snow Projections (~2055)



Glacier Dens and Future Snow Projections



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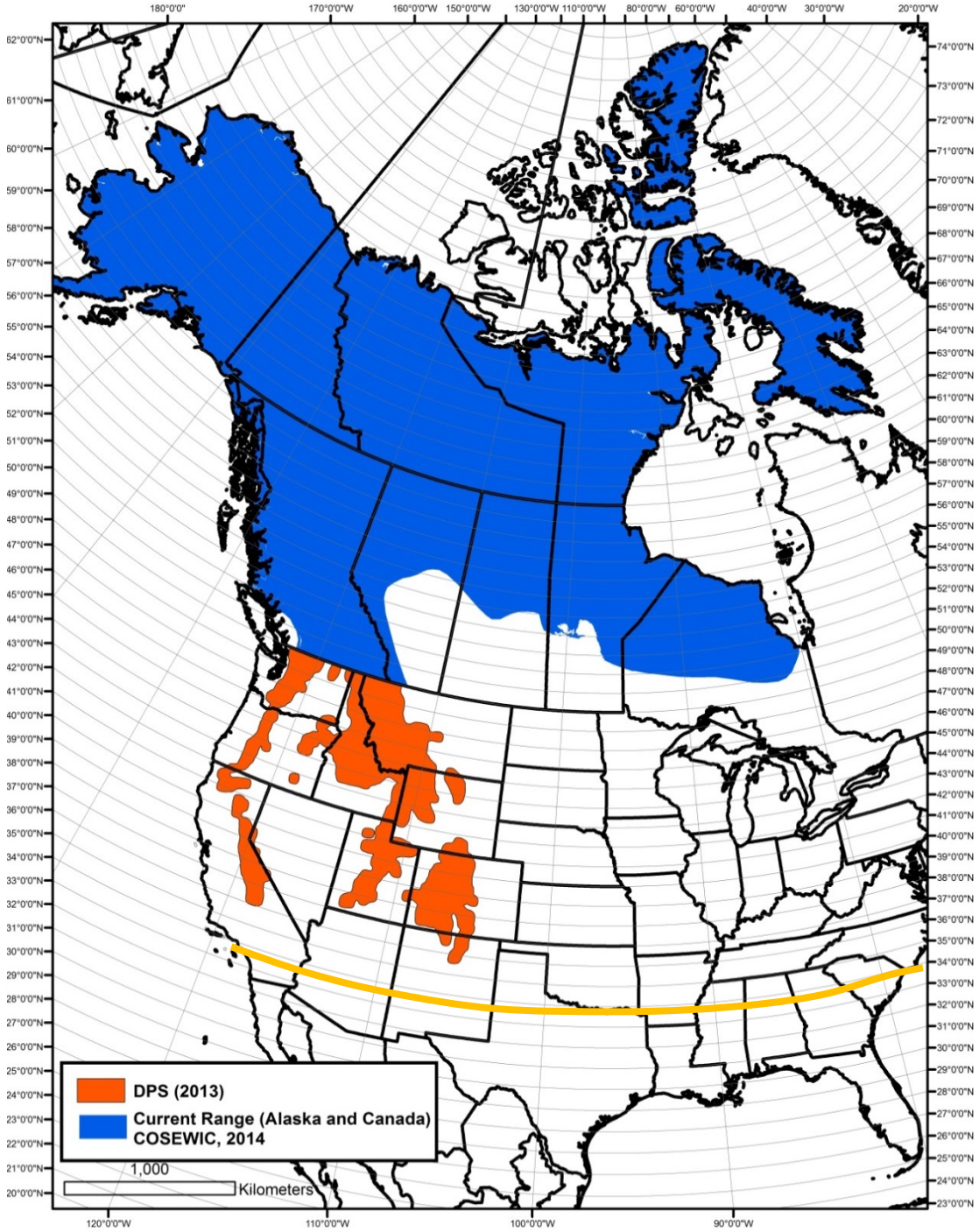
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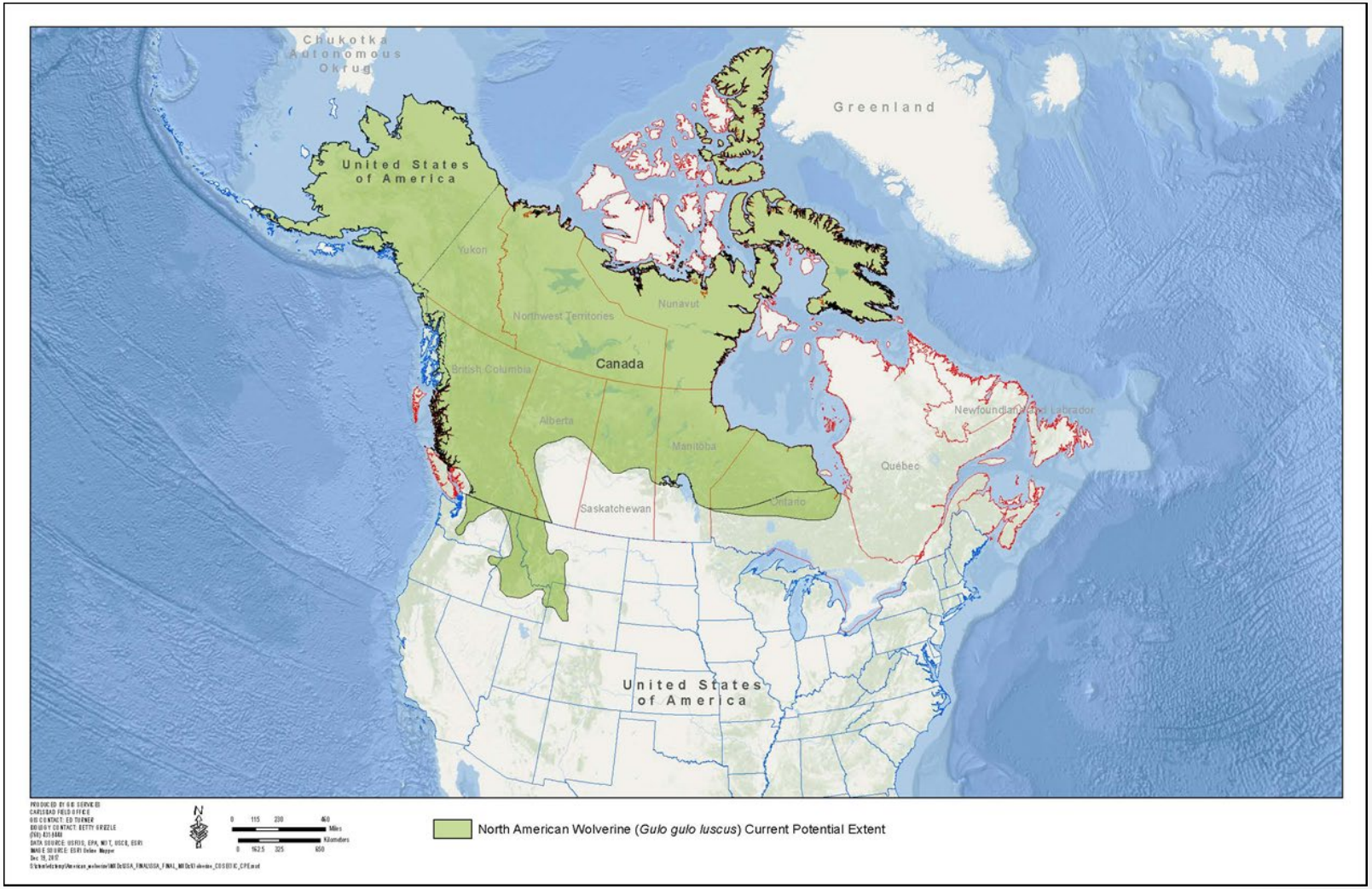


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Current Potential Extent – North America



Results for Glacier

- SCA (> 0.5m) declines in all scenarios
- Declines on April 15 range from 7-22%
- Declines on May 15 range from 12-50%
- Areas of frequent snow greater than 0.5m become concentrated at higher elevations, **largest decreases in lower elevations**
- **Below denning elevations, losses are large 40-70%, above that band, small losses occur, 2-8%**
- Climate and elevations of Glacier buffer significant changes in snow on May 1 above 1800m

From:	Turner, Ed
To:	Betty Grizzle
Subject:	COSEWIC with MEO and CPE acres
Date:	Wednesday, February 14, 2018 3:20:23 PM
Attachments:	CPE_COSEWIC_Acres.pdf MOE_Lower48_NT_Acres.pdf COSEWIC_CPE_NALAM.xlsx MEO_All_in_NaLAM.xlsx

---

Hi Betty, It gets a tad confusing, but not bad. On the historic map that looks at MOE with COSEWIC, there is one area north of the border (in Northern Territory) that is both considered Historic Range and Range. That means the value for that area gets calculated into both classes and is represented in the table.

The issue is the overall total for the area as a whole north of the boarder will be inflated by that amount.

The amount is 31,481,287.8861 acres (this is counted twice int he total for the MOE table.

On the COSEWIC / CPE table and map only the CPE in the lower 48 and the Range in the Northern Territory are represented so the total area north of the border is correct as that said area is blended in to the overall range.

I'm going to check the data to make sure COSEWIC didn't have any additional areas that were overlapping. If so I'll have to try and get those values too.

Hope this explains this a bit and sorry for any confusion.

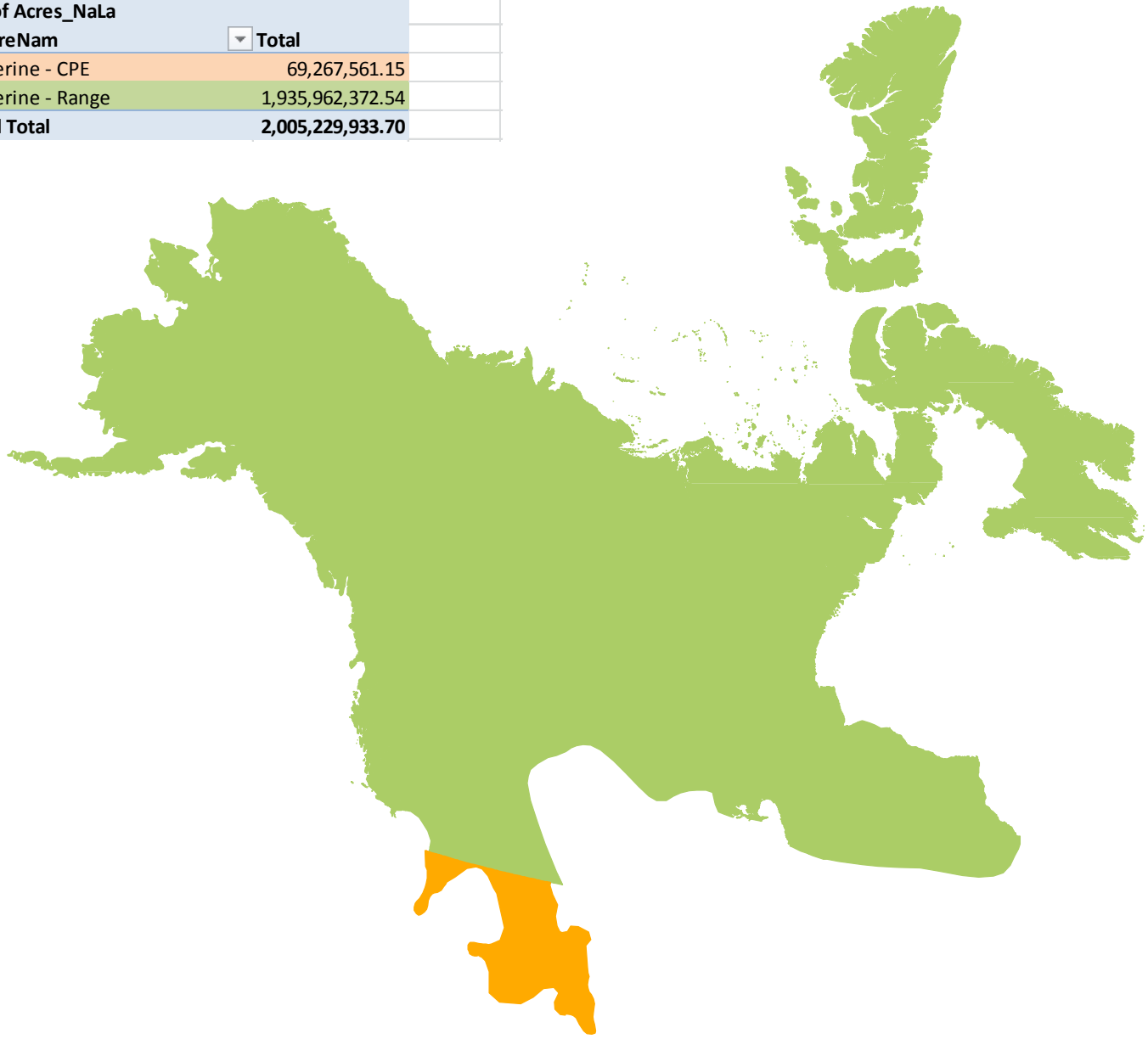
Ed

--
Edwin Turner, Geographer
GIS Mapping and Analysis Branch

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760.431.9440 x266

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Sum of Acres_NaLa FeatureNam	Total
Wolverine - CPE	69,267,561.15
Wolverine - Range	1,935,962,372.54
Grand Total	**2,005,229,933.70**

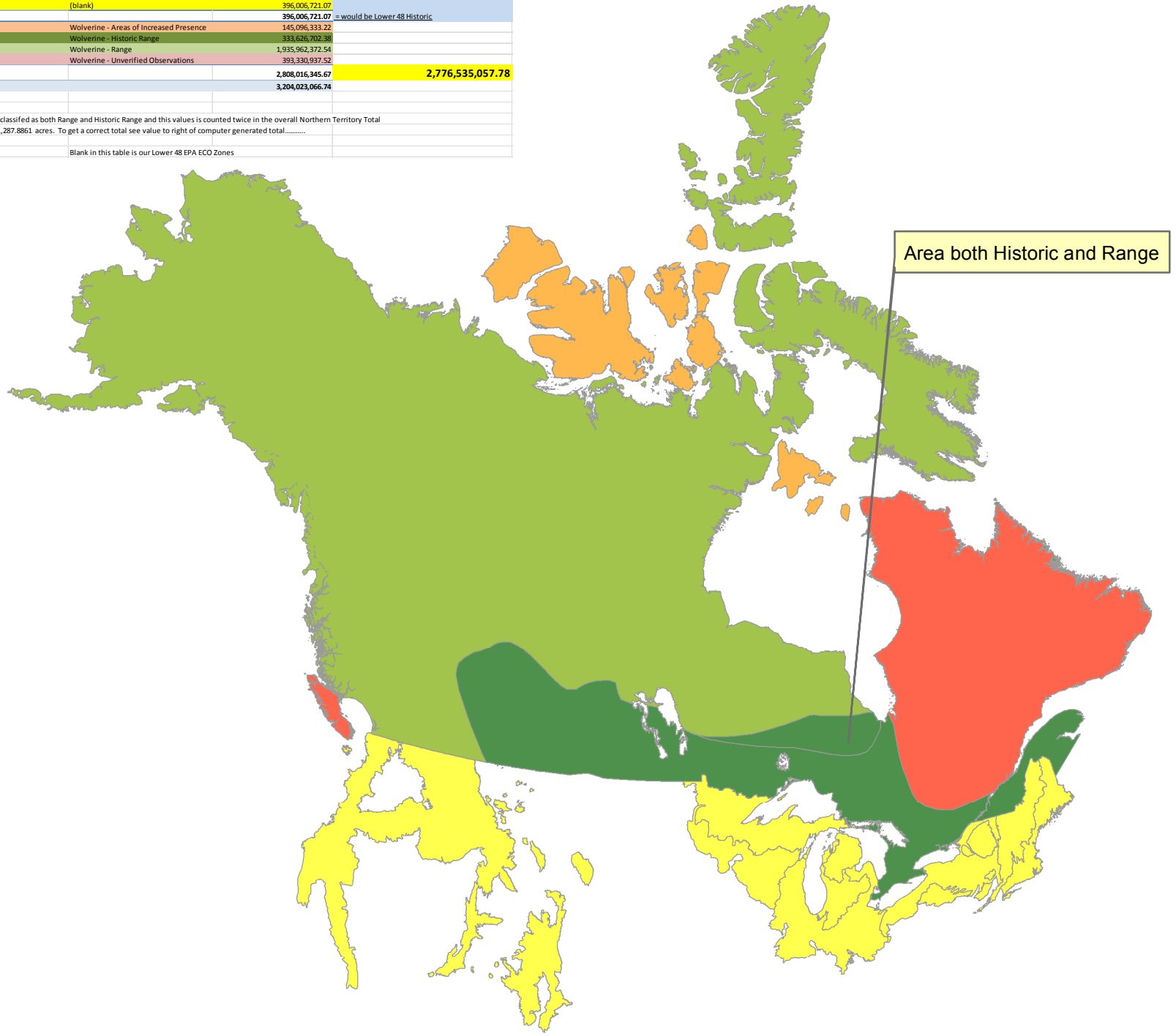


Sum of ACRES			
ZONE	FeatureNam	Total	
LOWER 48	(blank)	396,006,721.07	
LOWER 48 Total		396,006,721.07	= would be Lower 48 Historic
NORTHERN TERRITORY	Wolverine - Areas of Increased Presence	145,096,333.22	
	Wolverine - Historic Range	333,626,702.38	
	Wolverine - Range	1,935,962,372.54	
	Wolverine - Unverified Observations	393,330,937.52	
NORTHERN TERRITORY Total		2,808,016,345.67	2,776,535,057.78
Grand Total		3,204,023,066.74	

********* One specific area is classifed as both Range and Historic Range and this values is counted twice in the overall Northern Territory Total

****** The amount is 31,481,287.8861 acres. To get a correct total see value to right of computer generated total.....

Blank in this table is our Lower 48 EPA ECO Zones



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LOWER 48
LOWER 48 Total
NORTHERN TERRITORY
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Grand Total

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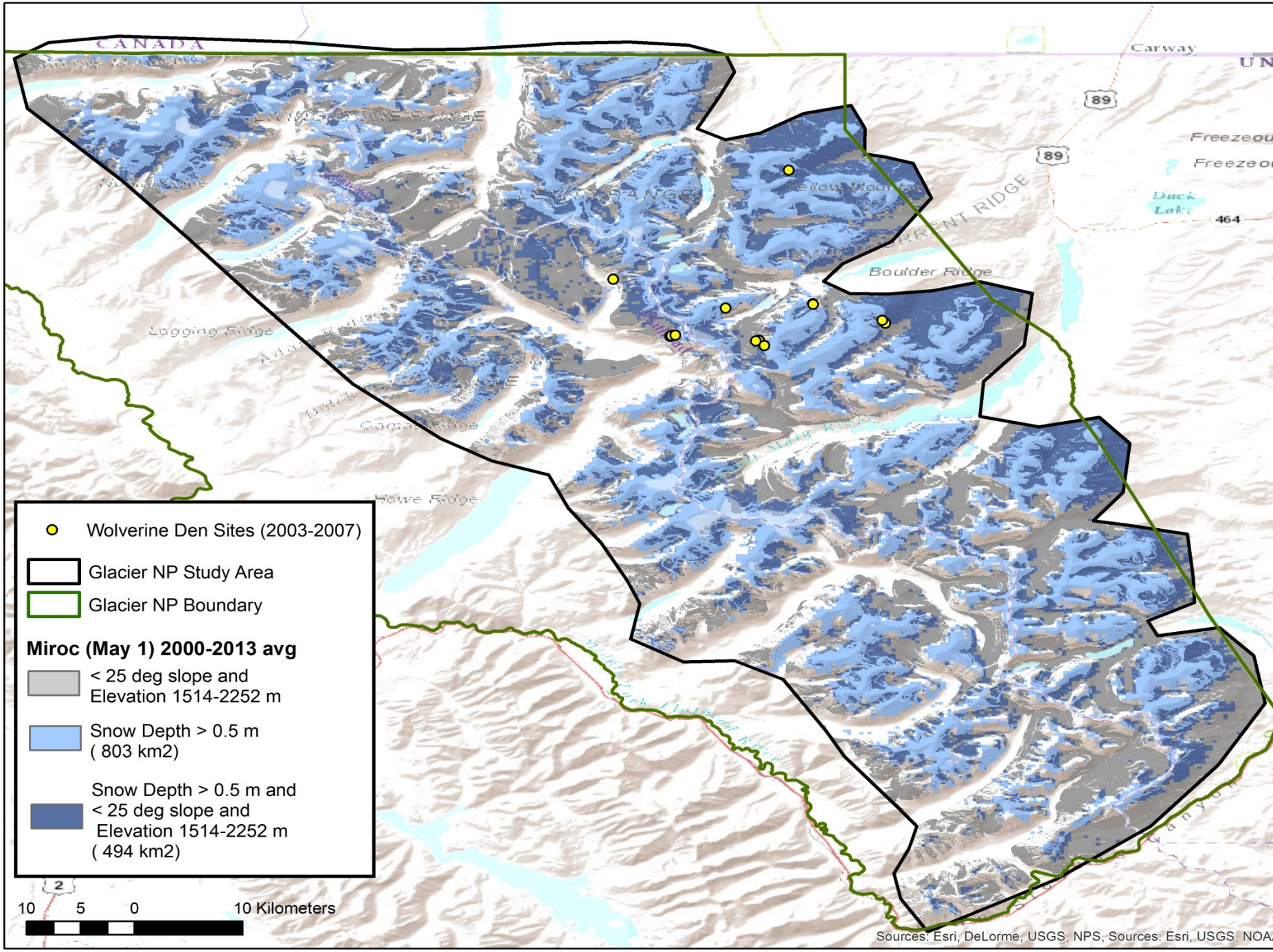
2,776,535,057.78

:al

2808016346
31481287.89
2776535058

From: Guinotte, John
To: Betty Grizzle
Subject: miroc glac slide
Date: Wednesday, February 14, 2018 3:29:59 PM
Attachments: miroc_glac_snow_slide.pptx

John Guinotte
Spatial Ecologist
Ecological Services
U.S. Fish and Wildlife Service
Mountain Prairie Region 6
134 Union Blvd., Lakewood, CO 80228
303-236-4264
john_guinotte@fws.gov


CANADA
Carway
Freezeout
Duck Lake
464
89
Boulder Ridge
Logging Ridge
Camas Ridge
Howe Ridge
Wolverine Den Sites (2003-2007)
Glacier NP Study Area
Glacier NP Boundary
Miroc (May 1) 2000-2013 avg
< 25 deg slope and Elevation 1514-2252 m
Snow Depth > 0.5 m (803 km2)
Snow Depth > 0.5 m and < 25 deg slope and Elevation 1514-2252 m (494 km2)
10 5 0 10 Kilometers
Sources: Esri, DeLorme, USGS, NPS, Sources: Esri, USGS, NOA

From: Grizzle, Betty
To: A J
Subject: Re: AK wolverine presentation
Date: Wednesday, February 14, 2018 4:02:15 PM

No, I will search for it though.

On Wed, Feb 14, 2018 at 1:22 PM, A J <222wsheridan@gmail.com> wrote:

> Betty
>
> Do you have this paper; I was a referee but did not get the final published paper.
>
> Wildlife Research 44(6-7):497-503. 2017
> https://doi.org/10.1071/WR17063
>
> **Temporal variation in the population characteristics of harvested wolverine (*Gulo gulo*) in northwestern Canada**
>
> Piia M. Kukka, Thomas S. Jung, Jean-François Robitaille and Fiona K. A. Schmiegelow

On Wed, Feb 14, 2018 at 11:11 AM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:

> No, I do not. There have been a few general news articles published by ADFG on their tracking efforts (e.g., AK Fish and Wildlife News, Dec 2014) but have not seen any formal publications.
>
> On Wed, Feb 14, 2018 at 10:52 AM, A J <222wsheridan@gmail.com> wrote:
>
> > Betty
> > Do you have any reports about wolverines written by Mike Harrington? There is nothing listed on the ADFG website for research reports.
> >
> > On Wed, Feb 14, 2018 at 9:34 AM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:
> >
> > > I will see if Mike will send me his presentation.
> > >
> > > On Wed, Feb 14, 2018 at 9:50 AM, A J <222wsheridan@gmail.com> wrote:
> > >
> > > > Thanks, Betty. I wish I could see his presentation but I'm in Fairbanks.
> > > >
> > > > Sent from my iPhone
> > > >
> > > > > On Feb 14, 2018, at 7:39 AM, Grizzle, Betty <betty_grizzle@fws.gov> wrote:
> > > > >
> > > > > Just an FYI - a colleague forwarded the attached announcement to me.

>
> --
>
> Betty J. Grizzle, D.Env., MPH
> Fish and Wildlife Biologist
> U.S. Fish and Wildlife Service
> Carlsbad Fish and Wildlife Office
> 2177 Salk Ave, Suite 250
> Carlsbad, CA 92008
> 760-431-9440, ext. 215
> 760-431-5901 fax
> <M. Harrington_ Wolverines.jpg>

--

Betty J. Grizzle, D.Env., MPH
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U.S. Fish and Wildlife Service
Carlsbad Fish and Wildlife Office
2177 Salk Ave, Suite 250
Carlsbad, CA 92008
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--

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From: Grizzle, Betty
To: Shoemaker, Justin; Marjorie Nelson; John Guinotte; Bush, Jodi
Subject: Revised PPT
Date: Wednesday, February 14, 2018 4:30:04 PM
Attachments: 20180214_Revised Briefing for Deputy Director_Wolverine 12mo finding.pptx

---

For your review: see attached, which incorporates revisions we discussed today.

--

Betty J. Grizzle, D.Env., MPH
Fish and Wildlife Biologist
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North American Wolverine (*Gulo gulo luscus*): SSA Summary/Listing Determination



Photo credit: Mark Packila; used with permission.

Briefing to Deputy Director
Mountain Prairie Regional Office
February 20, 2018

Wolverine Listing Decision Recommendation

- January 24, 2018, RMT Meeting held at Mountain Prairie Regional Office
  - RD and ARDs from Regions 1, 6, and 8
- Recommendation = ***Not T or E, not a valid DPS***
  - Add bullets re rationale

Listing History

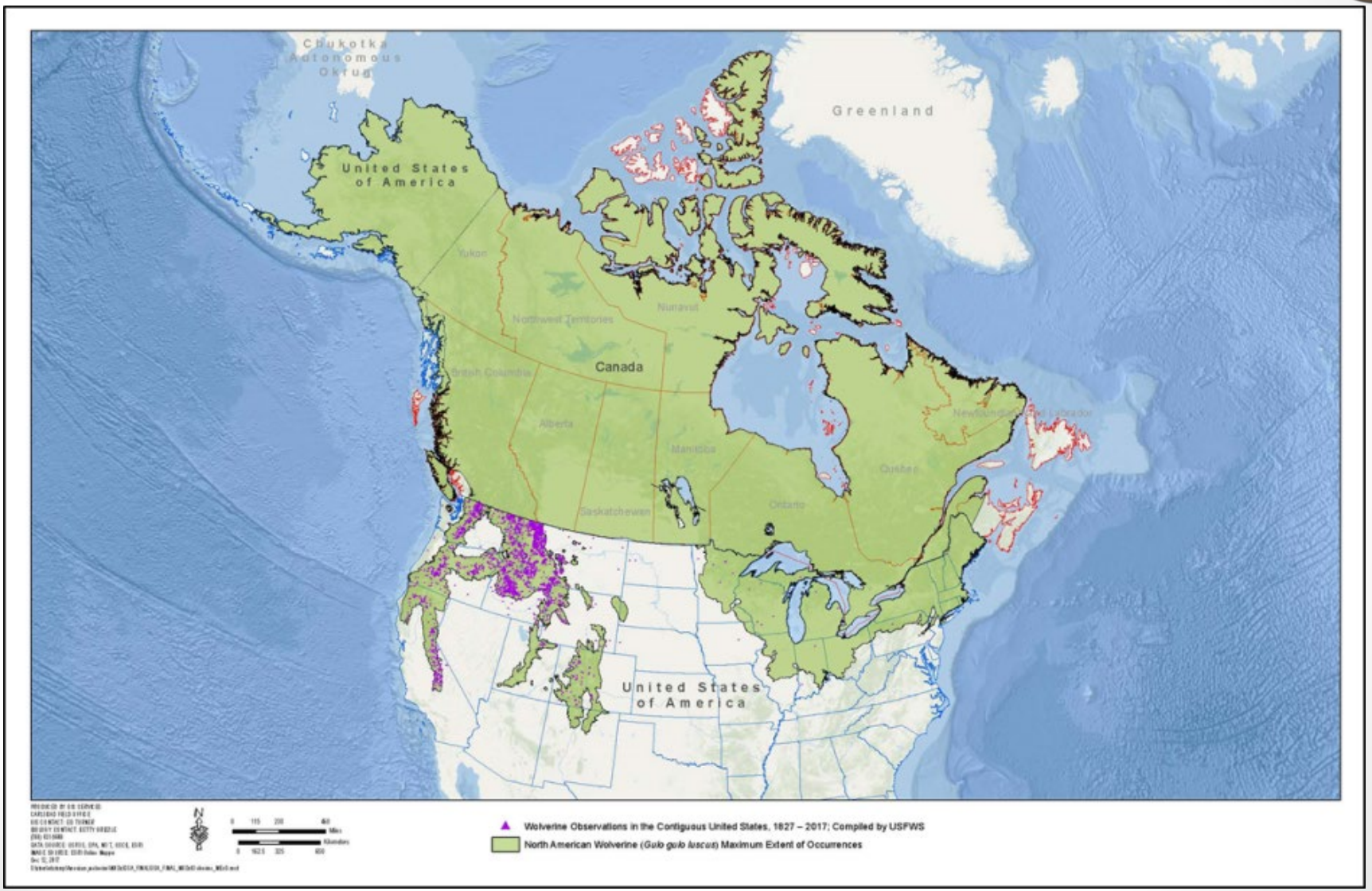
- **1995, 2003: 90- day Not Substantial Findings**
- **2008: Not Warranted 12-mo Finding (no DPS in lower U.S.)**
- **2010: 12-month Finding (DPS in lower 48)**
- **2013: Proposed Rule, with (2010) DPS**
  - Primary threat: synergistic effects of climate change
- **2014: Withdrawal**
  - Challenged; 2013 Proposed Rule reinstated
- **Current Action: Finalize Propose Rule or Withdrawal**
  - Service took a “fresh” look at science
  - Informed by SSA Report; publish FY 2018

SSA Report –New information and analysis

- Recent Scandinavian & Canadian studies providing new insights
  - Den sites located outside projected snow model (Copeland et al. 2010)
- New Canadian Wolverine Status Review
- Physiological discussion added; not well-documented in previous rules
- New range maps created
- Contacted many wolverine experts (both within and outside U.S.), State agencies, & other scientists
- New snow persistence/snow cover analysis

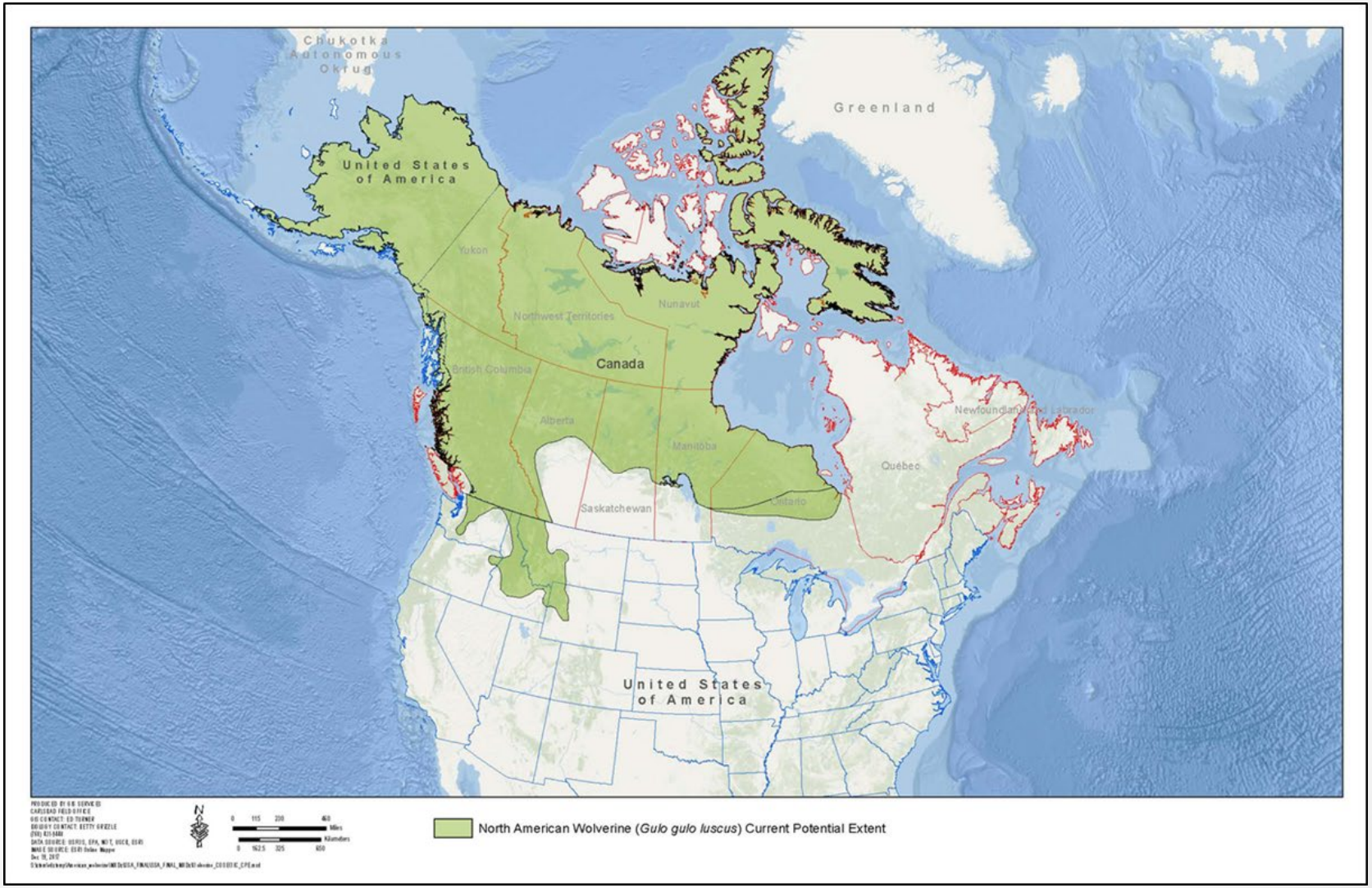
North American Wolverine Range Maps

- Maximum Extent of Occurrences (North America) (“historic range”)



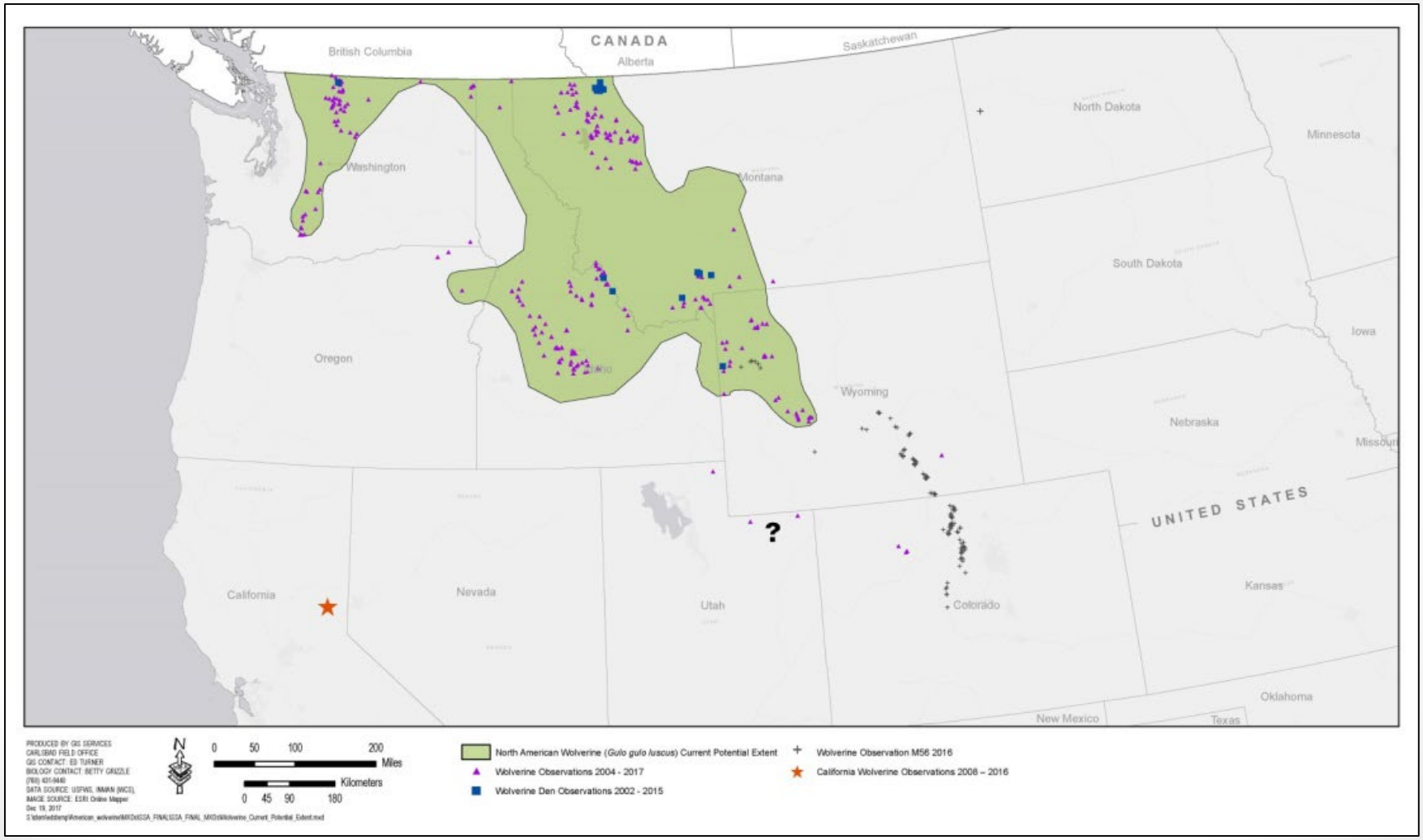
North American Wolverine Range Maps

- Current Extent of Occurrences (North America) (“current range”)



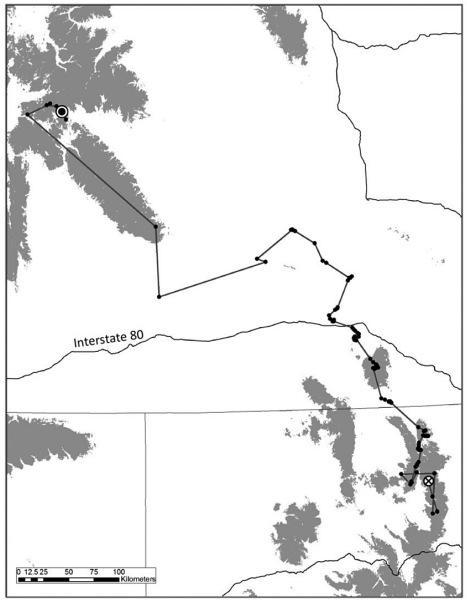
North American Wolverine Range Maps

- Current Potential Extent of Occurrences (Lower 48 only)



Species Description & Needs

- Key physical/ecological needs
  - Large territories, remote areas, high elevations
  - Access to variety of food, seasonal variability
  - Physical/Structural habitat features (e.g., talus slopes)
- Extensive movement and dispersal capabilities
  - Colorado, Utah, California, Michigan (all in 2000s)



Map of VHF locations of M56 - 2008 to 2009



Suspected wolverine M56, from Packila et al. 2017

Biological Status (current conditions)

- In North America: uncommon and at naturally low densities
  - Wolverines still recovering from decades of previously unregulated persecution
  - Wolverines now observed in areas previously extirpated (US & Canada)
- No population trend information for contiguous U.S.
- Contiguous US: 2013 modeled estimate = 318 individuals (range 249-626)
- Canada – over 10,000 adults
  - Stable or increasing in this northern part of No Am range
- AK population considered stable
- Movement between AK and Canada & lower 48 and Canada
- Multi-State Wolverine Occupancy Survey (2016, 2017)

Potential Stressors Evaluated

- Effects from roads
- Effects from wildland fire
- Disturbance – winter recreational activity, other infrastructure
- Disease and Predation
- Overutilization
  - Research activities
  - Harvesting and incidental trapping
- For Future Conditions
  - Effects of Climate Change
  - Cumulative Effects

Overall Assessment

- Known or likely potential stressors represent low risk to the species
- Legal trapping currently prohibited in contiguous U.S.
  - Trapping in Canada & Alaska appears to be sustainable
- Legal protections (T, E, Candidate) in several western states and Canada
  - Conservation plans (States and Tribes) in effect or planned
- Detailed look at snow obligate relationship and snow persistence related to climate change effects



Heinemeyer et al. 2017

Climate Change, Denning, and Snow

- New studies – Previous distribution model doesn't always predict wolverine occurrences or den sites
  - At least 80 natal dens in Scandinavia & several in Alberta, Canada, recorded outside model
- Limited information on how denning habitat affects **reproductive success**
- Denning habitat varies across range
- Re-evaluated assumptions on timing of natal den use
- **Many factors** influence den locations and shifts in den locations



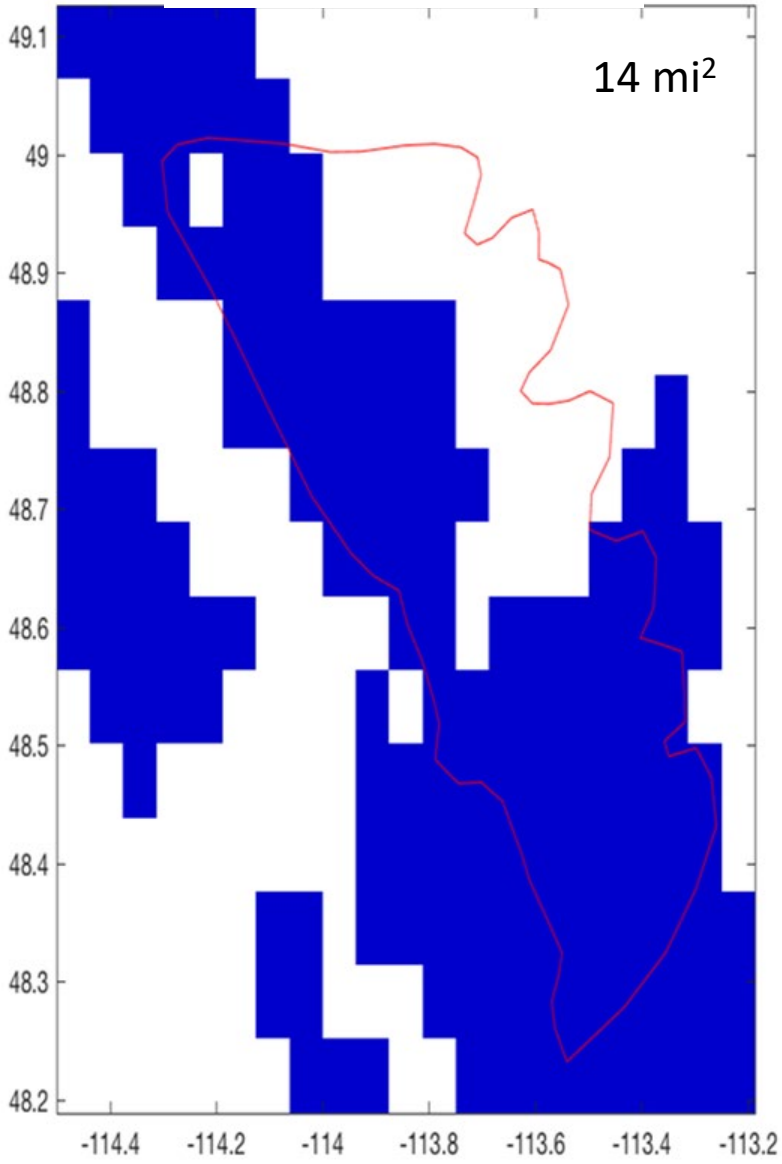
Denning "habitat" in Glacier NP; from Copeland and Yates 2008



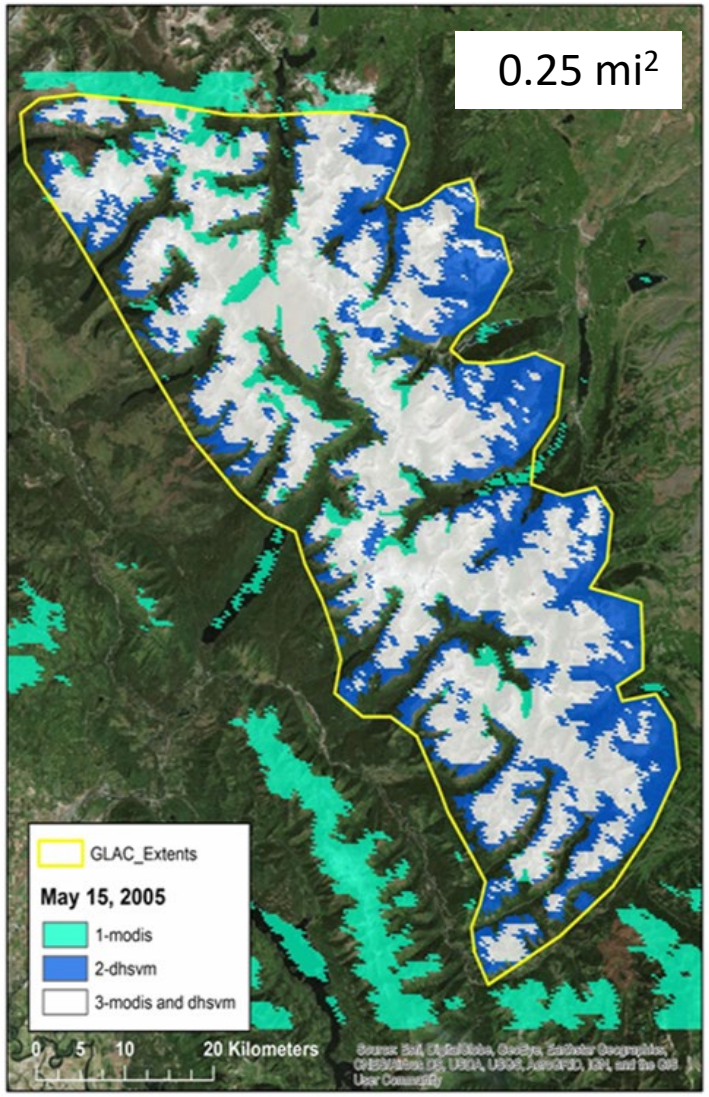
March 2, 2017 den in Sweden (lower left, under boulder)

Improved Snow Projections (2017)

McKelvey et al 2011

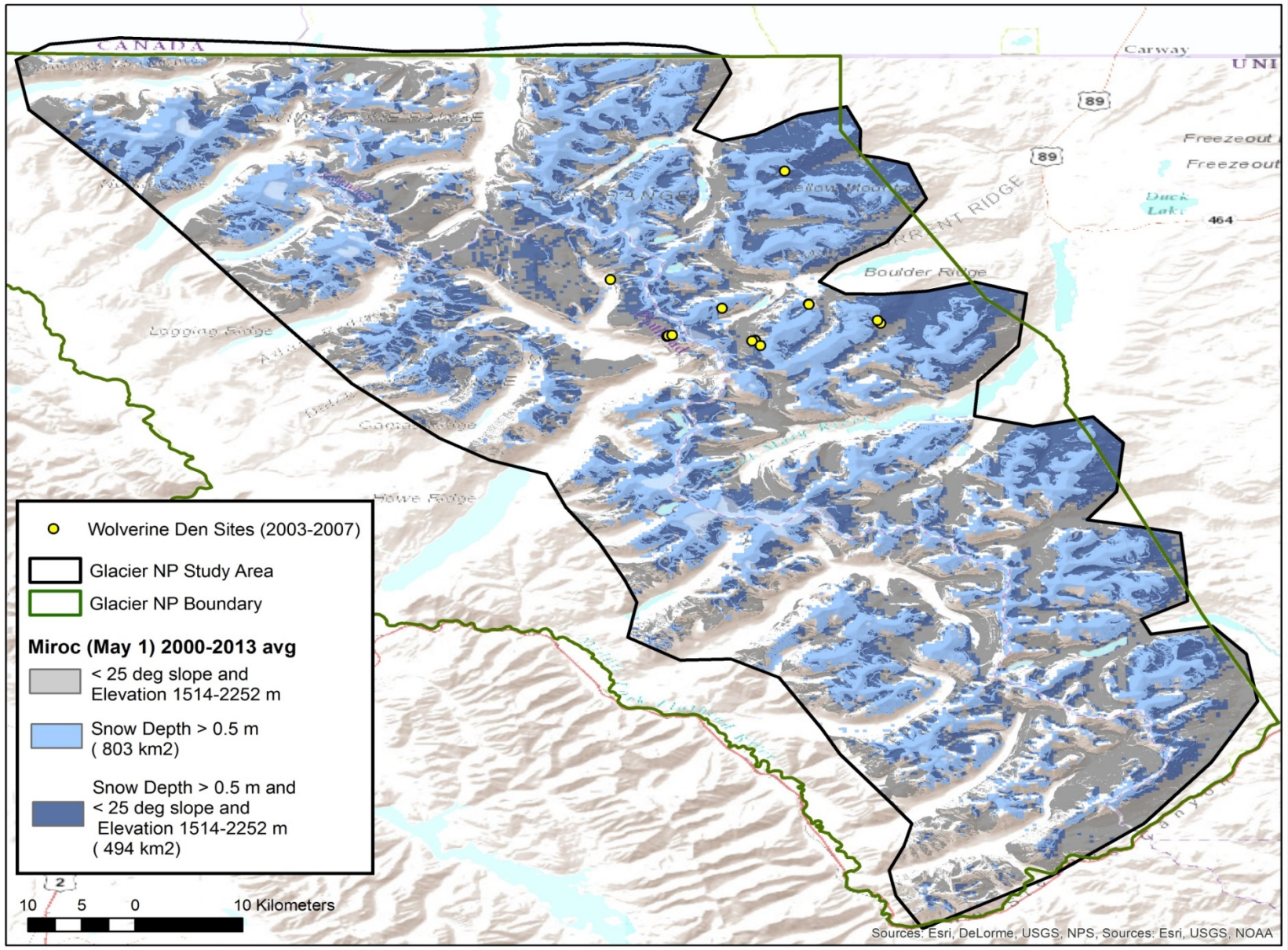


NOAA / CU 2017



(John Guinotte, FWS)

Glacier NP Future Snow Projections (~2055)



Comments Received

- Response to Oct 2016 NOI (proposed rule/status review)
  - Over 200 comments posted (from several States, Canada, U.K.)
  - Most, but not all, general public comments favored listing
- SSA Report Review
  - 11 States and Agency (Technical) Comments submitted
  - No Tribal comments submitted
  - Peer Reviewers – generally positive
  - Comments received reflected on-going debate on snow obligate characterization

DPS

- Previously – lower 48 DPS (both discrete and significant)
  - States and industry disputed validity
  - Based on “conservation status”
    - Effective population size, connectivity
- Evaluated new information
  - Discrete
    - Looked at trapping along border, not a significant barrier
  - Significant
    - Spatial evaluation of range to determine lower 48 extent, using detections, not estimates of range

Next Steps

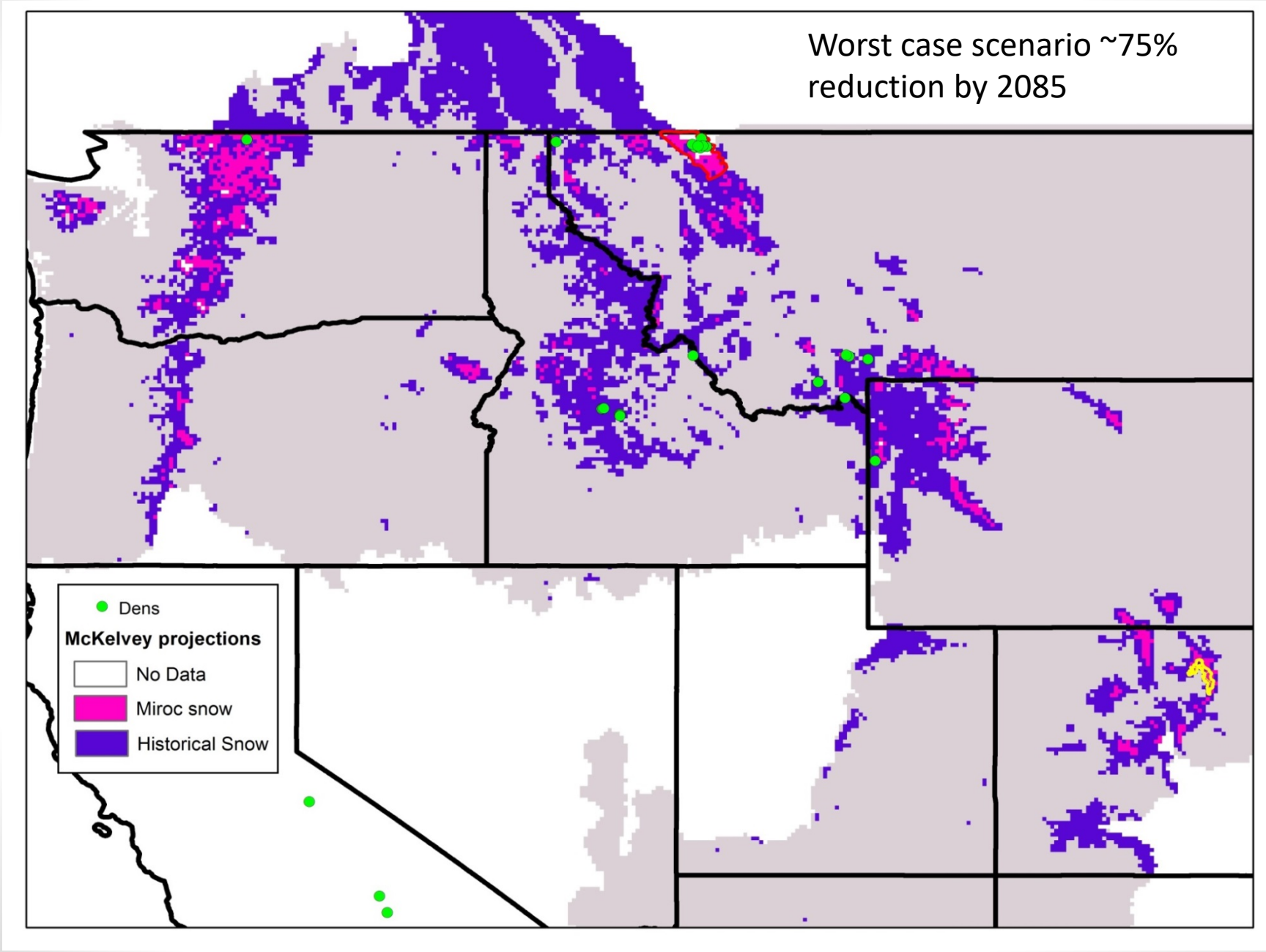


"Buddy" in California, May 2016;
Photo credit: Chris Stermer, CDFW

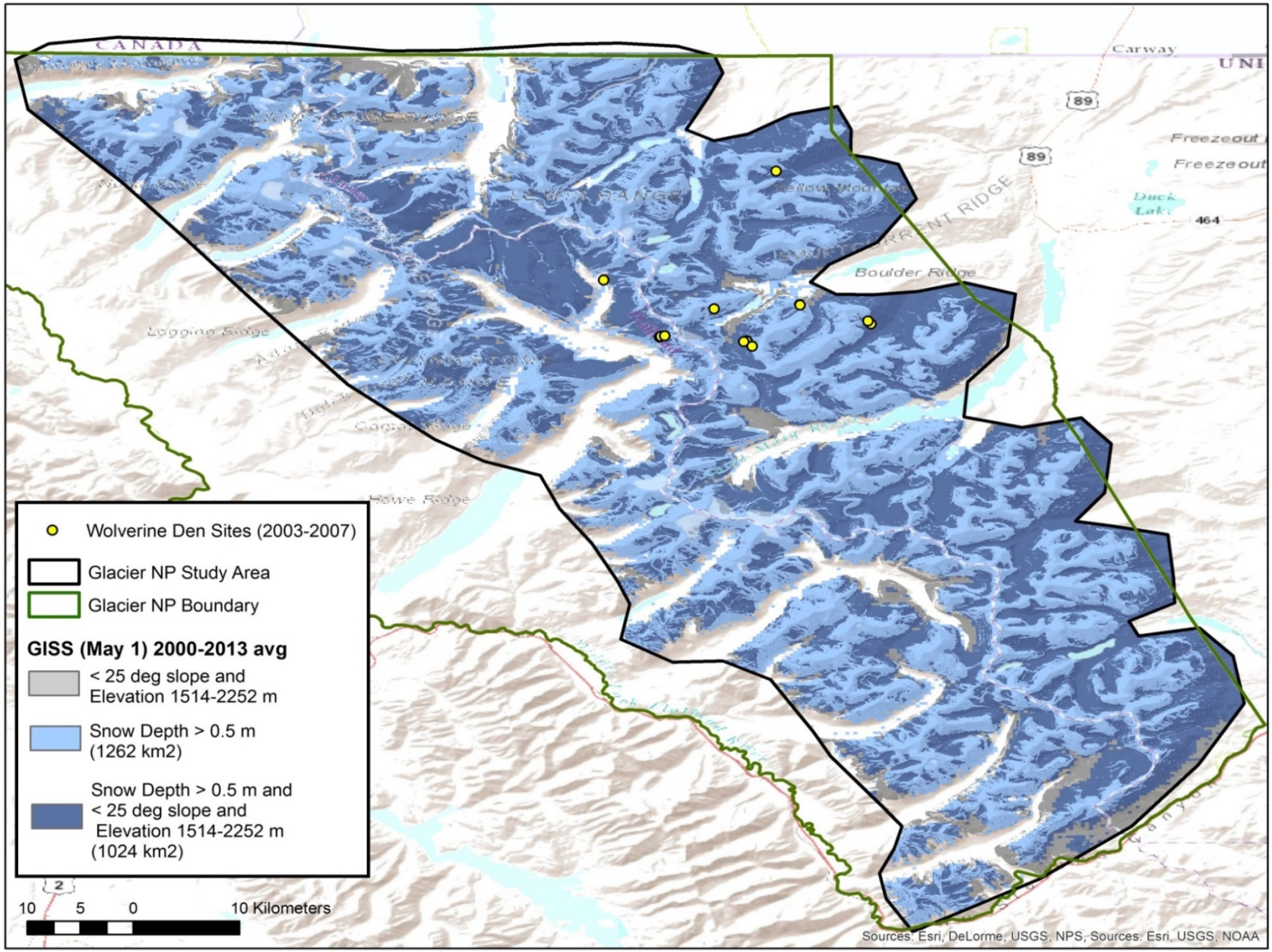
- Proposed approach
  - Stand alone, not warranted finding
  - No lower 48 DPS, but conducted full threats analysis
    - Using Sonoran bald eagle approach
- Timeline
  - FY 2018 workplan
  - Note: FR document preparation will need careful attention to detail given expected litigation

Extra Slides

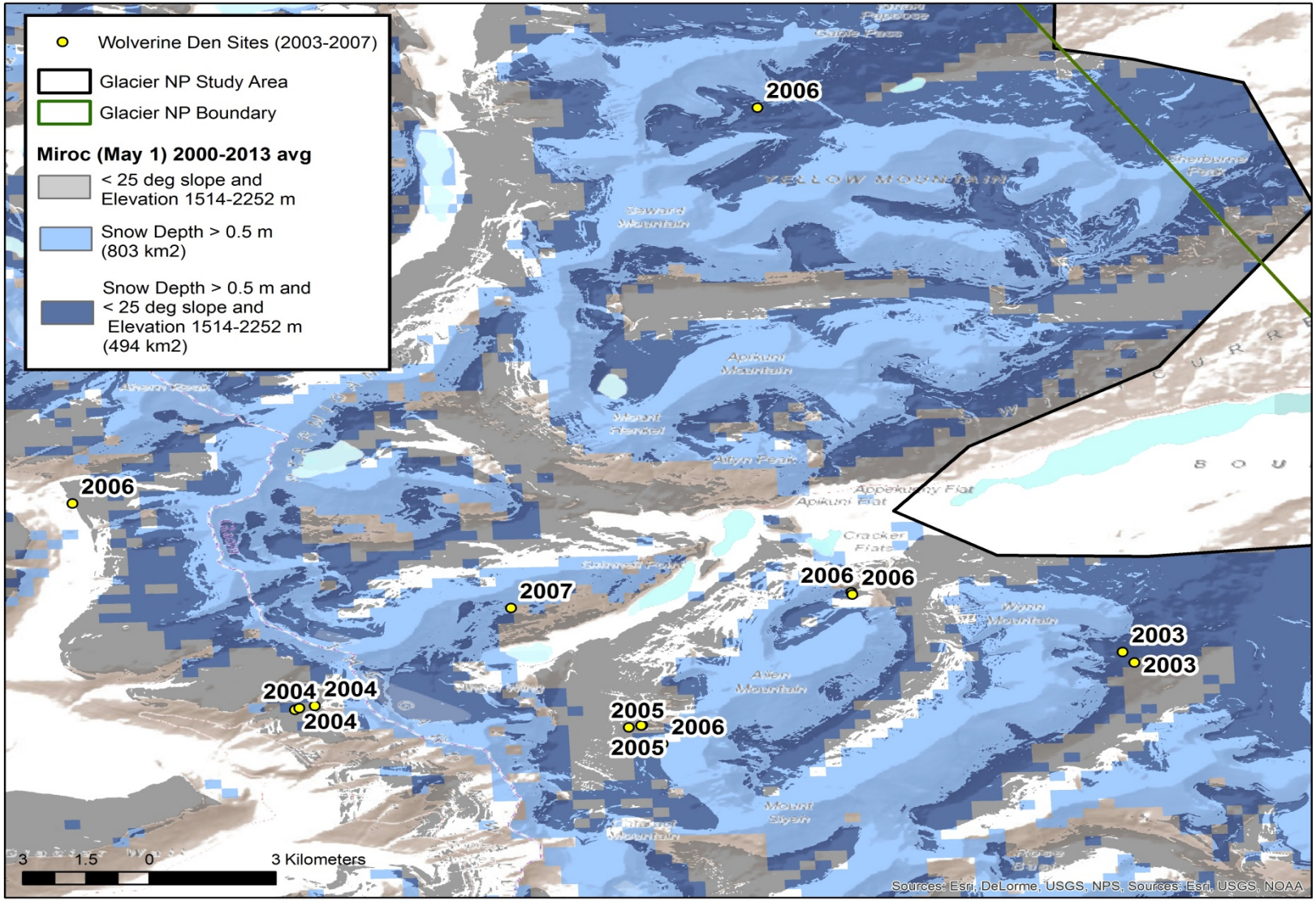
McKelvey et al (2011) snow projections



Glacier NP Future Snow Projections (~2055)



Glacier Dens and Future Snow Projections



From:	Turner, Ed
To:	Betty Grizzle
Cc:	Tony McKinney
Subject:	General COSEWIC assuming northern territory is Historic
Date:	Wednesday, February 14, 2018 5:22:01 PM
Attachments:	COSEWIC_General_Numbers_NT.xlsx General_COSEWIC_NT_Only_Acres.pdf

---

Betty, I took the areas from COSEWIC that makeup the Northern Territory and selected only the area considered "Range" by COSEWIC.

Took that and dissolved all internal polygon divisions of that area and retained a general acreage.

Next I took the COSEWIC Northern Territory area regardless of classification, considered it General Historic and dissolved all internal divisions, retained general acreage.

Please see map with table inset.

***** Note,** the general Northern Territory that is considered General Historic came out 7,089.60 acres less than the overall acreage I got when I built the MOE data combined with the COSEWIC. This could be the result of GIS processing which combines polygon areas and can cause slight shifting.

*** Difference is approx 0.000255% of the overall area in this genearl run to that of the MOE effort.

To be honest, I think working with these wide-span projects (North American Lambert) could just be causing some of this little issues. Like the acreage differences we see between projections when getting values in the Lower 48 vs the overall area including Alaska and Canada.

Hope this helps and we can run this by Tony if needed.

Ed

--
Edwin Turner, Geographer
GIS Mapping and Analysis Branch

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COSEWIC General Historic	**2,776,527,968.180000**
COSEWIC General Range	**1,935,962,372.540000**
	840,565,595.640000

***** Note General Historic overall differs slightly from the MOE COSEWIC
This can be the result of combining data and shifting of polylines.
Data here is simplified by dissolving all areas to simply represent Northern Te
When I combine the COSEWIC with our designed lower 48 area to create MC
Prior MOE Northern Territory total = 2776535057.7836

data combo.

erritory of General Historic and General Range
)E I get a COSEWIC Northern Territory 7089.60 acres larger than this general run.

COSEWIC General Historic	2,776,527,968.180000
COSEWIC General Range	1,935,962,372.540000
	840,565,595.640000

***** Note General Historic overall differs slightly from the MOE COSEWIC data combo.
This can be the result of combining data and shifting of polylines.
Data here is simplified by dissolving all areas to simply represent Northern Territory of General Historic and General Range
When I combine the COSEWIC with our designed lower 48 area to create MOE I get a COSEWIC Northern Territory 7089.60 acres larger than this general run.
Prior MOE Northern Territory total = 2776535057.7836

