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Dynamics of the North American Ice Sheet Complex during its inception and build-up to the Last Glacial Maximum

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ABSTRACT

The North American Ice Sheet Complex played a major role in global sea level fluctuations during the Late Quaternary but our knowledge of its dynamics is based mostly on its demise from the Last Glacial Maximum (LGM), a period characterised by non-linear behaviour in the form of punctuated ice margin recession, episodic ice streaming and major shifts in the location of ice divides. In comparison, knowledge of the pre-LGM ice complex is poorly constrained, largely because of the fragmentary nature of the evidence relating to ice sheet build-up. In this paper, we explore the inception and growth of ice (120–20 ka) using a glacial systems model which has been calibrated against a large and diverse set of data relating to the deglacial interval. We make use of calibration data prior to the LGM but its scarcity introduces greater uncertainty, which is partly alleviated by our large ensemble analysis. Results suggest that, following the last interglaciation (Oxygen Isotope Stage: OIS 5e), the ice complex initiated over the north-eastern Canadian Arctic and in the Cordillera within a few thousand years. It then underwent rapid growth to an OIS 5 maximum at ~110 ka (5d) and covered ~70% of the area occupied by the LGM ice cover (although only 30% by volume). An OIS 5 minimum is modelled at ~80 ka (5a), before a second phase of rapid growth at the start of OIS 4, which culminated in a large ice complex at ~65 ka (almost as large as at the LGM). Subsequent deglaciation was rapid (maximum modelled sea level contribution of >16 cm per century) and resulted in an OIS 3 minimum between ca 55–60 ka. Thereafter, the ice complex grew towards its LGM configuration, interrupted by several phases of successively less significant mass loss. Our results support and extend previous inferences based on geological evidence and reinforce the notion of a highly dynamic pre-LGM ice complex (e.g. with episodes of ± 10 m of eustatic sea level equivalent in <5 ka). Consistent with previous modelling, the fraction of warm-based ice increases towards the LGM from <20 to >50%, but even the thin OIS 5 ice sheets exhibit fast flow features (several 1000 m a^{-1}) in major topographic troughs. Notwithstanding the severe limitations imposed by the use of the 'shallow-ice approximation', we note that most fast flow-features generated prior to the LGM correspond to the location of 'known' ice streams during deglaciation, i.e. in major topographic troughs and over soft sediments at the southern and western margins. Moreover, the modelled flux of these 'ice streams' (*sensu lato*), appears to be non-linearly scaled to ice sheet volume, i.e. there is no evidence that decay phases were associated with significantly increased ice stream activity. This hypothesis requires testing using a model with higher-order physics and future modelling would also benefit from additional pre-LGM constraints (e.g. dated ice free/margin positions) to help reduce and quantify uncertainties.

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1. Introduction

The North American Ice Sheet Complex (NAISC) includes the Laurentide, Cordilleran and Innuitian Ice Sheets and other smaller ice caps (e.g. in Newfoundland and the Appalachians) and was the largest ice mass to grow and (almost) completely disappear during the last glacial cycle, thereby playing a major role in Late

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Pleistocene sea level change (Fulton and Prest, 1987; Shackleton, 1987; Lambeck and Chappell, 2001; Cutler et al., 2003). Our knowledge of its extent and dynamics, however, is based largely on reconstructions of its deglaciation since the Last Glacial Maximum (LGM), for which we now have a moderately well-constrained ice margin chronology in most, but not all, sectors (e.g. Dyke and Prest, 1987; Dyke et al., 2003; Dyke, 2004; Clark et al., 2009). Moreover, reconstructions of its behaviour during deglaciation attest to its dynamism and reveal major shifts in the location of ice divides (e.g. Dyke and Prest, 1987; Clark et al., 2000; Tarasov et al., 2012); episodes of ice streaming (e.g. Patterson, 1997, 1998; Winsborrow et al., 2004; Stokes et al., 2009) switches in ice sheet basal thermal regime (e.g. Dyke and Morris, 1988; Kleman and Hättestrand, 1999; Kleman and Glasser, 2007; Tarasov and Peltier, 2007) and abrupt episodes of mass loss that likely contributed to 'meltwater pulses' identified in the global sea level record (e.g. Hesse and Khodabakhsh, 1998; Hemming, 2004; Hesse et al., 2004; Tarasov and Peltier, 2005; Tarasov et al., 2012).

In contrast, comparatively little is known about its dynamics prior to the LGM, largely because most geological evidence on land has been erased by ice flow during the late stages (e.g. during final deglaciation). Thus, whilst ocean sediment records are particularly useful for investigating pre-LGM iceberg fluxes and meltwater events (e.g. Hesse and Khodabakhsh, 1998; Kirby and Andrews, 1999; Andrews and MacLean, 2003; Farmer et al., 2003; Rasmussen et al., 2003; Hemming, 2004; Hesse et al., 2004; Stokes et al., 2005), we have much less data to develop and/or test both conceptual and numerical models of ice sheet dynamics. Summarising a special issue of papers widely seen as a benchmark of knowledge on the Laurentide Ice Sheet (LIS), Andrews noted that "we have a fair knowledge of the Lateglacial history of the ice sheet back to about 20 ka; an imperfect knowledge of events between 20 and 40 ka; and sketchy knowledge of the events that led to the development of the ice sheet during marine isotope stages 5 through 4" (Andrews, 1987: p. 316).

Notwithstanding these limitations, impressive attempts have been made to link 'pockets' of evidence in the geological record that survived modification during deglaciation (e.g. relict landform and striae evidence; and till stratigraphic data) in order to reconstruct the approximate ice sheet configuration at broad time intervals prior to the LGM (e.g. Fulton, 1984, 1989; Dredge and Thorleifson, 1987; St-Onge, 1987; Vincent and Prest, 1987; Boulton and Clark, 1990a,b; Clark and Lea, 1992, synthesised in; Clark et al., 1993; Kleman et al., 2002, 2010). However, the fragmentary nature of the evidence precludes a robust chronology of ice sheet volume and dynamics and these reconstructions are, necessarily, tentative, and lack quantified uncertainty (e.g. especially in terms of ice volume). Thus, although we know that the ice complex underwent major fluctuations from the last major interglaciation (Oxygen Isotope Stage: OIS 5e) to the LGM (cf. Clark et al., 1993), the timing of these events is poorly constrained and the ice dynamics are largely unknown. This leaves several fundamental questions unanswered regarding its pre-LGM history. For example: where was its inception located and how did major ice domes/divides evolve through time? What was the magnitude and timing of major fluctuations in ice sheet volume; and how did rates of change compare to those during deglaciation? How did the basal thermal regime evolve? Where were the major ice streams located during the growth phase?

One way to address the above questions is through numerical ice sheet modelling, which can provide a continuous evolution of ice sheets through time. With this in mind, this paper presents the results of numerical modelling of the dynamics of the NAISC from its inception through to the LGM (120–20 ka). We do this using a previously-published Glacial Systems Model (GSM), which has

been calibrated against a large and diverse set of geophysical data (relating to relative sea level, marine limits, present day rates of uplift, palaeo-lake levels) for the deglacial interval (e.g. Tarasov and Peltier, 2004, 2007; Tarasov et al., 2012). Although few calibration data are available for intervals prior to the LGM (e.g. dated ice margin positions), we suggest that the dynamical self-consistency of the model implies that the robust deglacial calibration (see Tarasov et al., 2012) should increase the plausibility of the modelled pre-LGM chronology. The calibration provides a probabilistic output of ice sheet properties from effectively hundreds of thousands of model runs. This represents an advance over the small number of previous modelling studies that simulate the pre-LGM evolution of the ice complex but which have tended to lack substantial data-model calibration (e.g. Tarasov and Peltier, 1997a,b; Marshall et al., 2000; Bintanja et al., 2002) and/or deal with shorter time intervals (e.g. the LGM in Tarasov and Peltier, 1997a,b; 120–55 ka in Kleman et al., 2002).

Our focus here is on determining the spatial extent and ice volume through time and the associated changes in the dynamics of the ice sheet (e.g. in terms of the evolving basal thermal regime, velocity, etc.). This approach is not without its limitations (which we discuss below) but we believe it offers some significant insights into the pre-LGM history of the NAISC and generates some important hypotheses for future work to test.

2. Methods

The GSM is comprised of a 3D thermo-mechanically coupled ice-sheet model with visco-elastic bedrock response and a fully coupled diagnostic surface drainage solver; along with various smaller modules dealing with surface mass-balance, ice calving, ice-shelves, and basal drag (Tarasov and Peltier, 2004; Tarasov et al., 2012). The drainage solver is fully coupled in that evolving surface topography determines drainage, while pro-glacial lakes from the drainage solver affect the ice sheet via lacustrine calving, mini-ice shelves, and load effects on bed visco-elastic response. Drainage is recomputed every 100 years as a simple downslope 'depression fill' calculation as opposed to a dynamic solution of open channel hydraulic flow (Manning equation).

The model uses the 'shallow-ice approximation' (SIA), with a Weertman type power law (i.e. basal velocity is proportional to a power of driving stress) for basal sliding (exponent 3) and till-deformation (exponent 1). The thermodynamic solver for the ice is based on conservation of energy, only ignoring horizontal conduction due to the scales involved. The bed thermal model computes vertical heat conduction to a depth of 3 km and takes into account temperature offsets at exposed ground layers due to seasonal snow cover and varying thermal conductivity of thawed and frozen ground (see Tarasov and Peltier, 2007).

The GSM uses a 'basal flow factor' that is derived from a combination of the 'Sedmap' data-set (Laske and Masters, 1997), surficial geology map of Canada (Fulton, 1995) and seismic surveys from Hudson Bay (Josenhans and Zevenhuizen, 1990), see Fig. 1. Of most relevance is whether the value is equal to 0 or not, corresponding to whether fast flow due to basal sliding and/or subglacial sediment deformation is inhibited or permitted. An important caveat is that the model assumes that the sediment distribution is the same throughout the studied interval, which is convenient but potentially inaccurate (although to what extent for such a grid-scale has yet to be ascertained). It may be that the distribution of sediments was quite different at the start of the last glacial cycle, but there are some regions (e.g. Hudson Bay Lowlands) where the stratigraphic record indicates that fine-grained marine and lacustrine sediments were equally abundant after the previous interglaciation (e.g. Allard et al., 2012).

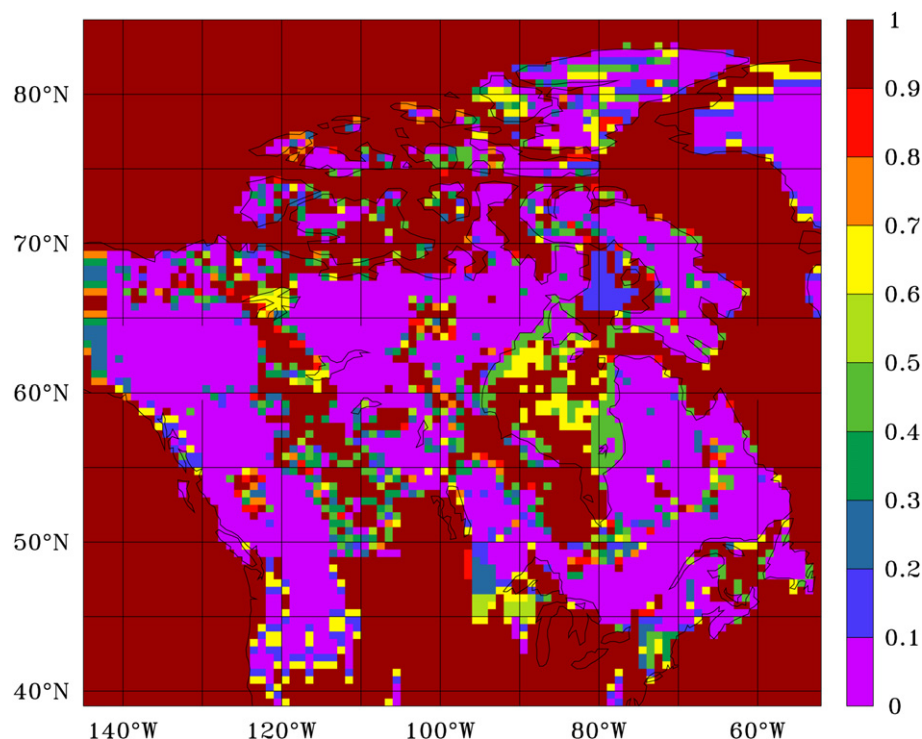


Fig. 1. Map of the 'basal flow factor' used as input for the Glacial Systems Model. A value of zero represents a grid cell where fast flow due to subglacial till deformation is not permitted. The strength of fast flow due to subglacial till deformation and hard-bedded sliding are both subject to calibration.

The availability of soft sediments is likely to influence the location of zones of fast flow (ice streams) in the model output. Thus, the lack of detailed knowledge of pre-LGM sediment distribution introduces greater uncertainty in terms of their location during ice sheet build-up. However, it should also be noted that a primary control on fast flow are the basal thermodynamics, which are an accurate predictive component of the model (although subject to uncertainties in climate forcing); and that soft deformable sediments are not an essential pre-requisite for ice streaming (cf. Stokes and Clark, 2003; Winsborrow et al., 2010). Additionally, the parameters controlling the strength of fast flow due to sliding and/or sub-glacial till-deformation are subject to calibration, which helps further reduce uncertainties.

The GSM is run at a 0.5° latitude by 1° longitude grid resolution and 39 calibration parameters determine the climate forcing, basal velocity, ice calving, etc. Model runs are initiated with no ice at 122 ka (except over the northwest tip of a stub of Greenland that falls within our study region and permits glaciation of Ellesmere and Axel Heiberg Islands) and results are from a high-probability subset of the calibrated model runs. Calibration involves parameter vectors (each consisting of a set of values for the 39 model calibration parameters) being obtained from Markov Chain Monte Carlo sampling with respect to the emulated fits to observational constraints. Prior parameter ranges are given in Tarasov et al. (2012) but their specification in such a large ensemble calibration is largely irrelevant as long as they are wide enough.

As noted above, the calibration dataset includes little pre-LGM geological constraint. Although glacial geologic data exist (e.g. multiple till stratigraphies that record multiple ice sheet advances/occupations of specific regions), they are generally rather poorly dated and, moreover, there are very few (if any?) dated ice margin positions, which would be of most use for model calibration, as in the deglacial calibration procedure (Tarasov et al., 2012). The

calibrated set of runs from Tarasov et al. (2012) were therefore subject to subsequent sieving (Table 1) to ensure ice volumes were within plausible ranges (taking into account the range of pre-LGM values for the other major ice sheets from ongoing and past calibrations). This small set of constraints alone proved significant enough to require modification of the climate forcing temporal dependence. Thus, the only change from the model setup described in Tarasov et al. (2012) is that a series of linear transformations were used to deform the temporal history of the climate weighting index, see Fig. 2, for interpolation between present-day observed climate and LGM climate from General Circulation Climate model runs (from the PMIP II database: Braconnot et al., 2007). This is necessary to attain an approximate phase match with the inferred sea level chronology of Waelbroeck et al. (2002) and results in a ca 5 ka shift at the start of the model runs (122 ka), which gradually deforms to match the LGM climate (Fig. 2).

It should also be noted that the climate index is not simply a prescribed mass balance forcing *per se*. The index is a weighting function for the interpolation between present-day observed and LGM climate and there are a number of other controls and

Table 1

Pre-LGM ice volume metric components. Maximum ice volume over the given time interval was tested against the constraint. Both the un-penalized range and maximum ice volume accepted are shown in m eustatic equivalent of ice. Runs with ice volumes below the un-penalized range are rejected. Weaker versions of the 37–33 and 50–42 ka constraints (respectively at 30 and 49 ka) were part of the original calibration.

Time (ka)	Un-penalised range (m)	Maximum accepted
115–105	21–36	40
93–84	8–34	40
67–57	47–65	71
50–42	49–63	67
37–33	50–65	70

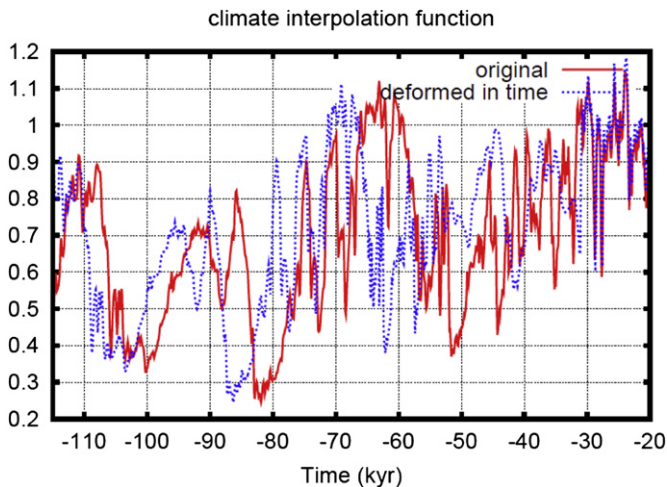


Fig. 2. Climate forcing interpolation index between present-day observed (value 0) and LGM climate (value 1). 'Original' is a glaciological inversion of the GRIP and GISPII records (updated from Tarasov and Peltier, 2003), now with the gicc05 chronology for GRIP (NGRIP Dating Group, 2006). This time series was used for the calibration. The 'deformed' time series is the temporally adjusted version to attain an approximate match to the inferred sea level record from Waelbroeck et al. (2002) and is an advance on previous work (e.g. Tarasov and Peltier, 1997a,b).

feedbacks that are subject to ensemble calibration (see Tarasov et al., 2012). Hence, the climate forcing is calibrated against data with respect to ice sheet constraints and 28 of the 39 calibration parameters control climate forcing and feedbacks in the model. Orbital configurations will, of course, vary over the last glacial cycle and, in particular, insolation represents a critical forcing for ice sheet inception. These variations are, to some extent absorbed by the climate index (albeit incompletely).

Ensemble means and variances presented below are generated assuming a Gaussian distribution of model likelihood given model misfit to observations. The key advantage of the ensemble approach is that it computes a probability distribution of past ice sheet evolution with explicit treatment of uncertainties in terms of both observational constraints and calibration parameters. It should be noted, however, that uncertainties due to inherent limitations in the GSM and parameterised form of the climate forcing are, at best, only partially quantified and so indicated uncertainty estimates below are unavoidably incomplete. Drawing conclusions from only a single model run (or small set thereof) would be highly speculative and so we focus here on analysis of ensemble means of a subset of 'best-fit' model runs that represent the likelihood of ice sheet properties (e.g. ice thickness, basal velocity) in any given location through time.

The model includes the entire NAISC (cf. Tarasov et al., 2012) for the obvious reason that all components are known to have been coalescent for at least part of the last glacial cycle. However, we primarily focus on the LIS (although not exclusively) because: (a) it covered a much larger area than the other components and was, therefore, more significant in terms of global ice volume/sea level fluctuations; (b) significantly more work has been undertaken on the LIS compared to the other components (both pre- and post-LGM), such that most calibration data from the deglacial interval relate to the LIS; and (c) the high relief and complex topography that characterises some components (e.g. the CIS) introduces considerable complexity and additional uncertainty for numerical modelling (and confidence in the results is correspondingly weaker), further exacerbated by the relatively large model grid cells.

3. Results

3.1. Ice sheet inception following the last interglaciation

The timing of inception is uncertain and is heavily dependent on the climate forcing interpolation function but, in order to explore the geographical location of ice sheet growth during the first few thousand years following the last interglaciation, we extracted ensemble probability plots of the likelihood of ice being >1 m thick. Results are presented in Fig. 3, which shows substantial ice accumulation from around 118 ka, with inception mainly on mountains and high plateaux of the central and eastern Canadian Arctic Archipelago (CAA) and over Foxe Basin. Inception then spread rapidly southward into Hudson Bay, south-westward into northern Keewatin, and westward across the remainder of the CAA. By 112 ka, it is probable that the ice complex covered most of northern Canada with a southern margin tracing the coast of Hudson Bay and across Keewatin to Great Slave and Great Bear Lake. A nearly full-bodied CIS developed over the same time interval and may have coalesced with the LIS in the north-east.

3.2. Ice sheet extent and volume

Fig. 4 shows snapshots of the modelled weighted ensemble average of basal velocity and ice sheet geometry for each of the major Oxygen Isotope Stages (OIS) since the last interglaciation (i.e. 5d, 5c, 5b, 5a, 4, 3 and 2). Fig. 5 shows the corresponding ice sheet volumes expressed as metres of sea level equivalent and the rate of ice volume change.

Following inception, our modelling shows the NAISC gradually expanded to cover most of mainland Canada by OIS 5d (110 ka, Fig. 4). Note, however, that although ice covered around 80% of the area that was later occupied by the larger ice sheets during OIS 4 (65 ka) and OIS 2 (25 ka), the modelled ice thicknesses are much less (eustatic sea level equivalent ca 25 m compared to 80 m at the LGM: Fig. 5). Most of the OIS 5d ice sheet was cold-based and the major ice domes/divides of the LGM configuration (e.g. Dyke and Prest, 1987) are less well developed (i.e. thinner over Quebec-Labrador, Foxe Basin and Keewatin).

During OIS 5c (100 ka) the ice cover shrank towards the north-eastern CAA and is depicted as a thin ice sheet that likely resembled a coalescence of cold-based plateau-ice fields, with ice thickness typically <1000 m and with an ice dome over Foxe Basin/north-east Keewatin. The ice cover thickened slightly and expanded south-westward again during OIS 5b (90 ka), with the Foxe-Keewatin dome remaining the most prominent feature, but it failed to reach the thickness and extent of the OIS 5d configuration at 110 ka. The transition to OIS 5a (80 ka) saw the ice sheet shrink to its minimal extent since OIS 5e, with a volume of just a few metres of sea level equivalent (Fig. 5) locked up in thin and mainly cold-based ice in and around the early inception grounds.

OIS 4 (65 ka) saw a rapid expansion of ice cover from around 80 ka to 65 ka, containing an additional volume equal to 60 m of sea level at the end of that 15 ka period (up to 10 m/thousand years: Fig. 5). The ice sheet configuration by 65 ka was almost identical to its LGM counterpart (compare to 25 ka) in terms of both its extent and the location of major ice divides (over Labrador, Foxe Basin/north-east Keewatin: cf. Dyke and Prest, 1987; Clark et al., 2000), but the difference here, compared to the build-up towards the LGM, is that the growth phase was continuous and rapid. There then followed an equally rapid period of mass loss to the start of OIS 3 (60 ka), which is mostly reflected in deglaciation of the southern margin and a major reduction in ice thickness. This deglaciation at the OIS 4/3 transition saw a peak sea level contribution of up to

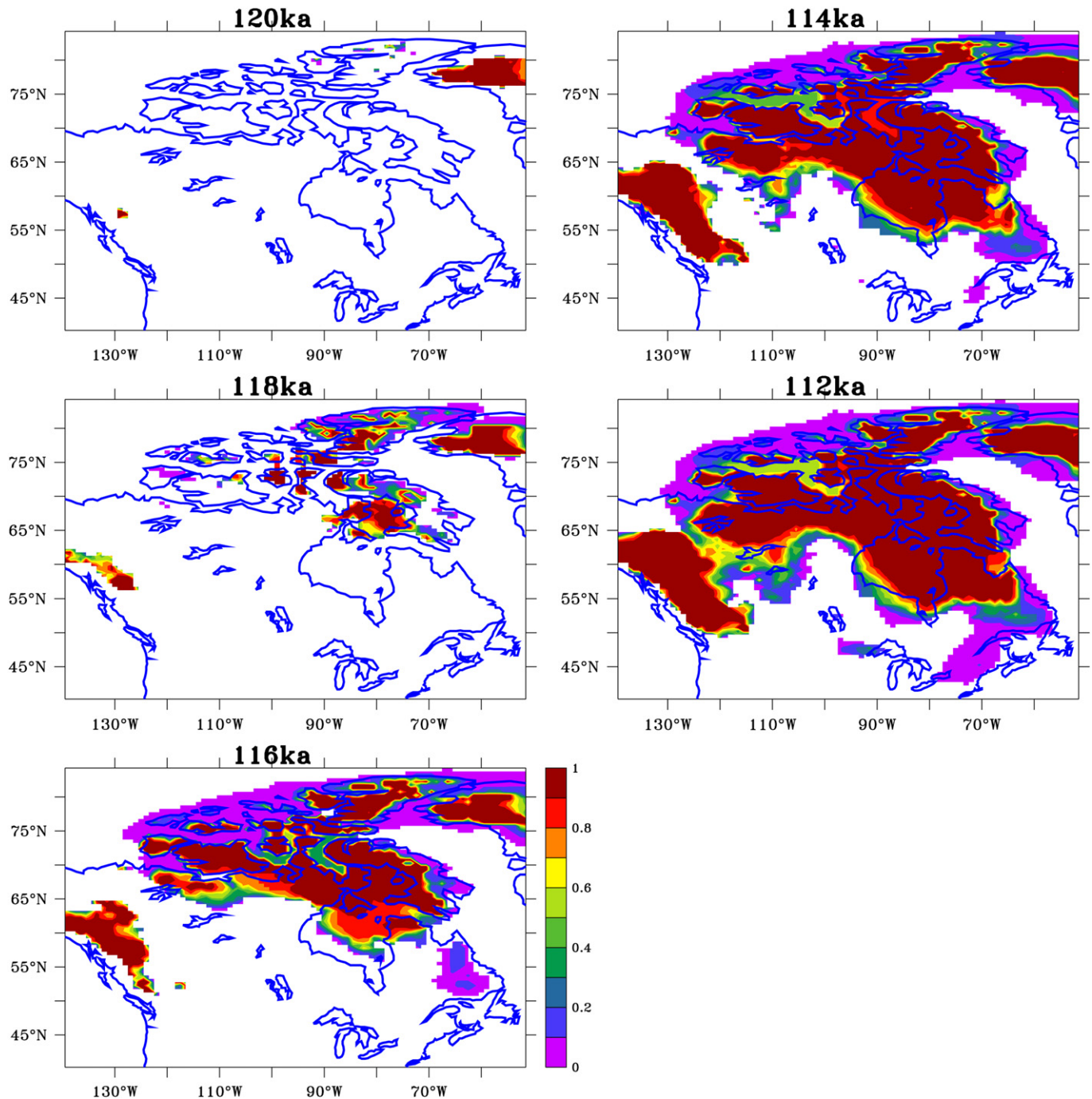


Fig. 3. Ensemble probability plots of ice thickness >1 m at 120 ka, 118 ka, 116 ka, 114 ka and 112 ka.

16 cm per century (Fig. 5). Several periods of ice growth and successively less significant ice shrinkage followed between 55 and 25 ka (Fig. 5), culminating in the LGM configuration of OIS 2 (~25 ka; cf. Clark et al., 2009).

3.3. Basal thermal regime and patterns in ice sheet velocity

The model calculates the evolving basal thermal regime, which strongly influences basal ice velocity and the drainage patterns. As in most numerical ice sheet models, ice velocities are dynamically determined based on the evolving driving stress and basal temperature of the ice, and on the presence of deformable

sediment. Rapid ice flow (e.g. ice streaming) can only occur when the basal temperature approaches the pressure melting point.

Fig. 6 shows a time series of the fraction of the bed occupied by warm-based ice from our new ensemble and from an 'old ensemble' (Tarasov and Peltier, 2007). A largely cold-based ice cover evidently evolves to a largely warm-based ice cover, as ice sheets expand and thicken, in agreement with previous work (Marshall and Clark, 2002). For the first 50–60 ka of glaciation, only 20–30% of the bed is warm, but this fraction increases to nearly 50% as the ice complex grows during OIS 4 (ca 65 ka). Following a reduction in the warm-based fraction around 60–55 ka, it then generally increases towards the LGM (25 ka), when it is >50%. The

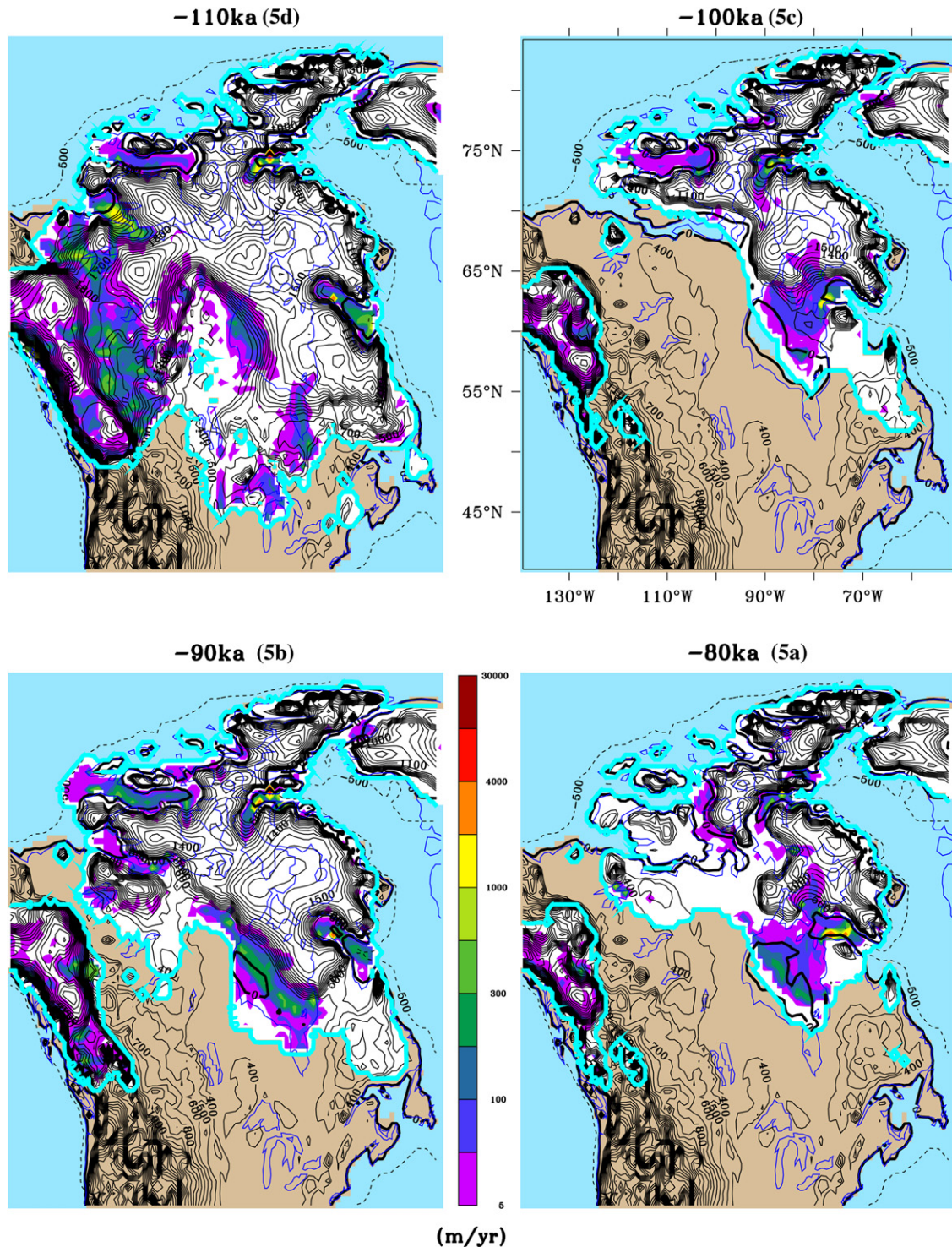


Fig. 4. Evolution of modelled weighted ensemble average of basal velocity for the North American Ice Sheet Complex during major OIS stages at 110 ka (5d), 100 ka (5c), 90 ka (5b), 80 ka (5a), 65 ka (4), 55 ka (early 3), 40 ka (mid 3) and 25 ka (2: ~LGM). For clarity, mean surface elevation contours are at 200 m intervals except for bathymetry and areas under 400 m. The thick light blue line is the ice margin. White areas are at or near to cold-based conditions.

results of the 'old' versus 'new' ensemble are broadly similar, although there is greater divergence for the period 55 to 30 ka. These differences in Fig. 6 are likely to reflect the deformation of the climate interpolation function (Fig. 1) and the additional constraints and parameters in the new calibration. The comparison is presented to underline the uncertainties in this and other modelling and the resultant contingent nature of the results.

The velocity patterns of the modelled ice complex (Fig. 4) closely resemble the patterns seen in modern-day ice sheets (e.g. Joughin et al., 1999; Bamber et al., 2000). Slow-flowing (cold-based) domes feed zones of intermediate velocity that are tributaries to discrete zones of rapidly-flowing ice ($>500 \text{ m a}^{-1}$). Even the relatively small ice sheets exhibit fast flow zones surrounded by cold-based ice. The relatively thin ice sheets between 100 and 80 ka, for example,

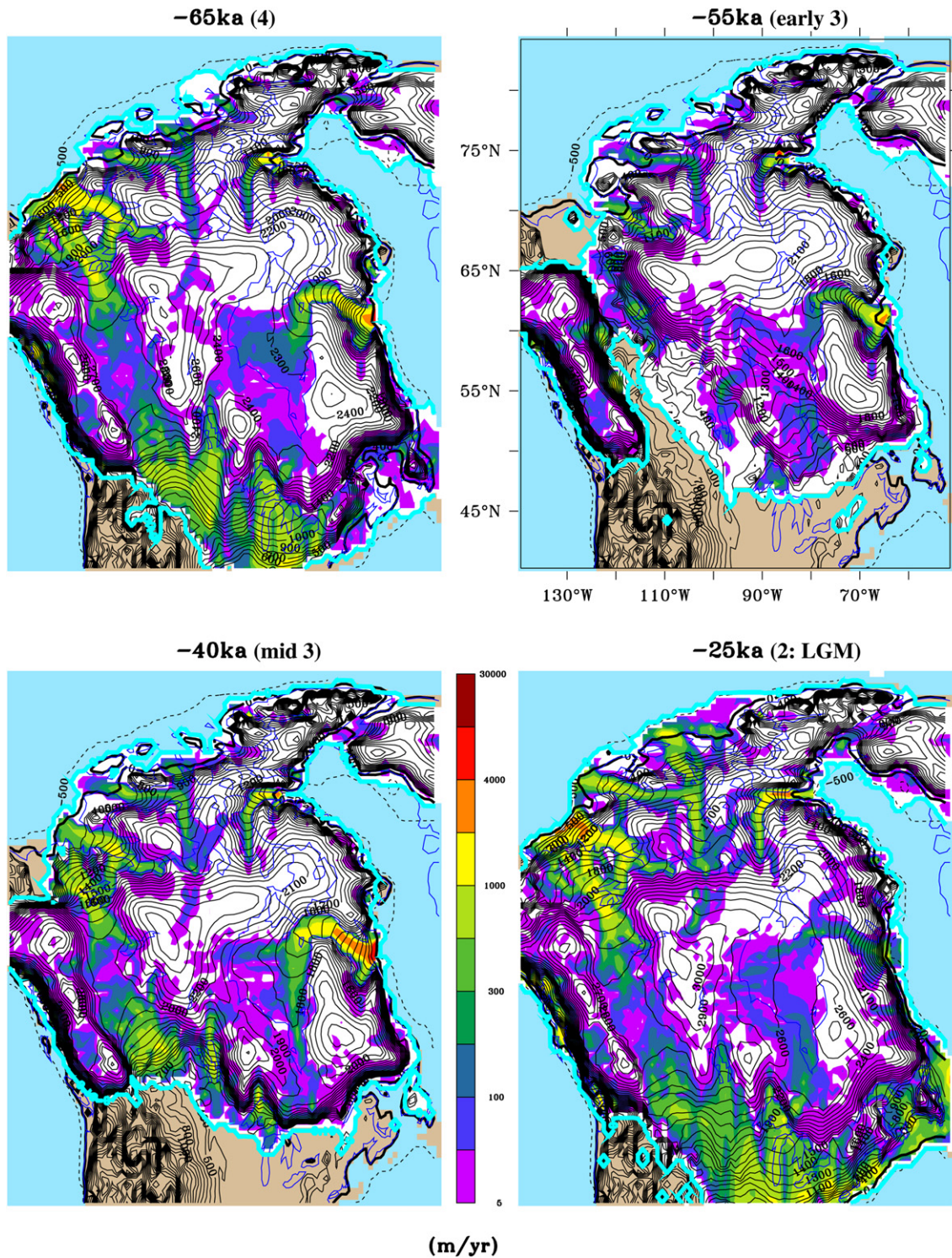


Fig. 4. (continued).

possess several arteries of rapidly-flowing ice, mostly restricted to marine margins that flow through topographic troughs, e.g. Hudson Strait, Lancaster Sound, etc (Fig. 4).

During more extensive glaciations (OIS 4 and 2), major zones of fast flow are located in the same marine troughs as in the smaller ice sheets but also occupy troughs further to the SE (e.g. Gulf of St Lawrence) and N/NW (e.g. Mackenzie River corridor). Broader zones of enhanced flow also formed inboard of the

southern terrestrial margin during periods of maximal ice coverage, with tributaries extending several hundred kilometres up-ice. These are particularly noticeable around the Great Lakes region and over the western prairies, some in association with proglacial lakes.

To show the temporal variability of these fast flow zones at a higher temporal resolution, Fig. 7 presents time series' of basal velocity at four locations where fast flow is generated by the model

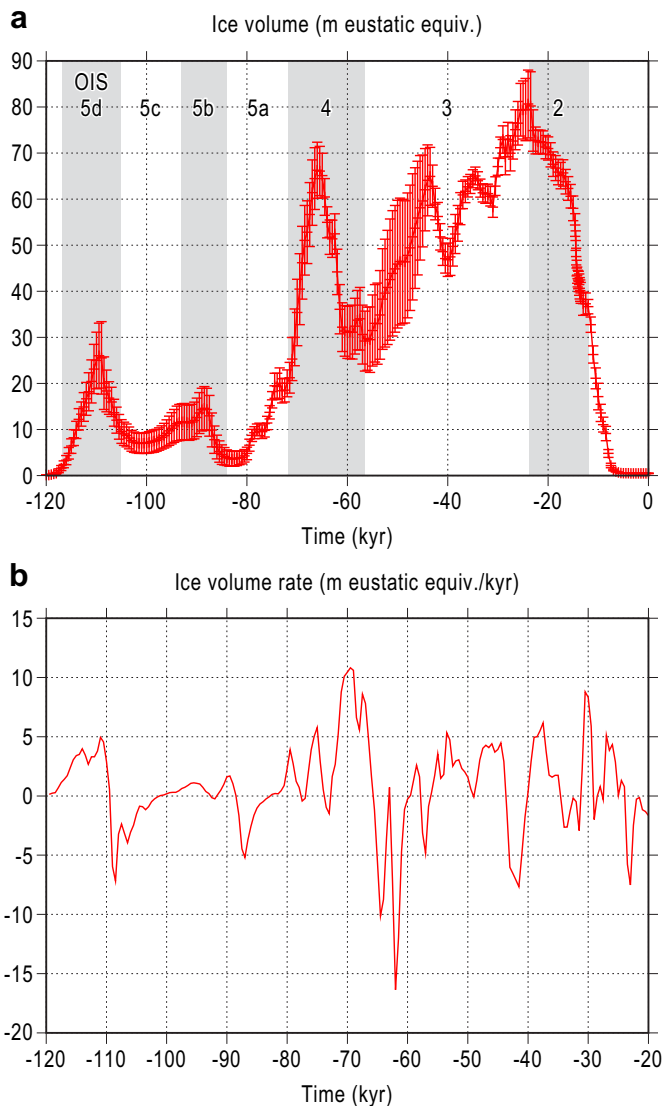


Fig. 5. (a) Modelled ice sheet volume of the North American Ice Sheet Complex from the last interglaciation to the LGM in eustatic equivalent metres of sea level (conversion factor of 25.19 m per 10^{15} m³ of ice) and, (b), corresponding ice sheet volume change expressed as rate of metres of sea level equivalent. All ages are calendar yr BP. Mean and two sigma deviation ranges from the calibrated ensemble are shown.

and where there is robust geological evidence for ice stream activity during final deglaciation (see inventories in Patterson, 1998; Stokes and Clark, 2001; Winsborrow et al., 2004; Stokes and Tarasov, 2010).

Hudson Strait (Fig. 7a) was a major conduit of fast flow during deglaciation and a dominant source of Heinrich events throughout the last glaciation (cf. MacAyeal, 1993; Andrews and MacLean, 2003; Hemming, 2004). The time series of basal velocity for this location denotes high velocities (>500 m a⁻¹) sustained from about 65 ka through to the LGM, but with a slowing down around 55 ka and, to a lesser degree 40 ka; a pattern which broadly matches ice sheet volume (Fig. 5a). Note that the near shut-down in Hudson Strait flow starting at 30 ka is an imposed feature of the model to dynamically facilitate the correct timing of Heinrich events 1 and 2. As detailed in Tarasov et al. (2012), the strength of the reduction in velocities (but not the timing) is under the control of calibrated parameters (i.e. the strength was not 'hand-tuned'). We stress that all other streaming is under free physical control and that the

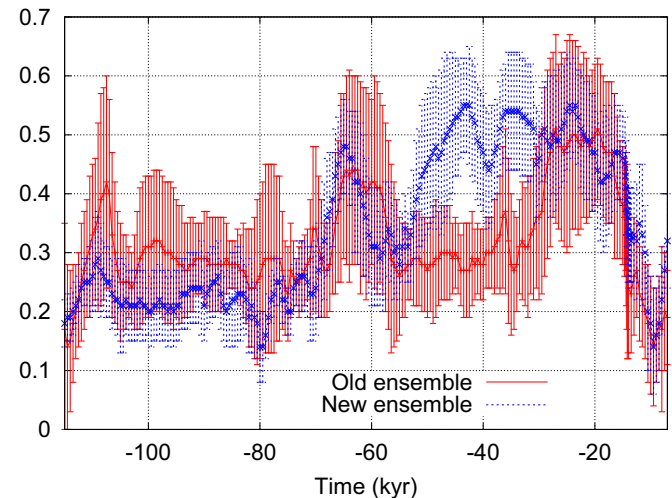


Fig. 6. Ensemble distribution of the fraction of warm-based ice over time. 'Old ensemble' is from Tarasov and Peltier (2007), which did not apply pre-LGM constraints on ice volume and used the 'original' climate interpolation function in Fig. 2, contrary to 'new ensemble' from current study.

dynamic facilitation in Hudson Strait (between 30 and 20 ka) is the only imposed feature in all model runs and is not important in the context of this paper, where much of the focus is on ice sheet dynamics prior to 30 ka.

Located at the north-western margin of the ice sheet, the M'Clure Strait Ice Stream was similar in size to Hudson Strait and has also been linked to major iceberg export events in the Arctic Ocean (Stokes et al., 2005, 2009). Model output indicates a similar onset and pattern of velocity to Hudson Strait at 65 ka but with comparatively lower velocities (Fig. 7b). This location exhibits less variability than Hudson Strait, as depicted by a narrower range of uncertainties and with a mean basal velocity typically falling within 500–1000 m a⁻¹ from 70 to 20 ka, whereas Hudson Strait often exceeds 2000 m a⁻¹. A similar pattern is seen in the Gulf of St Lawrence (Laurentian Channel Ice Stream: Shaw et al., 2006), but rapid ice velocities occur during maximal ice volumes from ca 30 ka and with some activity from 70 to 65 ka (Fig. 7c).

For comparison to the marine-based outlets, the time series of basal velocity for the land-terminating Des Moines lobe is also shown (Fig. 7d). This lobe is thought to have been the outlet of a large 'terrestrial ice stream' that may have operated several times since the LGM, with a prominent margin position between 30 and 20 ka, and a more recent re-advance around 14 ka attributed to ice streaming (Patterson, 1997). It is clear that rapid velocities are not a persistent feature but tend to only 'switch on' during glacial maxima (e.g. around 65–70 ka: Fig. 7d). We also note some possible cyclicity in this region from ca 50 ka, with the gradual build of streaming velocities often culminating in a sharp drop and then a further steady increase.

4. Discussion

4.1. Ice sheet inception and coalescence

The location of North American ice sheet inception has been debated since Flint's (1943) suggestion of a highland origin and windward growth for the LIS (see review in Ives et al., 1975; Stroeven et al., 2002). This hypothesis has gradually been replaced by a new paradigm that invokes 'instantaneous glacierisation' of Arctic and subarctic plateaux and uplands (e.g. Ives, 1957, 1962; Andrews and Mahaffy, 1976; Andrews et al., 1976; Andrews and

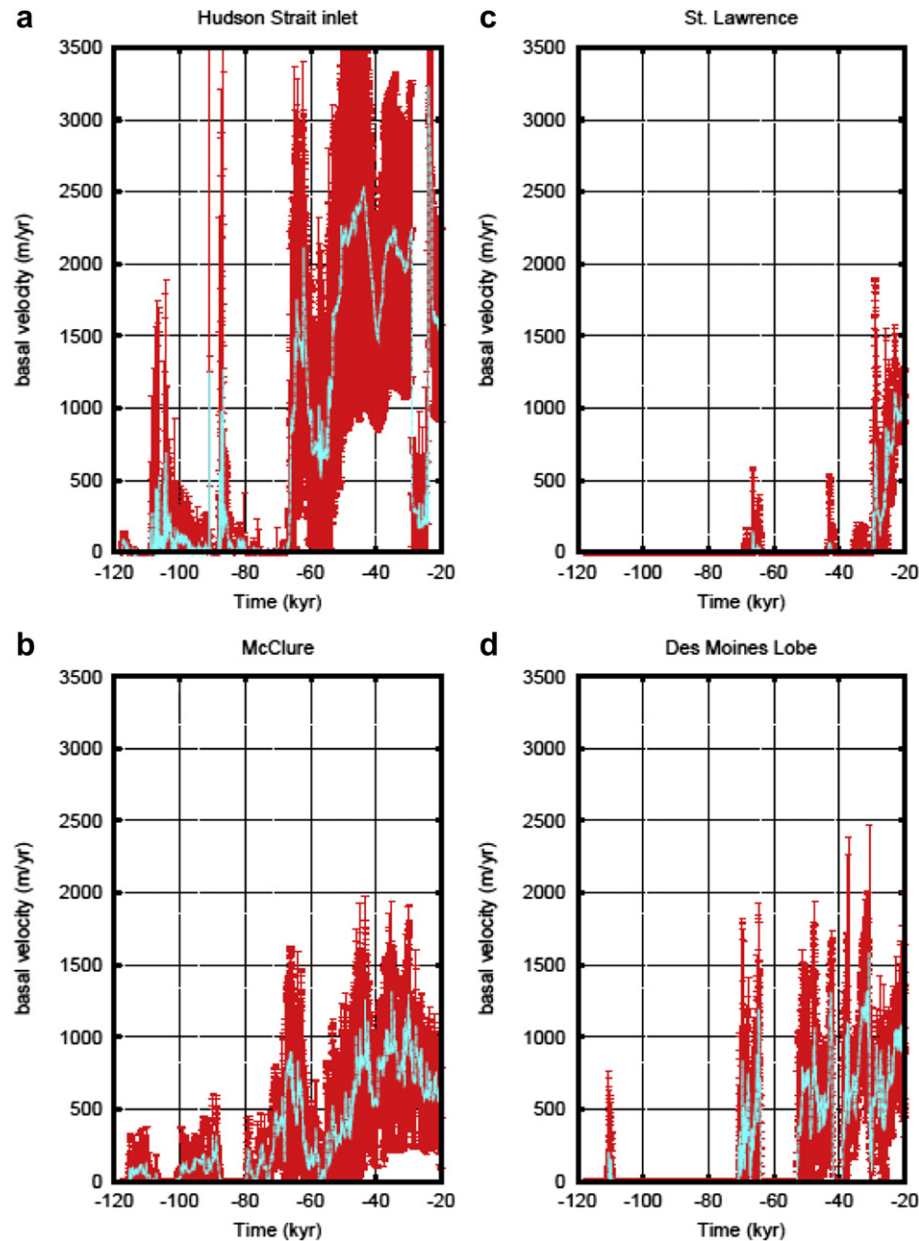


Fig. 7. Pre-LGM time series' of maximum basal velocity near the margin for four fast flow zones that correspond to the location of 'known' ice streams during deglaciation (cf. Patterson, 1998; Stokes and Clark, 2001; Winsborrow et al., 2004; see also Fig. 9): (a) Hudson Strait, (b), M'Clure Strait, (c) Gulf of St Lawrence, and (d), Des Moines Lobe. Weighted ensemble mean and one standard deviation uncertainties are shown. To minimize flux loss from ablation in the calculations, velocities were computed one grid cell in from the ice marginal cell (terrestrial) or at grid cells with grounded ice adjacent to floating ice. Fluxes were computed using basal ice velocities to isolate the contribution from streaming as opposed to ice deformation.

Barry, 1978; Dyke et al., 1989; Kleman et al., 2002; Wolken et al., 2005, 2008). This paradigm is based on the notion that these regions are close to the threshold of glaciation (cf. Williams, 1978a, 1979; Bromwich et al., 2002), as illustrated by the fact that thin plateau ice-fields spread over tens of thousands of square kilometres there during the Little Ice Age (e.g. Ives, 1962; Williams, 1978b). However, the precise timing of the last glacial epoch is not well known and albedo and temperature lapse-rate feedbacks must have acted to amplify relatively low-amplitude orbital cooling trends. In particular, global sea level lowering of at least 20 m and as much as ~75 m over 10 ka (cf. Andrews and Mahaffy, 1976) suggests a rapid growth that has, previously, not been easily captured by numerical models (cf. Bintanja et al., 2002; Marshall, 2002).

In terms of location, our results broadly agree with previous numerical modelling (e.g. Tarasov and Peltier, 1997a; Marshall and Clarke, 1999; Marshall et al., 2000; Bintanja et al., 2002; Kleman et al., 2002), that clearly support the location of inception on Arctic/subarctic plateaus along the eastern seaboard, with Ellesmere Island, Axel Heiberg, Devon and Baffin Islands seeding the proto-LIS/IIS within a few thousand years of the last interglaciation (Fig. 3). An embryonic dome over Quebec-Labrador formed a few thousand years later (cf. Andrews and Mahaffy, 1976; Vincent and Prest, 1987; Marshall et al., 2000). It had probably been established by 114 ka, but it is certainly apparent in the modelled ice sheet topography for 110 ka (Fig. 4) and supports inferences from striae and dispersal trains for an early ice build-up north of the St Lawrence River in the highlands of Quebec (cf. Veillette et al., 1999;

Veillette, 2004). A similarly early but short-lived build-up of the Appalachian ice complex is also shown during OIS 5 (at 110 ka Fig. 4), which has also been invoked based on till stratigraphic data (McDonald and Shilts, 1971; Lamothe et al., 1992; Clark et al., 1993). The early presence of ice in Hudson Bay/Strait is also in agreement with sedimentary sequences in the Hudson Bay Lowlands, where organic bearing units assumed to be of the last interglaciation are often capped (or completed) by a thick sequence of rhythmites, which suggest that the Hudson Bay watershed was blocked by ice, thereby creating a large proglacial lake prior to the next ice advance (e.g. Skinner, 1973; Allard et al., 2012).

Although there are issues that hinder the modelling of the CIS at this grid resolution (see Section 4.2.3), it is also noteworthy that the CIS is established and probably coalescent with the LIS by 112 ka. Ryder and Clague (1989) and Clark et al. (1993) comment that the timing of the first major ice sheet development in this region during the last glacial cycle (known as the ‘penultimate glaciation’) is uncertain (OIS 4 or 5?) but our modelling clearly supports a large CIS during OIS 5.

A key result of our modelling is the relatively large ice sheet at 110 ka, with the LIS coalescent with the CIS (Figs. 4 and 5). This agrees with inferences from one of the earliest numerical modelling studies, albeit using simplified physics (e.g. Andrews and Mahaffy, 1976), that suggested that large ice sheets can develop in about 10,000 years under certain conditions. Capturing the rapid growth of North American ice to a maximum at 110 ka, described as the “explosive ice sheet growth suggested by the marine record” (Marshall, 2002: p. 133), is a successful feature of our model and supports the assertion of Cutler et al. (2003) that ice sheet expansion need not be inherently slow and is often characterised by rapid intervals of growth. Our modelled rate of maximum sea level fall during growth of the OIS 5d ice sheet (between 120 and 110 ka) approaches 5 m/1000 yr (Fig. 5), which is almost identical to the figure (4.2 m/1000 yr) quoted by Andrews and Mahaffy (1976) in their modelling study. Furthermore, deglaciation of the OIS 5d ice sheet led to a peak sea level contribution of >5 m/1000 yr and we also note an equally rapid rise in sea level at the OIS 5b/5a transition, which has been identified in coral reef records (e.g. Cutler et al., 2003).

Despite their relatively large areal extent, these early ice sheets (e.g. 110 ka) were generally thin and largely cold-based. This may result from more limited precipitation, but is probably a reflection of the short time period over which they could grow (e.g. inception over Hudson Bay only starts between 116 and 114 ka) and the fact that it takes time to warm-up the base of the ice.

Following inception, our results depict ice episodically spreading from inception centres into Hudson Bay for the period 120–70 ka, when ice cover alternates between a ‘proto-Laurentide-Innuitian’ ice sheet and independent ice domes (cf. Marshall et al., 2000); although an ice-free corridor remains open between the LIS and the CIS for the period ~105 ka to ~75 ka. As noted by Marshall et al. (2000), this behaviour reflects both Milankovitch- and millennial-frequency climate oscillations, resulting in a dynamic ice sheet with migrating domes and divides (cf. Boulton and Clark, 1990a,b). Moreover, these embryonic domes acted as the seed points for major dispersal centres during longer or colder stadials and probably survived later inter-stadials as residual domes (cf. Kleman et al., 2010).

4.1.1. Inception over Foxe Basin?

Perhaps the most unexpected aspect of our modelled inception pattern is the very early nucleation of ice over Foxe Basin (ca 118 ka: Fig. 3) and then Hudson Bay. Generally, it has been assumed that these large marine basins were infilled by ice flowing from adjacent domes on land, but our modelling clearly depicts ice in Foxe Basin

at 118 ka before adjacent land has any substantial ice cover. This might suggest that marine processes (e.g. sea ice formation) can play an important role in ice sheet inception and, indeed, this is exactly what Denton and Hughes (1980) proposed in their Marine Ice Transgression Hypothesis (MITH). Under this scenario, climatic cooling leads to permanent sea-ice, whose albedo allows the regional snowline to descend to sea level. A mechanism to expand the snowline to low elevations is shared by proponents of the instantaneous glacierization hypothesis (e.g. Ives et al., 1975), but the MITH also allows for the sea ice to thicken into ice shelves that eventually ground in shallow marine embayments, thereby creating marine ice domes that expand into ice sheets (Hughes, 1987). In his theoretical treatment, Hughes (1987) specifically cited Foxe Basin as one of the first places where an ice shelf might ground in North America and then spread south into Hudson Bay.

The importance of sea ice formation in lowering regional albedo during glacial inception was also emphasised by Koerner (1980), who further noted that sea ice may have “allowed survival of the [North American] ice sheet through the warmer periods of the interstadials and promoted ice cap growth during ensuing periods, which might not have had the right conditions to start the ice sheet core itself” (p. 157). Using a GCM, Dong and Valdes (1995) also concluded that orbitally-induced changes in sea surface temperature and thus sea ice were likely to be a very important in initiating glaciation in the Canadian Arctic.

Although Hughes (1987) acknowledged that windward growth (Flint, 1943) and instantaneous glacierization (Ives, 1957) could have been important for inception patterns in other northern hemisphere ice sheets, he suggested that marine ice transgression was probably dominant in the Canadian Arctic Archipelago and our modelling also points towards the importance of marine processes (sea ice formation) in this region (Fig. 3). However, there are other factors which might also be important. These include the Sangamonian topography (adjusted so that the surface topography at the end of a model run is close to that observed for present-day), the degrees of freedom in the climate forcing, and perhaps the representation of ice calving and ice shelves. Thus, whilst the model potentially points towards the importance of marine processes in North American ice sheet inception, a more complete sensitivity analysis is required to reduce uncertainties in the causal explanation. In particular, the use of the SIA limits the models ability to fully capture marine ice sheet physics, which are likely to have been particularly important during OIS 5, when the ice sheet margins advances and retreats repeatedly over high latitude seaways (e.g. in Foxe Basin and Hudson Bay/Strait). As such, this finding should be regarded as preliminary.

4.2. Comparison to previous reconstructions of ice sheet build-up to the LGM

We now compare our model results with previous attempts to reconstruct the pre-LGM evolution of the NAISC and, particularly, the LIS. We refrain from a detailed comparison with localised records of pre-LGM ice sheet deposits (reviewed, for example, in Dredge and Thorleifson, 1987; St-Onge, 1987; Vincent and Prest, 1987; Clark et al., 1993), which is beyond the scope of our ice sheet-wide synthesis. Instead, we focus on comparison to previous reconstructions of the broad patterns of growth and decay.

4.2.1. Comparison to previous numerical modelling studies

Few studies have used numerical ice sheet models to examine the evolution of the LIS during the last glacial cycle (Tarasov and Peltier, 1997a,b; Marshall and Clarke, 1999; Marshall et al., 2000; Bintanja et al., 2002; Marshall and Clark, 2002). Typically, they focused on capturing the main features of the LGM ice sheet, presented as a time series of ice sheet properties (usually volume, but

sometimes extent, maximum thickness, warm-based fraction, etc.) from the last interglaciation (ca 120 ka) to present (e.g. Marshall et al., 2000). Early attempts using a simplified energy balance model tended to depict a monotonic growth towards the LGM (e.g. Tarasov and Peltier, 1997a) but more recent climate forcing prescribed by the GRIP (Greenland Ice Core Project) record has resulted in most models reproducing the higher-frequency growth and decay cycles (e.g. Marshall et al., 2000; Bintanja et al., 2002; Kleman et al., 2002; Marshall and Clark, 2002), that our modelling unsurprisingly replicates (Figs. 4 and 5).

In agreement with our results, most numerical modelling efforts (e.g. Marshall and Clarke, 1999; Marshall et al., 2000; Bintanja et al., 2002) simulate early ice sheet growth (OIS 5d?), albeit not always as rapid as in our modelling, followed by a minimum ice extent in OIS 5a (80 ka). All models reproduce growth to a large OIS 4 ice sheet, and most depict a minimum during OIS 3 (ca 55 ka), followed by punctuated growth towards the LGM. The major discrepancies amongst different numerical models largely emerge in terms of the precise timing of the various maxima/minima, which is likely a reflection of the specific treatment of the climate forcing. Indeed, the relatively rapid increases/decreases in ice volume produced in our model are largely attributable to the high frequency variability in the climate forcing index (Fig. 2), which percolate to varying extents (depending on phasing, calibration parameters, duration, intensity) into the ice volume curves. This is to be expected from a complex non-linear model which simulates the interactions and feedbacks of numerous processes, but it is more difficult to determine the potential contributions from different model processes/components (e.g. marine ice sheet dynamics, ice streams, subglacial sediment deformation) that might further modulate or even drive the rapid changes in ice volume. Elucidating those mechanisms would require a rigorous sensitivity analysis which is beyond the scope of the present paper. Rather, we now focus on the handful of studies that allow a rigorous assessment of our model results in capturing the spatial and temporal patterns of ice sheet dynamics, especially in terms of the geological evidence of ice extent within each OIS.

4.2.2. OIS 5

Based on a comprehensive review and synthesis of available stratigraphic records Clark et al. (1993) focus on when and where the LIS and CIS first developed and on tracing their growth to the LGM. In agreement with our modelling, they suggest that the LIS first developed over Keewatin, Quebec and Baffin Island during OIS 5 (Section 4.1), although our modelling suggests other large islands in the CAA and Foxe Basin were equally important. They also suggest that its encroachment into the western Canadian Arctic occurred late during stage 5 with a further possible advance into the St Lawrence Lowland. Their suggestion of early ice over the western CAA was based largely on the interpretation of till sheets on Banks Island, as well as equivalents on Melville Island and in the Tuktoyaktuk Coastlands, although these deposits are now considered to date to OIS 2 (England et al., 2009). Nevertheless, the presence of ice in the western CAA is certainly supported by our modelled ice sheet at 90 ka (OIS 5b). Following OIS 5b, Clark et al. (1993) suggest that the ice sheet retreated from these regions and they argued that the Hudson Bay Lowlands, the western Canadian Arctic, and the St Lawrence Lowlands may have been ice free during stage 5a (albeit with no firm chronological constraints), which broadly matches our reconstruction at 80 ka (Fig. 4).

Clark et al. (1993) do not discuss the possibility of a large (but thin) ice sheet during 5d (ca 110 ka: Fig. 4) but their overall conclusion that an OIS 5 ice sheet developed to ca 70% of its (full-glacial) OIS 2 extent, followed by significant deglaciation, is exactly what our modelling depicts. St-Onge (1987) reached a similar

conclusion in noting that where the ice sheet may have grown during OIS 5, most of the ice probably disappeared prior to major growth during OIS 4. A similar sequence is shown in our modelling of the OIS 5a/4 transition from ~80 to ~65 (see Fig. 4).

Further support for an extensive ice sheet in OIS 5 is also noted by Marshall et al. (2000) who point out that marine oxygen isotope records (e.g. Shackleton, 1987; Lambeck and Chappell, 2001) indicate rapid ice sheet growth in the early stages of glaciations and our modelling would also suggest rapid growth and establishment of the CIS during OIS 5. Indeed, elsewhere, Kleman et al. (1997) infer a similarly rapid growth of the Fennoscandian Ice Sheet (FIS) following the last interglaciation. They used mapped flow patterns alongside stratigraphic records and correlation to global sea level volumes and suggested that the FIS reached an OIS 5 maximum ca 110 ka (i.e. 5d, see also Lagerbäck, 1988a,b), before a relatively rapid recession by 100 ka. However, whilst the LIS reached a similar extent to the OIS 4 and 2 maxima, the OIS 5 FIS probably only covered approximately 50% of the LGM extent (see Fig. 11 in Lundqvist, 1992; Kleman et al., 1997). Kleman et al. (1997) also suggest that the FIS during OIS 5b was restricted in coverage and largely cold-based, which is similar to our reconstructed ice sheet during this period (e.g. 90 ka on Fig. 4) and they also highlight evidence of near-complete deglaciation during OIS 5a (e.g. Lagerbäck and Robertsson, 1988), which is in contrast to the NAISC (see e.g. 80 ka on Fig. 4).

Boulton and Clark (1990a,b) used geological evidence to reconstruct the pre-LGM behaviour of the LIS, but their study utilised the glacial lineation record and they inferred relative age relationships based on cross-cutting flow-sets, in addition to correlation with previously published till stratigraphic data (e.g. Andrews et al., 1983). Their synthesis was focussed largely on the Hudson Bay region and a key conclusion was that the major domes of the ice sheet were highly mobile during the last glacial cycle, with ice divides shifting by up to 1000–2000 km. Specifically, they identified major changes in the location of the Keewatin and Quebec-Labrador domes, either side of Hudson Bay, with periods of coalescence separated by at least one relatively long period of ice-free conditions in the James Bay Lowlands.

The first major period of coalescent domes illustrated by Boulton and Clark (1990a,b) is shown from around 110 ka, which would correspond to our large ice sheet at this time. However, whilst they, like us, favour a large OIS 5 ice sheet, our modelling shows a rapid deglaciation of this large ice sheet by 100 ka, whereas Boulton and Clark (1990a,b) only show it starting to recede from ca 100 ka. The timing of Boulton and Clark's (1990a,b) subsequent ice-free conditions in the Hudson Bay Lowlands is more uncertain but they depict it at around 75 ka, based on a period of marine inundation suggested by Andrews et al. (1983). Again, this is in broad agreement with our minimal ice extent from around 85 to 75 ka (Fig. 5). Andrews et al. (1983) further speculate that the southern shore of Hudson Bay was inundated at ca 105 ka. Our modelling (e.g. Fig. 4) does not support this interpretation for this time and suggests that if it did occur, it is more likely with our modelled configuration by 100 ka.

Kleman et al. (2002) present a plausible evolution of the NAISC during inception and build-up based on geological evidence and also in conjunction with numerical modelling. Their study covered the time interval from 120 ka to 55 ka, with a particular emphasis on the 90–60 ka period, which incorporates the large ice sheet growth during OIS 4. Their modelled ice sheet volume from 120 to 55 ka (see Fig. 1 in Kleman et al., 2002) mirrors the general pattern revealed in our analysis, with generally low ice sheet volumes from 120 to ~70 ka, followed by a period of major growth at the start OIS 4. They further suggest that this pattern captures the fundamental patterns of the ice sheet gleaned from the limited geological

evidence (in this case, three relict ice flow patterns represented by till lineations, striae, and till fabric data from northern Keewatin, the Interior Plains, and the Hudson Bay Lowlands). Both our and their reconstruction, for example, indicate a build-up of the ice sheet during OIS 5 with two stable dispersal centres ('nuclei') in the central Arctic/north-eastern Keewatin and Quebec-Labrador that survived inter-stadial warming. The major difference during this period is our modelled ice sheet expansion around 110 ka, whereas Kleman et al. (2002) depict minimal ice volumes ($\sim 2\text{--}3$ m of sea level equivalent) from 115 to 90 ka, as do Marshall and Clark (2002).

Later work by Kleman et al. (2010) also inferred minimal ice volumes during stage 5 based on a more comprehensive analysis of the glacial geological and geomorphological record, which they integrated with previously published stratigraphical and chronological evidence. Indeed, whereas we reconstruct a coalescent LIS and Cordilleran Ice Sheet at 110 ka (OIS 5d), Kleman et al. (2002, 2010) suggest this did not occur until ~ 64 ka (OIS 4). However, Kleman et al. (2002) acknowledge that their early 'central Arctic ice sheet' may have extended further south, which would be more in line with our reconstruction during stage 5, and that of Clark et al. (1993). Kleman et al. (2010) also discuss the possibility of a large ice sheet during a stadial in OIS 5 but they argue that 5d is unlikely because there was no residual ice left over in southern Quebec immediately after 5e to seed such a large ice sheet. Conversely, our modelling appears to show that residual ice is not an essential prerequisite for a large OIS 5d ice sheet (Figs. 3 and 4), but that marine processes (e.g. sea ice formation) could have played an important role (see discussion in Section 4.1).

4.2.3. OIS 4

During OIS 4, Clark et al. (1993) highlight several lines of evidence for a large ice complex that perhaps advanced across the continental shelf in Nova Scotia and south into New England. It may also be recorded in the Lake Ontario basin, and the western Arctic margin may have reached a similar extent to that recorded during its maximal position in OIS 5. Kleman et al. (2002) also note a massive increase in glaciated area at the OIS 5/4 transition. A large stage 4 ice sheet is shown by our modelling (e.g. 65 ka; Fig. 4) and, remarkably, our ice sheet at 65 ka is almost identical to the pre-Middle Wisconsinan maximum ice sheet extent portrayed by Vincent and Prest (1987), see Fig. 8a, which they outlined on the basis of an extensive review of glacial geologic evidence, albeit with minimum age control. Indeed, some of the evidence used to reconstruct this ice margin position has been subject to recent revision based, in particular, on new dating (e.g. England et al., 2009); but the overall conclusion of a large ice sheet is supported by our modelling, although not as extensive as the OIS 2 ice sheet (cf. Clark, 1992a).

We note that Boulton and Clark (1990a,b) show a much smaller LIS extent during OIS 4 (ca 60–70 ka). This was largely influenced by Andrews et al.'s (1983) assertion of ice free conditions along the southern shore of Hudson Bay at ~ 75 ka (C. Clark, pers. comm.). In line with Andrews et al. (1983), our modelling shows ice-free conditions in the Hudson Bay Lowlands from ca 100 to around 75 ka (see Fig. 4 and 5a) but followed by very rapid ice sheet growth to a near LGM-position at 65 ka. Boulton and Clark (1990a,b) simply depict a much slower growth from 75 ka, which helps explains the discrepancy between their much smaller ice sheet at 65 ka, compared to our modelled mid-Wisconsinan maximum. We also note that Kleman et al. (1997) appeal to an equally rapid build-up of the FIS during the climatic deterioration at the start of OIS 4. However, their OIS 4 FIS remains around 50% smaller than at the LGM, whereas our model of the LIS during OIS 4 is almost as large as it is at the LGM. Rapid build-up of ice sheets at the OIS 5/4 transition is also recorded in sea level records and Cutler et al. (2003) estimate

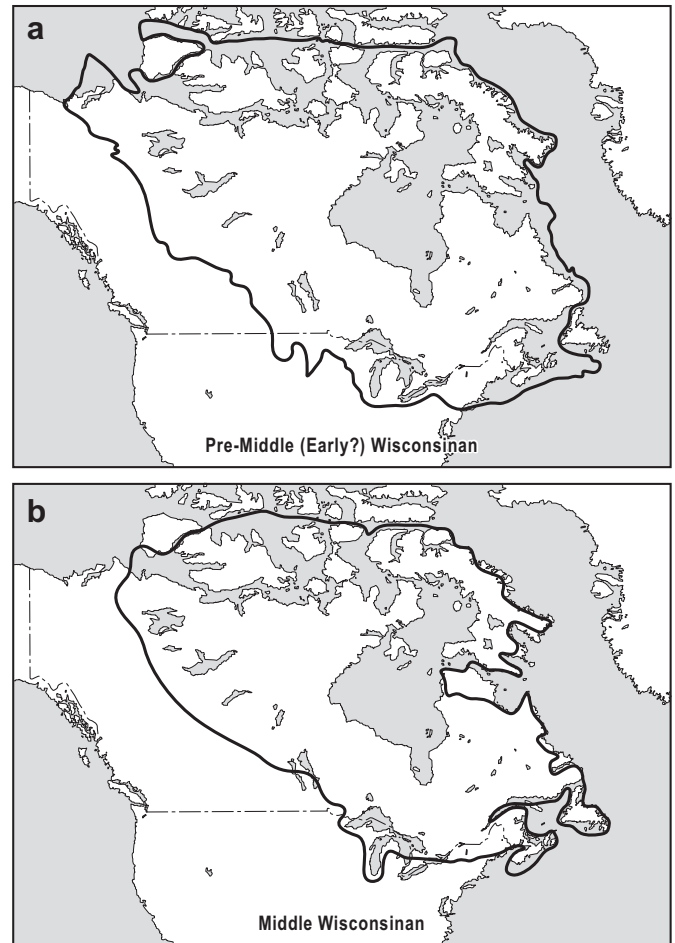


Fig. 8. (a) Possible extent of the Laurentide Ice Sheet in pre-Middle (Early?) Wisconsinan time based on an extensive review of geological evidence by Vincent and Prest (1987). Although some of the evidence used to reconstruct this ice margin has been subsequently re-interpreted (e.g. England et al., 2009), the margin position is almost identical to our modelled Early Wisconsinan ice sheet at 65 ka (see Fig. 4). (b) Possible maximum extent of the Middle Wisconsinan ice sheet based on an extensive review of geological evidence by Dredge and Thorleifson (1987). Dredge and Thorleifson (1987) acknowledge uncertainty over the precise extent of the ice sheet during OIS 3 and also discuss more minimal configurations. However, the maximum configuration shown here is entirely consistent with our relatively large ice sheet between ca 45 and 50 ka (Fig. 4).

that sea level fell 60 m in <6 ka (10.6 m/1000 yr), which is certainly in agreement with our estimates of the rapidity of ice sheet growth at the start of OIS 4 (Fig. 5).

Our OIS 4 ice maximum at ~ 65 ka is coincident with the oldest widely recognised Heinrich event (H6) (cf. Kirby and Andrews, 1999; Hemming, 2004) and Hemming (2004) notes that this was a time of generally high IRD abundance, perhaps analogous to the LGM, which further supports the presence of a large OIS 4 ice sheet. Indeed, Kleman et al. (2002) suggest that the limited infilling of ice in Hudson Bay up until ~ 65 ka in their reconstruction might be linked to the timing of the onset of Heinrich events (H6–H1; 67–64 to 15 ka; Hemming, 2004). They speculated that these intermittent massive iceberg discharge events could only develop when a large ice sheet was attained. Related to this, Kleman et al. (2002) further suggest that the LIS evolved from a largely cold-based to largely warm-based ice sheet, with the implication being that “its dynamics during the early part of the cycle may have differed fundamentally from that during the latter part” (p. 88). A similar conclusion was reached by Marshall and Clark (2002) who argued

that a substantial fraction (60–80%) of the ice sheet was frozen to the bed for the first 75 ka of the glacial cycle. Our analysis of basal thermal regime also shows a transition from a largely cold-based to a largely warm-based ice sheet, with only 20–30% of the bed warm-based between 120 and ca 70 ka (Fig. 6). We also note that the FIS is thought to have undergone a similar transition from a thinner, cold-based ice sheet to one with more extensive warm-based ice during the LGM (Kleman et al., 1997; Lambeck et al., 2010). For example, Kleman et al. (1997) suggest that the scarcity of flow traces from OIS 3 is likely to be a reflection of a frozen bed in the core area of the FIS that persisted throughout much of the last glacial cycle. However, an important difference is that, unlike the LIS, this core area was only partly consumed by inward transgressing warm-based zones during the final decay phase (cf. Kleman et al., 1997). As such, a major difference between the LIS and the FIS is that the latter, because of its more persistent cold-based core, preserves a much richer record of pre-LGM flow traces. A corollary of this is that the FIS may have been more “glaciologically stable” for the most part of its history (Kleman et al., 1997: p. 296).

Our modelling also supports a large CIS during stage 4 (e.g. at 65 ka in Fig. 4) but this is clearly in conflict with the long-standing view that there was a prolonged non-glacial interval from ca 60–30 ka (the Olympia/Boutellier Nonglacial Interval), when glaciation was probably confined to mountain areas throughout much of OIS 3 (see reviews in Ryder and Clague, 1989 and Clark et al., 1993). As noted above (Section 2), the complex topography of the Cordillera presents a serious problem for continental-scale ice sheet modelling, which necessitates large grid cells. For example, ablative valleys are not resolved sufficiently and this may partly explain the persistence of ice in this region, when geological/chronological constraints might suggest otherwise. In contrast, topographic variance over the LIS is much lower and model results are, correspondingly, far more credible.

4.2.4. OIS 3

Clark et al. (1993) suggest that recession from the OIS 4 position occurred early in stage 3, but that the ice sheet had likely started to re-advance again by 50 ka, e.g. in Illinois, the Lake Ontario basin, and possibly in the headwaters of the ancestral Mississippi River in Minnesota and Wisconsin. This rapid retreat from stage 4 and the re-advance of the southern margin is compatible with our reconstruction (see Figs. 4 and 5). Dredge and Thorleifson (1987) were more tentative than Clark et al. (1993) in their reconstruction of the Middle-Wisconsinan (OIS 3) ice sheet, and they highlight the difficulty of constraining the ice sheet extent based on the dating of deposits of that age. Instead, they offered three hypotheses that range from a minimum configuration with separate domes in Quebec, Keewatin and over Baffin Island (their hypothesis 2, intended to accommodate the youngest Wisconsinan marine episode in Hudson Bay proposed by Andrews et al., 1983) to a more maximal outline with an ice margin as far south at the Great Lakes and extending north-westward to the Mackenzie River corridor, covering most of the CAA (their hypothesis 1, see Fig. 8b, which is more in line with Clark et al., 1993). They consider these working hypotheses to be tested and, in so far as it is possible with our work, our modelling strongly supports a large ice complex (>50 m eustatic equivalent) from ~45 ka, which then gradually built up to the LGM, albeit punctuated with successively less significant periods of retreat (Fig. 5). Nevertheless, given the major fluctuations in ice sheet volume and extent during the relatively long OIS 3 (± 30 m eustatic equivalent), it is not surprising that the poorly dated geological deposits could support either an extensive or more minimal ice sheet (cf. Dredge and Thorleifson, 1987). Note, however, that our modelling and other more recent work (e.g. Dredge and Cowan, 1989; Clark

et al., 1993; Allard et al., 2012) does not support ice free conditions in Hudson Bay during OIS 3. Indeed, there is little evidence of marine sediments in the Hudson Bay Lowlands younger than the Sangaonian (Bell Sea) (cf. Allard et al., 2012).

Kleman et al. (2010) also reconstruct ice sheet extent in OIS 3 and we note almost identical configurations for this time period, with the southern margin extending into the Great Lakes region and with major domes over Quebec-Labrador and in north-eastern Keewatin (cf. Fig. 8c in Kleman et al., 2010 with our Fig. 4). Their late stage 3 depiction (Fig. 8b in Kleman et al., 2010) likely represents our last interstadial position around 30 ka (Fig. 5a), which Dyke et al. (2002) also suggest represents the most recent minimum ice extent prior to the LGM, although based on un-calibrated radiocarbon dating.

It is important to note that even our smallest modelled OIS 3 ice complex in North America was relatively extensive and consumed a global sea level equivalent of ca 30 m (e.g. ca 55 ka on Figs. 4 and 5a). This is in contrast to recent work on the FIS, which suggests that most of Scandinavia might have been ice-free during the early and middle parts of OIS 3 (cf. Helmens and Engels, 2010; Lambeck et al., 2010; Wohlfarth, 2010). Thus, if correct, a key implication is that the NAISC accommodated most of the ice from the Northern Hemisphere during OIS 3, although see the caveats above (Section 4.2.3) regarding the likely over-prediction of the CIS during this interval.

4.2.5. Summary: OIS 5 to OIS 3

Our numerical modelling depicts extensive ice sheets early during OIS 5 (5d) at 110 ka, which retreat rapidly by 100 ka, and to a minimum Wisconsinan post-OIS 5e position by ca 80 ka. Thereafter, large ice sheets in North America developed rapidly during OIS 4, reaching a maximum extent at ca 65 ka, before a rapid recession to an OIS 3 minimum ca 60–55 ka. This was followed by a more gradual period of ice sheet growth punctuated by episodes of successively less recession (minima at 40 and 30 ka) before a maximum LGM position is attained at ca 25 ka. This general evolution compares well with previous numerical modelling (Marshall and Clarke, 1999; Marshall et al., 2000; Bintanja et al., 2002) and only differs in terms of the precise timing of maxima/minima. However, some models fail to capture the rapid growth during OIS 5d, which is implied by the global sea level record (Marshall, 2002), and which our model captures partly due to the temporal deformation of the inferred GRIP temperature and partly because of the increased number of parametric controls (Fig. 1).

In terms of geological evidence, our modelled output is generally consistent with previous ice sheet-wide reconstructions that utilise geological evidence (most notably Vincent and Prest, 1987; Boulton and Clark, 1990a,b; Clark et al., 1993; Kleman et al., 2002, 2010). The only major discrepancies are the uncertainties about the precise timing of the OIS 5 maximum position (5d or 5b?: our results strongly favour 5d) and the lack of a large OIS 4 ice sheet in Boulton and Clark (1990a,b), which differs from most other reconstructions (e.g. Vincent and Prest, 1987; Clark et al., 1993). Thus, the consensus is that the pre-LGM ice sheet exhibited rapid episodes of growth and decay in the build-up to the LGM. Retreat from the stage 5d and stage 4 maxima, for example, contributed >7 and >16 m of eustatic sea level equivalent per thousand years, respectively. Indeed, comparisons suggest that the NAISC was more persistent and, relatively, experienced much larger volumetric fluctuations than the FIS, i.e. the OIS 5d and 4 maxima of the NAISC were almost as extensive as at the LGM, whereas those in Fennoscandia were, perhaps, more restricted; and the FIS may have almost disappeared altogether during OIS 3 (cf. Kleman et al., 1997; Lambeck et al., 2010; Wohlfarth, 2010).

4.3. Ice sheet dynamics (ice streaming?) during ice sheet growth and decay

Whilst our modelling supports previous inferences and extends our knowledge of the probable pre-LGM volume and extent of the NAISC, it may also offer some more tentative insight into ice sheet dynamics. One of the most dynamic features of an ice sheet are ice streams, rapidly-flowing arteries that play a key role in ice sheet mass balance through their ability to discharge large ice fluxes (typically several 10^5 km^3 per year). Observations of modern-day ice streams have revealed changes in their velocity (Howat et al., 2007) and position (Conway et al., 2002), highlighting their variability and heightening concern about their future contribution to sea level (Shepherd and Wingham, 2007). The danger of extrapolating these annual/decadal changes to draw conclusions about future ice sheet mass balance is, however, widely acknowledged (Conway et al., 2002; Nick et al., 2009); and this is one reason why numerous workers have extended the observational record through investigation of palaeo-ice streaming over millennial time-scales (e.g. see reviews in Stokes and Clark, 2001; Livingstone et al., 2012). An important consideration, however, is that although palaeo-ice streams have been implicated in Heinrich events dating back to $\sim 65 \text{ ka}$ (Hemming, 2004), most studies have focussed on their activity during deglaciation from the LGM (e.g. Conway et al., 1999; Lowe and Anderson, 2002; Kehew et al., 2005; Shaw et al., 2006; De Angelis and Kleman, 2007; Briner et al., 2009; Stokes et al., 2009; O'Cofaigh et al., 2010; Stokes and Tarasov, 2010). Thus, an assessment of their prevalence during the longer periods of ice sheet build-up would provide a useful comparison to their activity during ice sheet demise: are ice streams equally active during growth phases? How long does it take to initiate streaming in different settings and are ice streams located in similar locations to their counterparts during deglaciation from the LGM?

An obvious question to ask, therefore, is whether the fast-flow features generated by the model (see Section 3.3 and Fig's 4 and 7) are akin to ice streams and, as such, whether they might tell us something about their activity prior to the LGM (although there is no objective velocity threshold for ice streaming, we define it here as grounded ice flowing $>500 \text{ m/yr}$, i.e. typical of the lower-end of streaming velocities in modern-day ice sheets). Unfortunately, the use of the SIA and the relatively coarse resolution of the model preclude the ability to accurately reproduce the full stress balance of ice streams (see Hindmarsh, 2011; Kirchner et al., 2011). Thus, the extent to which the rapidly-flowing zones described above (Fig. 4) represent ice streams is questionable. Nevertheless, they evolve freely in the model as per the dynamically evolving basal temperature regime and stress balance. Moreover, Stokes and Tarasov (2010) indicate that this model successfully reproduces deglacial ice streams where geological evidence has been reported, despite the use of the SIA. Indeed, we note that the location of the fast flow-features generated in our pre-LGM modelling nearly always corresponds with the location of ice streams during deglaciation of the ice sheet, which have been identified based on a wealth of geological evidence, e.g. in major marine troughs and along the southern margin of the ice sheet (cf. Patterson, 1998; Stokes and Clark, 2001; Winsborrow et al., 2004; Stokes and Tarasov, 2010).

To illustrate the persistence of streaming velocities in both pre- and post-LGM settings in the NAISC, Fig. 9 shows the modelled basal velocity at 40 ka (OIS 3) with several major fast-flow zones corresponding to hypothesised locations from final deglaciation. In most cases, these patterns are likely the result of thermo-mechanical feedbacks, whereby thicker ice in topographic troughs enhances basal warming, increasing velocity, further increasing warming, further increasing velocity (Payne and

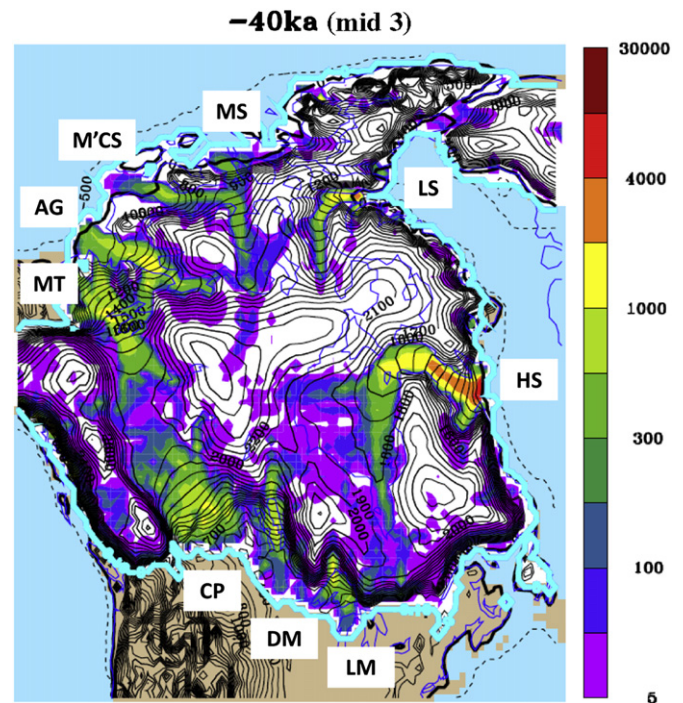


Fig. 9. Correspondence between modelled fast flow features at 40 ka and locations where ice streams are known to have existed during the deglacial interval. Anti-clockwise: HS = Hudson Strait (e.g. Andrews and MacLean, 2003), LS = Lancaster Sound (e.g. Klassen and Fisher, 1988; De Angelis and Kleman, 2007), MS = Massey Sound (e.g. England et al., 2006), M'CS = M'Clure Strait (e.g. Stokes et al., 2005), AG = Amundsen Gulf (e.g. Stokes et al., 2006), MT = Mackenzie Trough (e.g. Beget, 1987), CP = Canadian Prairies (e.g. O'Cofaigh et al., 2010), DM = Des Moines Lobe (e.g. Patterson, 1997), LM = Lake Michigan Lobe (e.g. Kehew et al., 2005).

Dongelmans, 1997). Thus, the main control on the location of these fast-flowing outlets appears to simply be the presence and supply of warm-based ice in association with soft sediment that floors these troughs. Rapid velocities experienced at the land-terminating margins (e.g. the southern margin) can also be attributed to the widespread availability of soft-sediments in these regions (cf. Clark, 1992b, 1994) that facilitated rapid flow and surge-type behaviour through sub-glacial till deformation and/or basal sliding (cf. Patterson, 1997, 1998), and sometimes in association with calving into proglacial lakes (cf. Stokes and Clark, 2004). Similar processes have been invoked to explain the location of episodic ice streams operating both pre- and post-LGM in the FIS (Houmark-Nielsen, 2010).

Given that there is some indication that the modelled fast flow zones are not spurious but may reflect the pre-LGM counterparts of post-LGM ice streams, we are reasonably confident that model output is likely to reflect the broad pattern and timing of the major pre-LGM ice stream activity, especially topographic ice streams, and can thus offer some preliminary insight into this aspect of ice sheet dynamics. It should also be emphasised that the patterns of basal velocity (e.g. Fig. 4) represent the mean of the ensemble of 'best-fit' model runs. In this sense, they offer a probabilistic assessment of where rapid velocities are most likely to occur and are, in effect, a smoothed history compared to individual runs (i.e. short-term spatial and temporal variability is averaged out). Indeed, given the sensitivity of millennial-scale ice stream oscillations to model numerics (Calov et al., 2010) and likely to missing processes such as basal hydrology, it is not the aim of this study to capture such behaviour.

The modelled ice sheet contains fast-flowing ice at all stages of growth from 110 to 20 ka (Fig. 4). Thus, ice streaming is likely to

have been a persistent feature in the pre-LGM ice sheets, with even the smallest and thinnest of them being drained by large topographic ice streams (note, for example, the Lancaster Sound Ice Stream in the very small ice complex at 110 ka). A more tentative conclusion is that the ice streams (*senus lato*) appear to increase in velocity and become more prevalent with ice sheet build-up (Figs. 4 and 7). This is shown in Fig. 10a, which plots the total 'ice stream' flux at the margin of the grounded ice sheet through time. The ice stream flux is relatively low from 115 to 70 ka but there is a notable peak around 65 ka (OIS 4). Thereafter, the flux drops before another peak around 45 ka, followed by a variable but significant increase towards the LGM. This pattern is mirrored by the relative area of the ice sheet occupied by 'streaming' cells, see Fig. 10b, showing that they occupy a greater proportion of the ice sheet through time, as a result of enlargement of existing fast flow zones and new zones emerging.

We suggest that the increase in streaming activity during and following OIS 4 (Fig. 10a) is due to the fact that smaller ice sheets prior to this were generally restricted to harder crystalline substrates, whereas their expansion on to softer substrates and into

greater contact with the oceans led to unstable behaviour and an increased activity of ice streams (cf. Clark, 1992b, 1994; Winsborrow et al., 2010). Larger ice sheets have also had more time to develop a larger fraction of warm based ice (Fig. 6). It is most instructive, however, to consider the total ice stream flux as a fraction of ice sheet volume. Fig. 11a shows this and indicates that ice stream flux at the margin exhibits strong high-frequency variation superimposed on a fairly steady increase towards LGM. As one would expect, individual runs display more variance than the ensemble mean, but ice streaming appears to be scaled (non-linearly) to ice sheet volume. This is clear from Fig. 11b, which reveals that an excess of stream flux during decay, as compared to growth, is only discernible for ice volumes above 55 m eustatic equivalent and grows with ice volume.

In summary, this is one of the first studies to assess the pan-ice sheet behaviour of ice streams prior to the LGM and we view it as necessarily exploratory. However, it suggests that ice streaming appears to be non-linearly scaled to ice sheet volume with a power between two and three, with only moderate dependence on whether the ice-sheet is growing or decaying (Fig. 11). This implies that ice streams are significant throughout both growth and decay and, despite millennial scale variability, an evolving population

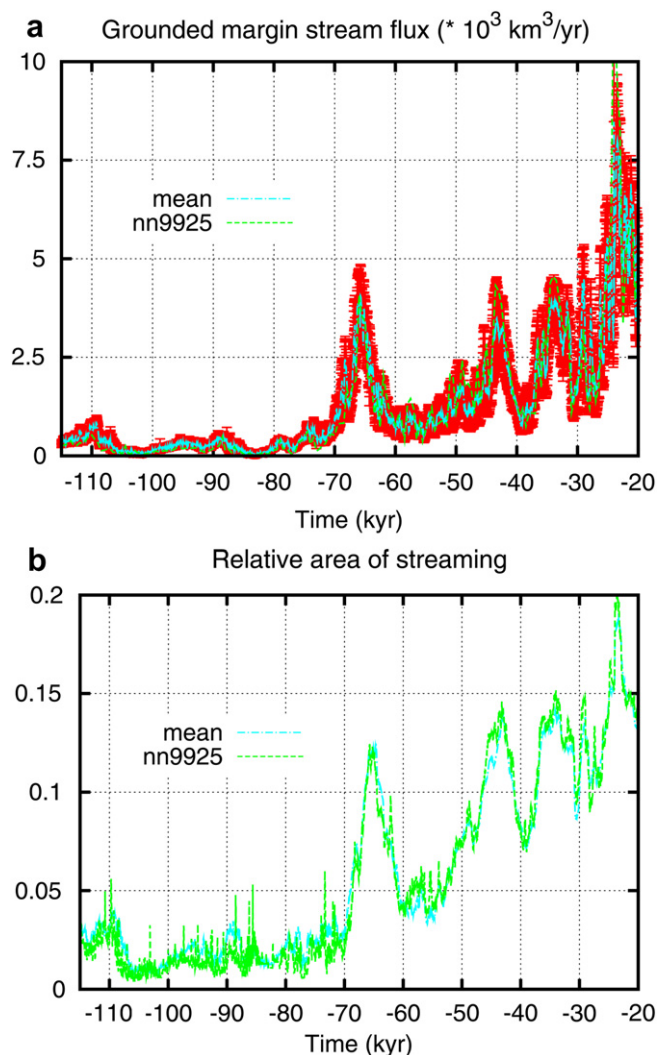


Fig. 10. (a) Total flux from all streaming ($>500 \text{ m/yr}$) cells at the margin of the grounded part of ice-sheet through time and, (b), relative area of streaming through time (grid cells $>500 \text{ m/yr}$). Both show ensemble mean, one standard deviation, and results for one of the best scoring model runs (nn9925). Note that streaming flux increases through time towards the LGM and is largely modulated by ice sheet volume, see also Fig. 11.

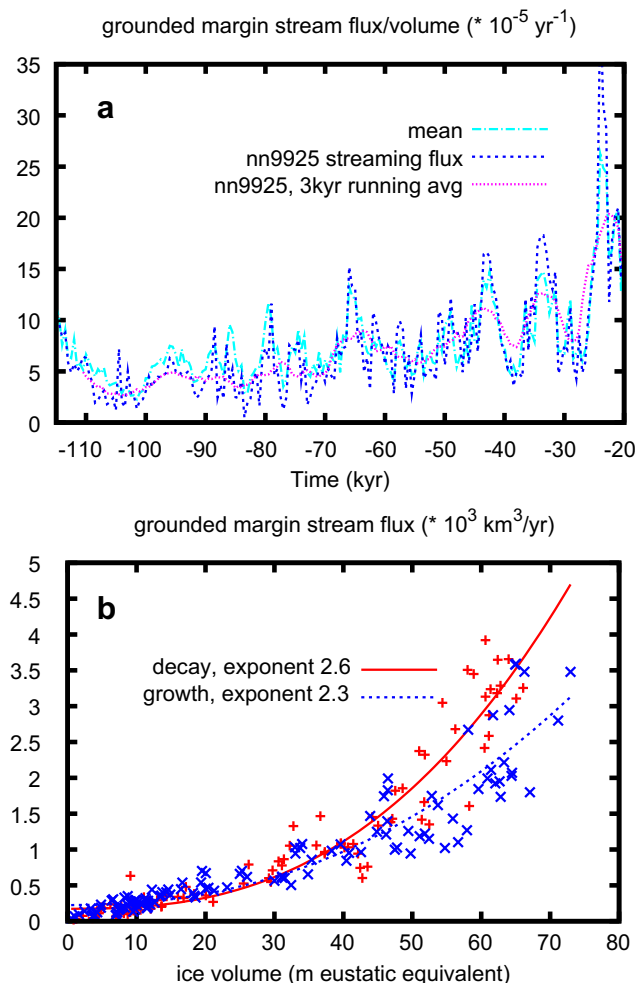


Fig. 11. (a) Total flux from all streaming ($>500 \text{ m/yr}$) cells at the margin of the grounded part of the ice sheet through time as a function of ice volume. In (b), mean ensemble results were decomposed into growth and decay sequences and the 29 ka to 20 ka interval was excluded because of the dynamic facilitation of H-events (see Section 3.3). These plots demonstrate the strong dependence of ice stream activity on ice sheet volume with no marked dependence on whether the ice sheet is in a growth or decay stage except at ice volumes greater than about 55 m eustatic equivalent.

deliver a relatively stable contribution to overall ice discharge over long (>5–10 ka) time-scales (Fig. 11a).

4.4. Future work

The results of our numerical modelling match well with previous attempts to reconstruct the pre-LGM inception and build-up of the NAISC and have generated some important new insights regarding its extent, volume, and dynamics. However, we identify five areas where further research would benefit/complement future modelling efforts:

- (1) Whilst it is reasonable to assume that ice streaming was a persistent feature of the pre-LGM LIS, a potentially novel outcome of our analysis is the hypothesis that ice stream activity is largely scaled to ice sheet volume. This represents an important context for our understanding of ice streaming and ice sheet dynamics at timescales >10 ka and we suggest that testing this hypothesis should be prioritised in future modelling that incorporates higher order physics.
- (2) A key uncertainty that prevents a more rigorous comparison is that there are few dates to constrain the geological evidence. As noted by Clark et al. (1993) our knowledge of the chronology of the LIS is mostly reliant on radiocarbon dating applied to the deglacial interval. However, it is possible to date deposits using other methods (e.g. U–Th, Ar/Ar) to bracket the extensive till stratigraphies in some areas and we note the importance of recent work in the Hudson Bay Lowlands in this regard (Allard et al., 2012). Such constraints on pre-LGM ice margin positions would thereby permit a more rigorous model calibration, such as has been applied to the deglacial interval (e.g. Tarasov et al., 2012), especially early (e.g. OIS 5 and 4) ice margin indicators (cf. Stroeve et al., 2002).
- (3) The results of our numerical modelling complement efforts that scrutinise pre-LGM iceberg-rafted debris (IRD) records in ocean sediment records (e.g. Farmer et al., 2003; Hemming, 2004; Roy et al., 2009). Our analysis highlights the importance of marine troughs as locations for large palaeo-ice streams (Fig. 7) but relatively few IRD records have been linked to specific ice stream catchments, although this is certainly possible through analysis of their provenance (see Andrews and MacLean, 2003; Hemming, 2004; Stokes et al., 2005). For example, Farmer et al. (2003) showed that the provenance of H3 is more closely associated with a south-eastern (St Lawrence estuary) outlet, as opposed to Hudson Strait, and we note a sharp increase in velocities of the Gulf of St Lawrence ice stream coincident with H3 at around 30 ka (see Fig. 7c). Thus, there is huge potential to test and compare model predictions against IRD records. Moreover, Hemming (2004) notes that there are few published data on IRD events prior to H6 (~60 ka) and she highlights this as a priority for future research (e.g. Roy et al., 2009). She draws attention, for example, to a study by Rasmussen et al. (2003) that identified 12 IRD peaks in a sediment core off Newfoundland throughout the last 130 ka whose provenance is uncertain. Results of our numerical modelling of several ice stream catchments suggests that ice occupied major marine troughs before 60 ka (e.g. at 110 ka and 65 ka) and, as such, could have delivered IRD to the ocean sediment record. The large ice sheet at 110 ka (Fig. 4a), for example, clearly depicts ice streams occupying several major marine troughs (e.g. Hudson Strait, M'Clure Strait, Amundsen Gulf) and their presence and activity could be confirmed in IRD records.
- (4) As noted above (e.g. with respect to the 'old' and 'new' ensemble results of the warm-based fraction: Fig. 6), the model

results are heavily influenced by the climate forcing. An obvious deficiency, therefore, is the uncertainty of the timing/phasing of climate evolution (along with possible regional asynchronies), which is poorly known the further back in time the modelling extends. Reducing and better capturing the uncertainty of past climate forcing is a non-trivial task but would clearly benefit pre-LGM ice sheet modelling.

- (5) Finally, better constraints on the pre-LGM ice volume chronology are needed. On-going calibration of all major ice sheets should better quantify bounds on pre-LGM partitioning of sea-level contributions. But there is correspondingly an urgent need for improving the far-field record of sea level change to further constrain this partitioning.

5. Conclusions

This paper presents a new analysis of the evolution and dynamics of the North American Ice Sheet Complex during its inception and build-up to the LGM using a 3D thermo-mechanical Glacial Systems Model that has undergone extensive calibration against a large and diverse set of geophysical data for the deglacial interval. A probabilistic assessment of ice sheet properties (thickness, basal velocity, etc) reveals major (sometimes abrupt) fluctuations in ice volume (± 10 s m of eustatic sea level equivalent in <5 ka) and a highly dynamic ice complex with major shifts in the location of ice divides. These rapid episodes of both growth and decay are consistent with sea level records (e.g. Cutler et al., 2003).

In agreement with previous work, the model reproduces ice sheet inception in Arctic/subarctic plateaux/uplands with ice masses growing and coalescing with a Quebec-Labrador dome within <10 ka after the last interglaciation. Preliminary findings also highlight the possibility of early inception over shallow marine basins, e.g. Foxe Basin (cf. Hughes, 1987). Thereafter, our numerical modelling broadly matches geological evidence (e.g. Clark et al., 1993; Kleman et al., 2002, 2010) and previous modelling attempts (e.g. Kleman et al., 2002; Marshall and Clark, 2002) of the evolution of the ice complex, supporting the notion of a large but thin OIS stage 5d (110 ka) ice sheet that covered 70–80% of the area occupied by subsequent full glacial ice sheets in OIS 4 and 2 (cf. Clark et al., 1993). The ice complex then retreated to a small, thin OIS 5 minimum (~80 ka) over the original inception grounds in north-eastern Canada, before a rapid period of growth saw a large OIS 4 ice complex at ~65 ka, which largely resembled the OIS 2 (LGM) ice complex, but with ~10 m less volume (eustatic equivalent). Subsequently, ice retreated rapidly to a minimum OIS 3 position ~60 ka, with a peak contribution to sea level >16 cm per century. The remainder of OIS 3 was characterised by several periods of growth and increasingly less significant retreat, culminating in the LGM (OIS 2) ice complex.

Consistent with previous work (e.g. Marshall and Clark, 2002), the fraction of the bed occupied by warm-based ice increases from 20 to 30% to >50% at the LGM, largely modulated by patterns in ice volume. However, even relatively small and largely cold-based ice sheets were drained by fast flow features (usually in major marine troughs). Despite limitations imposed by the use of the 'shallow-ice approximation' it is reasonable to conclude that ice streaming was a persistent feature during ice sheet build-up and most fast-flow features generated in pre-LGM ice sheets correspond to post-LGM ice stream locations (cf. Patterson, 1998; Winsborrow et al., 2004; Stokes and Tarasov, 2010). The extent of the marine margin is an important control on ice stream activity (cf. Winsborrow et al., 2010) and, where ice exists in deep topographic troughs, ice streaming generally persists (though not necessarily continuously), irrespective of climate forcing. Land-terminating ice streams appear more akin to ephemeral surge behaviour over soft

sediments and only occur when climate drives the ice sheet margin over such sediments (e.g. at the southern margin).

Taken together, a potentially novel insight from our analysis is the hypothesis that the overall area and flux of streams (cells $>500 \text{ m a}^{-1}$) is non-linearly proportional to ice sheet volume. This may be an artefact of our model physics and ice sheet mass balance but, if correct, the implication is that ice streams are significant during phases of both growth and decay and, overall, make a relatively stable contribution to mass loss, irrespective of whether the ice sheet is growing or shrinking. This, in turn, has implications for the behaviour of ice streams in modern-day ice sheets, where notable changes have recently been reported in their velocity and thickness (cf. Howat et al., 2007), but which lack a long-term context. We therefore suggest that further testing of this hypothesis should be a priority with a more rigorous treatment of ice streaming using higher-order physics with full inclusion of longitudinal/membrane stresses (cf. Hindmarsh, 2011; Kirchner et al., 2011). Future numerical modelling would also benefit from additional geological constraint data from the pre-LGM interval, such as the ongoing dating of till stratigraphies (e.g. Allard et al., 2012) and margin positions.

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