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Journal of Hydrology 196 (1997) 47–65

Journal of Hydrology

A parsimonious model for simulating flow in a karst aquifer

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Received 19 July 1996; revised 3 December 1996; accepted 5 December 1996

Abstract

This paper describes the hydrologic system associated with the Barton Springs portion of the Edwards aquifer and presents a lumped parameter model capable of reproducing general historical trends for measured water levels and spring discharge. Recharge to the aquifer was calculated based on flow loss studies of the creeks crossing the recharge zone and on estimates of the rate of diffuse infiltration of rainfall. Flow measurements on each creek above and below the recharge zone were used to develop a relationship between flow above the recharge zone and the rate of recharge. The five-cell groundwater model, each cell corresponding to one of the watersheds of the five main creeks crossing the recharge zone, was developed to support the management objectives of the City of Austin. The model differs from previous models in that the aquifer properties within cells are allowed to vary vertically. Each cell was treated as a tank with an apparent area and the water level of a single well in each cell was used to characterize the conditions in that cell. The simple representation of the hydrologic system produced results comparable to traditional groundwater models with fewer data requirements and calibration parameters. © 1997 Elsevier Science B.V.

1. Introduction

The Edwards aquifer is a karst system which lies in a broad arc across central Texas, USA. The portion of the aquifer located just south of the City of Austin is a hydrologically separate system (the darkly shaded area in Fig. 1), which discharges primarily at Barton Springs. This portion of the aquifer provides drinking water to about 35 000 residents in areas without access to the city drinking water system and provides recreational amenities at Barton Springs Pool, a municipal swimming pool formed by a dam just downstream

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PII S0022-1694(96)03339-2

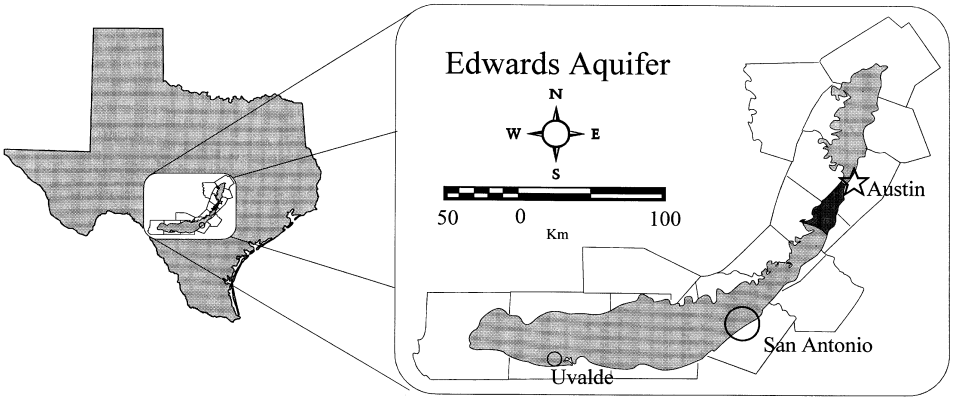

Fig. 1. Location of the Edwards aquifer.

from the spring. The Barton Springs salamander, which exists only in the vicinity of the springs, is also dependent on spring discharge for its survival.

The main goal of this research effort is the development of a computer model of the aquifer capable of predicting regional changes in water levels in the aquifer and discharge at the Barton Springs resulting from changes in the surface water systems. This paper describes the development and calibration of a parsimonious hydrologic model for the Barton Springs portion of the Edwards aquifer. Conventional groundwater models are overly complicated for this task given the uncertainties in parameterization and the difficulties associated with estimating changes in recharge characteristics resulting from development. A new type of lumped parameter model that allows vertical variation within cells is proposed and developed. This type of model should improve predictions in water table aquifers which are strongly stratified, while it retains the simplicity resulting from using lumped parameters.

2. Description of study area

The Edwards is a complex carbonate aquifer which exhibits numerous karst features. The Barton Springs portion of the Edwards aquifer covers an area of approximately 400 km^2 and is composed of Cretaceous age Edwards Limestone and Georgetown Limestone which dip generally to the east. It is underlain by the relatively impermeable Walnut Formation and bounded on the west by the Glen Rose Limestone. These rocks yield relatively little water compared with the Edwards. To the east, the water in the Edwards gradually becomes more saline, with the eastern boundary of the aquifer commonly considered to be the line where the concentration of total dissolved solids exceeds 1000 mg l^{-1} (approximately coincident with Interstate Highway 35). In the eastern portion of the study area, the aquifer is confined by the Del Rio Clay. To the south, a groundwater divide in the vicinity of Onion Creek separates the Barton Springs portion from the southern portion of the Edwards. Fig. 2 shows the extent of the aquifer in the study

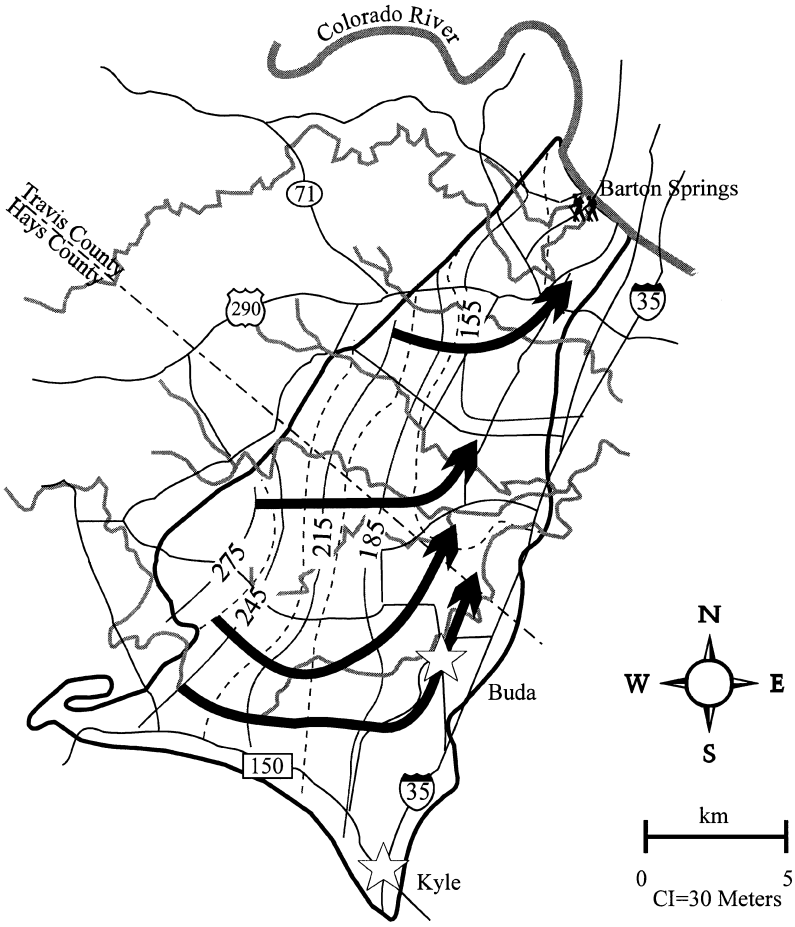

Fig. 2. Project study area (after Slade et al., 1985).

area and a typical potentiometric surface. The bold arrows indicate the general direction of water flow.

Numerous down-to-the-east normal faults with displacements as great as 60 m are present within the Barton Springs segment of the aquifer. Abundant caves, sinkholes, and enlarged fractures are evidence of the karstic nature of the aquifer. The general geology of the aquifer has been described in numerous reports including those by Brune and Duffin (1983), Rose (1972), Garner and Young (1976), Young (1977), Slade et al. (1986), and the Hauwert and Hanson (1995).

Flow in the aquifer moves from the west toward the east until the edge of the confined portion is reached where the flow moves generally northeast to discharge at Barton Springs. The western portion of the aquifer has the highest gradients and the least change in head from low to high flow conditions. Water levels fluctuate fewer than 3 m in this area. The eastern portion of the aquifer has the lowest gradients indicating extensive cavern development. Water levels in wells in the eastern portion of the aquifer are highly

correlated with each other and with flow at Barton Springs (Slade et al., 1986). The levels may vary as much as 27 m with changes in spring discharge.

Five main creeks supply most of the recharge to the Barton Springs portion of the Edwards. The watersheds of these creeks are divided into the contributing and recharge zones. The contributing zone consists of the portion of the watersheds of the creeks lying west of the aquifer and underlain by the Glen Rose Limestone. Development in this area will affect the volume and quality of baseflow and direct runoff which enters the creeks. To the east, the creeks flow over the outcrop of the Edwards Limestone where recharge to the aquifer occurs. This area is termed the recharge zone. Recharge in this area also occurs by direct infiltration of rainfall into the aquifer.

3. Literature review

The selection of the appropriate model to achieve the goals of estimating the impacts of urban development and other nonpoint sources of pollution is a major task. An appropriate conceptual model should be sufficiently simple so as to be amenable to mathematical treatment, but it should not be too simple so as to exclude those features which are of interest to the investigation at hand. The information should be available for calibrating the model, and the model should be the most economic one for solving the problem at hand (Bear, 1979).

To completely model a system requires a very detailed knowledge of the physical properties and the processes governing water movement. The virtue of a model rests in its ability to predict a general system from incomplete or partial data. The parsimonious model simplifies the representation of the physical structure or of the processes involved. This is especially appropriate in light of the extraordinary heterogeneity exhibited by karst aquifers.

Numerous types of models have been developed and used to predict water levels and spring discharge from karst aquifers. The simplest are black box models which contain no spatial information, but can predict spring discharge or other aquifer properties. Dreiss (1989) used time moment analysis to relate a time series of inputs (recharge) to a series of outputs (spring flow). Simple regression models also have been used to predict water levels in karst aquifers (Zaltsberg, 1984). The limitation of these types of models is that they lack predictive power.

Deterministic models for groundwater flow and transport may be physically based and may have either distributed or lumped parameters. Lumped models lack the spatial dimension in the equations describing flow and transport; consequently, only ordinary linear differential equations must be solved. These models offer the opportunity to simulate a given system with fewer data requirements for parameterization and calibration than their distributed counterparts. Lumped parameter models in groundwater applications generally have been single cell models such as those developed by Mercado (1976) and Gelhar and Wilson (1974). Karst aquifers also have been modeled as a series of linear reservoirs (Yurtsever and Payne, 1985).

A lumped parameter model for the San Antonio portion of the Edwards aquifer consisting of nine cells was described by Wanakule and Anaya (1993). Because of the large

cell size in their model relative to the number and distribution of conduits and other heterogeneities, they were able to represent the aquifer in each cell as a single equivalent porous medium. Lumped parameter models also have been described by Simpson (1988), who termed them discrete state compartment models. Campana and Mahin (1985) used the terminology to describe their 34-cell model of the southern Edwards which they employed to estimate the groundwater age distribution and aquifer properties.

Distributed parameter models are normally chosen to increase the accuracy of predictions or to achieve a high degree of spatial resolution. Several distributed parameter models using a single equivalent porous medium have been developed for the San Antonio portion of the Edwards (Maclay and Land, 1988; Thorkildson and McElhaney, 1992), although none perform better than the nine-cell model developed by Wanakule and Anaya. The most elaborate of these distributed parameter models was developed by Kuniansky and Holligan (1994) at the US Geological Survey (USGS). This is a finite element model of the Edwards/Trinity aquifer system containing over 7000 elements. Despite the high degree of spatial resolution, difficulties in generating input data have limited its usefulness.

Dual porosity distributed parameter models also have been developed for karst aquifers (Teutsch and Sauter, 1992). These models generally represent conduit and diffuse flow as separate systems linked by a transfer function. They have the advantage of being able to represent the fast transit and slow depletion often exhibited by karst aquifers, but at the cost of more than doubling the number of parameters required for calibration.

This review of karst aquifer models demonstrates the evolution in model complexity associated with attempts to increase the accuracy of predictions. The general tendency has been to increase the number of cells in the x–y plane while ignoring improvement which might be achieved by incorporating variation in the vertical direction. This approach has not been consistently successful. The more spatially detailed models have been difficult to calibrate and verify. In addition, input data must be developed for each cell; consequently, these models are not used to any great extent by regulatory agencies or other groups.

The goal of this modeling effort is the development of a model which is simple to calibrate and use, and yet achieves a high degree of accuracy. This modeling effort differs from preceding studies by retaining a simple spatial description of the aquifer, but allowing vertical variations in aquifer properties such as specific yield within cells. Since the variations are contained within the cell, not all the cells need have the same number of layers. Because water and solute movement within cells is not considered, the model retains the characteristic lack of a spatial dimension exhibited by lumped parameter models. This approach is appropriate for highly stratified aquifers under water table conditions. Caves and other solution features in the Edwards tend to develop at elevations near the water table (Kastning, 1983), so changes in water level may have a greater influence on storage and flow characteristics than do lateral changes.

4. Aquifer recharge and discharge

Development of a groundwater model requires the identification and quantification of all known sources of recharge and discharge. These inputs and outputs are related to the

state of the aquifer (i.e. head distribution) by a mathematical description of the aquifer. In the optimum case, this relationship is based on physical principles describing flow and storage in the aquifer and model parameters are derived from the aquifer properties. This section describes the quantification of the model inputs and the development of a parsimonious physically based description of the Barton Springs portion of the Edwards aquifer.

Water balance studies indicate that flow losses in the creeks crossing the Edwards outcrop are sufficient to supply all the known discharge at springs and well fields. Flow loss studies included manual gauging of stream segments in the recharge zone (Slade et al., 1986) and comparison of hourly flow records of gauging stations located upstream and downstream of the recharge zone on each creek. The locations of the watersheds of these creeks are shown in Fig. 3. All of the creeks except for Barton Creek exhibit similar recharge behavior. Below a threshold flow rate in each creek upstream of the recharge zone, all flow is lost to recharge. Once this threshold is exceeded, the recharge rate remains essentially constant despite the increases in water depth in the creek channel associated with higher flows. Fig. 4 demonstrates the relationship between flow and recharge for Onion Creek based on 2 years of hourly data. The relationship was developed by subtracting the daily average flow downstream of the recharge zone from the flow upstream of the recharge zone. The data were edited to exclude days when surface runoff to the creek between the two stations caused the flow downstream to be greater than the upstream flow. The edited data still contain numerous points showing less apparent recharge. This effect is the result of measured flow at the downstream station which was derived from perched water tables near the downstream edge of the recharge zone.

Computation of the rate of recharge from Barton Creek is more complicated than for the other creeks. The elevation of the bed of Barton Creek in its lower reaches is at about the

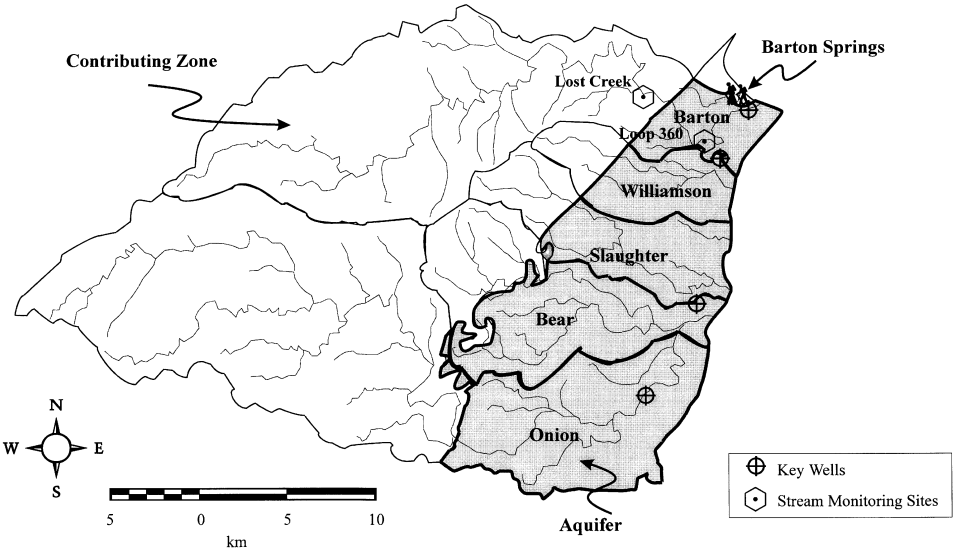

Fig. 3. Location of aquifer cells and key wells.

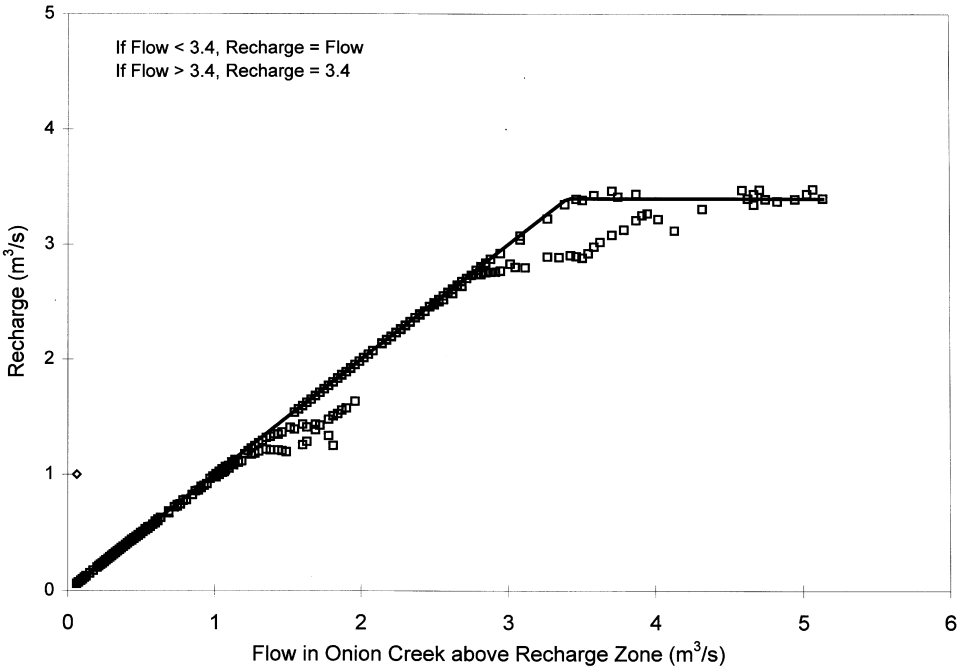

Fig. 4. Relationship between recharge and flow rate for Onion Creek.

same level as the average aquifer level in that location, so that segment of the Creek may either be gaining or losing water depending of the level of the aquifer. Other factors such as the location of recharge features, channel morphology, and geology may also affect the rate of recharge resulting from a given flow rate in the creek.

The rate of recharge in Barton Creek was estimated by comparing the flow rates above the recharge zone at the Lost Creek Boulevard monitoring station with the flow at Loop 360 during the period from 1989 through 1994 (monitoring locations shown in Fig. 3). The Loop 360 monitoring station is located approximately half way across the recharge zone. At this location the bed of the creek is always above the aquifer level, so that recharge above this point should not be significantly affected by aquifer level. For the purpose of this analysis, it was assumed that no recharge occurs in the reach between Loop 360 and Barton Springs Pool. The difference between the flow at Lost Creek Boulevard and Loop 360 equals the rate that recharge is occurring in that reach. The recharge rate is plotted against flow rate at the upstream station in Fig. 5. The open diamonds represent the difference in the two flow rates. Runoff to the creek between the two stations results in points which plot below the actual recharge rate for a given creek discharge. The solid points on the graph represents the recharge assigned to flows greater than 0.85 $m^3\ s^{-1}$.

The relationship between recharge and flow rate in the creek is similar to that exhibited by the other creek for flows of less than about 3.7 $m^3\ s^{-1}$. At higher flows, the rate of recharge increases dramatically. The highest recharge rate measured was 7 $m^3\ s^{-1}$, and this was assumed to be the maximum rate. The increase in the rate of recharge at higher flow

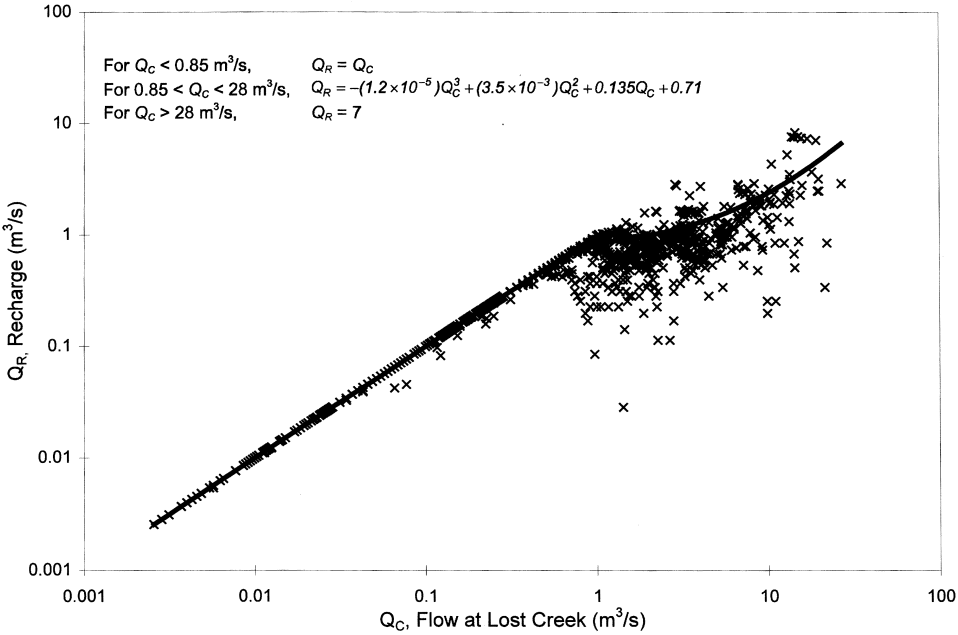

Fig. 5. Barton Creek recharge above Loop 360.

rates may be a function of channel morphology, differences in hydraulic conductivity between the base and banks of the channel, or scour during high flow rates which exposes recharge features in the bed of the creek. For the purposes of calculating recharge the following relationships were used:

For $Q_C < 0.85$ m^3 s^{-1}, $Q_R = Q_C$
For $0.85 < Q_C < 28$ m^3 s^{-1}, $Q_R = -(1.2 \times 10^{-5})Q_C^3 + (3.5 \times 10^{-3})Q_C^2 + 0.135\, Q_C + 0.71$
For $Q_C > 28$ m^3 s^{-1}, $Q_R = 7$

where Q_C is the flow rate at Lost Creek Boulevard (m^3 s^{-1}), and Q_R is the rate of recharge (m^3 s^{-1}).

Diffuse recharge was assumed to occur at a constant rate. This is a reasonable assumption when the thickness of the vadose zone is large. Throughout most of the Edwards recharge zone the water table lies more than 30 m below the land surface. The average rate of rainfall infiltration was estimated with the Groundwater Loading Effects of Agricultural Management Systems (GLEAMS) model developed by the US Department of Agriculture (Knisel, 1993). Using historical rainfall data from the period 1979–1993 and descriptions of the soil and vegetation types on the recharge zone, average infiltration was estimated to be about 50 mm year^{-1}, which is about 6% of the average annual precipitation. The daily infiltration was multiplied by the approximate surface area of the recharge zone over each cell to calculate the daily volume of infiltration.

The USGS has developed a rating curve to estimate discharge from Barton Springs based on the water level in well YD-58-42-903 which is located adjacent to the Springs.

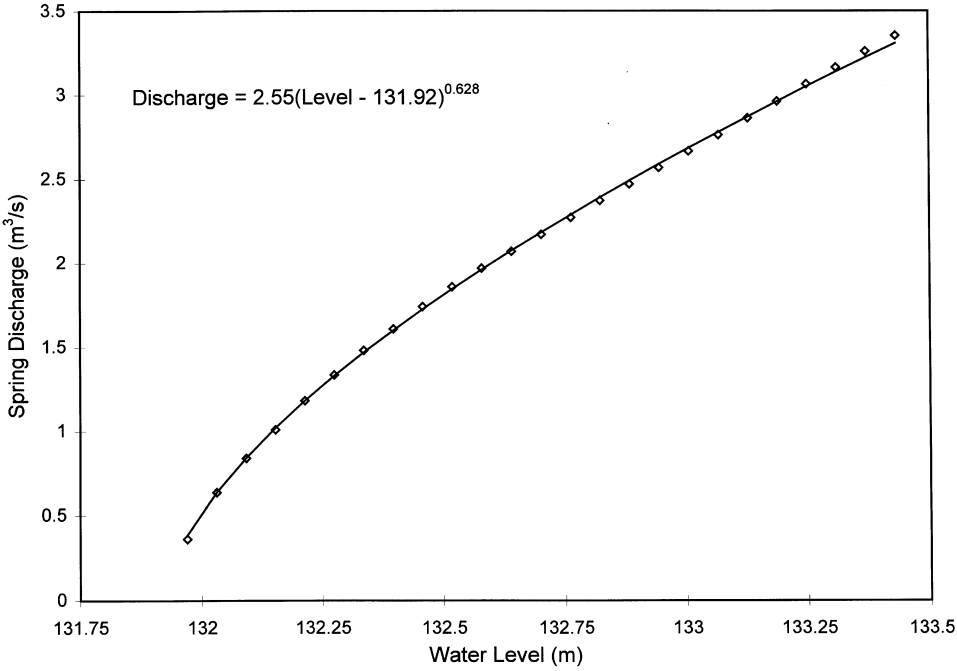

Fig. 6. Rating curve for Barton Springs.

Spring discharge in the model was calculated from the rating curve developed by the USGS. The rating curve and best fit line through the measured points are shown in Fig. 6. The equation for the line shown on the graph is given by:

$$Q = 2.55(x - 131.92)^{0.628}$$

where Q is Barton Springs discharge ($m^3\ s^{-1}$), and x is the water level in Well 58-42-903 (m). This function has a form similar to that which describes discharge from a tank through a submerged orifice. The only difference is in the value of the exponent which would be equal to 0.5 for orifice flow. The value 131.92 is the water surface elevation above mean sea level of Barton Springs Pool.

Wells penetrating the aquifer supply drinking water to approximately 30 000 residents of northern Hays and southern Travis Counties. The pumpage data collected by the Barton Springs/Edwards Aquifer Conservation District were analyzed to determine the location and volumes of the water supply wells. The data from 1994 were the most complete and were used for each year of the simulation. The average rate of pumpage was equal to about 0.14 $m^3\ s^{-1}$, which is 10% of the long-term average discharge from Barton Springs and equivalent to about 375 l day^{-1} per capita. Since the time step used in the computer simulation was 1 day, the monthly data were converted to average daily pumping rates and subtracted from the appropriate cell during each time step.

Discharge from the aquifer also occurs in the segment of Barton Creek between Loop 360 and Barton Springs (shown in Fig. 3) during periods of high aquifer water levels. The

volume and rates of discharge are unknown. Since the recharge from the creeks was larger than the discharge from Barton Springs and known well fields, the rate of discharge to the Creek was used as a calibration parameter to improve the spring flow prediction.

5. Model formulation and calibration

The model developed in this study is similar to that developed by Wanakule and Anaya (1993), in that relatively few cells are used to describe the aquifer which simplified calibration of the model. The response of individual wells to recharge events supports the validity of large cell sizes. Wells located miles from creeks providing recharge to the aquifer exhibit rapid increases in water levels simultaneously with wells near the creek beds. The Barton Springs portion of the Edwards aquifer receives the bulk of its recharge from the five main creeks which cross the recharge zone. These creeks are fairly evenly spaced which suggested the use of a five-cell model to predict the behavior of the aquifer.

A single well was chosen in each cell to represent conditions in that portion of the aquifer. The well chosen to represent the conditions in the Barton Creek cell (YD-58-42-903) is located adjacent to Barton Springs and is used by the USGS to estimate spring discharge. The wells chosen in the other cells are located along the eastern portion of the aquifer. This is the area that experiences the maximum range of groundwater elevations (up to 27 m). Each cell is treated as a tank which is assigned an effective area (equivalent to the product of specific yield and surface area). At the present time, there is no well appropriately located in the Slaughter cell with sufficient measurements to calibrate against. The locations of the cells and key wells used in the study are shown in Fig. 3.

There are a number of significant differences between this model and previous karst models. Rather than increasing the number of cells to obtain better simulated results, model predictions were improved by allowing vertical variation of aquifer properties within cells. In particular, specific yield and hydraulic conductivity of the cells are functions of elevation. A short time step (daily) was used in the model which facilitated the calculation of recharge, increased the accuracy of the model and allowed the governing equations to be solved explicitly. A schematic diagram of the model is shown in Fig. 7.

The model describes flow between the cells using Darcy's Law. The hydraulic conductivity was assigned to the boundaries between cells which was the method employed by Prickett and Lonnquist (1971), and the saturated thickness of the upstream cell was used to calculate the transmissivity. All external model boundaries were treated as no-flow boundaries, so there are only four boundaries where flow occurs. Flow rate across each internal boundary was calculated as:

$$Q_G = Kwb\left(\frac{\Delta h}{l}\right)$$

where Q_G is the groundwater flow rate across the boundary, w is the width of the boundary, Δh is the head difference across the boundary, b is the saturated thickness of the upstream cell, and l is the distance between the key wells in each cell. This was simplified in the

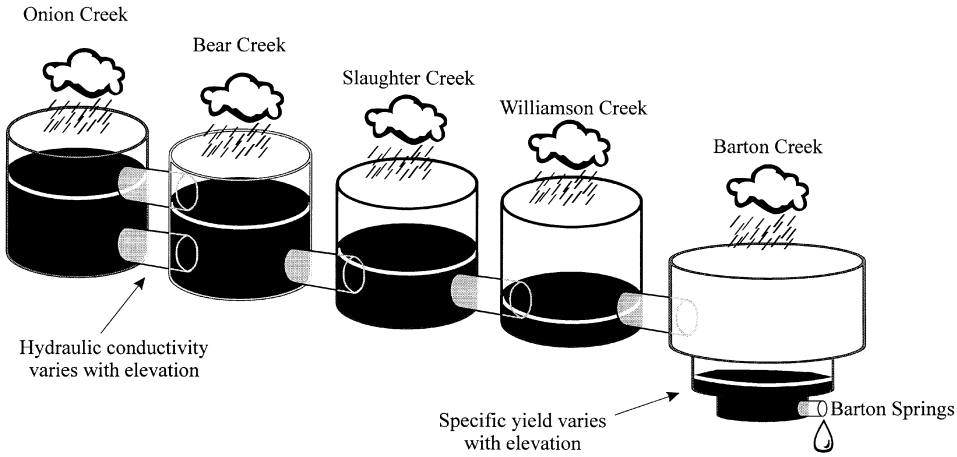

Fig. 7. Schematic diagram of the aquifer model.

model to:

$$Q_G = K'b\Delta h$$

where

$$K' = \frac{Kw}{l}$$

A reasonably good prediction of water levels in each of the cells could be obtained using this formulation; however, the model consistently over-predicted the water levels in the Onion cell during periods of peak recharge and high aquifer levels. The slope of the predicted recession curve closely matched the observed recession indicating that the parameter value describing cell storage was fairly accurate. By increasing the hydraulic conductivity of the cell boundary as the water level increased, water moved out of the cell at a faster rate, and the slope of the recession was largely unaffected. This change also resulted in better water level predictions in the adjoining cell. A simple two-layer representation of the hydraulic conductivity was sufficient to reproduce the measured water level fluctuations. A process of trial and error led to the choice of 183 m above mean sea level in the key well as the boundary between the zones of different conductivity. The following equation was used to describe flow between the Onion and Bear Creek cells when the water level exceeded this threshold value:

$$Q_G = K'_l(106)(\Delta h) + K'_u(h - 138)(\Delta h)$$

where K'_l and K'_u are the flow proportionality constants for the lower and upper sections, 106 m is the distance between the base of the cell and an elevation 183 m, h is the head in the Onion cell, and Δh is the difference in water level elevations between the Onion and Bear cells.

Since each aquifer cell is treated as a tank, a parameter is required to relate fluctuations in water surface elevation to changes in the amount of water in the cell. This parameter is described as the effective area of the tank and is physically equivalent to the product of the

average specific yield and surface area of the tank. The effective area of each cell was chosen to reproduce the spring flow recession and associated drop in aquifer water levels which occurred between August 1979 and January 1980. This is the same period chosen by Slade et al. (1985) for the calibration of their model.

The relationship between water level at the beginning of each time step and water volume is described by:

$$h = \frac{V}{A} + \zeta$$

where h is the water surface elevation of the cell, V is the volume of water in the in the cell, A is the effective area of the cell, and ζ is the elevation of the base of the cell above mean sea level.

If each of the cells had a specific yield independent of elevation, one would expect that the spring flow recession would be more rapid at the beginning. The data clearly demonstrate that the recession is not as rapid when the discharge from Barton Springs is greater than about 2.1 $m^3\ s^{-1}$. Several configurations were tested to reproduce this behavior. The most successful was the division of the Barton Creek cell into three zones. The effective area was assumed to take the form of a step function, assuming three discrete values. The elevations where these values change were estimated during the calibration process and have the physical representation of geologic layers with different specific yields.

The basal zone was defined to include the interval from the bottom of the aquifer (estimated to be about 106 m above mean sea level) to a measured elevation of 133.3 m in the representative well for that cell (which corresponds to a Barton Spring discharge of 2.1 $m^3\ s^{-1}$). This interval was assigned a smaller effective area which produced a more rapid spring flow recession.

A third zone in the Barton cell was defined based on the aquifer response to recharge when Barton Springs is discharging at fairly high flow rates. When Spring discharge exceeds about 3.1 $m^3\ s^{-1}$, additional recharge causes very little increase in water level in the well representing the Barton cell. To reproduce this behavior, a third zone with a higher effective area was required. The base of this zone was determined to be at an elevation of 132.4 m in the key well (corresponding to a spring discharge of 3.1 $m^3\ s^{-1}$).

The water volume which could be contained in the two lower sections when full was calculated. Since the volume in the cell is known, the layer containing the water surface is also known and the elevation can be calculated as:

$$h = \frac{V_t - V_{i-1}}{A_i} + \zeta$$

where h is the water surface elevation in the Barton cell, V_t is the total water volume in the cell, V_{i-1} is the total volume of all layers below the layer containing the water surface elevation, A_i is the effective area of the cell layer containing the water surface, and ζ_i is the elevation of the base of the cell layer containing the water surface.

During periods of high water levels in the aquifer, numerous ungauged springs supply baseflow to the section of Barton Creek between Loop 360 and Barton Springs. In the model, no baseflow was assumed to occur when aquifer levels were below that necessary to produce some arbitrary rate of discharge from Barton Springs. A minimum discharge

from Barton Springs of at least 2.25 $m^3\ s^{-1}$ for baseflow to occur resulted in the best calibration. When the predicted flow exceeded that rate, the following equation was used to estimate discharge to the Creek:

$$Q_B = 0.6 \times (Q_S - 2.25)$$

where Q_B is the baseflow discharge to Barton Creek and Q_S is the predicted discharge from Barton Springs.

The model calculates aquifer state based on a daily mass balance for each cell. For the purposes of calculating diffuse recharge volumes, the surface area of each cell is assumed to conform to the boundaries of the surface watershed of the creek supplying recharge to that portion of the aquifer. Because of the relatively short time step, the integration is done explicitly using Euler's method. The volume of each cell at the end of each time step except for Barton is calculate by the following formula:

$$V_{t+\Delta t} = V_t + S \times (q_i) \times \Delta t - P_{t+\Delta t} + \Delta t \times \sum Q_G|_{t+\Delta t}$$

where $V_{t+\Delta t}$ is the volume of water in the cell at the end of the next time step, V_t is the volume at the end of the preceding time step, Δt is the length of the time step, $\sum Q_G$ is the net groundwater flow rate into the cell from adjacent cells, $S(q_i)$ is the surface area of the cell times the rainfall infiltration rate, and $P_{t+\Delta t}$ is the volume pumped from the cell during the time step.

The mass balance for the Barton cell is calculated in a similar manner to that of the other cells except that terms expressing the volume of discharge at Barton Springs and baseflow to Barton Creek are included. The following equation is solved at every time step:

$$V_{t+\Delta t} = V_t + \Delta t \times (Q_W - Q_B - Q_S)|_{t+\Delta t} + S \times (q_i) \times \Delta t + P_{t+\Delta t}$$

where Q_W is the flow from the Williamson cell, Q_B which is the rate of baseflow discharge to the creek, and Q_S is the rate of Barton Spring discharge.

6. Model calibration

Accuracy of the model was judged using several criteria. One of the primary tests was the accuracy of model predictions for both spring discharge and water surface elevation during the Fall of 1979. During this period, Barton Spring discharge and numerous water level measurements for a number of the key wells used in the model are available. To determine aquifer properties during periods of extreme water levels, data from the period 1989 through 1994 were used. The most important criteria used to judge model accuracy during this period were spring discharge and water surface elevations in the cells most distant from the springs. The key wells in these cells had daily water level measurements for much of this period. The best fit was determined by comparing the sum of the squared error for water level and spring flow. Spring discharge alone proved to be a very poor predictor of overall model performance.

The measured and predicted discharge from Barton Springs for the calibration period is shown in Fig. 8. No measured discharge is available for the period from December 1991 through July 1992 because of a large flood which resulted in Barton Springs Pool being

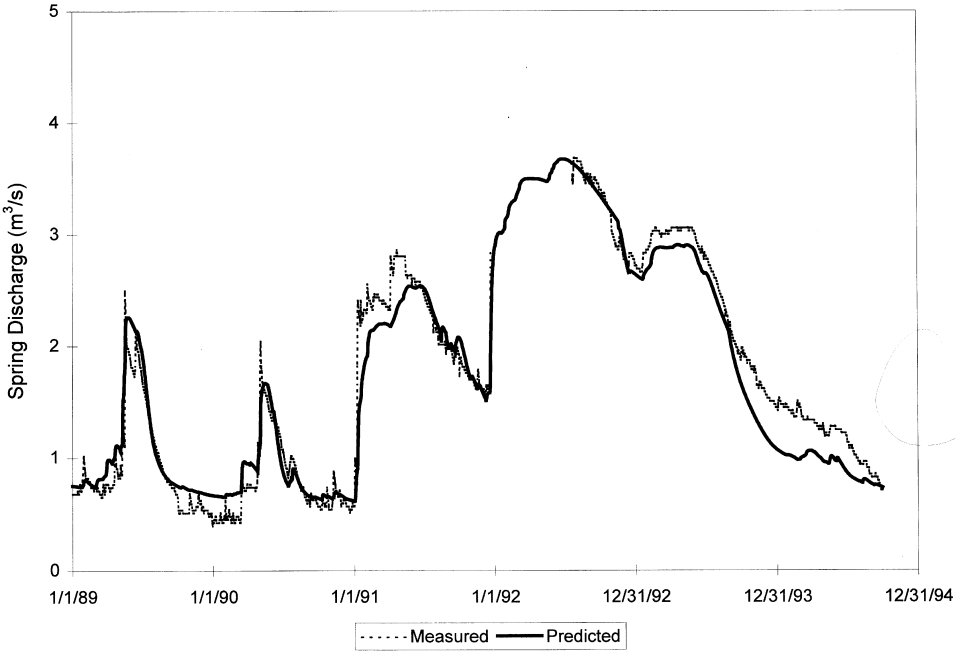

Fig. 8. Comparison of measured and predicted discharge for calibration period.

drained for repairs. The average flow predicted by the model during this time period was 1.61 $m^3\ s^{-1}$, while the observed flow was 1.68 $m^3\ s^{-1}$. The root mean squared error for the predicted values was 0.22 $m^3\ s^{-1}$. Since the prediction of Barton Springs discharge is based on the water level in the well used by the USGS for flow estimation, the figure also indicates the accuracy with which water level in that portion of the aquifer is predicted.

The representative well for the Onion Creek cell has numerous recorded water level measurements during the simulation period. In addition, in 1991, the Barton Springs/ Edwards Aquifer Conservation District installed monitoring equipment on that well to record daily water level measurements. A comparison with reported daily water levels in the Onion Creek cell for the period 5/91–9/95 is shown in Fig. 9. The average observed water level was 186.3 m, compared with the average predicted value of 185.6 m. The root mean squared error of the prediction was 4.0 m.

The key well for the Bear Creek cell also has daily water level measurements for much of this period and a comparison of measured with predicted values is shown in Fig. 10. The average measured level is 175.4 m, while the predicted level is 176.7. The root mean squared error for predictions in this cell, 7.8 m, is larger than for the Onion Creek cell. This is the result of water level changes resulting from the proximity of a local water supply well which are not reflected in the model prediction of the regional water level.

Daily water level measurements are now being made in the key well for the Williamson

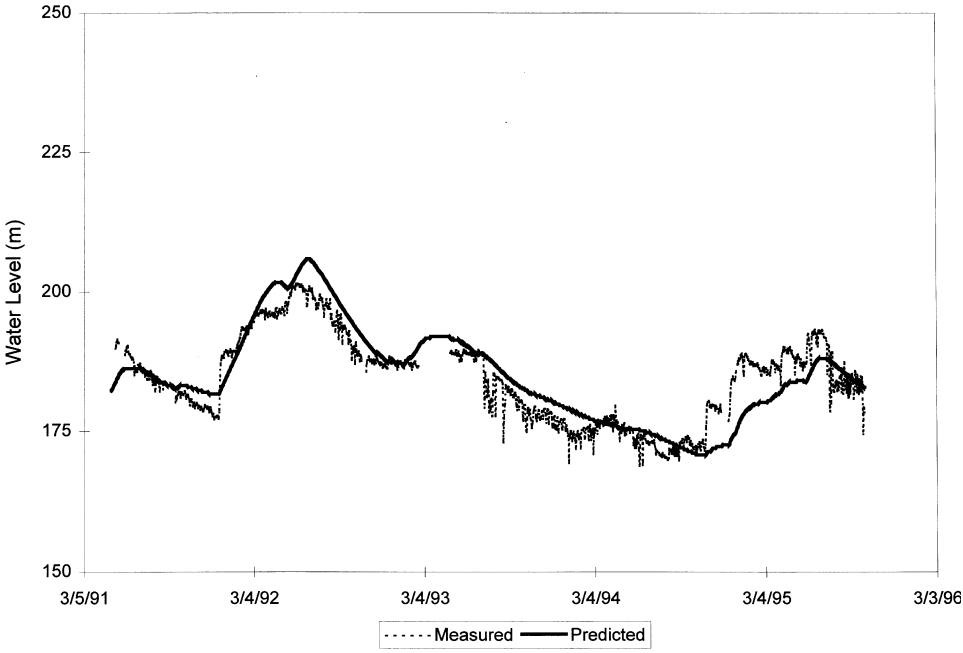

Fig. 9. Comparison of daily water levels in Onion Creek cell.

Creek cell; however, the period of record is so short that no meaningful comparison between measured and predicted values can be made for these date. Numerous discrete water level measurements were made in this well between 1979 and 1989 and a comparison of measured with predicted values is shown for this period in Fig. 11. There is currently no well appropriately located in the Slaughter Creek cell to provide water level measurements for comparison with model predictions.

The final calibration parameters for each cell are shown in Table 1 and Table 2. The choice of the representative well in each cell strongly affects the degree to which the parameters represent a measurable physical property. For instance, the well chosen for the Barton cell is located adjacent to Barton Springs. The proximity to the springs means that the range of water elevations recorded in the well is small (about 1.5 m) compared with the range measured in other parts of the aquifer being represented by the cell.

The small increases in water level resulting from large volumes of recharge mean that the effective area of the cell must be large and, consequently, the apparent specific yield is extremely high. Conversely, the wells chosen to represent conditions in each of the other cells were located in the eastern portion of the aquifer, where the range of recorded water levels is fairly large (up to 27 m). Therefore, a relatively small effective area produced the large changes in observed water level. If the effective area of these cells is divided by the surface area of the corresponding watersheds, the apparent specific yield is very low. The properties of the boundaries between cells are shown in Table 2. The parameter labeled ‘‘Flow length’’ is the distance between the key wells in each cell and is used to calculate the hydraulic gradient between cells.

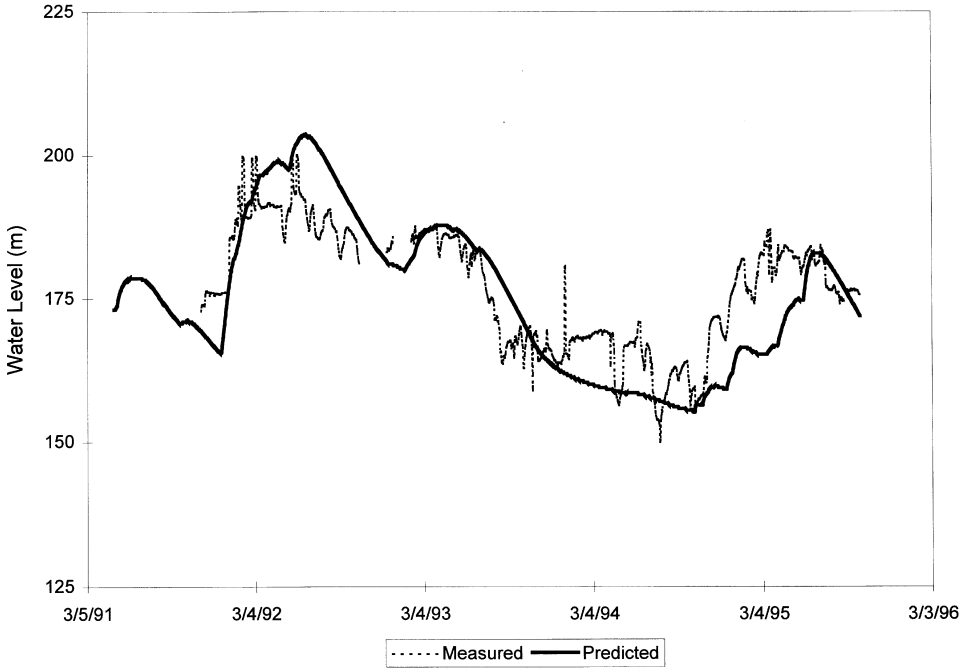

Fig. 10. Comparison of daily water levels in Bear Creek cell.

7. Conclusions

This study developed a new type of lumped parameter model for the Barton Springs portion of the Edwards aquifer. The model is capable of predicting regional water levels and discharge. A comparison of model predictions with historical data for the period August 1979–September 1995 demonstrates its accuracy. This simple representation of the hydrologic system produced results comparable to those of traditional groundwater models with fewer data requirements and calibration parameters.

This conceptual model of the aquifer appears to be successful because the majority of the Barton Springs portion of the Edwards is unconfined. Because of the horizontal stratification of the formation, vertical changes in aquifer properties have a greater influence on aquifer behavior than does horizontal variation. As water levels rise, caves, conduits, and other stratigraphic features which become submerged strongly affect flow and storage in the aquifer. The wide range of water levels which occur in this aquifer appear to amplify these differences in flow and storage characteristics.

When faced with the task of modeling an extremely complex flow system, the natural tendency is to develop a more complex model. However, this research shows that a very simple model can provide useful information about the behavior of such a system. In addition the model explicitly acknowledges the lack of detailed knowledge about the location of conduits and other flow paths by predicting only regional effects. While predictions made by more complex models are often given more validity by persons

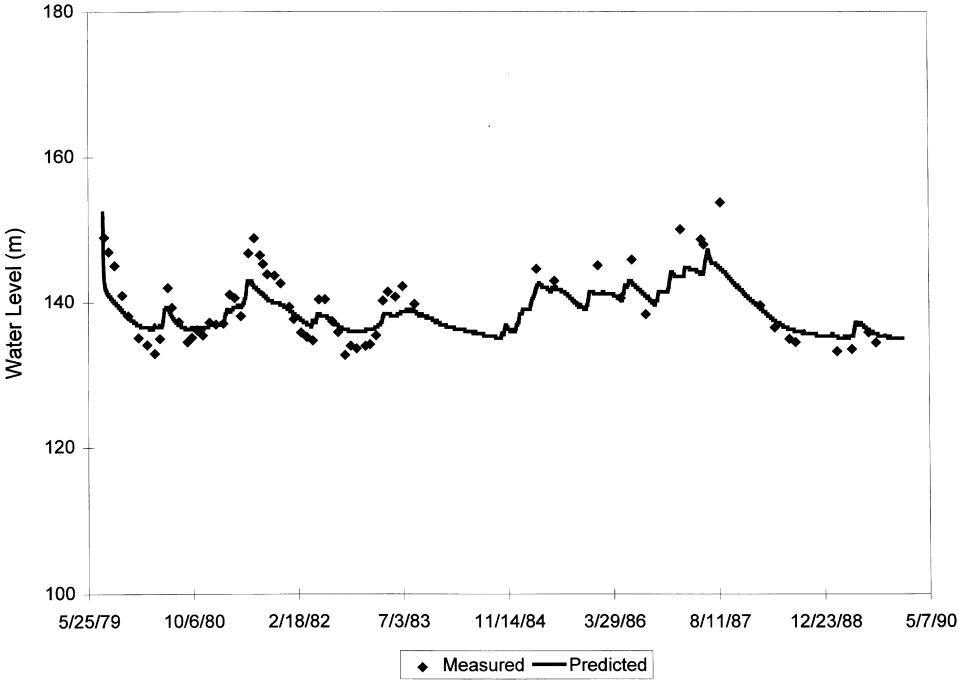

Fig. 11. Comparison of water levels in Williamson Creek cell.

unfamiliar with their use or development than might be warranted (especially true when the values of physical parameters such as specific yield or hydraulic conductivity may have been estimated from a sparse data set) this parsimonious model provides a useful management tool that is easy to use and understand, and whose predictions are not as subject to misinterpretation as those of a complex distributed parameter model.

This groundwater model, when used in conjunction with a surface water model, will allow a prediction of the hydrologic impact of the increase in runoff coefficient resulting from changes in land use patterns. The effect of potential runoff control structures on recharge and water levels also can be evaluated. Work is continuing on incorporating a

Table 1
Characteristics of the five aquifer cells

Cell	Interval (m above m.s.l.)	Effective area (m^2)	Actual area (m^2)	Specific yield (%)
Onion	76 and above	1180	137960	0.9
Bear	76 and above	220000	98100000	0.2
Slaughter	76 and above	130000	70420000	0.2
Williamson	61 and above	110000	351750000	0.2
Barton	106.7–132.7	2900000	343200000	6.7
	132.7–133.3	7430000		17.2
	133.3 and above	37160000		86.0

Table 2
Properties of the boundaries between cells

Cell boundary	Boundary width (m)	Flow length (m)	Hydraulic conductivity (m day^{-1})
Onion/Bear (76–183)	16100	7300	12
Onion/Bear (>183)	16100	7300	250
Bear/Slaughter	12900	5200	35
Slaughter/Williamson	10600	5200	37
Williamson/Barton	6400	3900	158

transport component into the model so that the impact of changes in land use on water quality in the aquifer and at Barton Springs can be estimated.

Acknowledgements

Numerous graduate students in The Department of Civil Engineering at The University of Texas in Austin, including Aaron Kam, James Kearley, and John Meadows, helped collect and analyze the data used to develop the model. J. Tyler Lehman was responsible for creating the graphics. This research was funded under a contract with the City of Austin, Texas.

References

Bear, J., 1979. Hydraulics of Groundwater. McGraw-Hill Publishing, New York, 567 pp.

Brune, G. and Duffin, G.L., 1983. Occurrence, Availability and Quality of Ground Water in Travis County, Texas. Texas Dept. of Water Resour. Rep. 276, 219 pp.

Campana, M.E. and Mahin, D.A., 1985. Model-derived estimates of groundwater mean ages, recharge rates, effective porosities and storage in a limestone aquifer. J. Hydrol., 76: 247–264.

Dreiss, S.J., 1989. Regional scale transport in a karst aquifer, 2. Linear systems and time moment analysis. Water Resour. Res., 25: 126–134.

Garner, L.E. and Young, K.P., 1976. Environmental geology of the Austin area: an aid to urban planning. University of Texas, Austin, TX, Bur. Econ. Geol., Rep. of Invest. 86, 39 pp.

Gelhar, L.W. and Wilson, J.L., 1974. Ground-water quality modeling. Ground Water, 12(6): 399–408.

Hauwert, N.M. and Hanson, J.A. (Editors), 1995. A Look at the Hydrostratigraphic Members of the Edwards Aquifer in Travis and Hays Counties, Texas. Austin Geological Society, 81 pp.

Kastning, E.H., 1983. Relict caves as evidence of landscape and aquifer evolution in a deeply dissected carbonate terrain: southwest Edwards Plateau, Texas, U.S.A. J. Hydrol., 61: 89–112.

Knisel, W.G., 1993. GLEAMS: Groundwater Loading Effects of Agricultural Management Systems. University of Georgia, Coastal Plain Experiment Station, Biological and Agricultural Engineering Department, Publication No. 5, 260 pp.

Kuniansky, E.L. and Holligan, K.Q., 1994. Simulation of flow in the Edwards–Trinity aquifer system and contiguous hydraulically connected units, West-Central, Texas. US Geological Survey Water-Resources Investigations Report 93-4039.

Maclay, R.W. and Land, L.F., 1988. Simulation of flow in the Edwards aquifer, San Antonio region, Texas, and refinement of storage and flow concepts. US Geol. Surv. Water Supply Paper 2336-A, 48 pp.

Mercado, A., 1976. Nitrate and chloride pollution of aquifers: a regional study with the aid of a single-cell model. Water Resour. Res., 12(4): 731–747.

Prickett, T.A. and Lonnquist, C.G., 1971. Selected digital computer techniques for groundwater resource evaluation. Illinois State Water Surv. Bull., 55: 1–62.

Rose, P.R., 1972. Edwards Group, Surface and Subsurface, Central Texas. University of Texas, Austin, TX, Bur. Econ. Geol., Rep. Invest. 74, 198 pp.

Simpson, E.S., 1988. The discrete state compartment model and its applications to flow through karstic aquifers. In: Karst Hydrogeology and Karst Environment Protection: In, Y. Daoxian (Editor), Proceedings of the 21st Congress of the International Association of Hydrogeologists, 10–15 October 1988, Guilin, China. Vol. 2, IAHS Press, pp. 671–676, Dublin, OH, 1992.

Slade, R.M., Jr., Ruiz, L.M. and Slagle, D., 1985. Simulation of the flow system of Barton Springs and associated Edwards aquifer in the Austin area, Texas. US Geol. Surv., Water Resour. Invest. Rep. 85-4299, 49 pp.

Slade, R.M., Jr., Dorsey, M.E. and Stewart, S.L., 1986. Hydrology and water quality of the Edwards aquifer associated with Barton Springs in the Austin, Texas, area. US Geol. Surv. Water Resour. Invest. Rep. 86-4036, 96 pp.

Teutsch, G. and Sauter, M., 1992. Groundwater modeling in karst terranes: scale effects, data acquisition, and field validation. In, J. Quinlan (Editor), Proceedings of the Third Conference on Hydrogeology, Ecology, Monitoring, and Management of Ground Water in Karst Terranes, 4–6 December 1991, Nashville, TN. pp. 17–34, National Groundwater Association.

Thorkildson, D. and McElhaney, P., 1992. Model refinement and applications for the Edwards aquifer (Balcones Fault Zone) in the San Antonio region, Texas. Texas Water Dev. Brd. Rep. 340, 31 pp.

Wanakule, N. and Anaya, R., 1993. A lumped parameter model for the Edwards aquifer. Texas Water Resour. Inst. Rep. TR-163, Texas A&M University, College Station, TX, 84 pp.

Young, K.P., 1977. Guidebook to the Geology of Travis County, Texas. University of Texas, Student Geological Society, 171 pp.

Yurtsever, Y. and Payne, B.R., 1985. Time-variant linear compartmental model approach to study flow dynamics of a karstic groundwater system by the aid of environmental tritium (a case study of south-eastern karst area in Turkey). In: G.A. Gultekin and Ivan Johnson (Editors), Karst Water Resources, Proceedings of the Ankara-Antalya Symposium, July 1985. pp. 545–561 IAHS, Washington, DC.

Zaltsberg, E.A., 1984. Forecast of karst water levels. In: A. Burger and L. Dubertret (Editors), Hydrogeology of Karstic Terraines. International Contributions to Hydrogeology, Heise, Hannover, pp. 40–42.