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Can we simulate regional groundwater flow in a karst system using equivalent porous media models? Case study, Barton Springs Edwards aquifer, USA

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Abstract

Various approaches can be used to simulate groundwater flow in karst systems, including equivalent porous media distributed parameter, lumped parameter, and dual porosity approaches, as well as discrete fracture or conduit approaches. The purpose of this study was to evaluate two different equivalent porous media approaches: lumped and distributed parameter, for simulating regional groundwater flow in a karst aquifer and to evaluate the adequacy of these approaches. The models were applied to the Barton Springs Edwards aquifer, Texas. Unique aspects of this study include availability of detailed information on recharge from stream-loss studies and on synoptic water levels, long-term continuous water level monitoring in wells throughout the aquifer, and spring discharge data to compare with simulation results. The MODFLOW code was used for the distributed parameter model. Estimation of hydraulic conductivity distribution was optimized by using a combination of trial and error and automated inverse methods. The lumped parameter model consists of five cells representing each of the watersheds contributing recharge to the aquifer. Transient simulations were conducted using both distributed and lumped parameter models for a 10-yr period (1989–1998). Both distributed and lumped parameter models fairly accurately simulated the temporal variability in spring discharge; therefore, if the objective of the model is to simulate spring discharge, either distributed or lumped parameter approaches can be used. The distributed parameter model generally reproduced the potentiometric surface at different times. The impact of the amount of pumping on a regional scale on spring discharge can be evaluated using a lumped parameter model; however, more detailed evaluation of the effect of pumping on groundwater levels and spring discharge requires a distributed parameter modeling approach. Sensitivity analyses indicated that spring discharge was much more sensitive to variations in recharge than pumpage, indicating that aquifer management should consider enhanced recharge, in addition to conservation measures, to maintain spring flow. This study shows the ability of equivalent porous media models to simulate regional groundwater flow in a highly karstified aquifer, which is important for water resources and groundwater management.

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1. Introduction

Numerical groundwater models are one of the most important predictive tools available for managing water resources in aquifers. These models can be used to test or refine different conceptual models, estimate hydraulic parameters, and, most importantly for water-resource management, predict how the aquifer might respond to changes in pumping and climate. Numerous models have been successfully developed in clastic aquifers; however, application of numerical models in karst aquifers is more problematic. Karst aquifers are generally highly heterogeneous. They are dominated by secondary (fracture) or tertiary (conduit) porosity and may exhibit hierarchical permeability structure or flow paths. These aquifers are likely to have a turbulent flow component, which may be problematic in that most numerical models are based on Darcy's law, which assumes laminar flow.

However, even with the above limitations, useful numerical flow models can be developed in karst aquifers, as long as their limitations are appreciated and respected. Quinlan et al. (1996) stated that: *"Although modeling of karstic processes is often possible and numerical flow models can sometimes simulate hydraulic heads, ground-water fluxes, and spring discharge, they often fail to correctly predict such fundamental information as flow direction, destination, and velocity."* Therefore, when discussing the relevance of numerical modeling in a karst aquifer, it is important to identify what type of model is being proposed: a flow model (hydraulic heads, groundwater fluxes, spring discharge) or a transport model (flow direction, destination, velocity). Many papers critical of numerical modeling (Huntoon, 1995; Quinlan et al., 1996) are specifically concerned about using numerical models to predict the direction and rate of solute transport, which are difficult to estimate a priori in even simple fractured systems.

It is not surprising that transport models do not perform well in karst aquifers, especially at local scales. Accurate transport predictions require in-depth knowledge of the distribution of the subsurface fracture and conduit systems. Transport of solutes in fractured rocks is an active research area (Bear et al., 1993; National Research Council, 1996). Often acknowledged in these studies is the difficulty of predicting, a priori, the direction and rate of solute transport through a fractured aquifer.

In fractured systems, as in karst systems, the concept of a representative elementary volume is used where size of the area of interest, or the cell in a model, becomes large enough to approximate equivalent porous media (Pankow et al., 1986; Neuman, 1987). Although accurate simulation of transport processes is still problematic, one may be able to model hydraulic heads, flow volumetrics, and general flow directions as supported by the characterization of the aquifer. Regional-scale models are much more likely to be successful than intermediate- or local-scale models (Huntoon, 1995).

It is important to consider the various modeling approaches available for simulating groundwater flow and contaminant transport in karst aquifers and to be aware of the advantages and limitations of each approach. One of the simplest approaches is the lumped parameter model, which has also been termed *black box* or *mixing cell model*. The spatial dimension in the equations is omitted in these models; therefore, only ordinary linear differential equations must be solved. The system is assumed to behave like an equivalent porous medium. These models generally result in good agreement between measured and simulated spring discharge (Yurtsever and Payne, 1986; Wanakule and Anaya, 1993; Barrett and Charbeneau, 1996; Zhang et al., 1996). The advantages of using lumped parameter models are that data requirements are minimal and simulations are rapid. The main disadvantage is lack of information on spatial variability in hydraulic head and directions and rates of groundwater flow.

Distributed parameter models are required to obtain detailed information on spatial variability in groundwater flow. Equivalent porous media distributed parameter models include single continuum and double continuum approaches. In many aquifers, single continuum models have proved adequate for simulating regional groundwater flow (Ryder, 1985; Kuniansky, 1993; Teutsch, 1993; Angelini and Dragoni, 1997; Keeler and Zhang, 1997; Greene et al., 1999 and Larocque et al., 1999). However, other studies have found the single continuum approach inadequate for simulating regional flow in highly karstified aquifers (Teutsch, 1993; Keeler and Zhang, 1997). An equivalent porous medium, dual continuum approach was used by Teutsch (1993) to model moderately to highly karstified systems. One continuum represents moderately karstified aquifer zones (diffuse flow, low conductivity, high storativity), and the other continuum represents highly karstified zones (high conductivity, low storativity). This double continuum approach is similar to the double porosity approach described in other studies, such as those in the oil industry (Warren and Root, 1963). Exchange or cross flow between the two continua is described by equations with exchange coefficients. Flow in both continua is assumed to be laminar. The double continuum model was used to simulate spring discharge, groundwater level fluctuations, and some tracer breakthrough curves in a karst system in southern Germany (Teutsch, 1993). This approach has also been used to delineate hypothetical well-head protection zones in a karst system in Florida (Knochemus and Robinson, 1996). In addition, the discrete fracture approach has been proposed for simulating flow in highly karstified systems; however, information on the location, geometry, and hydraulic properties of each fracture (deterministic models) or on the statistical attributes of fractures (stochastic models) is required. These large data requirements generally restrict the use of the discrete fracture approach to modeling local systems. Some studies have focused primarily on modeling of the conduit system (Halihan and Wicks, 1998; Jeannin, 2001).

A variety of factors are involved in choosing a suitable approach for simulating groundwater flow and transport in a karst aquifer. Teutsch and Sauter (1998) described four problem cases and appropriate modeling approaches—point and integral (or diffuse) source input and point (e.g. well) and integral (spring) observation output. Additional factors that are important in choosing a particular modeling approach include (1) degree of karstification, (2) objective of the modeling study, (3) availability of different types of data, and (4) availability of codes. (1) Karst aquifers have been classified as predominantly diffuse or conduit types, with a continuum between these end members. Equivalent porous media distributed parameter models have generally been applied to aquifers characterized as predominantly diffuse (Teutsch, 1993). Similar modeling approaches may also be appropriate for conduit systems if the conduits are fairly uniformly distributed and well interconnected. Some hydrologists suggest that carbonate aquifers have fairly similar structures and that the diffuse and conduit classification scheme simply reflects the bias in researchers' studies; i.e. those that focus on wells classify aquifers as predominantly diffuse, whereas those that focus on springs characterize the aquifers as conduit types (Davies et al., 1992). (2) The objective of the model is critical in determining an appropriate modeling approach. Examples of model objectives include evaluation of regional groundwater flow for water management, analysis of contaminant transport from point and nonpoint sources, and assessment of aquifer vulnerability to contamination. Equivalent porous media distributed and lumped parameter models have been used to simulate regional groundwater flow and non-point source contaminant transport (Barrett and Charbeneau, 1996). Double continuum equivalent porous media models have been used to simulate tracer transport and can potentially be used to simulate contaminant transport. Such models require detailed information on the location of subsurface conduits, which is often difficult to obtain. (3) Data availability may also constrain the type of modeling approach that can be used. In some cases there is information only on spring discharge (Angelini and Dragoni, 1997), whereas other sites may have detailed information on the flow system, including spring discharge, synoptic water levels, time series of water levels, tracer results for conduit delineation and flow velocities, spring temperature, and chemistry (Teutsch and Sauter, 1991). (4) Availability of codes may also affect choice of modeling approach. Although numerical codes for simulating karst genesis generally incorporate turbulent flow, it is generally not included in codes that simulate groundwater flow in karst systems on a regional scale.

The objective of this study was to evaluate the ability of equivalent porous media distributed and lumped parameter models to simulate regional groundwater flow in a karst aquifer and to assess the advantages and limitations of each approach. Different techniques for parameterizing the distributed model were evaluated, including trial and error and automated inverse methods. The models were applied to the Barton Springs segment of the Edwards aquifer, Texas. The Edwards aquifer has been hydrologically divided into three segments, the northern segment (north of the Colorado River), the central or Barton Springs segment, and the southern or San Antonio segment. The Barton Springs segment refers to the portion of the aquifer that discharges primarily to Barton Springs and is south of the Colorado River and north of a groundwater divide near Kyle, Texas. The primary management issue for this aquifer is maintaining spring flow during drought periods and assessing current and future pumpage effects on spring flow. Maintaining spring flow is a critical objective because the spring outlets are the sole habitat of the Barton Springs salamander, which is listed as an endangered species. The Barton Springs Edwards aquifer represents an excellent field laboratory for testing different models because there is accurate information on recharge using stream gage data, groundwater pumping, spring discharge (for decades), synoptic water levels measured at different times, and continuous water level monitoring records at eight wells distributed throughout the aquifer for up to 10 yr. These data can be used to test the reliability of the models. Although the Barton Springs Edwards aquifer is limited in area (330 km^2, Fig. 1), the term *regional* is used to describe the model to distinguish it from local-scale models. The boundaries of the aquifer are well defined: upper and lower confining layers, recharge zone, and approximate location of the groundwater divides. The lumped parameter model has been described in detail in Barrett (1996) and Barrett and Charbeneau (1996), and only aspects of the lumped parameter model that pertain to the code comparison are described in this paper.

1.1. Study area

This modeling study focuses on the Barton Springs segment of the Edwards aquifer within and adjacent to the city of Austin, Texas, that discharges into Barton Springs and Cold Springs and is hydrologically distinct from the rest of the Edwards aquifer. The model is approximately 330 km^2 in area. The Barton Springs Edwards aquifer (Fig. 1) constitutes the sole source of water to about 45,000 residents. Barton Springs pool also serves as a municipal swimming pool in Zilker Park, downtown Austin. Increased population growth and recent droughts (1996) have focused attention on groundwater resources and sustainability of spring flow.

Model boundaries are all hydrologic boundaries and include the Mount Bonnell fault to the west, which acts as a no-flow boundary (Senger and Kreitler, 1984); a groundwater divide in the south along Onion Creek; the “Bad-water Line” in the east; and the Colorado River (Town Lake) in the north (Fig. 1). Groundwater circulation in the Edwards aquifer decreases to the east and total dissolved solids (TDS) increase. The “Bad-water Line” marks the zone where TDS exceeds 1000 mg/l, which generally coincides with Interstate Highway 35 (IH35, Fig. 1). The groundwater divide in the south separates the Barton Springs segment from the San Antonio segment of the Edwards aquifer, which discharges into Comal Springs and San Marcos Springs.

The Edwards aquifer is up to 165 m thick and it is overlain by the Del Rio Formation, which is predominantly clay and forms a confining unit, and is underlain by the Glen Rose Formation (Fig. 1) (Hovorka et al., 1998). Northeast-trending faults in the study area are part of the Balcones Fault Zone. These faults consist of high-angle normal faults downthrown to the southeast. The faults have displacements of as much as 60 m. The Edwards aquifer is unconfined in the outcrop area where recharge occurs and in part of the section to the east, where it is overlain by the Del Rio Formation (Fig. 1). Farther to the east, the aquifer is confined by the Del Rio Formation. There is no recharge from the land surface to the unconfined section, where it is overlain by the Del Rio Formation. Approximately 80% of the aquifer is unconfined, and the remainder is confined (Slade et al., 1985). The study area is in the subtropical, humid climate zone (Larkin and Bomar, 1983). Mean annual precipitation is 825 mm, with major rainstorms occurring in spring and fall. Groundwater flows from west to east in the unconfined section of the aquifer and generally northeast in the confined section to discharge at Barton Springs.

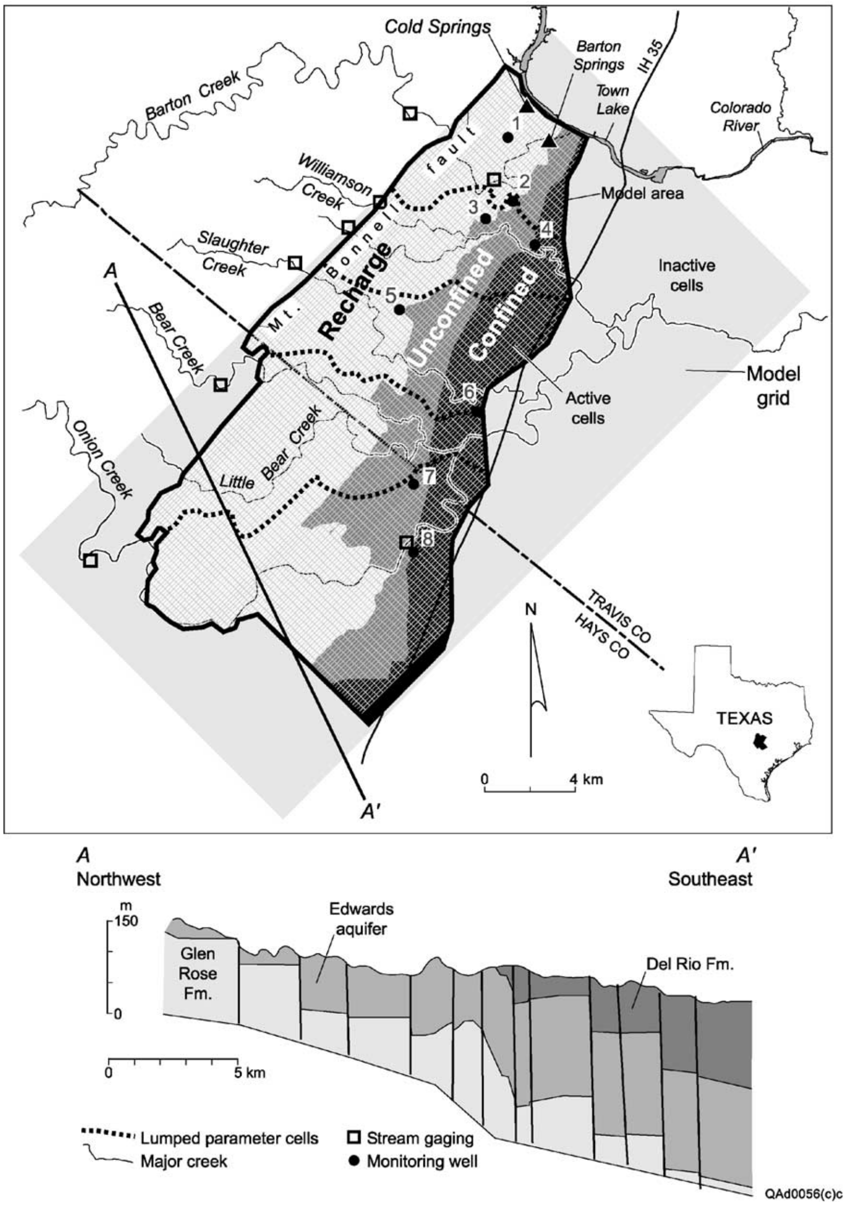

Fig. 1. Location of the study area, including cells for the lumped parameter model, stream gaging stations, and long-term monitoring wells. Recharge, unconfined, and confined zones also shown. The unconfined zone is overlain by the Del Rio Formation, which precludes recharge from the surface in this zone. Monitoring well numbers correspond to the following state well numbers (1, 58-42-8TW; 2, 58-58-216; 3, 58-50-221; 4, 58-50-301; 5, 58-50-411; 6, 58-50-801; 7, 58-58-123; 8, 58-58-101). Well YD 58-42-903 is < 100 m from Barton Springs. The grid for the distributed parameter model is shown where the cells are active. The crosssection shows faults and different units: Glen Rose Formation, Edwards aquifer, and Del Rio Formation (confining unit). The Edwards aquifer includes the Walnut, Kainer, Person, and Georgetown Formations. The city of Austin extends throughout the upper half of the model region. The Balcones Fault Zone includes numerous NE–SW-trending faults that were omitted from this figure for clarity.

Hydraulic gradients are steepest in the west portion of the aquifer and water level fluctuations are low (≤ 3 m). Hydraulic gradients are much lower in the east portion of the aquifer and water level fluctuations are up to 30 m.

The Barton Springs Edwards aquifer is highly karstified, as evidenced by abundant caves, sinkholes, and enlarged fractures. A total of 20 tracer tests were conducted by injecting a tracer in natural recharge features. These tests showed the predominant flow direction to be north to northeast to Barton Springs and Cold Springs (Hauwert et al., 2002). Travel times ranged from 0.8–1.6 km/d during low flow periods to 6.4–11.3 km/d during high flow periods. Additional evidence of karstification includes a ratio of high to low flow of 13:1 in Barton Springs; rapid response of water levels in many wells to recharge, indicating good interconnection; and uniform water levels in some wells, indicating that they penetrated conduits (e.g. well 6, Fig. 1). Exponential relationships (exponents ~ 0.5) between water levels in wells southeast of Barton Springs and spring discharge are attributed to turbulent flow (Barrett and Charbeneau, 1996, 1997; Worthington et al., 2001).

2. Methods

This section focuses primarily on the distributed parameter model because the lumped parameter model has been described in detail in Barrett (1996) and Barrett and Charbeneau (1996).

2.1. Groundwater recharge and discharge

The primary source of recharge is provided by seepage from streams in the recharge zone where the Edwards aquifer crops out. The Del Rio Formation to the east precludes recharge from occurring in the unconfined zone. Flow losses from the creeks are sufficient to account for most of the groundwater discharge in springs and from wells. Five major creeks (Barton, Williamson, Slaughter, Bear, and Onion) provide most of the recharge to this area (Fig. 1). Each of the creeks forms a separate cell for the lumped parameter model (Fig. 2). Recharge for Onion Creek was calculated by subtracting daily average flow downstream from that upstream of the recharge zone. Losses from Onion Creek to the southwest are considered negligible because this is a groundwater divide (Guyton and Associates, 1958). With the exception of Barton Creek, all flow is lost to aquifer recharge until a threshold flow (Slade et al., 1985) is exceeded. Once the threshold value is reached, recharge is assumed constant at that value. In the case of Barton Creek, the downstream gaging station is located within the recharge zone; therefore, recharge from this creek may be underestimated. For low flows (≤ 0.85 m^3/s at the upstream gaging station on Barton Creek), recharge is approximately equal to stream loss. For flows between 0.85 and 7.08 m^3/s, a quadratic relationship developed by Barrett and Charbeneau (1996) was used. Flows greater than 7.08 m^3/s were assigned this value for recharge because it was the highest measured recharge. Diffuse interstream recharge represents recharge outside the main channels and includes recharge from tributaries and precipitation. Diffuse interstream recharge was assumed equal to 15% of total recharge on the basis of water budget studies conducted by Slade et al. (1985) and modeling analysis by Barrett and Charbeneau (1996). Average annual recharge was calculated for the 19-yr period (1980–1998; 2.35 m^3/s) for the steady state simulations. Annual recharge varied markedly over the time period represented by the transient simulation (1989–1998; Table 1).

Groundwater discharge occurs primarily at Barton Springs, which consists of a series of springs in the Barton Springs Pool area in Barton Creek close to where it enters the Colorado River. Barton Springs discharge is calculated from a rating curve that relates water levels in well YD-58-42-903 (<100 m from Barton Springs) to spring discharge. Long-term discharge at Barton Springs is 1.5 m^3/s (1918–1999). Cold Springs, northwest of Barton Springs, discharges into the Colorado River but is not gaged because it is flooded by Town Lake. Discharge at Cold Springs has ranged from 3 to 28% of that at Barton Springs at different times.

Groundwater is also discharged through pumping wells. Monthly pumpage data are collected by the Barton Springs Edwards Aquifer Conservation District (BSEACD) and are available from 1989 through present. Annual pumpage ranged from 0.11 m^3/s (1990, 1991) to 0.18 m^3/s (1998) (Table 1).

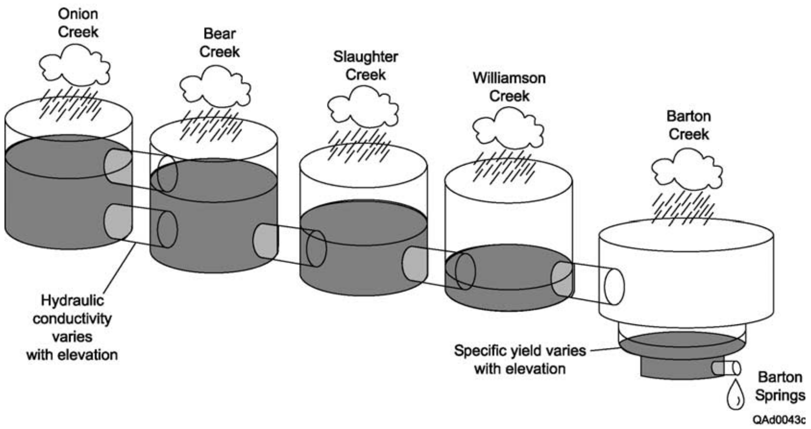

Fig. 2. Schematic of lumped parameter model cells and connections. The water levels in each of the cells do not represent conditions at any particular time. This figure is modified from Barrett and Charbeneau (1996, 1997).

The average annual pumpage from 1989 to 1998 was 0.14 m^3/s. One of the years with lowest pumpage (1991) corresponds to the year with highest precipitation. Annual pumpage ranged from 3% (1991, 1992) to 138% (1996) of recharge (Table 1).

2.2. Distributed parameter model

The code MODFLOW-96 (Harbaugh and McDonald, 1996) was used for distributed parameter model simulations. The model consists of 1 layer that has 120 rows and 120 columns and a total of 14,400 cells. Model rows were aligned parallel to the strike of the Edwards; the grid was therefore rotated 45° from north to south. Rectangular cells were 305 m long parallel to the strike of the faults and 152 m wide. Cell size was chosen to be small enough to reflect the availability of input data and to provide appropriate details in the output. Cells outside the model area were made inactive, resulting in 7043 active cells (Fig. 1).

Hydrologic boundaries were assigned to the model as described previously. Cells with layer thickness of less than 6 m were assigned as inactive. The structure of the top of the aquifer was based on ground-surface elevation in the unconfined recharge zone. East of the outcrop zone, the top of the aquifer corresponds to the base of the Del Rio Formation. The base of the aquifer corresponds to the base of the Walnut Formation, determined from recent studies by Small et al. (1996). The location of faults was also based on interpretations by Small et al. (1996). Recharge values were assigned to stream cells on the basis of analysis of flow losses in the streams. Recharge was uniformly

Table 1
Annual precipitation, recharge, pumpage for the 1989–1998 period. Recharge is also presented as a percentage of pumpage

Time (yr)	Precipitation (mm)	Recharge (m^3/s)	Pumpage (m^3/s)	Recharge as % of pumpage
1989	657	0.82	0.14	18
1990	722	0.59	0.11	19
1991	1326	3.99	0.11	3
1992	1170	4.77	0.13	3
1993	673	1.87	0.15	8
1994	1045	0.95	0.15	16
1995	863	2.35	0.15	6
1996	751	0.12	0.16	138
1997	1195	3.61	0.16	4
1998	993	4.35	0.18	4

distributed in each stream where the stream intersects the outcrop. Interstream recharge was 15% of the total stream recharge. Pumping was assigned to cells on the basis of the location of pumping wells. Unreported domestic (rural) pumpage was calculated from countywide estimates and was assigned to all active cells. Barton Springs and Cold Springs were represented as drains with elevations of 131.67 m for Barton Springs and 131.06 m for Cold Springs, and a high drain conductance value was used (93,000 m^2/d [1,000,000 ft^2/d]) to allow unrestricted discharge of water. The model layer was assigned as confined/unconfined. The model was set up to calculate transmissivity and storativity on the basis of saturated thickness. Initial head for steady state simulations was the top of the aquifer.

2.2.1. Calibration

Two basic steps were followed in modeling the aquifer: a steady state model was developed to determine the spatial distribution of hydraulic conductivity, and a transient model was run for a 10-yr period (1989–1998) by using monthly recharge and pumpage. Zonal distribution of hydraulic conductivity was determined using trial and error and automated inverse calibration procedures. Specific yield was calibrated using the transient model.

2.2.1.1. Steady state model calibration. Although steady state is never reached in a karst aquifer, a steady state model was developed to calibrate the distributed parameter model because calibration could focus on the hydraulic conductivity distribution and avoid having to consider storage parameters as would be required for transient calibration. The goal of the steady state calibration was to determine the distribution of hydraulic conductivity required to match the potentiometric surface during average discharge using average recharge as input to the model. Measured water levels in July and August (1999) were used to evaluate the steady state model calibration because spring discharge (~1.87 m^3/s) was close to average conditions (~1.5 m^3/s) at this time and the number of synoptic water level measurements (81) was greatest (Fig. 3). Spatial distribution of recharge among the streams and in the interstream settings was based on average recharge for a 19-yr period (1980–1998). Total recharge was reduced to equal the average

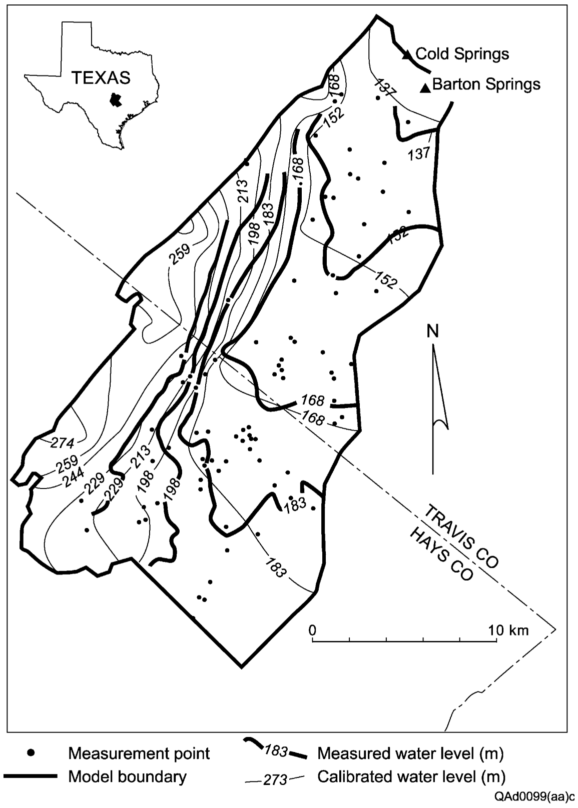

Fig. 3. Comparison of simulated water-level elevations with water levels measured in July/August 1999 for the steady state model.

spring discharge for Barton Springs and Cold Springs of 1.56 m^3/s and average pumpage for 1989–1998 (0.14 m^3/s). Recharge was assumed to be known and was not changed during the calibration. Faults with the greatest amount of offset (≥76 m; 250 ft) were simulated as horizontal flow barriers (Hsieh and Freckleton, 1993). Trial and error and automated inverse methods were used to estimate hydraulic conductivity.

There are measurements of hydraulic conductivities for the Barton Springs Edwards aquifer from aquifer tests throughout the study area; however, such point estimates of hydraulic conductivity are considered unreliable estimates of bulk hydraulic conductivity because of scaling issues. Hydraulic conductivities assigned to cells in an equivalent porous media model represent the matrix, fractures, and conduits within that cell and are dominated by the most conductive continuous features across the cell.

Because the area covered by the most conductive features (e.g. conduits) within a cell may be small relative to the total area of a cell, it is unlikely that a well will intersect these features. Therefore, hydraulic conductivities from well data will tend to underestimate, oftentimes dramatically, the actual hydraulic conductivity of a given area in a karst aquifer. It is therefore more appropriate to use model calibration to estimate distribution of hydraulic conductivity.

Initial attempts to use automated inverse methods to estimate hydraulic conductivity would not converge because the automated inverse procedure could not be used to define zones of varying hydraulic conductivity. Therefore, we used a combination of trial and error and automated inverse methods to optimize the hydraulic conductivity distribution. The trial and error procedure consisted of increasing the number of zones and varying the hydraulic conductivity of the zones until there was good correspondence between simulated and measured heads. Hydraulic gradients were used to assign zones and estimate hydraulic conductivities. Steep hydraulic gradients generally in the west portion of the model suggest low hydraulic conductivities, although water level data are limited in the extreme west and southwest portions of the model. In contrast, shallow hydraulic gradients near Barton Springs suggest high hydraulic conductivities because of flow convergence (Fig. 3). This relationship is generally found in many karst aquifers (Worthington et al., 2001) and has also been used to delineate the spatial distribution of hydraulic conductivity in other aquifers (Greene et al., 1999). The steep hydraulic gradients in the west could not be simulated by simply varying the spatial distribution of recharge in this region. The hydraulic conductivity distribution from the trial and error procedure was further refined with automated inverse modeling using UCODE (Poeter and Hill, 1998). Hydraulic heads measured in 81 wells in July/August 1999 were used in UCODE. Log transformation of hydraulic conductivity was used in UCODE.

2.2.1.2. Transient model calibration. Simulated heads and the calibrated distribution of horizontal hydraulic conductivity from the steady state model were used as input for the 10 yr (1989–1998) transient model. Monthly stress periods were used for transient simulations, which resulted in a total of 120 stress periods for the 10 yr simulation (1989–1998). Recharge and pumpage were changed for each stress period. Recharge rates were estimated from streamloss studies as discussed previously. Annual pumpage ranged from 0.11 m^3/s (1990, 1991) to 0.18 m^3/s (1998). The initial estimate of specific yield (0.01) was based on data from Slade et al. (1985). Specific storage was set at a value of $1.5 \times 10^{-6}\ m^{-1}$.

Initial transient simulations did not converge because simulated hydraulic heads in cells near the west-central portion oscillated between iterations. These cells were located in a zone where the base of the Edwards aquifer was much higher than surrounding areas. By lowering the base of some of these cells to values similar to those in adjacent areas, convergence was achieved. This lowering assumes that the underlying Glen Rose Formation is locally permeable and connected to the Edwards aquifer.

2.3. Lumped parameter model

The lumped parameter model developed by Barrett (1996) was used to simulate water levels and spring discharge for 1989–1998 for comparison with the distributed parameter model (Fig. 2). The use of different numbers of cells was evaluated, including one cell, seven cells, and, finally, five cells, each representing the major creeks in the study area (Figs. 1 and 2). A single well was chosen in each cell to represent water levels in that portion of the aquifer (monitoring wells 2, 4, 6, and 8 (Fig. 1) and well YD-58-42-903 immediately adjacent to Barton Springs). The model was calibrated using spring discharge and representative water-level fluctuations in each cell for 1989–1994. Each cell is treated as a tank, which is assigned an effective area (equivalent to the product of specific yield and surface area). Flow between cells was originally assumed to be turbulent, and the Chezy Manning equation was used to describe such flow. Although the turbulent flow assumption resulted in adequate simulations of spring discharge, hydraulic heads in representative wells were greatly overestimated by up to 100 m. The final model used Darcy's law to describe flow between cells. Hydraulic conductivity of the Onion Creek cell was allowed to vary with elevation to reduce overprediction of water levels during high recharge periods (Fig. 2). The area of the Barton Creek cell was reduced with depth to reproduce the Barton Springs recession, which is more gradual initially and increases with time. Because of the small number of cells, configuration of the model and the properties of each of the cells were successfully determined through trial and error.

2.4. Sensitivity analysis

Sensitivity of water levels in the calibrated distributed parameter model and spring discharge and water levels in the transient distributed and lumped parameter models to different aquifer parameters was evaluated. Sensitivity analysis quantifies uncertainty of the calibrated model to uncertainty in the estimates of the aquifer parameters, stresses, and boundary conditions (Anderson and Woessner, 1992, p. 246). The nonuniqueness of the calibrated model can be evaluated using sensitivity analysis. The hydrologic parameters that have the greatest impact on simulated water levels and spring discharge can also be identified through sensitivity analyses.

Sensitivity analyses were conducted on hydraulic conductivity, recharge, spring drain conductance, and pumpage in the steady state distributed parameter model, recharge, pumpage, specific yield, and specific storage in the transient distributed parameter model, and recharge and pumpage in the transient lumped parameter model. Each parameter was varied systematically, and the mean difference between simulated water levels from the sensitivity simulation relative to the calibrated base case simulation was calculated. Sensitivity of the lumped parameter model to variations in recharge and pumpage was also evaluated.

3. Results and discussion

3.1. Distributed parameter model

3.1.1. Steady state model

Calibration of the steady state model using trial and error resulted in 10 zones, with hydraulic conductivity ranging from 0.3 to 305 m/d. Hydraulic conductivities were generally low (0.3–1.3 m/d) in the outcrop area where the hydraulic gradient is steep. A low value of hydraulic conductivity (0. 3 m^3/d) was also assigned to the confined section near the “Bad-Water Line”. Much higher hydraulic conductivities (up to 305 m/d) were required in zones trending NNE toward Barton Springs. Hydraulic conductivities are generally high adjacent to major springs because of the confluence of conduits and increased dissolution in this zone. At the regional scale, measured water levels appear to be continuous. Comparison of simulated and measured heads yielded a root mean square (RMS) error of 11 m. Use of automated inversion with UCODE (Poeter and Hill, 1998) further reduced the RMS error to 7 m, which represents about 7% of the total head drop across the model ($\sim$100 m). The primary difference between trial and error and automated zonal hydraulic conductivity estimates was in the confined section to the southeast, where hydraulic conductivity was increased from 0.3 to 11.9 m/d. Two hydraulic conductivity zones near Barton Springs were highly correlated and were combined. The final distribution of hydraulic conductivity is shown in Fig. 4 based on trial and error and inverse modeling. The steady state model generally reproduced the potentiometric surface developed from water level measurements in July/August 1999 (Fig. 3). The scatter plot of simulated versus measured heads indicates that there is very little bias in the simulation results (Fig. 5). RMS error reflects uncertainties in both measured and simulated hydraulic heads. The heads were measured over a 2-month period. Measurement of synoptic water levels over a 2-month period is generally considered very short for most porous media aquifers but is fairly long for this karst aquifer, which is dynamic, and spring discharge decreased from 2.27 to 1.70 m^3/s during this time. Therefore, the measured heads may not accurately reflect average discharge of Barton Springs (1.56 m^3/s). Most of the head data were based on well locations and elevations obtained from 1:24,000 topographic maps, whereas some head data were based on global positioning system measurements and 1-m contour interval maps available for the city of Austin and vicinity. Simulated discharge was 1.47 m^3/s at Barton Springs, 0.09 m^3/s at Cold Springs, and 0.14 m^3/s from pumping wells.

Sensitivity analyses indicated that water levels were most sensitive to variations in recharge and hydraulic conductivity (Fig. 6). The impact of variations in pumpage and spring drain conductance was negligible. The insensitivity of the model to

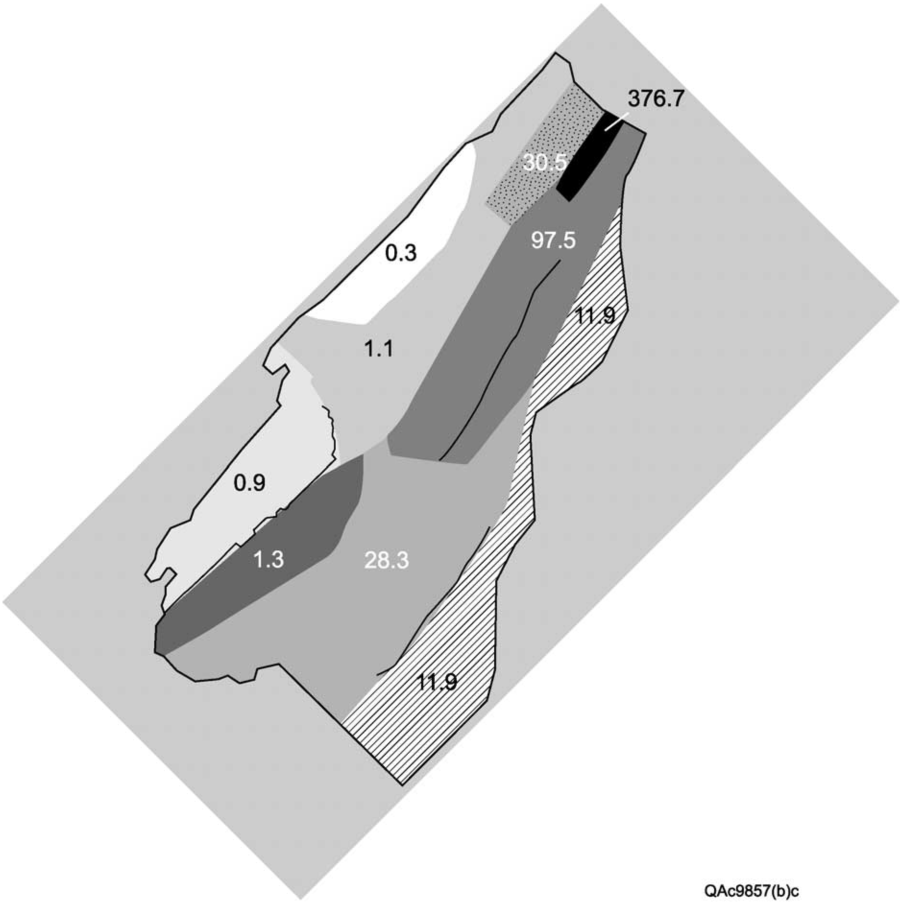

Fig. 4. Hydraulic conductivity (m/d) resulting from calibration of the steady state model. The faults shown on this map were simulated as horizontal flow barriers.

pumpage may result from average pumpage representing only a small fraction of the total water budget (i.e. 9% of total recharge). Simulated water levels increased with increasing recharge and with decreasing hydraulic conductivity. Reducing recharge by 25 and 50% of the calibrated value would not converge because many of the cells became dry.

3.1.2. Transient model

The initial transient simulation used a specific yield of 0.01 and a specific storage of 1.5×10^{-6} m^3/s and generally reproduced spring discharge fluctuations during the 10-yr period without any calibration. Many simulations were conducted at different times using various distributions of hydraulic conductivity that had minimal impact on spring discharge variations.

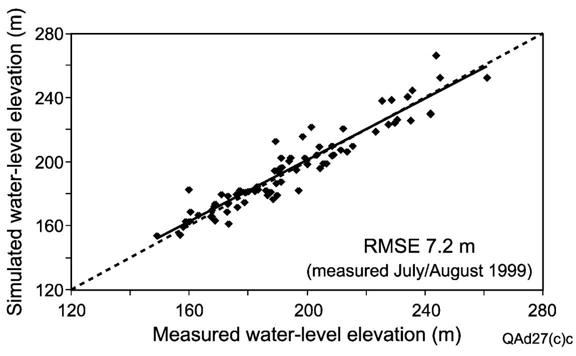

Fig. 5. Scatter plot of simulated versus measured water-level elevations for the steady state model developed using trial and error and automated inverse modeling.

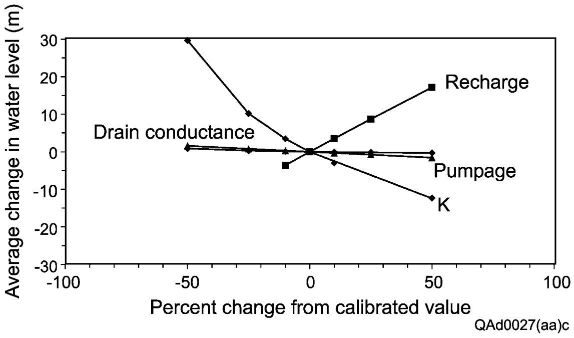

Fig. 6. Sensitivity of the predicted water levels of the steady state model to changes in recharge, hydraulic conductivity, pumpage, and spring drain conductance.

Simulation results were biased in that low flows were slightly overestimated and high flows generally underestimated. Because simulation of low flows is the primary goal of the model, we varied specific storage in the confined aquifer and specific yield in the unconfined aquifer to better reproduce measured low flows. Reducing specific yield from 0.01 to 0.005 increased the range of simulated spring discharges and provided more accurate simulations of low flow discharge. The resultant transient simulation was evaluated using three different criteria. (1) Simulated and measured spring discharges were compared (Figs. 7 and 8). (2) Scatter plots were developed for simulated and measured heads at different times (1994, 1996) (Fig. 9). (3) Simulated hydraulic heads were compared with hydrographs for some representative monitoring wells (Fig. 10).

Generally good agreement was obtained between measured and simulated discharge at Barton Springs (Figs. 7 and 8). The RMS error between measured and simulated discharge for the distributed parameter model is 0.35 m^3/s, which represents 11% of the discharge fluctuations measured at Barton Springs during that time. Data from an 8-month period from December 1991–July 1992 were omitted from the error calculations because of uncertainties related to the measured discharge data as a result of flooding. One of the main objectives of the model is to accurately simulate low flows in Barton Springs. The scatter plot suggests that on average there is no bias in the results; however, the scatter plot masks underpredictions and overpredictions at different times (Fig. 8(a)). Overprediction of low spring flows in 1989 and early 1990 is attributed to initial conditions (hydraulic head from steady state model) not being in equilibrium with boundary conditions (recharge and discharge) for the transient simulation. Good correspondence between measured and simulated discharge was found for 1990–1991 (Fig. 7). Simulated spring discharge generally underestimates measured discharge during the 1994 recession; however, both measured and simulated discharges have the same minimum value. In contrast, simulated discharge overestimates measured discharge during the 1996 recession. Slope of the simulated recession is more gradual than the measured recession, which is U shaped, and the timing of the minimum simulated discharge is later than that of the measured data. Peak discharges are underestimated in some cases (1990–1991), simulated accurately in other cases (1989, 1993, 1995), and overestimated in other cases (1991, 1992, 1997, 1998). During high flows, some of the discharge may be diverted to an ungaged spring and other smaller springs along Barton Creek, which are not accounted for in the model.

Scatter plots between measured and simulated water levels were developed for different times during the transient simulation (Fig. 9). The scatter plot for March/April 1994 shows that the model generally simulated water levels during low-flow conditions. The RMS error of 8.7 m represents 11% of the measured head drop in the model area. Comparison of measured and simulated water levels for July and August 1996 indicates that simulated water levels underestimate measured water levels by 11.2 m (10% of head drop) on average for this low-flow period. It is difficult to compare measured and simulated water levels during high flow periods because spring discharge is generally changing rapidly and synoptic water level measurements over 2 month time periods generally span large changes in spring discharge. In general, the model matches water levels for different times within 10% of the head drop in the model region.

The transient model generally reproduces water levels in many of the continuously monitored wells (Fig. 10). Qualitative agreement between measured and simulated water levels was better in the north part of the aquifer than in the south. Because well 5 is located adjacent to a cave (N. Hauwert, BSEACD, personal communication, 2000), its water levels

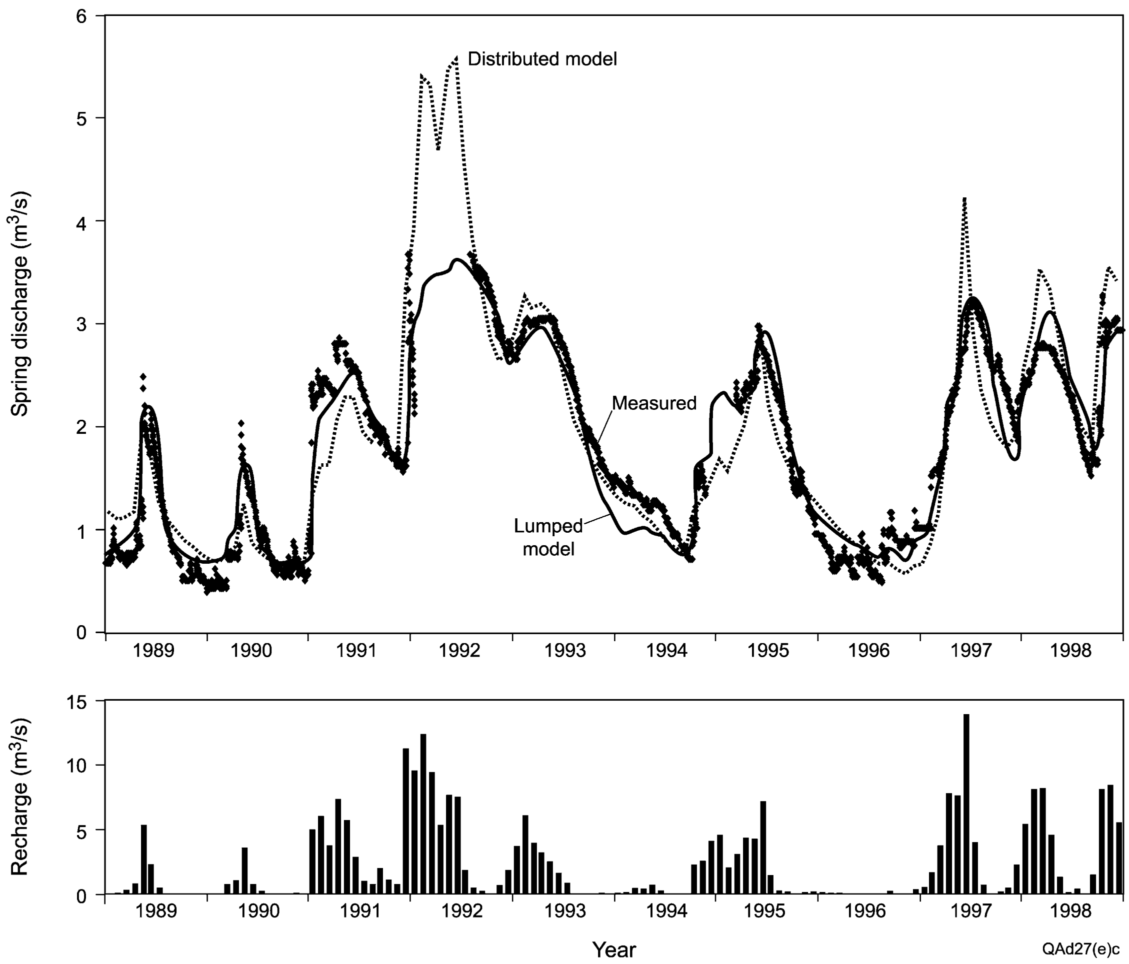

Fig. 7. Comparison of simulated and measured discharge at Barton Springs for 1989–1998. Monthly recharge input is also shown. Discharge values were unreliable in early 1992 because of flooding and also in early 1995.

remain fairly constant. These water levels are not reproduced in the simulation, which cannot represent flow in caves. Simulation of water level fluctuations in monitoring well 8 was poor, and attempts to improve simulations in this region by varying hydraulic conductivity were unsuccessful. Simulation of water levels in this region is discussed in more detail in Section 3.2 on sensitivity analysis and also in the comparison between lumped and distributed parameter models (Section 3.3).

Sensitivity analyses were generally restricted to the range of −10 to +50% of most parameters because simulations did not generally converge when parameters were reduced more than 10%. Groundwater recharge had the greatest impact on spring discharge and water levels in monitoring wells. Increasing recharge by 50% resulted in increasing the mean spring discharge by about the same amount. This result is expected because discharge equals recharge over long time periods (Table 2; Fig. 11). Increasing recharge had a greater impact on high flows than on low flows, and spring discharge was more variable as shown by the range and coefficient of variation of spring discharge (Table 2). These increases in spring discharge in response to increasing recharge may not be observed if high spring flows are diverted to ungaged springs along Barton Creek. Simulated water levels in monitoring wells displayed a similar

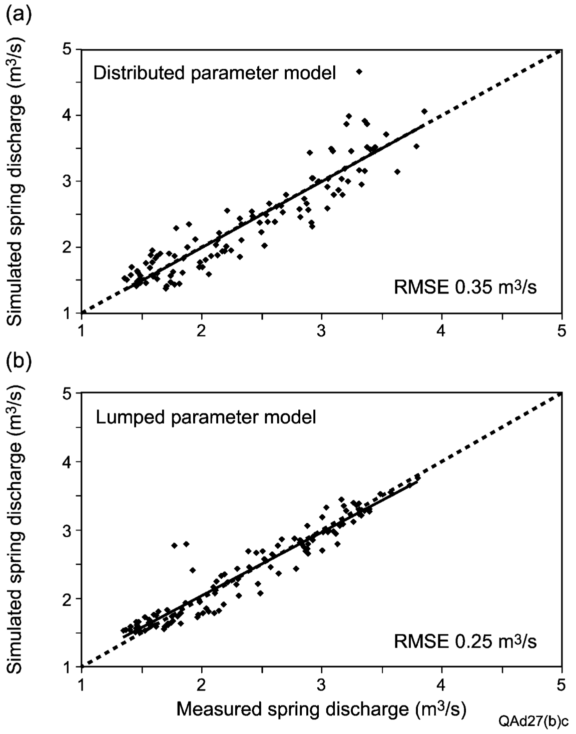

Fig. 8. Scatter plot of simulated versus measured spring discharge for 1989–1998 for (a) the distributed parameter model and (b) the lumped parameter model.

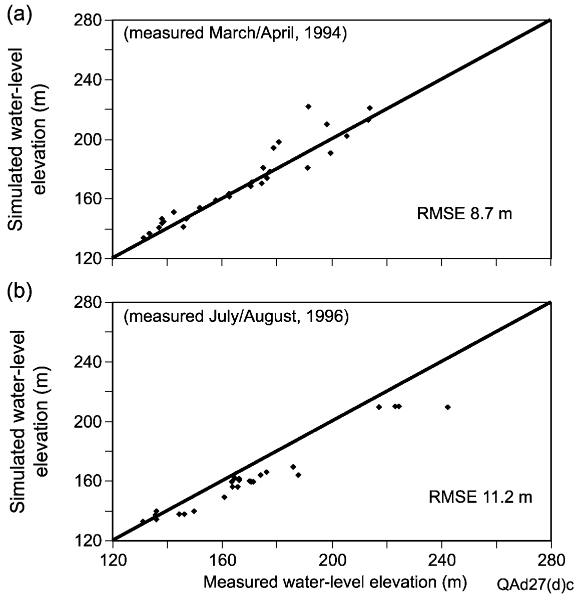

Fig. 9. Scatter plot of simulated versus measured water-level elevations for the transient simulation (a) March/April 1994 and (b) July/August 1996.

response to variations in recharge as spring discharge (Fig. 12). Decreasing recharge had the opposite effect of increasing recharge. Simulated spring discharge and water levels in wells were much less sensitive to variations in pumpage, specific yield, and specific storage (Table 2, Figs. 11 and 12). Increasing these parameters by up to 50% resulted in $\leq 2\%$ increase in mean spring discharge and $\leq 5\%$ change in monitored water levels. The low sensitivity to pumpage may result from low temporal variability in pumpage during the 10-yr simulation period (Table 1). These sensitivity analyses have important implications for future management of the aquifer and suggest that maintaining spring flow during droughts may require enhanced recharge. Focus on conservation may not be sufficient to maintain spring flow during periods of prolonged drought. Uncertainties in specific storage are greater than those of specific yield; therefore, an additional simulation was conducted to evaluate the impact of varying specific storage by a factor of 10. Increasing specific storage by 10 decreased mean spring discharge slightly but greatly reduced the range in spring discharge (Table 2). Increased specific storage does not simulate low spring discharges, which is critical for this study. Increasing specific storage by 10 had a similar effect on simulated water levels in the monitoring wells, which better replicate measured water level fluctuations in the monitoring wells (Fig. 12(c) and (d)). However, the study's emphasis on simulating low spring discharges over accurately simulating water levels in monitoring wells precludes using the higher specific storage in the final simulations.

3.2. Lumped parameter model

Generally good agreement was obtained between measured and simulated discharge at Barton Springs using the lumped parameter model (Fig. 7). The RMS error for the 1989–1998 period is 0.25 m^3/s, which represents 10% of the discharge fluctuations measured at Barton Springs during that time. The period 1989–1994 was used in model calibration, and the period 1994–1998 was used in model evaluation. Specific yield was varied with elevation in the Barton Springs cell to better simulate low flow discharge in Barton

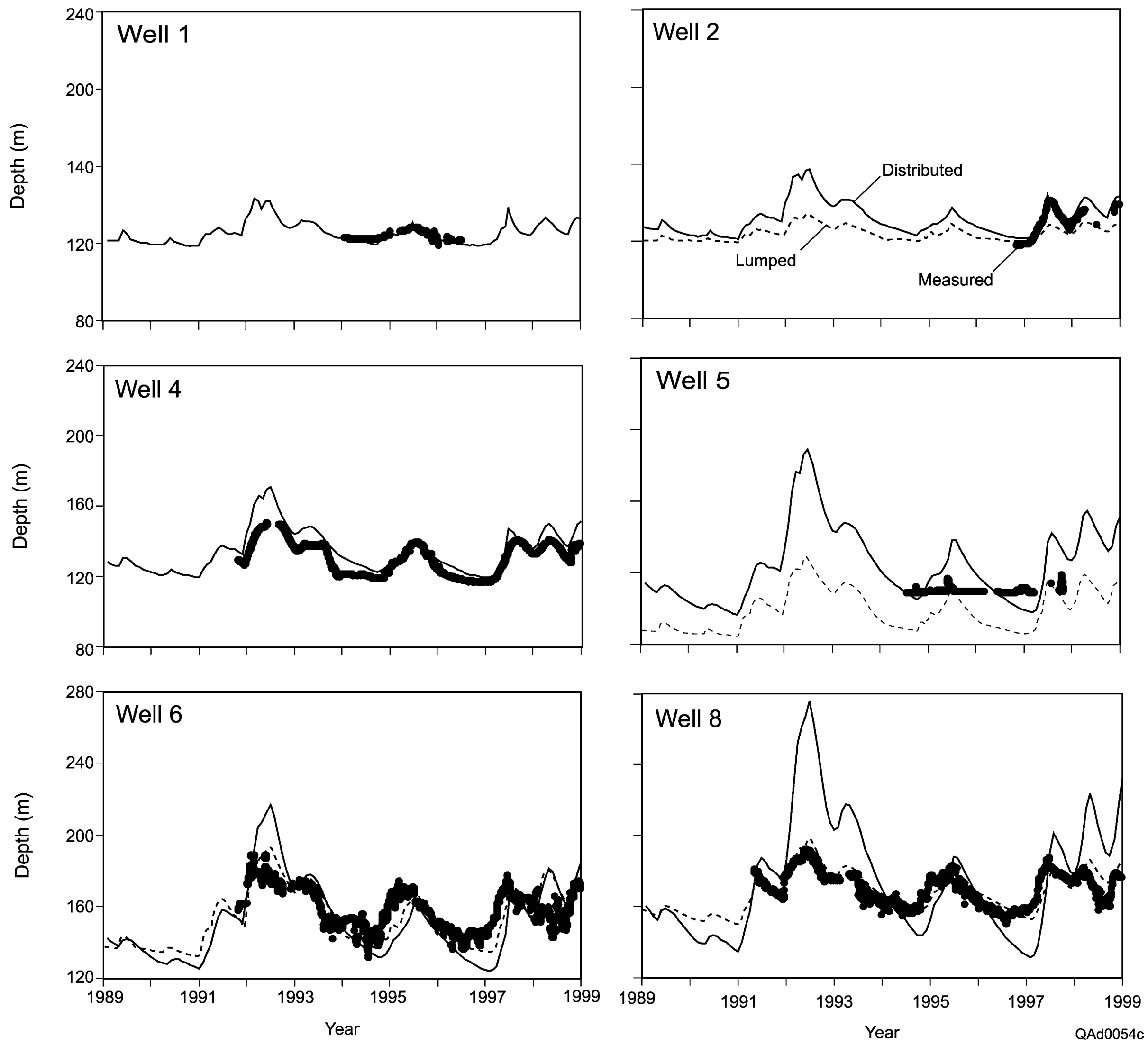

Fig. 10. Comparison of simulated and measured water-level elevation hydrographs in selected monitoring wells. For location of monitoring wells, see Fig. 1.

Springs. The lumped parameter model accurately simulates the period from 1994 to 1998. There is very little bias in the simulation results. The lumped parameter model also incorporated discharge at ungaged springs when discharge at Barton Springs exceeded 2.27 m^3/s, which accounts for the lower simulated spring flows using the lumped parameter model relative to those simulated using the distributed parameter model. The model also simulates water levels in key wells that were chosen to represent each of the cells in the model, particularly those in the south portion of the model (wells 6 and 8; Fig. 10). Hydraulic conductivity was varied with elevation in the southernmost cell (Onion Creek cell) to better simulate water level fluctuations in this cell. Results from sensitivity analyses for the lumped parameter model were similar to those for the distributed parameter model (Table 2). Simulation results were most sensitive to recharge and much less sensitive to pumpage variations. Sensitivity analyses indicate that

Table 2
Sensitivity of spring discharge to variations in recharge, pumpage, specific yield, and specific storage based on the distributed and lumped parameter models

	Mean (m^3/s)	Minimum (m^3/s)	Maximum (m^3/s)	Range (m^3/s)	Coefficient of variation
Distributed parameter model					
Calibrated value	1.914	0.538	5.550	5.012	0.017
Recharge (−10%)	1.722	0.510	4.870	4.361	0.017
Recharge (50%)	2.908	0.736	9.033	8.297	0.019
Pumpage (−10%)	1.928	0.555	5.581	5.026	0.017
Pumpage (+50%)	1.852	0.493	5.499	5.009	0.018
Specific yield (−10%)	1.923	0.510	5.663	5.154	0.018
Specific yield (+50%)	1.883	0.651	5.012	4.361	0.015
Specific storage (−10%)	1.920	0.538	5.862	5.324	0.018
Specific storage (+50%)	1.900	0.566	5.040	4.474	0.016
Specific storage (10×)	1.818	0.793	3.766	2.973	0.010
Lumped parameter model					
Calibrated value	1.836	0.632	3.643	3.011	0.912
Recharge (−50%)	1.300	0.508	3.068	2.559	0.683
Recharge (−10%)	1.749	0.610	3.546	2.936	0.880
Recharge (+50%)	2.179	0.735	4.079	3.344	1.024
Pumpage (−50%)	1.889	0.683	3.666	2.983	0.908
Pumpage (−10%)	1.847	0.642	3.648	3.005	0.911
Pumpage (+50%)	1.783	0.582	3.622	3.040	0.915

spring discharge did not vary as much as a function of recharge in the lumped parameter model relative to the distributed parameter model because the lumped parameter model includes ungaged springs that discharge at high flows.

3.3. Comparison of distributed and lumped parameter models

In this study results, simulated spring discharges using the lumped parameter model are similar to those using the distributed parameter model (Fig. 7). The lumped parameter model used daily recharge and pumpage inputs, whereas the distributed parameter model used monthly inputs; however, monthly recharge for the distributed parameter model was calculated using daily stream flow data. The similarity in simulation results for flows ≤ 3 m^3/s suggests that the model is generally insensitive to the time resolution of inputs within daily to monthly time-scales. The lumped parameter model also included additional ungaged springs that were active at high flow and which allowed the model to simulate discharge at Barton Springs more accurately during high flow. Future distributed parameter models should consider incorporating additional outflows during high flow conditions. The lumped parameter model more accurately simulated water level fluctuations in the monitoring wells in the south part of the study area than did the distributed parameter model. The lumped parameter model uses two layers to represent the southernmost cell. Using two layers rather than one layer may also help improve the simulation of water level fluctuations with the distributed parameter model and allow different parameters to be used during high flow conditions that would not impact the simulation of low spring discharges.

The main advantages of the lumped parameter model are its simplicity, ease of use, low data requirements, and rapid simulation runs. The lumped parameter model can readily be run from a spread sheet. It is also easy to evaluate different processes using the lumped parameter model, such as turbulent flow. Layering can be incorporated into individual

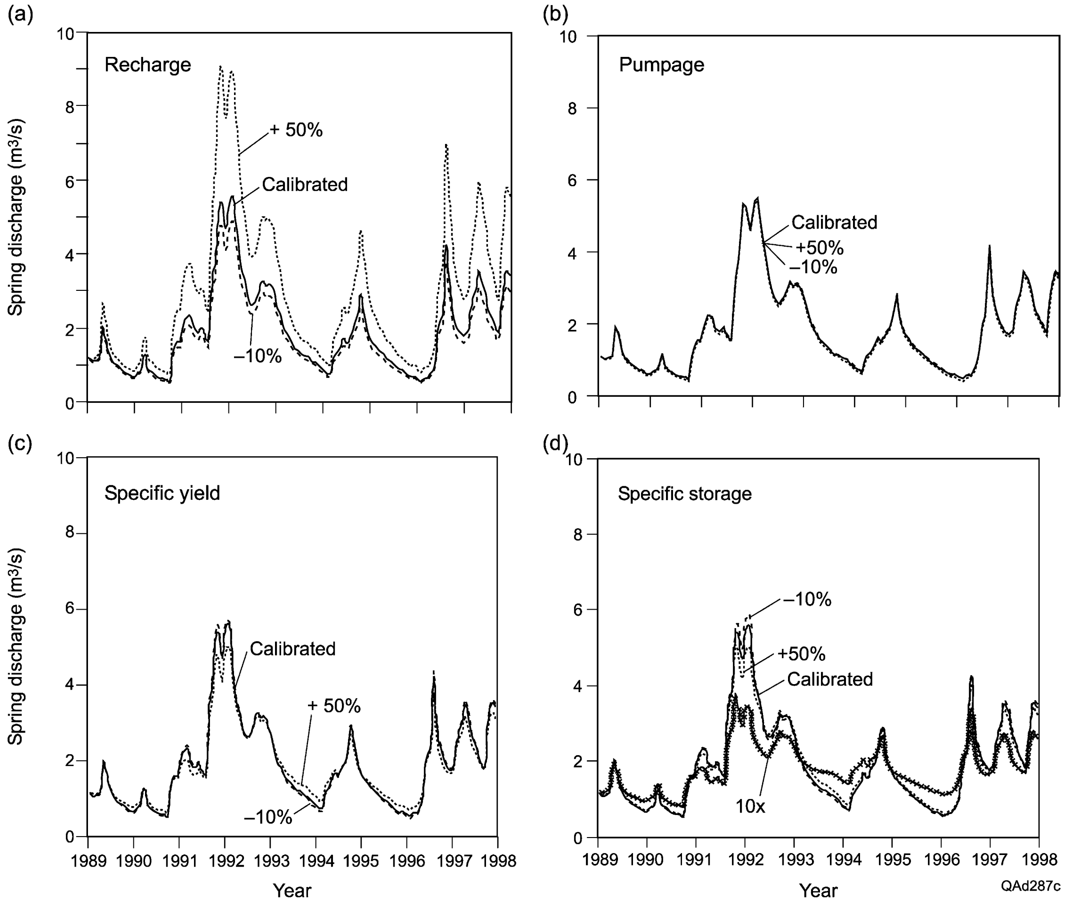

Fig. 11. Sensitivity of predicted spring discharge from the transient model to variations in (a) recharge, (b) pumpage, (c) specific yield, and (d) specific storage.

cells without much difficulty. The lumped parameter model may require a large range in flow conditions to accurately calibrate it.

In contrast to the lumped parameter model, the distributed parameter model is much more complex, is difficult to parameterize, has large data requirements, and requires much longer times to run. The hydraulic conductivity distribution for the distributed parameter model can generally be obtained from calibration using mean steady state flow conditions. It would be difficult to incorporate layering into the distributed parameter model of the Barton Springs model because layering would have to be incorporated into the entire model and problems can arise because of cells becoming dry. One of the advantages of the distributed parameter model is that the code on which it is based, MODFLOW, is widely used and tested. Parameterization was simplified for the Barton Springs model because recharge is fairly accurately known and only hydraulic conductivity had to be estimated. Because karst aquifers are generally very dynamic, obtaining representative synoptic water level maps can be difficult.

Choice of modeling approach will depend partly on the objective of the modeling study. If the primary objective of the modeling study is to accurately simulate spring discharge, then either lumped or distributed parameter models can be used. However, a distributed parameter model is

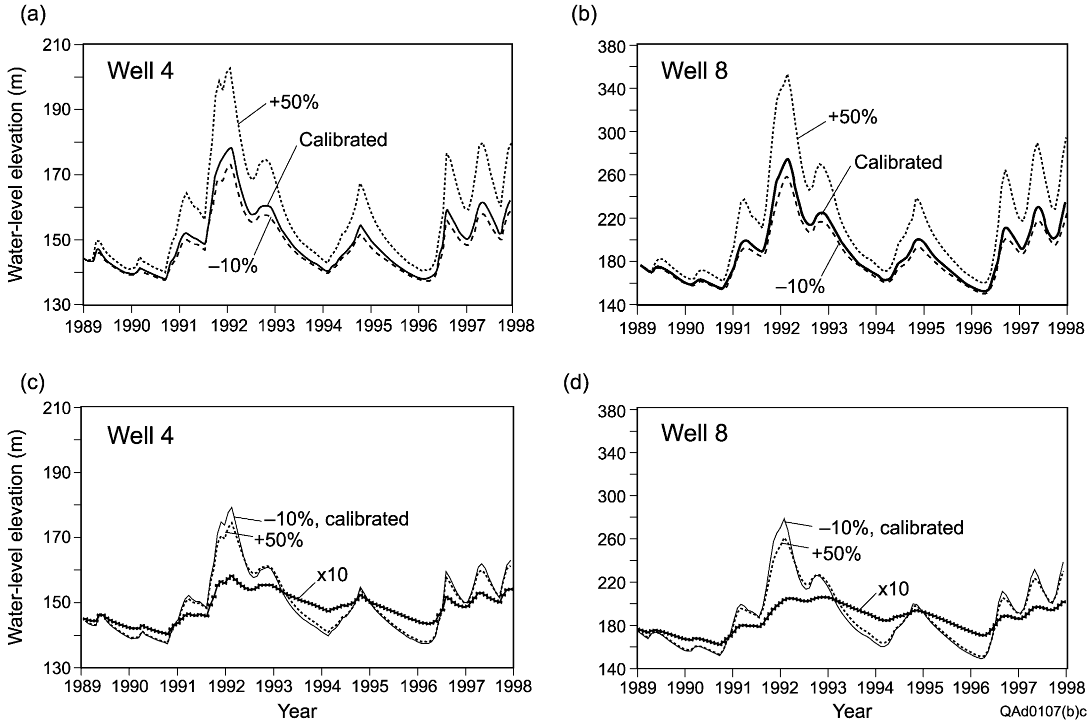

Fig. 12. Sensitivity of predicted water levels in representative wells to (a, b) recharge and (c, d) specific storage.

required to simulate the potentiometric surface, which is necessary to represent regional groundwater flow direction. Both lumped and distributed parameter models can simulate water level fluctuations in monitoring wells; however, the lumped parameter model is restricted to simulating a representative well in each cell. While the lumped parameter model can evaluate impact of pumpage on spring flow at a regional scale, it cannot be used to assess the effect of more local scale pumping on water levels in the aquifer. The distributed parameter model also cannot assess the effect of individual well pumpage on water levels but may be able to evaluate the impact of a large well field. Barrett and Charbeneau (1996) showed that the lumped parameter model can be used to evaluate the effect of non-point source contamination on spring water quality. The distributed parameter model should also be able to address non-point source contamination. Because neither model can simulate the local direction or rate of water flow, these models should not be used to assess the fate of point source contamination or to delineate protection zones for wells or springs.

3.4. Limitations of equivalent porous media models

It is critical to recognize the limitations of different modeling approaches. Unrealistic expectations and inappropriate applications of models can greatly reduce confidence in the use of numerical models. The equivalent porous media models developed in this study cannot be used to simulate local direction or rate of groundwater flow because major conduits are not explicitly represented in the models and because turbulent flow is not included. Objectives such as delineation of protection zones for wells and springs and simulation of point source contamination or aquifer tracer tests cannot be accomplished with these equivalent porous media models. These models should be restricted to evaluation of regional groundwater flow issues. It is questionable whether any modeling approach can predict direction and rate of groundwater flow from a point source because many tracer tests demonstrate that directions and rates of transport can be quite variable. Quinlan et al. (1996) suggested that tracer tests provide the best tools for delineating directions and rates of transport in karst systems. Although some studies have been able to calibrate models to reproduce tracer breakthrough curves (Teutsch, 1993), the level of detail required for input and calibration of such models far exceeds the level of data availability for most karst aquifers. Some have suggested that if information on direction and rates is obtained from tracer tests, modeling is superfluous.

3.5. Future studies

Future studies should consider a variety of improvements to the existing models, including additional data collection, different conceptual model design, and other factors. The current distributed parameter model assumes that recharge is distributed uniformly along streams; however, this is unrealistic. Recent field studies have been conducted to identify major features along the streams that would focus recharge. Future modeling should consider focusing recharge at different points on the basis of field data, and sensitivity of model output to various distributions should be examined. Parameterization of the distributed model depends on accurate information about hydraulic head. Future studies should improve the reliability of the head data by accurately locating wells and measuring the surface elevation. A greater number and wider distribution of head measurements would also improve parameterization of the distributed model. Inverse modeling could be used to guide location of additional head measurements. Synoptic water level maps should also be developed for high flow periods to evaluate model performance under these conditions.

The current distributed parameter model represents the aquifer using a single layer. Some of the transient simulations had difficulties converging because cells went dry. This problem was overcome in the current study by lowering the base of the model and assuming that the Edwards aquifer is hydraulically connected to the underlying Glen Rose Formation. Sensitivity analyses indicated that varying specific storage could improve simulations of water levels in the southern monitoring wells. Future modeling should consider representing the aquifer as two layers that would allow vertical variation in hydraulic properties and use of different specific storage values to reduce water level fluctuations during high flows without impacting simulation of low spring discharge. Variations in hydraulic conductivity with depth, particularly near the springs, could also improve simulation of low spring discharges, as shown by the lumped parameter model.

Distribution of major conduits based on dye tracing studies should be approximated in the model using zones of high hydraulic conductivity. The model grid should be refined near these zones to better represent the conduits. Inclusion of such high conductivity zones should improve simulations of troughs in the potentiometric surface (Fig. 3).

Both types of models should be updated to modify recharge values for Barton Creek since a new gaging station has been installed downstream of the outcrop zone. Recent dye tracing studies revealed that much of the flow in the northwest segment of the aquifer discharges at Cold Springs. This information should also be incorporated into revised models.

4. Conclusions

This study showed that equivalent porous media models could be used to simulate regional groundwater flow in this highly karstified aquifer. Both distributed and lumped parameter models generally reproduced temporal variations in discharge at Barton Springs with RMS errors of about 10% of the discharge fluctuations. Therefore, if the primary goal of the modeling study is to simulate spring discharge for water management purposes, either distributed or lumped parameter models can be used.

Parameterization of the distributed model for steady state was restricted to estimation of hydraulic conductivity because groundwater recharge could be accurately estimated from stream losses. The zonal distribution of hydraulic conductivity could be estimated from information on the hydraulic gradient, which resulted in low conductivities in the unconfined section of the aquifer and high conductivities near the spring where flow is concentrated. Automated inverse modeling was used to further improve the simulation of heads in the aquifer. RMS error for the steady state model of 7 m represents errors in measured and simulated heads. The transient model required very little calibration. Specific yield in the unconfined aquifer was reduced to better simulate low flow discharges in Barton Springs.

Sensitivity analyses indicated that the steady state distributed parameter model is most sensitive to variations in recharge and hydraulic conductivity and fairly insensitive to pumpage and spring drain conductance. These results suggest that future management should consider techniques such as enhanced recharge to buffer the system against future droughts.

Sensitivity analyses indicated that temporal variability in spring discharge is most sensitive to variations in recharge and relatively insensitive to variations in pumpage, specific yield, and specific storage in the -10 to $+50\%$ range. These results suggest that curtailing pumpage during droughts may not be adequate to maintain spring flow and that artificial recharge may also be required. Increasing specific storage by a factor of 10 greatly reduced temporal variability in spring discharge and water level fluctuations.

The distributed parameter model generally reproduced potentiometric surfaces at different times. RMS errors ranged from 8.7 to 11.2 m, which represent about 10% of the water level variations across the aquifer.

Both distributed and lumped parameter models simulated trends in water level fluctuations more accurately than absolute values of water levels in continuously monitored wells. The lumped parameter model could simulate water levels only in wells chosen to represent each cell in the model.

Impact of groundwater pumping on spring discharge can be evaluated using either distributed or lumped parameter modeling approaches; however, detailed evaluation of the effect of pumping on a more local scale, such as a large well field, requires a distributed parameter model.

The main limitations of equivalent porous media models is that they cannot accurately simulate direction or rate of water flow in the aquifer, which precludes them from simulating point source contamination or delineating aquifer protection zones. It is questionable whether any model can simulate these processes because of the complexity of karst systems.

Results of this study show the ability of equivalent porous media distributed and lumped parameter models to simulate regional groundwater flow, which is critical for managing water resources in karst aquifers and predicting the impact of future pumping and potential future drought conditions on spring flow.

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References

Anderson, M.P., Woessner, W.W., 1992. Applied Groundwater Modeling, Simulation of Flow and Advective Transport, Academic Press, New York, p. 381.

Angelini, P., Dragoni, W., 1997. The problem of modeling limestone springs: the case of Bagnara (North Apennines, Italy). Ground Water 35 (4), 612–618.

Barrett, M.E., 1996. A parsimonious model for simulation of flow and transport in a karst aquifer. PhD Thesis, The University of Texas at Austin, Texas, p. 180.

Barrett, M.E., Charbeneau, R.J., 1996. A parsimonious model for simulation of flow and transport in a karst aquifer. Center for Research in Water Resources, Technical Report 269, p. 149.

Barrett, M.E., Charbeneau, R.J., 1997. A parsimonious model for simulating flow in a karst aquifer. J. Hydrol. 196, 47–65.

Bear, J., Tsang, C.-F., de Marsily, G., 1993. Flow and Contaminant Transport in Fractured Rock, Academic Press, San Diego, p. 560.

Davies, G.J., Quinlan, J.F., Worthington, S.R.H., 1992. Carbonate aquifers; distinctions between conduit flow or diffuse flow should be abandoned in their classification and design of monitoring systems. Geol. Soc. Am., Abs. Programs 24 (7), 253.

Greene, E.A., Shapiro, A.M., Carter, J.M., 1999. Hydrogeologic characterization of the Minnelusa and Madison aquifers near Spearfish, South Dakota. US Geol. Surv., Water Resour. Inv. Rept. 98-4156, 64.

Guyton, W.F., and Associates, 1958. Recharge to the Edwards reservoir between Kyle and Austin. Report prepared for the City Water Board, San Antonio, TX, p. 9.

Halihan, T., Wicks, C.M., 1998. Modeling of storm responses in conduit flow aquifers with reservoirs. J. Hydrol. 208, 82–91.

Harbaugh, A.W., McDonald, M.G., 1996. User's documentation for MODFLOW-96, an update to the US Geological Survey modular finite-difference ground-water flow model. US Geol. Surv., Open-File Rep. 96-485, 56.

Hauwert, N.M., Johns, D.A., Sansom, J.W., Aley, T.J., 2002. Groundwater tracing of the Barton Springs Edwards aquifer, Travis and Hays counties, Texas. Gulf Coast Assoc. Geol. Soc. Trans. 52, 377–384.

Hovorka, S.D., Mace, R.E., Collins, E.W., 1998. Permeability structure of the Edwards aquifer, South Texas—Implications for aquifer management. University of Texas at Austin, Bur. Econ. Geol., Rep. Inv. No. 250, p. 55.

Hsieh, P.A., Freckleton, J.R., 1993. Documentation of a computer program to simulate horizontal-flow barriers using the US Geological Survey's modular three-dimensional finite-difference ground-water flow model. US Geol. Surv., Open File Rep. 92-477.

Huntoon, P.W., 1994. Is it appropriate to apply porous media groundwater circulation models to karstic aquifers? In: El-Kadi, A.I., (Ed.), Groundwater Models for Resources Analysis and Management, 1994 Pacific Northwest/Oceania Conference, Honolulu, HI, pp. 339–358.

Jeannin, P.-Y., 2001. Modeling flow in phreatic and epiphreatic karst conduits in the Holloch Cave (Muotatal, Switzerland). Water Resour. Res. 37, 191–200.

Keeler, R.R., Zhang, Y.K., 1997. Modeling of groundwater flow in a fractured-karst aquifer in the Big Springs Basin, Iowa. Geol. Soc. Am., Abs. Programs 29 (4), 25.

Knochemus, L.A., Robinson, J.L., 1996. Descriptions of anisotropy and heterogeneity and their effect on ground-water flow and areas of contribution to public supply wells in a karst carbonate aquifer system. US Geol. Surv., Water Supply Paper 2475, 47.

Kuniansky, E.L., 1993. Multilayer finite-element model of the Edwards and Trinity aquifers, Central Texas. In: Dutton, A.R., (Ed.), Multilayer Finite-Element Model of the Edwards and Trinity Aquifers, American Institute of Hydrology, pp. 234–249.

Larkin, T.J., Bomar, G.W., 1983. Climatic Atlas of Texas. Publication LP-192, Department of Water Resources, Austin, TX, p. 151.

Larocque, M., Banton, O., Ackerer, P., Razack, M., 1999. Determining karst transmissivities with inverse modeling and an equivalent porous media. Ground Water 37, 897–903.

National Research Council, 1996. Rock Fractures and Fluid Flow, Contemporary Understanding and Applications: National Research Council, National Academy Press, Washington, DC, p. 551.

Neuman, S.P., 1987. Stochastic continuum representation of fractured rock permeability as an alternative to the REV and fracture network concepts. In: Custodio, E., Gurgui, A., Lobo-Ferreira, J.P. (Eds.), NATO Advanced Workshop on Advances in Analytical and Numerical Groundwater Flow and Quality Modelling, NATO ASI Series, Series C: Mathematical and Physical Sciences, vol. 224. Reidel Publications, Dordrecht, pp. 331–362.

Pankow, J.F., Johnson, R.L., Hewetson, J.P., Cherry, J.A., 1986. An evaluation of contaminant migration patterns at two waste disposal sites on fractured porous media in terms of the equivalent porous medium (EPM) model. J. Contaminant Hydrol. 1, 65–76.

Poeter, E.P., Hill, M.C., 1998. Documentation of UCODE, A computer code for universal inverse modeling. US Geol. Surv., Water-Resour. Inv. Rep. 98-4080, 116.

Quinlan, J.F., Davies, G.J., Jones, S.W., Huntoon, P.W., 1996. The applicability of numerical models to adequately characterize ground-water flow in karstic and other triple-porosity aquifers. In: Ritchy, J.D., Rumbaugh, J.O. (Eds.), Subsurface Fluid-Flow (Ground-Water and Vadose Zone) Modeling, ASTM STP 1288, American Society for Testing and Materials, pp. 114–133.

Ryder, P.D., 1985. Hydrology of the Floridan aquifer system in west-central Florida. US Geol. Surv. Prof. Paper P 1403-F, F1–F63.

Senger, R., Kreitler, C., 1984. Hydrogeology of the Edwards Aquifer, Austin Area, Central Texas. University of Texas at Austin, Bur. Econ. Geol., Rept. Inv. No. 141, p. 35.

Slade, R.M., Ruiz, L., 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. Inv. Rep. 85-4299, 49.

Small, T.A., Hanson, J.A., Hauwert, N.M., 1996. Geologic framework and hydrogeologic characteristics of the Edwards Aquifer outcrop (Barton Springs segment), Northeastern Hays and Southwestern Travis counties, Texas. US Geol. Surv., Water Resour. Inv. Rep. 96-4306, 15.

Teutsch, G., 1993. An extended double-porosity concept as a practical modelling approach for a karstified terrain. Hydrogeol. Processes in Karst Terranes, Proc. of the Antalya Symp. And Field Seminar, Oct. 1990, Intl. Assoc. Hyd. Sci. Publ. No. 207, p. 281–292.

Teutsch, G., Sauter, M., 1991. Groundwater Modeling in Karst Terranes: Scale Effects, Data Acquisition and Field Observations, Proceedings of the Third Conference on Hydrogeology, Ecology, Monitoring, and Management of Ground Water in Karst Terranes, Nashville, Tennessee, National Ground Water Association, Dublin, Ohio, pp. 17–34.

Teutsch, G., Sauter, M., 1998. Distributed parameter modelling approaches in karst-hydrological investigations. Bull. d'Hydrogeologie No. 16, 99–109.

Wanakule, N., Anaya, R.A, 1993. A lumped parameter model for the Edwards aquifer. Texas Water Resources Institute, Technical Report No. 163, p. 84.

Warren, J.E., Root, P.J., 1963. The behaviour of naturally fractured reservoirs. Soc. Pet. Engng J. 3, 245–255.

Worthington, S.R.H., Schindel, G.M., Alexander, C.E.J., 2001. Aquifer scale properties for hydraulic characterization of carbonate aquifers. Geol. Surv. Am., Abs. Programs, A-411.

Yurtsever, Y., Payne, R.B., 1986. Mathematical Models Based on Compartmental Simulation Approach for Quantitative Interpretation of Tracer Data in Hydrological Systems, Proceedings of the 5th International Symposium of Underground Water Tracing, IGME, Athens, Greece, pp. 341–353.

Zhang, Y.-K., Bai, E.-W., Libra, R., Rowden, R., Liu, H., 1996. Simulation of spring discharge from a limestone aquifer in Iowa. Hydrogeol. J. 4, 41–54.