Diffuse-Flow Conceptualization and Simulation of the Edwards Aquifer, San Antonio Region, Texas

By R.J. Lindgren

In cooperation with the San Antonio Water System

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Abstract

A numerical ground-water-flow model (hereinafter, the conduit-flow Edwards aquifer model) of the karstic Edwards aquifer in south-central Texas was developed for a previous study on the basis of a conceptualization emphasizing conduit development and conduit flow, and included simulating conduits as one-cell-wide, continuously connected features. Uncertainties regarding the degree to which conduits pervade the Edwards aquifer and influence ground-water flow, as well as other uncertainties inherent in simulating conduits, raised the question of whether a model based on the conduit-flow conceptualization was the optimum model for the Edwards aquifer. Accordingly, a model with an alternative hydraulic conductivity distribution without conduits was developed in a study conducted during 2004–05 by the U.S. Geological Survey, in cooperation with the San Antonio Water System. The hydraulic conductivity distribution for the modified Edwards aquifer model (hereinafter, the diffuse-flow Edwards aquifer model), based primarily on a conceptualization in which flow in the aquifer predominantly is through a network of numerous small fractures and openings, includes 38 zones, with hydraulic conductivities ranging from 3 to 50,000 feet per day. Revision of model input data for the diffuse-flow Edwards aquifer model was limited to changes in the simulated hydraulic conductivity distribution. The root-mean-square error for 144 target wells for the calibrated steady-state simulation for the diffuse-flow Edwards aquifer model is 20.9 feet. This error represents about 3 percent of the total head difference across the model area. The simulated springflows for Comal and San Marcos Springs for the calibrated steady-state simulation were within 2.4 and 15 percent of the median springflows for the two springs, respectively. The transient calibration period for the diffuse-flow Edwards aquifer model was 1947–2000, with 648 monthly stress periods, the same as for the conduit-flow Edwards aquifer model. The root-mean-square error for a period of drought (May–November 1956) for the calibrated transient simulation for 171 target wells is 33.4 feet, which represents about 5 percent of the total head difference across the model area. The root-mean-square error for a period of above-normal rainfall (November 1974–July 1975) for the calibrated transient simulation for 169 target wells is 25.8 feet, which represents about 4 percent of the total head difference across the model area. The root-mean-square error ranged from 6.3 to 30.4 feet in 12 target wells with long-term water-level measurements for varying periods during 1947–2000 for the calibrated transient simulation for the diffuse-flow Edwards aquifer model, and these errors represent 5.0 to 31.3 percent of the range in water-level fluctuations of each of those wells. The root-mean-square errors for the five major springs in the San Antonio segment of the aquifer for the calibrated transient simulation, as a percentage of the range of discharge fluctuations measured at the springs, varied from 7.2 percent for San Marcos Springs and 8.1 percent for Comal Springs to 28.8 percent for Leona Springs. The root-mean-square errors for hydraulic heads for the conduit-flow Edwards aquifer model are 27, 76, and 30 percent greater than those for the diffuse-flow Edwards aquifer model for the steady-state, drought, and above-normal rainfall synoptic time periods, respectively. The goodness-of-fit between measured and simulated springflows is similar for Comal, San Marcos, and Leona Springs for the diffuse-flow Edwards aquifer model and the conduit-flow Edwards aquifer model. The root-mean-square errors for Comal and Leona Springs were 15.6 and 21.3 percent less, respectively, whereas the root-mean-square error for San Marcos Springs was 3.3 percent greater for the diffuse-flow Edwards aquifer model compared to the conduit-flow Edwards aquifer model. The root-mean-square errors for San Antonio and San Pedro Springs were appreciably greater, 80.2 and 51.0 percent, respectively, for the diffuse-flow Edwards aquifer model. The simulated water budgets for the diffuse-flow Edwards aquifer model are similar to those for the conduit-flow Edwards aquifer model. Differences in percentage of total sources or discharges for a budget component are 2.0 percent or

less for all budget components for the steady-state and transient simulations. The largest difference in terms of the magnitude of water budget components for the transient simulation for 1956 was a decrease of about 10,730 acre-feet per year (about 2 percent) in springflow for the diffuse-flow Edwards aquifer model compared to the conduit-flow Edwards aquifer model. This decrease in springflow (a water budget discharge) was largely offset by the decreased net loss of water from storage (a water budget source) of about 10,500 acre-feet per year.

Introduction

The Edwards aquifer in the Balcones fault zone of south-central Texas (fig. 1) is one of the most permeable and productive aquifers in the world. The aquifer consists of regionally extensive carbonate rocks that crop out within the Edwards Plateau and the Balcones fault zone and underlie the Gulf Coastal Plain. The northern aquifer boundary is defined by the updip limit of contiguous, outcropping rocks of the Edwards Group, Georgetown Formation, and their westward stratigraphic equivalents (Edwards rocks). The southern aquifer boundary usually is defined by what commonly is referred to as the freshwater/saline-water interface and defined on maps as the 1,000-milligrams per liter (mg/L) dissolved solids concentration line (fig. 2). This interface marks the northern boundary of the freshwater/saline-water transition zone, the zone of the aquifer between the 1,000- and 10,000-mg/L dissolved solids concentration lines (fig. 2). The San Antonio segment of the aquifer primarily includes parts of Kinney, Uvalde, Medina, Bexar, Comal, and Hays Counties that lie within and adjacent to the Balcones fault zone (fig. 1). This segment is bounded on the west and east by ground-water divides near Brackettville and Kyle, respectively, and contains the most productive and transmissive parts of the aquifer. The Barton Springs segment of the aquifer includes parts of Hays and Travis Counties and is bounded on the southwest by the ground-water divide near Kyle and on the northeast by the Colorado River. The San Antonio segment of the aquifer discharges primarily to Comal and San Marcos Springs, whereas the Barton Springs segment discharges primarily to Barton Springs (fig. 1).

A numerical ground-water-flow model (hereinafter, the original or conduit-flow Edwards aquifer model) of the karstic Edwards aquifer in south-central Texas was completed as part of a study conducted during 2000–2003 by the U.S. Geological Survey (USGS) and The University of Texas at Austin, Bureau of Economic Geology (BEG), in cooperation with the U.S. Department of Defense and the Edwards Aquifer Authority (Lindgren and others, 2004). Karst aquifers can be conceptualized as including a discrete conduit network with frequently turbulent flow conditions, with conduits representing the major flow paths in the aquifer. The conceptualization that served as the basis for the original Edwards aquifer model emphasizes conduit development and conduit flow. As conduit flow might be frequently turbulent and restricted to discrete pathways, MODFLOW (Harbaugh and others, 2000) is not designed for the simulation of conduit flow. Use of a distributed, porous-media model such as MODFLOW to simulate flow in a karst system is a simplification of the flow system. As a way to represent conduits, other than by use of a coupled-continuum pipe flow or dual-porosity or triple-porosity model (Birk and others, 2003; Liedl and others, 2003), conduits were simulated in the original Edwards aquifer model by narrow (one-cell, 0.25-mile [mi] wide), initially continuously connected zones with large hydraulic conductivities (Lindgren and others, 2004).

Although there is evidence to support the conduit-flow conceptualization, the degree to which conduits pervade the Edwards aquifer and influence ground-water flow remains uncertain. An alternative conceptualization, which can be called the diffuse-flow conceptualization, reflects the hypothesis that, although conduits likely are present, flow in the aquifer predominantly is through a network of small fractures and openings sufficiently numerous that the aquifer can be considered a porous-media continuum at the regional scale. Which is the more realistic conceptualization—in other words, whether conduit flow or diffuse flow predominates at the regional scale—is an open question. Development of the original Edwards aquifer model incorporating the conduit-flow conceptualization thus can be considered a test of one reasonable conceptualization. Previous numerical, distributed, porous-media, ground-water-flow models of the Edwards aquifer (Klemt and others, 1979; Maclay and Land, 1988; Thorkildsen and McElhaney, 1992; Scanlon and others, 2002) and other karstic carbonate aquifers (for example, the Floridan aquifer system in Florida and in parts of Georgia, South Carolina, and Alabama) (Knowles and others, 2002; Sepulveda, 2002; Payne and others, 2005) have reasonably simulated measured fluctuations in water levels in wells and springflow.

The hydraulic conductivity distribution for the conduit-flow Edwards aquifer model includes two components: (1) a base hydraulic conductivity distribution based on nonparametric geostatistics, stochastic simulation, and numerical flow simulation and (2) a network of conduits, simulated as continuously connected (other than a break in eastern Uvalde and southwestern Medina Counties), one-cell-wide (1,320 feet [ft]) zones with very large hydraulic conductivities (as much as 300,000 feet per day [ft/d]) (Lindgren and others, 2004). The simulated flow directions generally are toward the nearest conduit and subsequently along the conduits from the recharge zone into the confined zone and toward the major springs. However, the locations of the conduits were inferred from a qualitative study (Worthington, 2004) and are subject to some uncertainty. The effect of the uncertainty regarding the locations of conduits becomes more important as the size of the area of interest decreases, or, in other words, as the scale of the simulation decreases. Uncertainty also exists regarding the physical dimensions, connectivity, and hydraulic properties of conduits. The physical dimensions of the conduits in the conduit-flow Edwards aquifer model are constrained by the model cell dimensions, with most conduits being much smaller than the 0.25-mi dimensions of the model cells.

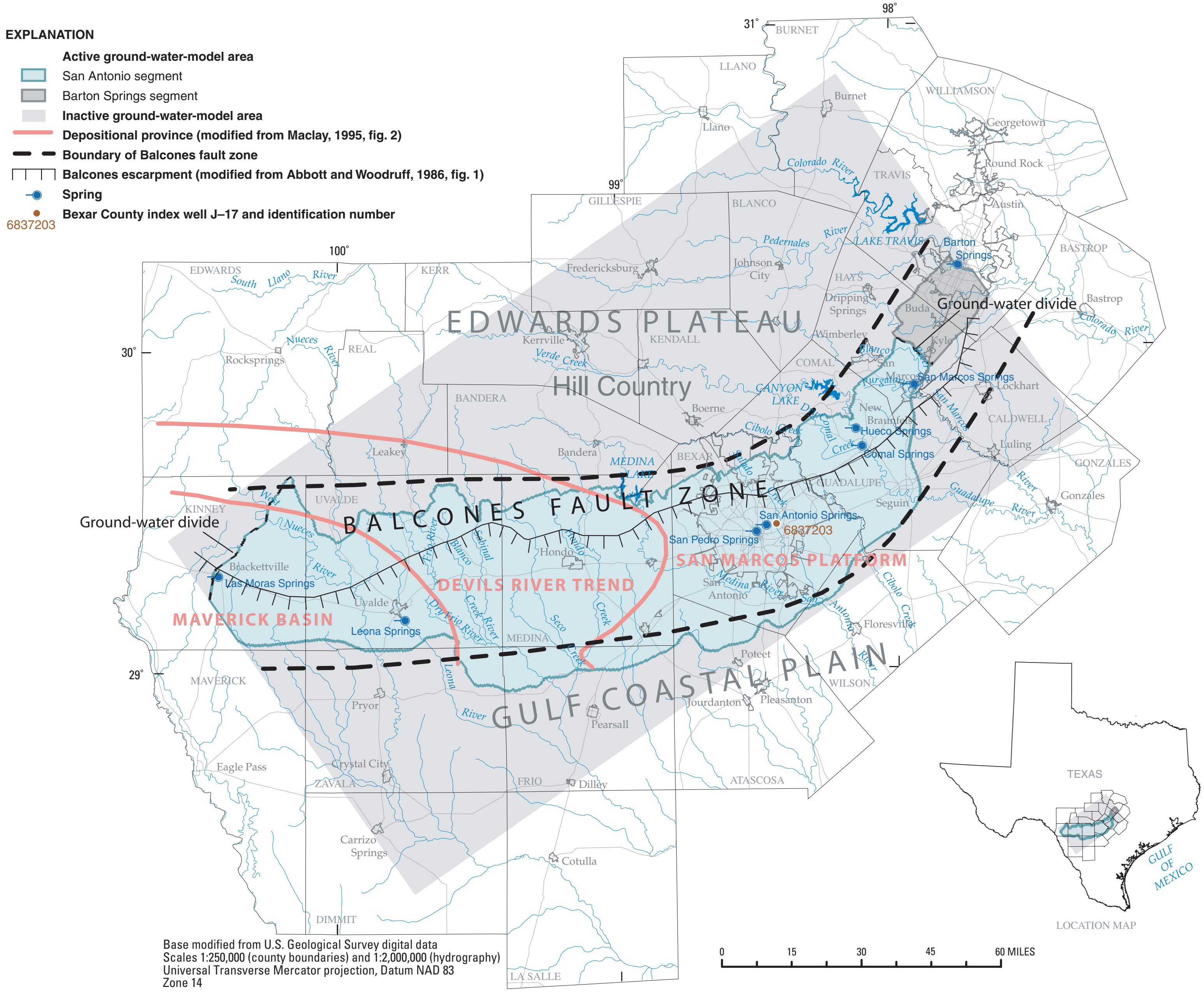


Figure 1. Location of ground-water-flow model area, Edwards aquifer segments, and physiographic regions, San Antonio region, Texas.

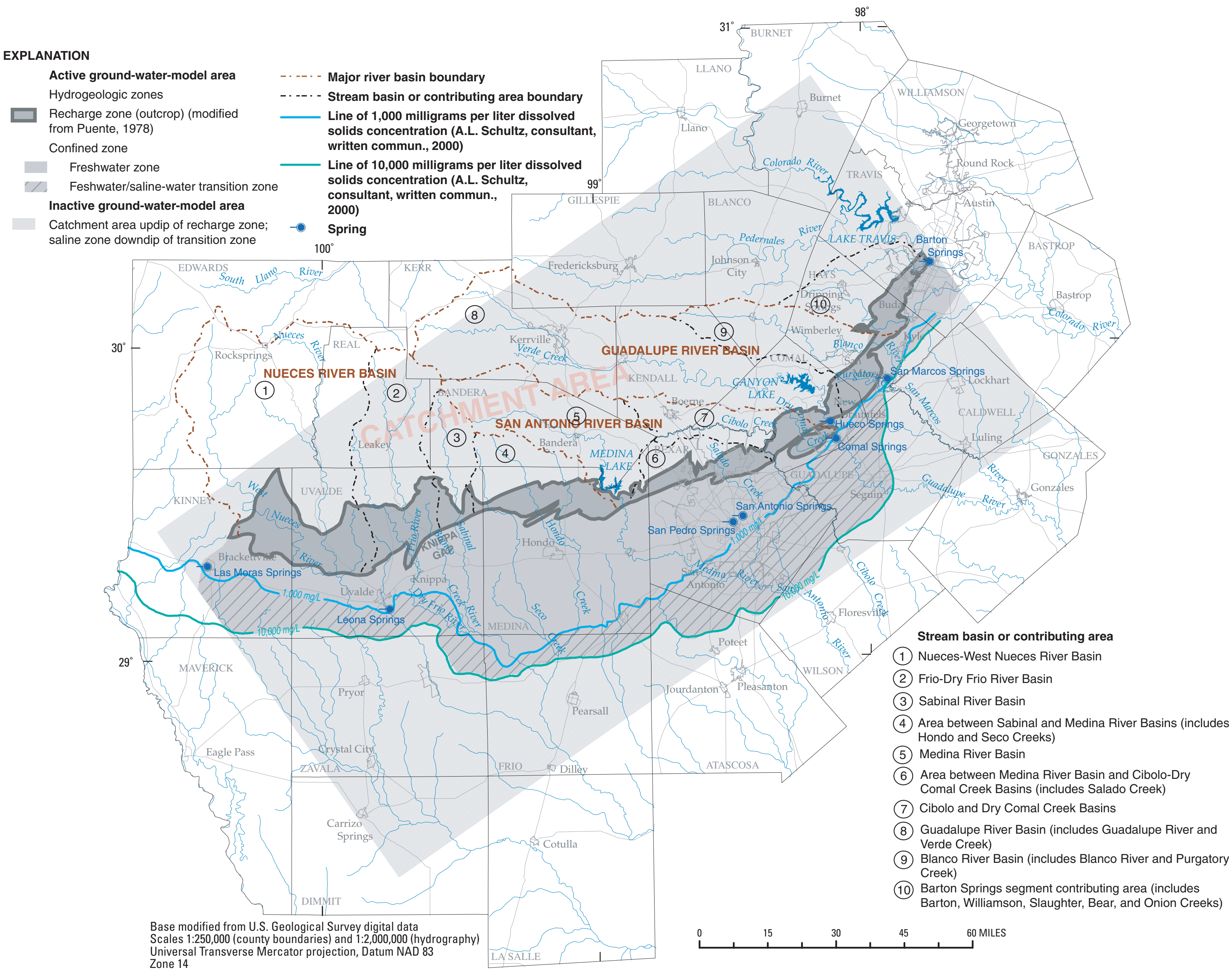


Figure 2. Hydrogeologic zones and catchment area (upper parts of stream basins that contribute recharge) of the Edwards aquifer, San Antonio region, Texas.

Calibration of the conduit-flow Edwards aquifer model indicated that zones of comparatively high hydraulic conductivity are needed to convey water toward the major springs and simulate the large measured springflows at Comal and San Marcos Springs (fig. 2). However, steady-state simulations for the conduit-flow Edwards aquifer model indicated that wider zones with relatively high hydraulic conductivity, but appreciably lower than the hydraulic conductivity used for the conduits, also could be used.

The uncertainties inherent in simulating conduits as one-cell-wide, continuously connected features raised the question of whether a model based on this mode of conduit simulation was the optimum model for the Edwards aquifer. Accordingly, a model with a hydraulic conductivity distribution without conduits explicitly simulated was considered. To develop an alternative, nonconduit-flow or diffuse-flow hydraulic conductivity distribution for the Edwards aquifer model, the USGS in cooperation with the San Antonio Water System, conducted a study during 2004–05. The objectives of this study were to (1) modify the hydraulic conductivity distribution of the conduit-flow Edwards aquifer model to replace conduits with broad zones of upscaled (adjusted from field-measurement scale to model-cell scale) hydraulic conductivity and (2) compare the hydrographs and residuals (differences between model-computed and measured values) for hydraulic heads and springflows for the conduit-flow Edwards aquifer model with those for the model containing the nonconduit, diffuse-flow hydraulic conductivity distribution (hereinafter, the diffuse-flow Edwards aquifer model). The hydraulic conductivity distribution that incorporates the nonconduit, diffuse-flow conceptualization of the aquifer emphasizes small-conduit and fracture flow rather than large, interconnected-conduit flow.

Purpose and Scope

This report describes the diffuse-flow conceptualization and simulation of the Edwards aquifer. Specifically, (1) the diffuse-flow hydraulic conductivity distribution with broad zones of upscaled hydraulic conductivity, (2) the calibration of the diffuse-flow Edwards aquifer model, and (3) the results of comparisons of the hydrographs and residuals for hydraulic heads and springflows for the conduit-flow Edwards aquifer model (with simulated one-cell-wide conduits) and the diffuse-flow Edwards aquifer model (with simulated broad zones of upscaled hydraulic conductivity). In addition, the simulated water budgets for the diffuse-flow Edwards aquifer model are presented, discussed, and compared with the water budgets for the conduit-flow Edwards aquifer model.

The Edwards aquifer area includes both the San Antonio and Barton Springs segments of the aquifer and all or parts of 11 counties in south-central Texas (fig. 1). The area of active model cells for the diffuse-flow Edwards aquifer model is the same as for the conduit-flow Edwards aquifer model, defined on the north by the northern limit of contiguous outcropping Edwards rocks (updip boundary of the recharge zone) and on the south by the 10,000-mg/L dissolved solids concentration line (downdip boundary of the freshwater/saline-water transition zone [the zone of brackish water between the 1,000- and 10,000-mg/L lines of dissolved solids concentration]) (fig. 2) (A.L. Schultz, consultant, written commun., 2000). Thus, the active model area includes the transition zone between freshwater and saline water (fig. 2). The diffuse-flow Edwards aquifer model was calibrated for steady-state (1939–46) and transient (1947–2000) conditions, with monthly stress periods, as was the conduit-flow Edwards aquifer model.

As for the conduit-flow Edwards aquifer model, although the Barton Springs segment of the aquifer is included in the model, calibration was not done for the diffuse-flow Edwards aquifer model for the northern part of the segment (Travis County). A numerical finite-difference ground-water-flow model recently was completed for the Barton Springs segment (Scanlon and others, 2002), and a duplication of that work was not considered necessary. However, calibration was done in the southern part of the Barton Springs segment (northern Hays County) because the simulated hydraulic heads and flows in that area influenced the location of the simulated ground-water divide near Kyle and the simulated hydraulic heads and flows in the adjoining part of the San Antonio segment of the aquifer.

Revision of model properties for the diffuse-flow Edwards aquifer model was limited to changes in the simulated hydraulic conductivity distribution. The other model properties, including the aquifer structure (top and bottom altitudes), fault locations and conductivity, recharge, pumpage, and drain properties representing the major springs, are the same as for the conduit-flow Edwards aquifer model.

There is uncertainty with regard to the capability of any numerical ground-water-flow model to simulate conditions in the real ground-water-flow system. To assess the importance of uncertainty associated with the conduit-flow Edwards aquifer model, a series of sensitivity tests was made to ascertain how the model results were affected by variations greater than and less than the calibrated values of input data (Lindgren and others, 2004). The sensitivity of the model results to variations in recharge, withdrawals, hydraulic conductivity, spring-orifice conductance for Comal and San Marcos Springs, and northern boundary inflow was tested with steady-state and transient simulations. Separate sensitivity tests were not made for the diffuse-flow Edwards aquifer model because (1) the only model input data that differs from the conduit-flow Edwards aquifer model is hydraulic conductivity, and (2) sensitivity tests were beyond the scope and time constraints of this study.

Acknowledgments

Information for this report was synthesized from the published reports and data of several agencies, companies, and institutions. Chief contributors of the information for this report are the USGS, the BEG, and the Texas Water Development Board (TWDB). The author is indebted to Robert Mace (TWDB), Larry Land (HDR, Engineering), and Alvin Schultz

(consultant) for their suggestions regarding the diffuse-flow hydraulic conductivity distribution.

Conceptualization of the Edwards Aquifer

The conceptualization of the Edwards aquifer includes a description of the geologic and hydrogeologic setting within which the aquifer occurs. The geologic and hydrogeologic setting for the Edwards aquifer is discussed in detail in Lindgren and others (2004). The conceptualization presented in that report, which is the basis for the conduit-flow Edwards aquifer model, emphasizes conduit development and conduit flow. The hydraulic conductivity distribution simulated in the Edwards aquifer model of this report is based on the diffuse-flow conceptualization. For this report, the discussion of the conceptualization of the Edwards aquifer will be restricted to the hydraulic conductivity and transmissivity of the aquifer and ground-water flow.

Hydraulic Conductivity and Transmissivity

Hydraulic conductivity and transmissivity of the Edwards aquifer reflect matrix, fracture, and conduit permeability and each varies over several orders of magnitude. Hovorka and others (1998) reported that hydraulic conductivity ranges from 10^{-3} to 10^{5} ft/d and transmissivity from 10^{-1} to 10^{7} feet squared per day (ft^2/d) on the basis of specific-capacity and other aquifer tests. Garza (1968, p. 31) estimated the transmissivity in the confined zone of the aquifer in the San Antonio region to be 1 to 2 million ft^2/d. On the basis of numerical modeling, Maclay and Land (1988) estimated transmissivities of more than 4.3 million ft^2/d in Comal County near Comal Springs in the confined freshwater zone of the aquifer; their smallest estimated transmissivity was 130 ft^2/d in the freshwater/saline-water transition zone. The transmissivities for most of the confined freshwater zone range from 430,000 to 2.2 million ft^2/d, but in the recharge (unconfined) zone transmissivities generally are less than 430,000 ft^2/d (Maclay and Land, 1988).

Hydraulic Conductivity Distribution Developed for Conduit-Flow Edwards Aquifer Model

Painter and others (2002) estimated hydraulic conductivity for the Edwards aquifer in the San Antonio region to provide initial values for the conduit-flow Edwards aquifer model. The broad objective was to provide representations of the areal distribution of vertically averaged hydraulic conductivities across the San Antonio region of the Edwards aquifer using the best available quantitative techniques.

Hydraulic conductivity in heterogeneous aquifers depends on the spatial scale of the measurement. Existing hydraulic conductivity measurements in the Edwards aquifer are derived mostly from single-well aquifer tests. These measurements must be modified or "upscaled" before being applied to the 0.25- by 0.25-mi cells of the Edwards aquifer model. Painter and others (2002) first derived a hydraulic conductivity distribution from a dataset (hereinafter, the Mace dataset) with a few data from multiple-well aquifer tests but with most of the data from single-well tests (Hovorka and others, 1998; Mace, 2000; Mace and Hovorka, 2000). An approach based on nonparametric geostatistics, stochastic simulation, and numerical flow simulation then was used to upscale and interpolate to the Edwards aquifer model grid. This constituted revision 1 of the hydraulic conductivity distribution. Revision 2 of the hydraulic conductivity distribution incorporated the use of measured hydraulic heads and an approach based on Bayesian statistics to infer hydraulic conductivity. However, revision 2 was a preliminary application of the Bayesian technique and was not used because of subsequent updates to the recharge and measured hydraulic-head datasets for the original Edwards aquifer model. Revision 2 is not discussed in this report. Revision 3 of the hydraulic conductivity distribution represents a further refinement of the approach and was the distribution used in the conduit-flow Edwards aquifer model.

Initially, the only manipulation of the Mace dataset by Painter and others (2002) was to geometrically average values when multiple values (representing different tests) existed for the same well. After this averaging, the dataset contained 653 values of hydraulic conductivity in the confined zone and 108 values in the recharge (unconfined) zone. Univariate statistical distributions of hydraulic conductivity data for the confined and unconfined zones of the Edwards aquifer are reasonably well approximated as lognormal, although the distribution for the confined zone does have a lower tail that is enhanced relative to the lognormal distribution (Painter and others, 2002, fig. 2-2). The mean and variance for the confined and unconfined zones of the aquifer are substantially different, with geometric means of 18.8 and 1.3 ft/d for the confined and unconfined zones, respectively. The logarithmic variance in hydraulic conductivity is 6.4 and 9.7 ft/d for the confined and unconfined zones, respectively. Data limitations cited by Painter and others (2002) for the single-well tests include data uncertainty, drawdown that is below the limit of measurement and recorded as zero (15 percent of the tests), and imprecise well locations.

Revision 1 of the hydraulic conductivity distribution was derived using a simulation approach that addresses data interpolation and the issue of scale consistency in hydraulic conductivity (Painter and others, 2002). Data interpolation was necessary because (1) the number of grid cells used in the Edwards aquifer model is much greater than the number of data points, and (2) the centers of grid cells do not necessarily correspond with well locations where measurements of hydraulic conductivity are available. Scale consistency is an issue because hydraulic conductivity in heterogeneous formations depends on the scale over which it is defined. A systematic bias toward lower hydraulic conductivity would be introduced by the unaltered application of local-scale hydraulic conductivity derived from aquifer tests to the 0.25-mi grid cells of the Edwards aquifer model. To address the scale dependencies and thereby avoid this systematic bias, a geostatistical approach was combined with numerical simulations in developing the hydraulic conductivity distribution. See Painter and others (2002) for a more detailed discussion of this method.

Hydraulic-head data imply considerable information about the underlying hydraulic conductivity distribution. In revision 3 of the hydraulic conductivity distribution, revision 1 was taken as a starting point and then modified to be more consistent with measured hydraulic-head data (Painter and others, 2002). Specifically, a recently developed Bayesian updating procedure (Woodbury and Ulrych, 1998, 2000) was used to update the hydraulic conductivity distribution. In this approach, the nonunique nature of the inverse problem is explicitly acknowledged, and results are given in terms of probability distributions for the hydraulic conductivity in each cell. In addition, the Bayesian method allows prior information of various types to be incorporated into the inversion procedure, which allows the previous work on upscaled hydraulic conductivity (revision 1) to be retained and used in the inversion. The model properties were assumed to be random, and the inversion approached from the viewpoint of probability theory, with Bayesian solutions being sought for the problem. See Painter and others (2002) for a more detailed discussion of this method. The estimated hydraulic conductivity for revision 3 ranges from 1 to 7,347 ft/d (Lindgren and others, 2004, fig. 8).

Hydraulic Conductivity Distribution Developed for Diffuse-Flow Edwards Aquifer Model

A hydraulic conductivity distribution based primarily on the diffuse-flow conceptualization was developed for use in the Edwards aquifer model. For the Barton Springs segment of the aquifer, the hydraulic conductivity distribution of Scanlon and others (2002) was used. The initial diffuse-flow hydraulic conductivity distribution that was developed for the San Antonio segment of the aquifer comprised 29 zones and was based on a synthesis of information from multiple sources, including previous reports, available aquifer tests, mapping of fractures/caverns observed in wells (A.L. Schultz, consultant, written commun., 2004), geologic structures, hydraulic gradients, and geochemistry. The primary basis for the diffuse-flow hydraulic conductivity distribution differed for each of the three hydrogeologic zones (recharge zone, confined freshwater zone, and confined freshwater/saline-water transition zone). The initial hydraulic conductivity zones and assigned values for the recharge zone were derived from Maclay and Land (1988, fig. 10), who reported relative transmissivity values by subareas, or zones, for the San Antonio segment of the Edwards aquifer. These estimates of relative transmissivity were made by Maclay and Small (1984) and were based on available geologic, hydrochemical, and hydrologic information. Eight zones were delineated in the recharge zone for the diffuse-flow hydraulic conductivity distribution, with assigned hydraulic conductivities ranging from 20 to 70 ft/d.

The initial hydraulic conductivity zones and assigned values for the confined freshwater zone of the Edwards aquifer were derived from revision 1 of the hydraulic conductivity distribution developed by Painter and others (2002) and described in the previous section. For the diffuse-flow hydraulic conductivity distribution, the cell-by-cell values for revision 1 of the hydraulic conductivity distribution were contoured, and hydraulic conductivity zones for the confined freshwater zone were delineated on the basis of the contour lines. An average hydraulic conductivity between the bounding contour lines for a delineated zone was assigned as the hydraulic conductivity for the zone. Fifteen zones were delineated in the confined freshwater zone for the diffuse-flow hydraulic conductivity distribution, with assigned hydraulic conductivity values ranging from 15 to 1,000 ft/d.

The confined freshwater/saline-water transition zone was delineated as a single zone, with an assigned hydraulic conductivity of 10 ft/d. The assignment of a uniform, comparatively low hydraulic conductivity for the freshwater/saline-water transition zone is consistent with previous models of Maclay and Land (1988), Scanlon and others (2002), and Lindgren and others (2004).

The preliminary diffuse-flow hydraulic conductivity distribution comprised 24 zones—8 for the recharge zone, 15 for the confined freshwater zone, and 1 for the freshwater/saline-water transition zone. Additionally, calibration of the original Edwards aquifer model indicated that zones of comparatively high hydraulic conductivity are needed to convey water toward the major springs and simulate the large measured springflows at Comal and San Marcos Springs. The revision 1 hydraulic conductivity distribution developed by Painter and others (2002) represents the best available mapping of hydraulic conductivity in the Edwards aquifer, on the basis of aquifer tests and upscaling and interpolation to the Edwards aquifer model grid. However, model simulation results indicate that using the revision 1 hydraulic conductivity distribution yields simulated springflows for Comal and San Marcos Springs that are much lower than measured flows. Further upscaling of hydraulic conductivity is required to simulate the high measured springflows. The required upscaling of the hydraulic conductivity can be accomplished by the insertion of broad, high hydraulic conductivity (HHC) zones within the model domain. The location, width, and hydraulic conductivities for these HHC zones were determined on the basis of the hydrogeology of the Edwards aquifer and known zones of high permeability in the aquifer.

The major areas of known high permeability in the aquifer include (1) areas in the central part of the aquifer in Medina and Bexar Counties (Maclay and Small, 1984; Maclay and Land, 1988, fig. 19; Hovorka and others, 1998, fig. 23; Painter and others, 2002) and (2) a relatively narrow zone between well J–17 (Bexar County index well; fig. 1) and Comal Springs (Worthington, 2004). For the diffuse-flow hydraulic conductivity distribution, an HHC zone was delineated in the central part of the aquifer in Medina and Bexar Counties, corresponding with the zones of high transmissivity mapped by

Maclay and Land (1988, fig. 19). The highest hydraulic conductivity in Bexar County might be in a zone less than 1.25-mi wide in eastern Bexar County and about 2 to 3 mi from the freshwater/saline-water interface, based on water-quality data (Worthington, 2004). A relatively broad zone of high transmissivity also was simulated by Maclay and Land (1988) in the confined zone of the aquifer in east-central Bexar County; and high matrix permeability was mapped by Hovorka and others (1998, fig. 12) in an overlapping area of east-central Bexar County.

High-permeability zones often are associated with proximity to major faults and structural grabens. Comal and San Marcos Springs are associated with faults and structural grabens within the very narrow zone of freshwater in the confined zone of the aquifer in Comal and Hays Counties. A marked increase in hydraulic conductivity occurs in a downgradient direction along the major flow path to Comal Springs (Worthington, 2004). Comparatively narrow HHC zones were delineated immediately upgradient from both Comal and San Marcos Springs in the confined zone of the aquifer, corresponding with narrow zones of high transmissivity simulated by Maclay and Land (1988).

Hunter channel (Maclay and Land, 1988, fig. 23), a narrow channel between major faults, lies between Comal and San Marcos Springs and contains extremely transmissive rocks. An HHC zone also was delineated between Comal and San Marcos Springs, reflecting the geologic structure and extremely transmissive rocks.

The major areas of known high permeability in the aquifer include areas along the freshwater/saline-water interface (Maclay and Small, 1984; Hovorka and others, 1998; Worthington, 2004). Relatively high porosity and permeability in the deepest parts of the aquifer near the freshwater/saline-water interface, anomalously high well yields, and sharp chemical gradients indicate that flow might be focused in this area. Therefore, an HHC zone was delineated along the freshwater/saline-water interface using the 1,000-mg/L dissolved solids concentration line as a guide. A highly permeable belt of rocks exists along segments of the freshwater/saline-water interface in areas where mixing ground water of two different chemical types increases the solution capacity of the water (Maclay and Small, 1984). Part of the freshwater/saline-water interface south of Knippa (fig. 2) might be associated with a syncline at the base of the Edwards aquifer, and conduit development resulting in increased permeability in this area might have preferentially occurred in this syncline (Worthington, 2004). The HHC zone delineated ranges from 3 to 13 cells wide and is located predominantly adjacent to and north of the 1,000-mg/L concentration line, except in Bexar County, where it includes areas both north and south of the 1,000-mg/L concentration line. The area between the 1,000- and 3,000-mg/L dissolved solids concentration lines was initially delineated as a single zone with a hydraulic conductivity of 500 ft/d.

The widths of the five delineated HHC zones (central Medina and Bexar Counties, immediately upgradient of Comal Springs, immediately upgradient of San Marcos Springs, between Comal and San Marcos Springs, and along the freshwater/saline-water interface) vary from as narrow as three model cells (0.75 mi) near the freshwater/saline-water interface and San Marcos Springs to as wide as about 5 to 10 mi. The confined freshwater zone of the Edwards aquifer is extremely narrow, or perhaps nonexistent, in sections between San Marcos and Comal Springs.

The magnitude of the hydraulic conductivity of the HHC zones was initially set at 20,000 ft/d, on the basis of the high end of the range of reported transmissivities. Garza (1968, p. 31) and Maclay and Land (1988) estimated maximum transmissivities of from 2 million to more than 4.3 million ft^2/d for the Edwards aquifer. Corresponding hydraulic conductivities for these maximum transmissivities would be on the order of 10,000 to 20,000 ft/d.

Ground-Water Flow

The ground-water-flow system of the Edwards aquifer in the San Antonio region includes the following components:

1. The catchment area (fig. 2) in the Edwards Plateau (fig. 1), where the rocks of the Edwards-Trinity and Trinity aquifers are exposed and receive direct recharge to the water table. Erosion has removed Edwards Group rocks in the Hill Country (fig. 1), as the southern margin of the plateau is known locally (hence Trinity aquifer rather than Edwards-Trinity aquifer in the Hill Country).
2. The recharge zone (fig. 2) in the northern and northeastern parts of the Balcones fault zone (fig. 1), where streams lose flow directly into the unconfined Edwards aquifer outcrop and where the aquifer receives direct recharge to the water table from infiltration of precipitation.
3. The confined zone (fig. 2) in the southern and southeastern part of the Balcones fault zone, which comprises the freshwater zone and the freshwater/saline-water transition zone.

A recent potentiometric-surface map of the Edwards aquifer (Roberto Esquilin, Edwards Aquifer Authority, written commun., 2004) indicates that water entering the catchment area and recharge zone moves from unconfined to confined zones of the aquifer through generally southeasterly flow paths. In the confined zone, the water moves under low hydraulic gradients through fractured, highly transmissive rocks toward the east and northeast, where it is discharged through springs and wells.

Ground-water flow in karst typically includes diffuse or matrix flow (slow flow system), flow through fractures, and flow through large conduits (fast flow system). The multimodal permeability distribution of the Edwards aquifer (Hovorka and others, 1998) implies that the fastest-moving water can travel many times faster than the largest volume of water. Hovorka and others (1998) developed empirical relations between porosity (total porosity, including fracture and solution-enhanced porosity) and permeability, and then used these relations to estimate matrix permeability from log (neutron and resistivity)-based porosity. On the basis of comparisons between mean matrix permeability and mean hydraulic conductivities estimated from aquifer tests, it is likely that the contribution of matrix permeability to regional-scale hydraulic conductivity is minor and that most Edwards aquifer water flows through fractures and conduits (Hovorka and others, 1998). Maclay and Land (1988) identified zones of high transmissivity, with values as much as 112 feet squared per second (ft^2/s) (9,676,800 ft^2/d), along flow paths leading to Comal Springs. The zone of highest transmissivity corresponds with the Comal Springs graben, a narrow graben containing extremely transmissive rocks. Hovorka and others (2004) and Worthington (2004) relate zones of high transmissivity to conduit development and preferential flow paths in the aquifer. Proximity to large faults and high dolomite content indicated by facies distribution and mapped by Rose (1972) indicate that conduits are major contributors to flow in the confined zone of the Edwards aquifer in Hays and Comal Counties (Hovorka and others, 1998). Maclay and Small (1984) assigned this zone the highest transmissivities in the aquifer.

Areas of high transmissivity and preferential flow paths in the Edwards aquifer might be associated with a number of features or processes, including (1) geologic structures, (2) geochemical solution processes, and (3) potentiometric surface troughs. Favorable structural location can be inferred to coincide with high transmissivity and preferential flow paths. For example, the known highest-yielding Edwards aquifer well (Rettman, 1991) is in a major structural trough. Large-scale structural troughs, grabens, and synclines, with high transmissivity and increased flow, occur in the Edwards aquifer. Worthington (2004) hypothesized that grabens and synclines are particularly favorable sites for development of conduits and identified nine major structural troughs in the San Antonio segment of the aquifer (Worthington, 2004, fig. 17). These include troughs in central Uvalde, Medina, and Bexar Counties, southwestern Medina County, within the recharge zone in Comal County, and near the freshwater/saline-water interface from west of Comal Springs to San Marcos Springs. Carbonate dissolution theory also indicates that grabens are a favorable location for conduit development. Maclay and Land (1988, table 6) listed 11 geologic gaps, grabens, and channels, many containing very transmissive rocks, that convey ground water. The conveying structures cited as exerting the most influence on regional ground-water flow are the Knippa gap, Leona Springs gap, Uvalde graben, Dry Frio-Frio River gap, and Hunter channel (Maclay and Land, 1988, fig. 23). A narrow zone within a graben north of San Marcos Springs is fully saturated and contains several wells that yield no measurable drawdown when pumped and several wells that yield water undersaturated with respect to calcite, which indicate a likely location of a conduit and rapid ground-water flow (Hovorka and others, 2004).

Faults can either increase or decrease total transmissivity (Hovorka and others, 1998). Some of the abundant, interconnected fractures in intensely fractured zones adjacent to faults have been enlarged, and they might focus flow parallel to faults. Where calcite cement fills breccia, cross-fault flow might be decreased. Stratigraphic offset of permeable zones along faults might also decrease the cross-fault flow (Maclay and Small, 1983, 1984). Holt (1959) observed nearly 100 ft of head difference across faults in northern Medina County, and George (1952) reported head differences of 6 to 26 ft across segments of major faults in unconfined, less-transmissive parts of the aquifer in Comal County. An inferred conduit between Comal Springs and eastern Bexar County is aligned parallel to major faults near Comal Springs but cuts across faults at a high angle to occur deeper in the subsurface in Bexar County (Hovorka and others, 2004). Maclay (1995) and Groschen (1996) characterized flow in the Edwards aquifer as being controlled laterally by barrier faults that locally compartmentalize the aquifer, especially toward the eastern part of the San Antonio segment. Maclay and Land (1988) hypothesized that large-throw faults segment the aquifer and divert flow in the recharge zone to the west before flow is redirected toward the east.

In southern Bexar County, there is a plume of water with electrical conductivity similar to or lower than that at Comal Springs, which implies the existence of one or more major conduits (Worthington, 2004, fig. 19). As noted in the previous section, the highest hydraulic conductivity in Bexar County might be in a zone less than 1.25-mi wide and about 2 to 3 mi from the freshwater/saline-water interface, on the basis of water-quality data (Worthington, 2004). High transmissivity and thus the potential for large flow near the interface might also be the case in Medina, western Uvalde, and eastern Kinney Counties, but a lack of hydraulic-head or chemistry data close to the freshwater/saline-water interface makes this assumption uncertain. Numerical models of dissolution of carbonate aquifers and tracer tests provide evidence that continuous conduits connect sinking streams and springs in carbonate aquifers (Worthington, 2004). Maclay and Small (1984) hypothesized that solution channels within the Edwards aquifer might be oriented parallel to the courses of streams recharging the Edwards aquifer and that vertical solution channels are well developed below segments of stream channels in the recharge zone.

A regionally extensive system of high-permeability zones is defined by broad troughs in the potentiometric surface (conduit indicators) in the confined zone of the Edwards aquifer. Indications of connections of the confined aquifer to the recharge zone are less well defined by troughs in the potentiometric surface. Worthington (2004) conceptualized a dendritic pattern of conduit connection from the recharge zone to the confined zone. Three approximately synoptic water-level maps constructed by Hovorka and others (2004, figs. 7, 8, 9) show two main trends: (1) a steep and fairly uniform gradient of more than 100 feet per mile (ft/mi) between the Edwards and Trinity aquifers (generally north-northwest to south-southeast) and (2) a gradual gradient from west to east ranging from 2.8 ft/mi in eastern Medina County to 1.2 ft/mi in the eastern part of the aquifer. Superimposed on these regional trends are a number of troughs and divides. Prominent in all three water-level maps is a wide trough that extends westward from central Bexar County to western Medina County. This trough is clearly defined in synoptic maps compiled by or for the Edwards Aquifer Authority (Roberto Esquilin, Edwards Aquifer Authority, written commun., 2003; Hovorka and others, 2004, fig. 10) and has been recognized as a zone of high transmissivity in previous models (Klemt and others, 1979; Maclay and Land, 1988; Painter and others, 2002). In all three water-level maps, the trough can be traced westward to Uvalde County. Westward into Kinney County, the trough becomes broad and poorly defined. Both Hovorka and others (2004, fig. 24) and Worthington (2004, fig. 21) infer the presence of conduits and major flow paths from western Medina County to central Bexar County, with an east-west trend that indicates structural influence. Hovorka and others (2004) hypothesize a complex of interconnected conduits, with about one-half of the segments parallel to faults and one-half of them crossing faults at an appreciable angle. The presence of conduits and major flow paths also is indicated in a potentiometric-surface trough that loops northward around the volcanic center of southeastern Uvalde County.

A steepening of the Edwards aquifer potentiometric-surface gradient occurs in eastern Uvalde County. Maclay and Land (1988) called the area of steepening the Knippa gap and interpreted it as a narrow opening within an extensive, complex barrier-fault system. More detailed structural mapping has complicated the interpretation of faults, and no structural restriction of flow is apparent coincident with the steeper gradient. Water-level maps show a complex, poorly defined flow pattern in central Uvalde County, probably reflecting structural complexity in this area that is further complicated by volcanic intrusions. From central Uvalde County, ground water flows southward and downdip toward southeastern Uvalde and southwestern Medina Counties. Worthington (2004, fig. 21) hypothesized the presence of conduits along this southeastward flow path partially within a syncline at the base of the Edwards aquifer. However, spatial and temporal data density is inadequate to accurately define the flow characteristics of this important region of the aquifer. Wells without measurable drawdown occur in disproportionate numbers near Leona Springs and along a zone trending northwestward from Leona Springs (Hovorka and others, 2004). Details of the connectivity of flow through this area are unclear; however, a network of conduits is inferred from the drawdown data.

Simulation of Ground-Water Flow

A conceptual model of the Edwards aquifer in the San Antonio region was formulated on the basis of an understanding of the hydrogeologic setting, aquifer characteristics, distribution and amount of recharge and discharge, and aquifer boundaries, as described by Lindgren and others (2004). A numerical model of ground-water flow, the conduit-flow Edwards aquifer model, was constructed on the basis of the conceptual model of the aquifer. The conceptual model and the numerical model for the diffuse-flow Edwards aquifer model are the same as those for the conduit-flow Edwards aquifer model (Lindgren and others, 2004), with the exception of the hydraulic conductivity distribution as described in the section "Hydraulic Conductivity Distribution Developed for Diffuse-Flow Edwards Aquifer Model."

Two types of simulations were done for the diffuse-flow Edwards aquifer model—a steady-state simulation that represents long-term average conditions when inflows to and outflows from the flow system are equal and a transient simulation that includes changes in ground-water storage over time.

Numerical Model Description

The FORTRAN computer-model code MODFLOW (McDonald and Harbaugh, 1988; Harbaugh and McDonald, 1996; Harbaugh and others, 2000), a modular finite-difference ground-water-flow code developed by the USGS, was used to simulate ground-water flow in the Edwards aquifer. The Edwards aquifer model uses the Basic, Output Control, Block-Centered-Flow, Recharge, Well, Horizontal-Flow Barrier, Drain, River, and LMG (numerical solver) modules, or MODFLOW packages, to simulate ground-water flow in the Edwards aquifer. The software Groundwater Vistas was used as a pre- and post-processor to facilitate data entry and allow analysis of model output (Environmental Simulations, Inc., 2002). A number of simplifying assumptions about the Edwards aquifer and boundary-condition specifications were required to mathematically represent the aquifer. For the original Edwards aquifer model (Lindgren and others, 2004), it was assumed that conduits could be represented in the model by narrow (one-cell-wide) zones with large hydraulic conductivities. This assumption was not incorporated in the diffuse-flow Edwards aquifer model, and conduit locations are not explicitly identified and simulated as one-cell-wide zones. However, hydrogeologic characteristics such as narrow grabens might result in narrow zones of comparatively high hydraulic conductivity.

A description of the model includes the structure and hydraulic properties of the aquifer, the boundary conditions imposed, and the stresses on the aquifer. Pertinent hydraulic and hydrologic characteristics of the aquifer for appropriate cells within the grid are needed to solve the governing partial-differential equation. Specific input properties required for the model include (1) active and inactive cells, (2) altitudes of the top and bottom of the layer, (3) fault locations and horizontal conductance, (4) horizontal hydraulic conductivity, (5) storativity, (6) river stage and riverbed (streambed) conductance, (7) hydraulic head and conductance of general-head boundary, (8) recharge rates, (9) ground-water withdrawal rates, (10) drain elevation and conductance, and (11) initial hydraulic heads. The properties for the diffuse-flow Edwards aquifer model are the same as for the original Edwards aquifer model, with the exception of the modified horizontal hydraulic conductivity. See Lindgren and others (2004) for a detailed discussion of the construction of the original Edwards aquifer model and the model properties.

Numerical ground-water-flow models of karst aquifers commonly require appreciable increases in measured hydraulic conductivity (upscaling) to make simulated hydraulic heads and springflows match measured hydraulic heads and springflows (Halihan and others, 2000; Scanlon and others, 2002; Hovorka and others, 2004). Hydraulic conductivity in a heterogeneous medium depends on the scale at which it is defined (scaling effect) (Painter and others, 2002). Single-well-based hydraulic conductivity measurements require upscaling to apply to areas the size of model cells. In a porous medium, upscaling generally is not needed and the geometric-mean hydraulic conductivity derived from aquifer tests can be used for aquifer simulations. However, for the Edwards aquifer there is a pronounced scaling effect, and hydraulic conductivities higher than those measured are needed to accurately simulate measured heads and springflows. Worthington (2004, fig. 11) demonstrated that the scaling effect for the Edwards aquifer is similar to that of other karst aquifers and, on the basis of this similarity, concluded that the Edwards aquifer has a well-developed conduit network. Upscaling in a numerical ground-water-flow model can be accomplished either by distribution of high permeability through a wide (multiple cells) zone (Klemt and others, 1979; Maclay and Land, 1988; Scanlon and others, 2002) or by addition of a smaller number of discrete, large-aperture conduits (Worthington, 2004). The diffuse-flow Edwards aquifer model incorporates upscaling by distribution of high permeability through a wide (multiple cells) zone, rather than by addition of a smaller number of discrete, large-aperture conduits.

The initial zoned hydraulic conductivity distribution for the diffuse-flow Edwards aquifer model comprised 29 zones and was derived primarily from (1) estimates of relative transmissivities for the recharge zone made by Maclay and Small (1984) on the basis of available geologic, hydrochemical, and hydrologic information, (2) the hydraulic conductivity distribution developed by Painter and others (2002) for the confined freshwater zone, and (3) geologic structure and geochemical information for the HHC zones. These factors were previously discussed in the “Hydraulic Conductivity and Transmissivity” section of the report.

Numerical Model Calibration

Model calibration is the process in which initial estimates of aquifer properties, stresses (recharge), and boundary conditions are adjusted until simulated hydraulic heads and flows acceptably match measured water levels and flows. Regarding flows, for the Edwards aquifer model, comparisons were made between measured and simulated springflows for Comal, San Marcos, Leona, San Antonio, and San Pedro Springs. Calibration and evaluation of the ground-water-flow model were conducted for steady-state (1939–46) and for transient (1947–2000) conditions. No storage terms are included in the steady-state simulation because change in storage in the aquifer is assumed to be zero. Transient simulations incorporate the storage property of the aquifer and are time-dependent. Changes in storage in the aquifer occur when the amount of water entering the aquifer and the amount of water leaving the aquifer are not equal.

During calibration of the diffuse-flow Edwards aquifer model, only the simulated hydraulic conductivity distribution was modified. All other properties were held constant at the same values as those used in the conduit-flow Edwards aquifer model. The input data incorporate refinements to published and unpublished data on hydraulic conductivity and transmissivity. The calibrated final distribution of hydraulic conductivity results from trial-and-error calibration. During calibration, the hydraulic conductivity data were adjusted (within ranges of variability in measured values) to minimize differences between measured and simulated hydraulic heads and springflows.

The calibration targets for the diffuse-flow Edwards aquifer model are the same as for the conduit-flow Edwards aquifer model and include measured water levels in wells and springflows. The water-level targets include (1) the averages of a series of measurements of water levels in multiple wells for a specified time period (steady-state calibration targets); (2) single measurements of water levels in multiple wells within a comparatively short time period, producing an areal distribution of hydraulic heads (potentiometric-surface map); and (3) a series of measurements of water level within a single well over time (hydrograph). Similarly, the springflow targets include (1) the median value of a series of measurements of springflow for a single spring for a specified time period (steady-state calibration target) and (2) a series of measurements of springflow for a single spring over time (transient calibration target; hydrograph). See Lindgren and others (2004) for a more detailed discussion of the calibration targets.

Steady-State Simulation

The steady-state calibration period for the diffuse-flow Edwards aquifer model was 1939–46, as it was for the conduit-flow Edwards aquifer model (Lindgren and others, 2004). Although hydrologic conditions in the Edwards aquifer have fluctuated temporally and spatially depending on the associated distributions of recharge and water use, long-term averages of recharge, discharge, and water-level data were compiled for a near-predevelopment (pre-1947) period to represent hydraulic equilibrium, or steady-state conditions. The period 1939–46 was chosen because (1) irrigation development was minimal, (2) average rainfall for the period was near the 30-year (1961–90) normal rainfall (National Oceanic and Atmospheric Administration, National Climatic Data Center, 1992), and (3) sufficient water-level and springflow information was available.

For the steady-state calibration, 144 wells were used as the calibration targets (pl. 1; table 1 at end of report). In addition to water levels in wells, median springflows for Comal, San Marcos, Leona, San Antonio, and San Pedro Springs for 1939–46 were used as targets for the steady-state calibration. The median springflows are 332, 152, 19, 10, and 6 cubic feet per second (ft^3/s) for Comal, San Marcos, Leona, San Antonio, and San Pedro Springs, respectively.

The steady-state simulation for the diffuse-flow Edwards aquifer model was calibrated by varying the configuration of the simulated hydraulic conductivity zones and the magnitude of the simulated hydraulic conductivities within those zones. The revisions to the initial simulated hydraulic conductivity distribution, as a result of the steady-state calibration, included

1. Changes in the magnitude of simulated hydraulic conductivities and the areal extent of some zones in the southern part of the Barton Springs segment of the aquifer (northern Hays County),
2. Increases in the simulated hydraulic conductivities for areas near streambeds in the recharge zone and predominantly decreases in the simulated hydraulic conductivities for the areas away from streambeds in the recharge zone,
3. Changes in the magnitude of simulated hydraulic conductivities and the areal extent of some zones in the northern part of the confined freshwater zone of the aquifer,
4. Increases in the hydraulic conductivities for some zones in the confined freshwater zone of the aquifer in eastern Uvalde and western Medina Counties,
5. Decreases in the simulated hydraulic conductivities for the freshwater/saline-water transition zone, and
6. Decreases in the simulated hydraulic conductivities for the HHC zone, delineated in the central part of the aquifer in Medina and Bexar Counties, that corresponds with the zones of high transmissivity mapped by Maclay and Land (1988, fig. 19).

Revisions were made to the configuration of some of the simulated hydraulic conductivity zones and to the simulated hydraulic conductivities for parts of the recharge zone of the Edwards aquifer. Hydraulic conductivity zones were delineated for the areas near streambeds in the recharge zone and assigned a value of 500 ft/d to better match measured hydraulic heads in and near the recharge zone. Maclay and Small (1984) hypothesized that solution channels within the Edwards aquifer might be oriented parallel to the courses of streams recharging the Edwards aquifer and that vertical solution channels are well developed below segments of stream courses in the recharge zone. Also, some of the initial hydraulic conductivity zones were further subdivided, and the hydraulic conductivities decreased to 10 ft/d for much of the recharge zone—in particular for the northern part and areas of zero or small saturated thickness. As discussed in Lindgren and others (2004), the simulated aquifer thickness in parts of the recharge zone includes Trinity aquifer rocks, which have lower hydraulic conductivity than those of the Edwards aquifer.

Revisions were made to the configuration of some of the simulated hydraulic conductivity zones and to the simulated hydraulic conductivities for parts of the confined freshwater zone of the aquifer. In particular, revisions were made in the northern part of the confined freshwater zone of the aquifer. For example, a hydraulic conductivity zone with a value of 10 ft/d was expanded in north-central Medina County, and the hydraulic conductivity of two zones in north-central Bexar County was increased to 1,500 ft/d and the areal extent of the zones was enlarged. These revisions were made to obtain a better match between measured and simulated hydraulic heads in and near the recharge zone. Other changes to the hydraulic conductivity distribution in the confined freshwater zone of the aquifer included (1) increases in hydraulic conductivities to 2,000–7,500 ft/d for some zones in eastern Uvalde and western Medina Counties, in particular the elongate south- and southeast-trending zones, and (2) the addition of a zone with a hydraulic conductivity of 2,000 ft/d specified for the area to the north, northwest, and northeast of Comal Springs.

Revisions were made to the configuration of the simulated hydraulic conductivity zones and to the simulated hydraulic conductivities for the freshwater/saline-water transition zone. The hydraulic conductivity for the HHC zone delineated along the freshwater/saline-water interface was decreased from 20,000 to 10,000 ft/d in Uvalde and Medina Counties. In Bexar County, the initially single HHC zone was divided into two zones with hydraulic conductivities of 12,500 and 15,000 ft/d. The area between the 1,000- and 3,000-mg/L dissolved solids concentration lines, initially delineated as a single zone with a hydraulic conductivity of 500 ft/d, also was divided into multiple zones with varying hydraulic conductivities. A new zone was delineated and the hydraulic conductivity was (1) decreased to 250 ft/d for the area in Uvalde, northeastern Zavala, and northwestern Frio Counties, (2) decreased to 100 ft/d in north-central Frio, eastern Medina, northwestern Atascosa, and western Bexar Counties, (3) maintained at 500 ft/d in eastern Bexar County, and (4) increased to 1,100 ft/d in western Comal County (southwest of Comal Springs).

In addition to the changes for the HHC zone delineated along the freshwater/saline-water interface, revisions were made to the configuration of the simulated hydraulic conductivity zones and to the simulated hydraulic conductivities for other HHC zones. The HHC zone initially delineated as a single zone in the central part of the aquifer in Medina and Bexar Counties (corresponding with the zones of high transmissivity mapped by Maclay and Land [1988, fig. 19]) was divided into two zones with hydraulic conductivities of 10,000 and 12,500 ft/d, decreased from an initial value of 20,000 ft/d. Also, a third zone was delineated in the immediate vicinity of Comal Springs and an increased hydraulic conductivity of 50,000 ft/d assigned. Numerical ground-water-flow models for both the San Antonio (Maclay and Land, 1988) and Barton Springs (Scanlon and others, 2002) segments of the Edwards aquifer assigned the highest hydraulic conductivities to zones nearest to the major springs. The hydraulic conductivity for the HHC zone west of San Marcos Springs was maintained at 20,000 ft/d, whereas the hydraulic conductivity for the HHC zone between Comal and San Marcos Springs was decreased to 12,500 ft/d. An additional

HHC zone, with a hydraulic conductivity of 20,000 ft/d, was added eastward from San Marcos Springs.

The steady-state simulation calibration results include a comparison of simulated hydraulic heads and springflows with average measured water levels and median measured springflows for 1939–46. The calibrated steady-state simulation generally reproduces the spatial distribution of measured water levels. Simulated hydraulic heads were within 30 ft of measured water levels at 130 of the 144 wells used as calibration targets for the steady-state simulation (table 1). The difference was less than 20 ft at 122 of the 144 wells. The largest difference between measured and simulated hydraulic heads was 116.9 ft for a well west of the Guadalupe River near the recharge zone in Comal County. Six of the 14 residuals greater than 30 ft occur in wells in and near the recharge zone (unconfined conditions). Residuals greater than 30 ft also occur in the confined freshwater zone of the aquifer at three wells in southern Hays County, at two wells southwest of the Bexar County index well, and at one well each in southern Comal and eastern Medina Counties. The goodness-of-fit between measured and simulated hydraulic heads was quantified using the mean absolute difference, mean algebraic difference, and root-mean-square (RMS) error. The mean absolute difference between measured and simulated hydraulic heads, computed as the sum of the absolute values of the differences divided by the number of wells, is 11.3 ft (table 2, at end of report). The mean algebraic difference between measured and simulated hydraulic heads, computed as the algebraic sum of the differences divided by the number of wells, is 1.5 ft, which indicates that positive differences were approximately balanced by negative differences. The graph of simulated relative to measured hydraulic heads (fig. 3) indicates little bias in the steady-state simulation results.

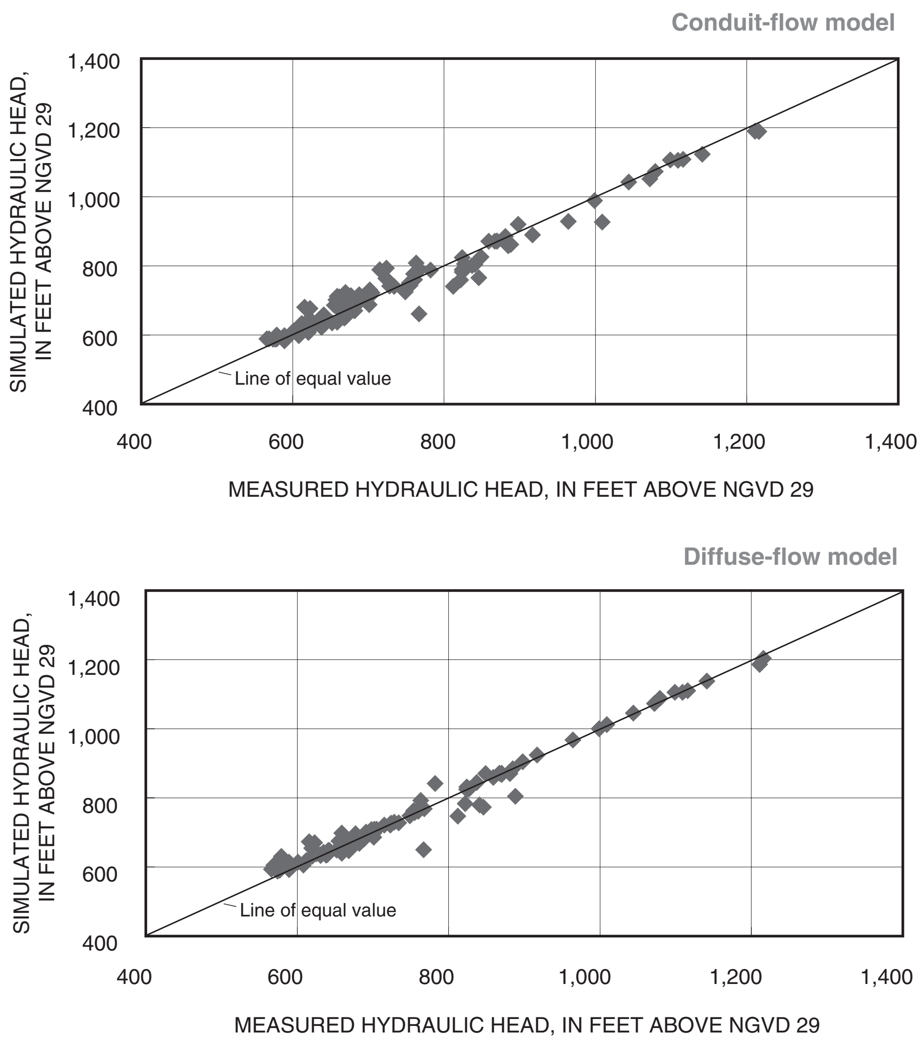


Figure 3. Simulated relative to measured hydraulic heads, steady-state simulation, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas.

The RMS error is derived from the residuals between the measured and simulated hydraulic heads, as given in the following equation:

$$RMS = \left[\frac{1}{n}\sum_{i=1}^{n}(h_m - h_s)_i^2\right]^{0.5}, \qquad (1)$$

where

RMS is the root-mean-square error [L],

n is the number of calibration points,

h_m is the measured hydraulic head at point i [L], and

h_s is the simulated hydraulic head at point i [L].

The RMS error for the 144 target wells of the calibrated steady-state simulation is 20.9 ft (table 2). This error represents about 3 percent of the total measured head difference across the model area (650 ft). For comparison, calibration guidelines adopted by the TWDB Groundwater Availability Modeling (GAM) program specify that the RMS error should be less than 10 percent of the total head difference across the model area (Texas Water Development Board, 2004).

The simulated diffuse-flow-model springflows for Comal and San Marcos Springs for the calibrated steady-state simulation are within about 2.4 and 15 percent of the median springflows for the two springs, respectively (table 3, at end of report). GAM calibration guidelines specify that simulated flows, such as springflow or stream-aquifer leakage, should be within 10 percent of the measured flows. The combined simulated springflows for San Antonio and San Pedro Springs were 31 percent greater than the measured springflows. However, the simulated flows probably reflect local hydrogeologic conditions. Little local-scale data were available for calibration of these two relatively small springs. The simulated springflow for Leona Springs was about twice the measured springflow. However, this discrepancy is reasonable because the reported discharge for Leona Springs might not account for all the discharge from the Edwards aquifer to the Leona gravels (Green, 2004).

Transient Simulation

The transient simulation includes changes in ground-water storage over time that result from pumping and other hydraulic stresses. The distribution of aquifer storativity resulted, therefore, from the calibration of the transient simulation for the original Edwards aquifer model. For the diffuse-flow Edwards aquifer model, however, only the hydraulic conductivity distribution was changed during model calibration, as indicated previously. Simulations for steady-state and transient calibration were run sequentially, and adjustments were made to the appropriate input data until the final versions of both simulations were numerically consistent (residuals minimized for steady-state and transient simulations using the same model properties) representations of the Edwards aquifer flow system.

The transient calibration period for the original Edwards aquifer model was 1947–2000. The transient simulation used 648 monthly stress periods and a single time step per stress period. Multiple time steps (as many as 10) per stress period were used initially, but the number of time steps used was found to have no effect on the model results. Therefore, to minimize the length of the transient-simulation run times, only one time step per stress period subsequently was used.

The transient calibration targets included both a series of measurements over time of water level within a single well (hydrograph) and synoptic sets of water levels in multiple wells within comparatively short time periods. Synoptic sets of water levels in multiple wells during periods of below-normal (drought) and above-normal rainfall were selected to maximize the range of hydrologic conditions included in the transient calibration. The period of below-normal rainfall selected was August 1956, during the 1950s drought, when the lowest water levels of record were recorded. Water-level measurements for May–November 1956 were included as targets for the drought period to have a greater number and better areal distribution of wells with measurements; however, most of the water-level measurements were for August 1956. Therefore, the simulated hydraulic heads for the corresponding (August 1956) model stress period (stress period 117) were used to calculate the hydraulic head residuals. For the drought period, 171 wells were used as transient calibration targets (pl. 1; table 4, at end of report). The period of above-normal rainfall selected was November 1974–July 1975, a period of near record-high water levels in wells. Although most of the water-level measurements were for February 1975, the expansion of the time period of water-level measurements used was necessary to have a greater number and areal distribution of wells. The simulated hydraulic heads for the model stress period corresponding with February 1975 (stress period 339) were used to calculate the hydraulic head residuals. For the above-normal rainfall period, 169 wells were used as transient calibration targets (pl. 1; table 5, at end of report).

In addition to water levels in wells, springflows for Comal, San Marcos, Leona, San Antonio, and San Pedro Springs were used as calibration targets for the transient model calibration. Springflows for time periods that lack measurements were estimated by interpolation or from relations between springflows and water levels at nearby index wells.

The transient simulation for the diffuse-flow Edwards aquifer model was calibrated by varying the simulated hydraulic conductivities. Minor adjustments to the simulated hydraulic conductivities were needed during the transient calibration, particularly in the recharge zone. Calibration of the transient simulation for periods of greatly above-normal rainfall and recharge necessitated increases in hydraulic conductivities for parts of the recharge zone to avoid simulated hydraulic heads above land surface during these periods. This increase of hydraulic conductivities was most prevalent in Kinney, Uvalde, and Medina Counties. The final calibrated hydraulic conductivity distribution for the diffuse-flow Edwards aquifer model used for the steady-state and transient conditions is shown in figure 4. The distribution comprises 38 zones with hydraulic conductivities ranging from 3 to 50,000 ft/d.

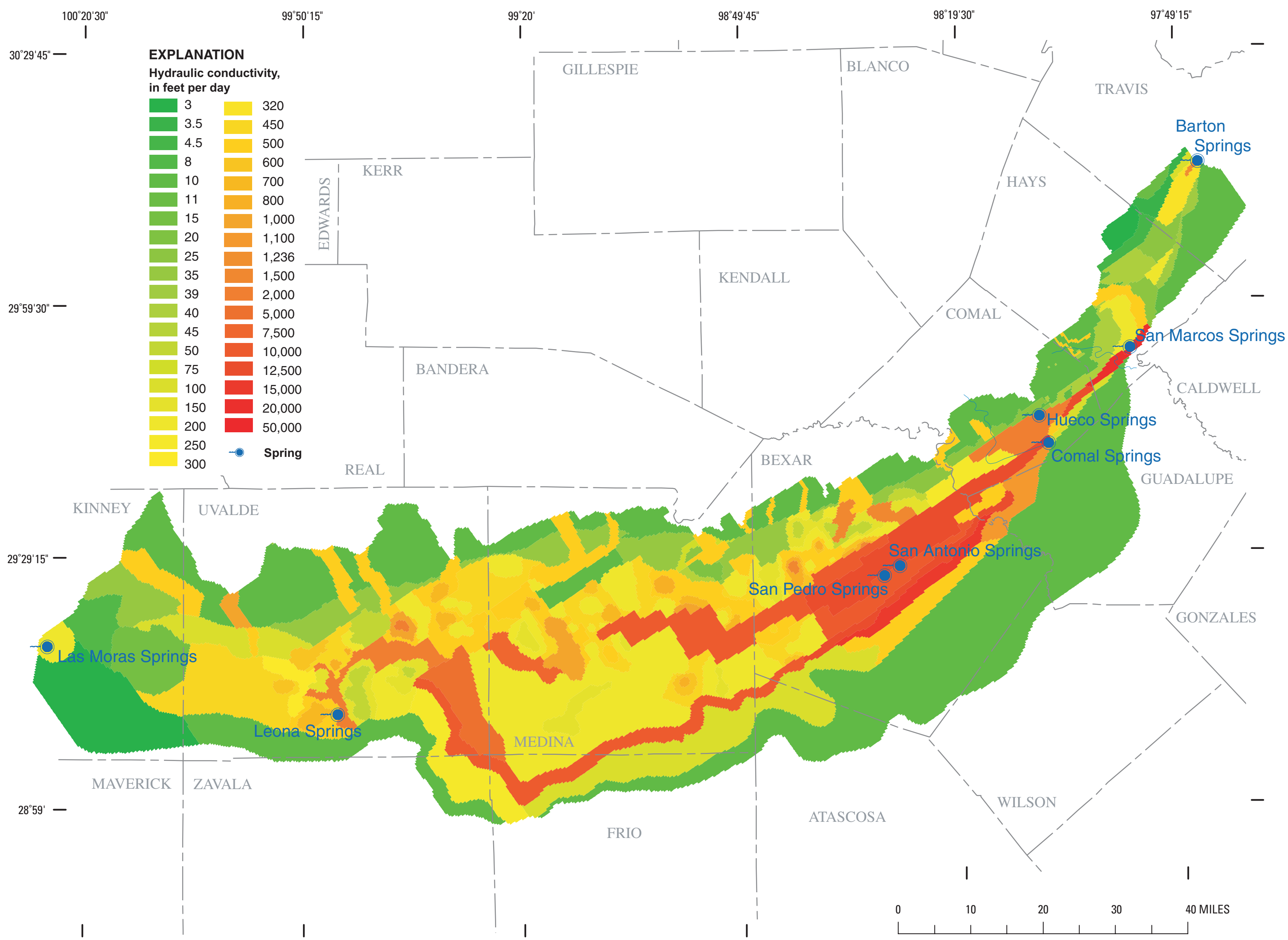


Figure 4. Simulated distribution of hydraulic conductivity for calibrated diffuse-flow Edwards aquifer model, San Antonio region, Texas.

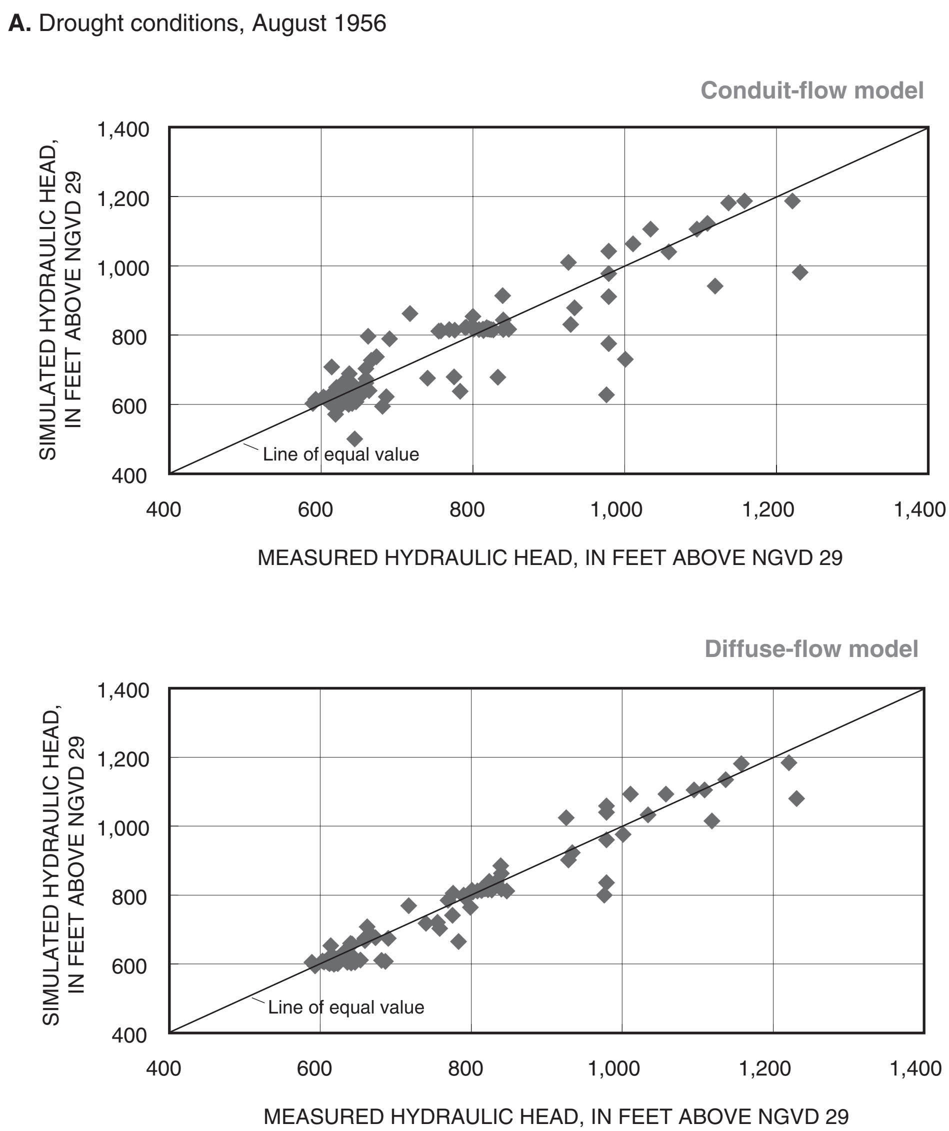


Figure 5. Simulated relative to measured hydraulic heads for (A) August 1956 (stress period 117) and (B) February 1975 (stress period 339), transient simulation, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas.

The final transient simulation generally reproduces the spatial distribution of measured water levels for the periods of drought and above-normal rainfall. Quantitative measures of goodness-of-fit between measured and simulated hydraulic heads (mean absolute difference, mean algebraic difference, and RMS error) were computed for the periods of drought and above-normal rainfall. The closest-match simulated hydraulic heads for the period of drought (May–November 1956) were within 30 ft of measured water levels at 141 of the 171 wells for which water-level data were available (table 4). The difference was less than 20 ft at 131 of the 171 wells. Differences were greater than 100 ft for five of the wells, and the largest difference was 175.8 ft for a well in the recharge zone in Medina County. All five of these wells are in or near the recharge zone in Medina and Uvalde Counties. Residuals greater than 30 ft occurred predominantly in wells in and near the recharge zone and in isolated individual wells in the confined freshwater zone of the aquifer in northeastern Kinney, central and eastern Uvalde, north-central and eastern Bexar, and southern Hays Counties. The isolated individual wells with large residuals generally are near wells with smaller (less than 30 ft) residuals; therefore, the large residuals are anomalous and might be caused by local hydrogeologic conditions not represented in the diffuse-flow Edwards aquifer model. The mean absolute difference between measured and simulated hydraulic heads is 20.5 ft. The corresponding mean algebraic difference is -8.1 ft,

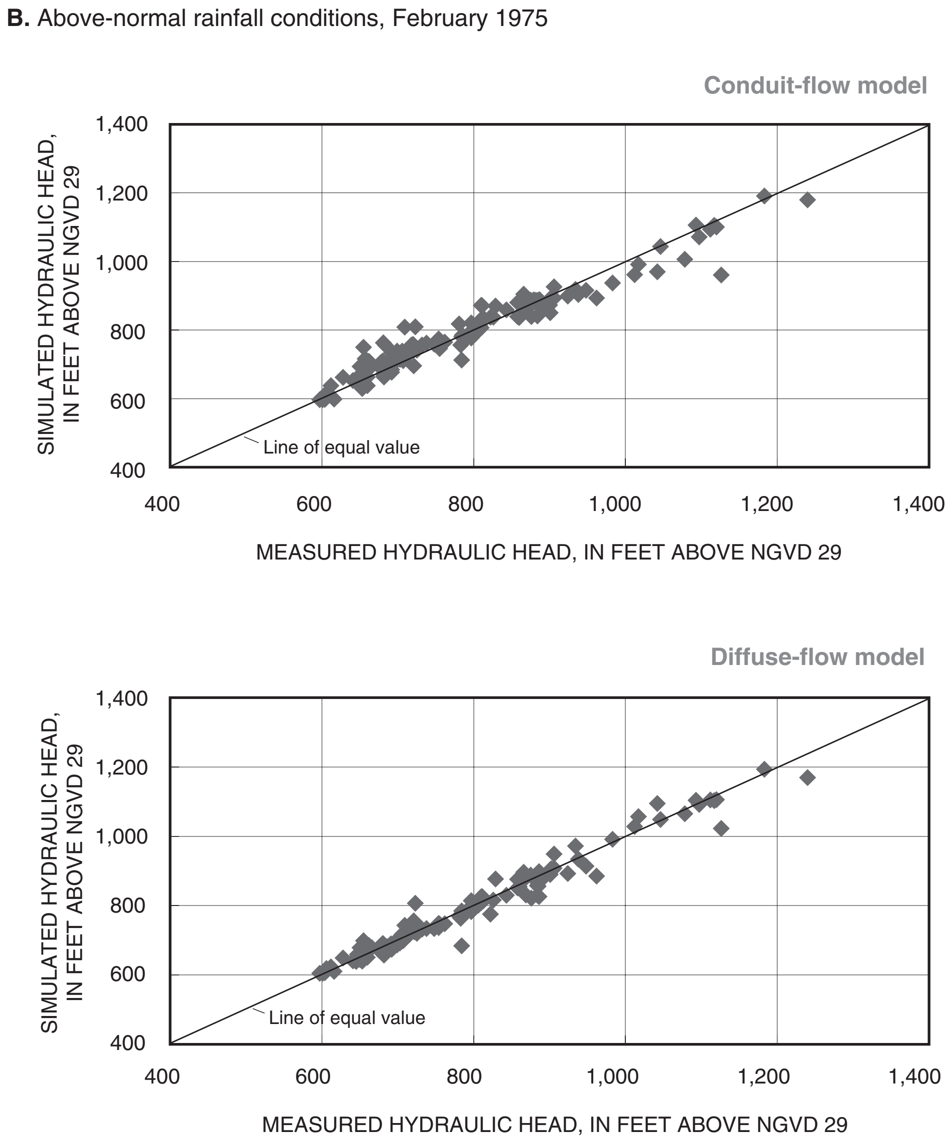


Figure 5. Continued.

which might indicate a small bias toward simulated hydraulic heads too low relative to measured hydraulic heads during the period of drought. The graph of simulated relative to measured hydraulic heads indicates little bias in the simulation results for the period of drought (fig. 5). The RMS error is 33.4 ft (table 2) This error represents about 5 percent of the total measured head difference across the model area (712 ft).

The final transient simulation generally reproduces the spatial distribution of measured water levels for the period of above-normal rainfall. The closest-match simulated hydraulic heads for the period of above-normal rainfall (November 1974–July 1975) were within 30 ft of measured water levels at 138 of the 169 wells for which water-level data were available (table 5). The difference was less than 20 ft at 125 of the 169 wells. Differences were greater than 100 ft for two wells, and the largest difference was 103.3 ft for a well in the recharge zone in Uvalde County. Wells with residuals greater than 30 ft occur predominantly in and near the recharge zone. Isolated individual wells with relatively large residuals (greater than 30 ft) occur in central and south-central Uvalde, central and north-central Bexar, and southern Hays Counties. The mean absolute difference between measured and simulated hydraulic heads is 17.2 ft (table 2). The corresponding mean algebraic difference is 0.5 ft, which indicates that positive differences were approximately balanced by negative differences. The graph of simulated relative to measured hydraulic heads indicates little bias in the simulation results for the period of above-normal rainfall (fig. 5). The RMS error is 25.8 ft (table 2). This error represents about 4 percent of the total measured head difference across the model area (663 ft).

As indicated by the mean algebraic difference between simulated and measured hydraulic heads (-8.1 ft), the model might tend to simulate lower-than-measured hydraulic heads during periods of drought. On the basis of the RMS errors and the mean algebraic differences, the model provides a reasonable, conservative simulation of water levels for varying hydrologic conditions.

The transient calibration results also include a comparison of simulated springflows and hydraulic heads with a series of measurements of springflow and of water levels within individual wells over time (hydrograph). Selected hydrographs comparing measured and simulated hydraulic heads for 11 of 12 target wells with long-term water-level measurements and springflows for five springs are included in this report. The 12 target wells are distributed throughout the model area (pl. 1) and are representative of the results of the transient simulation. The transient simulation for 1947–2000 acceptably reproduces measured fluctuations in hydraulic heads over time in the Edwards aquifer (figs. 6–8). The match between measured and simulated hydraulic heads generally is closer for wells completed in the confined zone of the aquifer than for those in and near the recharge zone. The RMS error ranged from 6.3 to 30.4 ft in 12 wells with water-level measurements for varying periods during 1947–2000 (table 6, at end of report), and these errors represent 5.0 to 31.3 percent of the range in water-level fluctuations of each of those wells. The three wells with RMS errors greater than 20 ft are all in Hays County. The smallest RMS error (6.3 ft) was for well 6823701 in the confined freshwater zone of the aquifer in Comal County, and the largest (30.4 ft) was for well 6702103 in the confined freshwater zone of the aquifer in Hays County (table 6).

Generally acceptable agreement also was obtained between measured and simulated flow at springs (figs. 9, 10). The RMS errors for Comal, San Marcos, Leona, San Antonio, and San Pedro Springs ranged from 348,280 cubic feet per day (ft^3/d) (4.0 ft^3/s) for San Pedro Springs to 3,448,867 ft^3/d (39.9 ft^3/s) for San Antonio Springs (table 6). The RMS errors for the five springs, as a percentage of the range of discharge fluctuations measured at the springs, varied from 7.2 percent for San Marcos Springs and 8.1 percent for Comal Springs to 28.8 percent for Leona Springs. The mean algebraic differences between simulated and measured spring discharges are -127,334 ft^3/d (-1.5 ft^3/s) and 543,298 ft^3/d (6.3 ft^3/s) for Comal and San Marcos Springs, respectively, indicating that positive differences were approximately balanced by negative differences (table 6). Simulated high discharges during the 1970s and 1990s for Comal Springs and during the late 1960s for San Marcos Springs are lower than the measured high discharges (fig. 9), whereas during other periods of measured high flows, the simulated and measured discharges generally are in close agreement. The recessions in simulated flows for San Marcos Springs generally are more gradual than are those in the measured data, and for some time periods the simulated low discharges of the recessions do not match the measured low discharges. Therefore, the simulated springflows tend to overestimate the measured springflows during many low-flow periods, resulting in the positive mean algebraic difference for San Marcos Springs (table 6). The simulated spring discharge for Leona Springs often is greater than the measured discharge during periods of zero or low discharge and is anomalously high following the very large recharge events of the early 1990s (fig. 10). However, as noted at the end of the previous section, reported discharge for Leona Springs might appreciably underestimate the actual discharge because of unmeasured discharge from the Edwards aquifer to the Leona gravels. The simulated spring discharges for San Antonio Springs tend to overestimate the measured discharges and for San Pedro Springs tend to underestimate the measured discharges during peak discharge periods (fig. 10). In general, the model reasonably simulates springflows for different hydrologic conditions; the match is closer for Comal and San Marcos Springs than for the smaller springs (Leona, San Antonio, and San Pedro Springs).

Water Budget

In addition to hydraulic heads, the water budget for the model is computed for each stress period. A water budget in the context of the model is an accounting of inflow to, outflow from, and storage change in the aquifer. For steady-state conditions, inflow (sources) to the aquifer equals outflow (discharges) from the aquifer. For transient conditions, changes in storage likely occur. For a balanced transient water budget—that is, total sources equal total discharges equal total flow through the aquifer—positive changes (gains) in storage must be included as discharges, and negative changes (losses) in storage must be included as sources. Sources of water to the Edwards aquifer include (1) recharge from leakage from streams and infiltration of rainfall in the recharge zone and (2) inflow through the northern and northwestern aquifer boundaries. Also, a small amount of leakage from the Colorado River to the aquifer occurs during periods of low water levels. Discharge from the Edwards aquifer includes (1) springflow (drain discharge), (2) withdrawals by wells, and (3) leakage to the Colorado River from the aquifer.

Steady-State Simulation

The steady-state simulation water budget indicates that recharge accounts for 93.6 percent of the sources of water to the Edwards aquifer, and inflow through the northern and northwestern model boundaries contributes 6.4 percent (table 7, at end of report). Most of the flow into the model area through the northern and northwestern model boundaries occurs through the northern boundary (87.9 percent). The largest discharges from the Edwards aquifer in the steady-state simulation water budget are springflow (73.4 percent) and withdrawals by wells (25.7 percent) (table 7). Discharge from the aquifer to the Colorado River (as stream-aquifer leakage) is a minor component of the steady-state budget (0.9 percent).

Transient Simulation

In the simulated water budget for the transient simulation for 1956 (table 7), positive changes (gains) in storage for stress periods are included as discharges, and negative changes (losses) in storage are included as sources. This convention is consistent with making the water budget (total sources equal total discharges equal total flow through the aquifer) balance. The 1956 water budget represents drought conditions.

The principal source of water to the Edwards aquifer (excluding change in storage) for the transient simulation is recharge, constituting about 60 percent of the sources of water to the Edwards aquifer during 1956 (table 7). Inflow through the northern and northwestern model boundaries contributed a relatively small amount of water. Subsurface inflow through the northern and most of the northwestern model boundaries was simulated as a constant flux, and therefore no variation occurs on a monthly and annual basis. Although the amount of water contributed by boundary inflow was relatively small, it constituted about 40 percent of the sources (excluding change in storage) to the aquifer during 1956, because of the greatly reduced recharge during this drought period. A very small amount of leakage from the Colorado River to the aquifer occurred during 1956. This leakage occurred because hydraulic heads in the aquifer, although generally above river stage, are below river stage during periods of low water levels in the aquifer.

The principal discharges from the Edwards aquifer for the transient simulation are springflow and withdrawals (pumpage) (table 7). During 1956, representing drought conditions, the greatest discharge was withdrawals (pumpage) (76.1 percent), followed by springflow (23.2 percent); this is the reverse of steady state for which the greatest discharge is springflow, followed by withdrawals (pumpage). During 1956 and for steady-state, discharge from the aquifer to the Colorado River is a small component of the budget. Withdrawals (pumpage) was the largest budget component (excluding change in storage) during 1956, with low rainfall resulting in low recharge and increased pumpage. In contrast, for steady-state the much greater rainfall and corresponding greater recharge (steady-state recharge about 9.6 times greater than in 1956) resulted in pumpage being a proportionately smaller component of the budget. Springflow was the largest discharge from the aquifer for steady state, much greater than in 1956 (by a factor of about 4.9), because of higher water levels in the aquifer. Water levels in the aquifer depend on both recharge and pumpage. Steady-state pumpage was approximately one-half of the pumpage during 1956, and recharge was greater by a factor of 9.6, which resulted in higher water levels.

During 1956, the change in storage (net water released from storage) was much greater than recharge, accounting for 75.0 percent of the total flow (including change in storage) compared to 15.0 percent for recharge (table 7). The amount and percentage of water released from storage is large during 1956 (drought conditions) because recharge is small and more water is required from storage to meet the pumpage demand. During 1956, the largest net releases from storage occurred during June–August, the period of largest withdrawals by wells and comparatively low recharge. No net addition to storage occurred during 1956, which indicated the aquifer was being depleted of water during the entire year.

Comparison of Simulations

Hydraulic heads and springflows simulated by the diffuse-flow Edwards aquifer model for selected observation wells and springs were compared to the corresponding hydraulic heads and springflows simulated by the conduit-flow Edwards aquifer model. Both the steady-state and transient simulated hydraulic heads and springflows were used for the comparisons. The results of the comparison between the simulated hydraulic heads and springflows for the diffuse-flow Edwards aquifer model and the conduit-flow Edwards aquifer model are shown in hydrographs for selected observation wells and springs (figs. 11, 12) and summarized in table 6. The mean absolute difference, mean algebraic difference, and RMS error of the residuals for the diffuse-flow Edwards aquifer model and the conduit-flow Edwards aquifer model are tabulated. These statistics are used as quantitative measures of the goodness-of-fit between the measured and simulated hydraulic heads and springflows.

The mean absolute difference and the RMS error of the residuals for hydraulic heads for the diffuse-flow Edwards aquifer model are appreciably smaller than those for the conduit-flow Edwards aquifer model for the three synoptic water-level time periods (steady state, drought, and above-normal rainfall) (table 2). The RMS errors for the conduit-flow Edwards aquifer model are 27, 76, and 30 percent greater than for the diffuse-flow Edwards aquifer model for the steady-state, drought, and above-normal rainfall synoptic time periods, respectively. The improved statistical match between the simulated and measured hydraulic heads also is evident from a comparison of the graphs of simulated relative to measured hydraulic heads, as indicated by less deviation from the match line (line of equal value) of the data in the graphs for the diffuse-flow Edwards aquifer model (figs. 3, 5). The improvements in the match between simulated and measured hydraulic head were for wells predominantly in or near the recharge zone of the aquifer. The improvements were achieved primarily by adjustments to the configuration of hydraulic conductivity zones and the assigned hydraulic conductivities for the hydraulic conductivity zones in and near the recharge zone. The general pattern of lower simulated hydraulic heads for the diffuse-flow Edwards aquifer model than for the conduit-flow Edwards aquifer model is indicated by the consistently lower mean algebraic differences for the diffuse-flow Edwards aquifer model for all three synoptic water-level time periods (table 2).

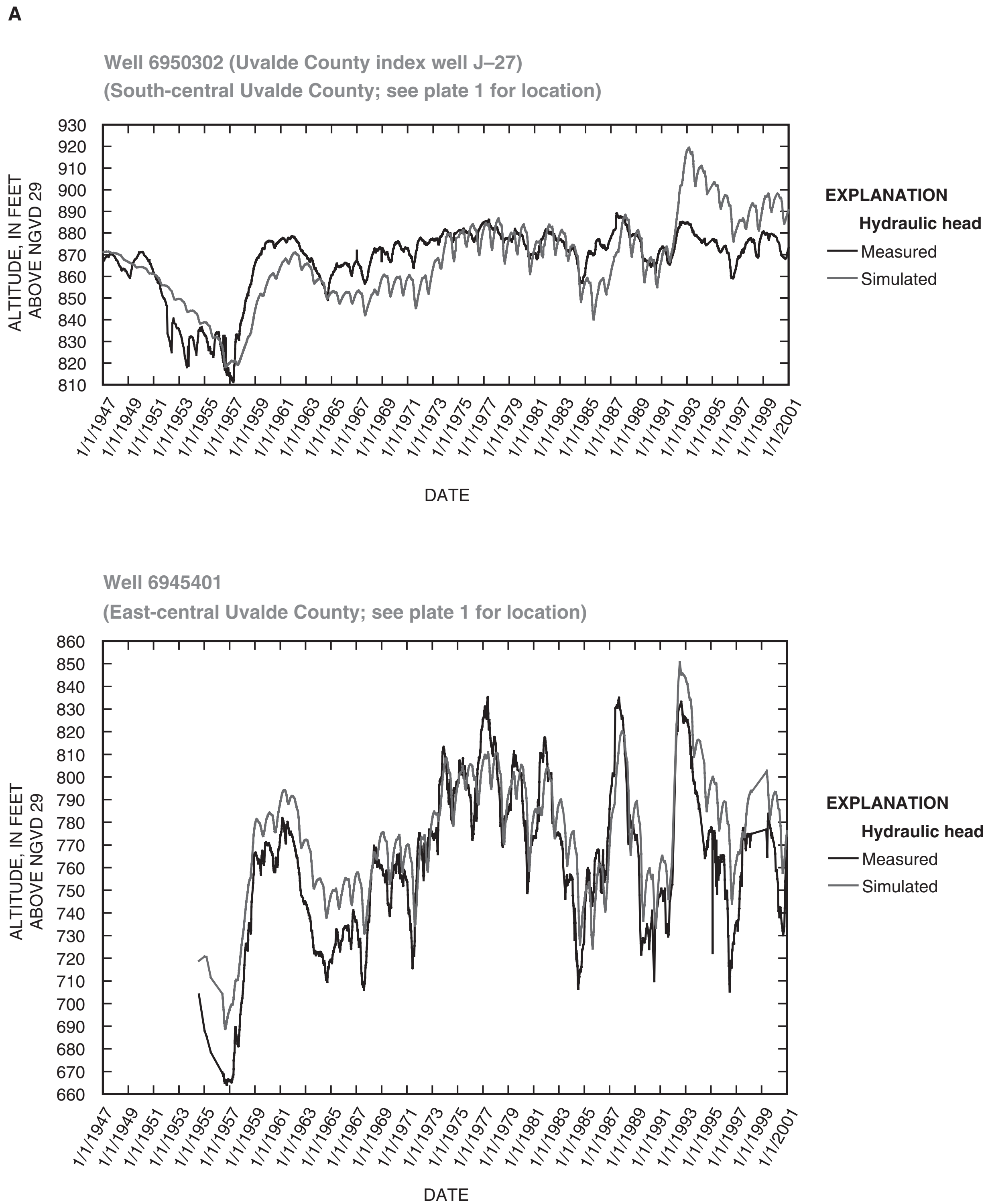
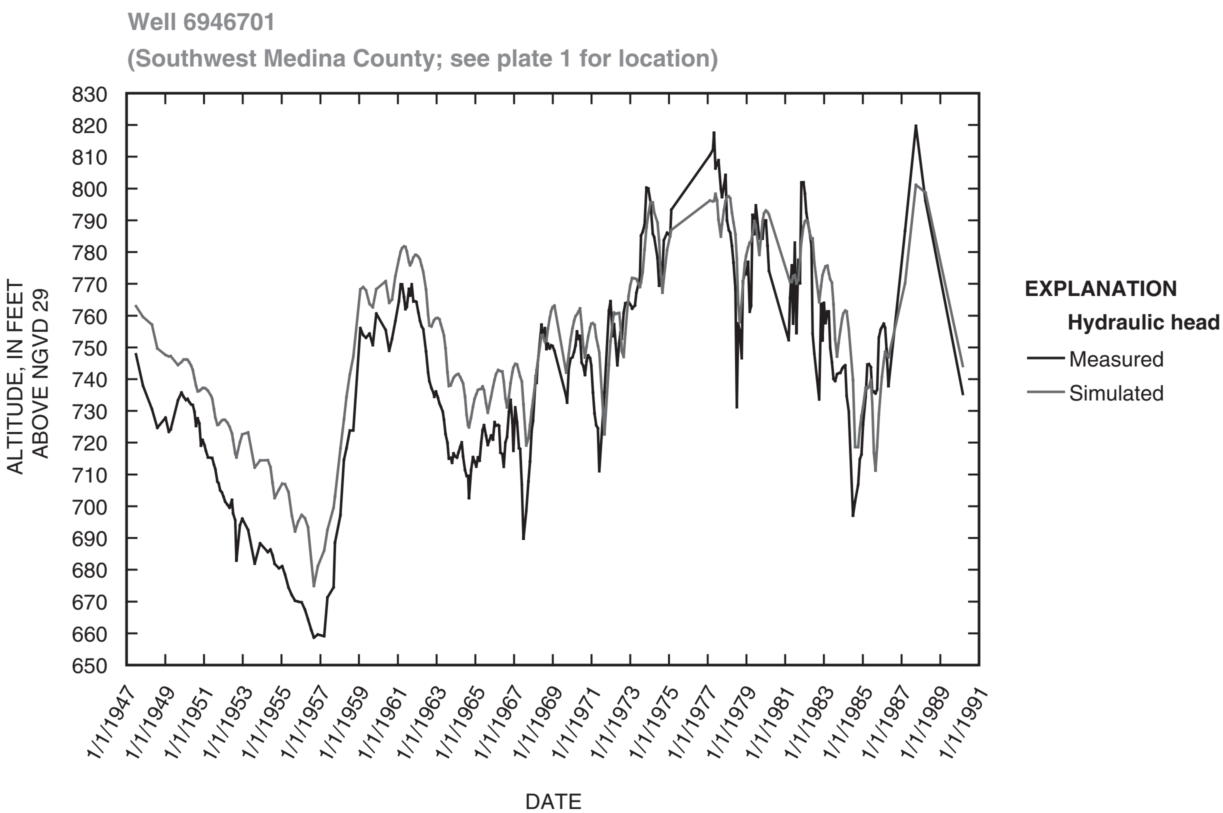
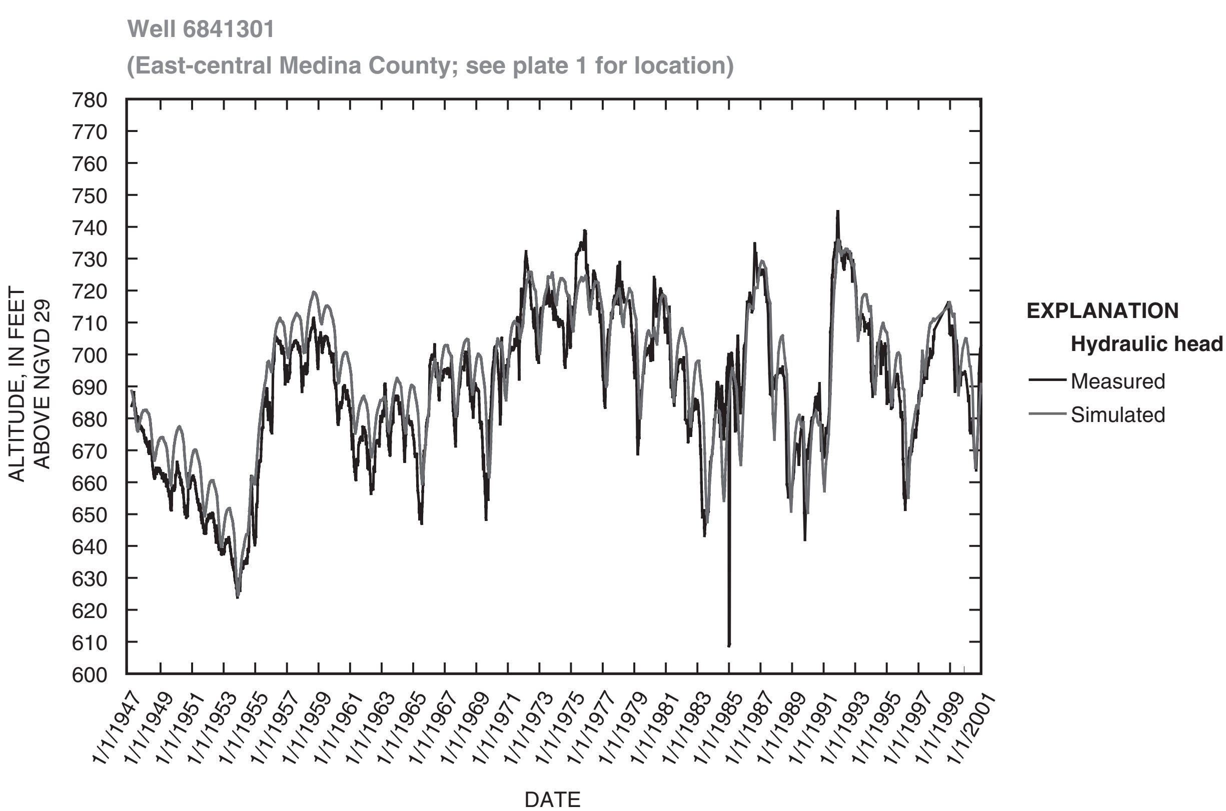


Figure 6. Measured and simulated water levels (hydraulic heads) for Edwards aquifer wells in (A) Uvalde County and (B) Medina County, diffuse-flow Edwards aquifer model, San Antonio region, Texas.

B





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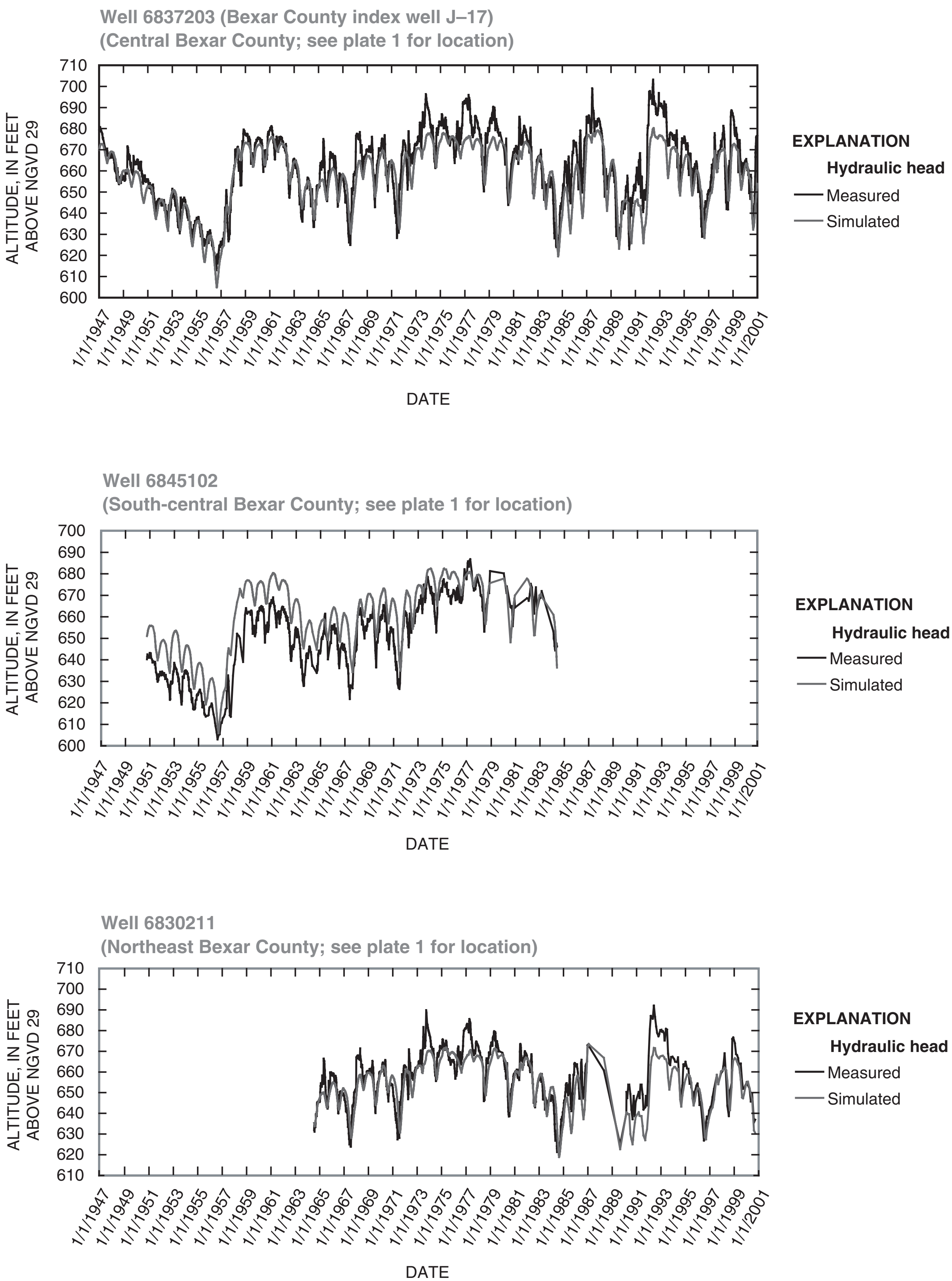


Figure 7. Measured and simulated water levels (hydraulic heads) for Edwards aquifer wells in Bexar County, diffuse-flow Edwards aquifer model, San Antonio region, Texas.

The simulated steady-state springflows for the diffuse-flow Edwards aquifer model and the conduit-flow Edwards aquifer model are generally similar (table 3). The differences are 3 ft^3/s for San Marcos and San Pedro Springs, 2 ft^3/s for Comal Springs, and approximately zero for Leona and San Antonio Springs. The differences are 0.6 and 2.0 percent of the median steady-state springflows for Comal and San Marcos Springs, respectively.

For the transient simulation, the mean absolute difference and RMS error of the residuals for the diffuse-flow Edwards aquifer model are smaller than those for the conduit-flow Edwards aquifer model for six of the 12 target wells with long-term water-level measurements (table 6). Conversely, those statistics for the diffuse-flow Edwards aquifer model are larger than those for the conduit-flow Edwards aquifer model for six of the 12 target wells. The largest reductions in the RMS error for the diffuse-flow Edwards aquifer model occurred for two wells in Bexar County (-11.1 and -8.0 ft). The largest increases in the RMS error occurred for two wells in Hays County (15.0 and 15.1 ft). In Comal, Medina, and Uvalde Counties the differences in the RMS errors are 7.2 ft or less and are equally distributed between reductions and increases. The mean algebraic differences for the three target wells in Hays County for the diffuse-flow Edwards aquifer model are higher than those for the conduit-flow Edwards aquifer model (table 6), indicating a general increase in hydraulic heads. In contrast, the mean algebraic differences for the three target wells in Bexar County are lower for the diffuse-flow Edwards aquifer model, indicating a general decrease in hydraulic heads. The mean algebraic differences for the target wells in Comal, Medina, and Uvalde Counties are equally distributed between reductions and increases for the diffuse-flow Edwards aquifer model compared to those of the conduit-flow Edwards aquifer model. Hydrographs also illustrate the differences between measured and simulated hydraulic heads and springflows for the diffuse-flow Edwards aquifer model compared to those of the conduit-flow Edwards aquifer model. The hydrographs for the Bexar County index well (6837203) and the Uvalde County index well (6950302) generally illustrate lower water levels simulated by the diffuse-flow Edwards aquifer model than those simulated by the conduit-flow Edwards aquifer model (fig. 11).

The residual statistics (mean absolute difference and RMS error of the residuals) in some cases are not a complete measure of the goodness-of-fit between the measured and simulated hydraulic heads. The hydrograph for the Bexar County index well indicates that the conduit-flow Edwards aquifer model might be more responsive to the magnitude of fluctuations in measured water levels than the diffuse-flow Edwards aquifer model (fig. 11), although the residual statistics for the diffuse-flow Edwards aquifer model are similar to those for the conduit-flow Edwards aquifer model (table 6). The simulated peaks in the hydrograph for the Bexar County index well for the diffuse-flow Edwards aquifer model generally are lower than for the conduit-flow Edwards aquifer model. During the periods of highest water levels (periods of highest precipitation and recharge) and largest peaks in the hydrograph, the differences in simulated heads between the two models are greater than for periods with lower peaks in the hydrograph.

The goodness-of-fit between measured and simulated springflows, as indicated by the statistics in table 6 and the hydrographs of Comal and San Marcos Springs in figure 12, is similar for Comal, San Marcos, and Leona Springs for the diffuse-flow Edwards aquifer model and the conduit-flow Edwards aquifer model. The mean absolute difference and RMS error were lower for Comal and Leona Springs and minimally higher for San Marcos Springs for the diffuse-flow Edwards aquifer model. The RMS errors for Comal and Leona Springs were 15.6 and 21.3 percent less, respectively, whereas the RMS error for San Marcos Springs was 3.3 percent greater. The differences in simulated springflows between the diffuse-flow Edwards aquifer model and the conduit-flow Edwards aquifer model are greatest during the periods of greatest springflows (fig. 12). The RMS errors for San Antonio and San Pedro Springs were appreciably greater, 80.2 and 51.0 percent, respectively, for the diffuse-flow Edwards aquifer model (table 6). The appreciable increases in the RMS errors for San Antonio and San Pedro Springs might be caused by the movement of large volumes of water through a broader HHC zone, as opposed to the one-cell-wide conduits for the conduit-flow Edwards aquifer model. San Antonio and San Pedro Springs are relatively distant from the simulated one-cell-wide conduits in the conduit-flow Edwards aquifer model, but they are within an HHC zone in the diffuse-flow Edwards aquifer model. The mean algebraic differences indicate that the springflows are less for all the springs, except San Antonio Springs, for the diffuse-flow Edwards aquifer model compared to those for the conduit-flow Edwards aquifer model. This, coupled with the observation that greater differences occur during the periods of greatest springflows, might indicate that the diffuse-flow Edwards aquifer model is somewhat less responsive than the conduit-flow Edwards aquifer model.

The simulated water budgets for the diffuse-flow Edwards aquifer model are similar to those for the conduit-flow Edwards aquifer model. For the steady-state simulation, the largest differences in percentage of total sources or discharges for a budget component are 0.3 percent for two discharges, springflow and stream-aquifer leakage (table 7). For 1956 during the transient simulation, the largest differences in percentage of total sources or discharges for a budget component are an increase of 1.8 percent for withdrawals (pumpage) (increase of 749.7 acre-feet per year [acre-ft/yr]) and a decrease of 2.0 percent for springflow (decrease of 10,732.1 acre-ft/yr). In addition, the magnitude (decrease of 10,497.6 acre-ft/yr) and percentage (decrease of 0.9 percent) for the net change in storage (net loss of water from storage) were less for the diffuse-flow Edwards aquifer model. The relatively small increase in withdrawals (pumpage) for the diffuse-flow Edwards aquifer model was due to fewer model cells going dry in response to the simulated drought conditions and the deactivation of wells in those dry model cells.

A

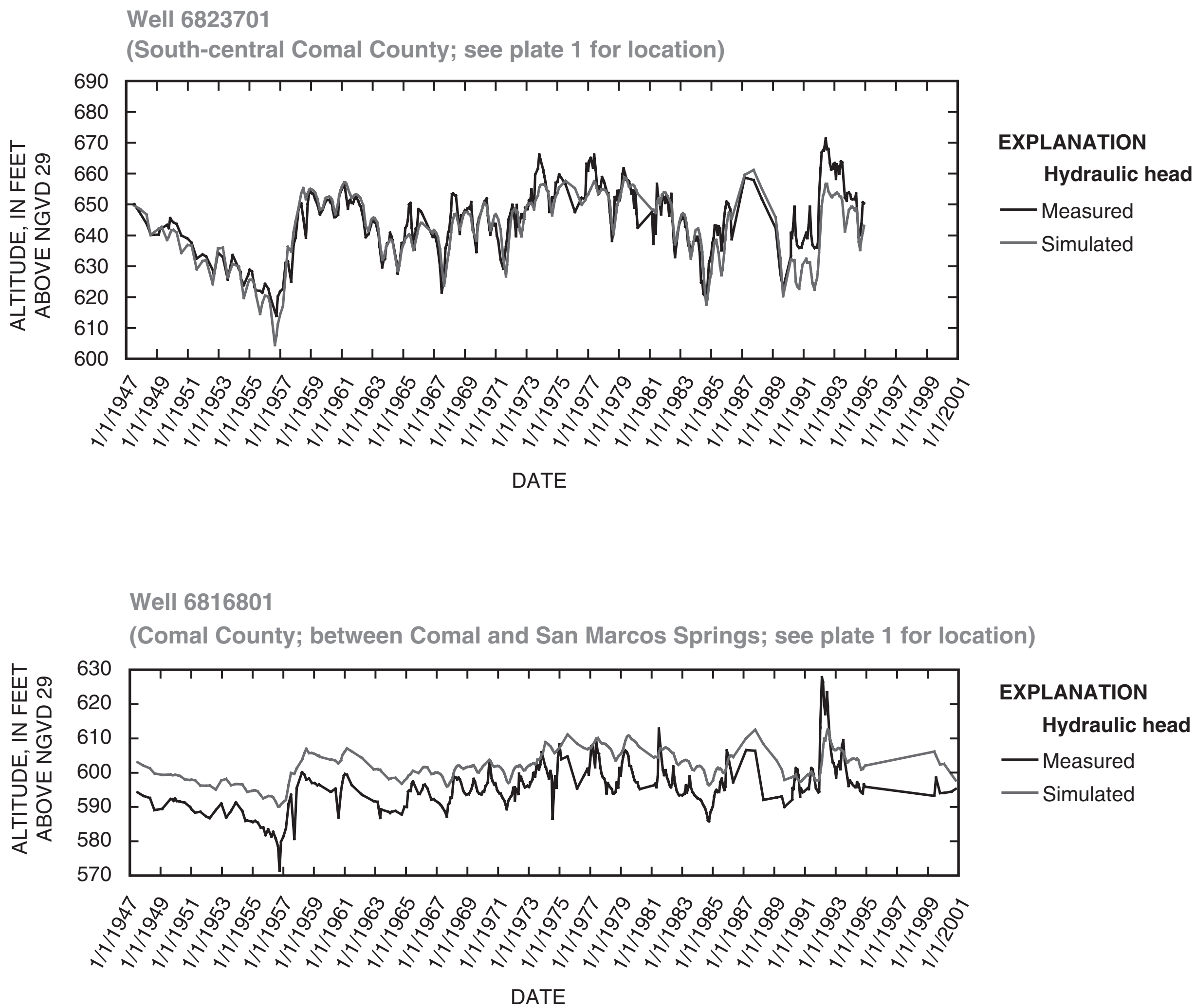
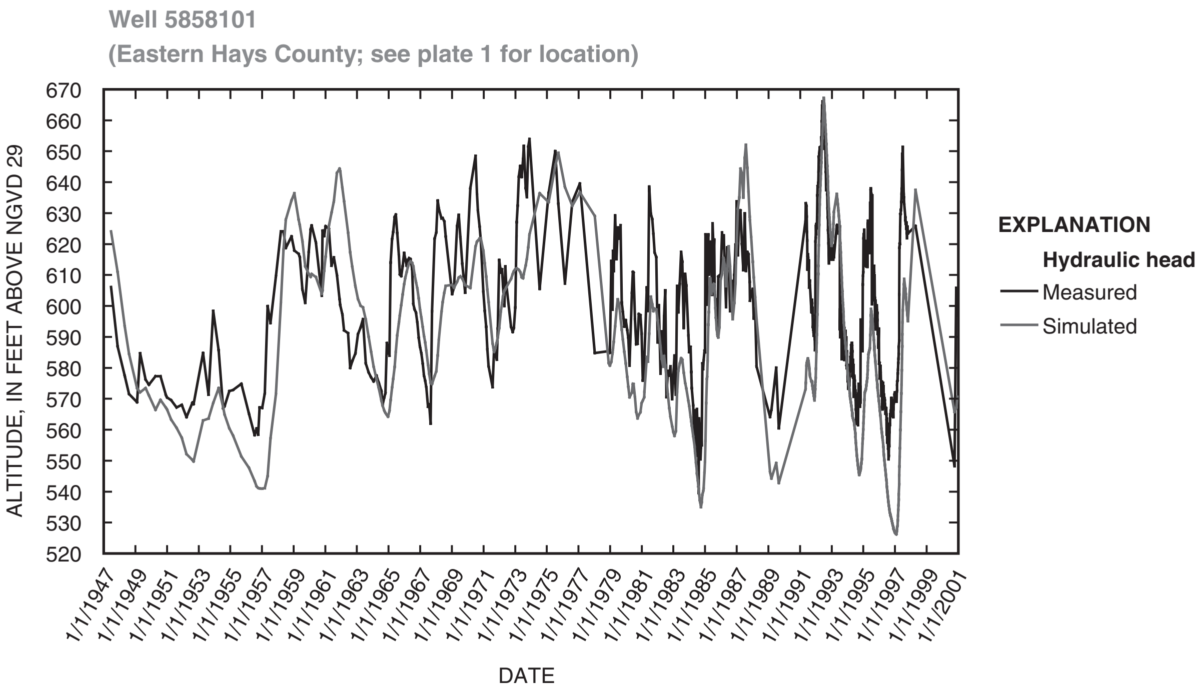
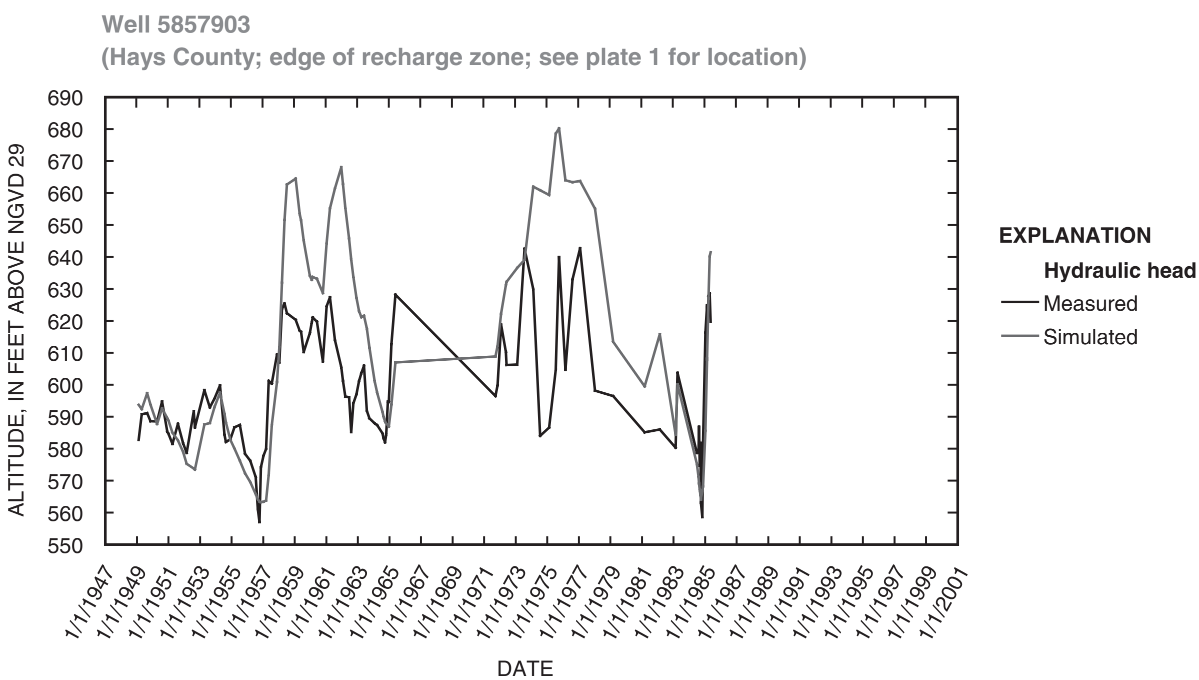


Figure 8. Measured and simulated water levels (hydraulic heads) for Edwards aquifer wells in (A) Comal County and (B) Hays County, diffuse-flow Edwards aquifer model, San Antonio region, Texas.

The largest change in terms of the magnitude of water budget components was the decrease of about 10,730 acre-ft/yr in springflow for the diffuse-flow Edwards aquifer model compared to springflow for the conduit-flow Edwards aquifer model (table 7). This decrease in springflow (a water budget discharge) was largely offset by the smaller net loss of water from storage (a water budget source) of about 10,500 acre-ft/yr (table 7). The decrease in springflow for the diffuse-flow Edwards aquifer model compared to that for the conduit-flow Edwards aquifer model might be due to the general pattern of lower simulated hydraulic heads for the diffuse-flow Edwards aquifer model compared to that for the conduit-flow Edwards aquifer model, indicated by the lower mean algebraic differences for the diffuse-flow Edwards aquifer model (table 2). Smaller springflows for all the springs, except San Antonio Springs, for the diffuse-flow Edwards aquifer model compared to springflows for the original Edwards aquifer model also are indicated by the lower mean algebraic differences for the diffuse-flow Edwards aquifer model as compared to the differences for the conduit-flow Edwards aquifer model (table 6).

B





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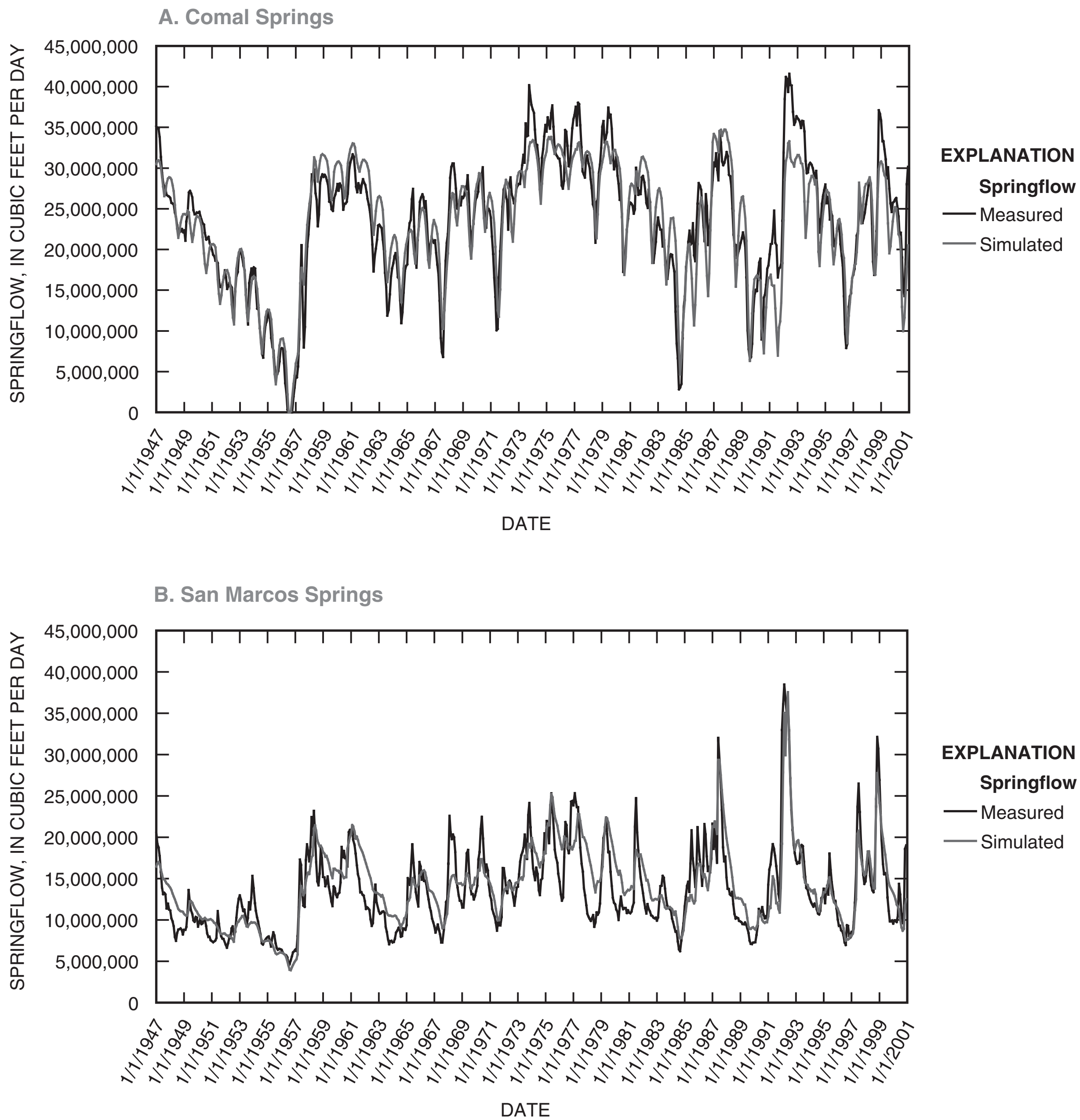


Figure 9. Measured and simulated springflows for (A) Comal Springs and (B) San Marcos Springs, 1947–2000, diffuse-flow Edwards aquifer model, San Antonio region, Texas.

Model Limitations

All numerical ground-water-flow models are simplifications of the real system and, therefore, have limitations. Limitations generally result from assumptions used to develop the conceptual and numerical models, limitations in the quality and quantity of the input data, and scale at which the model can be applied. In addition, a combination of input to the model different from that used in the calibrated simulations could produce the same result; the solution is non-unique. A summary of important limitations and uncertainties associated with conceptual and numerical models, input data, and scale of application is given in the text box on page 29. Lindgren and others (2004) contains a more detailed discussion of the model limitations.

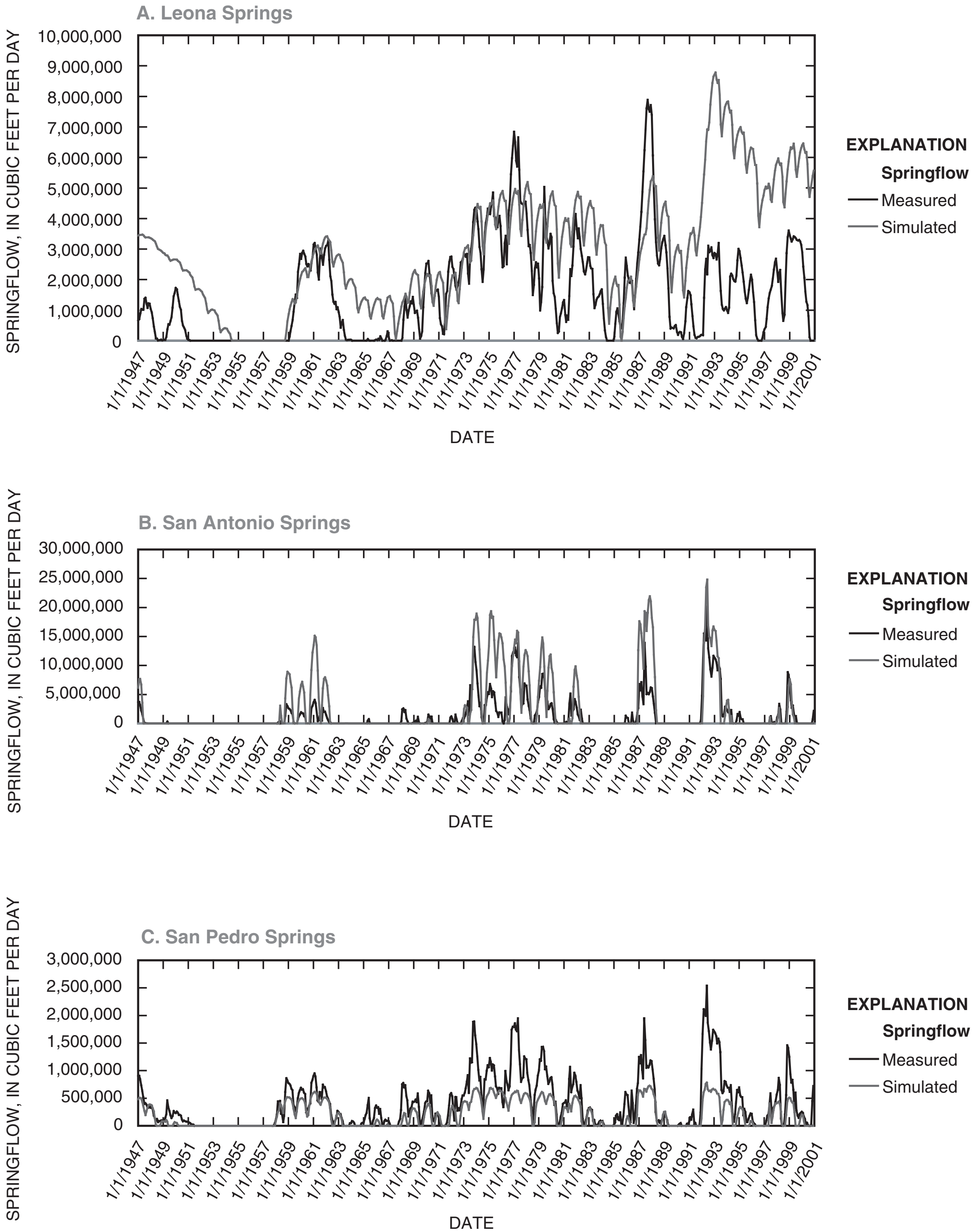


Figure 10. Measured and simulated springflows for (A) Leona Springs, (B) San Antonio Springs, and (C) San Pedro Springs, 1947–2000, diffuse-flow Edwards aquifer model, San Antonio region, Texas.

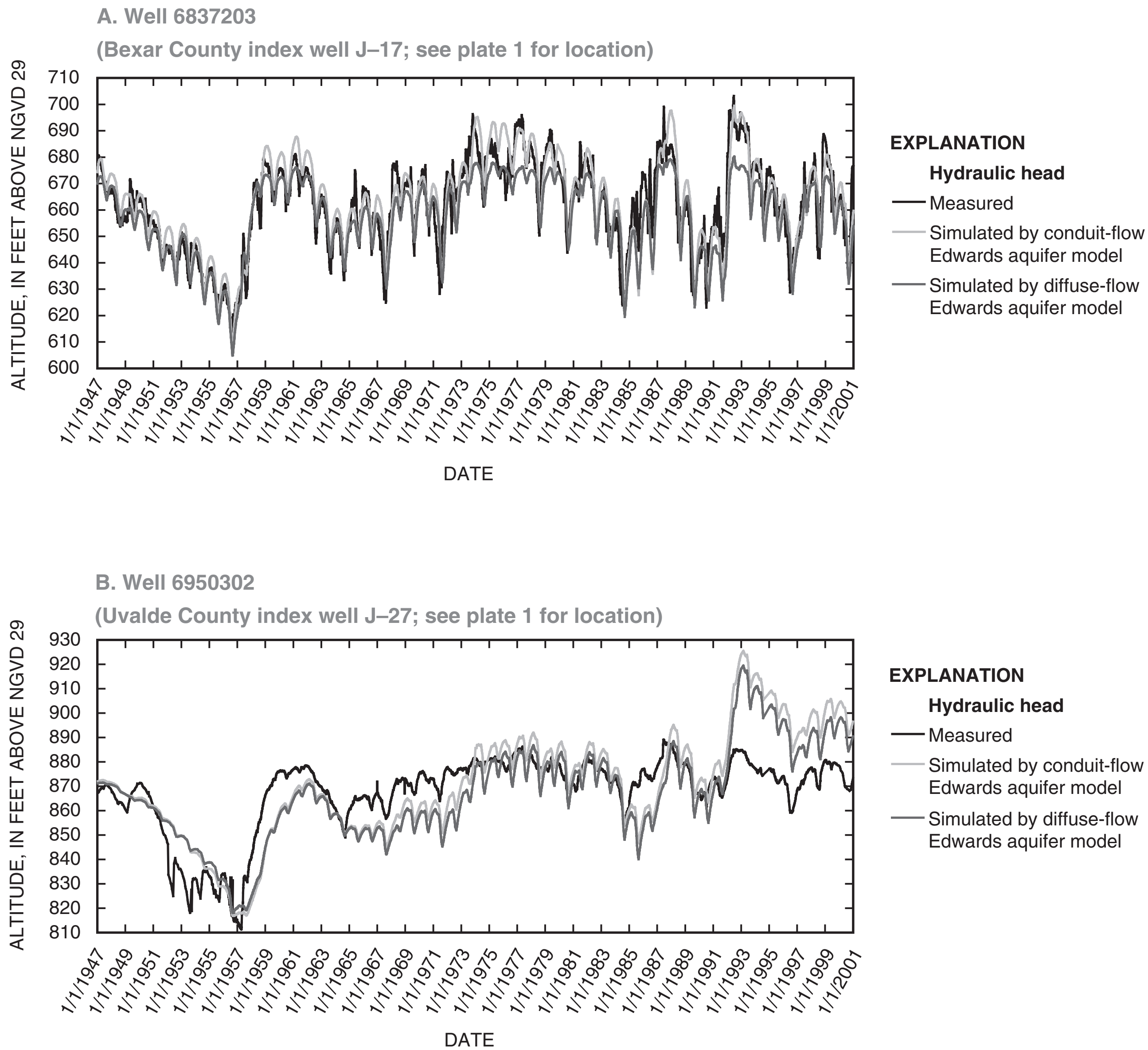


Figure 11. Measured and simulated water levels (hydraulic heads) for (A) Bexar County index well (J–17, 6837203) and (B) Uvalde County index well (J–27, 6950302), 1947–2000, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas.

An additional limitation results from the calibration technique, trial-and-error calibration, that was used for both the conduit-flow Edwards aquifer model and the diffuse-flow Edwards aquifer model. Manual trial-and-error adjustment of input data does not give information on the degree of uncertainty in the final input data selection (confidence intervals on input data estimates cannot be determined), nor does it guarantee the statistically best solution. Also, the trial-and-error process is influenced by the modeler's expertise and biases and is therefore a subjective solution. An alternative calibration technique is an automated statistically-based solution of the inverse problem, generally called parameter estimation. Parameter estimation quantifies the uncertainty in estimates of input data and gives the statistically most appropriate solution for the given input data, provided it is based on an appropriate statistical model of errors (Anderson and Woessner, 1992). However, the use of parameter estimation was beyond the scope of the study of this report.

Summary of important limitations and uncertainties associated with conceptual and numerical models, input data, and scale of application

1. Assumptions for conceptual and numerical models

    a. Use of a distributed, porous media model to simulate flow in a karst system results in

    - Inability to simulate rapid, potentially turbulent flow in conduits
    - Inability to simulate travel times for contaminants in the aquifer system

    b. Discretization of the model grid

    - Vertical: one model layer
    - Horizontal: relatively coarse cell size

    c. Temporal discretization for transient simulation

    - Monthly stress periods

    d. Representation of boundary conditions

    - Placement of the southern model boundary at the 10,000-mg/L dissolved solids concentration line
    - Use of a constant-flux boundary condition for the northern model boundary for the transient simulation
    - Assumption that the effectiveness of a fault as a barrier to flow is proportional to the fault displacement

2. Limitations of input data

    a. Datasets based on scanty information for some parameters and in some areas

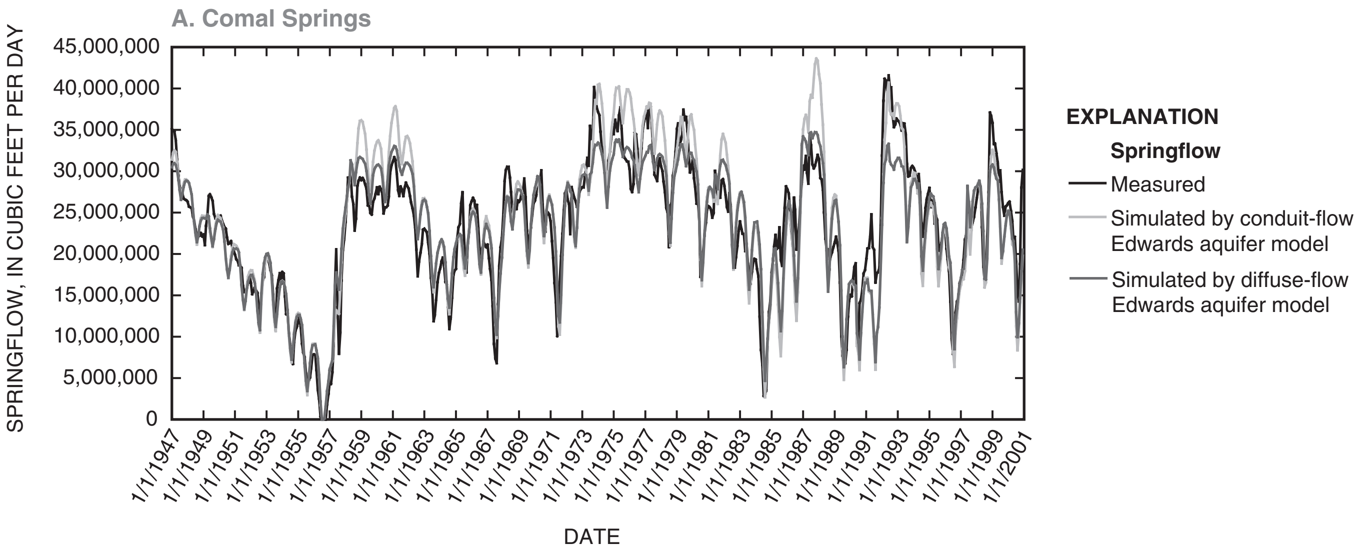
    - Parameters and data based on scanty information: (1) storativity distribution and (2) water-level data
    - Area with scanty information: (1) recharge zone, (2) Kinney County, and (3) south of the 1,000-mg/L dissolved solids concentration line

    b. Data of uncertain accuracy that warrants further analysis

    - Recharge
    - Withdrawals by wells
    - Location and characteristics of high-permeability zones or conduits

3. Scale of application

    a. The Edwards aquifer model is regional in scale, and therefore its application to local, site-specific issues is not appropriate



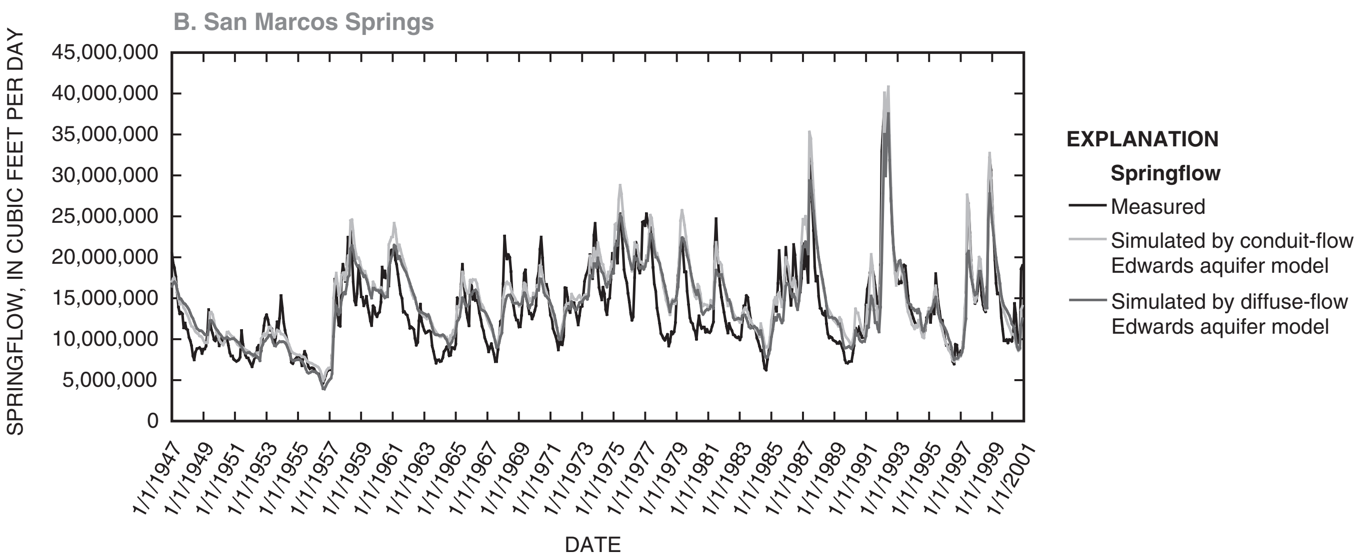


Figure 12. Measured and simulated springflows for (A) Comal Springs and (B) San Marcos Springs, 1947–2000, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas.

Summary

A numerical ground-water-flow model (hereinafter, the conduit-flow Edwards aquifer model) of the karstic Edwards aquifer in south-central Texas was completed as part of a study conducted during 2000–2003 by the U.S. Geological Survey (USGS) and The University of Texas at Austin, Bureau of Economic Geology, in cooperation with the U.S. Department of Defense and the Edwards Aquifer Authority. The conceptualization that served as the basis for the conduit-flow Edwards aquifer model emphasizes conduit development and conduit flow, and included simulating conduits as one-cell-wide, continuously connected features. Uncertainties regarding the degree to which conduits pervade the Edwards aquifer and influence ground-water flow, as well as uncertainties inherent in simulating conduits as one-cell-wide, continuously connected features, raised the question of whether a model based on the conduit-flow conceptualization was the optimum model for the Edwards aquifer. Accordingly, a model with a hydraulic conductivity distribution without conduits was considered. To develop an alternative, non-conduit or diffuse-flow hydraulic conductivity distribution for the Edwards aquifer model, the

USGS, in cooperation with the San Antonio Water System, conducted a study during 2004–05.

Calibration of the Edwards aquifer model with the alternative hydraulic conductivity distribution (hereinafter, the diffuse-flow Edwards aquifer model) included changes to the simulated hydraulic conductivity distribution only. All other properties were held constant at the same values used in the conduit-flow Edwards aquifer model. The hydraulic conductivity distribution for the diffuse-flow Edwards aquifer model, based primarily on a conceptualization in which flow in the aquifer predominantly is through a network of numerous small fractures and openings, includes 38 zones with hydraulic conductivities ranging from 3 to 50,000 ft/d. The initial hydraulic conductivity distribution for the diffuse-flow Edwards aquifer model was derived primarily from (1) estimates of relative transmissivities made by Maclay and Small (1984), based on available geologic, hydrochemical, and hydrologic information, for the recharge zone; (2) the hydraulic conductivity distribution developed by Painter and others (2002) for the freshwater confined zone; and (3) geologic structure and geochemical information used to delineate zones of high hydraulic conductivity. The final calibrated distribution of hydraulic conductivity results from the outcome of trial-and-error calibration, by varying the configuration of some of the simulated hydraulic conductivity zones and the simulated hydraulic conductivities.

Calibration and evaluation of the diffuse-flow (and the conduit-flow) Edwards aquifer model were conducted for steady-state (1939–46) and transient (1947–2000) conditions. Simulated hydraulic heads were within 30 ft of measured water levels at 130 of 144 wells used as targets for the steady-state simulation. The difference was less than 20 ft at 122 of the 144 wells. The RMS error for the calibrated steady-state simulation is 20.9 ft. This error represents about 3 percent of the total head difference across the model area. The simulated springflows for Comal and San Marcos Springs for the calibrated steady-state simulation were within 2.4 and 15 percent of the median springflows for the two springs, respectively.

The transient calibration period for the diffuse-flow Edwards aquifer model was 1947–2000, with 648 monthly stress periods, the same as for the conduit-flow Edwards aquifer model. The closest-match simulated hydraulic heads for a period of drought (May–November 1956) for the calibrated transient simulation were within 30 ft of measured water levels at 141 of the 171 wells for which water-level data were available. The difference was less than 20 ft at 131 of the 171 wells. The RMS error is 33.4 ft, which represents about 5 percent of the total head difference across the model area (712 ft). The closest-match simulated hydraulic heads for a period of above-normal rainfall (November 1974–July 1975) for the calibrated transient simulation were within 30 ft of measured water levels at 138 of the 169 wells for which water-level data were available. The difference was less than 20 ft at 125 of the 169 wells. The RMS error is 25.8 ft, which represents about 4 percent of the total head difference across the model area.

The RMS error ranged from 6.3 to 30.4 ft in 12 target wells with long-term water-level measurements for varying periods during 1947–2000 for the calibrated transient simulation for the diffuse-flow Edwards aquifer model, and these errors represent 5.0 to 31.3 percent of the range in water-level fluctuations of each of those wells. The three wells with RMS errors greater than 20 ft are all in Hays County. The RMS errors for the five major springs in the San Antonio segment of the aquifer for the calibrated transient simulation, as a percentage of the range of discharge fluctuations measured at the springs, varied from 7.2 percent for San Marcos Springs and 8.1 percent for Comal Springs to 28.8 percent for Leona Springs. The simulated spring discharges for San Antonio Springs tend to overestimate the measured discharges and for San Pedro Springs tend to underestimate the measured discharges during peak discharge periods.

In addition to hydraulic heads, the water budget for the model is computed for each stress period. A water budget in the context of the model is an accounting of inflow to, outflow from, and storage change in the aquifer. The steady-state simulation water budget indicates that recharge accounts for 93.6 percent of the sources of water to the Edwards aquifer, and inflow through the northern and northwestern model boundaries contributes 6.4 percent. The largest discharges from the Edwards aquifer in the steady-state simulation water budget are springflow (73.4 percent) and withdrawals by wells (25.7 percent).

The principal source of water to the Edwards aquifer (excluding change in storage) for the transient simulation is recharge, constituting about 60 percent of the sources of water to the Edwards aquifer during 1956, a drought period. Although the amount of water contributed by boundary inflow was relatively small, it constituted about 40 percent of the sources (excluding change in storage) to the aquifer during 1956, because of the greatly reduced recharge during this drought period. The principal discharges from the Edwards aquifer for the transient simulation during 1956 were withdrawals (pumpage) (76.1 percent), followed by springflow (23.2 percent). During 1956, the change in storage (net water released from storage) was much greater than recharge, accounting for 75.0 percent of the total flow (including change in storage) compared to 15.0 percent for recharge. The amount and percentage of water released from storage is large during 1956 (drought conditions) because recharge is small and more water is required from storage to meet the pumpage demand.

Hydraulic heads and springflows simulated by the diffuse-flow Edwards aquifer model for selected observation wells and springs were compared to the corresponding hydraulic heads and springflows simulated by the conduit-flow Edwards aquifer model. The mean absolute difference and the RMS error of the residuals for hydraulic heads for the diffuse-flow Edwards aquifer model are appreciably smaller than those for the conduit-flow Edwards aquifer model for the three synoptic water-level time periods (steady state, drought, and above-normal rainfall). The RMS errors for the conduit-flow Edwards aquifer model are 27, 76, and 30 percent greater than those for the diffuse-flow Edwards aquifer model for the steady-state, drought, and above-normal rainfall synoptic time periods, respectively. The improvements in the match between simulated and measured hydraulic head were predominantly for wells in or near the recharge zone of the aquifer. The general pattern of lower simulated hydraulic heads for the diffuse-flow Edwards aquifer model compared to that for the conduit-flow Edwards aquifer model is indicated by the consistently lower mean algebraic differences for the diffuse-flow Edwards aquifer model for all three synoptic water-level time periods.

The simulated steady-state springflows for the diffuse-flow Edwards aquifer model and the conduit-flow Edwards aquifer model generally are similar. Differences are 0.6 and 2.0 percent of the median steady-state springflows for Comal and San Marcos Springs, respectively. For the transient simulation, the mean absolute difference and RMS error of the residuals for the diffuse-flow Edwards aquifer model are smaller than those for the conduit-flow Edwards aquifer model for six of the 12 target wells with long-term water-level measurements and larger for the six other wells. The largest reductions in the RMS error for the diffuse-flow Edwards aquifer model occurred for two wells in Bexar County (-11.1 and -8.0 ft). The largest increases in the RMS error occurred for two wells in Hays County (15.0 and 15.1 ft). The hydrographs for the Bexar County index well (6837203) and the Uvalde County index well (6950302) generally illustrate lower water levels simulated by the diffuse-flow Edwards aquifer model than those simulated by the conduit-flow Edwards aquifer model.

The goodness-of-fit between measured and simulated springflows is similar for Comal, San Marcos, and Leona Springs for the diffuse-flow Edwards aquifer model and the conduit-flow Edwards aquifer model. The RMS errors for Comal and Leona Springs were 15.6 and 21.3 percent less, respectively, whereas the RMS error for San Marcos Springs was 3.3 percent greater for the diffuse-flow Edwards aquifer model compared to the conduit-flow Edwards aquifer model. The RMS errors for San Antonio and San Pedro Springs were appreciably greater, 80.2 and 51.0 percent, respectively, for the diffuse-flow Edwards aquifer model. The mean algebraic differences indicate that the springflows are less for all the springs, except San Antonio Springs, for the diffuse-flow Edwards aquifer model compared to those for the conduit-flow Edwards aquifer model. This, coupled with the observation that greater differences occur during the periods of greatest springflows, might indicate that the diffuse-flow Edwards aquifer model is somewhat less responsive than the conduit-flow Edwards aquifer model.

The simulated water budgets for the diffuse-flow Edwards aquifer model are similar to those for the conduit-flow Edwards aquifer model. For the steady-state simulation, the largest differences in percentage of total sources or discharges for a budget component are 0.3 percent for two discharges, springflow and stream-aquifer leakage. For 1956 for the transient simulation, the largest differences in percentage of total sources or discharges for a budget component are an increase of 1.8 percent for withdrawals (pumpage) (increase of 749.7 acre-ft/yr) and a decrease of 2.0 percent for springflow (decrease of 10,732.1 acre-ft/yr). In addition, the magnitude (decrease of 10,497.6 acre-ft/yr) and percentage (decrease of 0.9 percent) for the net change in storage (net loss of water from storage) were less for the diffuse-flow Edwards aquifer model.

The largest change in terms of the magnitude of water budget components was the decrease of about 10,730 acre-ft/yr in springflow for the diffuse-flow Edwards aquifer model compared to springflow for the conduit-flow Edwards aquifer model. This decrease in springflow (a water budget discharge) was largely offset by the smaller net loss of water from storage (a water budget source) of about 10,500 acre-ft/yr. The decrease in springflow for the diffuse-flow Edwards aquifer model compared to that for the conduit-flow Edwards aquifer model might be due to the general pattern of lower simulated hydraulic heads for the diffuse-flow Edwards aquifer model compared to that for the conduit-flow Edwards aquifer model.

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Table 1. Steady-state simulation target wells and residuals, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas.

[Measured and simulated water levels in altitude above National Geodetic Vertical Datum of 1929. TWDB, Texas Water Development Board; ID, identification number; DD.MMSS, degrees.minutes/seconds; residual, simulated water level minus measured water level; mean absolute difference, sum of absolute values of residuals divided by number of wells; mean algebraic difference, algebraic sum of residuals divided by number of wells; RMS error, root-mean-square error]

TWDB well ID	County	Latitude (DD.MMSS)	Longitude (DD.MMSS)	Measured water level (feet)	Conduit-flow Edwards aquifer model (Lindgren and others, 2004)		Diffuse-flow Edwards aquifer model	
					Simulated water level (feet)	Residual (feet)	Simulated water level (feet)	Residual (feet)
7038601	Kinney	100.2569	29.4392	1,216	1,188.7	-27.3	1,203.4	-12.6
7038903	Kinney	100.2731	29.4122	1,211	1,189.7	-21.3	1,185.8	-25.2
7038906	Kinney	100.2833	29.3792	1,141	1,123.2	-17.8	1,137.9	-3.1
7045502	Kinney	100.4250	29.3136	1,109	1,105.6	-3.4	1,105.3	-3.7
7045602	Kinney	100.4125	29.3269	1,099	1,106.1	7.1	1,105.5	6.5
7046101	Kinney	100.3428	29.3436	1,116	1,108.2	-7.8	1,110.3	-5.7
7046302	Kinney	100.2622	29.3367	1,079	1,073.0	-6.0	1,087.7	8.7
7046901	Kinney	100.2667	29.2703	1,044	1,042.6	-1.4	1,045.0	1.0
7047402	Kinney	100.2431	29.3236	1,072	1,052.1	-19.9	1,073.2	1.2
7047802	Kinney	100.2078	29.2631	999	989.1	-9.9	999.7	.7
7048701	Kinney	100.1139	29.2911	964	928.7	-35.3	967.6	3.6
6935601	Uvalde	99.6297	29.4311	898	920.2	22.2	904.3	6.3
6935804	Uvalde	99.7017	29.3950	1,009	927.2	-81.8	1,012.1	3.1
6941701	Uvalde	99.9711	29.2603	917	889.8	-27.2	923.9	6.9
6943204	Uvalde	99.6744	29.3628	881	885.4	4.4	870.0	-11.0
6943404	Uvalde	99.7367	29.3042	859	871.0	12.0	859.0	0
6943604	Uvalde	99.6586	29.2933	888	861.9	-26.1	804.6	-83.4
6944705	Uvalde	99.6028	29.2644	763	808.2	45.2	792.9	29.9
6950302	Uvalde	99.7867	29.2103	870	872.0	2.0	871.1	1.1
6950304	Uvalde	99.7531	29.2294	870	871.7	1.7	867.6	-2.4
6950609	Uvalde	99.7825	29.1875	867	871.1	4.1	871.1	4.1
6825708	Medina	98.9772	29.5153	885	858.3	-26.7	884.4	-.6
6825912	Medina	98.8881	29.5017	782	788.0	6.0	841.9	59.9
6833101	Medina	98.9839	29.4747	824	823.8	-.2	824.4	.4
6833107	Medina	98.9764	29.4644	841	804.6	-36.4	779.9	-61.1
6833208	Medina	98.9269	29.4658	822	759.7	-62.3	783.7	-38.3
6833209	Medina	98.9253	29.4764	824	782.6	-41.4	825.2	1.2
6833210	Medina	98.9356	29.4858	849	826.2	-22.8	870.2	21.2
6833211	Medina	98.9500	29.4744	846	766.2	-79.8	773.3	-72.7
6833303	Medina	98.9025	29.4981	837	800.5	-36.5	844.3	7.3
6833503	Medina	98.9431	29.4519	728	741.3	13.3	729.7	1.7
6833604	Medina	98.8881	29.4219	705	722.3	17.3	709.5	4.5
6834106	Medina	98.8528	29.4917	754	745.8	-8.2	761.5	7.5
6834706	Medina	98.8661	29.4153	698	720.2	22.2	707.9	9.9
6841605	Medina	98.9011	29.3139	703	715.2	12.2	704.7	1.7
6842224	Medina	98.8189	29.3592	677	711.1	34.1	696.7	19.7
6842226	Medina	98.8244	29.3656	659	712.0	53.0	697.8	38.8
6842504	Medina	98.8147	29.3094	691	708.7	17.7	700.9	9.9
6938602	Medina	99.2725	29.4328	824	789.2	-34.8	831.5	7.5
6939507	Medina	99.1947	29.4531	827	805.2	-21.8	825.8	-1.2
6939903	Medina	99.1336	29.4042	760	776.5	16.5	761.4	1.4
6940405	Medina	99.1022	29.4347	768	789.8	21.8	768.2	.2
6945601	Medina	99.4089	29.3328	761	760.3	-.7	768.7	7.7
6946701	Medina	99.3525	29.2639	754	750.1	-3.9	762.6	8.6
6947514	Medina	99.1989	29.2922	812	740.9	-71.1	747.5	-64.5
6954401	Medina	99.3342	29.1864	755	743.1	-11.9	757.5	2.5
6822701	Bexar	98.3739	29.6517	749	724.7	-24.3	748.2	-.8
6822702	Bexar	98.3489	29.6250	663	684.8	21.8	670.6	7.6

Table 1. Steady-state simulation target wells and residuals for the conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas—Continued.

TWDB well ID	County	Latitude (DD.MMSS)	Longitude (DD.MMSS)	Measured water level (feet)	Conduit-flow Edwards aquifer model (Lindgren and others, 2004)		Diffuse-flow Edwards aquifer model	
					Simulated water level (feet)	Residual (feet)	Simulated water level (feet)	Residual (feet)
6826901	Bexar	98.7536	29.5236	734	740.6	6.6	728.0	-6.0
6827514	Bexar	98.6803	29.5619	724	793.9	69.9	727.2	3.2
6827515	Bexar	98.6747	29.5575	723	763.6	40.6	720.9	-2.1
6827701	Bexar	98.7356	29.5122	702	730.2	28.2	708.8	6.8
6827702	Bexar	98.7267	29.5036	698	720.5	22.5	699.9	1.9
6828201	Bexar	98.5797	29.5942	715	789.1	74.1	721.5	6.5
6828507	Bexar	98.5514	29.5422	679	701.4	22.4	678.9	-.1
6828704	Bexar	98.6008	29.5250	701	688.4	-12.6	698.4	-2.6
6829207	Bexar	98.4267	29.6164	688	716.5	28.5	680.3	-7.7
6829411	Bexar	98.4847	29.5642	674	688.6	14.6	671.4	-2.6
6829502	Bexar	98.4500	29.5708	674	683.8	9.8	669.9	-4.1
6829604	Bexar	98.4094	29.5511	668	667.4	-.6	668.0	0
6829605	Bexar	98.3872	29.5500	668	663.3	-4.7	667.1	-.9
6829815	Bexar	98.4242	29.5031	682	670.0	-12.0	668.7	-13.3
6829916	Bexar	98.4078	29.5103	669	669.1	.1	668.3	-.7
6830101	Bexar	98.3553	29.5919	667	663.5	-3.5	664.3	-2.7
6830612	Bexar	98.2889	29.5431	670	651.9	-18.1	664.5	-5.5
6830706	Bexar	98.3706	29.5083	667	667.1	.1	667.6	.6
6830801	Bexar	98.3139	29.5175	671	657.7	-13.3	666.0	-5.0
6835202	Bexar	98.6906	29.4889	691	707.9	16.9	694.6	3.6
6835312	Bexar	98.6558	29.4717	690	700.5	10.5	686.4	-3.6
6835807	Bexar	98.7075	29.3778	685	696.7	11.7	686.3	1.3
6836113	Bexar	98.5861	29.4664	671	689.1	18.1	677.7	6.7
6836505	Bexar	98.5461	29.4531	674	684.4	10.4	675.2	1.2
6836506	Bexar	98.5711	29.4297	676	685.9	9.9	676.2	.2
6836606	Bexar	98.5072	29.4328	668	680.7	12.7	672.9	4.9
6836706	Bexar	98.6242	29.4061	682	691.6	9.6	680.8	-1.2
6836709	Bexar	98.6244	29.3822	675	689.2	14.2	675.9	.9
6836713	Bexar	98.5992	29.3878	673	686.8	13.8	675.4	2.4
6836812	Bexar	98.5819	29.4094	673	686.8	13.8	675.4	2.4
6836813	Bexar	98.5667	29.3922	675	684.7	9.7	674.9	-.1
6836910	Bexar	98.5225	29.4011	674	682.3	8.3	673.9	-.1
6836930	Bexar	98.5314	29.3939	675	683.1	8.1	674.2	-.8
6837114	Bexar	98.4900	29.4703	672	675.8	3.8	671.3	-.7
6837126	Bexar	98.4692	29.4653	681	671.4	-9.6	670.4	-10.6
6837204	Bexar	98.4272	29.4717	675	674.2	-.8	669.9	-5.1
6837407	Bexar	98.4633	29.4364	671	678.4	7.4	671.7	.7
6837408	Bexar	98.4789	29.4267	671	679.7	8.7	672.2	1.2
6837409	Bexar	98.4956	29.4183	616	680.7	64.7	672.8	56.8
6837411	Bexar	98.4894	29.4278	659	680.1	21.1	672.5	13.5
6837517	Bexar	98.4303	29.4386	623	676.5	53.5	670.7	47.7
6837606	Bexar	98.4108	29.4456	677	675.5	-1.5	670.1	-6.9
6837707	Bexar	98.4928	29.4006	675	683.2	8.2	673.0	-2.0
6837715	Bexar	98.4889	29.3983	666	683.2	17.2	673.0	7.0
6837716	Bexar	98.4878	29.4142	665	680.6	15.6	672.7	7.7
6842314	Bexar	98.7825	29.3706	692	705.9	13.9	694.1	2.1
6842315	Bexar	98.7611	29.3739	689	702.9	13.9	692.0	3.0
6843108	Bexar	98.7272	29.3528	688	698.9	10.9	689.6	1.6
6843611	Bexar	98.6636	29.3264	673	693.1	20.1	679.9	6.9
6843812	Bexar	98.6839	29.2806	679	695.8	16.8	686.3	7.3
6843813	Bexar	98.6700	29.2703	688	695.1	7.1	685.1	-2.9

Table 1. Steady-state simulation target wells and residuals for the conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas—Continued.

TWDB well ID	County	Latitude (DD.MMSS)	Longitude (DD.MMSS)	Measured water level (feet)	Conduit-flow Edwards aquifer model (Lindgren and others, 2004)		Diffuse-flow Edwards aquifer model	
					Simulated water level (feet)	Residual (feet)	Simulated water level (feet)	Residual (feet)
6844220	Bexar	98.5714	29.3494	676	686.7	10.7	675.4	-0.6
6844221	Bexar	98.5692	29.3458	673	686.8	13.8	675.4	2.4
6844222	Bexar	98.5719	29.3436	675	687.0	12.0	675.5	.5
6844223	Bexar	98.5678	29.3431	676	686.9	10.9	675.4	-.6
6844405	Bexar	98.6100	29.3050	671	690.6	19.6	677.3	6.3
6844602	Bexar	98.5378	29.3294	655	686.3	31.3	674.9	19.9
6845102	Bexar	98.4972	29.3742	660	683.9	23.9	673.6	13.6
6815904	Comal	98.1625	29.7503	640	634.7	-5.3	635.7	-4.3
6816703	Comal	98.0928	29.7531	601	611.1	10.1	613.5	12.5
6816704	Comal	98.1097	29.7639	612	632.9	20.9	618.5	6.5
6816801	Comal	98.0528	29.7861	594	594.5	.5	602.7	8.7
6816804	Comal	98.0628	29.7778	608	598.4	-9.6	605.3	-2.7
6822301	Comal	98.2578	29.7119	677	713.5	36.5	667.7	-9.3
6822303	Comal	98.2517	29.7133	662	710.2	48.2	659.2	-2.8
6822502	Comal	98.2964	29.6975	670	722.7	52.7	676.5	6.5
6822503	Comal	98.3161	29.6844	701	720.9	19.9	686.3	-14.7
6822601	Comal	98.2858	29.6772	657	701.6	44.6	661.2	4.2
6822803	Comal	98.3225	29.6392	666	689.1	23.1	672.1	6.1
6822804	Comal	98.3306	29.6311	673	685.2	12.2	670.0	-3.0
6823101	Comal	98.2106	29.7403	668	657.2	-10.8	647.8	-20.2
6823102	Comal	98.2250	29.7261	767	661.0	-106.0	650.1	-116.9
6823206	Comal	98.1819	29.7483	659	636.7	-22.3	639.4	-19.6
6823209	Comal	98.1792	29.7206	635	644.7	9.7	642.5	7.5
6823210	Comal	98.2078	29.7147	643	647.5	4.5	647.3	4.3
6823212	Comal	98.1728	29.7433	637	633.9	-3.1	638.4	1.4
6823220	Comal	98.2081	29.7419	668	648.7	-19.3	647.1	-20.9
6823306	Comal	98.1528	29.7464	630	633.7	3.7	634.9	4.9
6823308	Comal	98.1453	29.7475	631	633.7	2.7	633.5	2.5
6823309	Comal	98.1347	29.7183	625	640.3	15.3	637.8	12.8
6823507	Comal	98.1989	29.6889	641	658.0	17.0	647.7	6.7
6823604	Comal	98.1633	29.6700	638	622.4	-15.6	633.9	-4.1
6823705	Comal	98.2378	29.6344	652	635.4	-16.6	649.4	-2.6
6824106	Comal	98.1133	29.7344	620	636.1	16.1	631.7	11.7
6830216	Comal	98.3286	29.6203	671	678.2	7.2	663.2	-7.8
6830313	Comal	98.2572	29.6167	619	639.6	20.6	654.3	35.3
5857301	Hays	97.8903	30.0936	621	607.7	-13.3	650.7	29.7
5858101	Hays	97.8422	30.0836	589	598.5	9.5	614.2	25.2
5858104	Hays	97.8486	30.1042	589	583.8	-5.2	592.3	3.3
5858703	Hays	97.8542	30.0278	579	600.0	21.0	630.1	51.1
6701307	Hays	97.8869	29.9675	569	588.9	19.9	604.7	35.7
6701702	Hays	97.9650	29.8958	577	594.5	17.5	592.8	15.8
6701807	Hays	97.9192	29.9008	574	587.6	13.6	588.1	14.1
6702103	Hays	97.8725	29.9889	579	589.4	10.4	615.4	36.4
6709102	Hays	97.9758	29.8508	577	588.0	11.0	590.9	13.9
6816605	Hays	98.0042	29.8289	566	588.6	22.6	593.9	27.9
Statistics:								
Mean absolute difference						19.4		11.3
Mean algebraic difference						4.5		1.5
RMS error						26.5		20.9

Table 2. Comparison of the residuals for hydraulic heads, by synoptic water-level time period (steady-state, below-normal rainfall [drought], and above-normal rainfall), conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas.

[residual, simulated water level minus measured water level; MAE (mean absolute difference), sum of absolute values of residuals divided by number of wells; ME (mean algebraic difference), algebraic sum of residuals divided by number of wells; RMS, root-mean-square error]

Synoptic time period	Hydraulic-head residuals (feet)					
	Conduit-flow Edwards aquifer model (Lindgren and others, 2004)			Diffuse-flow Edwards aquifer model		
	MAE	ME	RMS	MAE	ME	RMS
Steady state	19.4	4.5	26.5	11.3	1.5	20.9
Drought	31.6	-7.6	58.7	20.5	-8.1	33.4
Above-normal rainfall	23.5	3.5	33.5	17.2	.5	25.8

Table 3. Measured and simulated springflows for steady-state simulation, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas.

[Simulated, model-computed springflow for steady-state simulation]

Year	Springflow (cubic feet per second)									
	Comal Springs		San Marcos Springs		Leona Springs		San Antonio Springs		San Pedro Springs	
	Mean	Median	Mean	Median	Mean	Median	Mean	Median	Mean	Median
	Measured									
1939	297	301	98	103	21	20	4	0	5	4
1940	275	274	107	107	18	18	0	0	3	3
1941	340	339	182	170	21	23	28	29	9	9
1942	346	333	153	135	26	27	29	23	9	8
1943	337	325	133	134	21	20	13	10	7	6
1944	342	331	183	178	11	10	11	8	6	6
1945	357	342	189	176	11	13	23	16	8	7
1946	356	358	181	173	3	2	16	9	7	6
Mean (1939–46)	332	326	153	147	16	17	16	12	7	6
Median (1939–46)	341	332	167	152	20	19	14	10	7	6
	Simulated									
Conduit-flow Edwards aquifer model	342		172		40		16		8	
Diffuse-flow Edwards aquifer model	340		175		40		16		5	

Table 4. Transient simulation target wells and residuals for drought conditions, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas.

[Measured and simulated water levels in altitude above National Geodetic Vertical Datum of 1929. TWDB, Texas Water Development Board; ID, identification number; DD.MMSS, degrees.minutes/seconds; residual, simulated water level minus measured water level; mean absolute difference, sum of absolute values of residuals divided by number of wells; mean algebraic difference, algebraic sum of residuals divided by number of wells; RMS error, root-mean-square error]

TWDB well ID	County	Latitude (DD.MMSS)	Longitude (DD.MMSS)	Measured water level (feet)	Conduit-flow Edwards aquifer model (Lindgren and others, 2004)		Diffuse-flow Edwards aquifer model	
					Simulated water level (feet)	Residual (feet)	Simulated water level (feet)	Residual (feet)
7038501	Kinney	100.3083	29.4328	1,221	1,186.5	-34.5	1,183.5	-37.5
7038901	Kinney	100.2606	29.4114	1,158	1,186.5	28.5	1,180.8	22.8
7038906	Kinney	100.2833	29.3792	1,137	1,181.3	44.3	1,134.5	-2.5
7045502	Kinney	100.4250	29.3136	1,109	1,122.0	13.0	1,105.1	-3.9
7045602	Kinney	100.4125	29.3269	1,095	1,105.6	10.6	1,105.3	10.3
7046901	Kinney	100.2667	29.2703	1,034	1,106.0	72.0	1,033.0	-1.0
7047101	Kinney	100.2108	29.3572	1,058	1,039.8	-18.2	1,093.1	35.1
7047201	Kinney	100.1750	29.3636	1,011	1,063.0	52.0	1,092.8	81.8
7047301	Kinney	100.1461	29.3417	979	1,041.9	62.9	1,039.8	60.8
7047303	Kinney	100.1617	29.3531	979	977.4	-1.6	1,058.3	79.3
7047501	Kinney	100.1711	29.3206	926	1,009.9	83.9	1,024.5	98.5
6933601	Uvalde	99.8972	29.4264	1,231	980.9	-250.1	1,080.0	-151.0
6933901	Uvalde	99.8783	29.3936	1,119	941.9	-177.1	1,014.8	-104.2
6935701	Uvalde	99.7461	29.3917	979	911.0	-68.0	960.2	-18.8
6935804	Uvalde	99.7017	29.3950	934	879.2	-54.8	923.4	-10.6
6935901	Uvalde	99.6456	29.3869	717	862.6	145.6	769.5	52.5
6937401	Uvalde	99.4908	29.4575	799	826.0	27.0	764.6	-34.4
6941101	Uvalde	99.9606	29.3442	1,001	729.9	-271.1	976.3	-24.7
6941701	Uvalde	99.9711	29.2603	839	913.5	74.5	885.1	46.1
6941903	Uvalde	99.9089	29.2603	840	843.5	3.5	862.6	22.6
6942101	Uvalde	99.8561	29.3419	929	830.6	-98.4	902.0	-27.0
6942603	Uvalde	99.7550	29.3083	800	854.2	54.2	807.9	7.9
6942604	Uvalde	99.7708	29.3125	801	816.6	15.6	814.0	13.0
6942911	Uvalde	99.7614	29.2883	802	818.5	16.5	809.1	7.1
6942912	Uvalde	99.7758	29.2661	814	816.4	2.4	813.8	-.2
6943106	Uvalde	99.7389	29.3550	847	816.8	-30.2	811.7	-35.3
6943204	Uvalde	99.6744	29.3628	791	822.9	31.9	794.3	3.3
6943404	Uvalde	99.7367	29.3042	790	821.9	31.9	799.6	9.6
6943406	Uvalde	99.7494	29.3083	776	814.2	38.2	805.6	29.6
6943501	Uvalde	99.6906	29.3292	769	815.9	46.9	784.6	15.6
6943911	Uvalde	99.6500	29.2892	755	811.4	56.4	720.7	-34.3
6944705	Uvalde	99.6028	29.2644	662	796.3	134.3	708.2	46.2
6945401	Uvalde	99.4681	29.3192	666	727.8	61.8	688.7	22.7
6950101	Uvalde	99.8506	29.2406	833	678.6	-154.4	838.6	5.6
6950202	Uvalde	99.7981	29.2372	818	821.6	3.6	819.6	1.6
6950204	Uvalde	99.8247	29.2203	821	817.9	-3.1	824.3	3.3
6950302	Uvalde	99.7867	29.2103	818	819.0	1.0	818.1	.1
6950305	Uvalde	99.7550	29.2261	813	817.0	4.0	814.1	1.1
6950306	Uvalde	99.7522	29.2322	808	816.1	8.1	812.2	4.2
6950406	Uvalde	99.8667	29.1806	824	816.1	-7.9	840.1	16.1
6950507	Uvalde	99.8289	29.1861	820	819.0	-1.0	826.8	6.8
6950609	Uvalde	99.7825	29.1875	819	818.0	-1.0	819.2	.2
6950612	Uvalde	99.7533	29.1789	822	816.5	-5.5	815.5	-6.5
6950901	Uvalde	99.7739	29.1664	814	813.2	-.8	818.8	4.8
6950902	Uvalde	99.7669	29.1653	840	815.8	-24.2	817.7	-22.3
6951407	Uvalde	99.7475	29.1819	827	815.0	-12.0	815.1	-11.9
6952401	Uvalde	99.6250	29.1972	758	812.4	54.4	703.6	-54.4

Table 4. Transient simulation target wells and residuals for drought conditions, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas—Continued.

TWDB well ID	County	Latitude (DD.MMSS)	Longitude (DD.MMSS)	Measured water level (feet)	Conduit-flow Edwards aquifer model (Lindgren and others, 2004)		Diffuse-flow Edwards aquifer model	
					Simulated water level (feet)	Residual (feet)	Simulated water level (feet)	Residual (feet)
6826702	Medina	98.8556	29.5122	614	708.6	94.6	653.4	39.4
6833303	Medina	98.9025	29.4981	783	637.6	-145.4	665.7	-117.3
6833604	Medina	98.8881	29.4219	628	660.7	32.7	625.0	-3.0
6841202	Medina	98.9297	29.3353	615	633.0	18.0	625.8	10.8
6841301	Medina	98.8797	29.3544	627	635.5	8.5	624.7	-2.3
6841403	Medina	98.9886	29.3203	640	631.8	-8.2	627.9	-12.1
6842504	Medina	98.8147	29.3094	626	639.2	13.2	621.2	-4.8
6932702	Medina	99.0939	29.5258	976	628.1	-347.9	800.2	-175.8
6932801	Medina	99.0797	29.5350	979	775.5	-203.5	835.7	-143.3
6938904	Medina	99.2653	29.3833	690	789.5	99.5	674.2	-15.8
6939507	Medina	99.1947	29.4531	740	676.0	-64.0	717.7	-22.3
6939903	Medina	99.1336	29.4042	659	703.2	44.2	670.8	11.8
6940101	Medina	99.1228	29.4906	775	679.2	-95.8	741.0	-34.0
6940405	Medina	99.1022	29.4347	673	737.7	64.7	676.1	3.1
6940901	Medina	99.0008	29.4094	637	688.8	51.8	629.4	-7.6
6945601	Medina	99.4089	29.3328	663	640.1	-22.9	680.0	17.0
6946601	Medina	99.2786	29.3319	659	674.3	15.3	668.3	9.3
6946701	Medina	99.3525	29.2639	659	661.8	2.8	674.9	15.9
6947301	Medina	99.1600	29.3597	634	665.9	31.9	640.2	6.2
6947402	Medina	99.2097	29.3331	642	652.8	10.8	657.7	15.7
6948402	Medina	99.1069	29.3247	636	656.3	20.3	634.8	-1.2
6954401	Medina	99.3342	29.1864	659	648.6	-10.4	673.3	14.3
6955202	Medina	99.2006	29.2269	640	661.0	21.0	659.4	19.4
6822702	Bexar	98.3489	29.6250	620	649.5	29.5	608.5	-11.5
6826804	Bexar	98.7978	29.5242	638	620.1	-17.9	631.1	-6.9
6826901	Bexar	98.7536	29.5236	632	632.6	.6	617.2	-14.8
6827515	Bexar	98.6747	29.5575	653	628.9	-24.1	611.6	-41.4
6827608	Bexar	98.6442	29.5436	644	500.6	-143.4	611.7	-32.3
6827702	Bexar	98.7267	29.5036	625	637.7	12.7	614.0	-11.0
6828201	Bexar	98.5797	29.5942	686	622.7	-63.3	608.0	-78.0
6828507	Bexar	98.5514	29.5422	620	600.2	-19.8	603.0	-17.0
6828705	Bexar	98.6247	29.5014	623	591.9	-31.1	605.9	-17.1
6828908	Bexar	98.5350	29.5183	621	594.8	-26.2	605.2	-15.8
6829207	Bexar	98.4267	29.6164	681	594.6	-86.4	610.8	-70.2
6829304	Bexar	98.4106	29.5975	623	608.1	-14.9	601.6	-21.4
6829411	Bexar	98.4847	29.5642	641	602.8	-38.2	602.9	-38.1
6829508	Bexar	98.4203	29.5775	619	571.2	-47.8	604.9	-14.1
6829604	Bexar	98.4094	29.5511	616	601.4	-14.6	604.8	-11.2
6829701	Bexar	98.4636	29.5369	617	602.1	-14.9	604.9	-12.1
6829709	Bexar	98.4617	29.5028	623	591.6	-31.4	604.6	-18.4
6829916	Bexar	98.4078	29.5103	636	599.8	-36.2	604.6	-31.4
6829917	Bexar	98.4078	29.5100	616	604.0	-12.0	604.6	-11.4
6830101	Bexar	98.3553	29.5919	617	604.0	-13.0	604.9	-12.1
6830404	Bexar	98.3606	29.5456	610	607.3	-2.7	604.7	-5.3
6830513	Bexar	98.3222	29.5475	617	607.5	-9.5	604.7	-12.3
6830514	Bexar	98.3194	29.5789	623	607.0	-16.0	604.7	-18.3
6830612	Bexar	98.2889	29.5431	618	606.4	-11.6	604.7	-13.3
6830707	Bexar	98.3706	29.5083	618	606.8	-11.2	604.7	-13.3
6830801	Bexar	98.3139	29.5175	616	608.4	-7.6	604.7	-11.3
6830802	Bexar	98.3139	29.5175	615	607.5	-7.5	604.7	-10.3

Table 4. Transient simulation target wells and residuals for drought conditions, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas—Continued.

TWDB well ID	County	Latitude (DD.MMSS)	Longitude (DD.MMSS)	Measured water level (feet)	Conduit-flow Edwards aquifer model (Lindgren and others, 2004)		Diffuse-flow Edwards aquifer model	
					Simulated water level (feet)	Residual (feet)	Simulated water level (feet)	Residual (feet)
6830901	Bexar	98.2669	29.5128	646	607.5	-38.5	604.7	-41.3
6834301	Bexar	98.7789	29.4753	629	607.4	-21.6	620.9	-8.1
6834602	Bexar	98.7900	29.4386	633	628.4	-4.6	621.2	-11.8
6834603	Bexar	98.7586	29.4169	624	627.6	3.6	614.1	-9.9
6835202	Bexar	98.6906	29.4889	624	618.3	-5.7	612.6	-11.4
6835311	Bexar	98.6503	29.4711	626	618.0	-8.0	608.5	-17.5
6835504	Bexar	98.6714	29.4392	620	609.7	-10.3	608.8	-11.2
6835807	Bexar	98.7075	29.3778	620	609.4	-10.6	611.4	-8.6
6835901	Bexar	98.6381	29.4147	617	615.7	-1.3	608.6	-8.4
6835911	Bexar	98.6506	29.3944	626	608.2	-17.8	609.3	-16.7
6836302	Bexar	98.5364	29.4722	621	609.4	-11.6	605.7	-15.3
6836303	Bexar	98.5094	29.4697	589	603.4	14.4	605.4	16.4
6836410	Bexar	98.6236	29.4558	619	602.6	-16.4	607.7	-11.3
6836507	Bexar	98.5442	29.4264	624	606.5	-17.5	605.9	-18.1
6836602	Bexar	98.5022	29.4411	615	610.0	-5.0	605.1	-9.9
6836604	Bexar	98.5411	29.4336	620	610.8	-9.2	605.9	-14.1
6836706	Bexar	98.6242	29.4061	620	609.3	-10.7	608.3	-11.7
6836911	Bexar	98.5314	29.4064	616	608.1	-7.9	605.8	-10.2
6836913	Bexar	98.5197	29.3806	615	612.1	-2.9	606.2	-8.8
6837114	Bexar	98.4900	29.4703	617	613.3	-3.7	605.0	-12.0
6837203	Bexar	98.4322	29.4708	621	602.1	-18.9	604.8	-16.2
6837204	Bexar	98.4272	29.4717	617	608.5	-8.5	604.8	-12.2
6837407	Bexar	98.4633	29.4364	616	608.1	-7.9	604.9	-11.1
6837409	Bexar	98.4956	29.4183	610	610.7	.7	605.1	-4.9
6837512	Bexar	98.4294	29.4383	618	611.3	-6.7	605.1	-12.9
6837513	Bexar	98.4281	29.4400	618	609.2	-8.8	605.1	-12.9
6837514	Bexar	98.4183	29.4344	615	609.0	-6.0	605.1	-9.9
6837606	Bexar	98.4108	29.4456	618	609.5	-8.5	605.0	-13.0
6837707	Bexar	98.4928	29.4006	618	609.6	-8.4	605.9	-12.1
6838109	Bexar	98.3467	29.4664	614	613.0	-1.0	604.8	-9.2
6838301	Bexar	98.2653	29.4725	611	609.0	-2.0	605.8	-5.2
6842314	Bexar	98.7825	29.3706	625	608.2	-16.8	616.3	-8.7
6842315	Bexar	98.7611	29.3739	624	623.0	-1.0	614.9	-9.1
6843205	Bexar	98.6978	29.3456	623	620.6	-2.4	612.1	-10.9
6843404	Bexar	98.7142	29.2961	613	619.4	6.4	616.8	3.8
6843507	Bexar	98.6961	29.2939	622	621.7	-.3	615.6	-6.4
6843605	Bexar	98.6533	29.3269	603	621.0	18.0	608.5	5.5
6843611	Bexar	98.6636	29.3264	617	617.9	.9	609.3	-7.7
6843807	Bexar	98.6719	29.2750	628	618.2	-9.8	613.8	-14.2
6843812	Bexar	98.6839	29.2806	625	620.5	-4.5	615.0	-10.0
6844213	Bexar	98.5614	29.3578	616	621.0	5.0	606.6	-9.4
6844405	Bexar	98.6100	29.3050	614	614.0	0	608.2	-5.8
6845102	Bexar	98.4972	29.3742	605	617.3	12.3	606.1	1.1
6815903	Comal	98.1414	29.7575	618	613.3	-4.7	599.4	-18.6
6815904	Comal	98.1625	29.7503	612	605.2	-6.8	601.5	-10.5
6815906	Comal	98.1597	29.7786	633	606.5	-26.5	617.2	-15.8
6816703	Comal	98.0928	29.7531	593	614.5	21.5	594.7	1.7
6816801	Comal	98.0528	29.7861	578	599.3	21.3	590.2	12.2
6816804	Comal	98.0628	29.7778	596	589.0	-7.0	591.4	-4.6
6822301	Comal	98.2578	29.7119	649	591.8	-57.2	614.8	-34.2

Table 4. Transient simulation target wells and residuals for drought conditions, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas—Continued.

TWDB well ID	County	Latitude (DD.MMSS)	Longitude (DD.MMSS)	Measured water level (feet)	Conduit-flow Edwards aquifer model (Lindgren and others, 2004)		Diffuse-flow Edwards aquifer model	
					Simulated water level (feet)	Residual (feet)	Simulated water level (feet)	Residual (feet)
6822601	Comal	98.2858	29.6772	628	640.2	12.2	610.1	-17.9
6823208	Comal	98.1936	29.7347	618	633.5	15.5	603.8	-14.2
6823212	Comal	98.1728	29.7433	618	605.1	-12.9	602.0	-16.0
6823302	Comal	98.1389	29.7161	614	606.1	-7.9	601.3	-12.7
6823306	Comal	98.1528	29.7464	612	605.0	-7.0	600.3	-11.7
6823307	Comal	98.1594	29.7461	616	605.0	-11.0	600.8	-15.2
6823308	Comal	98.1453	29.7475	613	604.9	-8.1	599.8	-13.2
6823310	Comal	98.1253	29.7306	603	605.0	2.0	599.9	-3.1
6823507	Comal	98.1989	29.6889	620	613.4	-6.6	603.9	-16.1
6823701	Comal	98.2164	29.6486	615	611.2	-3.8	604.6	-10.4
6823809	Comal	98.1928	29.6594	588	604.6	16.6	604.4	16.4
6824102	Comal	98.1069	29.7444	610	604.3	-5.7	598.2	-11.8
6830208	Comal	98.3194	29.6097	615	615.3	.3	605.0	-10.0
6830217	Comal	98.3006	29.6181	614	615.1	1.1	605.0	-9.0
6830313	Comal	98.2572	29.6167	576	605.7	29.7	604.6	28.6
5857301	Hays	97.8903	30.0936	596	551.6	-44.4	559.3	-36.7
5857902	Hays	97.8958	30.0083	574	582.1	8.1	576.6	2.6
5857903	Hays	97.8861	30.0381	561	575.9	14.9	564.1	3.1
5858101	Hays	97.8422	30.0836	560	548.3	-11.7	541.7	-18.3
6701307	Hays	97.8869	29.9675	565	584.8	19.8	583.8	18.8
6701701	Hays	97.9639	29.8956	568	586.2	18.2	586.1	18.1
6701808	Hays	97.9194	29.9014	582	585.0	3.0	585.0	3.0
6702103	Hays	97.8725	29.9889	519	584.6	65.6	577.1	58.1
6816301	Hays	98.0214	29.8714	601	589.2	-11.8	603.6	2.6
Statistics:								
Mean absolute difference						31.6		20.5
Mean algebraic difference						-7.6		-8.1
RMS error						58.7		33.4

Table 5. Transient simulation target wells and residuals for above-normal rainfall conditions, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas.

[Measured and simulated water levels in altitude above National Geodetic Vertical Datum of 1929. TWDB, Texas Water Development Board; ID, identification number; DD.MMSS, degrees.minutes/seconds; residual, simulated water level minus measured water level; mean absolute difference, sum of absolute values of residuals divided by number of wells; mean algebraic difference, algebraic sum of residuals divided by number of wells; RMS error, root-mean-square error]

TWDB well ID	County	Latitude (DD.MMSS)	Longitude (DD.MMSS)	Measured water level (feet)	Conduit-flow Edwards aquifer model (Lindgren, 2004)		Diffuse-flow Edwards aquifer model	
					Simulated water level (feet)	Residual (feet)	Simulated water level (feet)	Residual (feet)
7038501	Kinney	100.3083	29.4328	1,240	1,179.8	-60.2	1,169.7	-70.3
7038902	Kinney	100.2617	29.4131	1,183	1,191.1	8.1	1,193.6	10.6
7045505	Kinney	100.4219	29.3106	1,117	1,105.2	-11.8	1,103.3	-13.7
7045603	Kinney	100.4125	29.3278	1,093	1,105.7	12.7	1,103.5	10.5
7046101	Kinney	100.3428	29.3436	1,120	1,100.2	-19.8	1,105.8	-14.2
7046201	Kinney	100.3042	29.3453	1,112	1,094.0	-18.0	1,105.1	-6.9
7046302	Kinney	100.2622	29.3367	1,097	1,071.9	-25.1	1,092.3	-4.7
7046901	Kinney	100.2667	29.2703	1,046	1,043.9	-2.1	1,048.6	2.6
7047501	Kinney	100.1711	29.3206	1,017	991.4	-25.6	1,057.1	40.1
6933901	Uvalde	99.8783	29.3936	1,126	960.7	-165.3	1,022.7	-103.3
6935804	Uvalde	99.7017	29.3950	1,042	970.2	-71.8	1,094.7	52.7
6936601	Uvalde	99.5158	29.4256	829	870.2	41.2	876.5	47.5
6937402	Uvalde	99.4700	29.4411	860	836.9	-23.1	855.2	-4.8
6941101	Uvalde	99.9606	29.3442	1,012	961.8	-50.2	1,028.1	16.1
6941502	Uvalde	99.9483	29.3183	983	937.2	-45.8	991.0	8.0
6941504	Uvalde	99.9336	29.2967	934	918.3	-15.7	971.8	37.8
6941701	Uvalde	99.9711	29.2603	938	904.0	-34.0	933.9	-4.1
6941901	Uvalde	99.8803	29.2850	906	893.1	-12.9	907.9	1.9
6942601	Uvalde	99.7508	29.3150	879	888.3	9.3	878.7	-.3
6942709	Uvalde	99.8625	29.2731	904	892.0	-12.0	902.4	-1.6
6942901	Uvalde	99.7589	29.2544	867	887.0	20.0	876.9	9.9
6942907	Uvalde	99.7533	29.2828	881	887.3	6.3	876.2	-4.8
6943103	Uvalde	99.7100	29.3478	924	899.0	-25.0	892.2	-31.8
6943106	Uvalde	99.7389	29.3550	962	893.5	-68.5	885.4	-76.6
6943202	Uvalde	99.6994	29.3592	948	916.3	-31.7	913.9	-34.1
6943301	Uvalde	99.6503	29.3650	866	904.7	38.7	896.3	30.3
6943603	Uvalde	99.6364	29.2956	811	871.0	60.0	826.1	15.1
6943804	Uvalde	99.6917	29.2767	885	882.9	-2.1	856.1	-28.9
6943902	Uvalde	99.6333	29.2761	886	869.7	-16.3	826.6	-59.4
6943903	Uvalde	99.6508	29.2694	884	877.3	-6.7	829.7	-54.3
6943910	Uvalde	99.6400	29.2503	876	839.4	-36.6	821.9	-54.1
6944301	Uvalde	99.5117	29.3494	808	811.9	3.9	815.9	7.9
6944402	Uvalde	99.5933	29.3297	810	872.3	62.3	825.3	15.3
6944701	Uvalde	99.5942	29.2697	808	827.6	19.6	814.5	6.5
6944703	Uvalde	99.5950	29.2586	810	806.9	-3.1	809.0	-1.0
6944704	Uvalde	99.6108	29.2547	797	821.1	24.1	814.3	17.3
6944804	Uvalde	99.5733	29.2903	813	834.9	21.9	814.3	1.3
6945401	Uvalde	99.4681	29.3192	806	798.5	-7.5	801.4	-4.6
6949302	Uvalde	99.8803	29.2442	902	891.7	-10.3	906.3	4.3
6950101	Uvalde	99.8506	29.2406	900	889.7	-10.3	898.0	-2.0
6950202	Uvalde	99.7981	29.2372	887	887.4	.4	882.6	-4.4
6950204	Uvalde	99.8247	29.2203	887	887.7	.7	885.5	-1.5
6950302	Uvalde	99.7867	29.2103	881	886.2	5.2	880.4	-.6
6950306	Uvalde	99.7522	29.2322	884	886.3	2.3	877.0	-7.0

Table 5. Transient simulation target wells and residuals for above-normal rainfall conditions, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas—Continued.

TWDB well ID	County	Latitude (DD.MMSS)	Longitude (DD.MMSS)	Measured water level (feet)	Conduit-flow Edwards aquifer model (Lindgren, 2004)		Diffuse-flow Edwards aquifer model	
					Simulated water level (feet)	Residual (feet)	Simulated water level (feet)	Residual (feet)
6950403	Uvalde	99.8367	29.1908	875	886.6	11.6	888.2	13.2
6950408	Uvalde	99.8603	29.2042	886	888.2	2.2	898.0	12.0
6951104	Uvalde	99.7467	29.2358	883	886.3	3.3	875.3	-7.7
6951202	Uvalde	99.6719	29.2331	843	858.6	15.6	830.1	-12.9
6951401	Uvalde	99.7347	29.1786	858	880.3	22.3	876.1	18.1
6951602	Uvalde	99.6600	29.1756	826	837.2	11.2	815.9	-10.1
6952201	Uvalde	99.5781	29.2189	803	795.0	-8.0	800.2	-2.8
6952402	Uvalde	99.6236	29.1986	723	809.5	86.5	806.6	83.6
7040901	Uvalde	100.0075	29.3944	1,078	1,006.0	-72.0	1,065.6	-12.4
6833601	Medina	98.8903	29.4350	738	761.9	23.9	734.0	-4.0
6833604	Medina	98.8881	29.4219	732	756.6	24.6	731.7	-.3
6834506	Medina	98.8128	29.4314	720	758.2	38.2	735.0	15.0
6841301	Medina	98.8797	29.3544	722	742.8	20.8	724.3	2.3
6841305	Medina	98.8897	29.3672	722	746.2	24.2	725.4	3.4
6842106	Medina	98.8536	29.3547	725	741.2	16.2	719.7	-5.3
6842504	Medina	98.8147	29.3094	716	735.2	19.2	718.4	2.4
6849813	Medina	98.9392	29.1650	717	739.0	22.0	737.9	20.9
6937301	Medina	99.3836	29.4664	884	841.4	-42.6	858.7	-25.3
6938601	Medina	99.2831	29.4383	901	850.3	-50.7	889.8	-11.2
6938901	Medina	99.2653	29.4147	858	839.6	-18.4	850.7	-7.3
6938902	Medina	99.2744	29.3756	799	801.0	2.0	788.1	-10.9
6939503	Medina	99.1928	29.4358	869	858.6	-10.4	830.7	-38.3
6939504	Medina	99.1942	29.4503	893	870.0	-23.0	882.1	-10.9
6939505	Medina	99.2025	29.4492	900	872.9	-27.1	899.5	-.5
6940101	Medina	99.1228	29.4906	906	926.5	20.5	948.7	42.7
6940901	Medina	99.0008	29.4094	748	759.6	11.6	734.2	-13.8
6946601	Medina	99.2786	29.3319	784	780.8	-3.2	784.3	.3
6946701	Medina	99.3525	29.2639	793	782.7	-10.3	789.0	-4.0
6946901	Medina	99.2831	29.2836	791	778.2	-12.8	785.1	-5.9
6947204	Medina	99.2067	29.3606	785	785.6	.6	778.2	-6.8
6947302	Medina	99.1639	29.3689	754	773.7	19.7	748.8	-5.2
6947604	Medina	99.1389	29.2969	762	765.0	3.0	748.2	-13.8
6947701	Medina	99.2489	29.2561	797	776.4	-20.6	783.1	-13.9
6948102	Medina	99.0864	29.3508	754	763.8	9.8	736.4	-17.6
6955501	Medina	99.1989	29.1764	783	756.1	-26.9	763.5	-19.5
6956501	Medina	99.0517	29.1736	755	745.8	-9.2	747.0	-8.0
6827505	Bexar	98.6797	29.5600	781	817.7	36.7	769.6	-11.4
6827512	Bexar	98.6831	29.5625	822	837.8	15.8	775.0	-47.0
6828404	Bexar	98.6039	29.5417	721	696.1	-24.9	756.0	35.0
6828507	Bexar	98.5514	29.5422	655	749.4	94.4	699.0	44.0
6828705	Bexar	98.6247	29.5014	707	710.5	3.5	699.2	-7.8
6828901	Bexar	98.5308	29.5389	692	734.0	42.0	685.3	-6.7
6828909	Bexar	98.5261	29.5072	784	712.7	-71.3	683.7	-100.3
6828910	Bexar	98.5381	29.5356	699	738.9	39.9	686.5	-12.5
6829103	Bexar	98.4869	29.5894	709	809.0	100.0	743.3	34.3
6829304	Bexar	98.4106	29.5975	657	716.1	59.1	681.7	24.7
6829506	Bexar	98.4411	29.5736	681	708.1	27.1	678.6	-2.4
6829605	Bexar	98.3872	29.5500	685	679.5	-5.5	674.6	-10.4

Table 5. Transient simulation target wells and residuals for above-normal rainfall conditions, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas—Continued.

TWDB well ID	County	Latitude (DD.MMSS)	Longitude (DD.MMSS)	Measured water level (feet)	Conduit-flow Edwards aquifer model (Lindgren, 2004)		Diffuse-flow Edwards aquifer model	
					Simulated water level (feet)	Residual (feet)	Simulated water level (feet)	Residual (feet)
6829701	Bexar	98.4636	29.5369	690	702.0	12.0	679.2	-10.8
6829811	Bexar	98.4475	29.5136	692	684.2	-7.8	675.5	-16.5
6829913	Bexar	98.3792	29.5306	684	678.6	-5.4	674.5	-9.5
6829914	Bexar	98.4036	29.5075	684	685.0	1.0	675.1	-8.9
6830101	Bexar	98.3553	29.5919	682	682.0	0	672.4	-9.6
6830211	Bexar	98.3278	29.6050	676	692.8	16.8	670.6	-5.4
6830220	Bexar	98.3153	29.6008	660	663.2	3.2	669.1	9.1
6830514	Bexar	98.3194	29.5789	682	663.1	-18.9	670.2	-11.8
6830705	Bexar	98.3667	29.5403	692	677.0	-15.0	674.1	-17.9
6830802	Bexar	98.3139	29.5175	682	670.8	-11.2	673.2	-8.8
6830807	Bexar	98.2956	29.5247	682	667.1	-14.9	672.6	-9.4
6834602	Bexar	98.7900	29.4386	718	756.5	38.5	733.5	15.5
6834603	Bexar	98.7586	29.4169	706	730.6	24.6	708.5	2.5
6835202	Bexar	98.6906	29.4889	706	738.8	32.8	717.0	11.0
6835312	Bexar	98.6558	29.4717	700	727.6	27.6	703.1	3.1
6835807	Bexar	98.7075	29.3778	706	721.0	15.0	699.8	-6.2
6836104	Bexar	98.6006	29.4906	680	709.3	29.3	690.7	10.7
6836105	Bexar	98.6014	29.4875	698	710.0	12.0	690.2	-7.8
6836407	Bexar	98.6239	29.4261	703	714.7	11.7	692.3	-10.7
6837103	Bexar	98.4869	29.4972	681	705.2	24.2	681.3	.3
6837203	Bexar	98.4322	29.4708	687	692.4	5.4	677.8	-9.2
6837409	Bexar	98.4956	29.4183	688	700.1	12.1	681.3	-6.7
6837511	Bexar	98.4292	29.4389	666	695.4	29.4	678.6	12.6
6837606	Bexar	98.4108	29.4456	686	693.7	7.7	677.8	-8.2
6837707	Bexar	98.4928	29.4006	686	703.1	17.1	681.5	-4.5
6843611	Bexar	98.6636	29.3264	691	715.4	24.4	690.1	-.9
6843807	Bexar	98.6719	29.2750	699	717.5	18.5	695.5	-3.5
6844109	Bexar	98.5858	29.3378	662	709.0	47.0	685.1	23.1
6844214	Bexar	98.5481	29.3558	689	705.8	16.8	683.8	-5.2
6845102	Bexar	98.4972	29.3742	677	703.9	26.9	682.2	5.2
6845301	Bexar	98.3994	29.3711	664	695.8	31.8	679.5	15.5
6845901	Bexar	98.3817	29.2564	650	694.3	44.3	678.6	28.6
6815902	Comal	98.1397	29.7592	645	653.2	8.2	639.3	-5.7
6815903	Comal	98.1414	29.7575	641	652.9	11.9	640.5	-.5
6816501	Comal	98.0433	29.7972	597	596.6	-.4	604.6	7.6
6816602	Comal	98.0239	29.7936	601	597.3	-3.7	604.6	3.6
6816701	Comal	98.1044	29.7572	612	637.8	25.8	623.2	11.2
6816703	Comal	98.0928	29.7531	606	615.5	9.5	618.6	12.6
6816801	Comal	98.0528	29.7861	604	597.8	-6.2	606.9	2.9
6816803	Comal	98.0736	29.7697	616	598.3	-17.7	610.7	-5.3
6822301	Comal	98.2578	29.7119	681	763.0	82.0	689.4	8.4
6822501	Comal	98.2931	29.6800	684	755.6	71.6	678.0	-6.0
6822903	Comal	98.2597	29.6267	647	647.6	.6	659.5	12.5
6823101	Comal	98.2106	29.7403	682	684.3	2.3	658.2	-23.8
6823202	Comal	98.2006	29.7111	652	672.1	20.1	657.4	5.4
6823206	Comal	98.1819	29.7483	655	657.2	2.2	649.3	-5.7
6823220	Comal	98.2081	29.7419	682	672.9	-9.1	657.1	-24.9
6823302	Comal	98.1389	29.7161	628	662.3	34.3	648.8	20.8

Table 5. Transient simulation target wells and residuals for above-normal rainfall conditions, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas—Continued.

TWDB well ID	County	Latitude (DD.MMSS)	Longitude (DD.MMSS)	Measured water level (feet)	Conduit-flow Edwards aquifer model (Lindgren, 2004)		Diffuse-flow Edwards aquifer model	
					Simulated water level (feet)	Residual (feet)	Simulated water level (feet)	Residual (feet)
6823306	Comal	98.1528	29.7464	650	652.9	2.9	644.1	-5.9
6823701	Comal	98.2164	29.6486	660	638.0	-22.0	650.5	-9.5
6823706	Comal	98.2289	29.6317	656	643.7	-12.3	653.7	-2.3
6823807	Comal	98.1714	29.6658	653	629.1	-23.9	640.0	-13.0
6823808	Comal	98.1819	29.6572	648	632.5	-15.5	643.8	-4.2
6824102	Comal	98.1069	29.7444	609	654.0	45.0	637.2	28.2
6824104	Comal	98.1017	29.7472	616	654.0	38.0	636.9	20.9
6824105	Comal	98.0883	29.7383	636	617.2	-18.8	619.3	-16.7
6830208	Comal	98.3194	29.6097	674	695.9	21.9	670.4	-3.6
6830312	Comal	98.2822	29.6128	657	652.9	-4.1	664.7	7.7
5849911	Hays	97.8894	30.1328	629	594.6	-34.4	671.7	42.7
5857902	Hays	97.8958	30.0083	614	606.3	-7.7	648.1	34.1
5857903	Hays	97.8861	30.0381	587	619.7	32.7	662.5	75.5
5858101	Hays	97.8422	30.0836	637	610.5	-26.5	635.0	-2.0
5858104	Hays	97.8486	30.1042	623	595.9	-27.1	608.9	-14.1
5858406	Hays	97.8558	30.0614	624	620.0	-4.0	653.8	29.8
6701203	Hays	97.9206	29.9619	598	602.7	4.7	620.5	22.5
6701304	Hays	97.8761	29.9844	586	593.5	7.5	631.3	45.3
6701305	Hays	97.8872	29.9675	578	592.9	14.9	620.0	42.0
6701401	Hays	97.9642	29.9500	600	628.6	28.6	631.9	31.9
6701501	Hays	97.9475	29.9236	577	617.3	40.3	607.7	30.7
6701701	Hays	97.9639	29.8956	579	605.6	26.6	596.9	17.9
6701809	Hays	97.9286	29.9119	587	598.5	11.5	594.3	7.3
6702103	Hays	97.8725	29.9889	598	593.6	-4.4	634.5	36.5
6709102	Hays	97.9758	29.8508	580	590.2	10.2	593.0	13.0
6709110	Hays	97.9819	29.8431	590	590.4	.4	593.7	3.7
6808601	Hays	98.0278	29.9458	810	680.0	-130.0	772.2	-37.8
6816301	Hays	98.0214	29.8714	606	640.6	34.6	689.6	83.6
6816605	Hays	98.0042	29.8289	588	590.9	2.9	596.6	8.6
Statistics:								
Mean absolute difference						23.5		17.2
Mean algebraic difference						3.5		.5
RMS error						33.5		25.8

Table 6. Comparison of the residuals for hydraulic heads and springflows for selected observation wells and springs, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas.

[TWDB, Texas Water Development Board; ID, identification number; residual, simulated water level minus measured water level; MAE (mean absolute difference), sum of absolute values of residuals divided by number of wells; ME (mean algebraic difference), algebraic sum of residuals divided by number of wells; RMS, root-mean-square error; C, confined conditions at well site; U, unconfined conditions at well site]

TWDB well ID	Period of measurements	County	Hydrologic condition	Hydraulic-head residuals (feet)					
				Conduit-flow Edwards aquifer model (Lindgren and others, 2004)			Diffuse-flow Edwards aquifer model		
				MAE	ME	RMS	MAE	ME	RMS
6945401	1954–2000	Uvalde	C	11.4	.4	14.2	17.2	12.7	19.9
6950302	1947–2000	Uvalde	C	11.7	1.7	15.0	11.1	-2.3	13.8
6841301	1950–2000	Medina	C	15.9	14.9	17.8	8.5	5.4	10.6
6946701	1947–90	Medina	C	10.6	-1.4	13.5	15.1	11.6	17.3
6837203	1947–2000	Bexar	C	6.2	1.5	8.0	6.7	-4.4	8.7
6830211	1964–2000	Bexar	C/U	12.8	4.8	15.4	5.5	-2.7	7.4
6845102	1950–84	Bexar	C	22.3	22.3	23.4	11.1	10.7	12.3
6816801	1947–2000	Comal	U	3.4	-.6	5.0	7.0	6.3	7.7
6823701	1947–94	Comal	C	6.7	-1.1	8.7	4.4	-2.1	6.3
5857903	1949–85	Hays	C/U	8.0	2.1	11.0	18.9	11.8	26.0
5858101	1947–2000	Hays	C	26.2	-18.6	29.2	21.0	-15.4	25.9
6702103	1947–77	Hays	C	11.1	-8.9	15.3	27.1	27.1	30.4

Spring name	Period of measurements	County	Springflow residuals (cubic feet per day)					
			Conduit-flow Edwards aquifer model (Lindgren and others, 2004)			Diffuse-flow Edwards aquifer model		
			MAE	ME	RMS	MAE	ME	RMS
Comal	1947–2000	Comal	3,083,452	577,905	3,967,067	2,654,312	-127,334	3,348,776
San Marcos	1947–2000	Hays	2,145,232	1,296,041	2,691,442	2,209,134	543,298	2,780,848
Leona	1947–2000	Uvalde	2,025,127	1,881,722	2,884,793	1,697,922	1,452,740	2,271,649
San Antonio	1947–2000	Bexar	830,745	-418,464	1,914,358	1,573,585	1,120,551	3,448,867
San Pedro	1947–2000	Bexar	153,317	4,359	230,676	207,910	-199,293	348,280

Table 7. Simulated annual water budget for the steady-state simulation and for 1956 for the transient simulation, conduit-flow and diffuse-flow Edwards aquifer models, San Antonio region, Texas.

[Recharge includes leakage from streams through streambeds and infiltration of precipitation in interstream areas. Boundary inflow includes inflow through general-head and specified-flow boundary condition cells at northern and northwestern model boundaries. Stream-aquifer leakage is between Edwards aquifer and Colorado River at northeastern model boundary. Subtotal comprises source or discharge components exclusive of changes in storage. Total includes changes in storage. Negative net change in storage indicates net loss of water from storage (storage accounted for as source). NA, not applicable]

Budget component and time period	Conduit-flow Edwards aquifer model (Lindgren and others, 2004)			Diffuse-flow Edwards aquifer model			Difference
	Flow rate (acre-feet/ year)	Percentage of total sources or discharges	Percentage of subtotal for sources or discharges	Flow rate (acre-feet/ year)	Percentage of total sources or discharges	Percentage of subtotal for sources or discharges	Percentage of total sources or discharges
Sources							
Recharge							
Steady-state	583,985.9	93.5	NA	585,154.8	93.6	NA	0.1
1956	60,459.5	14.5	60.4	61,111.7	15.0	60.1	.5
Boundary inflow							
Steady-state	40,265.3	6.5	NA	40,265.3	6.4	NA	-.1
1956	39,457.3	9.5	39.4	40,354.3	9.9	39.7	.4
Stream-aquifer leakage							
Steady-state	0	0	NA	0	0	NA	0
1956	179.0	.04	.2	204.2	.05	.2	.01
Subtotal							
1956	100,095.8	24.1	100.0	101,670.2	25.0	100.0	NA
Total sources							
Steady-state	624,251.2	100.0	NA	625,420.1	100.0	NA	NA
1956	416,036.3	100.0	NA	407,113.1	100.0	NA	NA
Discharges							
Withdrawals (pumpage)							
Steady-state	160,592.5	25.7	NA	160,593.2	25.7	NA	0
1956	309,177.7	74.3	74.3	309,927.4	76.1	76.1	1.8
Springflow							
Steady-state	460,169.3	73.7	NA	459,363.9	73.4	NA	-.3
1956	104,983.6	25.2	25.2	94,251.5	23.2	23.2	-2.0
Stream-aquifer leakage							
Steady-state	3,488.7	.6	NA	5,462.3	.9	NA	.3
1956	1,871.4	.4	.4	2,933.0	.7	.7	.3
Subtotal							
1956	416,032.7	100.0	100.0	407,111.8	100.0	100.0	NA
Total discharges							
Steady-state	624,250.5	100.0	NA	625,419.5	100.0	NA	NA
1956	416,032.7	100.0	100.0	407,111.8	100.0	100.0	NA
Net change in storage							
1956	-315,940.5	75.9	NA	-305,442.9	75.0	NA	-.9

Information regarding water resources in Texas is available at
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