# ATTACHMENT E
# USGS SIR 2013-5042 Groundwater Model Report

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Prepared in cooperation with the city of Wichita, Kansas,
as part of the *Equus* Beds Groundwater Recharge Project

# Simulation of Groundwater Flow, Effects of Artificial Recharge, and Storage Volume Changes in the *Equus* Beds Aquifer near the City of Wichita, Kansas Well Field, 1935–2008

Scientific Investigations Report 2013–5042

U.S. Department of the Interior
U.S. Geological Survey

**Cover.** Aerial photograph of city of Wichita Aquifer Storage and Recovery phase II facility (photograph by Tom Mason, city of Wichita, February 1, 2012).

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**U.S. Department of the Interior**
**U.S. Geological Survey**

**U.S. Department of the Interior**
SALLY JEWELL, Secretary

**U.S. Geological Survey**
Suzette M. Kimball, Acting Director

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# Conversion Factors and Datums

Inch/Pound to SI

| Multiply | By | To obtain |
|---|---|---|
| | Length | |
| foot (ft) | 0.3048 | meter (m) |
| mile (mi) | 1.609 | kilometer (km) |
| | Area | |
| acre | 0.004047 | square kilometer ($km^2$) |
| | Volume | |
| gallon (gal) | 0.003785 | cubic meter ($m^3$) |
| acre-foot (acre-ft) | 1,233 | cubic meter ($m^3$) |
| cubic feet ($ft^3$) | .02832 | cubic meter ($m^3$) |
| | Flow rate | |
| foot per day (ft/d) | .3048 | meter per day (m/d) |
| acre-foot per day (acre-ft/d) | 1,233 | cubic meter per day ($m^3$/d) |
| cubic foot per day ($ft^3$/d) | .02832 | cubic meter per day ($m^3$/d) |
| gallon per day (gal/d) | .003785 | cubic meter per day ($m^3$/d) |
| | Hydraulic conductivity | |
| foot per day (ft/d) | .3048 | meter per day (m/d) |
| | Hydraulic gradient | |
| foot per mile (ft/mi) | .1894 | meter per kilometer (m/km) |
| | Transmissivity* | |
| foot squared per day ($ft^2$/d) | .09294 | meter squared per day ($m^2$/d) |
| | Leakance | |
| foot per day per foot [(ft/d)/ft] | 1 | meter per day per meter [(m/d)/m] |

Temperature in degrees Celsius (°C) may be converted to degrees Fahrenheit (°F) as follows:

°F=(1.8×°C)+32

Temperature in degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) as follows:

°C=(°F-32)/1.8

Vertical coordinate information is referenced to the North American Vertical Datum of 1988 (NAVD 88)

Horizontal coordinate information is referenced to the North American Datum of 1983 (NAD 83)

Altitude, as used in this report, refers to distance above the vertical datum.

*Transmissivity: The standard unit for transmissivity is cubic foot per day per square foot times foot of aquifer thickness [($ft^3$/d)/$ft^2$]ft. In this report, the mathematically reduced form, foot squared per day ($ft^2$/d), is used for convenience.

# Acknowledgments

The study presented in this report was prepared in cooperation with the city of Wichita as part of the *Equus* Beds Aquifer Storage and Recovery project. The authors thank personnel from the city of Wichita and *Equus* Bed Groundwater Management District No. 2 for providing data in support of the model. We thank Washburn University for the use of the HiPACE computer cluster and the University of Nebraska at Lincoln for the use of resources at the Holland Computer Center.

We also thank Brian Klager of the U.S. Geological Survey Kansas Water Science Center for preparing input files, figures, and tables, and assisting with analysis of output data.

# Simulation of Groundwater Flow, Effects of Artificial Recharge, and Storage Volume Changes in the *Equus* Beds Aquifer near the City of Wichita, Kansas Well Field, 1935–2008

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## Abstract

The *Equus* Beds aquifer is a primary water-supply source for Wichita, Kansas and the surrounding area because of shallow depth to water, large saturated thickness, and generally good water quality. Substantial water-level declines in the *Equus* Beds aquifer have resulted from pumping groundwater for agricultural and municipal needs, as well as periodic drought conditions. In March 2006, the city of Wichita began construction of the *Equus* Beds Aquifer Storage and Recovery project to store and later recover groundwater, and to form a hydraulic barrier to the known chloride-brine plume near Burrton, Kansas. In October 2009, the U.S. Geological Survey, in cooperation with the city of Wichita, began a study to determine groundwater flow in the area of the Wichita well field, and chloride transport from the Arkansas River and Burrton oilfield to the Wichita well field.

Groundwater flow was simulated for the *Equus* Beds aquifer using the three-dimensional finite-difference groundwater-flow model MODFLOW-2000. The model simulates steady-state and transient conditions. The groundwater-flow model was calibrated by adjusting model input data and model geometry until model results matched field observations within an acceptable level of accuracy. The root mean square (RMS) error for water-level observations for the steady-state calibration simulation is 9.82 feet. The ratio of the RMS error to the total head loss in the model area is 0.049 and the mean error for water-level observations is 3.86 feet. The difference between flow into the model and flow out of the model across all model boundaries is -0.08 percent of total flow for the steady-state calibration. The RMS error for water-level observations for the transient calibration simulation is 2.48 feet, the ratio of the RMS error to the total head loss in the model area is 0.0124, and the mean error for water-level observations is 0.03 feet. The RMS error calculated for observed and simulated base flow gains or losses for the Arkansas River for the transient simulation is 7,916,564 cubic feet per day (91.6 cubic feet per second) and the RMS error divided by (/) the total range in streamflow (7,916,564/37,461,669 cubic feet per day) is 22 percent. The RMS error calculated for observed and simulated streamflow gains or losses for the Little Arkansas River for the transient simulation is 5,610,089 cubic feet per day(64.9 cubic feet per second) and the RMS error divided by the total range in streamflow (5,612,918/41,791,091 cubic feet per day) is 13 percent. The mean error between observed and simulated base flow gains or losses was 29,999 cubic feet per day (0.34 cubic feet per second) for the Arkansas River and -1,369,250 cubic feet per day (-15.8 cubic feet per second) for the Little Arkansas River. Cumulative streamflow gain and loss observations are similar to the cumulative simulated equivalents. Average percent mass balance difference for individual stress periods ranged from -0.46 to 0.51 percent. The cumulative mass balance for the transient calibration was 0.01 percent.

Composite scaled sensitivities indicate the simulations are most sensitive to parameters with a large areal distribution. For the steady-state calibration, these parameters include recharge, hydraulic conductivity, and vertical conductance. For the transient simulation, these parameters include evapotranspiration, recharge, and hydraulic conductivity.

The ability of the calibrated model to account for the additional groundwater recharged to the *Equus* Beds aquifer as part of the Aquifer Storage and Recovery project was assessed by using the U.S. Geological Survey subregional water budget program ZONEBUDGET and comparing those results to metered recharge for 2007 and 2008 and previous estimates of artificial recharge. The change in storage between simulations is the volume of water that estimates the recharge credit for the aquifer storage and recovery system.

The estimated increase in storage of 1,607 acre-ft in the basin storage area compared to metered recharge of 1,796 acre-ft indicates some loss of metered recharge. Increased storage outside of the basin storage area of 183 acre-ft accounts for all but 6 acre-ft or 0.33 percent of the total. Previously estimated recharge credits for 2007 and 2008 are 1,018 and 600 acre-ft, respectively, and a total

estimated recharge credit of 1,618 acre-ft. Storage changes calculated for this study are 4.42 percent less for 2007 and 5.67 percent more for 2008 than previous estimates. Total storage change for 2007 and 2008 is 0.68 percent less than previous estimates. The small difference between the increase in storage from artificial recharge estimated with the groundwater-flow model and metered recharge indicates the groundwater model correctly accounts for the additional water recharged to the *Equus* Beds aquifer as part of the Aquifer Storage and Recovery project. Small percent differences between inflows and outflows for all stress periods and all index cells in the basin storage area, improved calibration compared to the previous model, and a reasonable match between simulated and measured long-term base flow indicates the groundwater model accurately simulates groundwater flow in the study area.

The change in groundwater level through recent years compared to the August 1940 groundwater level map has been documented and used to assess the change of storage volume of the *Equus* Beds aquifer in and near the Wichita well field for three different areas. Two methods were used to estimate changes in storage from simulation results using simulated change in groundwater levels in layer 1 between stress periods, and using ZONEBUDGET to calculate the change in storage in the same way the effects of artificial recharge were estimated within the basin storage area. The three methods indicate similar trends although the magnitude of storage changes differ.

Information about the change in storage in response to hydrologic stresses is important for managing groundwater resources in the study area. The comparison between the three methods indicates similar storage change trends are estimated and each could be used to determine relative increases or decreases in storage. Use of groundwater level changes that do not include storage changes that occur in confined or semi-confined parts of the aquifer will slightly underestimate storage changes; however, use of specific yield and groundwater level changes to estimate storage change in confined or semi-confined parts of the aquifer will overestimate storage changes. Using only changes in shallow groundwater levels would provide more accurate storage change estimates for the measured groundwater levels method.

The value used for specific yield is also an important consideration when estimating storage. For the *Equus* Beds aquifer the reported specific yield ranges between 0.08 and 0.35 and the storage coefficient (for confined conditions) ranges between 0.0004 and 0.16. Considering the importance of the value of specific yield and storage coefficient to estimates of storage change over time, and the wide range and substantial overlap for the reported values for specific yield and storage coefficient in the study area, further information on the distribution of specific yield and storage coefficient within the *Equus* Beds aquifer in the study area would greatly enhance the accuracy of estimated storage changes using both simulated groundwater level, simulated groundwater budget, or measured groundwater level methods.

# Introduction

In October 2009, the U.S. Geological Survey (USGS), in cooperation with the city of Wichita, began a study to determine groundwater flow in the area of the Wichita well field and chloride transport from the Arkansas River and Burrton oilfield to the Wichita well field. The primary study area includes the *Equus* Beds aquifer near the city of Wichita supply wells and encompasses the Aquifer Storage and Recovery (ASR) project phase I, II, and III artificial recharge areas. The groundwater model used to calculate recharge credits for the ASR project from 2007 through 2010 was a modified version of the model developed by the USGS (Myers and others, 1996) to determine groundwater flow and chloride transport from the Arkansas River and Burrton oilfield. The approach in the study presented in this report was to update and recalibrate the existing Myers and others (1996) model with new hydrologic data; however, in 2010 the USGS and a peer review panel selected by the city of Wichita determined that the boundary conditions of the original USGS model (Myers and others, 1996), although appropriate for the original purpose, were not adequate for chloride transport and recharge credit accounting near Wichita's well field. Data from existing groundwater-flow models (Spinazola and others, 1985; Myers and others, 1996; Pruitt, 1993) and 2008 lithologic and hydrologic data were used to construct a new model to simulate groundwater flow in the *Equus* Beds aquifer. Thus, the development and calibration of a new groundwater-flow model to simulate groundwater flow and artificial recharge credit accounting with appropriate boundary conditions was deemed necessary and is the focus of this report. Changes made to the new model include model edges that are located far from the area of interest near the Wichita well field and the Burrton oilfield to reduce their effect on simulated groundwater flow, smaller model cell sizes to reduce potential errors for simulation of chloride transport, areally distributed recharge based on data from multiple weather stations and relative permeability of soils, and more accurate location and representation of pumping wells. Simulation of chloride transport from the Arkansas River and Burrton oilfield to the Wichita well field using the new groundwater-flow model is not described in this report.

The easternmost extension of the High Plains aquifer system (National aquifer code N100HGHPLN) consists of alluvial deposits of sand and gravel interbedded with clay or silt (Williams and Lohman, 1949), and is referred to in this report as the *Equus* Beds aquifer for the Pleistocene-age horse fossils present in the aquifer sediments. The *Equus* Beds aquifer is unconfined in the study area (Spinazola and others, 1985), and is a substantial water-supply source for Wichita, Kansas (Kans.) and the surrounding area because of shallow depth to water [as shallow as 10 feet (ft) near the Arkansas River], a saturated thickness of as much as 250 ft (Myers and others, 1996; Hansen and Aucott, 2004), and generally good water quality (Ziegler and others, 1999). The general direction of groundwater movement within the study area is to the east (Aucott and others, 1998). Numerous irrigation wells also withdraw water from the aquifer within the boundaries of *Equus* Beds Groundwater Management

District No. 2 (GMD2) (*Equus* Beds Groundwater Management District No. 2, 1995) (fig. 1).

The well field was developed by the city of Wichita in the *Equus* Beds aquifer during the 1940s and 1950s (fig. 1). As of 2008, there were 55 active city production wells in the Wichita *Equus* Beds well field and about 40 percent of the water-supply needs for the city of Wichita came from the *Equus* Beds aquifer (Debra Ary, city of Wichita, oral commun., 2010). The city of Wichita began using water from Cheney Reservoir in 1965 to supplement its supply from the *Equus* Beds aquifer. The proportion of the water supply obtained from Cheney Reservoir increased from 20 percent in 1965 to 44 percent in 1994. From 1995 through 2010, water from Cheney Reservoir ranged from 51 to 69 percent of Wichita's water supply (Ziegler and others, 2010).

The increased reliance on surface water from Cheney Reservoir was part of Wichita's Intergrated Local Water Supply Plan, implemented in 1993 (Wichita, 2011), and described as the Integrated Resource Plan in Warren and others (1995). This plan was initiated to ensure the city's water-supply needs are met through 2050 by promoting conservation, increasing water use from Cheney Reservoir, and decreasing pumping from city wells in the Wichita *Equus* Beds well field.

From 1900 to 1940, water from the *Equus* Beds aquifer was withdrawn for municipal and industrial use near Hutchinson and Wichita; however, no cone of depression in the water table map for 1940 is apparent, which indicates storage had not been greatly affected by well pumping before 1940 (Spinazola, and others, 1985; Williams and Lohman, 1949). For this study, the water table of 1940 is assumed to represent predevelopment conditions.

Substantial water-level declines in the *Equus* Beds aquifer have resulted from pumping groundwater for agricultural and municipal needs, as well as periodic drought conditions since 1940. The lowest water levels to date were recorded in October 1992 and were as much as 50 ft lower than the predevelopment (1940) water levels in some locations (Hansen and Aucott, 2001, 2004; Hansen, 2007). Water-level declines caused concern about the adequacy of the city's future water supply.

Another concern is saltwater migration into the aquifer. Sources of saltwater include the Arkansas River, oilfield brines that leaked from surface disposal pits or injection wells in the Burrton oilfield area (fig. 1), municipal wastewater facility discharges, and mineralized water from the underlying Wellington Formation (Ziegler and others, 1999; Whittemore, 2007). Declining water levels may accelerate migration of saltwater from the Burrton oilfield to the northwest and from the Arkansas River to the southwest into the freshwater of the *Equus* Beds aquifer (fig. 1) (Hansen, 2007).

## Description of Study Area

The study area (model area) consists of 1,844 square miles in Harvey, Kingman, Marion, McPherson, Reno, Rice, and Sedgwick Counties in south-central Kansas (fig. 1). Most of the area is within GMD2 and includes the major cities of Hutchinson and Wichita. Smaller cities within the study area include Burrton, Halstead, Newton, Sedgwick, and Valley Center.

The study area lies in the Arkansas River section of the Central Lowland physiographic province and has little relief except for an area of sand dunes north and northeast of Hutchinson. The land slopes gently toward the major streams in the area (Schoewe, 1949). Land-surface altitudes range from 1,270 to 1,650 ft above the North American Vertical Datum of 1988 (NAVD 88) and generally slope from northwest to southeast. The two major rivers that flow through the area, the Arkansas and the Little Arkansas, flow from northwest to southeast. Land-surface altitude and streams in the study area are shown in figure 2.

The study area has a continental climate with large variations in seasonal temperatures. Temperature extremes for the period of weather records at Wichita, Kans. range from more than 110 degrees to less than -20 degrees Fahrenheit ($^oF$). Temperatures above 90 $^oF$ occur an average of 63 days per year, whereas cold temperatures below zero occur about 2 days per year (National Oceanic and Atmospheric Administration, 2008). Mean annual precipitation at Wichita is about 30 inches (in). Most precipitation occurs during spring and summer. The wettest years have recorded more than 50 in. and the driest years less than 15 in. (National Oceanic and Atmospheric Administration, 2008).

Agriculture is the main land use in the study area. Field crops include corn, grain sorghum, soybeans, alfalfa, and wheat. Livestock production is primarily cattle. (Kansas Department of Agriculture, 2006). Irrigation from the *Equus* Beds aquifer supplements rainfall for agricultural production needs.

Extensive salt deposits are commercially mined in the study area. Rock salt was first discovered in Kansas in 1887. Several shaft mines and solution mines have been developed in Reno and Sedgwick Counties. In 2000, salt mines in Kansas produced 2,944,000 tons of rock salt, 75 percent of which is mined in Reno County (Kansas Geological Survey, 2011a, 2011b). Oil and natural gas production is an important industry in the study area. Cumulative oil production in Harvey, Reno, and Sedgwick Counties is more than 255 million barrels (bbl) and cumulative natural gas production is more than 136 billion cubic ft (Kansas Geological Survey, 2011c).

## Previous Studies

The *Equus* Beds aquifer has been extensively studied because it is a critical source of water to municipalities, industry, and agriculture in the study area. The geology and groundwater resources of the *Equus* beds were described by Williams and Lohman (1949) and the aquifer near the Wichita well field was described by Williams and Lohman (1949), Stramel (1956, 1962a, 1962b, 1967), and Petri and others (1964).

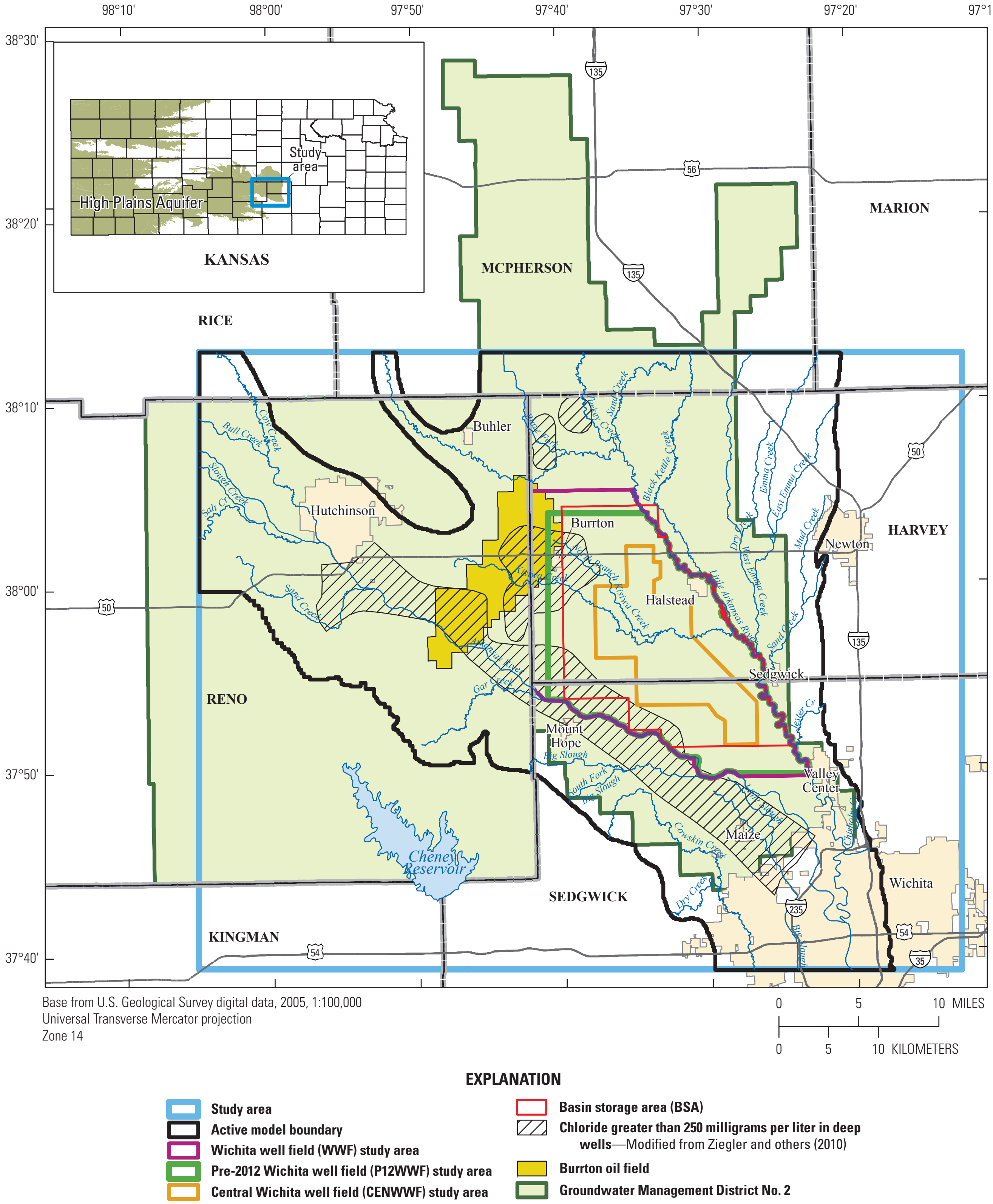
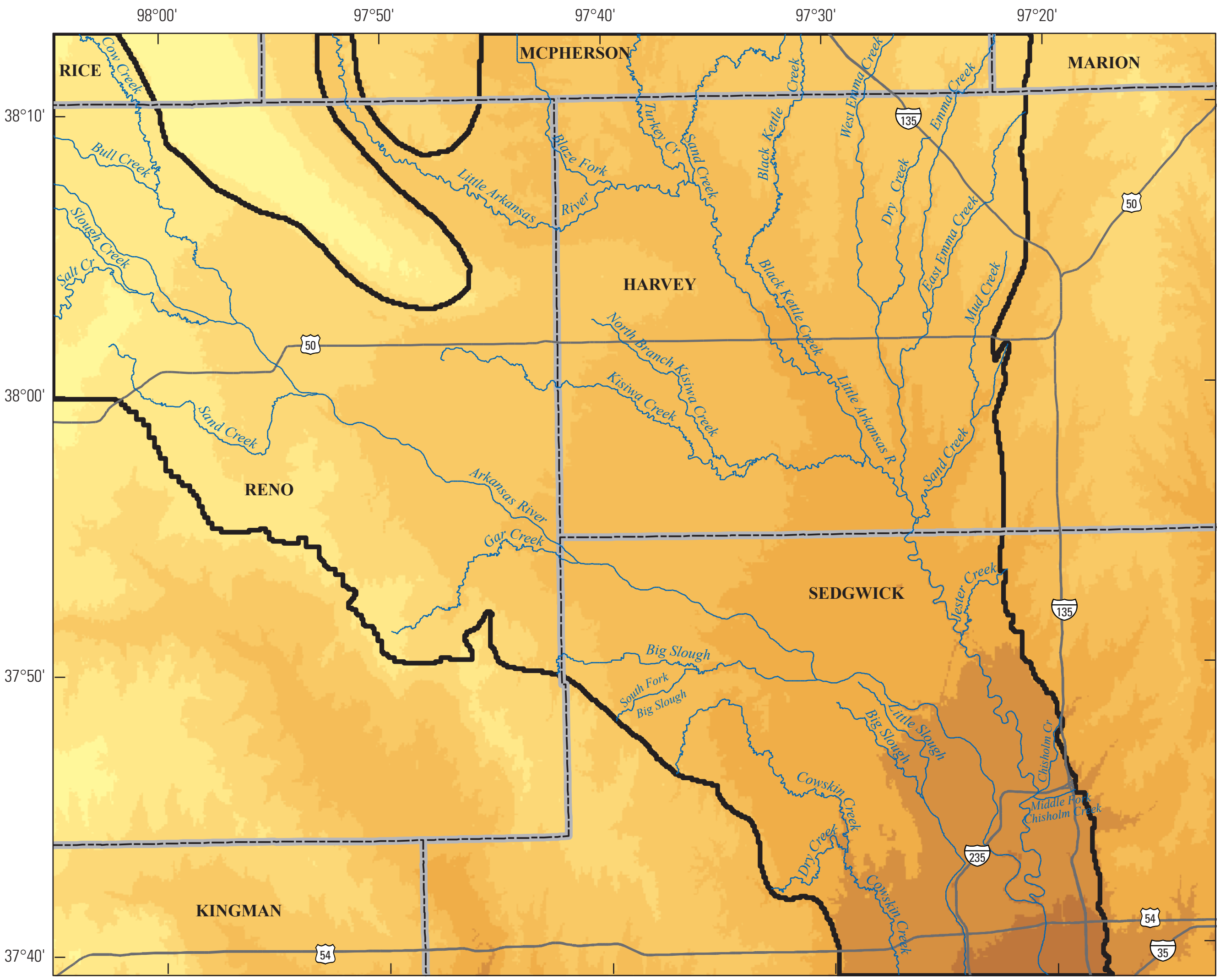



**Figure 1.** Location of study area.




Base from U.S. Geological Survey digital data, 2005, 1:100,000
Universal Transverse Mercator projection
Zone 14
Datum is North American Datum of 1983

0 5 10 MILES
0 5 10 KILOMETERS

**EXPLANATION**

**Land surface altitude**—In feet above North American Vertical Datum of 1988

- 1,601 to 1,650
- 1,551 to 1,600
- 1,501 to 1,550
- 1,451 to 1,500
- 1,401 to 1,450
- 1,351 to 1,400
- 1,301 to 1,350
- 1,251 to 1,300
- 1,201 to 1,250
- 1,151 to 1,200
- 1,112 to 1,150

**Active model boundary**

**Figure 2.** Land surface altitude and streams in the study area.

Water levels that have been collected by the city of Wichita since 1940 are on file with the city of Wichita, Kans. and are stored in the National Water Information System (NWIS) database at *http://waterdata.usgs.gov/ks/nwis* (U.S. Geological Survey, 2009a). Water-level data also have been collected by GMD2 since 1978 from wells in the *Equus* Beds aquifer (*Equus* Beds Groundwater Management District No. 2, 1995). Annual water-level data have been collected in the study area since 1937 by the Kanss Department of Agriculture–Division of Water Resources (KDA-DWR), USGS, and Kansas Geological Survey (KGS). The data on file with the USGS in Lawrence, Kans. are stored in the NWIS database (U.S. Geological Survey, 2009a); data on file with the KGS, including annual water-level data collected by KGS and KDA-DWR since 1997, are stored in the WIZARD database (Kansas Geological Survey, 2009a). Historical and near-real-time data and reports associated with the USGS work on the *Equus* Beds aquifer include Ziegler and others (1999); Ziegler and others (2010). Aucott and Myers (1998); Aucott and others (1998); Hansen and Aucott (2001, 2004, 2010); and Hansen (2007, 2009a, 2009b) published water-level-decline maps for the study area and discussed the changes in storage volume for noteworthy past and recent periods of time.

Several groundwater-flow models of the *Equus* Beds aquifer have been developed and used to describe groundwater flow, solute transport, or both in the *Equus* Beds aquifer. Sophocleous (1983) simulated chloride transport in the *Equus* Beds aquifer, Spinazola and others (1985) developed a model to simulate groundwater flow and chloride transport in the *Equus* Beds aquifer and underlying Wellington Formation, and Myers and others (1996) developed a model to simulate the interaction between the Arkansas River and groundwater in the *Equus* Beds aquifer and to use particle tracking to simulate chloride transport.

## Artificial Recharge

The water supply for the city of Wichita from the *Equus* Beds well field and Cheney Reservoir must be increased to meet future water needs through 2050 (Warren and others, 1995; J. Blain, oral commun., 2005). Stramel (1956, 1962a, 1962b, 1967) proposed the artificial recharge of the *Equus* Beds aquifer with streamflow runoff during periods of abundant precipitation. This artificially recharged water then could be recovered by pumping from the aquifer during periods of drought. Stramel proposed using a variety of techniques including water spreading, recharge pits or ponds, recharge wells, and induced recharge from streams by pumping. Stramel (1962b) also suggested investigation of the relation between streamflow or stage and water quality for the Arkansas River, Little Arkansas River, and Kisiwa Creek as sources for artificial recharge. Wichita's Integrated Local Water Supply Plan (Wichita, 2011) also calls for investigating *Equus* Beds recharge using excess water from the Little Arkansas River.

The city of Wichita initiated the *Equus* Beds Groundwater Recharge Demonstration Project in 1995 to test artificial recharge as a method for increasing water supply and preventing water-quality degradation (Ziegler and others, 1999). The purpose of the Demonstration Project was to investigate the feasibility of artificial recharge and its effects on the water quantity and quality of the *Equus* Beds aquifer. The project was a cooperative effort between the city of Wichita, the USGS, and the Bureau of Reclamation (U.S. Department of the Interior), with additional participation from the GMD2 and the U.S. Environmental Protection Agency (USEPA). The USGS roles in the cooperative study were to document changes in hydrologic and water-quality conditions in the study area, to identify the probable causes of the changes, and to develop a baseline condition for evaluating the effects of larger full-scale artificial recharge. Project work was coordinated with the Kansas Department of Health and Environment (KDHE), the Kansas Water Office (KWO), and the KDA-DWR. Burns and McDonnell Engineering Consultants (Kansas City, Missouri) and Mid-Kansas Engineering Consultants (Wichita, Kans.) provided engineering expertise and project management. The construction, maintenance, and operation of the recharge facilities were performed by the city of Wichita.

Diversion sites were constructed near the towns of Halstead and Sedgwick (fig. 3) to divert water from the Little Arkansas River for the Recharge Demonstration Project (Ziegler and others, 1999; Schmidt and others, 2007). At each site, water from the river was diverted when streamflow exceeded base-flow requirements established by KDA-DWR permit conditions. Different methods of diverting river water and recharging the aquifer were used at each site. At the Halstead diversion site, water could be pumped from a well adjacent to the Little Arkansas River when streamflow exceeded 42 cubic feet per second ($ft^3/s$ ) (minimum streamflow requirement established by KDA-DWR) from April 1 through September 30, and 20 $ft^3/s$ from October 1 through March 31 (Burns and McDonnell, 1996). This water was recharged to the aquifer through recharge basins, trenches, or injection wells at the Halstead recharge site. During this time, the number of days per year that minimum streamflow requirements in the Little Arkansas River were large enough to allow withdrawal of water for recharge from the well adjacent to the Little Arkansas River ranged from 99 days in 2002 to 349 days in 1999 (U.S. Geological Survey, 2011).

At the Sedgwick recharge site, water could be withdrawn directly from the Little Arkansas River when streamflow exceeded 40 $ft^3/s$ (minimum streamflow requirement) regardless of season. This water was treated to decrease turbidity and total fecal coliform bacteria, and to remove organic compounds before being recharged to the aquifer through recharge basins. Recharge activities at the Sedgwick site continued from April 1998 to November 2000. During this time, the number of days per year that minimum flow requirements in the Little Arkansas River were large enough to allow withdrawal of recharge water from the river ranged from 290 days in 2000 to 365 days

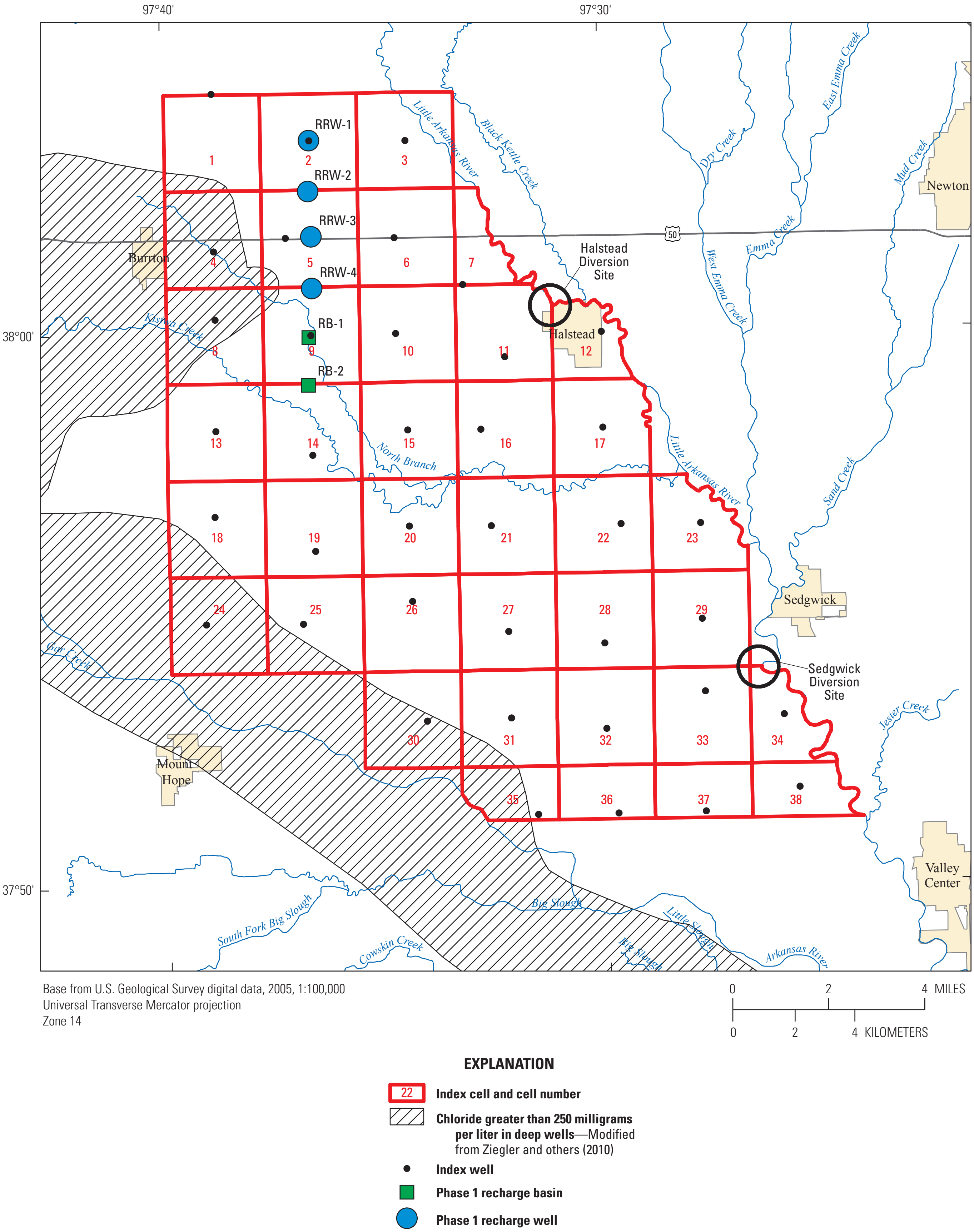



**Figure 3.** Location of aquifer storage and recovery project area.

in 1999 (U.S. Geological Survey, 2011). Total Demonstration Project recharge from 1995 through 2002 was more than 148.7 million cubic feet ($ft^3$) [3,416 acre-feet (acre-ft)].

In March 2006, the city of Wichita followed up the success of the Demonstration Project by starting construction of the *Equus* Beds ASR project to artificially recharge the *Equus* Beds aquifer on a larger scale. Other entities involved with the ASR project included GMD2 (Halstead, Kans.), KDA-DWR, KDHE, KWO, Bureau of Reclamation, USEPA, USGS, and various local interest groups and private consulting and engineering firms. Phase I of the ASR project was completed in 2006 and large-scale artificial recharge of the aquifer began at the phase I sites in March 2007. The phase I sites (fig.3) use water from the Little Arkansas River—pumped from the river directly or from wells in the riverbank that induce recharge from the river—as the source of artificial recharge to the *Equus* Beds aquifer. Water diversion from this river for artificial recharge only is allowed when flows are greater than minimums set by the State of Kansas (Kansas Department of Agriculture, Division of Water Resources, 2007).

The purpose of the *Equus* Beds ASR Project is to store and later recover groundwater and to form a hydraulic barrier to a known chloride-brine plume near Burrton, Kans. (figs. 1 and 3). A basin storage area (BSA) divided into index cells was defined by the city of Wichita and a groundwater-flow model has been used to calculate recharge credits for each index cell (Burns and McDonnell, 2008, 2009). Recharge credits indicate the volume of water Wichita has recharged to the aquifer, the movement of recharged water between index cells, and the amount of water Wichita can remove at a later date from index cells that contain recharged water. For each index cell, the groundwater model will determine the effect of natural and artificial recharge, groundwater inflow and outflow, evaporation and transpiration, groundwater diversions from nondomestic wells, infiltration and discharge to streams, calculated recharge credits, and surface water diversions.

## Purpose and Scope

The purpose of this report is to describe the development and calibration of a three-layer, groundwater-flow model of the *Equus* Beds aquifer in the area surrounding the Arkansas and Little Arkansas rivers near the city of Wichita well field. The report also presents results of simulated groundwater flow from 1935 (predevelopment conditions) to 2008 within the study area. The report summarizes the simulated effects of the Demonstration Project and phase I of the ASR project (Kansas Underground Injection Control Area Permit Class V Injection Well, Kansas Permit No. KS-05-079-001). The study described in this report is part of a long-term cooperative study (since 1940) between the city of Wichita and USGS to describe the water quantity and quality conditions in the *Equus* Beds aquifer and the Little Arkansas River, and more recently, the potential effects of artificial recharge on water resources in south-central Kansas.

Much of the data used as model input were collected during previous studies of the *Equus* Beds aquifer (Spinazola and others, 1985; Myers and others, 1996). Updated data include land-surface altitude; river stage for the Arkansas and Little Arkansas rivers; rainfall; lithologic and soils data; groundwater levels from 1935 to 2008; aquifer test data; groundwater pumpage from irrigation, industrial, and production wells located in the study area; and times and rates of artificial recharge.

This report includes maps showing the distribution of hydraulic properties for each model layer, locations of water-level data, simulated and observed water levels, and simulated groundwater flow and storage changes with and without phase I artificial recharge operation for 2006 through 2008.

# Groundwater Flow System

This section describes the groundwater flow system in the study area and includes discussion of the thickness and lithology of the Quaternary and Permian geologic deposits. Surface water hydrology is discussed and the major streams and tributaries are listed. The hydrology and conceptual model of groundwater flow is described and includes the areal extent and aquifer thickness, hydraulic properties of the aquifer, aquifer boundaries, recharge, discharge, groundwater level, groundwater-flow directions, and interactions of groundwater and surface water.

## Geology

The *Equus* beds are located in flat Quaternary deposits of unconsolidated sand, silt, and gravel deposits forming the Arkansas River Lowlands and Wellington-McPherson Lowlands (Zeller, 1968). The Quaternary alluvial deposits are as much as 330 ft thick in the study area and were derived from the Rocky Mountains to the west (Arkansas River Lowlands) and the High Plains to the north (Wellington-McPherson Lowlands). Eolian dune sand deposits occur in the northern part of the study area adjacent to the Little Arkansas River (fig. 4). The dunes are stable, vegetated, and are primarily made of well sorted, moderately well rounded, fine to medium sand and may locally include clay and silt.

The *Equus* beds consist primarily of sand and gravel interbedded with clay or silt but locally may consist primarily of clay with thin sand and gravel layers (Lane and Miller, 1965a; Myers and others, 1996). The middle part of the deposits generally has more fine-grained material than the lower and upper parts (fig. 5) (Lane and Miller, 1965b; Myers and others, 1996); however, areas of high hydraulic conductivity exist in all parts of the aquifer. The *Equus* beds overlie the Wellington Formation, part of the Lower Permian Sumner Group.

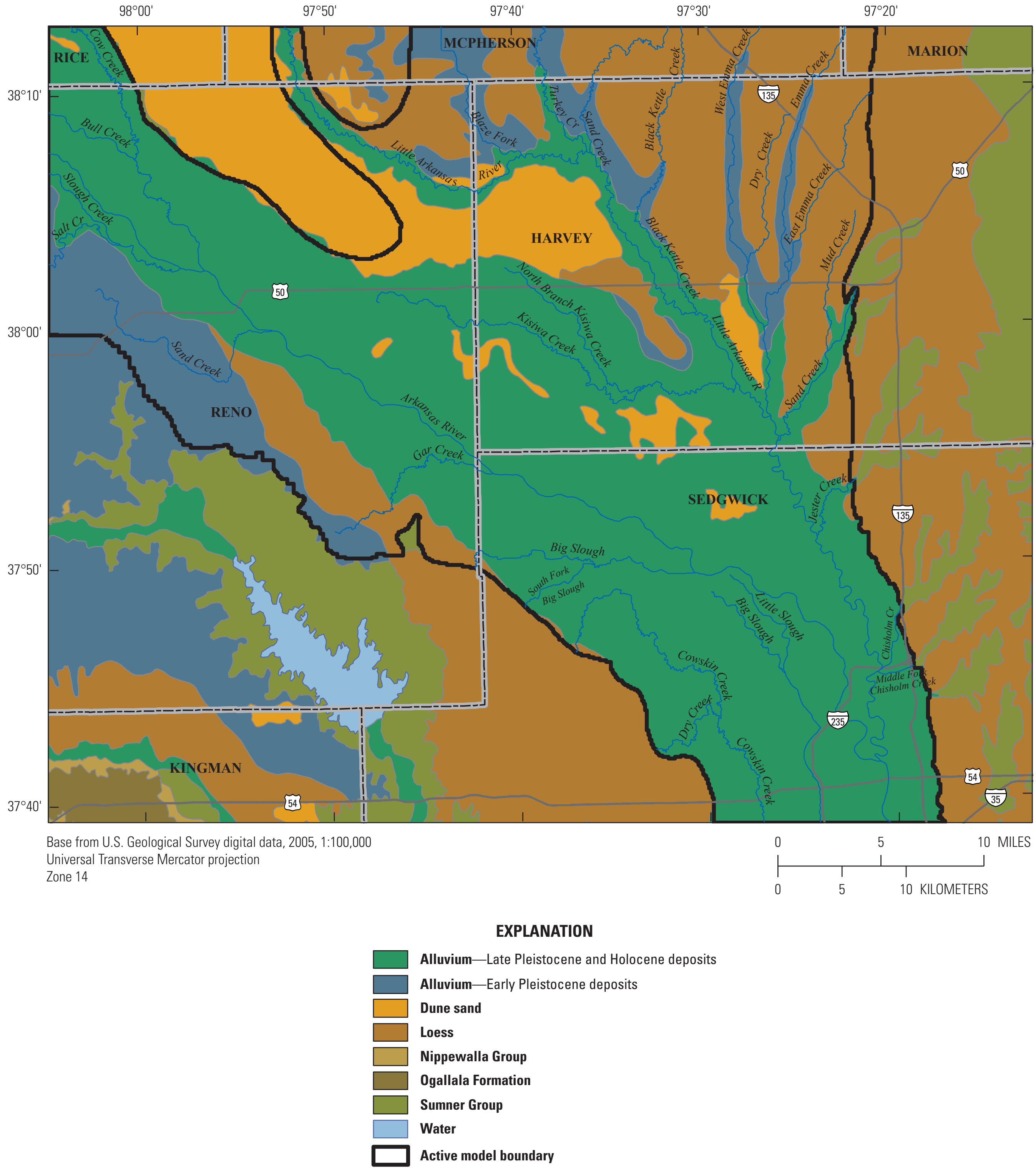



**Figure 4.** Surficial geology in the study area.

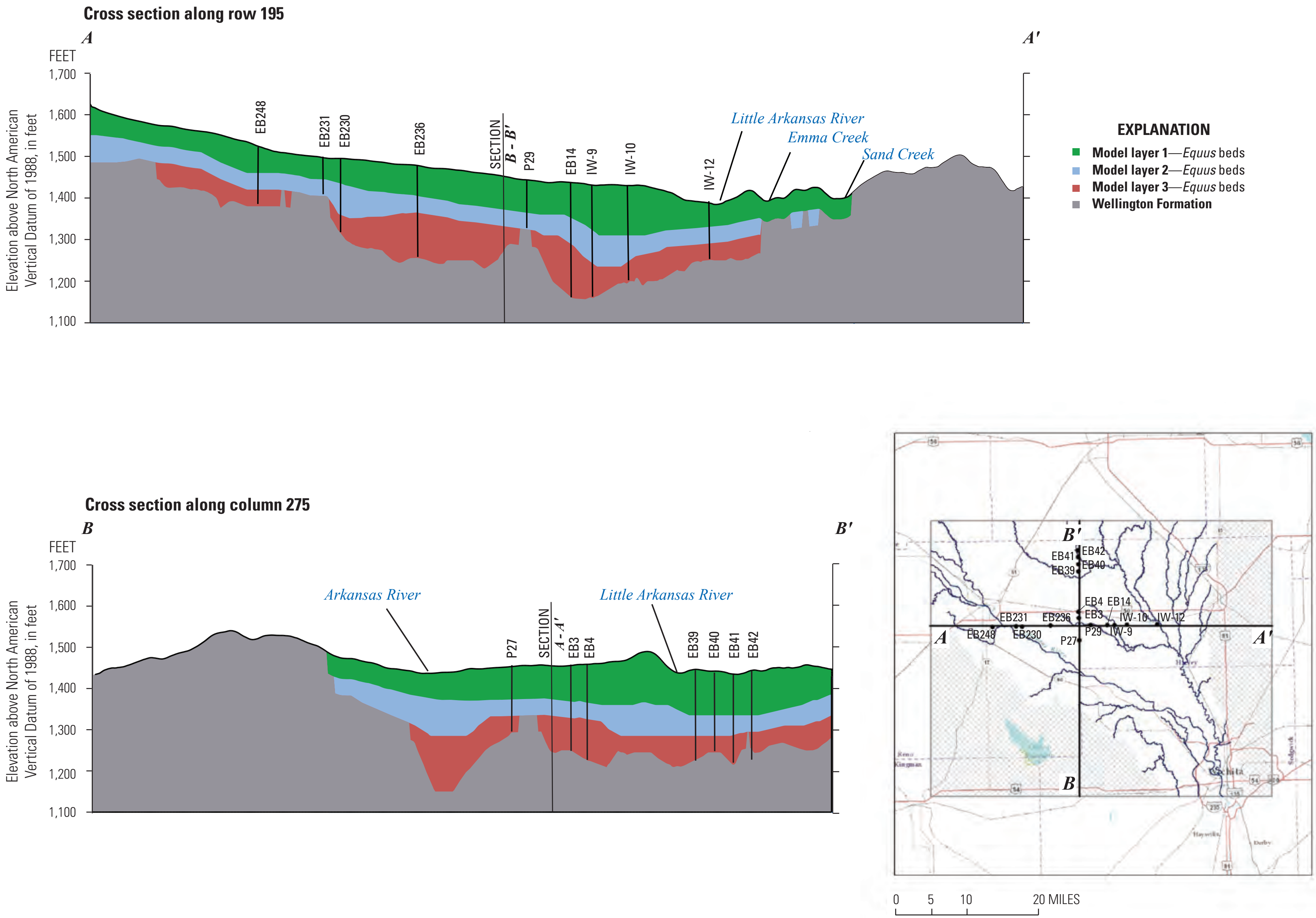



**Figure 5.** Geologic cross section in the study area from selected well lithologies.

The Wellington Formation of the Early Sumner Group of Permian age underlies the Quaternary deposits in the study area and forms the bedrock confining unit below these deposits. The Wellington Formation is about 700 ft thick (Bayne, 1956) and consists of three members—the upper shale member, about 200 ft thick; the Hutchinson Salt Member, about 300 ft thick; and the lower anhydrite member, about 200 ft thick (Myers and others, 1996). Dissolution of the Hutchinson Salt Member resulted in subsidence of the overlying upper shale member, formation of low areas in the bedrock, and concurrent accumulation of alluvial deposits that now compose the *Equus* Beds aquifer (fig. 5) (Myers and others, 1996).

Outcrops of the Wellington Formation immediately east of the study area primarily are shale with some limestone, dolomite, siltstone, gypsum, and anhydrite. The Hutchinson Salt Member of the Wellington Formation is present only in the subsurface, is composed primarily of halite (rock salt), and thickens to approximately 400 ft in areas west of the study area. The eastern margin of the rock salt is an active dissolution front and extends from just southwest of Wichita to near Salina, Kans. (Gogel, 1981). Although generally the freshwater *Equus* Beds aquifer and the underlying saltwater Wellington aquifer are separated by a confining unit, there is some exchange of freshwater and saltwater between the two, leading to replenishment of unsaturated water in the saltwater aquifer and saltwater contamination of the freshwater aquifer. Oil and gas deposits are located beneath many parts of the study area in rocks of Ordovician through Pennsylvanian age with accompanying oil wells that are drilled through the *Equus* Beds aquifer (Williams and Lohman, 1949).

## Surface Water

The primary streams in the study area that are simulated within the groundwater-flow model are the Arkansas and Little Arkansas Rivers, and tributaries including Turkey, Blaze Fork, Sand, Slough, Big Slough, Little Slough, Emma, Kisiwa, Chisholm, Cowskin, Salt, Jester, Black Kettle, Cow, Bull, and Gar Creeks (fig. 6). The major streams are hydraulically connected to the aquifer. Within the study area, the Arkansas River is hydraulically connected to the aquifer and, depending on streamflow, recharge, and well pumping, can be a gaining or a losing stream. The Little Arkansas River primarily is a gaining stream within the study area. The 12 streamflow gages located within the study area are shown on figure 6. Streamflow data from these stations were used to estimate the river stage and the flux of water into and out of the aquifer through the streams.

## Hydrology and Conceptual Model of Groundwater Flow

Groundwater flow in the *Equus* Beds aquifer is affected largely by areal recharge, streams, topography, and the hydraulic properties of the aquifer material. These processes and properties were used to construct a conceptual model of groundwater flow that identifies the hydrologic processes that need to be simulated in the groundwater-flow model. The hydrology and conceptual model of the *Equus* Beds aquifer presented in this report are based on previous studies of groundwater flow (Spinazola and others, 1985; Myers and others, 1996; Pruitt, 1993).

### Areal Extent and Aquifer Thickness

The *Equus* Beds aquifer is bounded at the top by the water table, laterally by the extent of the *Equus* beds deposits, and at the base by shales of the Wellington Formation. The *Equus* Beds aquifer extends north and west from the study area and arbitrary model boundaries were established at these edges of the study area. Depth to bedrock data (Kansas Geological Survey, 2009b) were used to define model geometry in the model area. Depth to bedrock and locations of depth to bedrock data are shown in figure 7. The thickness and extent of the aquifer in the study area are shown in figure 8. The greatest aquifer thickness is about 330 ft, but average thickness is about 110 ft.

### Hydraulic Properties of the Aquifer

Hydraulic properties of the aquifer include hydraulic conductivity, specific yield, and storage coefficient. Hydraulic conductivity is the capacity of the aquifer to transmit water and is measured as the volume of water at the existing kinematic viscosity that will move in unit time under a unit hydraulic gradient through a unit area at a right angle to the direction of flow (American Society of Civil Engineers, 1985). Reported hydraulic conductivity values for the *Equus* Beds aquifer range from 55 to 1,000 feet per day (ft/day) (Reed and Burnett, 1985); 50 to 1,200 ft/day (Myers, 1996); and 5 to 750 ft/day (Spinazola and others, 1985). The specific yield, the storage coefficient in an unconfined aquifer, is a measure of the ratio of the volume of water that will drain because of gravity to the volume of saturated aquifer. For the *Equus* Beds aquifer, the specific yield has been reported to be between 0.08 and 0.35 (Williams and Lohman, 1949; Reed and Burnett, 1985; Spinazola and others, 1985; Fetter, 1988; Myers and others, 1996). The storage coefficient (for confined conditions) has been reported to be between 0.0004 and 0.16 (Reed and Burnett, 1985).

The distribution within the aquifer of clay, silt, sand, and gravel controls the distribution of hydraulic properties of the *Equus* Beds aquifer. Clay and silt, for example, have smaller hydraulic conductivity values than sand and gravel.

### Aquifer Boundaries

The shale within the Wellington Formation forms the lower bedrock boundary of the *Equus* Beds aquifer in the study area and is considered a no-flow boundary in the

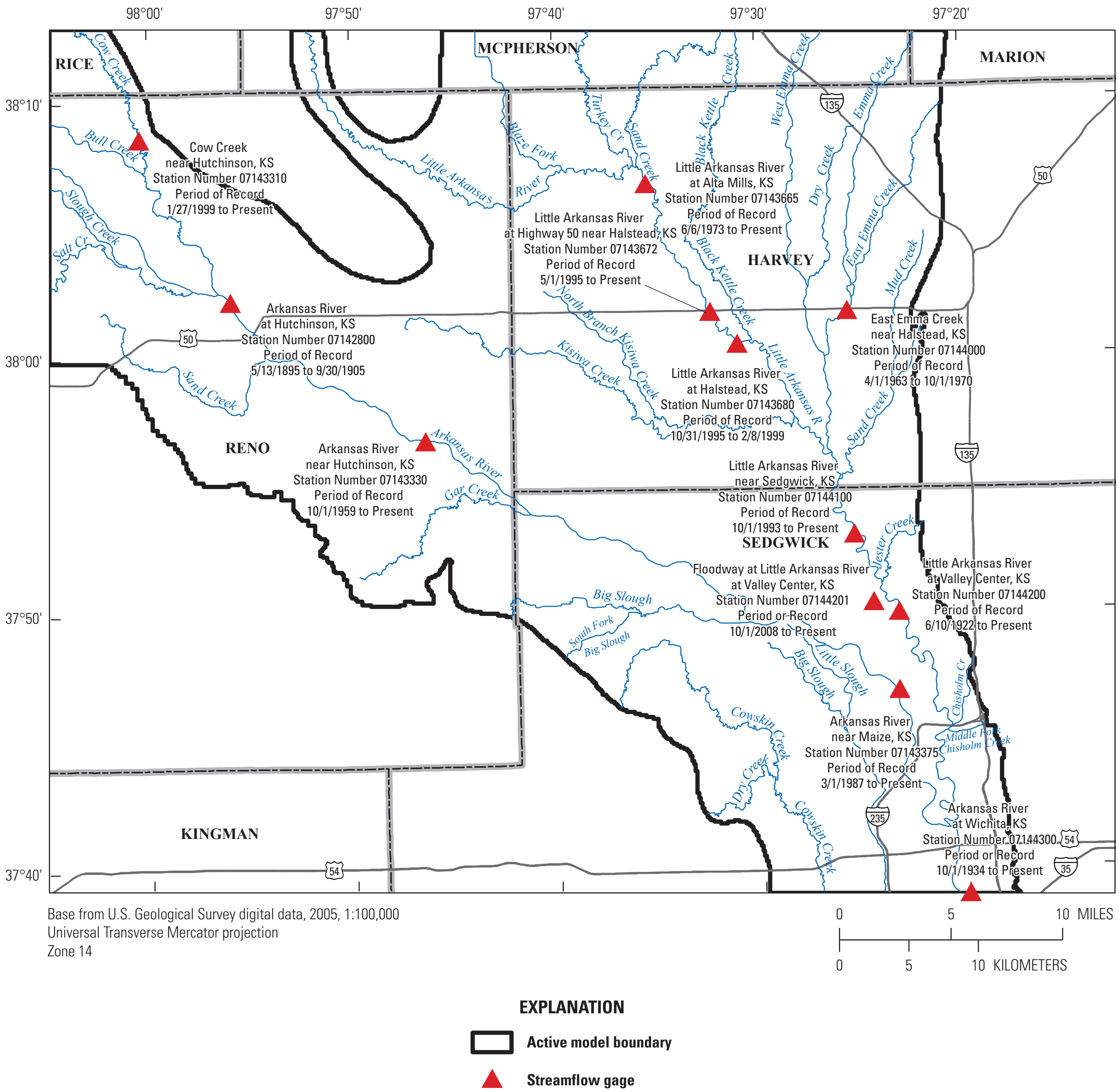



**Figure 6.** Streams and streamflow gages in the study area.

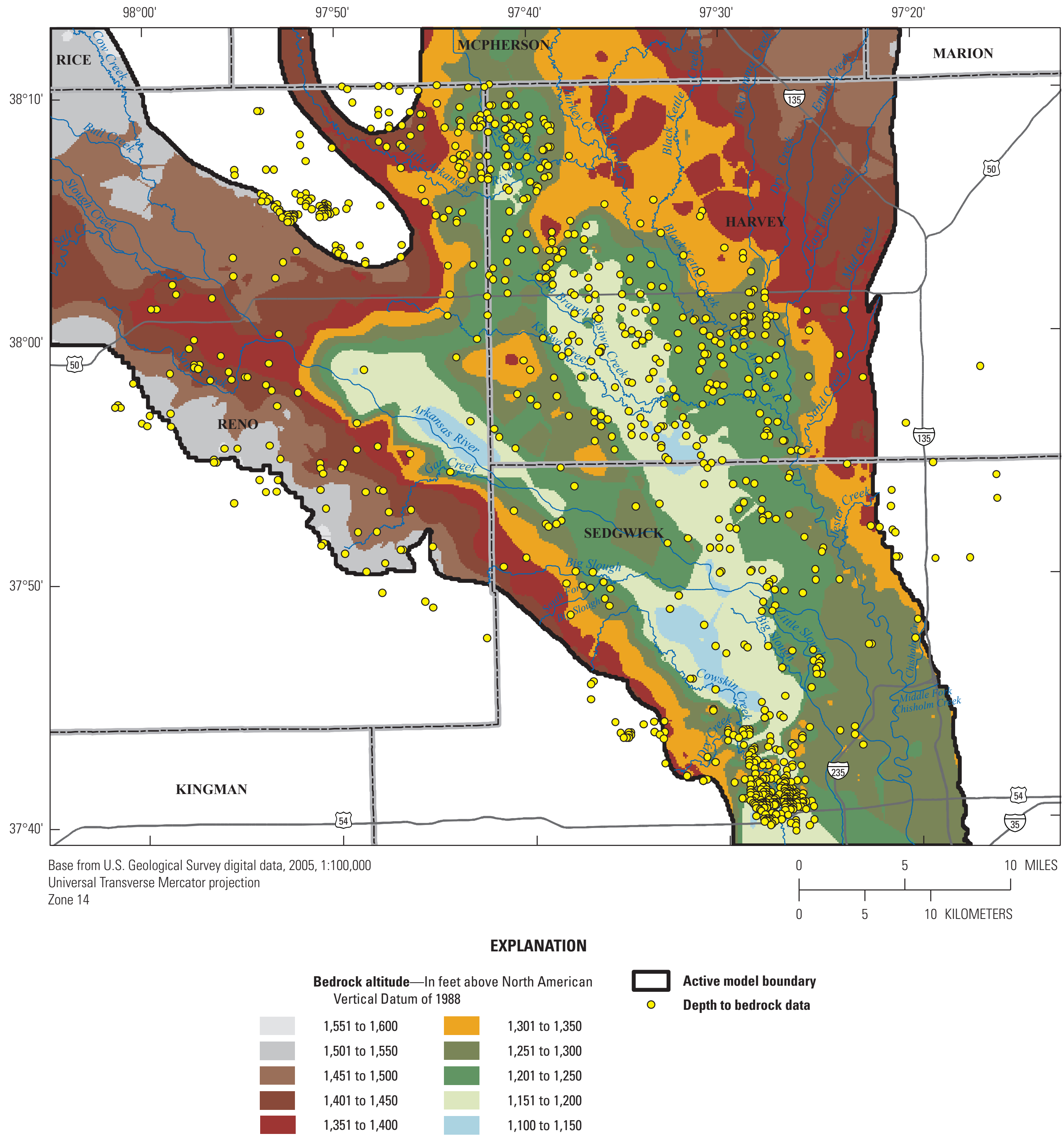



**Figure 7.** Bedrock altitude and locations of wells with depth to bedrock data.

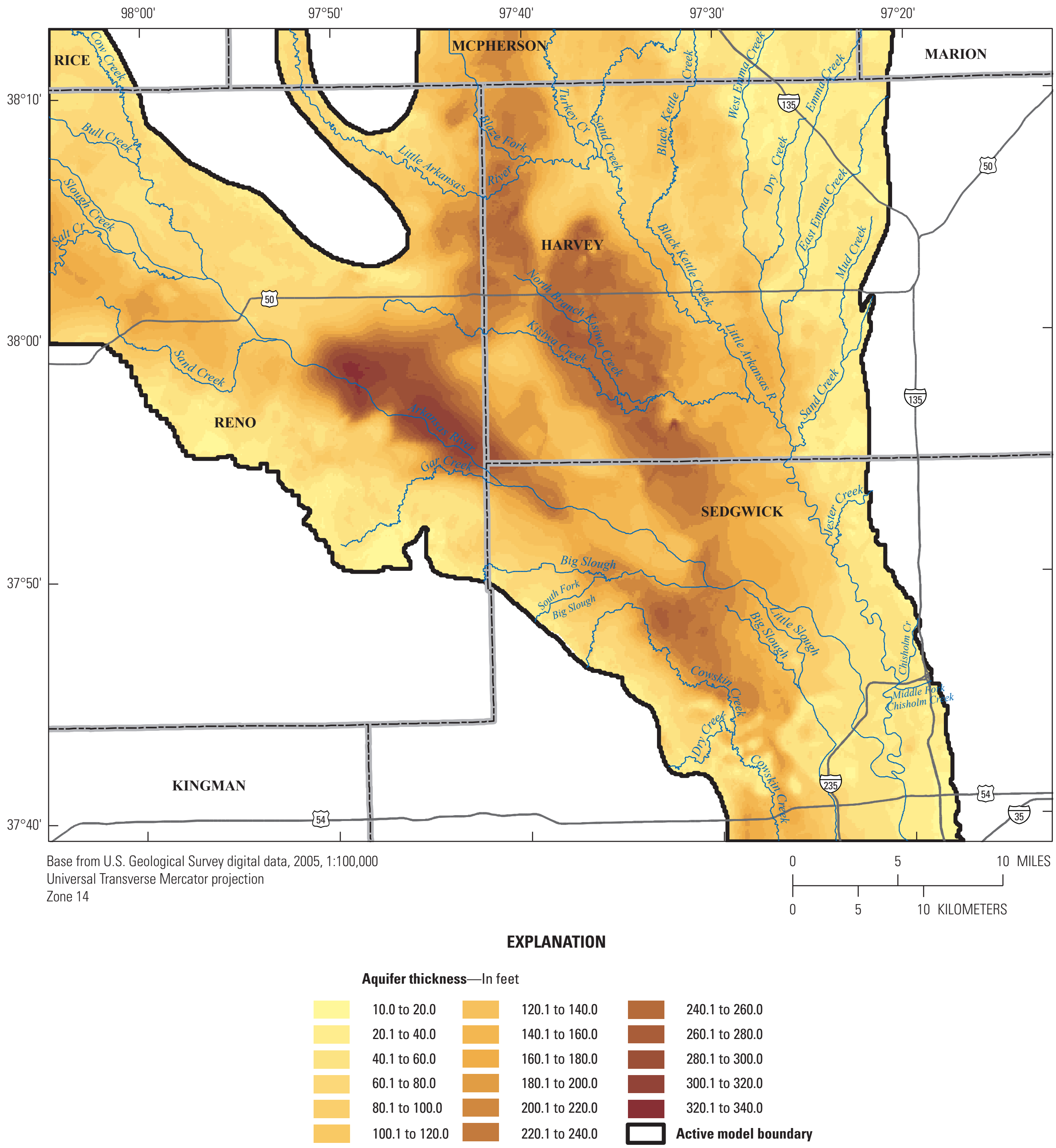



**Figure 8.** Thickness of *Equus* Beds aquifer.

groundwater-flow model, because hydraulic conductivity, calculated from vertical leakance values for the confining bed between the *Equus* Beds and Wellington aquifers (Spinazola and others, 1985), ranges from 1.5 x $10^{-6}$ ft/day to 2.42 x $10^{-4}$ ft/day and is several orders of magnitude less than the hydraulic conductivity of the overlying *Equus* Beds aquifer. The lateral extent of the *Equus* beds formation on the southern and eastern edges of the study area also is considered a no-flow boundary in the model.

Lateral boundaries of the *Equus* Beds aquifer at the northern, western, and southern edges of the study area are not physical hydraulic boundaries, but were chosen as model boundaries based on the study objectives and area of interest. Groundwater flow into the system is simulated through the upgradient northern and western lateral boundaries of the study area and flow out of the system is simulated through the downgradient southern boundary of the study area using head-dependent fluxes (fig. 1).

## Recharge

The water table is the boundary across which recharge from precipitation flows into the aquifer. Areally distributed recharge occurs when the rate of precipitation or snowmelt exceeds the rate of runoff and evapotranspiration. When recharge exceeds discharge, water levels rise in the aquifer; the converse also is true.

Topography generally has little effect on the areal distribution of recharge, although low-lying areas in the study area may have larger recharge rates caused by collected runoff. Rather, the vertical hydraulic conductivity of soils directly controls the rate of infiltration in most areas. Aquifer recharge is greater beneath ponded areas or soils with larger vertical hydraulic conductivities than beneath soils with smaller vertical hydraulic conductivities. Therefore, soil variability affects the areal distribution of recharge to the aquifer.

Recharge in the study area has been estimated in previous studies to be between 10 and 20 percent of precipitation (Hansen, 1991). Assuming 30 inches of rainfall per year (National Oceanic and Atmospheric Administration, 2008), recharge to the aquifer in the study area based on these percentages would range from 3 to 6 inches per year. For the active model area, total recharge would be between 21.2 and 42.4 million $ft^3$/day or 487 and 973 acre-feet per day (acre-ft/day).

## Discharge

In 2008, 157 production wells, 986 irrigation wells, and 104 industrial wells were in operation in the active model area. Water pumped from the aquifer in the study area by production wells in 2008 totaled almost 4,370,000 $ft^3$/day (100.3 acre-ft/day) (Kansas Department of Agriculture–Division of Water Resources, unpub. data, 2009). Cumulative pumping and rate of pumping from the *Equus* Beds aquifer by industrial, irrigation, production, artificial recharge, and artificial withdrawal wells from 1935 through 2008 are shown in figure 9. From 1939 through 2008, total cumulative pumping from the *Equus* Beds aquifer in the study area was almost 273 billion $ft^3$ (6.26 million acre-ft). Total pumping steadily increased from 1939 to 1991 and then stabilized from 1992 through 2008 to between 12 and 18 million $ft^3$/day (275 to 413 acre-ft/day). Municipal and industrial pumping decreased but irrigation pumping increased from 1990 through 2008.

Evapotranspiration removes water from the aquifer when the water table is near land surface and within the root zone of vegetation. Evapotranspiration is a combination of evaporation and uptake of water by plants and is greatest during the growing season. Evapotranspiration was estimated by Spinazola and others (1985) to be 3.5 inches per year. For the active model area, total estimated evapotranspiration is about 24.7 million $ft^3$/day or 568 acre-ft/day.

## Groundwater Level

Groundwater-level data have been collected periodically from more than 100 wells by city of Wichita personnel using standard water-level measurement techniques that are similar to USGS methods described in Cunningham and Schalk (2011). Data collection began just before the beginning of city pumpage from the aquifer in 1940; water levels in most wells have been measured at least quarterly. These data are on file in paper and electronic form with the city of Wichita Water and Sewer Department in Wichita, Kans., and are stored in the USGS National Water Information System (NWIS) database (U.S. Geological Survey, 2009a).

During 2001 and 2002, 38 pairs of index wells in the study area were installed by Burns and McDonnell Engineering Consultants and Clarke Well Equipment, Inc. (Great Bend, Kans.) for the city of Wichita (Debra Ary, city of Wichita, written commun., September 25, 2009). Each pair of index wells consists of a well completed in the upper part of the aquifer and another well completed in the lower part of the aquifer. These wells were designed for use by the city to monitor water quality and water levels in the aquifer and any changes that might occur as a result of the ASR project. These wells also were used to determine if there are any water-quality differences between the shallow and deep parts of the aquifer. These wells were added to the water-level monitoring network in the study area in 2002. Water levels in the index wells were measured quarterly by GMD2 and occasionally by the USGS; all index well water-level measurements used in this report were measured by GMD2. The data collected by the USGS are stored in the NWIS database (U.S. Geological Survey, 2009a). The data collected by GMD2 are stored in the Kansas Geological Survey's (KGS's) Water Information Storage and Retrieval Database (WIZARD) (*http://www.kgs.ku.edu/Magellan/WaterLevels/index.html*).

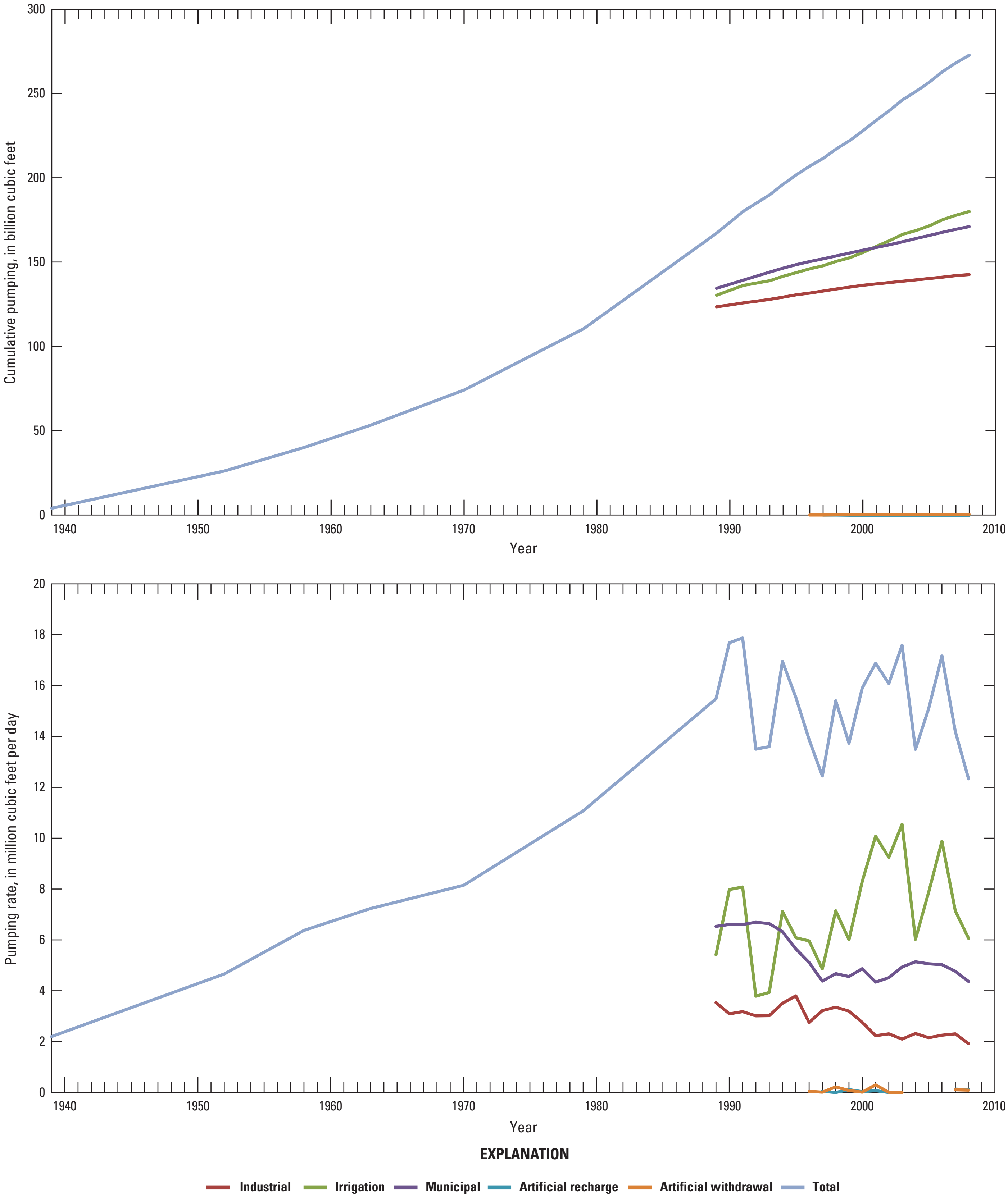



**Figure 9.** Cumulative pumping and pumping rate for the *Equus* Beds aquifer by industrial, irrigation, production, artificial recharge and artificial withdrawal wells in the study area from 1939 through 2008.

The drought during the 1950s and groundwater pumpage from the aquifer near the Wichita well field for production and agricultural use between 1940 and 1957 caused a substantial water level decline in and near the Wichita well field (Hansen and Aucott, 2004). Increased irrigation pumpage during the 1970s and 1980s caused further declines in groundwater levels (Myers and others, 1996; Aucott and Myers, 1998). Most of the water-level declines were caused by groundwater pumpage but the effects of climate on recharge also have affected water levels (Hansen and Aucott, 2003). Groundwater level altitudes in parts of the aquifer near the Wichita well field (fig. 1) increased by more than 20 ft between 1992 and 2006. Other factors contributing to water-level increases include subsurface inflow, streamflow losses, and irrigation return flow (Myers and others, 1996).

In areas where the aquifer is well connected hydraulically, shallow and deep groundwater levels are similar; however, in areas where the aquifer is semi-confined, substantial differences in shallow and deep groundwater levels exist (Hansen and Aucott, 2003). The dune sands in the northwest part of the study area (fig. 4) contain layers of silt and clay that limit the downward movement of water (Myers and others, 1996), as indicated by the existence of interdune ponds (Williams and Lohman, 1949) and shallow water levels in closely spaced wells that are 27 ft higher than deeper water levels (Williams and Lohman, 1949). Although downward movement of groundwater is limited in this area, the existence of a groundwater mound in the *Equus* beds deposits below the sand dune area indicates recharge through the sand dunes is larger than in surrounding areas (Myers and others, 1996).

### Groundwater-Flow Directions

Groundwater flow within the aquifer in the area between the Arkansas and Little Arkansas Rivers generally is west to east and groundwater flow in the area north of the Little Arkansas River is from north to south. Groundwater level maps from 1940 and 1989 (Myers and others, 1996) illustrate the general flow of groundwater in the *Equus* Bed aquifer at these times (figs. 10 and 11). Groundwater withdrawals create localized cones of depression around each well or well field that may alter regional groundwater to flow toward the wells. A cone of depression generally has the shape of an inverted cone with the lowest part centered at the pumping well. Although cones of depression around wells are not visible at the scale shown in figures 10 and 11, increased well pumping in 1989 along and west of the Little Arkansas River has lowered the water table and altered groundwater flow directions compared to 1940 conditions.

### Groundwater/Surface-Water Interaction

Long-term withdrawal of groundwater from the *Equus* Beds aquifer lowered groundwater-levels in the area of the Wichita well field (Hansen and Aucott, 2003). Seepage runs conducted in the 1980s and 1990 indicate that the Arkansas River in the study area either gained or lost water in the upper reach (upstream from the point midway between the streamflow gages near Maize (07143375) and Hutchinson (07143330), fig. 6) but lost water in the lower reach that is adjacent to the area of lowered groundwater levels (Myers and others, 1996).

Streamflow, estimated base flow (Lim and others, 2005; Sloto and Crouse, 1996), and the difference in base flow are shown for the Arkansas River for the streamflow gages near Hutchinson (07143330) and Maize (07143375) from December 1989 through 2008 in figure 12 and for the Little Arkansas River for the streamflow gages at Alta Mills and Valley Center from December 1989 through 2008 in figure 13. As shown in figure 12, the Arkansas River is a gaining stream between the streamflow gages at Hutchinson (07143330) and Maize (07143375) during high flows most likely associated with times of increased recharge to the *Equus* Beds aquifer; however, during low flow most likely associated with decreased recharge, the Arkansas River is a losing stream in this reach. As shown in figure 13, the Little Arkansas River is a gaining stream between the streamflow gages at Alta Mills and Valley Center from 1989 through 2008 with base flow increasing during high flows most likely associated with times of increased recharge to the *Equus* Beds aquifer and decreasing during low flows most likely associated with periods of decreased recharge.

# Methods

This section describes the methods used to simulate groundwater flow and includes discussion of the computer software and the equation used to simulate groundwater flow. Spatial and temporal discretization of the finite-difference groundwater-flow model is discussed and the hydrogeologic framework is described. Parameter values, associated model zones, and the hydraulic properties of the *Equus* Beds aquifer are listed and illustrated. Boundary conditions required to simulated groundwater flow are discussed including recharge, evapotranspiration, streamflow, well pumping, and lateral boundaries of the aquifer where groundwater flows into, or out of, the model area. Observations of head and streamflow are described. The techniques used and criteria for model calibration are discussed including initial conditions, steady-state calibration, transient calibration, and parameter sensitivity. A comparison between simulated and measured groundwater level change for selected times is presented and the limitations of the model are listed and discussed.

## Groundwater Flow Simulation

Groundwater flow was simulated for the *Equus* Beds aquifer using the three-dimensional finite-difference

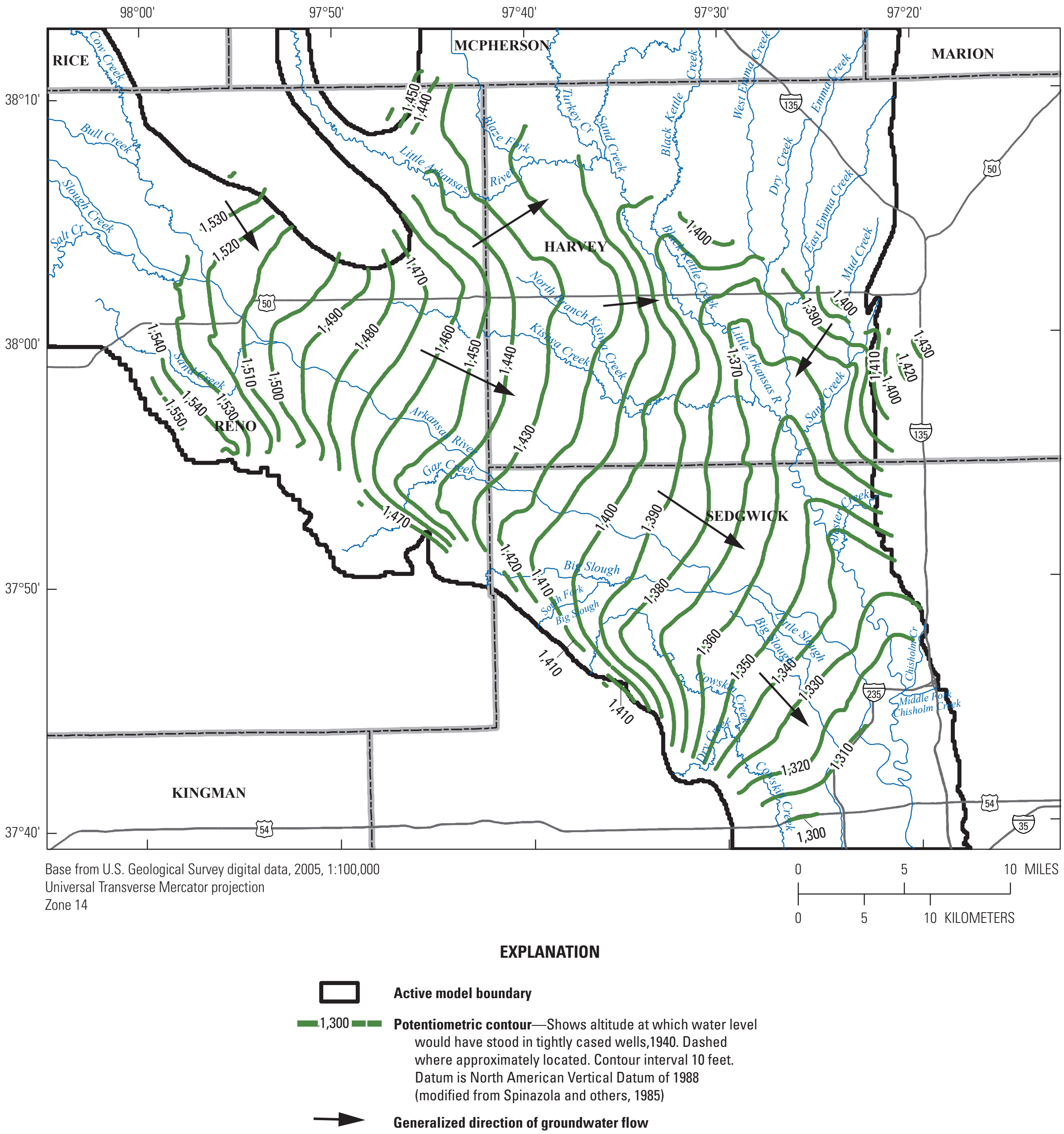
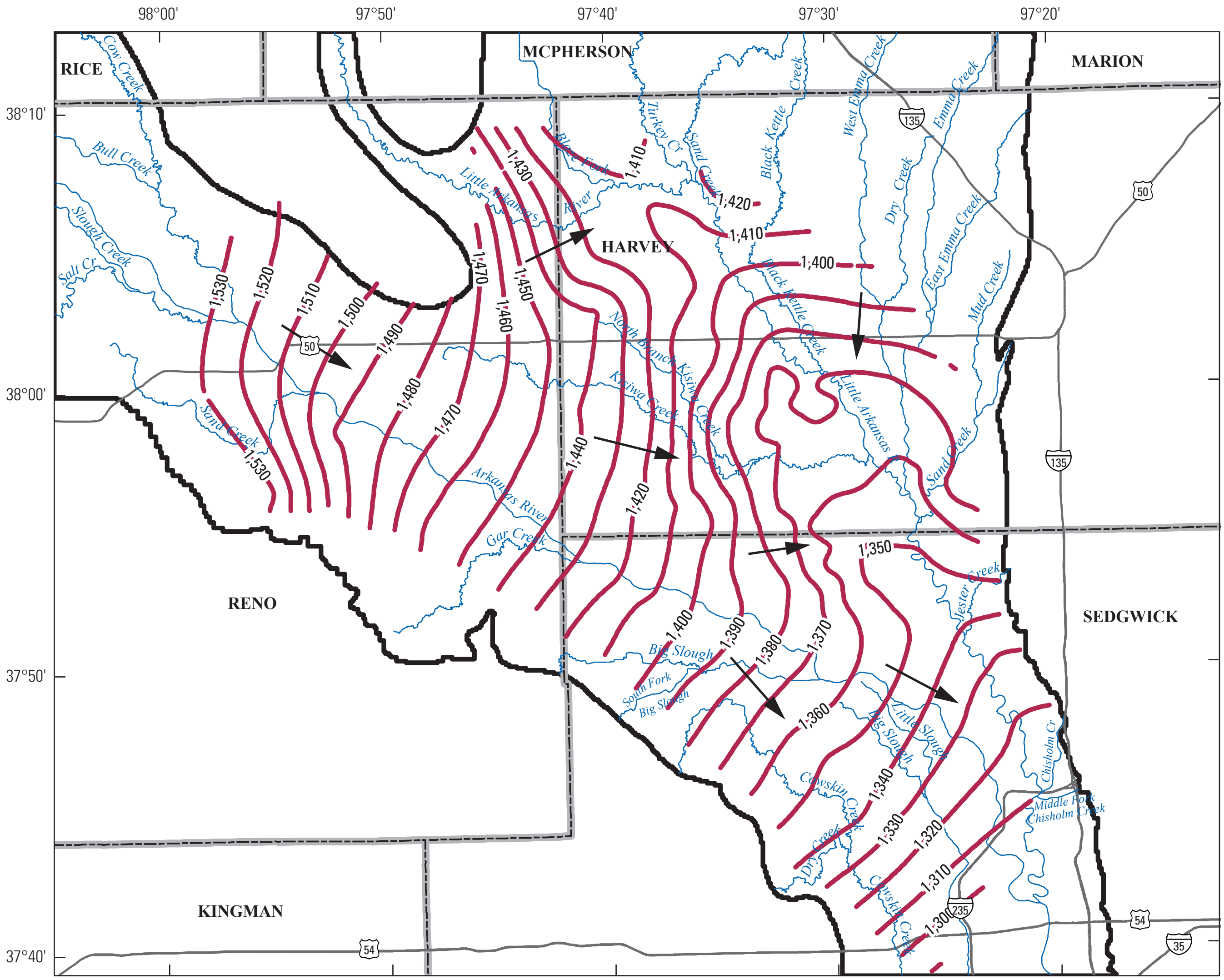



**Figure 10.** *Equus* Beds aquifer groundwater altitude, 1940.




Base from U.S. Geological Survey digital data, 2005, 1:100,000
Universal Transverse Mercator projection
Zone 14

0 5 10 MILES
0 5 10 KILOMETERS

**EXPLANATION**

**Active model boundary**

1,300 **Potentiometric contour**—Shows altitude at which water level would have stood in tightly cased wells, 1989. Dashed where approximately located. Contour interval 10 feet. Datum is North American Vertical Datum of 1988—(modified from Myers and others, 1996)

**Generalized direction of groundwater flow**

**Figure 11.** *Equus* Beds aquifer groundwater altitude, 1989.


**Figure 12.** Streamflow, estimated base flow, and the difference between base flow for the Arkansas River streamflow gages near Hutchinson (07143330) and Maize (07143375), December 1989 through 2008.

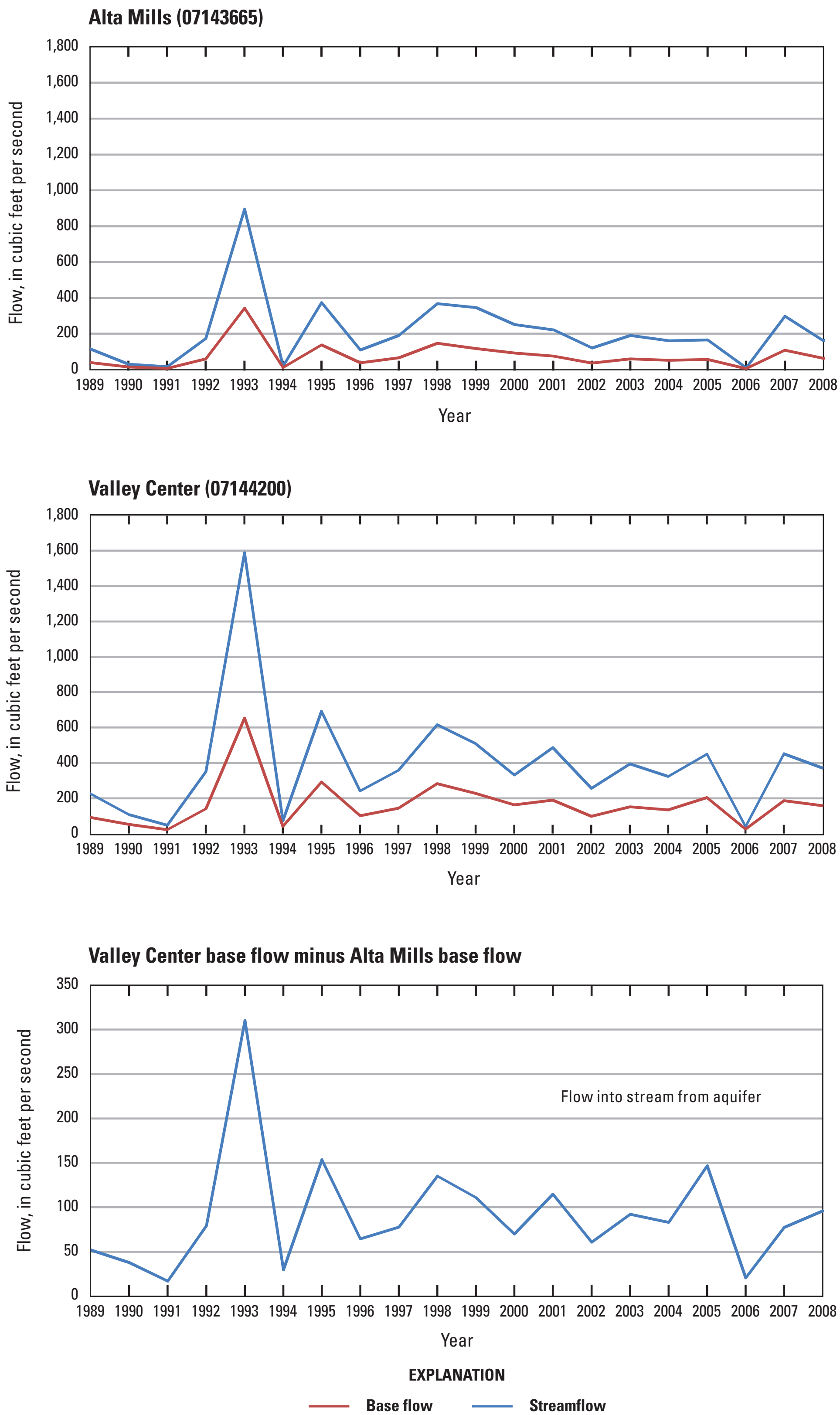



**Figure 13.** Streamflow, estimated base flow, and the difference between base flow for the Little Arkansas River streamflow gages at Alta Mills (07143665) and Valley Center (07144200), December 1989 through 2008.

groundwater-flow model MODFLOW-2000 (Harbaugh and others, 2000). MODFLOW-2000 is a modified version of MODFLOW (McDonald and Harbaugh, 1988) that incorporates the use of parameters to define model input and the calculation of parameter sensitivities. In addition, the code incorporates the modification of parameter values to match observed heads, flows, or advective transport using the observation, sensitivity, and parameter-estimation processes described by Hill and others (2000).

Three-dimensional simulation of groundwater flow in the *Equus* Beds aquifer was necessary to accurately determine the hydraulic-head distribution in the aquifer. Substantial differences in shallow and deep groundwater levels exist (Hansen and Aucott, 2003) in areas where the aquifer is semi-confined. Discharge from the aquifer to rivers may vary according to river size, depth of the streambed, or streambed conductance. Groundwater flow may be divided into smaller flow subsystems because of the degree of interaction between groundwater, the well fields, and the larger and smaller rivers in the study area. Pumping from the well fields located near a river can induce flow from the river and cause groundwater flow beneath the river.

The following equation was the governing equation used in MODFLOW-2000 to approximate groundwater flow rates in three dimensions:

$$\frac{\partial}{\partial x}\left(K_x \frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_y \frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_z \frac{\partial h}{\partial x}\right)-W=Ss\frac{\partial h}{\partial t} \quad (1)$$

where

$K_x$, $K_y$, and $K_z$ are the values of hydraulic conductivity along the $x$, $y$, and $z$ coordinate axes and are assumed to be parallel to the major axes of hydraulic conductivity, in feet per day;

$h$ is the potentiometric head, in feet;

$W$ is a volumetric flux per unit volume and represents sources or sinks, or both, of water, such as well discharge, leakage through confining units, streambed leakage, recharge, and water removed from the aquifer by drains, per day;

$Ss$ is the specific storage of the porous material, per foot; and

$t$ is time, in days.

## Initial Conditions

Initial conditions for the transient calibration simulation were obtained from the steady-state calibration simulation. These initial conditions included the head distribution simulated using average hydrologic conditions from 1935 through 1939 and recharge, evapotranspiration, streamflow, general head boundary conditions, and well pumping.

## Spatial and Temporal Discretization

The modeled area covers almost 1,845 square miles including the entire study area shown in figure 1. The model has uniform cells 400 ft per side and contains 963,900 cells in 510 rows, 630 columns, and 3 model layers. Model layer 1 is the topmost, model layer 2 is the middle layer, and model layer 3 is the bottom layer. Model layer thickness and areal extent are shown for model layer 1 in figure 14, model layer 2 in figure 15, and model layer 3 in figure 16. The regular grid spacing facilitated data input from a Geographic Information System (GIS) and analysis of model output by the GIS. The small size of each cell limits the error associated with particle tracking and solute transport, which are potential uses for the model. Cells containing sinks that do not discharge at a rate large enough to consume all the water entering the cell introduce uncertainty into the computed path of the imaginary particle. The irregular shape of the active model boundary reduced the number of active cells in the model to 369,346 with 177,572 active cells in model layer 1 (fig. 14); 123,265 active cells in model layer 2 (fig. 15); and 68,509 active cells in model layer 3 (fig. 16).

The model simulates steady-state and transient conditions. Steady-state conditions were simulated using average hydrologic conditions from 1935 through 1939 that include recharge, evapotranspiration, streamflow, flow across model boundaries, and well pumping. Transient conditions including recharge, evapotranspiration, streamflow, flow across model boundaries, and well pumping were simulated from 1935 to 2008 using 26 stress periods. Stress periods 1 through 7 simulate groundwater flow from 1935 through 1989 and stress period lengths are from Myers and others (1996). Yearly stress periods 8 through 26 simulate groundwater flow from 1990 through 2008 and allow simulation of changes in areally distributed recharge based on average annual precipitation. Stress periods, time steps, and time-step multipliers are listed in table 1 for all stress periods.

**Table 1.** Transient groundwater simulation stress periods, stress period start date, stress period length, time steps, and time-step multipliers.

| Stress period | Stress period start date | Stress period length | Time steps | Time-step multiplier[1] |
|---|---|---|---|---|
| 1 | January 1, 1935 | 5 years | 50 | 1.01 |
| 2 | January 1, 1940 | 13 years | 50 | 1.01 |
| 3 | January 1, 1953 | 6 years | 50 | 1.01 |
| 4 | January 1, 1959 | 5 years | 50 | 1.01 |
| 5 | January 1, 1964 | 7 years | 50 | 1.01 |
| 6 | January 1, 1971 | 9 years | 50 | 1.01 |
| 7 | January 1, 1980 | 10 years | 50 | 1.01 |
| 8 to 26 | January 1, 1990 | 1 year | 50 | 1.01 |

[1]The time-step multiplier is used by MODFLOW to calculate a geometric increase in the length of each time step within a stress period.

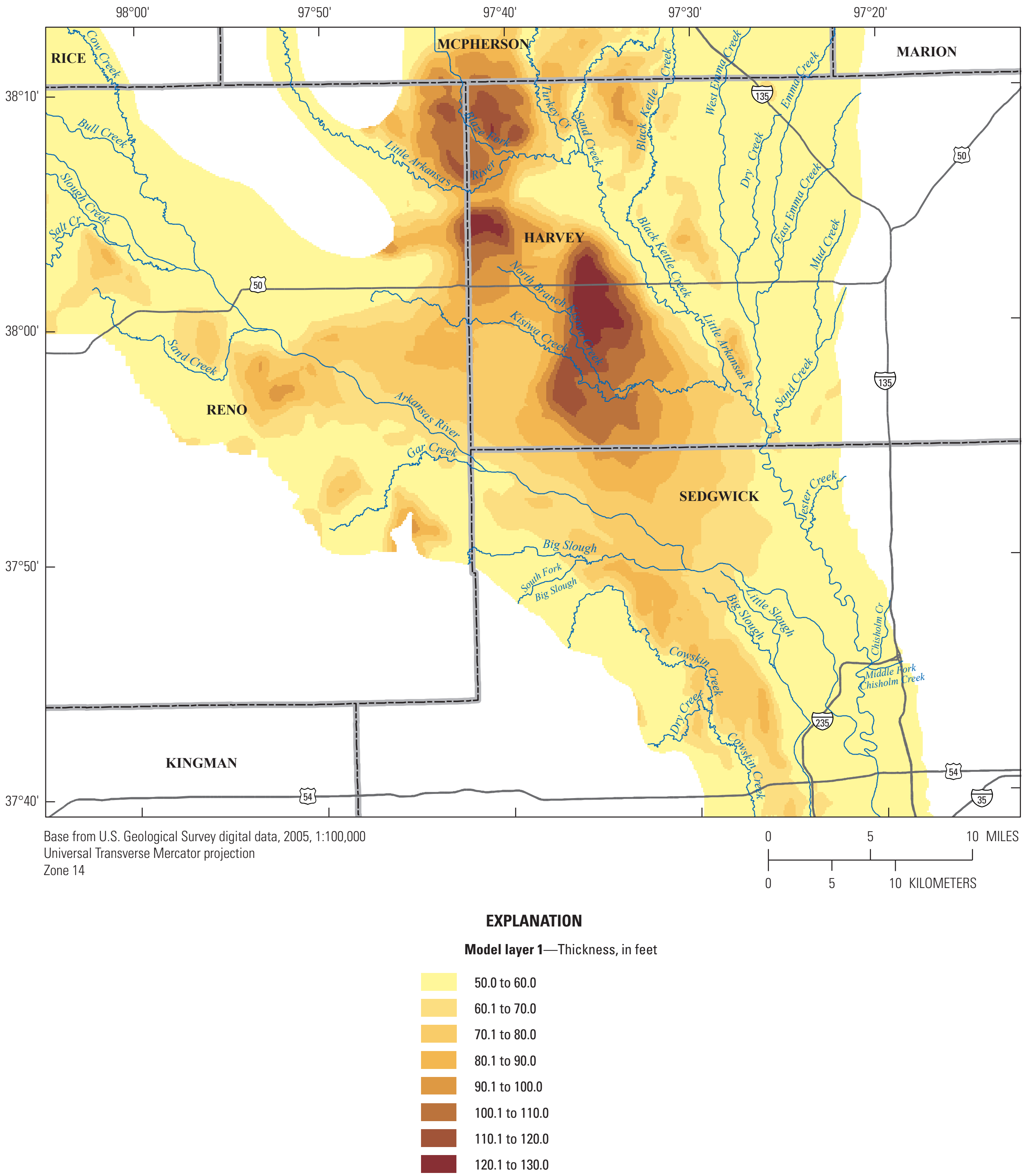



**Figure 14.** Active cells and thickness for model layer 1.

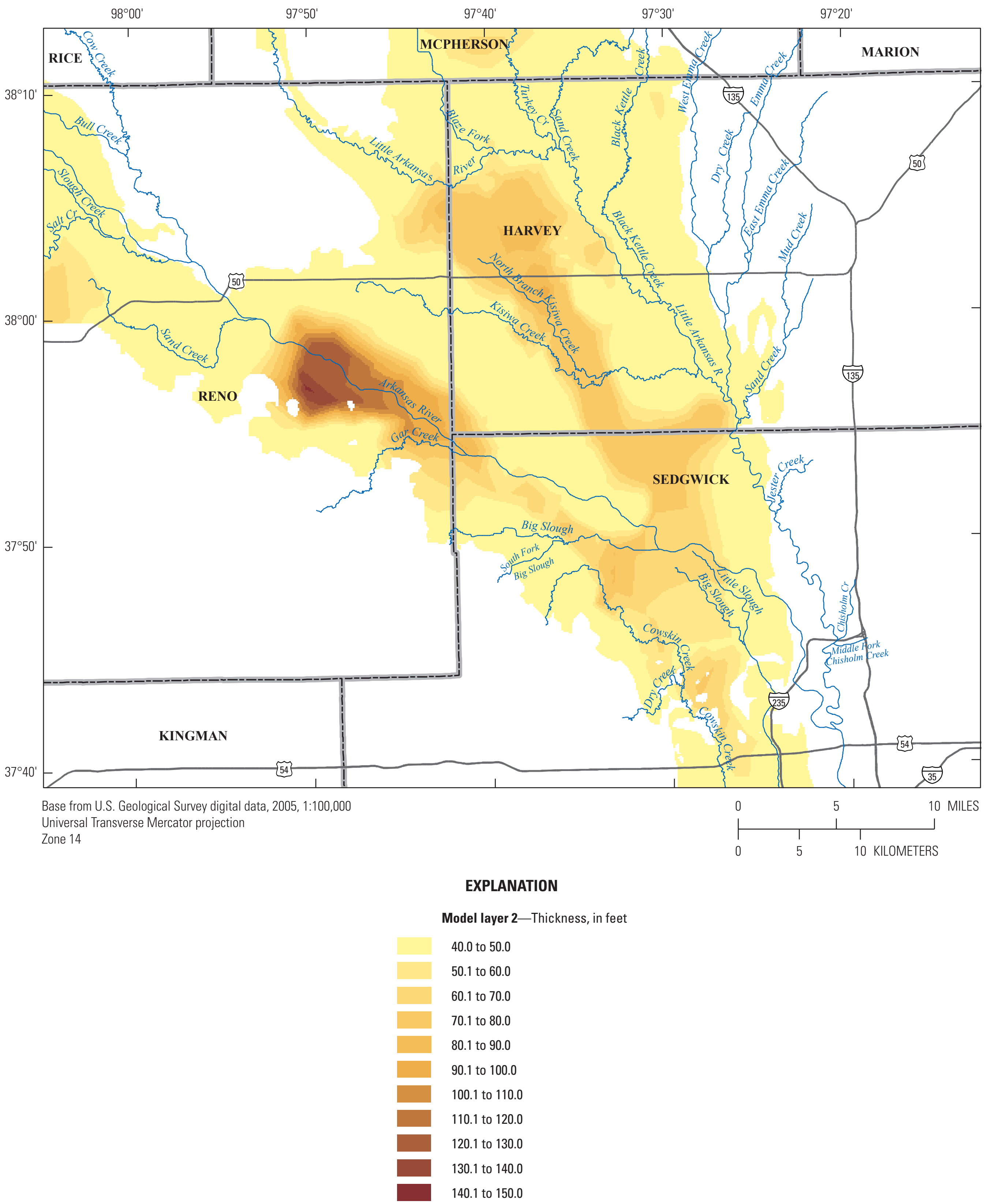
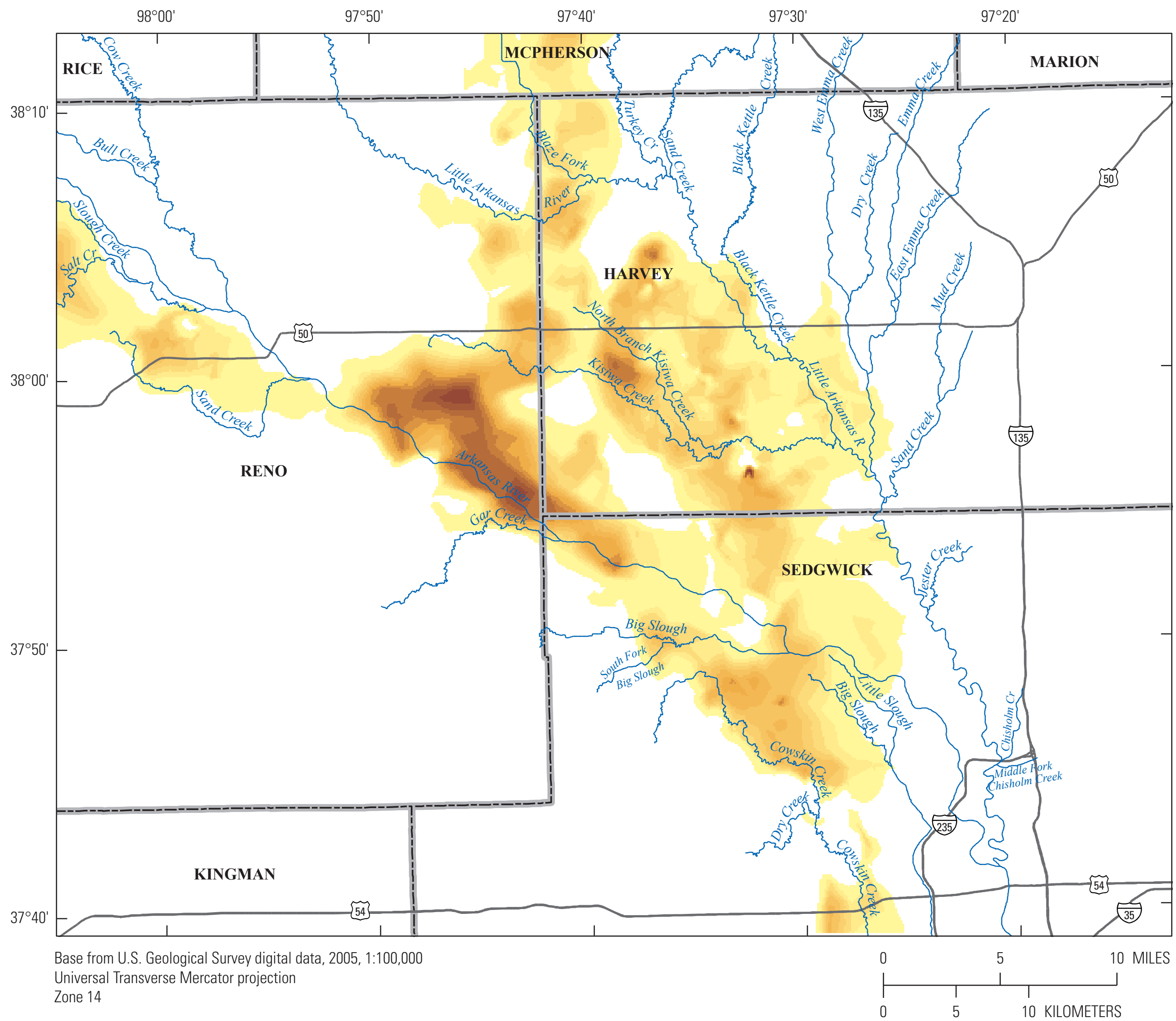



**Figure 15.** Active cells and thickness for model layer 2.




**EXPLANATION**

**Model layer 3**—Thickness, in feet

- 40.0 to 50.0
- 50.1 to 60.0
- 60.1 to 70.0
- 70.1 to 80.0
- 80.1 to 90.0
- 90.1 to 100.0
- 100.1 to 110.0
- 110.1 to 120.0
- 120.1 to 130.0
- 130.1 to 140.0
- 140.1 to 150.0
- 150.1 to 160.0
- 160.1 to 170.0

**Figure 16.** Active cells and thickness for model layer 3.

## Hydrogeologic Framework

A layer of finer grained deposits separate the upper and lower parts of the *Equus* Beds aquifer in the study area and is represented in model layer 2 (fig. 5). To determine the top and bottom of each model layer, the altitude where lithology changed between the coarser and finer grained material was identified in lithologic logs of wells and boreholes in the study area. These altitudes were then interpolated between well and borehole locations to define model layer top and bottom altitudes. Lithologic data also was used to determine depth to bedrock in the model area. Locations of lithologic data used to determine model layer top and bottom altitudes are shown in figure 17 (Kansas Geological Survey, 2009b). Model layer 1 corresponds to the upper part of the aquifer primarily composed of sand and gravel interbedded with clay and silt (fig. 5). Model layer 2 corresponds to the middle part of the aquifer where fine-grained deposits are more prevalent than in model layers 1 and 3. Model layer 3 corresponds to deep parts of the aquifer and has similar lithology to model layer 1. All three model layers are present in the area of interest around the Wichita well field (figs. 14–16).

## Model Input Variables

Values for the various hydraulic properties of the aquifer were assigned in the model using model input variables referred to as parameters in this report. Groups of cells in each model layer were assigned to zones and a uniform value of a hydraulic property, the parameter value, was assigned to each zone. The hydraulic properties include horizontal hydraulic conductivity, vertical hydraulic conductivity, specific storage, specific yield, recharge, evapotranspiration, streambed hydraulic conductivity, flow across the model boundaries, and well pumping. Lithologic descriptions recorded during the installation of wells and boreholes are the most numerous and have the greatest areal extent of all data types in the study area. The areal distribution of clay, silt, sand, and gravel within the aquifer was used to areally distribute values for horizontal and vertical hydraulic conductivity. Model cells with similar lithologic properties were grouped together and a separate value for horizontal or vertical hydraulic conductivity was assigned to each group. Specific yield and specific storage were uniformly distributed within each model layer. The areal distribution of soil permeability (Juracek, 2000) was used to areally distribute recharge rate as a percent of rainfall. Evapotranspiration was distributed uniformly across the model. Streambed hydraulic conductivity values were assigned to model cells that contained simulated streams or drains. The value for flow across the model boundaries was assigned to model cells on the edge of the model where simulated flow entered or exited the model. Well pumping was assigned to model cells that contained a simulated pumping well. Values for all parameters except well pumping were adjusted during manual steady-state and transient calibration of the model. Parameter names, hydraulic property, model layer and zone numbers, and final calibrated values, units, and comments are listed in table 2 at the back of this report.

## Hydraulic Properties

Numerous wells and boreholes with lithologic data are widely distributed within the study area (fig. 17). Initial values of hydraulic conductivity were based on the distribution of lithology within each model layer. The major lithologies consist of clay, silt, sand, and gravel. For each borehole or well location with lithologic data, a number was assigned to each major lithology type: clay equals 1, silt equals 2, sand equals 3, and gravel equals 4. At each location with lithologic data, the percent of the thickness of each lithology within a given model layer was calculated, multiplied by the corresponding lithologic value, and the result was summed. For example, for a location with 50 percent of the thickness as sand (3) and 50 percent of the thickness as gravel (4), the resulting composite lithologic value would be 3.5. This number was then used as a semi-quantitative composite lithologic value for the model layer at each location and used to interpolate model layer lithology between locations. The distribution of composite lithology is shown for each model layer in figures 18, 19, and 20.

The areal distribution of hydraulic conductivity values from previously calibrated groundwater-flow models of the *Equus* Beds aquifer (Myers and others, 1996; Spinazola and others, 1985) was compared to the composite lithologic distribution calculated for this study. Larger hydraulic conductivity values corresponded to areas with predominately coarser lithologies and smaller hydraulic conductivity values corresponded to areas with predominately finer lithologies. The initial distribution of hydraulic conductivity for each model layer was created by assigning the larger hydraulic conductivity values to the coarser lithologies and the smaller values to the finer lithologies.

The final distribution of hydraulic conductivity for each model layer is shown in figures 21, 22, and 23. Hydraulic conductivity ranges from 0.25 to 1,200 ft/day in model layer 1; from 5 to 600 ft/day in model layer 2; and from 10 to 800 ft/day in model layer 3. Hydraulic conductivity was reduced in the dune sand area in model layer 1 to more accurately simulate groundwater levels in the low permeability sand dune deposits. Initial hydraulic conductivity values were altered during calibration to more closely match observed and simulated groundwater levels.

The simulated flow of water between model cells in adjacent model layers is controlled by the vertical conductance term. Vertical conductance, or leakance, is calculated within MODFLOW from the thickness of each model layer between model nodes (the center of the cell to the edge of the cell in the vertical direction) and the vertical hydraulic conductivity of each model layer (McDonald and Harbaugh, 1988). The vertical conductance terms between cells of adjacent model layers simulate the presence of vertical anisotropy caused by interbedding of clay, silt, and fine sand deposits.

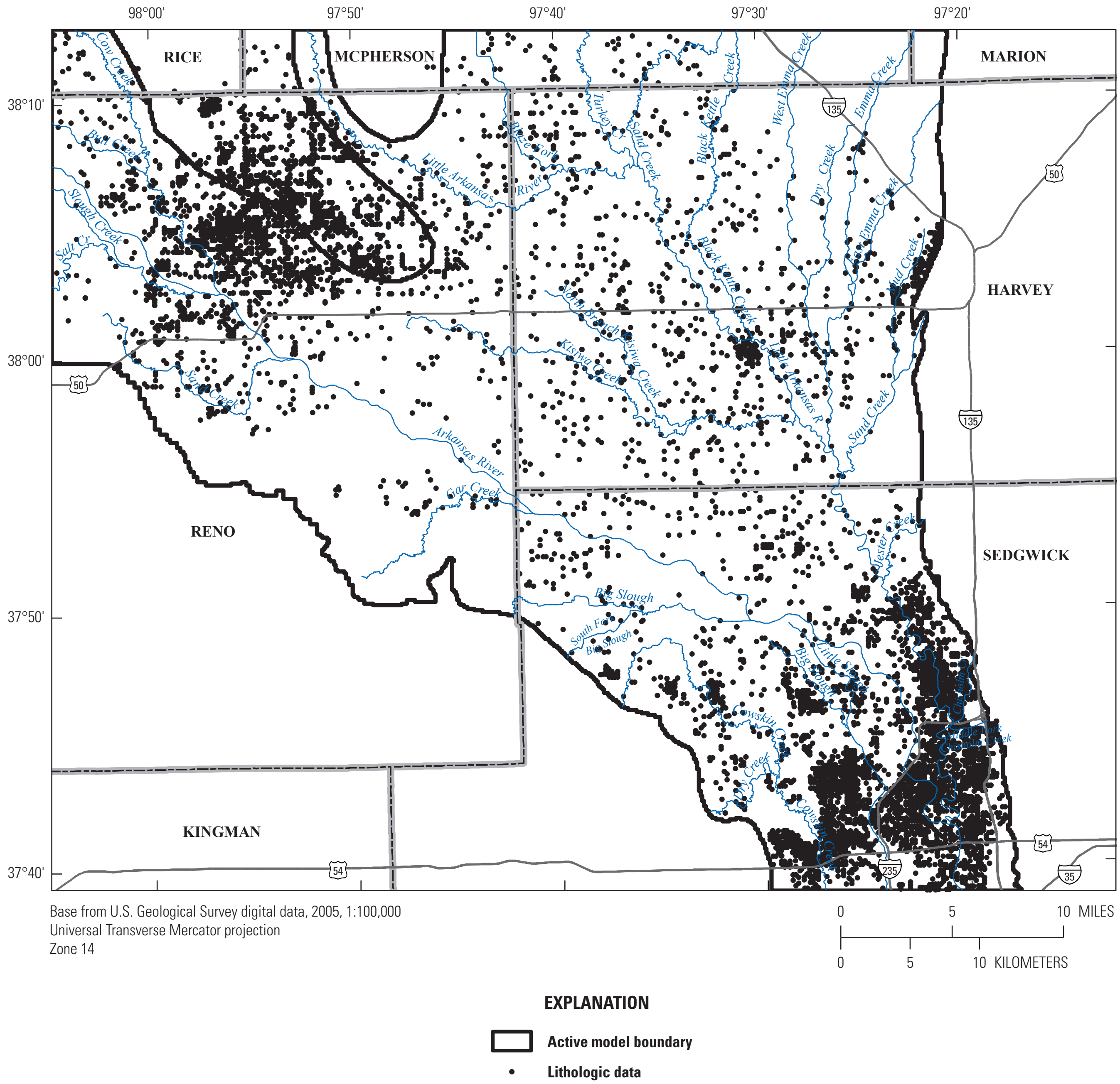



**Figure 17.** Location of lithologic data used to determine model layer top and bottom altitudes and distribution of lithology within each model layer.

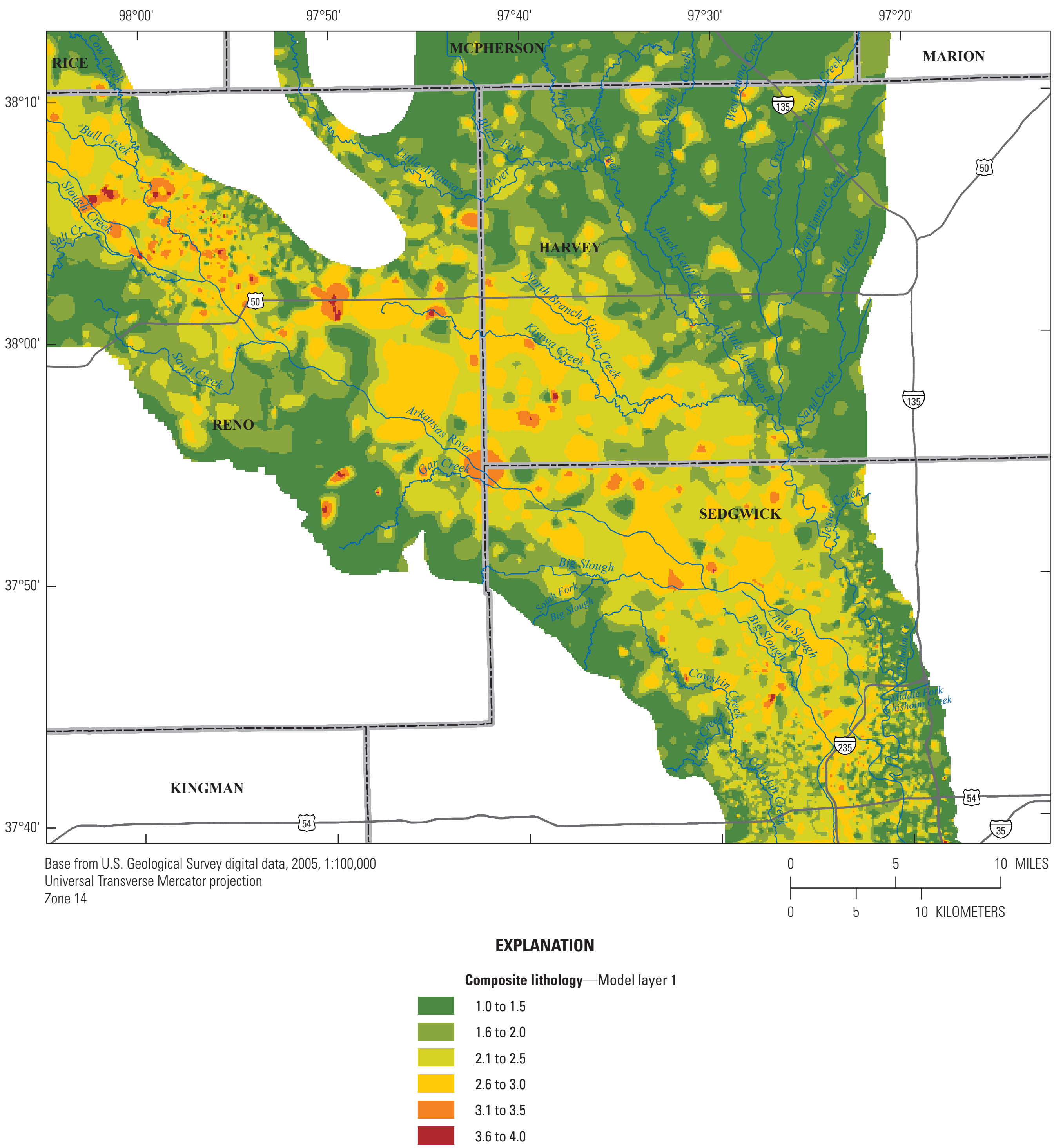



**Figure 18.** Distribution of composite lithologic value for model layer 1.

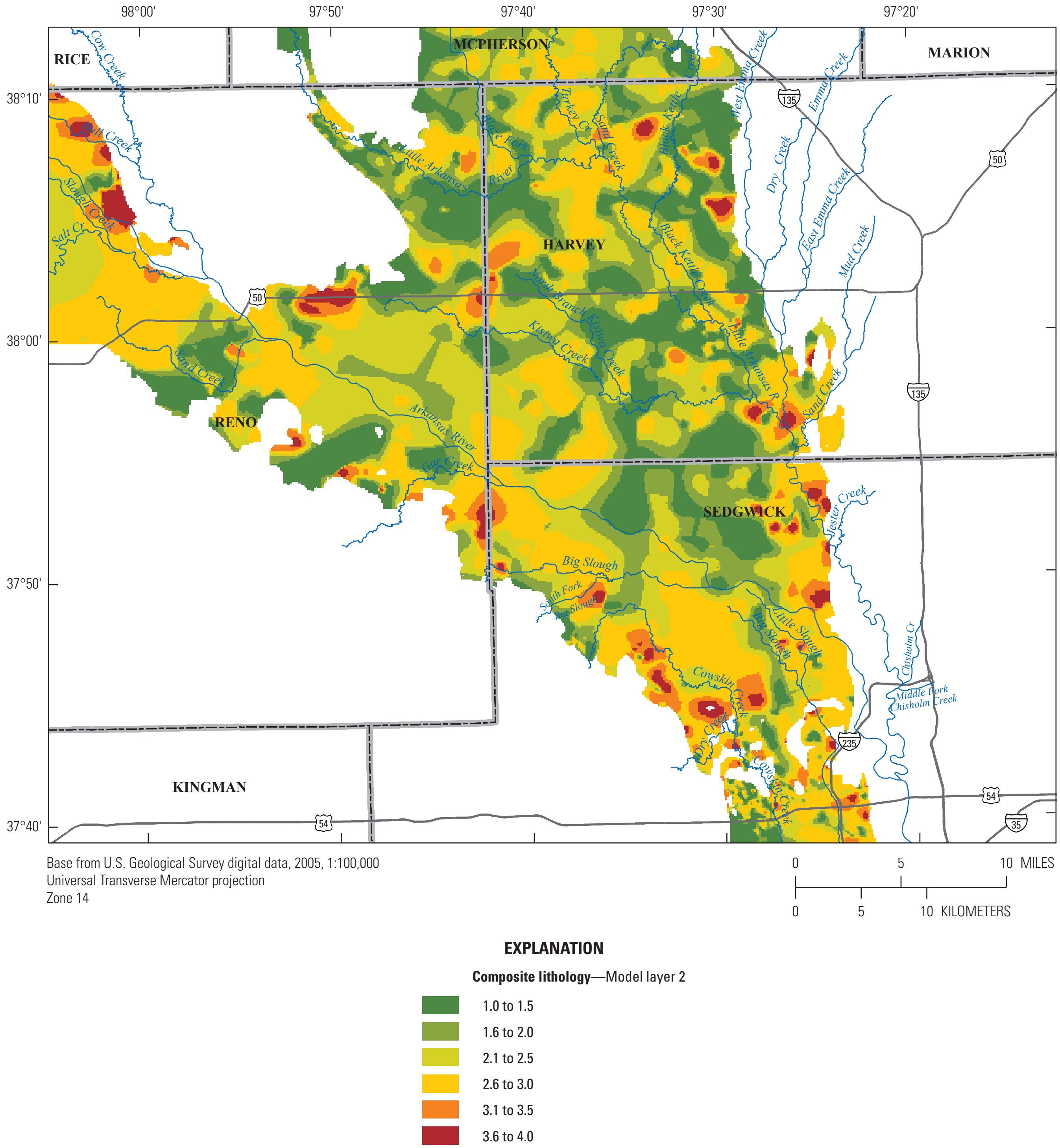



**Figure 19.** Distribution of composite lithologic value for model layer 2.

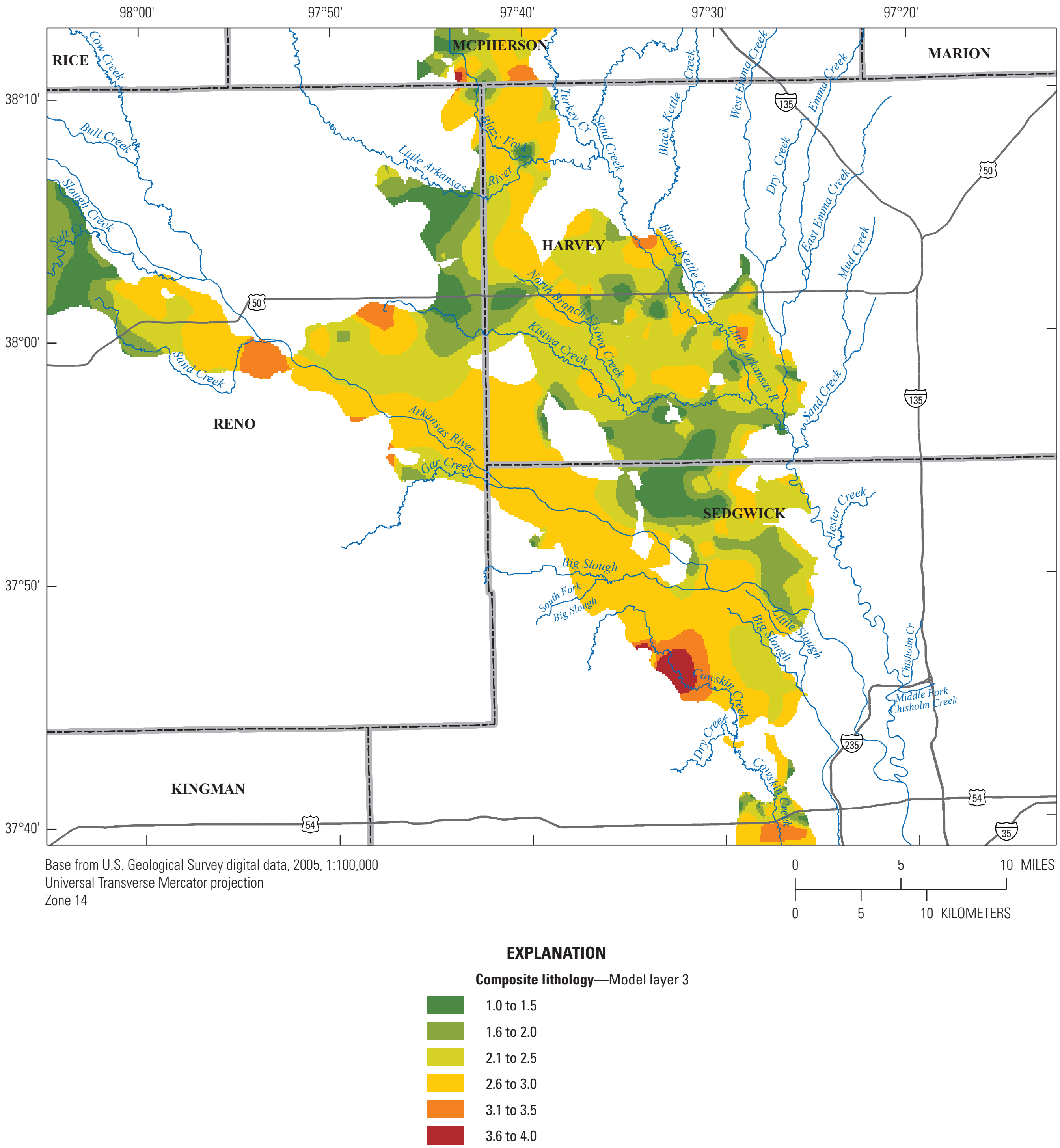
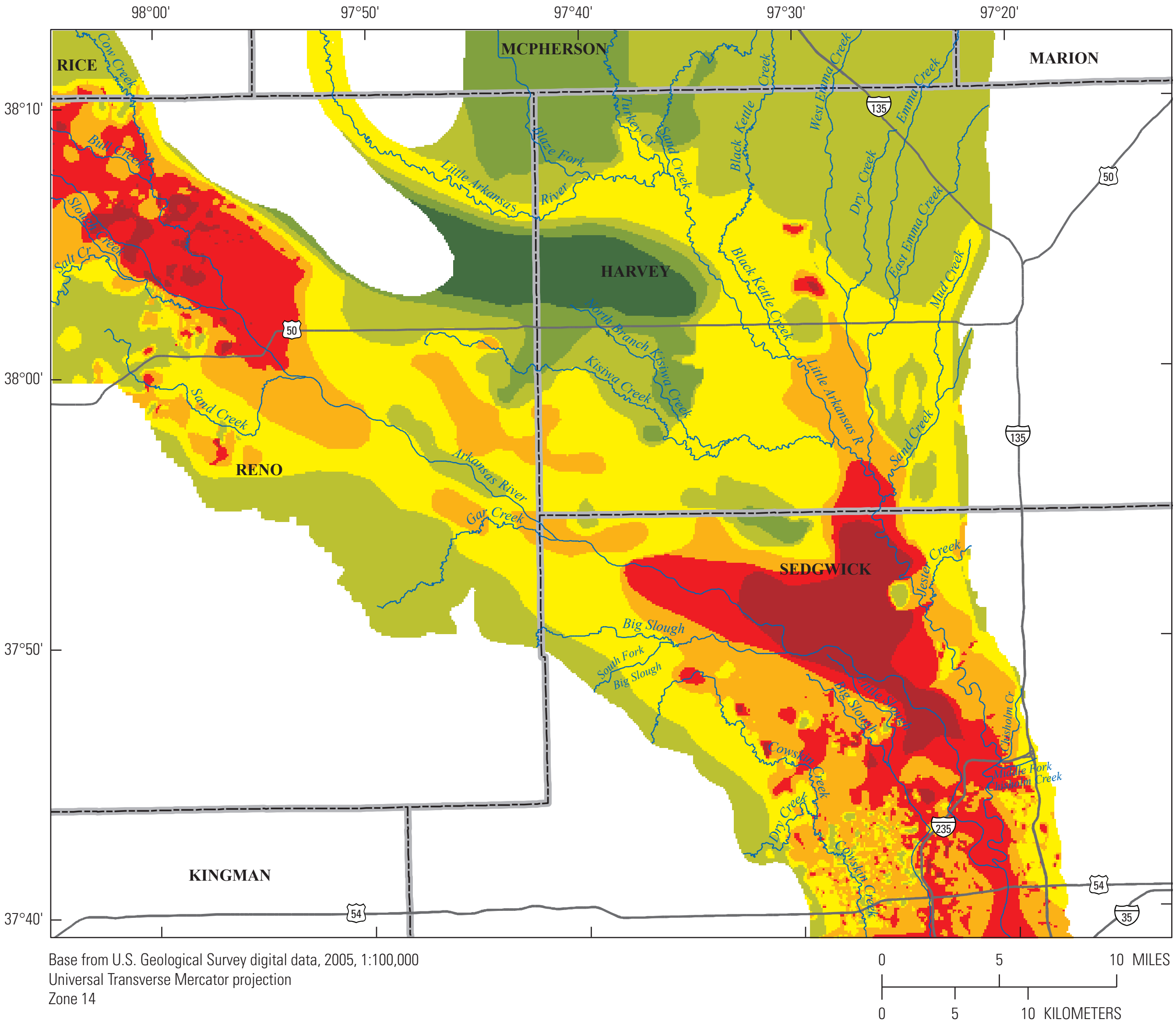



**Figure 20.** Distribution of composite lithologic value for model layer 3.




**EXPLANATION**

**Hydraulic conductivity**—Model layer 1, in feet per day and parameter name

- 0.25, L1-Z1
- 1, L1-Z2
- 40, L1-Z3
- 220, L1-Z4
- 600, L1-Z5
- 800, L1-Z6
- 1,200, L1-Z7

**Figure 21.** Distribution of hydraulic conductivity for model layer 1.

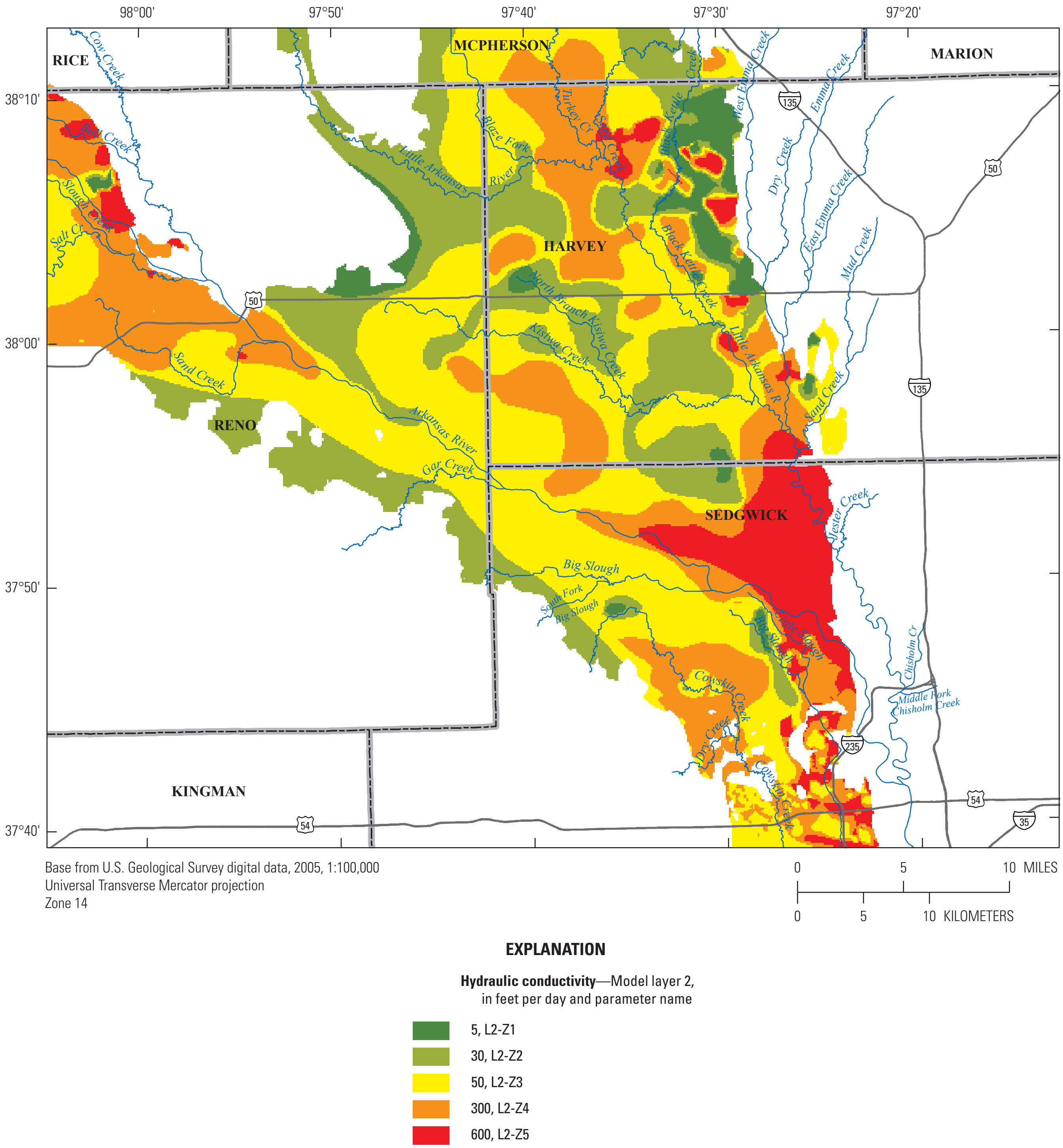



**Figure 22.** Distribution of hydraulic conductivity for model layer 2.

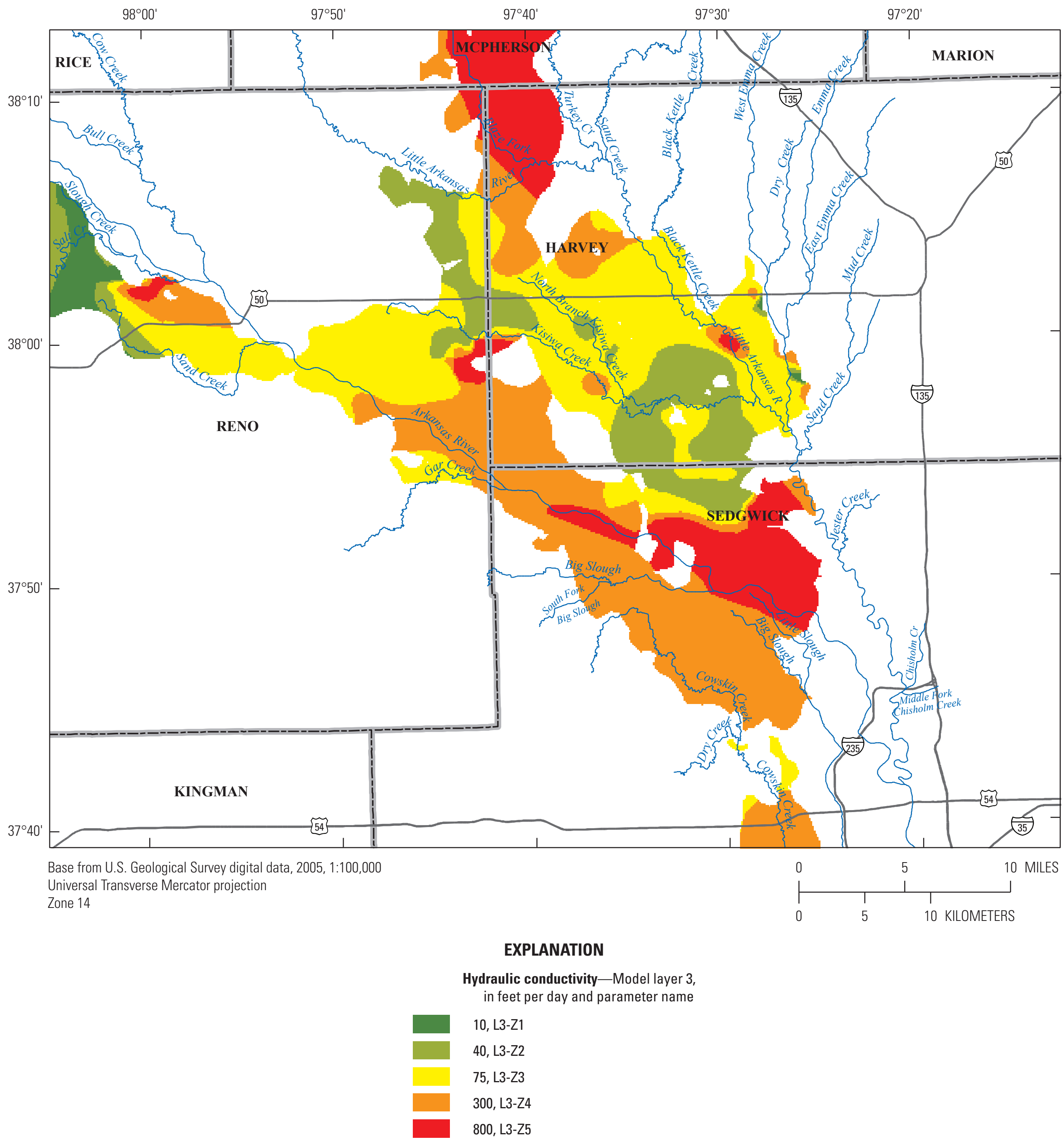



**Figure 23.** Distribution of hydraulic conductivity for model layer 3.

The ratio of horizontal hydraulic conductivity to vertical hydraulic conductivity within model layer 1 ranges from 10 to 500 and is shown in figure 24. Larger values indicate smaller vertical hydraulic conductivity. Small vertical hydraulic conductivity values were assigned to account for vertical anisotropy caused by thin layers of clay, silt, and fine-grained sand in parts of the study area (Myers and others, 1996). The ratio of horizontal hydraulic conductivity to vertical hydraulic conductivity between adjacent cells within model layers 2 and 3 was set at 10 to account for vertical anisotropy caused by thin layers of clay, silt, and fine-grained sand.

A specific yield of 0.15 was used for model layers 1, 2, and 3 to represent conditions where water is released from storage as water drains from the aquifer. A storage coefficient of 0.0005 was used for model layers 1, 2, and 3 to represent conditions where water is released from storage because of expansion of the water or compaction of the aquifer material and not actual drainage of water from the aquifer. All model layers were defined in MODFLOW-2000 as convertible and each required a specific yield and a confined storage coefficient value as model input.

## Boundary Conditions

Model boundary conditions are used to specify flow into and out of the model domain. Sources of flow into and out of the aquifer include recharge, evapotranspiration, gaining and losing streams, pumping wells, and artificial recharge wells and basins. The groundwater-flow model simulates the water table as a free surface, where its position is not fixed but varies with time (Franke and others, 1984). Specified flux boundaries, where the volume of water that flows across the boundary is a function of time, position, and head, and varies as a function of flow, include the lateral boundary of the *Equus* Beds aquifer, bedrock (no flow boundaries), and recharge from precipitation. Head-dependent flux boundaries where water flow varies as a function of head and conductance include flow across lateral boundaries of the model, evapotranspiration, gaining and losing streams, pumping wells, and artificial recharge wells and basins.

### Recharge

The water table is the surface across which areally distributed recharge enters the aquifer. Recharge to the model was applied to the top-most active cell in each vertical column and varied temporally as a function of average precipitation for each stress period and spatially as a percentage of precipitation.

Annual precipitation data for 1938 through 2008 for six Cooperative and Weather Bureau Army Navy (WBAN) weather stations in and near the study area were used to estimate the precipitation for the study area. Average precipitation for each stress period and periods of data from weather stations used in the model are listed in table 3 at the back of this report. Average annual precipitation for weather stations near the Wichita well field is shown in figure 25. Average precipitation calculated from weather stations was evenly distributed across the model for each stress period.

The areal distribution of soil permeability (Juracek, 2000) was used for the initial distribution of recharge rate as a percent of rainfall. Soil permeability was divided into six groups shown in figure 26. Soils with low permeability were assigned small values of recharge as a percent of precipitation and soils with large permeability were assigned large values. The initial distribution of recharge as a percent of precipitation was altered during the course of model calibration to more closely match simulated and observed groundwater levels. The final distribution of recharge as a percentage of precipitation for each recharge zone is shown in figure 27.

### Evapotranspiration

Evapotranspiration is simulated in the model as removal of water from the saturated aquifer through plant transpiration and evaporation. Evapotranspiration is set to a maximum rate when the water table is at land surface and is set to zero (extinction depth) when the water table is more than a specified depth below the land surface (set at 10 ft). Evapotranspiration varies linearly with changes in the water table between the two surfaces. Maximum average evapotranspiration was calculated for each stress period using the Hamon equation (Hamon, 1961; Alkaeed and others, 2006). The Hamon equation uses only saturated vapor pressure, mean daily air temperature, and average number of daylight hours per day as input. Evapotranspiration was estimated for 1935 through 2008 using mean monthly air temperature and saturation vapor pressure from the Cooperative Weather Station at Newton, Kans. (station 145744). Daily values of maximum evapotranspiration were used to calculate evapotranspiration for each stress period in feet/day. The Hamon equation is:

$$ET_o = \frac{2.1 H_t^2 e_s}{\left(T_{mean} + 273.2\right)} \tag{2}$$

where

$ET_o$ is the evapotranspiration for the stress period,
$H_t$ is the average number of daylight hours per day for the stress period,
$e_s$ is the saturation vapor pressure in millimeters per day at the mean daily air temperature for the stress period, and
$T_{mean}$ is the mean daily air temperature (ºC) for the stress period.

and

$$e_s = 6.112 \cdot \exp[17.67 \cdot (T)/(T+243.5)] \tag{3}$$

where

$T$ is the mean daily air temperature for the stress period (Rogers and Yau, 1989).

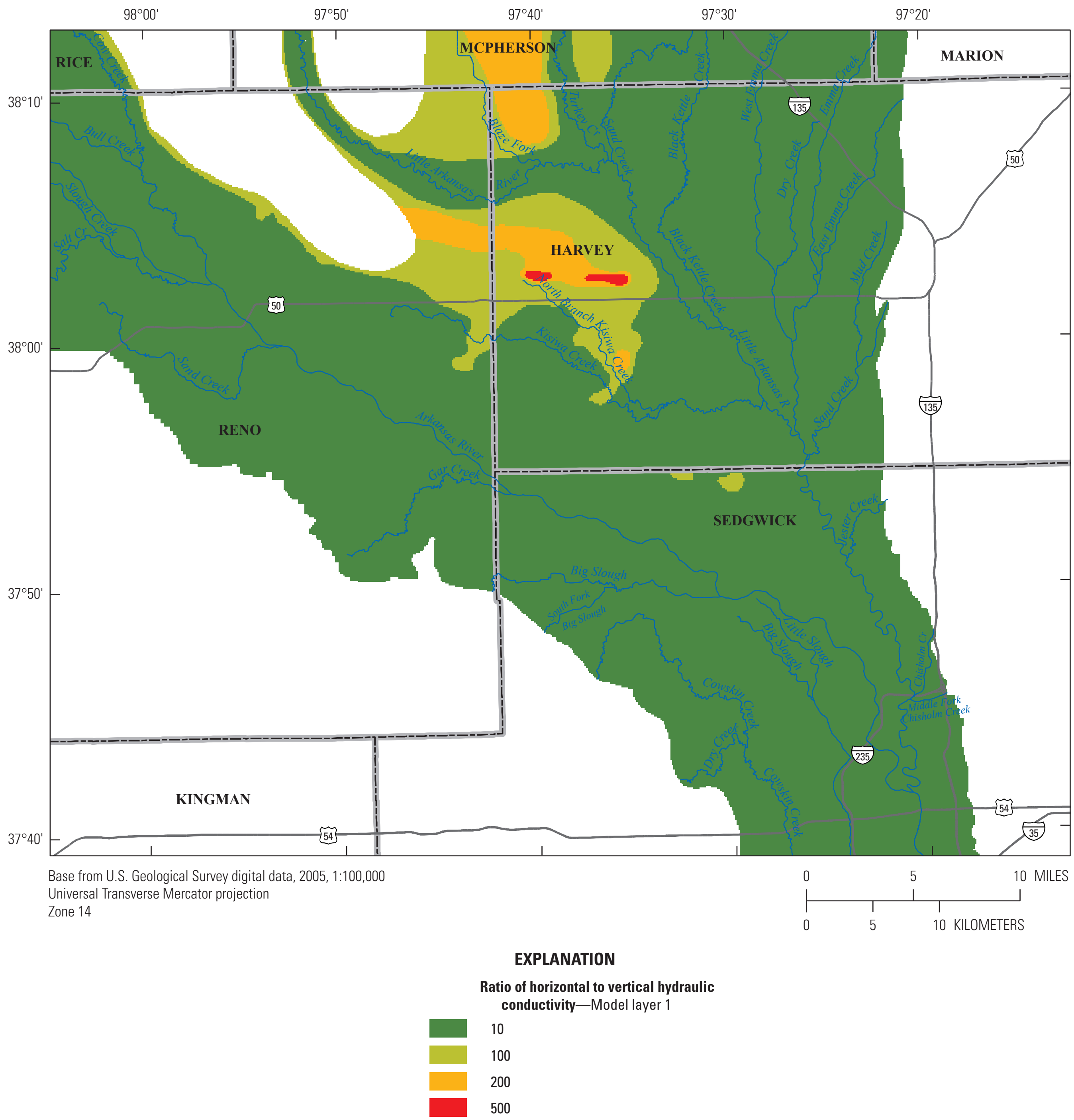



**Figure 24.** Distribution of the ratio of horizontal to vertical hydraulic conductivity for model layer 1.

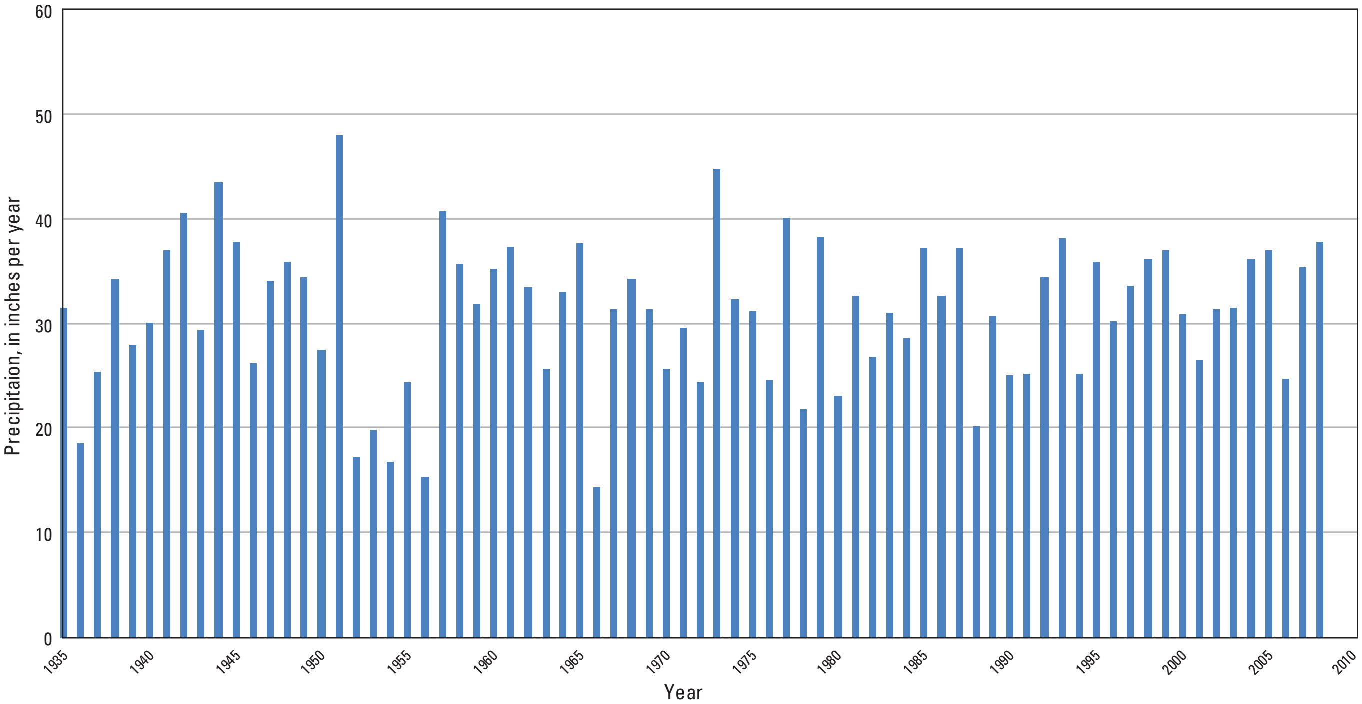



**Figure 25.** Average annual precipitation in inches per year for weather stations at Hutchinson (143930), Mt. Hope (145539), Newton (145744), Sedgwick and Halstead( 143366) and Wichita (148830) near the Wichita well field (National Oceanic and Atmospheric Administration, 2008).

The average number of daylight hours per day was calculated for a point near the center of the model area, at Halstead, Kans. (longitude 97°31′00″ W, latitude 38°00′00″ N). Sunrise and sunset times were obtained from the Astronomical Applications Department, U.S. Naval Observatory (*http://aa.usno.navy.mil/data/docs/RS_OneYear.php*). Temperature data from Newton, Kans. were used for the computation because a complete record was available from National Climate Data Center Archives (*http://lwf.ncdc.noaa.gov/oa/climate/stationlocator.html)*. Initial estimated evapotranspiration was altered during calibration to more closely match observed and simulated groundwater levels.

## Streams

The Arkansas River, Little Arkansas River, and their tributaries are represented in the model as head-dependent flux boundaries. The Arkansas River, Little Arkansas River, and Cow Creek (near Hutchinson, Kans.) were simulated in MODFLOW-2000 using the River Package and the smaller streams and tributaries were simulated using the Drain Package (McDonald and Harbaugh, 1988). All rivers and drains are within model layer 1.

Flow into or out of the aquifer at each of the cells where a river is simulated is a function of the river stage with respect to the altitude of the potentiometric surface, the hydraulic conductivity of the streambed material, the cross-sectional area of flow between the stream and the aquifer, and the altitude of the water table with respect to the altitude of the streambed (McDonald and Harbaugh, 1988). Stream stages in the Arkansas and Little Arkansas Rivers were recorded at streamflow gages (fig. 6) hourly (U.S. Geological Survey, 2009a) and average annual stage was calculated for each gage. The average annual altitude of the river surface used in each stress period of each simulation was assigned to each model cell with a stream by interpolating the specified river surface altitude between gaging stations. Each stream was assigned a single value for streambed hydraulic conductivity. The area of the stream within each model cell was calculated and the streambed hydraulic conductivity value was multiplied by the area of the stream and then divided by the thickness of the streambed to determine the streambed conductance. Streambed thicknesses are unknown and were assigned an arbitrary value of 1 ft. Initial streambed conductances were altered during calibration to more closely match observed and simulated flow between the streams and the aquifer.

Flow into or out of the aquifer at each of the cells where a drain is simulated is a function of the altitude of the potentiometric surface, the hydraulic conductivity of the drain bed material, the cross-sectional area of flow between the drain and the aquifer, and the altitude of the water table with respect to the altitude of the drain bed (McDonald and Harbaugh, 1988). Each stream simulated as a drain was assigned a single value for streambed hydraulic conductivity. The area of the stream within each model cell was calculated and the initial streambed hydraulic conductivity value was multiplied by the area of the stream and then divided by the thickness of the streambed to determine the streambed conductance.

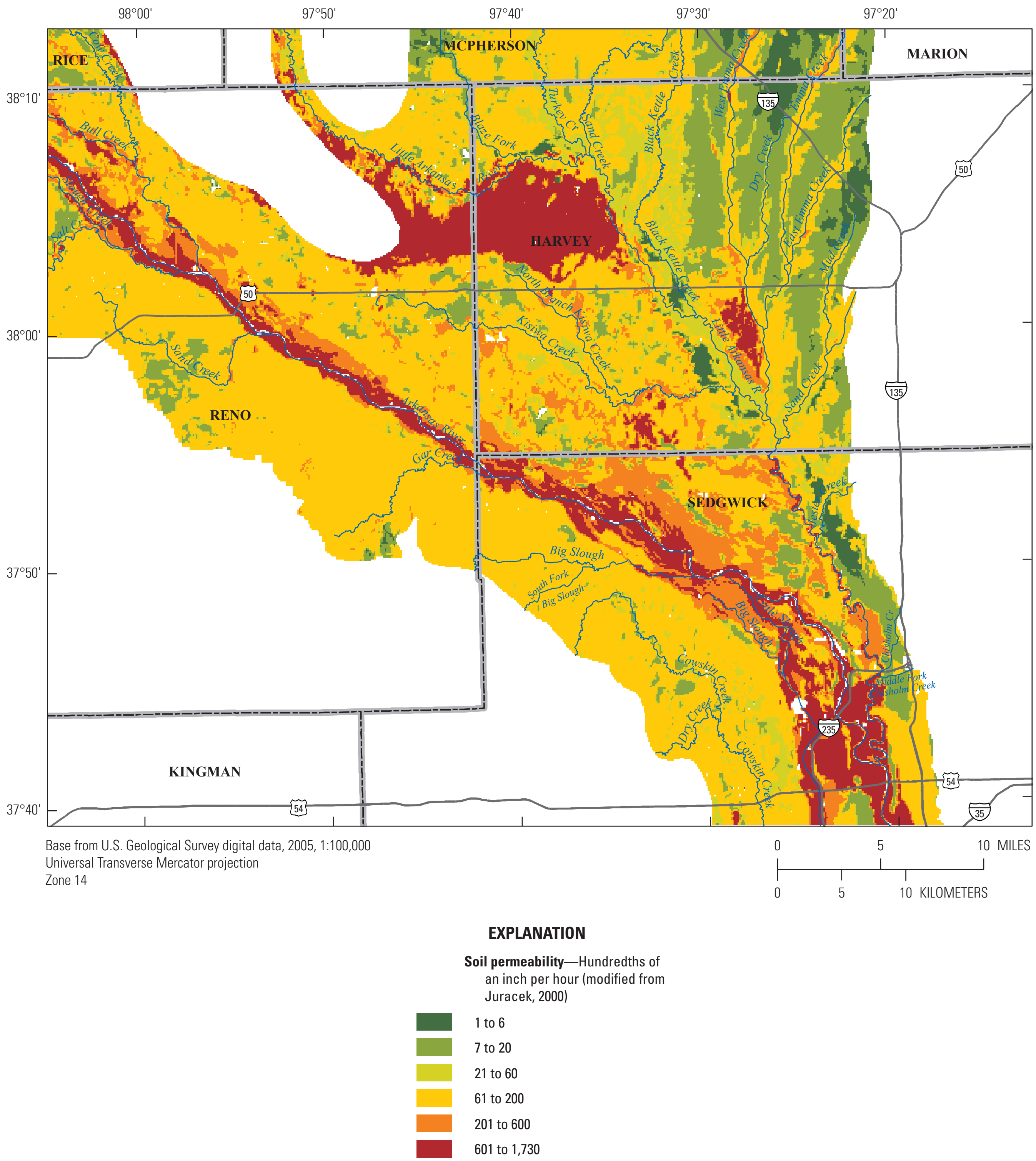



**Figure 26.** Soil permeability.

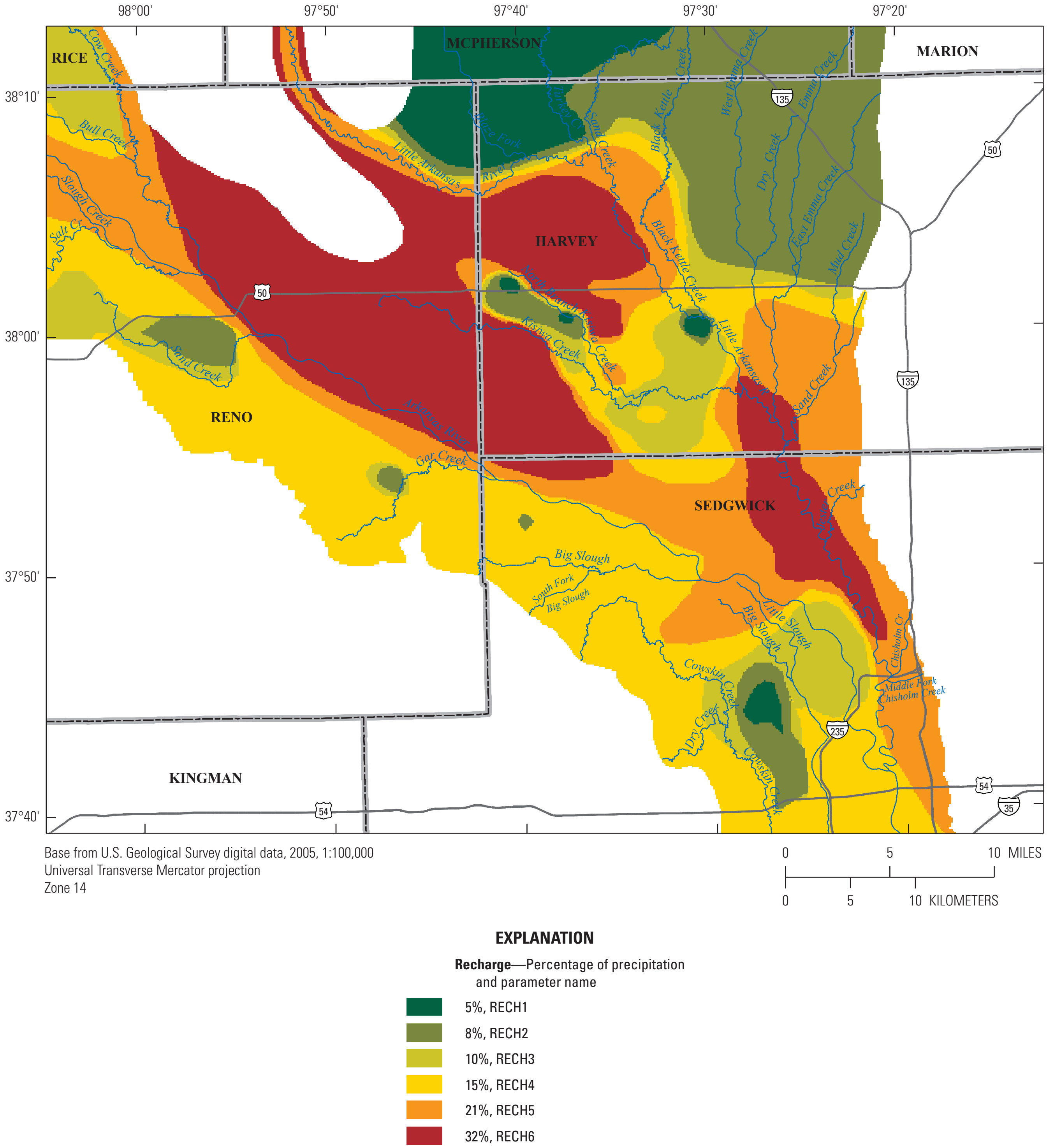



**Figure 27.** Recharge as a percent of precipitation.

Streambed thicknesses are unknown and were assigned an arbitrary value of 1 ft. Initial streambed conductances for drains were altered during calibration to more closely match observed and simulated flow from the aquifer to the drains. Simulated rivers and drains are shown in figure 28.

### Wells

Pumping wells are internal boundaries of the model where water was removed at a specified rate equal to the discharge of each well. The total volume of water withdrawn annually from the aquifer by pumping from irrigation, production, and industrial wells was obtained from each water supplier when available or from the KDA-DWR Water Rights Information System database (Kansas Department of Agriculture–Division of Water Resources, unpub. data, 2009). The depth of each pumping well was based on the screened interval, when known, or the depth of the well. The Multinode Well Package was used to simulate all industrial, irrigation and production well pumping (Harbaugh and others, 2000). The MultiNode Well Package vertically distributes pumping between model layers from each well based on the top and bottom altitudes of the screened interval and the hydraulic properties of each model layer.

Groundwater pumpage data for 1935 to 1979 were obtained from Spinazola and others (1985) and Myers and others (1996). Groundwater pumpage for the stress periods from 1935 through 1979 was distributed in the model based on the spatial and temporal distribution of pumping in Spinazola and others (1985). The model cells from Spinazola and others (1985) are 1 mile on each side and pumping was assigned to the center of each cell. Pumping wells were placed in the current model to coincide with the center of each cell in the model from Spinazola and others (1985). Pumping was distributed vertically across all model layers by using the MultiNode Well Package. Locations of simulated pumping wells for 1935 through 1979 are shown in figure 29.

Annual groundwater pumpage data for industrial, irrigation, and production wells in the study area for 1988 through 2008 were obtained from the KDA-DWR (Kelly Emmons, Kansas Department of Agriculture, Division of Water Resources, written commun., June 5, 2009, and August 31, 2009). Groundwater pumpage for the stress period from 1980 through 1989 was distributed in the model using well locations and pumping rates from 1989. Groundwater pumpage for the stress periods from 1990 through 2008 was distributed in the model using well locations and average annual pumping rates.

Monthly pumpage data for Wichita's production wells for 1990 through 1993 and 1995 through 2008 (Megan Schmeltz, city of Wichita, written commun., September 25, 2009) and monthly artificial recharge data for phase I ASR sites for 2007 through 2008 were obtained from the city of Wichita (U.S. Geological Survey, 2011). Monthly pumping rates were used to calculate an annual rate used for the Wichita wells. The city of Wichita also provided annual artificial-recharge data for 2002 through 2005 for the *Equus* Beds Groundwater Recharge Demonstration sites (U.S. Geological Survey, 2011). Locations for Wichita's production wells and the phase I ASR artificial-recharge wells were provided by KDA-DWR. Locations of the *Equus* Beds Artificial-recharge Demonstration Project recharge sites were those previously determined by the USGS.

The pumping wells and artificial-recharge sites were assigned to the model-grid cell they plotted within based on decimal-degree locations provided by KDA-DWR. Each well was evaluated individually to determine the altitude of the bottom of the well and the screened interval. The top and bottom altitudes were used in the MultiNode Well Package to vertically distribute well pumping across model layers for each well.

The depth of each well was determined using one of the following methods. Where data were available, the altitude of the bottom of the screened interval was used. For unknown screened intervals, the altitude of the bottom of the well was used. If the altitude of the bottom of the screened interval or depth of the well were unknown, and aquifer information provided by KDA-DWR indicated the well was in the *Equus* Beds aquifer, well depth was assigned as the depth of the lowest model layer in the cell that contained the well. If the well was not in the *Equus* Beds aquifer, it was excluded from use.

The top of the screened interval for each well was determined using one of the following methods. If the screened interval was known, the top altitude was used. If the screened interval was unknown, the top of the screened interval was arbitrarily set at 20 ft below land surface. For shallow pumping wells located in model layer 1, the top of the screened interval was arbitrarily set at 10 ft below land surface. Locations of simulated pumping wells for 1980 through 2008 are shown in figure 30.

Industrial pumpage was assumed to be at a constant rate throughout the year. The annual volume of pumpage divided by the number of days in the year was used to calculate a pumpage rate in cubic feet per day.

Two modifications were made to the annual irrigation pumpage data obtained from KDA-DWR. Irrigation pumpage that was unmetered (pumpage reported as the number of hours the pump ran multiplied by a pump rate) was considered over-reported and was reduced by varying annual percentages. Comparisons of pumpage at selected wells before and after metering indicated that unmetered pumpage was over-reported by about 20 percent before 1990 (Andy Lyon, Kansas Department of Agriculture, Division of Water Resources, written commun., July 2010). The KWO estimates the percentage by which the annual reported unmetered irrigation water is greater than actual irrigation (Kansas Water Office and Kansas State Board of Agriculture, Division of Water Resources, 1989) using the following equation:

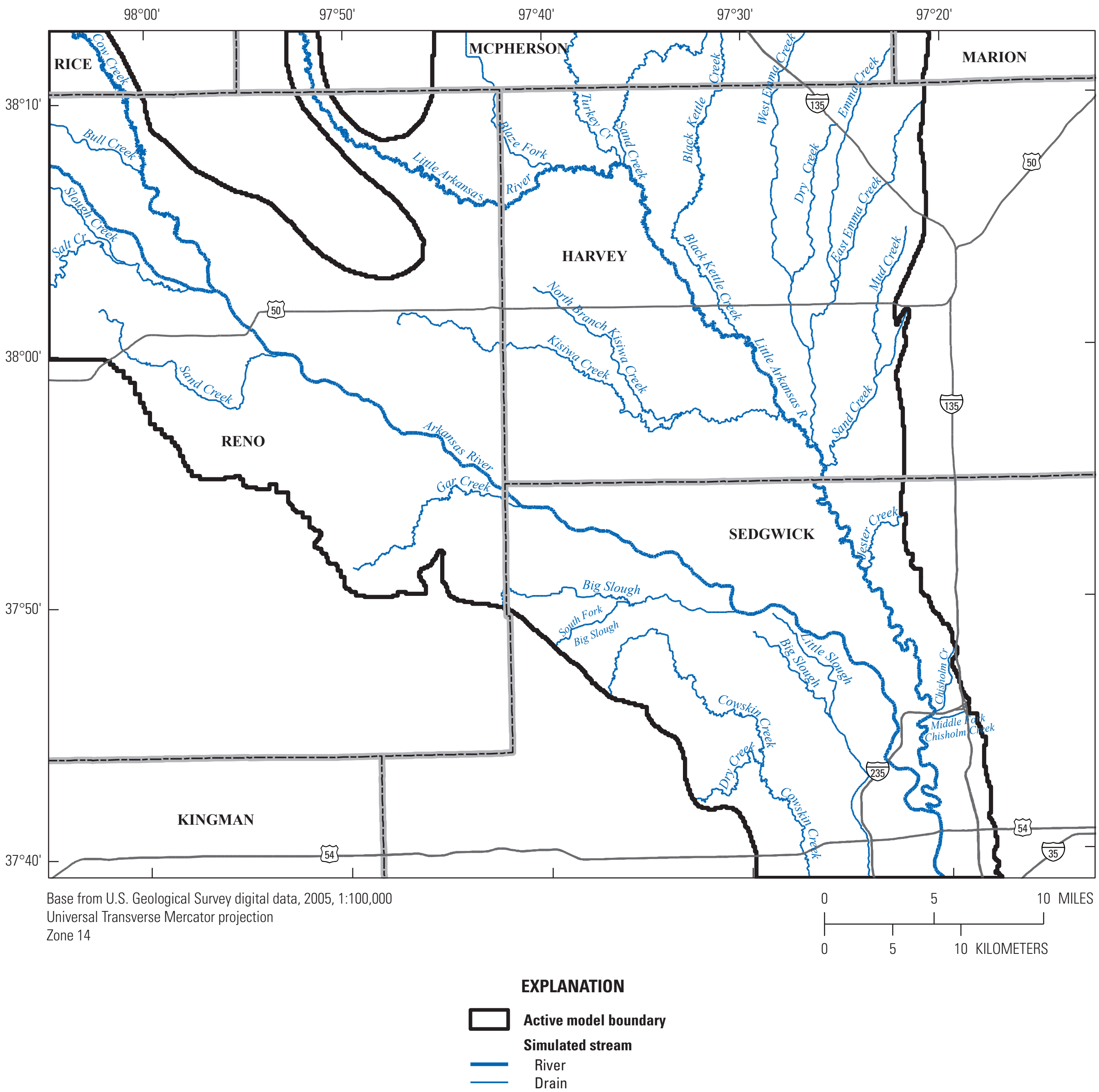



**Figure 28.** Streams simulated as rivers or drains.

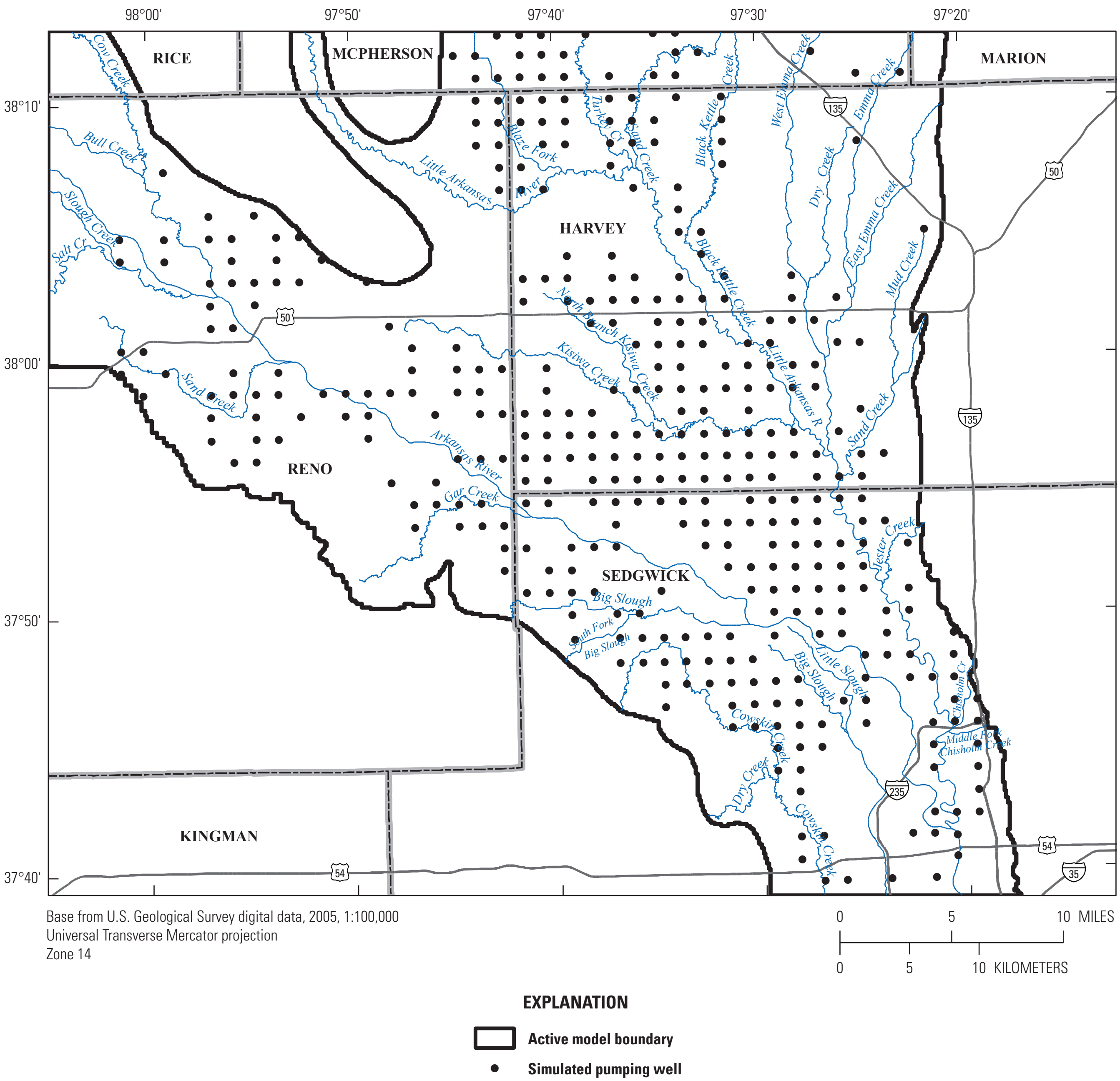



**Figure 29.** Simulated pumping wells, 1935 through 1979.

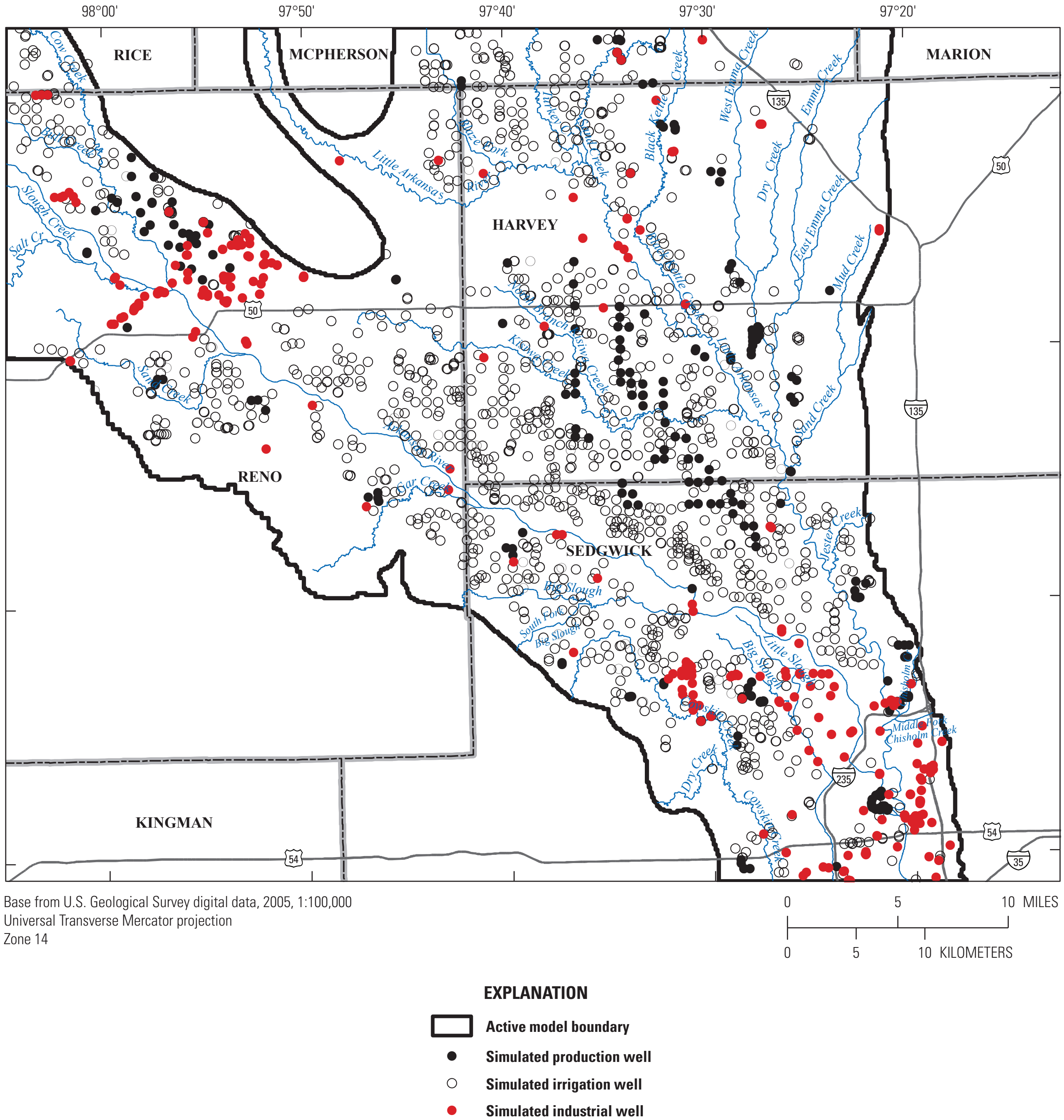



**Figure 30.** Simulated pumping wells, 1980 through 2008.

$$\text{Percent of unmetered irrigation over-reported} = 100x[((Viu/Aiu) - (Vim/Aim))xAiu]/Viu \quad (4)$$

where

*Viu* = volume of unmetered irrigation water,
*Aiu* = area irrigated by unmetered irrigation water,
*Vim* = volume of metered irrigation water, and
*Aim* = area irrigated by metered irrigation water.

Using this method for the years 1989 through 2008, the amount by which the unmetered irrigation pumpage was estimated to be over-reported within the model area is listed in table 4.

**Table 4.** Estimated over-reporting of unmetered irrigation pumpage by year.

| Year | Estimated overreporting of unmetered irrigation pumpage, in percent | Year | Estimated overreporting of unmetered irrigation pumpage, in percent |
|---|---|---|---|
| 1989 | 17.17 | 1999 | 17.18 |
| 1990 | 1.33 | 2000 | 3.52 |
| 1991 | 5.19 | 2001 | 1.90 |
| 1992 | 24.20 | 2002 | 6.01 |
| 1993 | 22.71 | 2003 | 0 |
| 1994 | 6.96 | 2004 | 18.31 |
| 1995 | 11.03 | 2005 | 8.81 |
| 1996 | 17.55 | 2006 | 0.34 |
| 1997 | 27.29 | 2007 | 16.73 |
| 1998 | 11.27 | 2008 | 19.72 |

For the multiyear stress periods simulating 1935 through 1979, the amount of irrigation water that was over-reported was estimated at 20 percent. For the stress period from 1980 through 1989, the 1989 value (17.17 percent) was used.

Irrigation return flow is the part of the applied irrigation pumpage that is not consumed and recharges the aquifer. For the preparation of data for the model, the amount of irrigation return flow was considered to vary by the type of irrigation system used. Estimated return flow by system type is the same as that estimated by the KGS and used by KDA-DWR for the Middle Arkansas River Basin model (Andrew Lyon, Kansas Department of Agriculture, Division of Water Resources, written commun., July 2010). These percentages are similar to those reported in the Irrigation Guide for Kansas (National Resources Conservation Service, 2006). The irrigation system types used by KDA-DWR were grouped into the return-flow groups and are listed in table 5.

Since 1991, the percent of irrigation in the model area that was assigned to flood, center pivot-impact, and center pivot-LDN (low-impact drop nozzle) was estimated using a modification of a method used by KDA-DWR for their Middle Arkansas River Basin model (Andrew Lyon, Kansas Department of Agriculture, Division of Water Resources, written commun., July 2010). This method used data from a study done by Kansas State University that estimated the percentage of acres irrigated by gravity (flood), sprinkler, and microirrigation (drip) methods for the years 1970 through 2000 (Lamm and Brown, 2004). A ratio of the acres irrigated using sprinkler methods divided by the acres irrigated using gravity methods was computed for each year for the entire State of Kansas. The acres irrigated using drip methods during 1970 through 2000 were negligible (less than 1 percent in 2000). Using the data from KDA-DWR, the ratio of the acres irrigated using methods other than flood to the acres irrigated using the flood method for 1991 through 2008 for the model area was calculated and a relation between the state-wide ratio and the model area ratio was developed from the 10 years of overlap (1991 through 2000) between the datasets. In general, the ratio of center pivot to flood irrigation was lower for the whole State of Kansas than for the active part of the study area, probably in part because of the lack of large surface-water irrigation districts in the area. For the Middle Arkansas River Basin model, the state-wide ratio of center pivot to flood irrigation was multiplied by 2.5 to estimate the ratio of center pivot to flood irrigation was in their model area. A multiplier of 1.5 gave a better fit for the data for the active model area of this study. The 1.5 multiplier was applied to the state-wide ratio of center pivot to flood irrigation for 1970 through 1990 to estimate the ratio for the active model area. From this ratio, the percentage of acres irrigated by flood and nonflood methods in the active model area was estimated for 1970 through 1990. For the years before 1970, the ratio in 1970 was reduced by 0.03 each year through 1967 and by 0.02 each year for 1955 through 1966 to account for changes in irrigation methods with time. For the Middle Arkansas River Basin model, all irrigation before 1955 was assumed to use flood methods (Andrew Lyon, Kansas Department of Agriculture, Division of Water Resources, written commun., July 2010). This assumption also was used for the *Equus* Beds aquifer model. For the multiyear stress periods before 1990, the average percentage of acres estimated as irrigated by flood and nonflood methods was used to estimate the amount of irrigation return flow. Irrigation return flow calculated for each well was then subtracted from that well's pumping to obtain the net amount of groundwater pumpage. Although irrigation pumpage was assumed to occur only in May through August, annual irrigation pumping rates were calculated and used in the simulation.

Municipal pumpage was assumed to occur throughout the year. Monthly data, when available (for the city of Wichita municipal wells for 1990 through 1993 and 1995 through 2008), were used to determine annual pumping rates instead of using the annual rates from KDA-DWR. The changes made to the production pumpage data supplied by the city of Wichita are summarized in table 6, located at the back of the report. Average annual pumping rates were used for all other production wells.

**Table 5.** Estimated return flow from irrigation by irrigation system types.

[KDA-DWR, Kansas Department of Agriculture–Division of Water Resources]

| Return-flow system type | Estimated return flow (percent) | KDA-DWR irrigation system type |
|---|---|---|
| Flood | 25 | Flood |
| Center-pivot high-impact nozzle | 9 | Unreported |
| | | Center pivot-standard |
| | | Sprinkler other |
| | | Other |
| Center-pivot low-impact drop nozzle | 7 | Drip |
| | | Center-pivot low-impact drop nozzle |
| | | Drip and other |
| Combination | 12.2 | Center pivot and flood (assumed 80-percent center-pivot-standard and 20-percent flood) |

Some of the Wichita production wells were redrilled, causing substantial changes in screen and well depths. Information from NWIS and Wichita (Rich Robinson, city of Wichita, written commun., December 2009) about the well and screen depths, and information available from KDA-DWR was used to more accurately assign well and screen depth for each well in each stress period.

Annual volumes of artificial recharge in gallons for the *Equus* Beds Demonstration Recharge sites and at each of the phase I ASR sites (U.S. Geolgoical Survey, 2011) were available. These volumes were converted to cubic feet and then divided by the total number of days in the year to get the artificial-recharge rate in cubic feet per day as used in the model.

Some of the *Equus* Beds Recharge Demonstration and phase I ASR project's artificial-recharge sites that are not wells (for example, basins or trenches; fig. 3) cover parts of adjacent model-grid cells; however, all of the artificial recharge for these sites was assigned to the cell that contained the point location previously used as the location of the site. The error associated with assigning artificial recharge to one cell instead of all the cells that intersect the recharge basins is assumed to be small because the recharge basins do not extend more than one cell from the point location previously used as the location of the site. Because the model treats sites where water is pumped into or out of the aquifer as wells, the artificial recharge was distributed to the entire cell. If recharge wells were drilled into a recharge basin (for example, at the Recharge Demonstration basins at Halstead) and the amount of recharge at each well was unavailable, the total amount was divided equally among them.

### Head-Dependent Boundaries

The *Equus* Beds aquifer extends beyond the model boundary in several areas, and thus the model boundary does not represent the actual physical or groundwater flow boundaries of the aquifer. These boundaries were simulated in the model as general head boundaries, a form of the head-dependent flux boundary that allows groundwater to enter or exit the model proportional to the difference between the water level in the model and the water level assigned to the boundary multiplied by a conductance term that limits the rate of flow (McDonald and Harbaugh, 1988). These boundaries were located as far as practical from the Wichita well field to limit boundary effects on model results. Water levels along the boundary were assigned to each general head boundary cell based on an assumed water table value located 20 miles outside the model. General head boundary conductances were calculated by multiplying the hydraulic conductivity of each general head boundary cell by the length and width of the cell divided by the distance to the location of the assumed water-table value (20 miles). General head boundaries are shown in figure 31.

### Head and Streamflow Gain and Loss Observations

Groundwater-level observations and streamflow gain and loss observations were compared to simulated groundwater levels and streamflow gains and losses using the Head Observation Package and the River Observation Package (Harbaugh and others, 2000) for the steady-state and transient groundwater calibration simulations. Groundwater-level observation data, including groundwater level altitude, well location within the model, and time of observation, were calculated and entered into the Head Observation Package.

Groundwater-level data and associated well-construction and aquifer information available from the USGS NWIS database (U.S. Geological Survey, 2009a; U.S. Geological Survey, unpub. data, 2009) and the Kansas Geological Survey's WIZARD database (Kansas Geological Survey, 2009) were compiled for wells in the study area. Groundwater levels commonly are recorded as depth below land surface. To convert them to groundwater altitudes, they were subtracted from the land-surface altitude determined for the well. If a land-surface altitude was not determined for the well, one was estimated

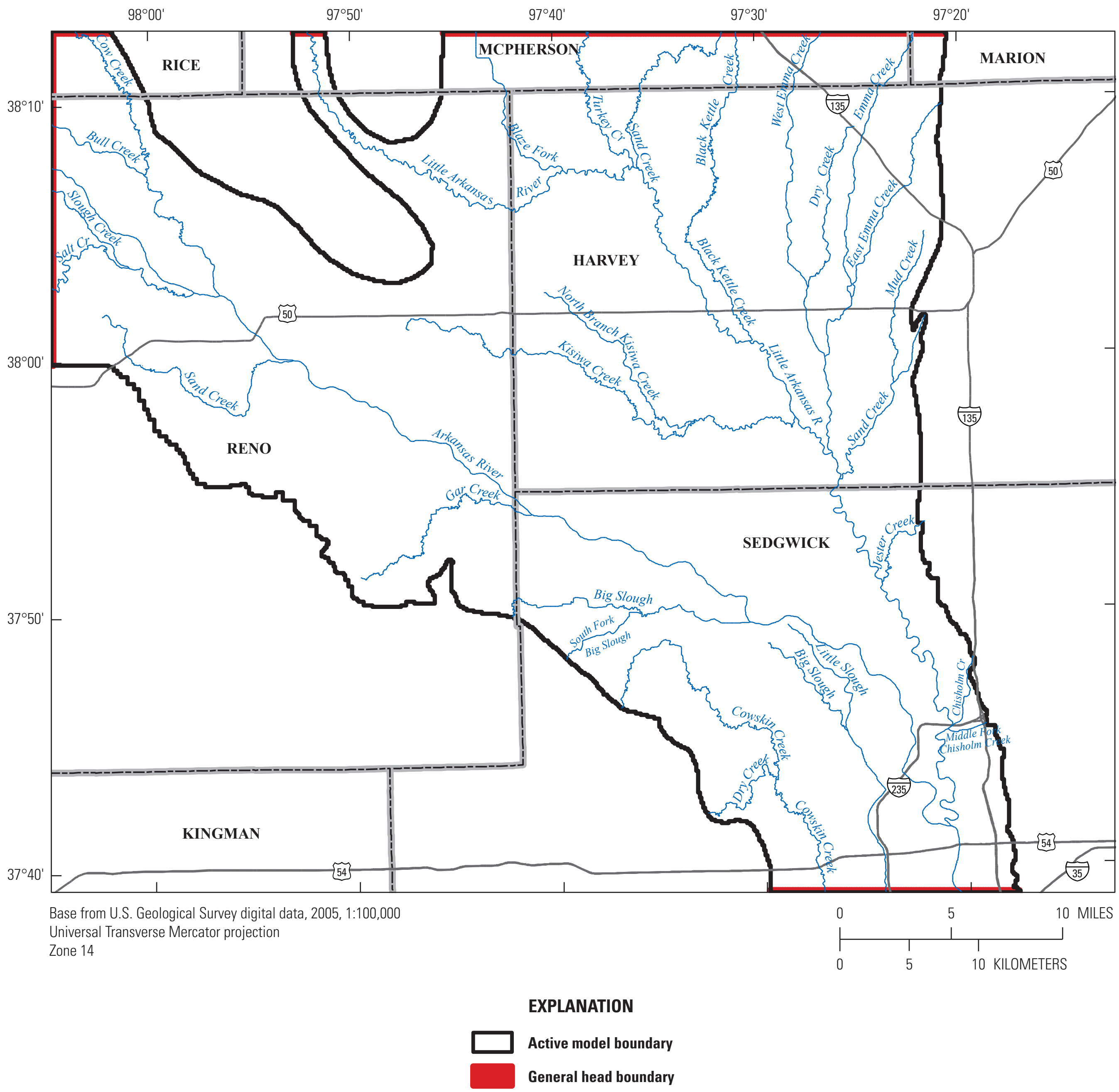



**Figure 31.** Locations of general head boundaries for the groundwater model.

from the National Elevation Datum database (U.S. Geological Survey, 2009b). Wells and their associated groundwater altitudes were assigned to a model layer based on the altitude of the bottom of the well's screened interval, or, if the screened interval was unavailable, the altitude of the bottom of the well. If neither of these data values were available, then the well was not used. Hydrogeologic information stored in NWIS was used to confirm that a well was open to the *Equus* Beds aquifer. If a well was in the aquifer but was 5 or fewer ft deeper than the bottom of the modeled aquifer, the well was retained in the dataset and the water-level altitude assigned to the bottom layer of the model.

Streamflow gain or loss observation data, including a list of model cells for each stream reach and flow into or out of the aquifer along the stream reach, and time of observation were calculated and entered into the River Observation Package. Streamflow measurements at USGS streamflow gages on the Arkansas River near Maize (07143375) and Hutchinson (07143330), and on Little Arkansas River at Valley Center (07144200) and Alta Mills (07143665) (fig. 6) were used to estimate base flow (gains from and losses to the aquifer) for each model stress period when measurements from each pair of gages were available using hydrograph separation (Lim and others, 2005). Mean base flow for each stress period was calculated for each gage. Streamflow gains and losses for each stress period were then calculated by subtracting upstream gage base flow from downstream gage base flow. Streamflow gain is caused by discharge of water from the aquifer to the stream and is represented by a negative number. Conversely, streamflow loss results from water flow from the streams into the *Equus* Beds aquifer and is represented by a positive number.

## Geometric Multigrid Solver

The groundwater flow equation was solved by the geometric multigrid method (Wilson and Naff, 2004), a method for solving the groundwater flow equation. Closure criteria are set to stop the interative solver for head and flow residual. The head closure criterion was set to 0.01 ft and the flow residual criterion was set to 1,000.0 $ft^3$/day.

## Model Calibration

The groundwater-flow model was calibrated by adjusting model input data until model results matched field observations within an acceptable level of accuracy (Konikow, 1978). Both steady-state and transient hydraulic head and streamflow data were used to calibrate the model. Steady-state conditions occur when inflow to the system equals outflow from the system. Calibration to steady-state conditions was used to assess the conceptual model of groundwater flow and simulated boundary conditions, and estimate hydraulic conductivity values and recharge rates. Transient conditions occur when inflow does not equal outflow and is balanced by water flow into or out of the aquifer from storage. Calibration to transient conditions refined the model hydraulic properties determined from the steady-state calibration and provided estimates of storage properties of the aquifer.

Calculation of parameter sensitivities was used for the steady-state predevelopment simulation to indicate the relative importance of each model input variable. Parameter values from the steady-state simulation were used as a starting point for manual calibration of the transient simulation. Hydraulic properties adjusted during the calibration process include horizontal hydraulic conductivity, vertical hydraulic conductivity between model layers, specific storage, specific yield, recharge rates, evapotranspiration, streambed hydraulic conductivity, and general head boundary conductance. After each change in one of these parameters, the simulated groundwater levels and streamflow gains and losses were compared to observed values. The difference between simulated and observed values is called the residual. Parameter estimation (Harbaugh and others, 2000) was attempted for the transient simulation; however, nonconvergence for the transient parameter-estimation simulations prevented its use. The nonconvergence was most likely caused by nonlinear groundwater flow, heterogeneous hydraulic properties of the *Equus* Beds aquifer, and complexity of the transient simulation.

The model accuracy was estimated using several methods. The root mean square (RMS) error between observed and simulated hydraulic head as well as observed and simulated streamflow gains or losses were calculated for each well and stream observation for the entire simulation. Model accuracy was increased by minimizing the RMS error during the calibration process. The RMS error measures the absolute value of the variation between measured and simulated hydraulic heads at control points or the variation between measured and simulated streamflow along stream reaches. The equation to calculate the RMS error is:

$$\text{RMS error} = \sqrt{\frac{e_1^2 + e_2^2 + e_3^2 + ...e_n^2}{n}}\ , \tag{5}$$

where

$e$ is the difference between the observed and simulated values, and

$n$ is the number of observations.

Water-table altitudes range from about 1,500 to about 1,300 ft above NAVD 88 in the main part of the model area between Hutchinson and Wichita, Kans. (or 200 ft of head loss, excluding the dune sand area) (Myers and others, 1996). The ratio of the RMS error to the total head loss in the model area is a measure of the amount of model error in the overall model response. A value less than 10 percent is a generally accepted threshold (Anderson and Woessner, 1992). Thus, for this study, the RMS error divided by the total head loss should be less than 20 ft (10 percent of the 200 ft of head loss in the model area).

The mean error between observed and simulated hydraulic head and between observed and simulated streamflow gains and losses was calculated for each well and each stream observation for the entire simulation. In keeping with the MODFLOW-2000 convention, simulated results were subtracted from observed values. Negative errors indicated the simulated results were too large (simulated result needs to decrease), positive errors indicated the simulated results were too small (simulated result needs to increase). Model accuracy increased the closer the value of the mean error was to zero. The mean error measured the average difference between measured and simulated hydraulic heads at control points or the variation between measured and simulated streamflow gains and losses along stream reaches, and indicated if simulated results were higher or lower than measured observations.

The accuracy of water-level measurements also was one of the criteria used to assess values of the RMS and mean errors used to determine if the model calibration was acceptable. Most groundwater levels used for calibration were measured with a steel tape or an electric water-level measuring tape to the nearest 0.01 ft. Historical water levels for wells were measured or estimated using unknown techniques. For these water-level measurements, the accuracy is assumed to be within 1 ft. The measuring-point altitudes for most wells used in this study were obtained using standard surveying or global positioning system methods. The accuracy of these altitudes is between 0.01 and 0.5 ft. The measuring-point altitude of a few wells in the study area was estimated from USGS 7.5-minute topographic maps. The vertical accuracy of land-surface altitudes from these maps is one-half of the contour interval. The contour interval on topographic maps is 5 or 10 ft and the accuracy of measuring-point altitudes for these wells is 2.5 or 5 ft, respectively; therefore, the largest possible error in measurement of water-level altitudes is approximately 5 ft.

Water levels measured in monitoring wells located near pumping wells are closely related to the rate of pumping. The use of an average pumping rate instead of the actual pumping rate can introduce substantial error between a simulated and measured water level. The most likely instance when this would occur is when average annual pumping rates are used. Typical well-field pumping consists of increasing and decreasing pumping rates by turning wells on or off to meet water-supply demand. If the water level was measured when the nearby well was pumping, the simulated water levels will be greater than the measured water level. If the well was not pumping, the simulated water levels will be too low. This type of error is not quantified easily but could be several feet if the measured well is close to the pumping well. The maximum possible error for water-level measurements is the sum of the maximum errors caused by water-level measurement errors, measuring-point altitude errors, and well pumping. The chance that the maximum error would occur at any well is small. A combination of errors of varying value and sign is more likely to occur.

River stage is measured at USGS streamflow gages to the nearest 0.01 ft. Streamflow measurement accuracy is plus or minus 2 percent of the actual value for “excellent” measurements, plus or minus 5 percent for “good” measurements, and plus or minus 8 percent for “fair” measurements (Rantz and others, 1982). An estimate of the error associated with the calculation of base flow was made using the assumption that all streamflow measurements were “good” and each measurement was within 5 percent of the actual value. Estimated base flow for each gage was multiplied by 0.05 to obtain an estimate of the error in base flow from the error in each streamflow measurement. The largest base flow error from measurement is represented by two conditions, subtracting a high upstream measurement from a low downstream measurement, and subtracting a low upstream measurement from a high downstream measurement. These two conditions were used to calculate the largest and smallest measurement error for estimated base flow observations. For the Arkansas River streamflow gain or loss observations the largest estimated base flow error from streamflow measurements is almost 12,375,000 $ft^3$/day (143 $ft^3$/s), the smallest is almost 556,000 $ft^3$/day (6 $ft^3$/s), and the mean is almost 3,615,000 $ft^3$/day (42 $ft^3$/s). For the Little Arkansas River estimated base flow observations, the largest estimated base flow error from measurements is more than 4,300,000 $ft^3$/day (50 $ft^3$/s), the smallest is more than 132,000 $ft^3$/day (2 $ft^3$/s), and the mean is almost 1,088,000 $ft^3$/day (13 $ft^3$/s).

The amount of error associated with the method used to estimate base flow is unknown but may be substantial. Estimates of base flow may be affected by streamflows that result from regulation. These may include flows from sewage treatment facilities, flood control reservoirs , and water-supply diversions. Also, base flow estimates are related to the hydrologic conditions of the period of record used in the analysis. Base flow estimated during a dry or wet period will be biased toward those conditions (Sloto and Crouse, 1996). Knowledge of errors associated with observation data is important for choosing an appropriate calibration target and for preventing calibration of the model to an error substantially smaller than the errors associated with the measurement of the observed data.

For the Arkansas River, estimated base flow observations near Maize (07143375) at river mile 772.2 ranged from almost 4,370,000 to almost 139,290,000 $ft^3$/day (51 to 1,612 $ft^3$/s), and at Hutchinson (07143330) at river mile 800.3 from more than 6,655,000 to more than 108,193,000 $ft^3$/day (77 to 1,252 $ft^3$/s). Observed base flow was calculated for each reach and for all base flow observations, and the minimum was subtracted from the maximum to calculate the range of observed base flow. The range of observed base flow on the Arkansas River between Maize (07143375) and Hutchinson (07143330), 28.1 river miles in length was 35,778,000 $ft^3$/day (414 $ft^3$/s). For the Little Arkansas River, estimated base flow observations at Valley Center (07144200) at river mile 17.5 ranged from more than 2,052,000 to more than 56,528,000 $ft^3$/day (24 to 654 $ft^3$/s), and at Alta Mills (07143665) at river mile 50.1 from more than 473,000 to more than 29,700,000 $ft^3$/day (5 to 344 $ft^3$/s). The range of observed base flow on the Little Arkansas River between Valley Center and Alta Mills, 32.6 river miles in length was almost 25,371,000 $ft^3$/day (294 $ft^3$/s). The ratio of the RMS error to the total range in observed base flow is a measure of the amount of base flow error in the overall model response. Accounting for errors in base flow from streamflow measurements and errors associated with base flow estimation using hydrograph separation, an arbitrary value of 25 percent was chosen as an acceptable ratio of RMS error for simulated base flow to total range in estimated base flow. The RMS error to total range in observed base flow should be less than 8,944,500 $ft^3$/day (103.5 $ft^3$/s) for the Arkansas River and less than 6,342,750 $ft^3$/day (73.4 $ft^3$/s) for the Little Arkansas River.

### Steady-State Calibration

The steady-state hydraulic head data were obtained from historic groundwater level data from 284 wells in the study area. Well locations are shown in figure 32 and the well number, date of observation, observed water level, and simulated water level of each well used in the steady-state calibration are listed in table 7 at the back of this report. Head observation data were collected between 1935 and 1939. Concurrent streamflow measurements between gage pairs are unavailable for the Arkansas and Little Arkansas Rivers before 1959, thus the steady-state simulation could not be calibrated to streamflow gains or losses. Values for river stage, recharge, evapotranspiration, and well pumping averaged from 1935 to 1939 were assumed to approximate steady-state conditions. In reality, river stage, recharge, evapotranspiration, and well pumping were variable during this time and groundwater levels responded to these changes. Because the amount of well pumping was relatively small and constant, and groundwater level and river stage measurements from one gage each on the Arkansas and Little Arkansas Rivers are available, this period is the best estimation of pre-development conditions for model calibration. The RMS error for the steady-state calibration simulation is 9.82 ft. The ratio of the RMS error to the total head loss in the model area is 0.049 (9.82 ft divided by 200 ft) or 4.9 percent. The level of accuracy of the simulation in representing the steady-state hydraulic-head distribution was acceptable because it is less than 10 percent of the change in groundwater level across the model, and is close to the assumed groundwater level measurement errors previously discussed. The mean error (observed–simulated) for 284 water-level observations is 3.86 ft.

The location of wells and calibration residuals calculated as the simulated head minus observed head in ft at each well with an observation is shown in figure 32. For most of the modeled area, simulated head is within 5 ft of observed head. Simulated heads are more than 5 ft greater than observed heads near the Wichita well field. The larger simulated heads are assumed to be the result of observations recorded in 1939 when well pumping was larger but simulated pumping was lower because pumping was averaged from 1935 through 1939. The observed and simulated groundwater level maps from 1940 are shown in figure 33.

### Steady-State Groundwater Flow Budget

Inflows and outflows to the groundwater model were recorded for the steady-state calibration simulation and are listed in table 8. Total simulated flow through the groundwater system was more than 49 million $ft^3$/day. Major inflows to the system as a percent of total flow were recharge (64.7 percent) and river leakage (30.5 percent). Major outflows from the system were river leakage (51.8 percent), evapotranspiration (38.8 percent), drains (4.6 percent), and well pumping (4.6 percent). The difference between inflows and outflows, called the mass balance, indicates the ability of the numerical model to solve the groundwater flow equation such that numerical errors are small. The difference between flows into and out of the model was -0.08 percent of total flow for the steady-state calibration simulation.

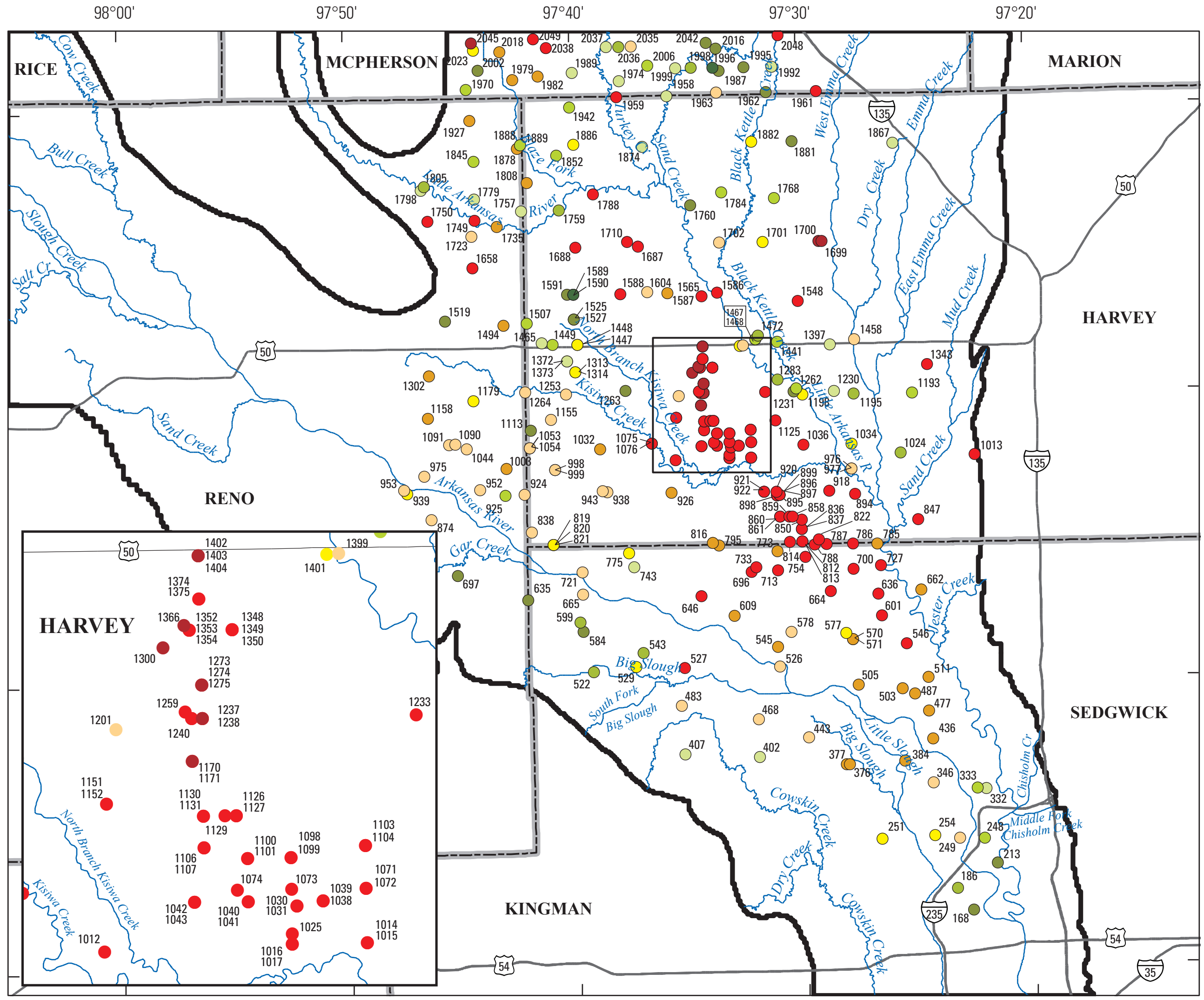
### Transient Calibration

Hydraulic-head data for the transient calibration were obtained from 346 wells in the study area (fig. 34). The well number, date of observation, observed water level, and simulated water level for each well used in the transient calibration are listed in table 9 at the back of this report. Wells were selected to include all model layers and a wide distribution in the model. A total of 3,677 water-level observations from 1935 through 2008 were used for the transient calibration. The RMS error for all water-level observations is 2.48 ft for the transient calibration. This value is less than the maximum measurement errors and indicates the acceptability of the calibrated model. The ratio of the RMS error to the total head loss in the model area (2.48/200) is 0.0124, or 1.24 percent. The mean error for all water level observation wells used in the transient calibration is 0.03 ft.

**Table 8.** Steady-state calibration simulation flow budget.

[$ft^3$/day, cubic feet per day; acre-ft/day, acre feet per day; --, not applicable]

| Budget component | Flow rate, in $ft^3$/day | Flow rate, in acre-ft/day | Percent of total flow |
|---|---|---|---|
| | Inflow | | |
| Head dependent boundaries | 2,320,409 | 53.3 | 4.7 |
| Recharge | 31,855,858 | 731.3 | 64.7 |
| River leakage | 15,024,649 | 344.9 | 30.5 |
| Well pumping | 0 | 0.0 | 0.0 |
| Total in | 49,200,916 | 1,129.5 | 100 |
| | Outflow | | |
| Head dependent boundaries | 1,167,715 | 26.8 | 0.2 |
| Evapotranspiration | 18,569,682 | 426.3 | 38.8 |
| Drains | 2,129,863 | 48.9 | 4.6 |
| River leakage | 25,165,966 | 577.7 | 51.8 |
| Well pumping | 2,204,735 | 50.6 | 4.6 |
| Total out | 49,237,960 | 1,130.3 | 100 |
| Total in - out | 37,044 | 0.9 | -- |
| Percent difference | -0.08 | -0.08 | -- |




Base from U.S. Geological Survey digital data, 2005, 1:100,000
Universal Transverse Mercator projection
Zone 14

0 5 10 MILES
0 5 10 KILOMETERS

**EXPLANATION**

**Active model boundary**

Monitoring well and number—Simulated head minus observed head, in feet

| | |
|---|---|
| -36.8 to -20.0 | 0.1 to 1.0 |
| -19.9 to -5.0 | 1.1 to 3.0 |
| -4.9 to -3.0 | 3.1 to 5.0 |
| -2.9 to -1.0 | 5.1 to 20.0 |
| -0.9 to 0.0 | 20.1 to 48.7 |

**Figure 32.** Monitoring well locations and residuals (simulated head minus observed head, in feet) for the steady-state calibration simulation.

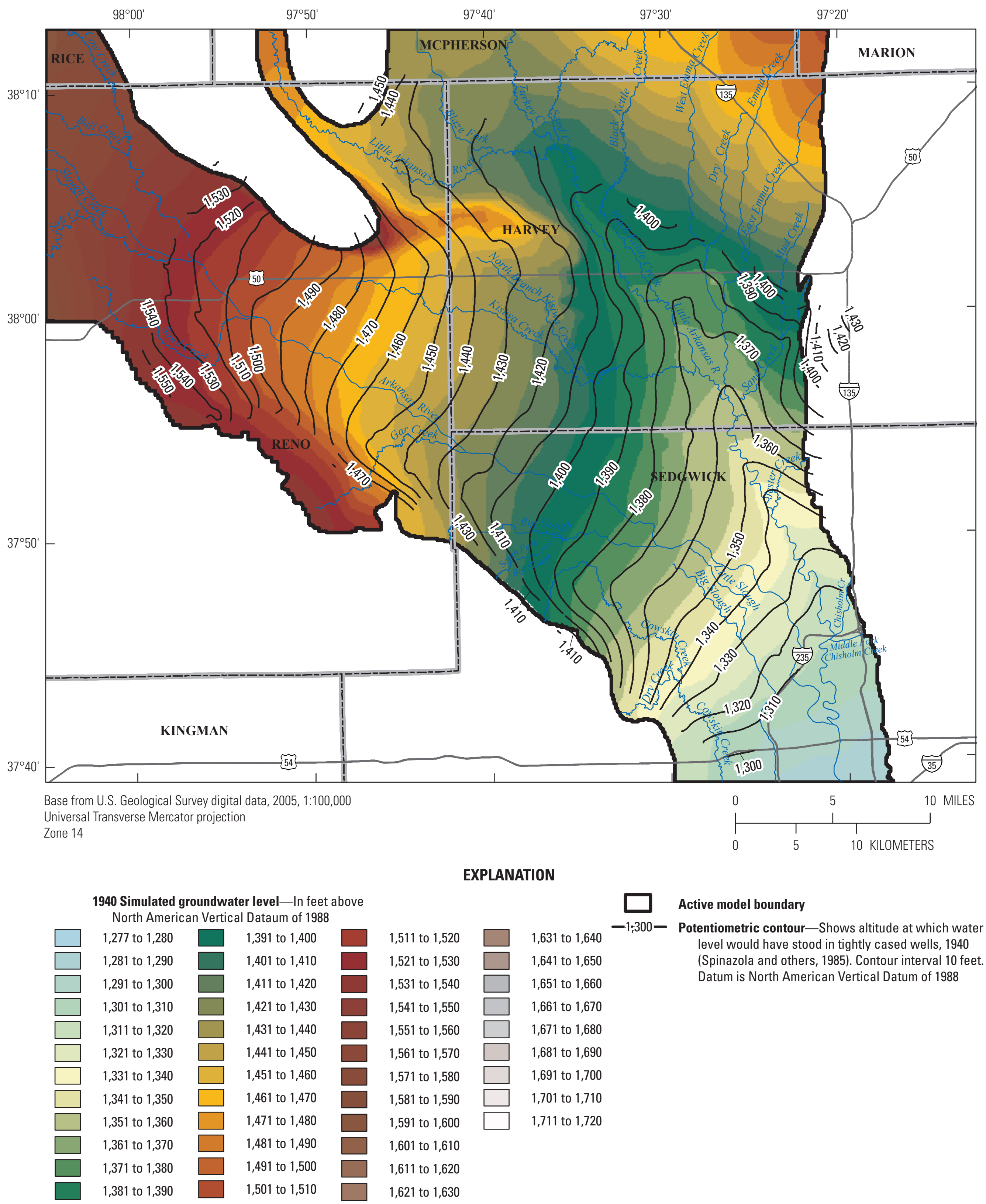
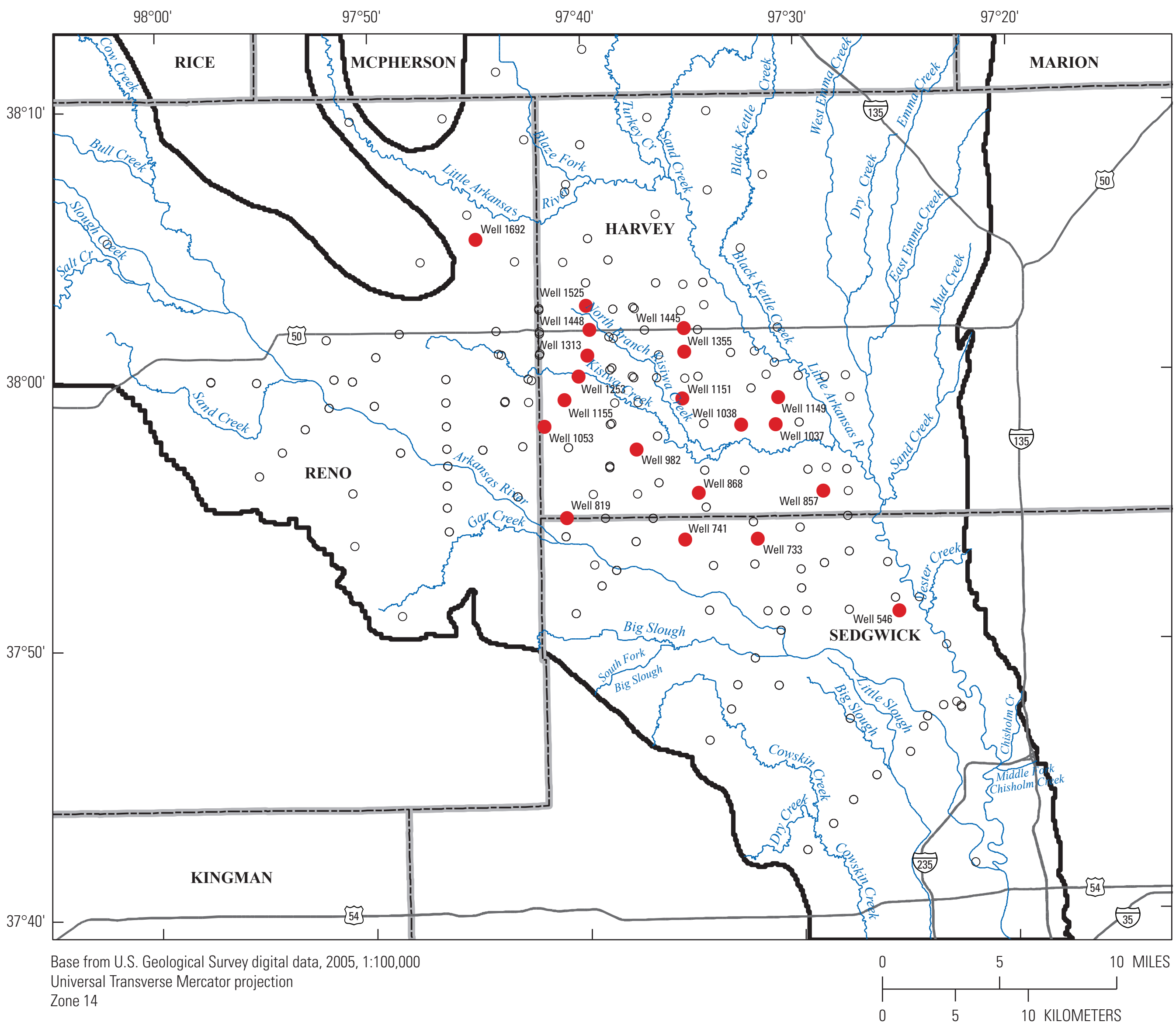



**Figure 33.** Observed and simulated *Equus* Beds aquifer groundwater levels, 1940.




**EXPLANATION**

**Active model boundary**

**Monitoring well and number—Hydrographs shown on fig. 40**

**Monitoring well**

**Figure 34.** Monitoring well locations used for the transient calibration simulation.

Simulated versus observed groundwater levels closely match the one to one line and are plotted in figure 35. The modeled area was divided into six zones (fig. 36) with calibration statistics calculated for each zone to allow assessment of model calibration for different model areas. Zone 1 is the basin storage area and contains the index wells and artificial-recharge accounting index cells (used by the city to monitor water levels in the aquifer and any changes that might occur as a result of the ASR project, fig. 3), zone 2 is near the Burrton area, zone 3 is near the Arkansas River, zone 4 is the dune sand area, zone 5 is the upland area south of the Arkansas River, and zone 6 is the upland area north and east of the Little Arkansas River. These zones roughly correspond to similar zones presented in Myers and others, (1996), except for zone 4. Calibration zones, RMS error, the RMS error divided by the head loss, and mean error for each calibration zone are listed in table 10. Monitoring well locations used for calibration in the transient simulation are indicated by zone on figure 36.

The areal distribution of mean error, the average difference between observed and simulated groundwater levels for all wells and for the entire transient simulation period, can reveal areas of the model that consistently over- or under-simulate groundwater levels. The mean error for wells in each model layer is shown in figures 37, 38, and 39. The distribution of mean error for simulated groundwater levels in model layer 1 does not indicate a spatial bias in most of the modeled area (fig. 37). In the area south of the Arkansas River, near Mount Hope (fig 1), simulated groundwater altitudes are greater than observed and simulated groundwater altitudes are less than observed in the dune sand area north of Burrton. For model layer 2, simulated groundwater altitudes are slightly less than observed to the southwest of Burrton and along the Arkansas River between Hutchinson to just upstream from Mount Hope (fig. 38). No spatial bias in mean error is apparent for the rest of model layer 2. The distribution of mean error for model layer 3 (fig. 39) indicates simulated groundwater altitudes are slightly less than observed to the southwest of Burrton and along the Arkansas River between Hutchinson to just upstream from Mount Hope, as was indicated in model layer 2. North of the Little Arkansas River, between Blaze Fork and Turkey Creek simulated groundwater altitudes are greater than observed in layer 2.

Comparison of simulated and observed well hydrographs is used to assess the response of simulated groundwater levels to temporal changes in stresses to the aquifer. Simulated and observed groundwater levels are shown for 20 selected wells in figure 40. Multiyear stress periods were simulated from 1935 through 1989 and annual stress periods were simulated from 1990 through 2008. Multiyear trends in the hydrographs are illustrated by the overall trends from 1935 through 2008. Simulated water levels follow the observed long-term trends for most wells, indicating the model adequately simulates long-term changes to groundwater levels resulting from sustained stresses on the aquifer such as long-term rate of groundwater withdrawal, gains from and losses to streams, or long-term trends in recharge.

Some differences in long-term trends are apparent in the simulated versus observed hydrographs for wells 733, 741, 819, 868, 1053, 1149, 1155, 1253, and 1525 in the multiyear stress period for 1953 through 1958 (fig. 40). All of these wells show simulated water levels went down or the rate of decrease was faster during the 1950s but the observed water levels went up or the rate of decrease was slower. The most likely explanation for this is that average rainfall assigned to the 1953 to 1958 stress period is less than during 1958 when observed groundwater levels were measured. Average rainfall for the 1953–58 stress period was about 25 inches per year; however, in 1957, rainfall was almost 40 inches per year and in 1958 rainfall was almost 36 inches per year (fig. 25). The lower simulated values were caused by using the average rainfall rate for the stress period. The larger observed values resulted from water levels that were used as observations and measured in 1958 after they had increased in response to the larger than average rainfall for 1957 and 1958.

Annual trends are illustrated in the hydrographs between 1990 and 2008 when annual pumping, annual stream flow, and annual recharge were simulated. Simulated short-term trends follow observed water level trends for most wells. Differences between annual simulated and observed water levels are most

**Table 10.** Root Mean Square error, the ratio of Root Mean Square error to head loss, and mean error for each transient calibration zone.

[--, not applicable]

| Calibration zone | Calibration zone from Myers and others (1996) | Root mean square error, in feet | Ratio of root mean square error to head loss, in feet | Mean error, in feet (negative value indicates simulated is larger than observed) | Mean absolute difference (Myers and others, 1996) |
|---|---|---|---|---|---|
| 1 (Basin Storage Area) | 5 | 2.74 | 0.014 | -0.199 | 6.76 |
| 2 (Burrton Area) | 1 | 2.45 | 0.012 | -0.055 | 5.76 |
| 3 (Arkansas River) | 2 | 1.5 | 0.008 | 0.518 | 2.47 |
| 4 (Sand Dunes) | -- | 2.09 | 0.01 | 1.58 | -- |
| 5 (South Uplands) | 3 | 1.39 | 0.007 | 0.167 | 2.15 |
| 6 (North Uplands) | 4 | 8.35 | 0.042 | -6.258 | 6.76 |

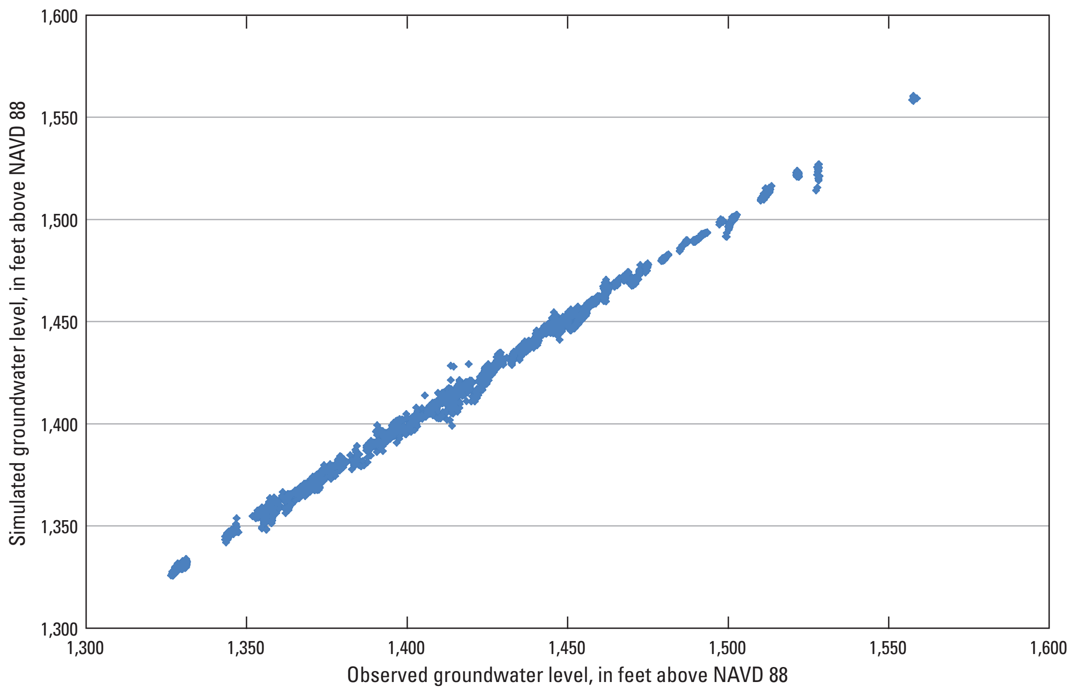



**Figure 35.** Simulated versus observed groundwater levels for the transient calibration.

apparent for wells 857, 868, 1448, and 1692 (fig. 40), where large short-term variations in the observed water levels are not well simulated although the overall trends are similar. These differences between simulated and observed short-term water levels for wells are most likely caused by observed water levels measured after large stresses such as precipitation events that occur in a shorter time interval than the model stress periods or heterogeneities in the hydraulic properties of the aquifer at these locations that are not incorporated into the model.

The RMS error calculated for observed and simulated base flow gains or losses for the Arkansas River for the transient simulation was 7,916,564 $ft^3$/day (91.6 $ft^3$/s) and the RMS error divided by the total range in streamflow (7,916,564/37,461,669 $ft^3$/day) is 22 percent. The RMS error calculated for observed and simulated streamflow gains or losses for the Little Arkansas River for the transient simulation was 5,610,089 $ft^3$/day (64.9 $ft^3$/s) and the RMS error divided by the total range in streamflow (5,612,918/41,791,091 $ft^3$/day) is 13 percent. The RMS values are less than the maximum measurement errors and the RMS error divided by the total range in streamflow are less than 25 percent, indicating the acceptability of the simulated streamflow gains or losses in the transient calibrated model. The mean error between observed and simulated base flow gains or losses was 29,999 $ft^3$/day (0.34 $ft^3$/s) for the Arkansas River and -1,369,250 $ft^3$/day (-15.8 $ft^3$/s) for the Little Arkansas River. Observed and simulated streamflow gains or losses for each stress period are listed in table 11 at the back of this report. Comparison of observed and simulated cumulative streamflow gains or losses indicate how well the model simulates long-term streamflow gains or losses. Cumulative streamflow gain and loss observations are similar to the cumulative simulated equivalents and are shown for the Arkansas River and Little Arkansas River in figure 41.

### Transient Groundwater Flow Budget

Inflows and outflows to the groundwater model were recorded for each stress period of the transient simulation. Average flow rates and cumulative flows for the transient calibration simulation are listed for each stress period in table 12 at the back of the report. Cumulative inflows to the system as a percent of total flow from largest to smallest were recharge (67 percent), river leakage (27 percent), head-dependent boundaries (4 percent), and storage (2 percent). Cumulative outflows from the system from largest to smallest were river leakage (42 percent), evapotranspiration (34 percent), well pumping (16 percent), drains (4 percent), storage (2 percent), and head-dependent boundaries (2 percent). Average percent mass balance difference for individual stress periods ranged from -0.46 to 0.51 percent. The cumulative mass balance for the transient calibration was 0.01 percent.

### Parameter Sensitivity

A sensitivity analysis was performed to assess the response of the model to changes in various input parameter values. When the model is sensitive to an input parameter, small changes to the parameter value cause large changes in hydraulic head. If a change of parameter value does not change the simulated hydraulic head distribution, the model is

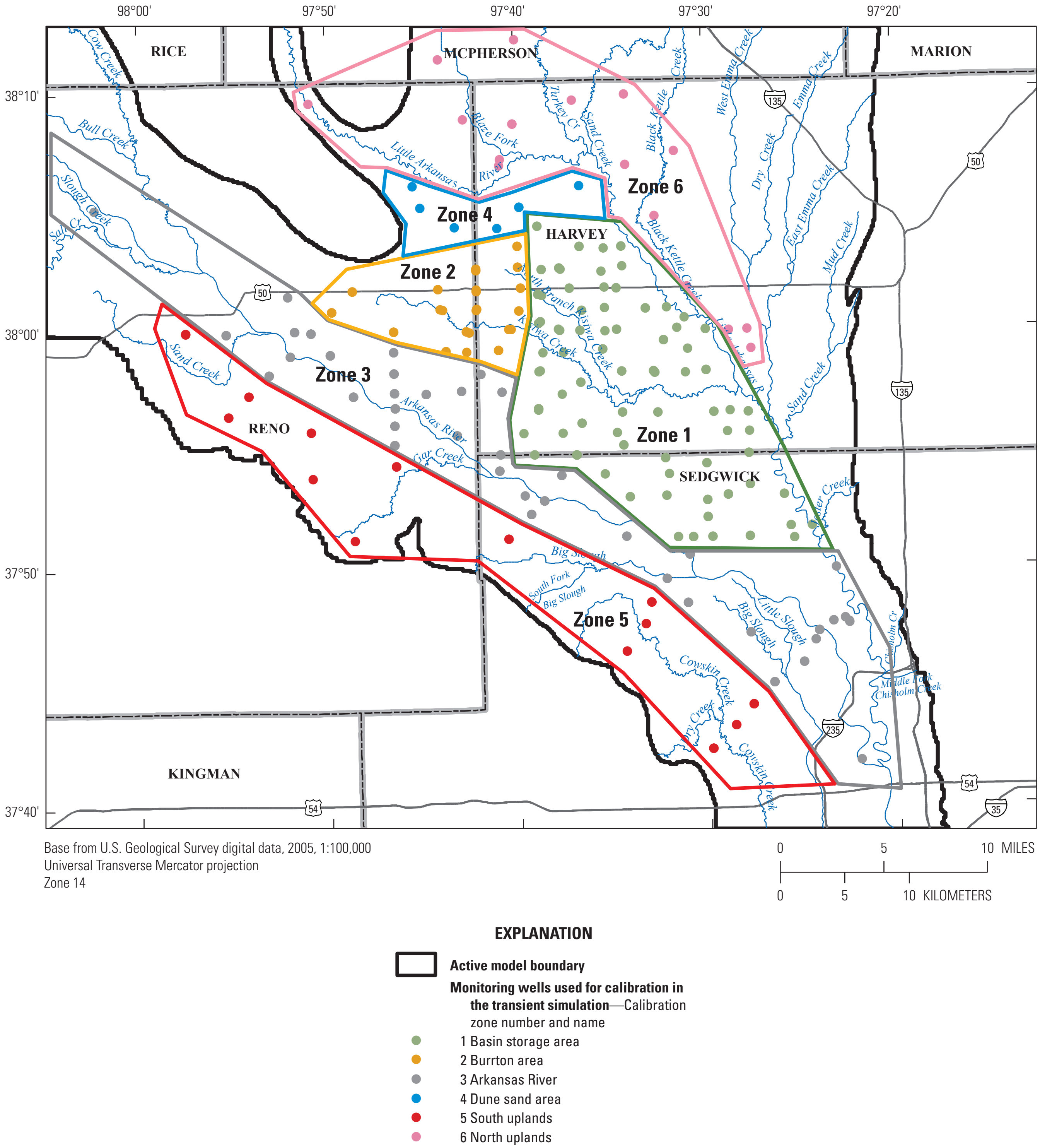



**Figure 36.** Calibration zones for monitoring wells used for calibration in the transient simulation.

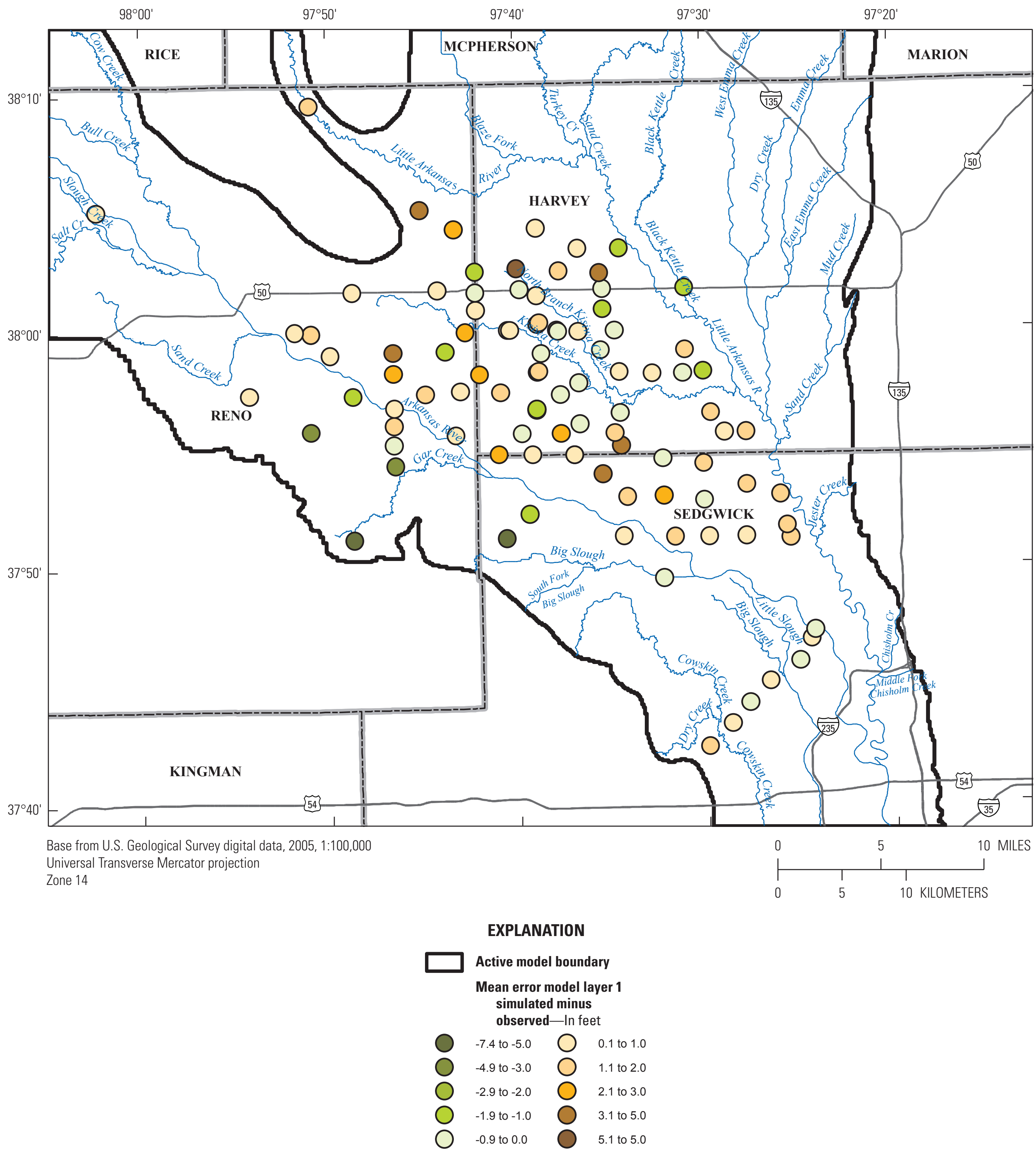



**Figure 37.** Mean error between observed and simulated water levels from wells in model layer 1 for the transient calibration period.

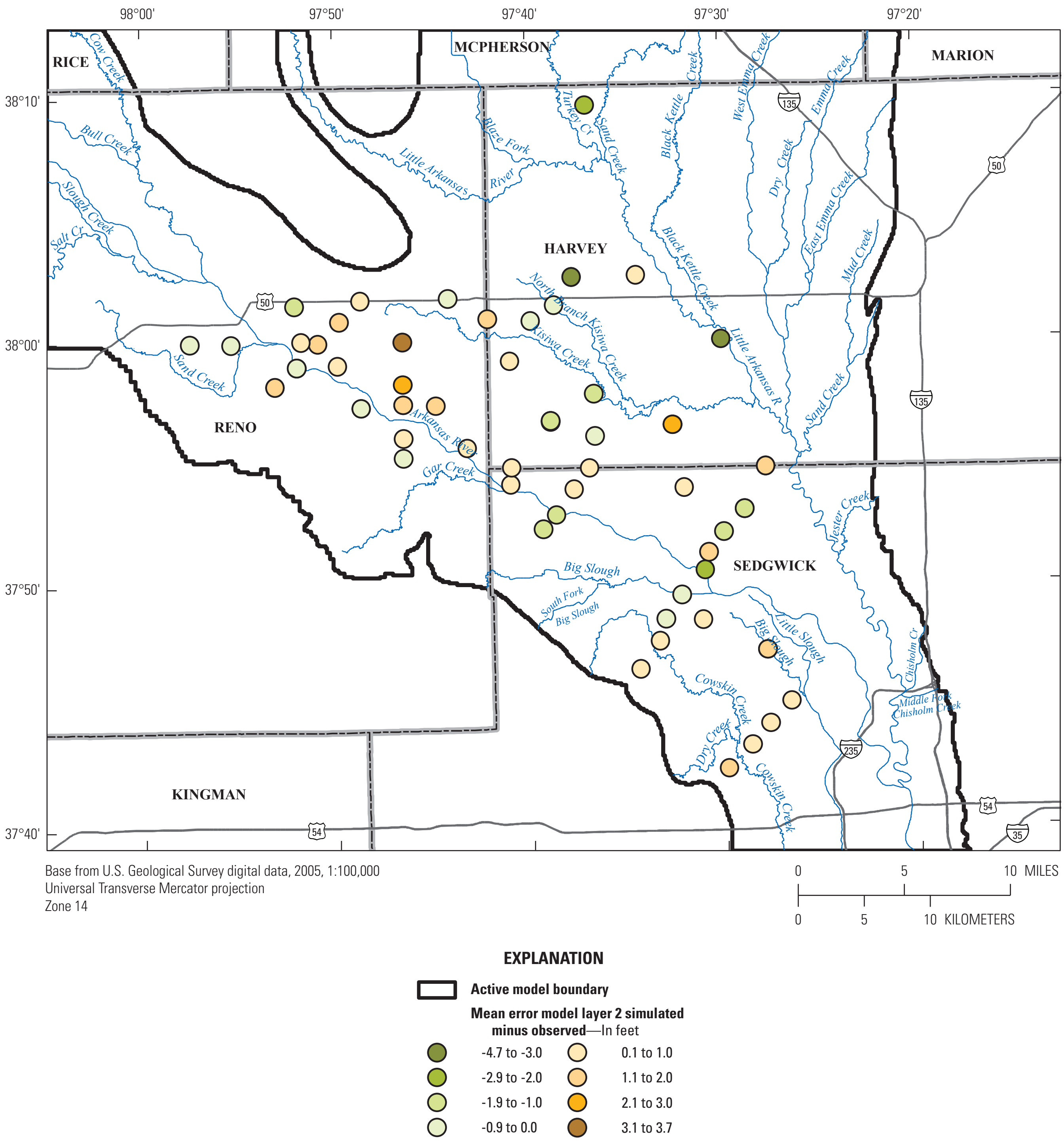



**Figure 38.** Mean error between observed and simulated water levels from wells in model layer 2 for the transient calibration period.

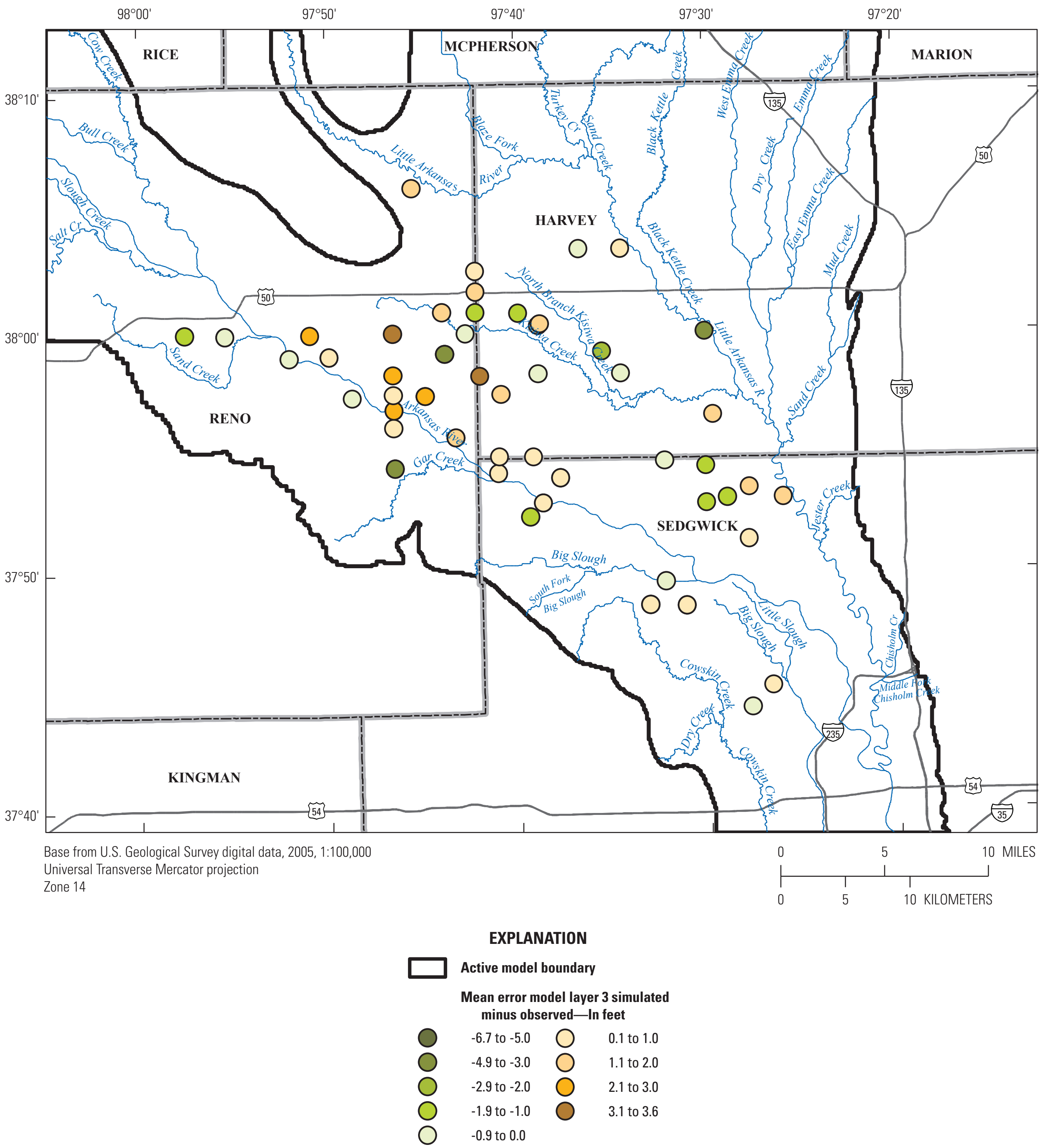



**Figure 39.** Mean error between observed and simulated water levels from wells in model layer 3 for the transient calibration period.

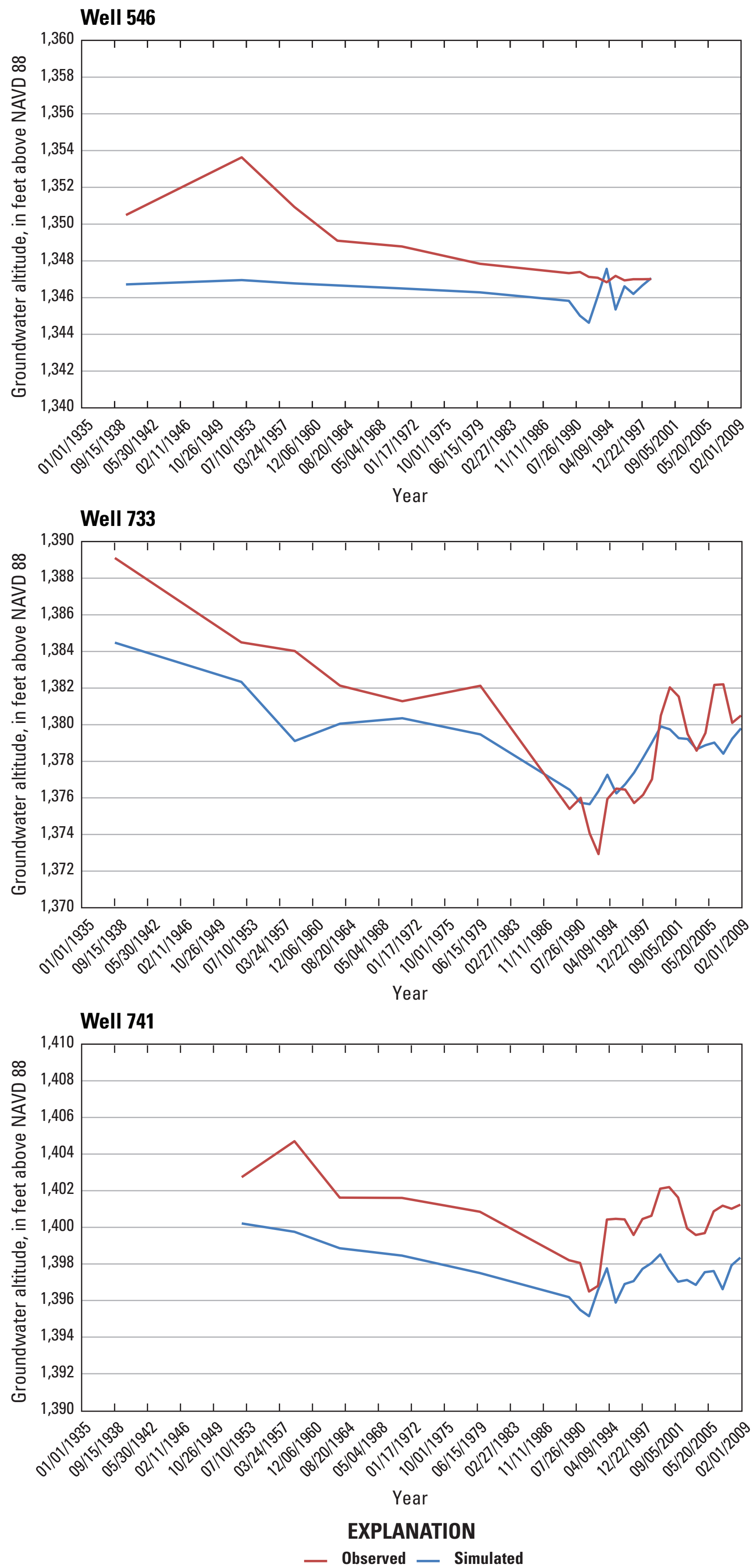



**Figure 40.** Simulated and observed groundwater levels for selected wells.

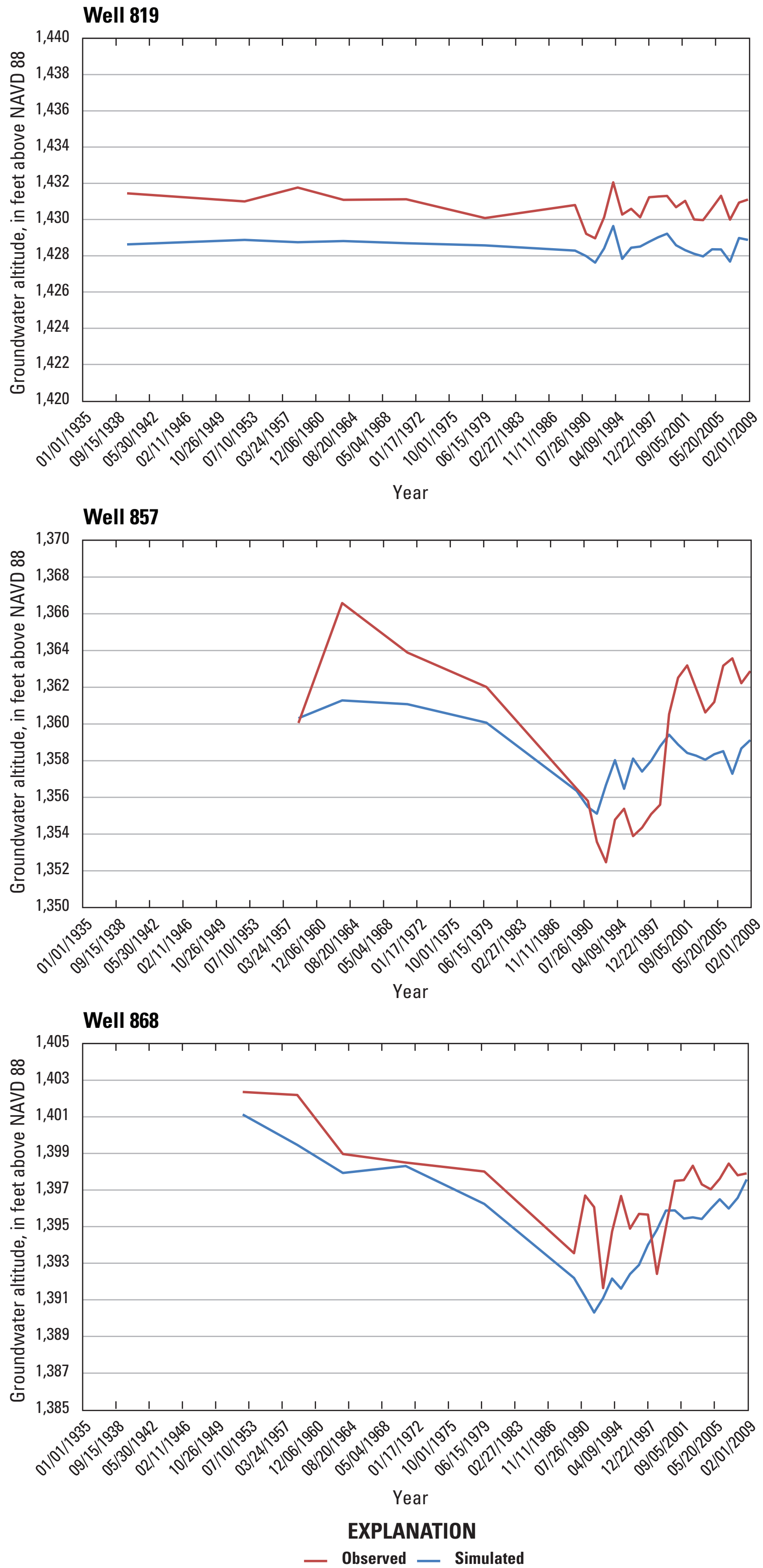



**Figure 40.** Simulated and observed groundwater levels for selected wells.—Continued

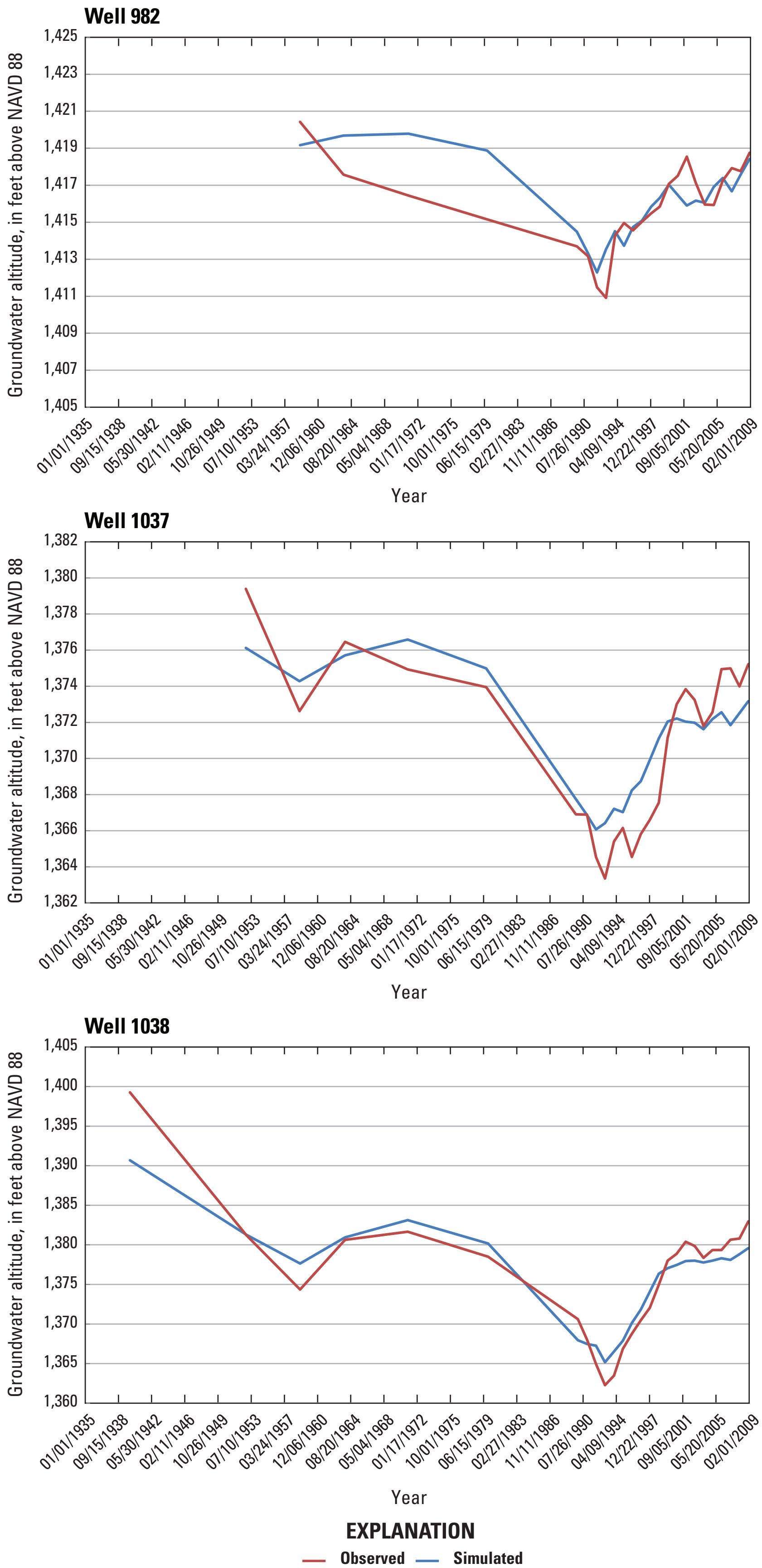



**Figure 40.** Simulated and observed groundwater levels for selected wells.—Continued

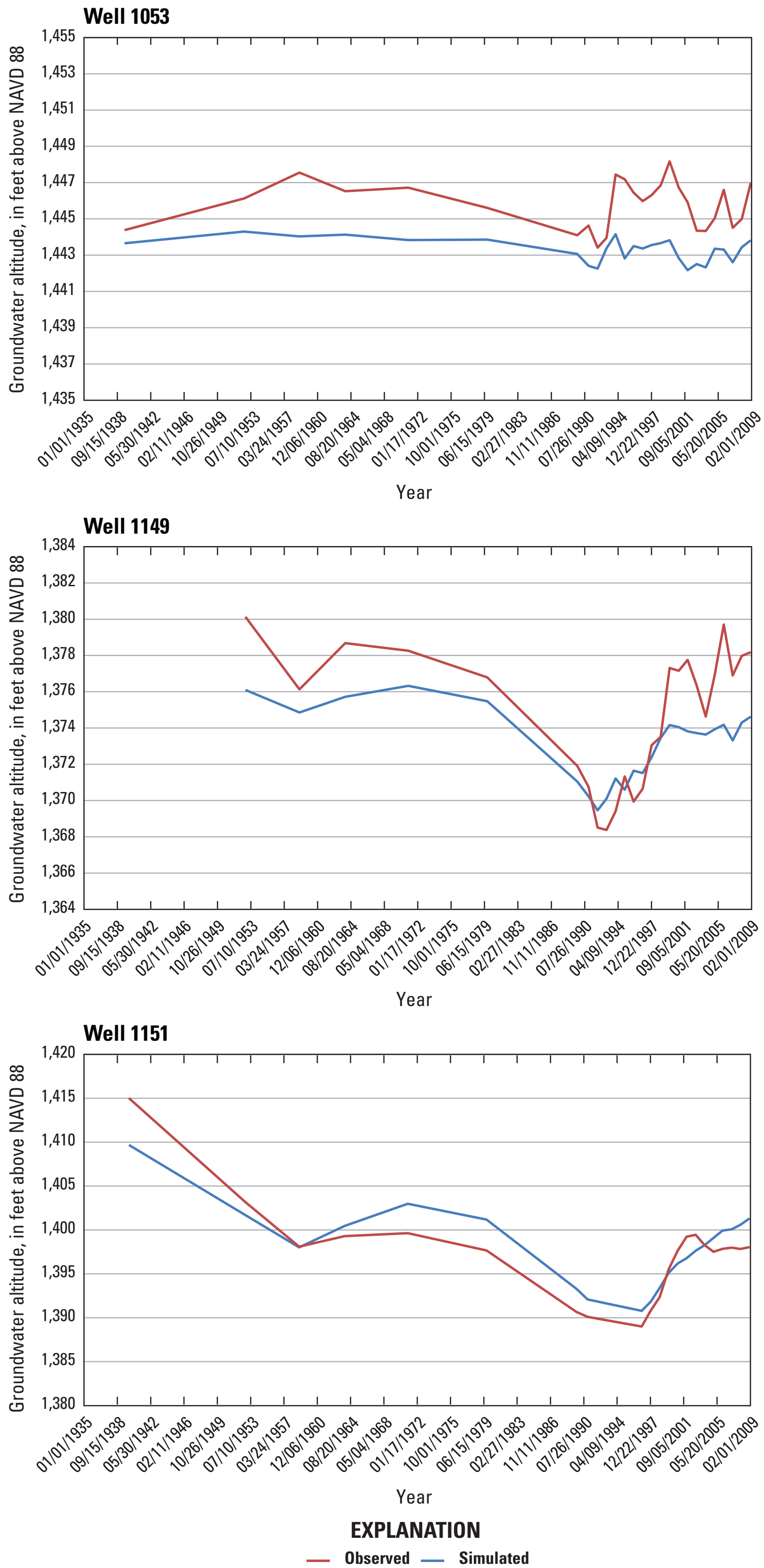



**Figure 40.** Simulated and observed groundwater levels for selected wells.—Continued

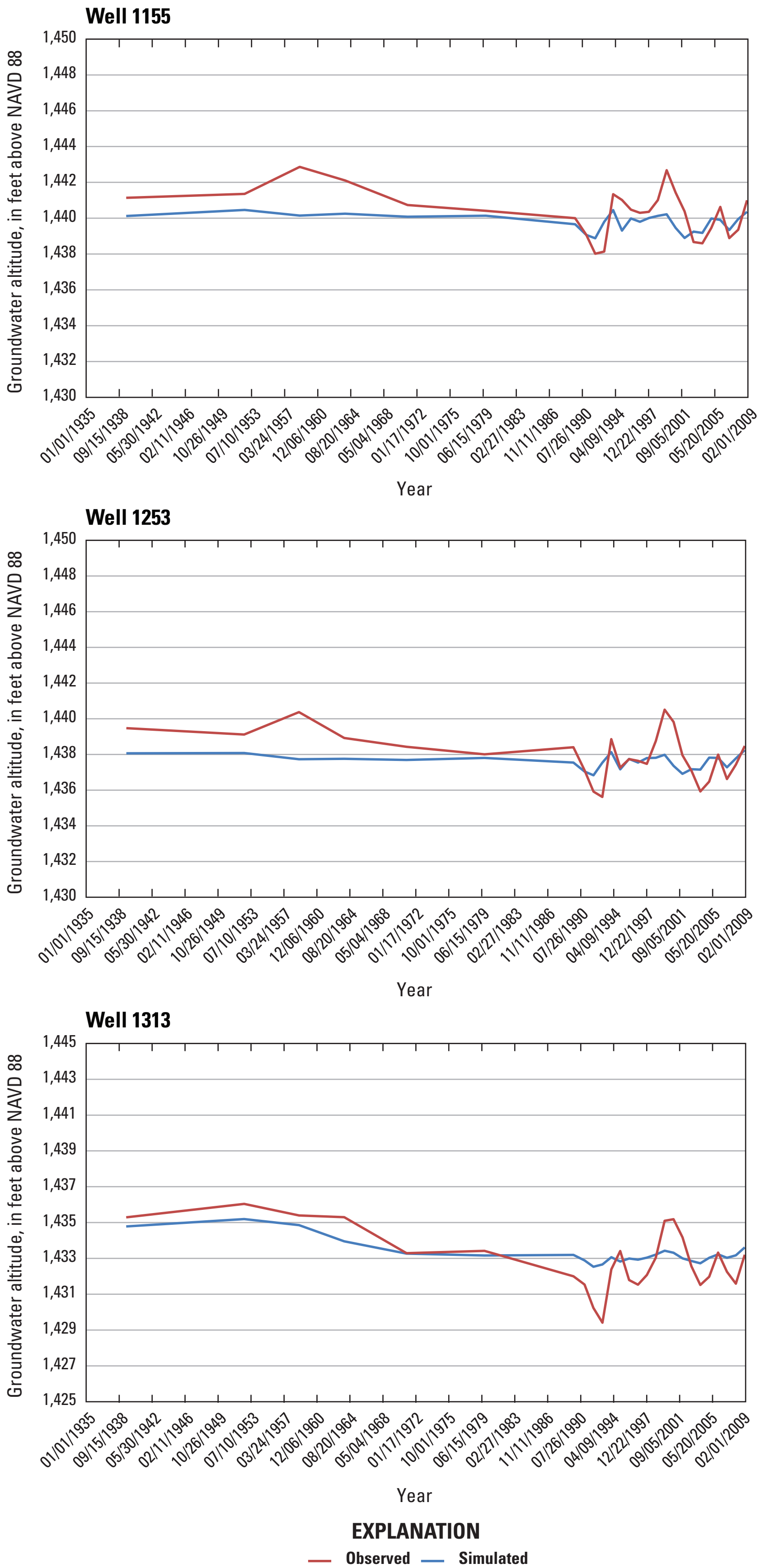



**Figure 40.** Simulated and observed groundwater levels for selected wells.—Continued

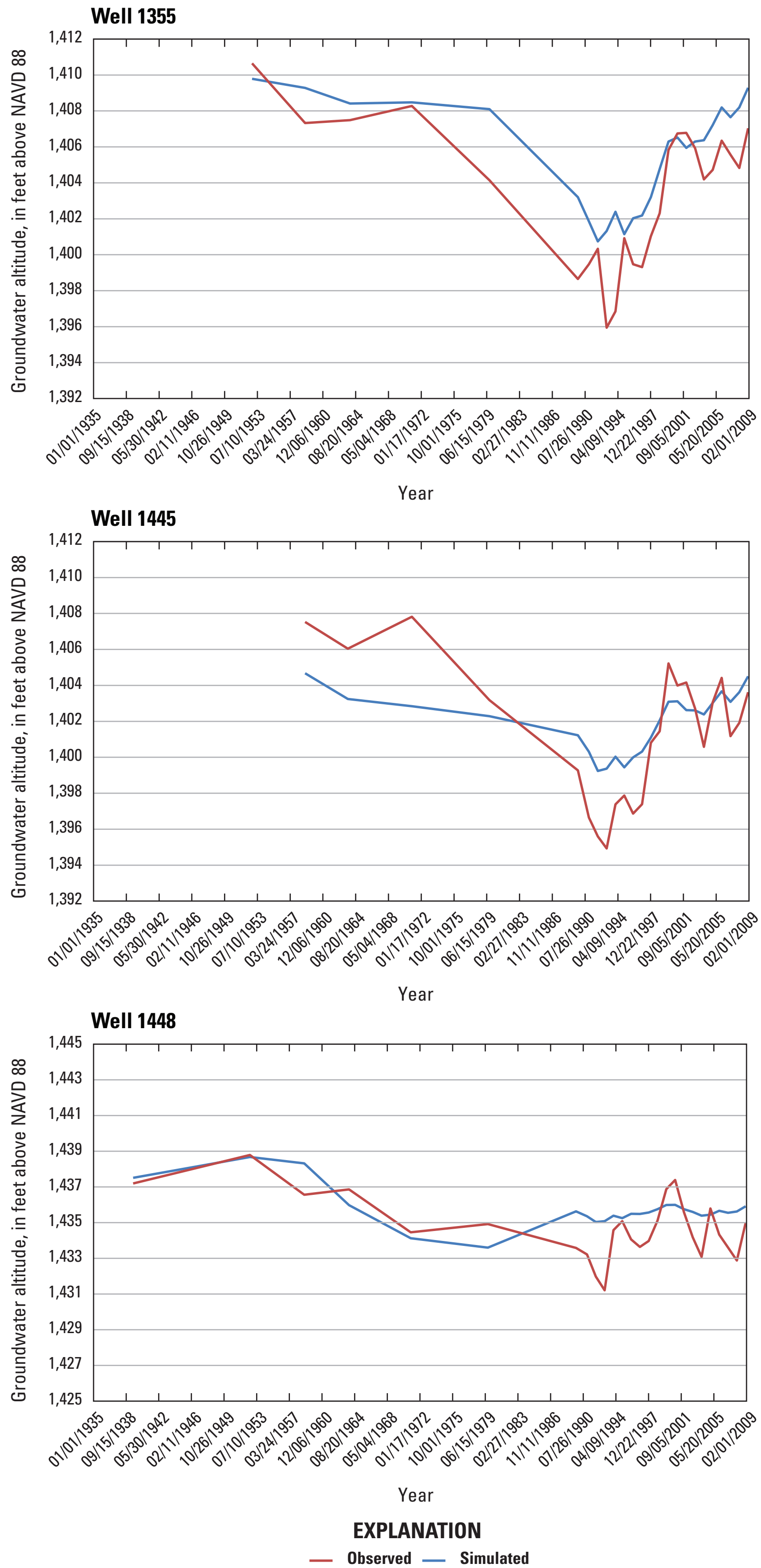



**Figure 40.** Simulated and observed groundwater levels for selected wells.—Continued

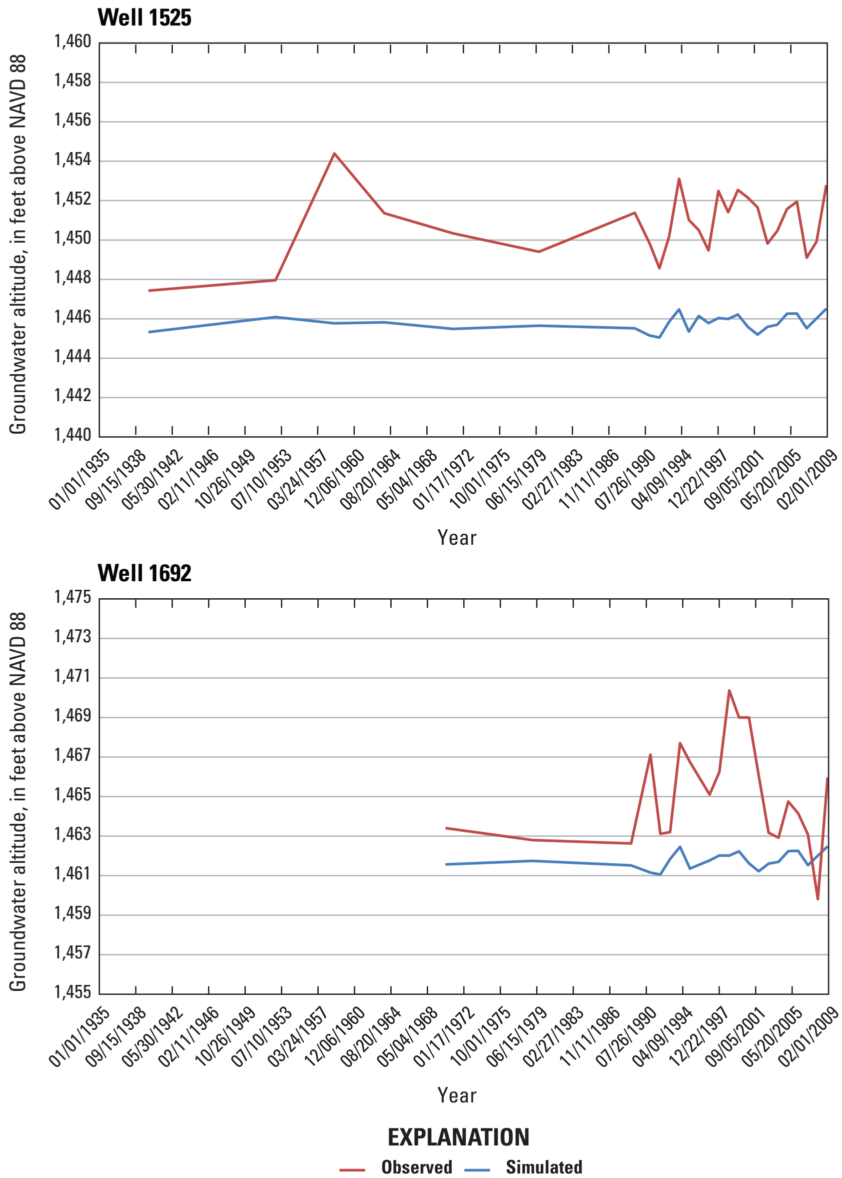



**Figure 40.** Simulated and observed groundwater levels for selected wells.—Continued

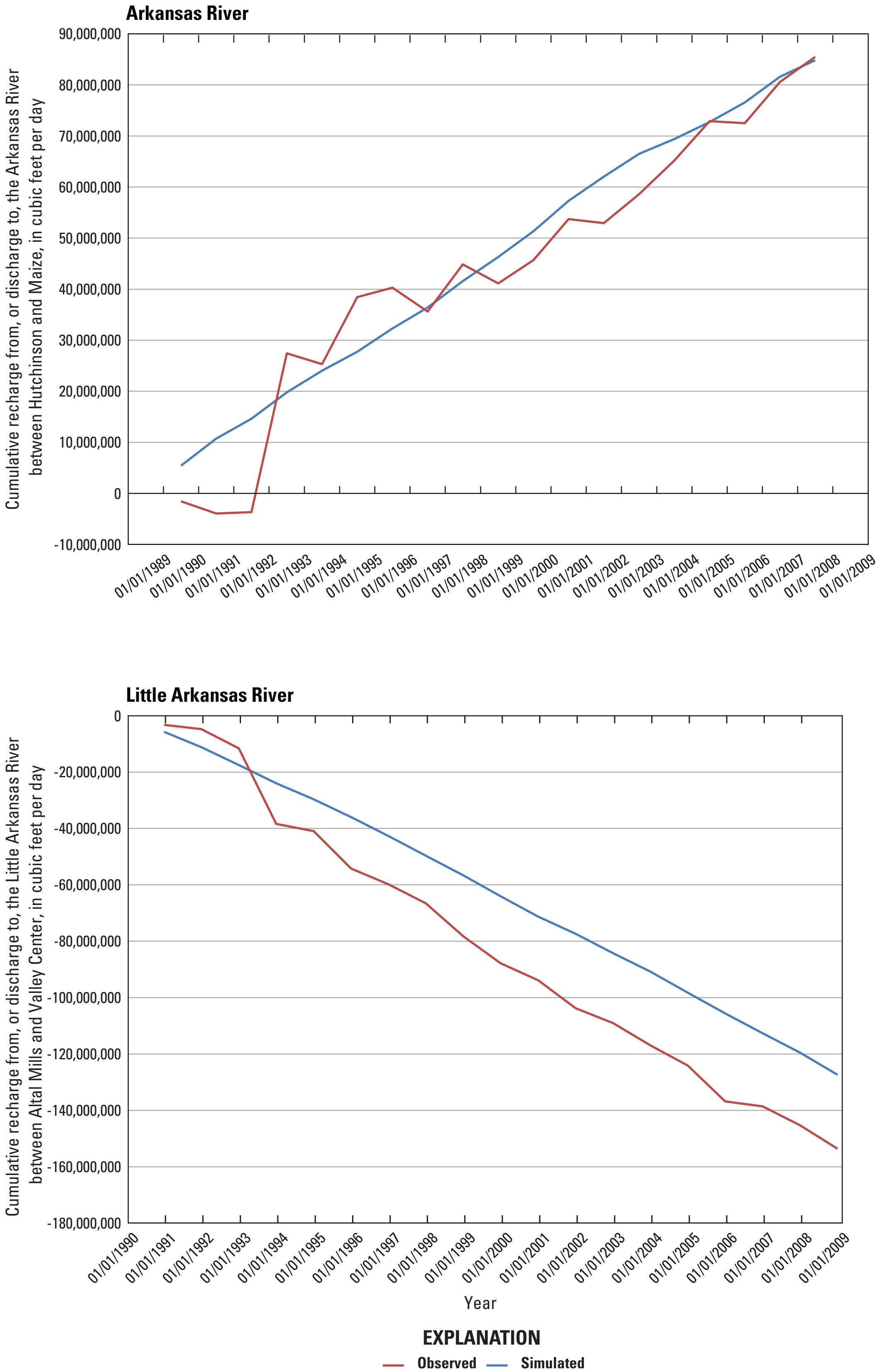



**Figure 41.** Observed and simulated cumulative streamflow gains and losses for the Arkansas and Little Arkansas Rivers for the transient simulation.

considered insensitive to that parameter. In addition, calculated sensitivities depend on the existence of observation data. If observations are not available in an area of the model, changes to a parameter may cause large changes to hydraulic head or flows, but the sensitivity of those parameters will not reflect the large effect they may have.

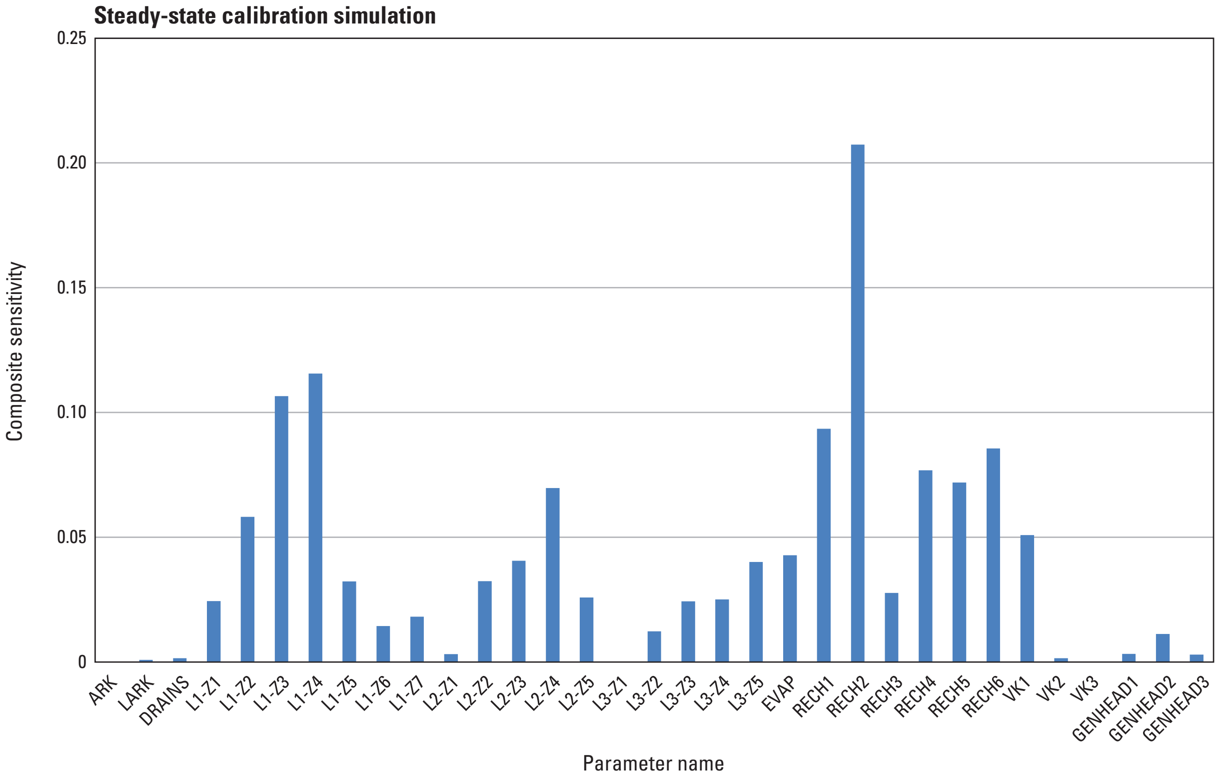
Composite scaled sensitivities were calculated by MODFLOW-2000 using dimensionless scaled sensitivities for all observations. The relative values of composite scaled sensitivities are used to indicate the total amount of information provided by the observations for the estimation of a parameter (Hill, 1998). Composite scaled sensitivities for selected parameters are shown for the steady-state and transient calibration simulations in figure 42. The model is more sensitive to a parameter with a large composite sensitivity value than to a parameter with a small value.

Composite sensitivities are smaller for the steady-state calibration simulation compared to the transient calibration simulation because there are fewer observations available in the steady-state simulation. In both simulations, parameters with larger composite sensitivities have a large areal distribution. For the steady-state simulation, the 10 parameters with the largest composite sensitivities are RECH2, L1-Z4, L1-Z3, RECH1, RECH6, RECH4, RECH5, L2-Z4, L1-Z2, and VK1. Recharge (fig. 27) and vertical conductance in model layer 1 (fig. 24) affect heads in all areas of the model. The hydraulic conductivity parameter zones L1-Z4, L1-Z3, and L1-Z2 (fig. 21) are present in the area of the Wichita well field and basin storage area. Hydraulic conductivity zone L2-Z4 (fig. 22) also is widely distributed. For the transient calibration simulation, the 10 parameters with the largest composite sensitivities are EVAP, RECH6, RECH5, L1-Z7, RECH4, L2-Z5, L1-Z6, L1-Z5, L1-Z4, and L3-Z5. For the transient calibration simulation, evapotranspiration and recharge affect heads in all areas of the model, and, as was indicated for the steady-state calibration simulation, hydraulic conductivity zones with a large distribution also have large composite sensitivities. The larger composite sensitivities for hydraulic conductivity parameters L2-Z5 and L3-Z5 are most likely because they are located near the Arkansas and Little Arkansas rivers and their value affects flow between the rivers and the aquifer.

One-percent scaled sensitivities are calculated by MODFLOW-2000 and approximately equal the amount that the simulated values would change if the parameter values increased by one percent (Hill, 1998). For observations related to flows between the aquifer and streams, positive sensitivities indicated an increase in flow from the river to the aquifer (or decrease in flow from the aquifer to the river); negative sensitivities indicated an increase in flow from the aquifer to the river or decrease in flow from the river to the aquifer. For groundwater-level observations, positive sensitivities indicate an increase in head with an increase in parameter value; negative sensitivities indicate a decrease in head with an increase in parameter value. Different 1-percent sensitivities for groundwater-level observations are caused by the proximity of the well to the area of the model that the parameter affects and its value. One-percent scaled sensitivities from the transient calibration simulation are shown in figure 43 for selected parameters and groundwater-level observations (observed and simulated hydrographs shown in fig. 40) and stream observations (observed and simulated hydrographs shown in fig. 41).

Scaled 1-percent sensitivities were positive for recharge for all groundwater observations and increasing recharge resulted in increased simulated groundwater levels. Scaled 1-percent sensitivities were positive and negative for hydraulic conductivity for groundwater-level observations, indicating the response of groundwater levels to changes in hydraulic conductivity is complex. Increasing hydraulic conductivity typically lowers groundwater levels and decreasing hydraulic conductivity raises groundwater levels; however, the opposite effect can occur when flow into and flow out of the aquifer is affected by aquifer hydraulic conductivity. For example, for well 733, increasing hydraulic conductivity in model layer 1 for L1-Z4, L1-Z5, and L1-Z6 causes simulated groundwater levels to increase, but increasing hydraulic conductivity for L1-Z1, L2-Z2, or L3-Z3 causes simulated groundwater levels to decrease. Increasing hydraulic conductivity in model layer 1 most likely increases the amount of recharge to model layer 2 where well 733 is screened, whereas increasing hydraulic conductivity in model layers 2 and 3 allows groundwater to flow more quickly in these model layers, thus lowering groundwater levels. Simulated groundwater levels increase near wells 1037 and 1038 when L1-Z4 is increased but decrease when L1-Z5 is increased. For these well locations, L1-Z4 defines the hydraulic conductivity in the area and there are numerous pumping wells. Increasing the hydraulic conductivity increases groundwater flow to the pumping wells and reduces drawdown in the area. Hydraulic conductivity parameter L1-Z5 is located adjacent to the Little Arkansas River. Increasing the value for L1-Z5 increases groundwater discharge to the Little Arkansas River and lowers groundwater levels in the area. For wells 1525 and 1692 that are located in the dune sand area, increasing recharge raises groundwater levels and and increasing evapotranspiration lowers groundwater levels. In this area, low values of hydraulic conductivity (fig. 21) limit the downward movement of groundwater, and recharge and evapotranspiration have the most effect on groundwater levels.

Flow between the Arkansas River and the *Equus* Beds aquifer is affected most by changes in recharge (fig. 27) and hydraulic conductivity (figs. 21, 22, and 23). Increasing recharge either increases flow from the aquifer to the Arkansas and Little Arkansas Rivers or decreases flow from the rivers to the aquifer. Increasing evapotranspiration has a large effect on the Arkansas River but a small effect on the Little Arkansas River. This is most likely the result of larger rates of evapotranspiration near the Arkansas River because of shallow depth to groundwater. Evapotranspiration is less near the Little Arkansas River because well pumping has increased depth to groundwater. Increasing hydraulic conductivity in areas near the rivers increases the rate of water flow between the rivers



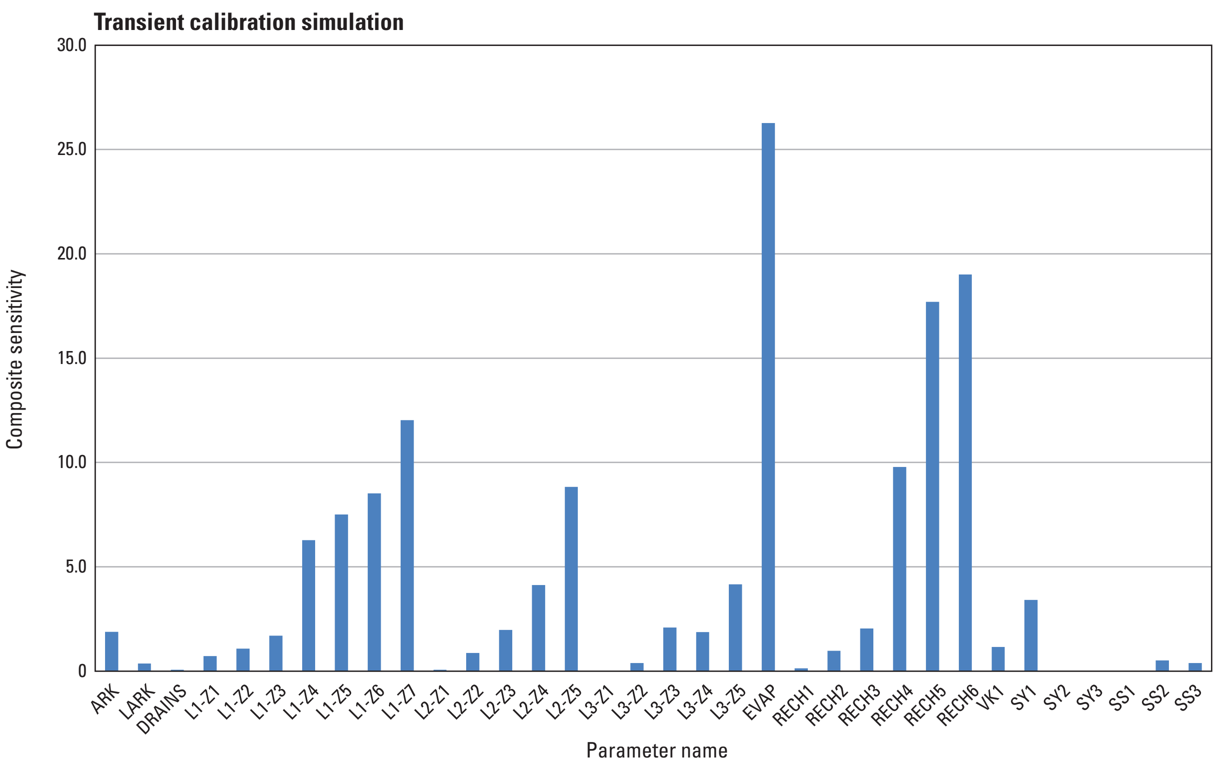




**Figure 42.** Composite scaled sensitivities for model input parameters from the steady-state and transient calibration simulations.

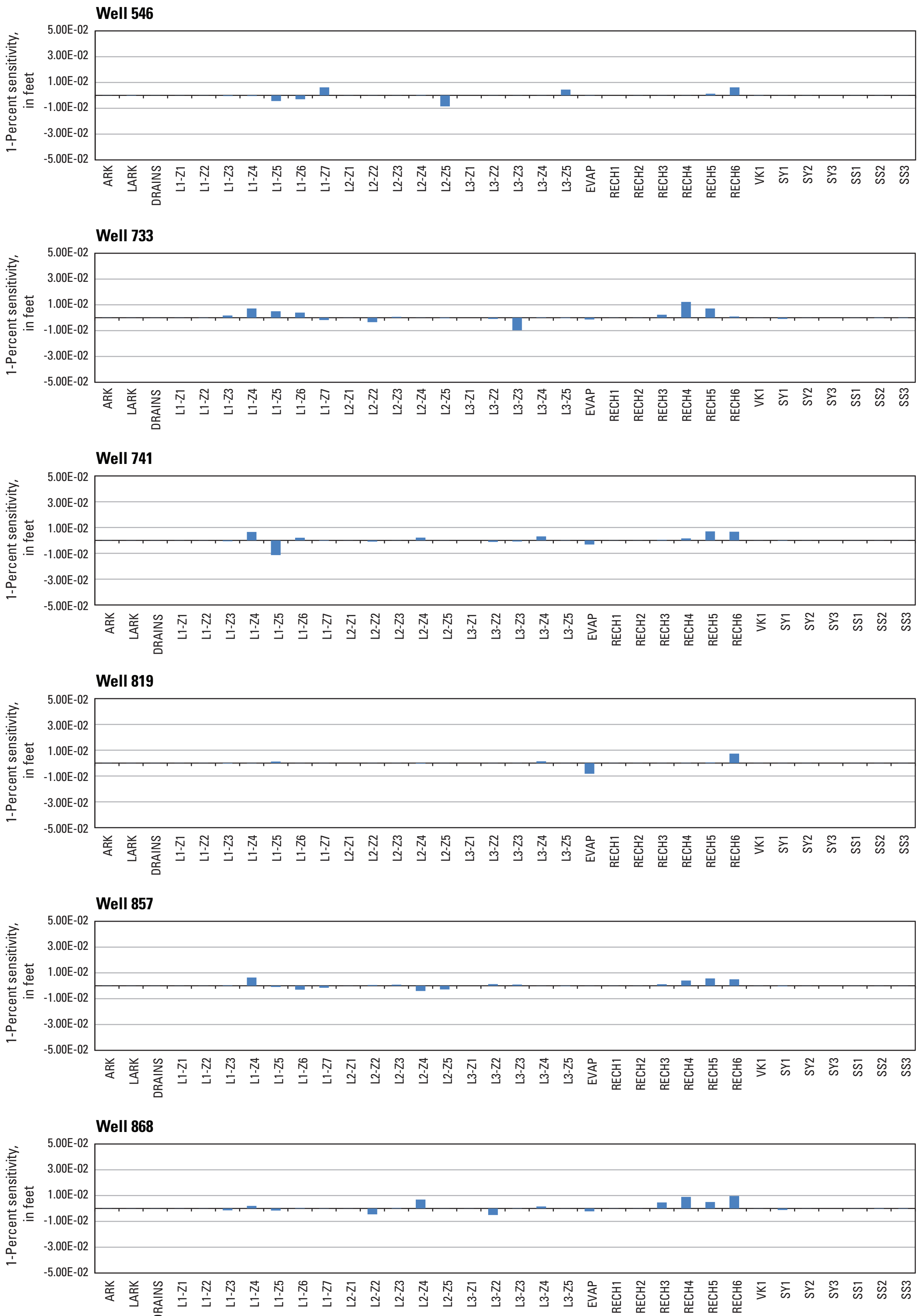



**Figure 43.** One-percent scaled sensitivities of parameters for selected stream and groundwater-level observations for the transient calibration simulation.

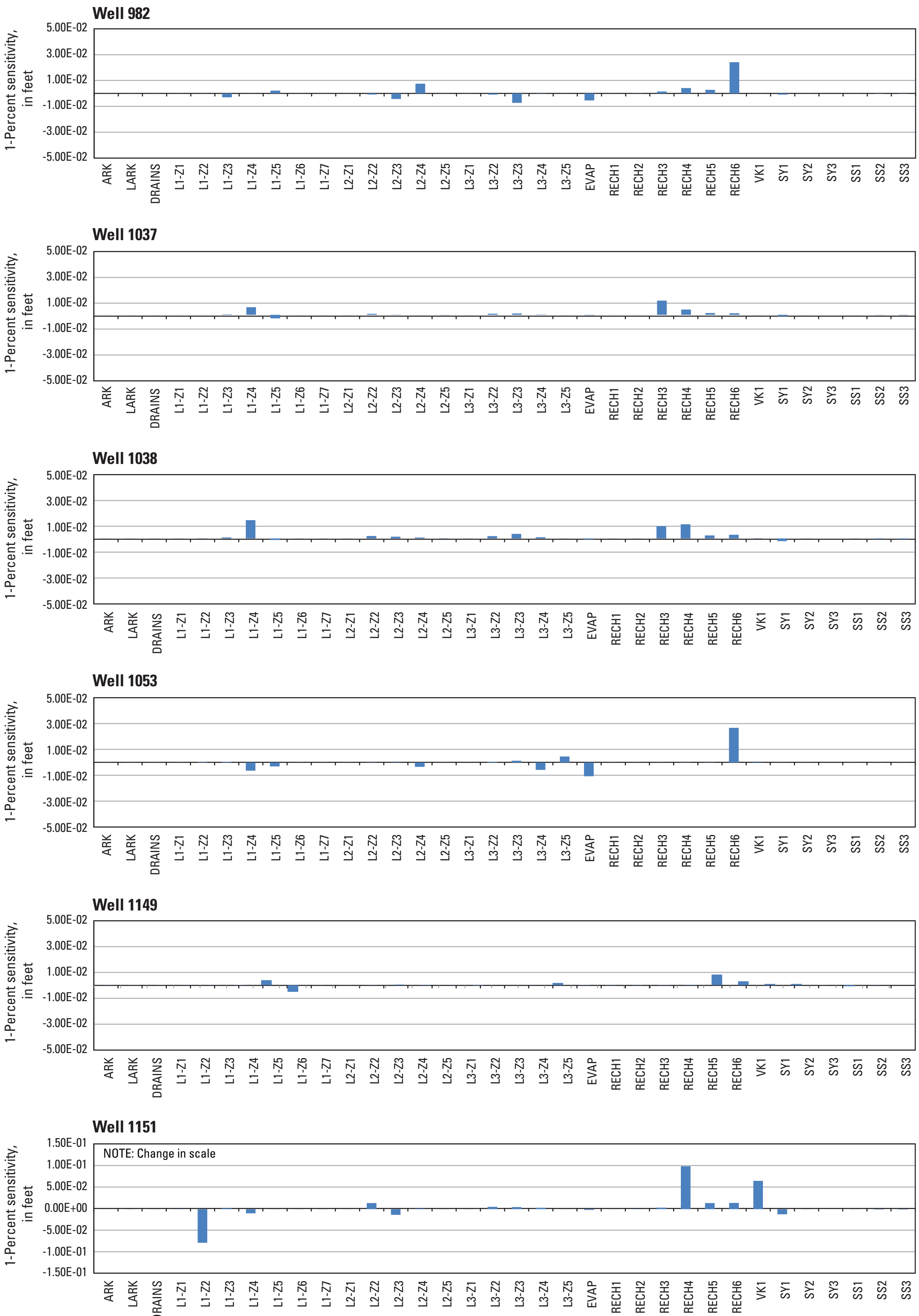



**Figure 43.** One-percent scaled sensitivities of parameters for selected stream and groundwater-level observations for the transient calibration simulation.—Continued

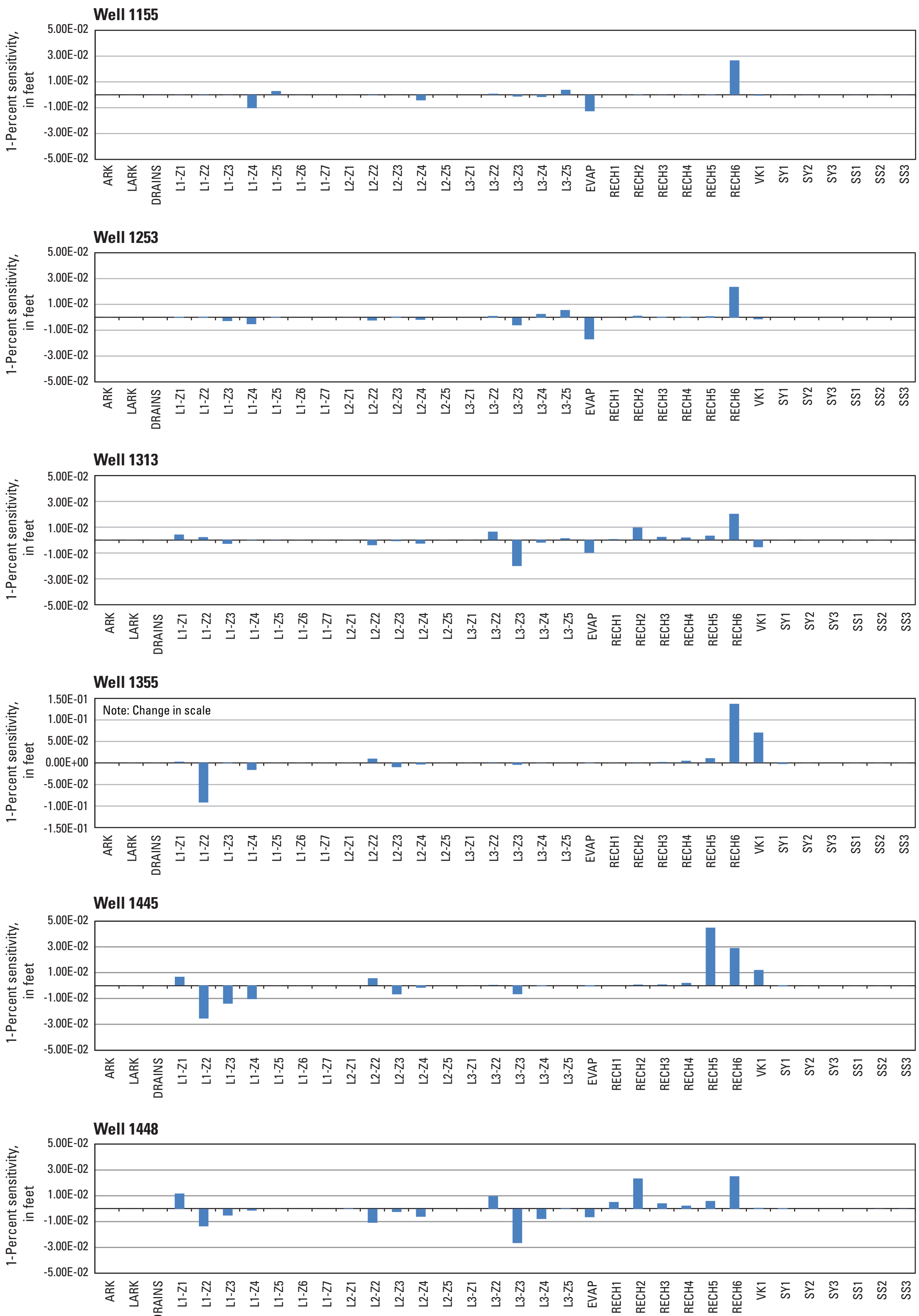



**Figure 43.** One-percent scaled sensitivities of parameters for selected stream and groundwater-level observations for the transient calibration simulation.—Continued

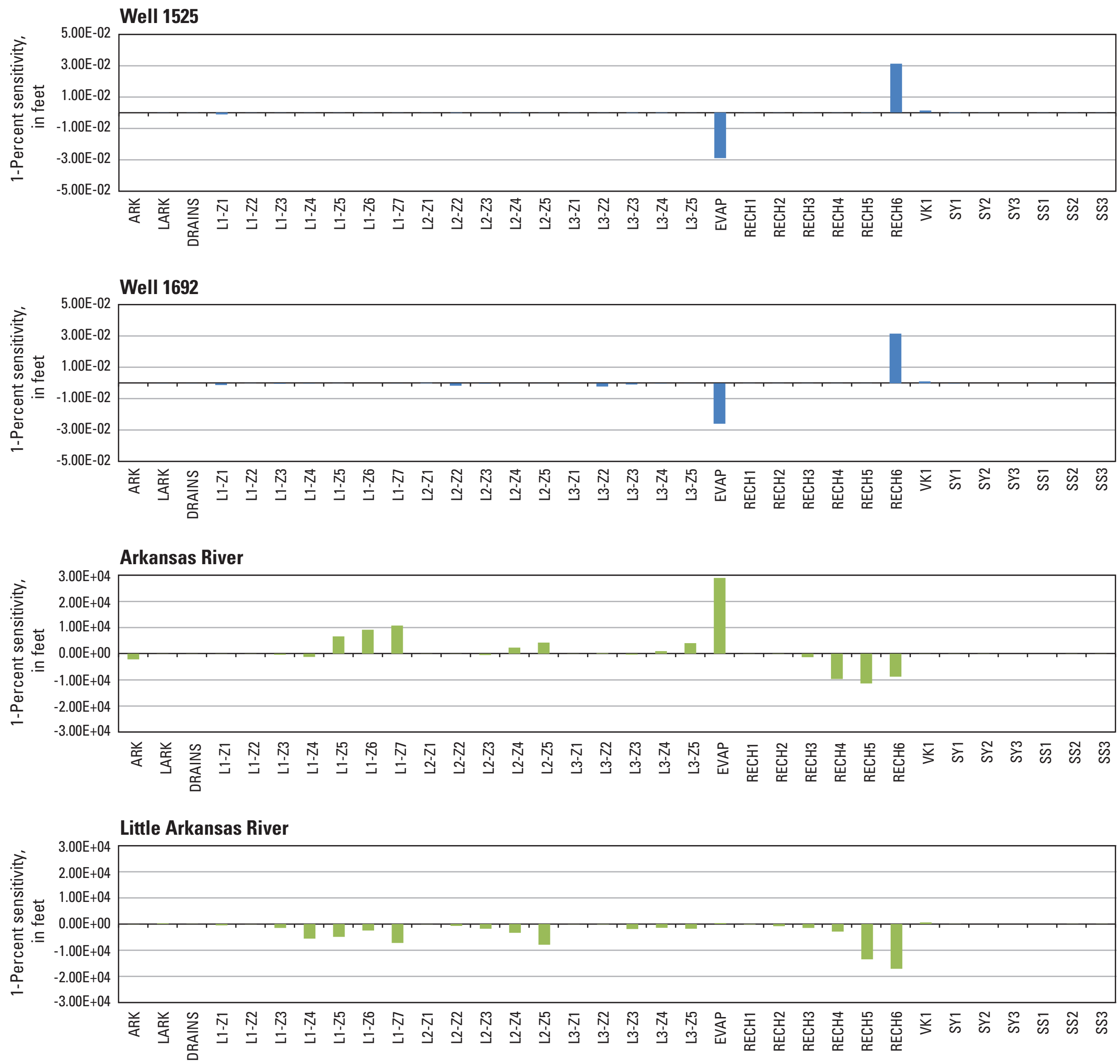



**Figure 43.** One-percent scaled sensitivities of parameters for selected stream and groundwater-level observations for the transient calibration simulation.—Continued

and the aquifer. For the Arkansas River, flow from the river to the aquifer increased with increasing hydraulic conductivity but for the Little Arkansas River, flow from the aquifer to the river is increased.

## Model Limitations

A groundwater model is a simplification of actual conditions. The accuracy of the groundwater model results depend on the accuracy of the input data and the accuracy of the equations used to characterize groundwater flow. The groundwater-flow model for this study was constructed with available hydrologic data to simulate groundwater flow in the *Equus* Beds aquifer in the study area. To correctly interpret model results, the following limitations of the model should be considered.

1. Model parameters such as hydraulic conductivity and recharge are applied uniformly to groups of model cells or zones. The assumption of uniformity likely is inaccurate because geologic materials and factors affecting groundwater flow are typically nonuniform.

2. The groundwater-flow model was discretized using a grid with cells measuring 400 ft by 400 ft. Model results were evaluated on a relatively large scale and cannot be used for detailed analyses such as simulating water-level drawdown near a single well. A grid with smaller cells would be needed for such detailed analysis.

3. The time discretization is too coarse to capture episodic floods or heavy precipitation events.

4. Recharge rates, evapotranspiration rates, specific yield, storage coefficient, and streambed conductance values are artifacts of model calibration. Field measurements of these parameters would provide more reliable values as model input.

5. The unsaturated zone, a part of the groundwater flow system overlying the aquifer, is not simulated.

6. Average annual rates for production well pumping were used in the groundwater-flow model. Average pumping rates may introduce error if water-level observations from monitoring wells located close to pumping wells are obtained when the wells are pumping at a rate that is different than the average used in the model. In this case, matching the simulated water levels to observed water levels during calibration may either overestimate or underestimate hydraulic properties of the aquifer near the monitoring well.

# Artificial-Recharge Accounting

The ability of the calibrated model to account for the additional water recharged to the *Equus* Beds aquifer as part of the ASR project was assessed using the USGS subregional water budget program ZONEBUDGET (Harbaugh, 1990), and by comparing those results to metered recharge for 2007 and 2008 and previous estimates of artificial recharge (Burns and McDonnel, 2008, 2009). A programming error in MODFLOW-2000 with respect to ZONEBUDGET calculations was corrected in MODFLOW-2005 (Harbaugh, 2005), which was used for artificial-recharge accounting simulations in this report. Model input is identical although there are small differences in components of the flow budget output between the two programs. ZONEBUDGET used the cell by cell flow data from MODFLOW to calculate water-flow budgets for each index cell (fig. 3) of the BSA within the groundwater model. Phase I of the ASR project was completed in 2006 and large-scale artificial recharge of the aquifer began at the phase I sites in March 2007. Groundwater flow was simulated from 1935 through 2008 and groundwater flow budgets were calculated for 2007 and 2008 for each index cell in the BSA. Initial conditions for the accounting simulations were obtained from the steady-state calibration simulation. For 1935 through 2006, the stress periods and stresses from the transient calibration simulation were used as model input. For 2007 and 2008, stress periods and stresses from the transient calibration simulation were used as model input for the artificial-recharge (AR) simulation and stress periods, and stresses from the transient calibration simulation, except for artificial-recharge well pumping, were used as model input for the no artificial-recharge (NAR) simulation. To calculate the effects of artificial recharge on groundwater flow in the BSA, results from the NAR simulation were subtracted from results from the AR simulation. With identical model input, except for artificial-recharge operation, the difference in simulated flows estimates the change in flows caused by artificial recharge. For transient groundwater flow, the rate of outflow equals the rate of inflow plus the rate that water is released from storage. The change in storage between the AR and NAR simulations is the volume of water that estimates the recharge credit for the aquifer storage and recovery system. Simulated groundwater flow budgets of the total modeled area for the AR and NAR simulations for 2006 through 2008 are listed in table 13. The amount of artificial recharge applied to each recharge basin and well in 2007 and 2008 and the BSA index cell where the basin or well is located is listed in table 14.

The change in storage between AR and NAR simulations for 2007 was 1,107 acre-ft and metered recharge was 963 acre-ft for the total model area. For 2008 the simulated change in storage was 684 acre-ft and metered recharge was 833 acre-ft. Total simulated change in storage was 1,790 acre-ft and total metered recharge was 1,796 acre-ft. Increased well pumping (inflow from artificial recharge and outflow from well pumping) is the largest difference between the AR and NAR simulations for 2007 and 2008 followed by changes in storage and river flows. Although pumping was larger in the AR simulation because of diversion wells located next to the Little Arkansas River near Halstead, Kans., the increased pumping was offset largely by increased flow into the model from the Little Arkansas River and decreased flow to the Little Arkansas River as groundwater that would have discharged to the river was intercepted by the pumping wells. The increased storage resulting from artificial recharge in the model was in the BSA where phase 1 artificial-recharge sites are located.

**Table 13.** Simulated groundwater flow budgets of the total modeled area for the artificial recharge accounting simulations, 2006–08.

[STO, Storage; RIV, River leakage; HDB, Head dependent boundaries; RCH, recharge; WEL, Well pumping; TOT, total; DRN, Drain; EVT, evapotranspiration; DIF, difference between inflow and outflow; PCT, percent difference; --, not applicable]

| Year | Inflow, in acre feet | | | | | | Outflow, in acre feet | | | | | | | DIF | PCT |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | STO | RIV | HDB | RCH | WEL | TOT | STO | DRN | RIV | EVT | HDB | WEL | TOT | | |
| Simulation with artificial recharge (AR) | | | | | | | | | | | | | | | |
| 2006 | 68,723 | 119,841 | 4,341 | 284,036 | 60 | 477,001 | 265 | 16,011 | 185,706 | 132,453 | 423 | 143,869 | 478,727 | -1,727 | -0.36 |
| 2007 | 531 | 159,983 | 4,329 | 342,131 | 1,159 | 508,134 | 74,858 | 17,813 | 142,505 | 151,827 | 471 | 120,037 | 507,512 | 622 | 0.12 |
| 2008 | 4,426 | 123,142 | 4,194 | 365,564 | 1,012 | 498,339 | 37,891 | 20,674 | 183,996 | 151,489 | 478 | 104,352 | 498,880 | -541 | -0.11 |
| Simulation with no artificial recharge (NAR) | | | | | | | | | | | | | | | |
| 2006 | 68,723 | 119,841 | 4,341 | 284,036 | 60 | 477,001 | 265 | 16,011 | 185,706 | 132,453 | 423 | 143,869 | 478,727 | -1,726 | -0.36 |
| 2007 | 538 | 159,740 | 4,330 | 342,131 | 29 | 506,768 | 73,757 | 17,810 | 142,993 | 151,788 | 471 | 119,114 | 505,934 | 834 | 0.16 |
| 2008 | 4,425 | 123,126 | 4,194 | 365,564 | 25 | 497,335 | 37,206 | 20,671 | 184,566 | 151,399 | 478 | 103,562 | 497,882 | -546 | -0.11 |
| Difference between simulations (AR–NAR) | | | | | | | | | | | | | | | |
| 2006 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 1 | 0 | -- |
| 2007 | -6 | 243 | | 0 | 1,130 | 1,366 | 1,101 | 2 | -488 | 39 | 0 | 923 | 1,578 | -211 | -- |
| 2008 | 1 | 15 | | 0 | 988 | 1,004 | 684 | 4 | -570 | 90 | 0 | 790 | 998 | 5 | -- |
| Total difference between simulations | | | | | | | | | | | | | | | |
| | -5 | 259 | | 0 | 2,118 | 2,370 | 1,785 | 6 | -1,058 | 129 | 0 | 1,714 | 2,576 | -206 | -- |

**Table 14.** Artificial recharge applied to each recharge basin and well for 2007 and 2008.

[--, not applicable]

| Artificial recharge site | Index cell | Gallons | Cubic feet | Acre-feet |
|---|---|---|---|---|
| **2007** | | | | |
| RB-1 | 9 | 17,679,732 | 2,010,186 | 46 |
| RB-2 | 14 | 55,799,656 | 6,344,421 | 146 |
| RRW1 | 2 | 42,039,084 | 4,779,844 | 110 |
| RRW2 | 5 | 72,554,308 | 8,249,425 | 189 |
| RRW3 | 5 | 78,903,186 | 8,971,292 | 206 |
| RRW4 | 5 | 101,798,312 | 11,574,468 | 266 |
| 2007 Total | -- | 368,774,278 | 41,929,635 | 963 |
| **2008** | | | | |
| RB-1 | 9 | 18,100,800 | 2,058,061 | 47 |
| RB-2 | 14 | 64,246,416 | 7,304,817 | 168 |
| RRW1 | 2 | 35,940,324 | 4,086,415 | 94 |
| RRW2 | 5 | 63,229,576 | 7,189,203 | 165 |
| RRW3 | 5 | 61,825,212 | 7,029,527 | 161 |
| RRW4 | 5 | 75,579,944 | 8,593,440 | 197 |
| 2008 Total | -- | 318,922,272 | 36,261,462 | 833 |
| Total | -- | 687,696,550 | 78,191,097 | 1,796 |

Knowledge of the change in groundwater storage from operation of the ASR is important as is the location of stored groundwater between index cells in the BSA. As water is recharged to the aquifer, it flows downgradient to the eventual point of discharge. To recover the recharged water to the aquifer without impacting existing water rights, the city of Wichita needs to know the location of the recharged water. Groundwater flow budgets for each index cell in the BSA are identical for the 2006 AR and NAR simulations. Groundwater flow budgets for each index cell for the AR and NAR simulations and the difference in flow budgets between simulations from 2007 and 2008 are listed in table 15, at the back of this report. As previously discussed, changes in groundwater storage between the AR and NAR simulations estimate the amount of recharge from operation of the ASR. Changes in storage for the AR and NAR simulations and previous estimates of artificial-recharge credits are listed for each index cell for the BSA in table 16.

As indicated in table 16, index cells where phase 1 recharge sites are located (2, 5, 9, and 14), or index cells near phase 1 recharge sites (1, 3, 4, 6, 8, 10, and 13) have the largest increase in storage. As expected, index cells located farthest away from recharge sites had little change in storage. Storage increases estimated in this study for 2007 and 2008 are 973 and 634 acre-ft, respectively, for a total increase in storage of 1,607 acre-ft in the basin storage area. The estimated increase in storage of 1,607 acre-ft compared to metered recharge of 1,796 acre-ft (table 14) indicates some loss of metered recharge. As listed in table 16, ASR operation caused 183 acre-ft of increased storage outside of the BSA, which accounts for all but 6 acre-ft or 0.33 percent of the total.

Previously estimated recharge credits for 2007 and 2008 (Burns and McDonnell Engineering Consultants, 2008, 2009) are 1,018 and 600 acre-ft, respectively, with a total estimated recharge credit of 1,618 acre-ft in the basin storage area. Storage changes calculated for this study are 4.42 percent less for 2007 and 5.67 percent more for 2008 than previous estimates. Total storage change for 2007 and 2008 is 0.68 percent less than previous estimates.

The small difference between the increase in storage from artificial recharge estimated with the groundwater-flow model and metered recharge indicates the groundwater model correctly accounts for the additional water recharged to the *Equus* Beds aquifer as part of the ASR project. Small percent differences between inflows and outflows for all stress periods and all index cells in the BSA, improved calibration compared to the previous model (table 10), and a reasonable match between simulated and measured long-term base flow indicates the groundwater model accurately simulates groundwater flow in the study area.

# Storage Volume Changes

The change in groundwater level through recent years compared to the August 1940 groundwater level map has been documented and used to assess the change of storage volume of the *Equus* Beds aquifer in and near the Wichita well field using a specific yield of 0.20 (Aucott and Myers, 1998a, 1998b; Hansen and Aucott, 2001, 2004, 2010; and Hansen, 2007). The specific yield is the volume of water that an unconfined aquifer releases from storage per unit surface area of aquifer per unit decline in the water table (Freeze and Cherry, 1979). This information has been used to assess changes to the hydrology of the aquifer and the effect of recharge, well pumping, and operation of the ASR system. Storage volumes have been estimated for three different areas: the central Wichita well field (CENWWF), the pre-2012 Wichita well field (P12WWF), and the current (2012) study area, the Wichita well field (WWF) (fig. 1). Two methods were used to estimate changes in storage from simulation results using the specific storage value from the groundwater flow model of 0.15. The first method used the simulated change in groundwater levels in model layer 1 between stress periods, and the second method used ZONEBUDGET to calculate the change in storage within the area of interest in the same way the effects of artificial recharge were estimated for index cells within the BSA. Simulated storage changes for these areas, changes in the groundwater flow budget, and previous storage estimates are compared for each area. A third method to estimate changes in storage used measured groundwater levels and is described in Hansen and Aucott (2010).

**Table 16.** Storage loss, storage gain, and storage change between artificial recharge and no artificial recharge simulations and previous estimates of artificial recharge credits.

[BSA, Basin Storage Area; -, not available]

| BSA Index cell | 2007 | | | 2008 | | | 2006 to 2008 | | | 2007 | 2008 | Total |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | Storage loss | Storage gain | Storage change | Storage loss | Storage gain | Storage change | Storage loss | Storage gain | Storage change | Recharge credit (Burns and McDonnel, 2008, 2009) | | |
| | Acre feet | | | | | | | | | | | |
| 1 | -1 | 37 | 39 | 0 | 18 | 18 | -1 | 55 | 56 | 0 | 0 | 0 |
| 2 | 1 | 93 | 93 | 0 | 34 | 34 | 1 | 127 | 127 | 123 | 60 | 183 |
| 3 | 0 | 6 | 5 | -2 | 17 | 18 | -1 | 22 | 24 | 177 | 190 | 367 |
| 4 | 0 | 66 | 67 | 0 | 74 | 74 | 0 | 140 | 141 | 0 | 0 | 0 |
| 5 | 0 | 278 | 278 | 0 | 105 | 105 | 0 | 384 | 383 | 197 | 104 | 301 |
| 6 | 0 | 43 | 43 | -1 | 37 | 38 | -1 | 80 | 81 | 52 | 9 | 61 |
| 7 | 0 | 2 | 2 | -1 | 3 | 4 | -1 | 5 | 6 | 6 | -6 | 0 |
| 8 | 0 | 39 | 39 | 0 | 67 | 67 | 0 | 105 | 105 | 0 | 0 | 0 |
| 9 | -1 | 237 | 238 | 1 | 136 | 135 | 0 | 372 | 373 | 121 | 97 | 218 |
| 10 | 0 | 23 | 23 | 0 | 34 | 34 | 0 | 57 | 57 | 45 | 16 | 61 |
| 11 | 0 | 3 | 3 | 0 | 7 | 8 | 0 | 10 | 11 | 17 | 8 | 25 |
| 12 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 1 | -1 | 0 |
| 13 | 0 | 11 | 11 | 0 | 19 | 19 | 0 | 30 | 30 | 0 | 0 | 0 |
| 14 | -1 | 78 | 78 | 0 | 57 | 57 | -1 | 135 | 135 | 241 | 53 | 295 |
| 15 | 0 | 9 | 9 | 0 | 18 | 18 | 0 | 27 | 27 | 22 | 21 | 43 |
| 16 | 0 | 2 | 2 | 0 | 4 | 4 | 0 | 6 | 6 | 9 | 12 | 21 |
| 17 | 0 | 1 | 1 | 0 | 0 | 0 | 0 | 1 | 1 | 1 | 5 | 6 |
| 18 | 0 | 2 | 2 | 0 | 3 | 3 | 0 | 6 | 6 | 0 | 0 | 0 |
| 19 | 0 | 5 | 5 | 0 | 7 | 7 | 0 | 13 | 13 | 0 | 0 | 0 |
| 20 | 0 | 3 | 3 | 0 | 7 | 7 | 0 | 10 | 10 | 2 | 5 | 7 |
| 21 | 0 | 2 | 2 | 0 | 1 | 1 | 0 | 3 | 3 | 2 | 10 | 12 |
| 22 | 0 | 2 | 2 | 0 | 0 | 0 | 0 | 1 | 1 | 0 | 4 | 4 |
| 23 | 0 | 1 | 1 | 0 | 0 | -1 | 0 | 0 | 0 | 0 | 1 | 1 |
| 24 | 0 | 1 | 1 | 0 | 0 | 0 | 0 | 1 | 1 | 0 | 0 | 0 |
| 25 | 0 | 2 | 2 | 0 | 0 | 0 | 0 | 2 | 2 | 0 | 2 | 2 |
| 26 | 0 | 2 | 2 | 0 | 0 | 0 | 0 | 2 | 2 | 0 | 1 | 1 |
| 27 | 0 | 1 | 1 | 0 | 0 | 0 | 0 | 2 | 2 | 0 | 5 | 6 |
| 28 | 0 | 2 | 2 | 0 | 0 | 0 | 0 | 1 | 1 | 0 | 2 | 2 |

**Table 16.** Storage loss, storage gain, and storage change between artificial recharge and no artificial recharge simulations and previous estimates of artificial recharge credits.—Continued

[BSA, Basin Storage Area; -, not available]

| BSA Index cell | 2007 | | | 2008 | | | 2006 to 2008 | | | 2007 | 2008 | Total |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | Storage loss | Storage gain | Storage change | Storage loss | Storage gain | Storage change | Storage loss | Storage gain | Storage change | Recharge credit (Burns and McDonnel, 2008, 2009) | | |
| | Acre feet—Continued | | | | | | | | | | | |
| 29 | 0 | 2 | 2 | 0 | -2 | -2 | 0 | 1 | 1 | 0 | 1 | 1 |
| 30 | 0 | 1 | 1 | 0 | -1 | -1 | 0 | 1 | 0 | 0 | 1 | 1 |
| 31 | 0 | 2 | 2 | 0 | -1 | -1 | 0 | 1 | 1 | 0 | 0 | 0 |
| 32 | 0 | 3 | 4 | 0 | -2 | -2 | 0 | 1 | 1 | 0 | 1 | 1 |
| 33 | 0 | 4 | 4 | 0 | -3 | -3 | 0 | 1 | 1 | 0 | 0 | 0 |
| 34 | 0 | 1 | 1 | 0 | -1 | -1 | 0 | 0 | 0 | 0 | 0 | 0 |
| 35 | 0 | 1 | 1 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| 36 | 0 | 1 | 1 | 0 | -1 | -1 | 0 | 1 | 0 | 0 | 0 | 0 |
| 37 | 0 | 2 | 2 | 0 | -2 | -2 | 0 | 0 | 0 | 0 | 0 | 0 |
| 38 | 0 | 1 | 1 | 0 | -1 | -1 | 0 | 0 | 0 | 0 | 0 | 0 |
| Outside BSA | -3 | 131 | 134 | 2 | 51 | 49 | -1 | 182 | 183 | -- | -- | -- |
| Total BSA | -3 | 970 | 973 | -1 | 633 | 634 | -4 | 1,603 | 1,607 | 1,018 | 600 | 1,618 |
| Total model area | -6 | 1,101 | 1,107 | 1 | 684 | 684 | -5 | 1,785 | 1,790 | -- | -- | -- |

Estimates of the change in saturated volume using simulated groundwater levels were calculated by recording the groundwater level for each stress period for each model cell, subtracting the groundwater level from one stress period to the next, multiplying the groundwater level change by the area of the model cell, and summing the volumes of all the model cells for each of the three areas of interest. The change in storage is the saturated volume change multiplied by the specific yield of 0.15.

Simulated groundwater level change maps were calculated by subtracting the simulated groundwater levels in model layer 1 for each stress period from simulated groundwater levels for model layer 1 at the end of stress period 1. The end of stress period 1 coincides with December 31, 1939, and is used as an analogue to the August 1940 groundwater level data for comparison to previous estimates. Storage volume estimated from previous studies for January 1993 and January 2006 are compared to simulated results for December 31, 1992, and December 31, 2005, respectively. Simulated and measured maps of changes in groundwater level since December 31, 1939, for December 31, 1992, December 31, 2005, and December 31, 2007, are shown in figures 44, 45, and 46.

The overall pattern of groundwater level change is similar between the simulated and measured groundwater level change maps. The simulated groundwater level change maps show less decline in the north part of the area compared to measured groundwater level changes for all times. This is caused by high simulated groundwater levels because of the low vertical hydraulic conductivities in this area, and lower measured groundwater levels because of the use of groundwater observations from below the layer of low vertical hydraulic conductivities for the measured groundwater level change maps. Simulated groundwater-level change from December 31, 1939, to December 31, 1992, (fig. 44) is about 10 ft less near the center of the greatest measured groundwater level change. Simulated groundwater level change from December 31, 1939, to December 31, 2005, (fig. 45) and from December 31, 1939, to December 31, 2008, (fig. 46) is about 5 ft less near the center of the greatest measured groundwater level change, and the total area of simulated groundwater level decline is slightly smaller. Several important differences between the simulated and measured groundwater level change maps exist in addition to those previously described. Simulated maps are based on average annual pumping rates and average annual recharge rates used within each stress period. If wells are pumping at a rate greater than average, water levels measured in wells close to those pumping wells will have a greater depth to water than would be the case if pumping were at an average annual rate. Often times, wells are pumped at a greater rate during summer or when rainfall is low and simulated depth to groundwater using average annual pumping rates will be less. Also, high rainfall or dry periods during the year will affect recharge and changes in water levels caused by those events will not be reflected in simulated water levels based on average annual recharge.

Simulated changes in saturated volumes and storage volumes, for the Central Wichita Wellfield, the pre-2012 Wichita Wellfield, and the current (2012) Wichita Wellfield study areas for each stress period and since December 31, 1939, December 31, 1992, and December 31, 2005, are listed in table 17 at the back of the report.

Changes in simulated storage were also calculated using a similar method to that used for artificial-recharge accounting. ZONEBUDGET (Harbaugh, 1990) used the cell by cell flow data from MODFLOW to calculate water-flow budgets for each area of interest. Groundwater flow budgets for each stress period are listed in table 18 for the CENWWF, in table 19 for the pre-2012 Wichita Wellfield, and in table 20 for the current (2012) Wichita Well field at the back of the report. Increases in storage are represented in the simulated groundwater flow budgets as increases in flow to storage or flow out of the groundwater system and into storage. Storage estimates were calculated using a specific yield of 0.15 and a storage coefficient of 0.0005. The storage coefficient defines the volume of water that an aquifer releases from storage per unit surface area of the aquifer per unit decline in the component of hydraulic head normal to that surface (Freeze and Cherry, 1979). Simulated changes in storage calculated from groundwater flow budgets for the Central Wichita Wellfield, the pre-2012 Wichita Wellfield, and the current (2012) Wichita Well field study areas for each stress period since December 31, 1939, December 31, 1992, and December 31, 2005, are listed in table 21 at the back of the report. The simulated groundwater flow budget for the pre-2012 Wichita Well field area is shown in figure 47.

Storage changes calculated from simulated groundwater flow budgets, simulated groundwater levels, and measured groundwater levels from Hansen and Aucott (2010) for the pre-2012 Wichita well field area from 1939 through 2008 are listed in table 22 at the back of the report and shown from 1990 through 2008 in figure 48. Simulated well pumping (fig. 47) and changes in storage (fig. 48) for the pre-2012 Wichita Well field shows the relation between changes in well pumping, recharge, and changes in estimated storage. In 1993 and from 1995 to 2000, decreased well pumping and increased recharge resulted in increased storage (flow out of the groundwater system and into storage).

The three methods used to estimate storage indicate similar trends although the magnitude of storage changes differ (fig. 48). The average ratio of storage changes estimated from simulated groundwater flow budgets using a specific yield of 0.15 to storage changes estimated from water level measurements using a specific yield of 0.20 is 0.63, and the average ratio of storage changes estimated from simulated groundwater levels using a specific yield of 0.15 to storage changes estimated from water level measurements using a specific yield of 0.20 is 0.45. Most of the difference can be explained by the use of a specific yield of 0.15 in the simulated estimates whereas the measured estimate uses a specific yield of 0.2. Thus, a 25 percent lower estimate for the simulated methods compared to the measured groundwater method is to be

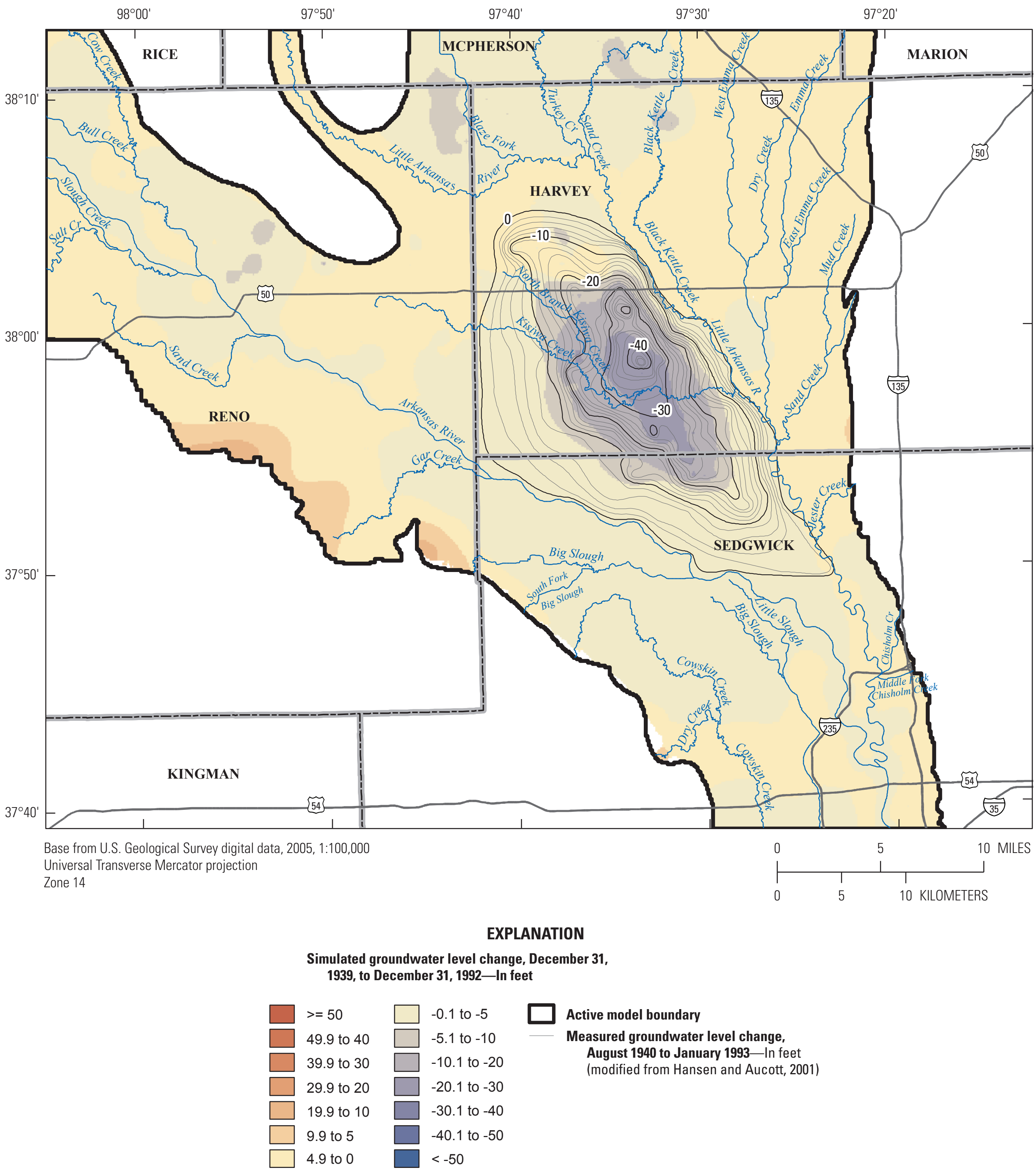
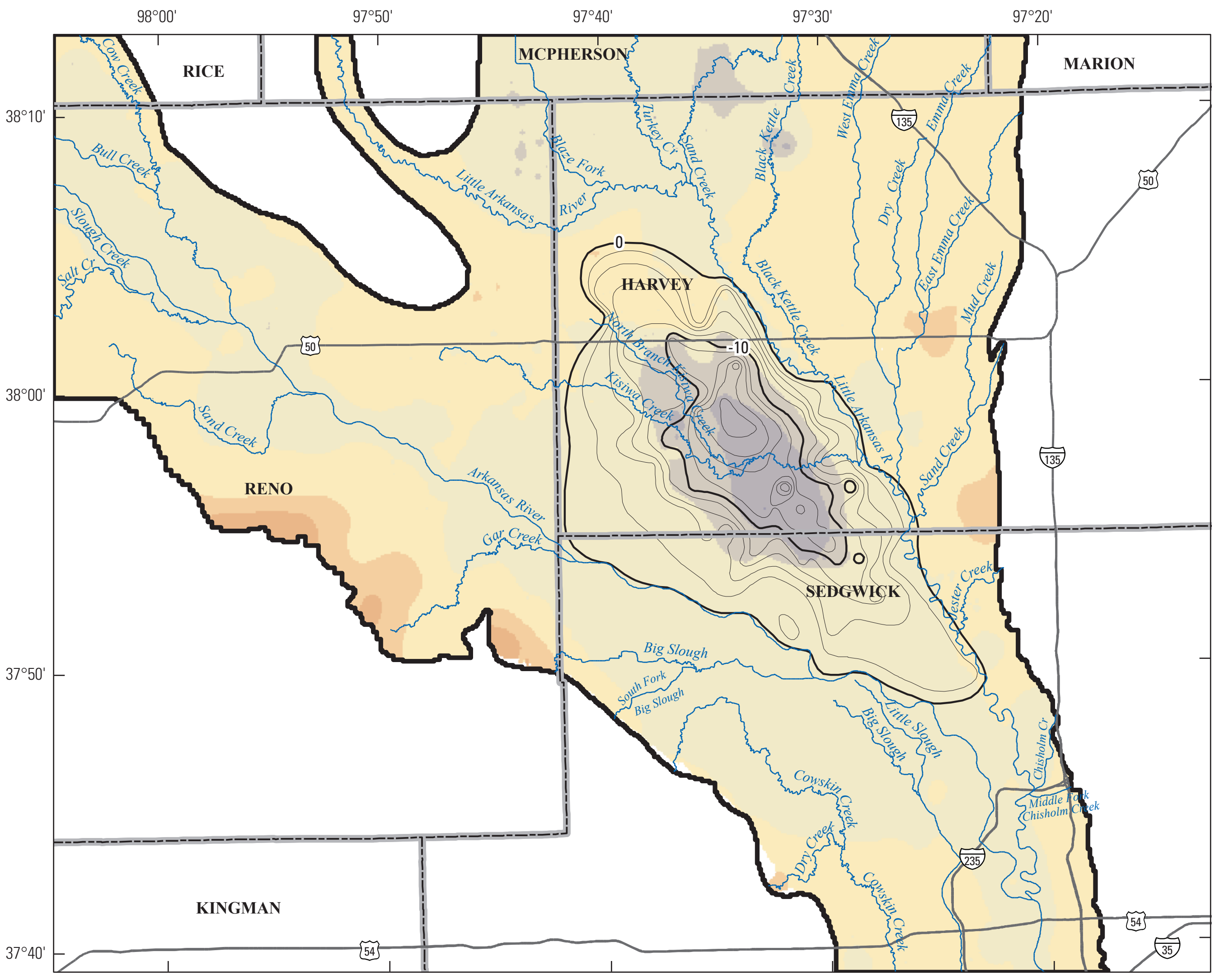



**Figure 44.** Simulated (December 31, 1939, to December 31, 1992) and measured (August 1940 to January 1993) groundwater level change.




Base from U.S. Geological Survey digital data, 2005, 1:100,000
Universal Transverse Mercator projection
Zone 14

**EXPLANATION**

**Simulated groundwater level change December 31, 1939, to December 31, 2005—In feet**

| | |
|---|---|
| >= 50 | -0.1 to -5 |
| 49.9 to 40 | -5.1 to -10 |
| 39.9 to 30 | -10.1 to -20 |
| 29.9 to 20 | -20.1 to -30 |
| 19.9 to 10 | -30.1 to -40 |
| 9.9 to 5 | -40.1 to -50 |
| 4.9 to 0 | < 50 |

**Active model boundary**

**Measured groundwater level change, August 1940 to January 2006**—In feet (modified from Hansen, 2007)

**Figure 45.** Simulated (December 31, 1939, to December 31, 2005) and measured (August 1940 to January 2006) groundwater level change.

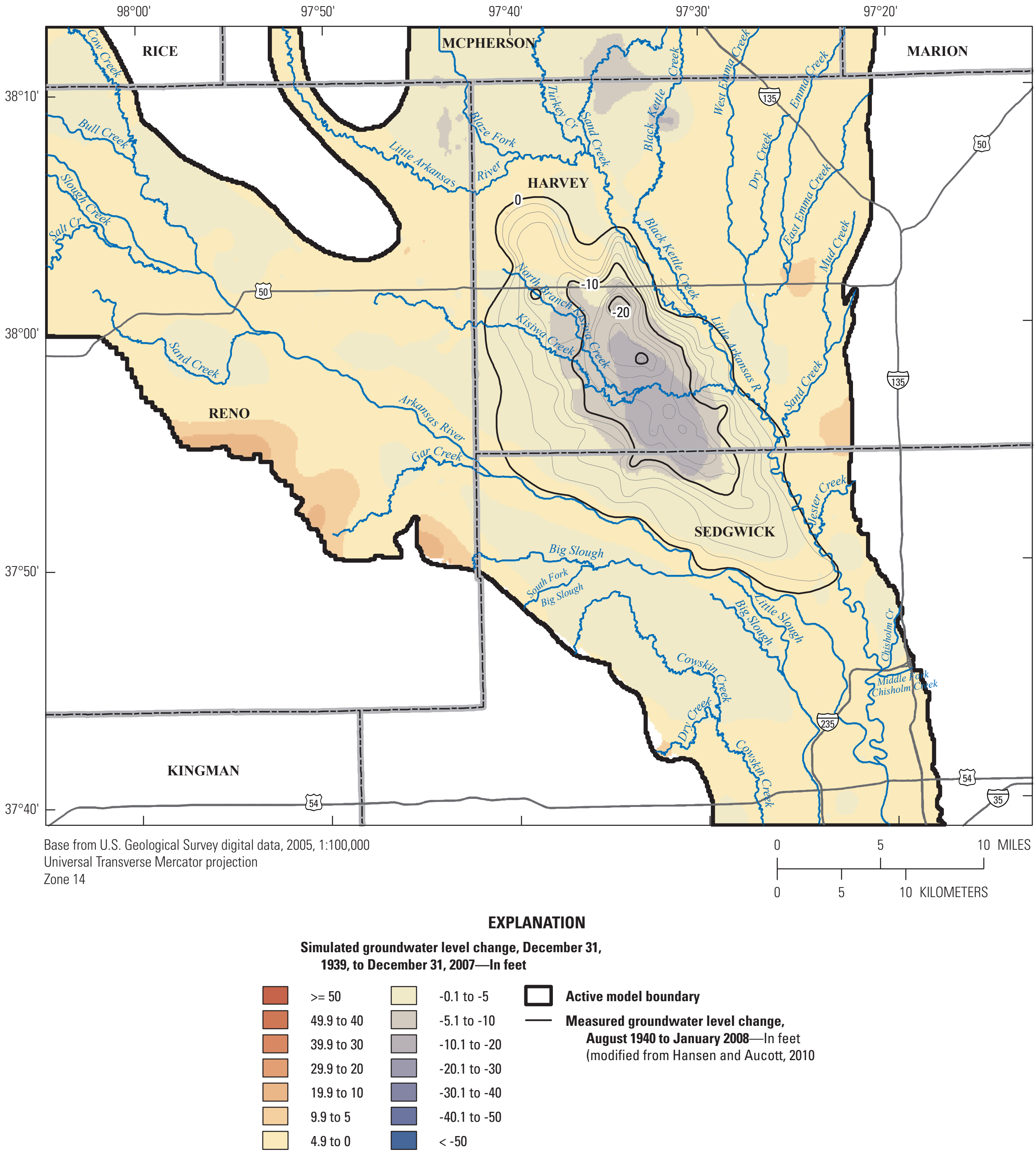



**Figure 46.** Simulated (December 31, 1939, to December 31, 2007) and measured (August 1940 to January 2008) groundwater level change.

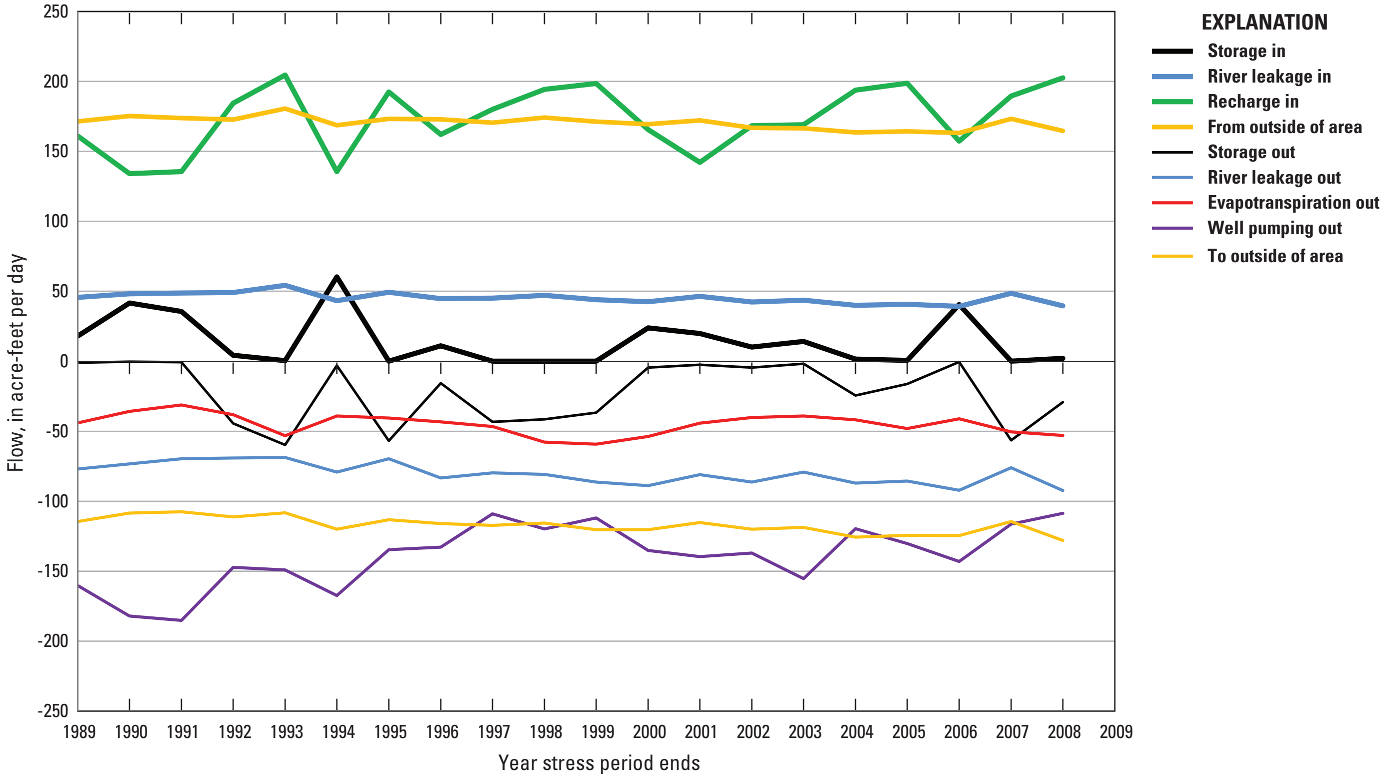



**Figure 47.** Simulated groundwater flow budget for the pre-2012 Wichita well field area, 1989 to 2008.

expected. The average ratio of storage changes estimated from simulated groundwater flow budgets using a specific yield of 0.15 to storage changes estimated from water level measurements using a specific yield of 0.15 is 0.84. The average ratio of storage changes estimated from simulated groundwater levels using a specific yield of 0.15 to storage changes estimated from water level measurements using a specific yield of 0.15 is 0.59.

Changes in estimated storage from the measured groundwater level method are greater than estimates from the simulated methods, in part, because the measured groundwater level method used water levels from the part of the aquifer that is below the low vertical hydraulic conductivity deposits of the dune sand area (fig 3). Including these water levels increased the estimate of storage change. Water levels are more variable from below the sand dune deposits than within them but changes in water levels in this part of the aquifer are most likely not caused by dewatering, and the use of specific yield for unconfined aquifers likely over-estimates changes in storage. In this part of the aquifer a confined storage value (0.0005) would be more appropriate. In contrast, both simulation methods include water levels from the dune sand area that are higher in altitude and water level changes (translated into changes in storage) that are much less variable because of the low vertical hydraulic conductivity of these deposits.

The average ratio of storage changes estimated from simulated groundwater levels to storage changes estimated from simulated groundwater budgets is 0.71. Storage estimated from the simulated groundwater levels used only water level changes from the upper part of the *Equus* Beds aquifer that included the low vertical hydraulic conductivity deposits of the dune sand area (fig. 4). In contrast, estimated storage changes from the groundwater flow budgets include storage changes from the upper part of the *Equus* Beds aquifer (including the dune sand area) and from lower parts of the aquifer (model layers 2 and 3). Both simulation methods include water levels from the dune sand area that are higher in altitude and water level changes (translated into changes in storage) that are much less variable because of the low vertical hydraulic conductivity of these deposits.

Information about the change in storage in response to hydrologic stresses is important for managing groundwater resources in the study area. The comparison between the three methods indicates similar storage change trends are estimated and each method is valid for estimating relative increases or decreases in storage. Use of groundwater level changes that do not include storage changes that occur in confined or semi-confined parts of the aquifer will slightly underestimate storage changes; however, use of specific yield and groundwater level changes to estimate storage change in confined or semi-confined parts of the aquifer will overestimate storage changes. Using only changes in shallow groundwater levels would provide more accurate storage change estimates for the measured groundwater levels method.

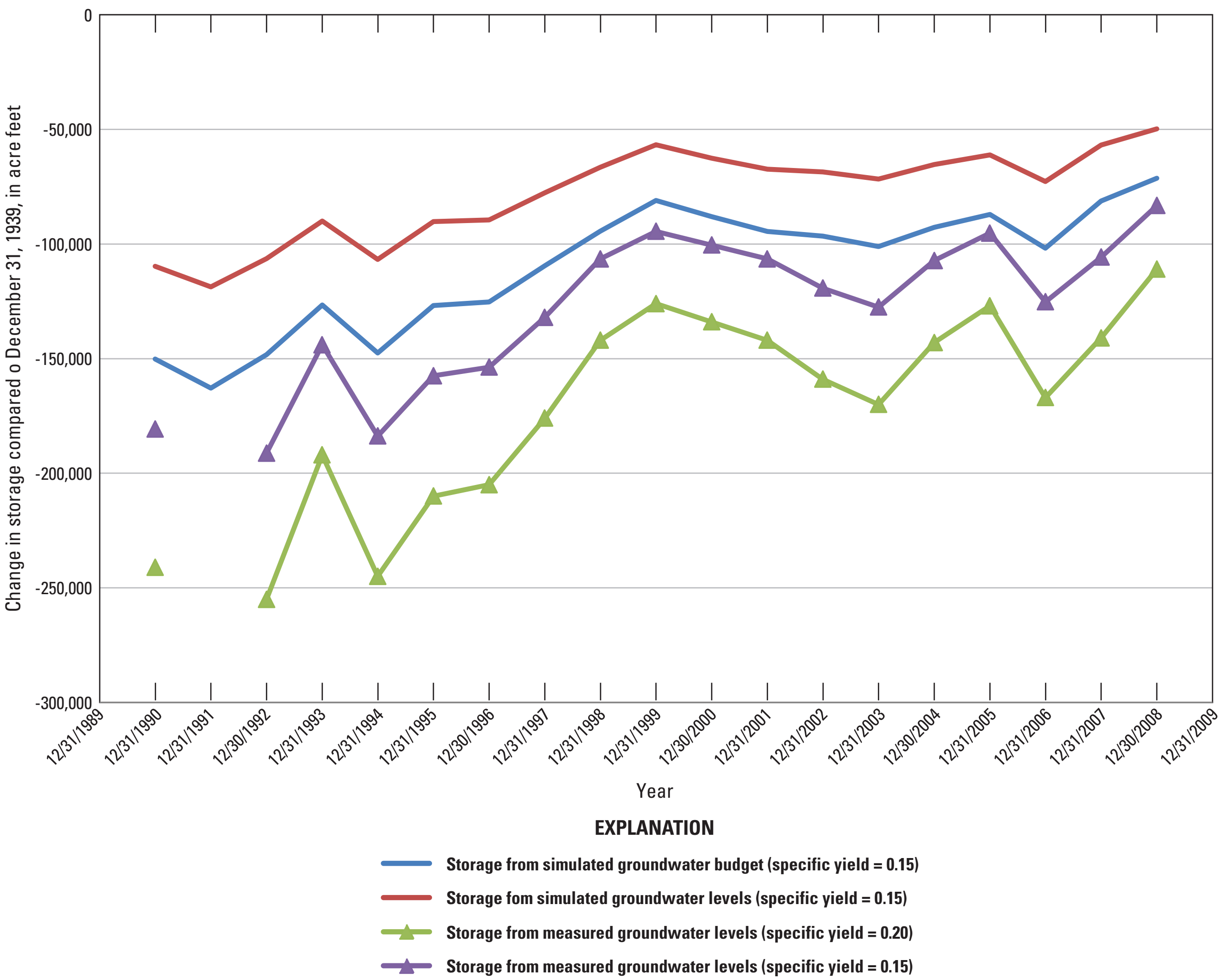



**Figure 48.** Storage changes calculated from simulated groundwater flow budgets, simulated groundwater levels, and measured groundwater levels (Hansen and Aucott, 2010) for the pre-2012 Wichita well field area from 1990 through 2008.

The value used for specific yield also is an important consideration when estimating storage. For the *Equus* Beds aquifer, the reported specific yield ranges between 0.08 and 0.35 (Williams and Lohman, 1949; Reed and Burnett, 1985; Spinazola and others, 1985; Fetter, 1988; Myers and others, 1996) and the storage coefficient (for confined conditions) ranges between 0.0004 and 0.16 (Reed and Burnett, 1985). Considering the importance of the value of specific yield and storage coefficient to estimates of storage change over time, and the wide range and substantial overlap for the reported values for specific yield and storage coefficient in the study area, further information on the distribution of specific yield and storage coefficient within the *Equus* Beds aquifer in the study area would greatly enhance the accuracy of estimated storage changes using both simulated groundwater level, simulated groundwater budget, or measured groundwater level methods.

## Summary

The *Equus* Beds aquifer is a primary water-supply source for Wichita, Kansas, and the surrounding area because of shallow depth to water, large saturated thickness, and generally good water quality. The well field developed by the city of Wichita in the *Equus* Beds aquifer during the 1940s and 1950s is one of the primary sources of water for the city and the surrounding area. Substantial water-level declines in the *Equus* Beds aquifer have resulted from pumping groundwater for agricultural and municipal needs, as well as periodic drought conditions. The lowest water levels to date were recorded in October 1992 and were as much as 50 ft lower than the predevelopment (1940) water levels in some locations. Water-level declines caused concern about the adequacy of the city's future water supply. Declining water levels likely represent a diminished water supply and may accelerate migration of saltwater from the Burrton oil field to the northwest and from the Arkansas River to the southwest into the freshwater of the *Equus* Beds aquifer.

In March 2006, the city of Wichita began construction of the *Equus* Beds Aquifer Storage and Recovery (ASR) project to store and later recover groundwater and to form a hydraulic barrier to the known chloride-brine plume near Burrton, Kans. Large-scale artificial recharge of the aquifer began at the phase I sites in March 2007. The phase I sites use water from the Little Arkansas River as the source of artificial recharge to the *Equus* Beds aquifer. In October 2009, the USGS, in cooperation with the city of Wichita, began a study to determine groundwater flow in the area of the Wichita well field and chloride transport from the Arkansas River and Burrton oilfield to the Wichita well field.

Groundwater flow was simulated for the *Equus* Beds aquifer using the three-dimensional finite-difference groundwater-flow model MODFLOW-2000. The primary study area includes the *Equus* Beds aquifer near the city of Wichita supply wells and encompasses the ASR phase I, II, and III artificial-recharge areas. The modeled area is almost 1,845 square miles and contains the entire study area. The model simulates steady-state and transient conditions. Steady-state conditions were simulated using average hydrologic conditions from 1935 through 1939. Transient conditions were simulated from 1935 to 2008. Hydrologic processes simulated include recharge, evapotranspiration, rivers, general head boundary conditions, and well pumping.

The groundwater-flow model was calibrated by adjusting model input data and model geometry until model results matched field observations within an acceptable level of accuracy. The steady-state calibration was used to test the conceptual model of groundwater flow, test the appropriateness of simulated boundary conditions, and obtain approximate hydraulic conductivity values and recharge rates. The transient calibration was used to fine tune the model hydraulic properties determined from the steady-state calibration and determine storage properties of the aquifer.

The root mean square (RMS) error for all water-level observations for the steady-state calibration simulation is 9.82 ft. The ratio of the RMS error to the total head loss in the model area is 0.049 (9.82 ft divided by 200 ft) or 4.9 percent. The mean error (observed minus simulated) for all 284 water-level observations is 3.86 ft. The difference between flow into the model and flow out of the model across all model boundaries was -0.08 percent of total flow for the steady-state calibration.

The RMS error for all water-level observations for the transient calibration simulation is 2.48 ft. The ratio of the RMS error to the total head loss in the model area is 0.0124 (2.48 ft divided by 200ft) or 1.24 percent. The mean error for all water level observation wells used in the transient calibration simulation is 0.03 ft. Simulated water levels follow the observed long-term trends for all wells, indicating the model adequately simulates long-term changes to groundwater levels resulting from sustained stresses on the aquifer such as overall rate of groundwater withdrawal, gains from and losses to streams, or long-term trends in recharge. Simulated short-term trends follow observed water level trends for most wells. Differences between simulated and observed short term water levels most likely are caused by observed water levels measured after large stresses such as precipitation events that occur in a shorter time interval than the model stress periods or heterogeneities in the aquifer material at these locations that are not incorporated into the model.

The RMS error calculated for observed and simulated base flow gains or losses for the Arkansas River for the transient simulation is 7,916,564 $ft^3$/day (91.6 $ft^3$/s) and the RMS error divided by the total range in streamflow (7,916,564/37,461,669 $ft^3$/day) is 22 percent. The RMS error calculated for observed and simulated streamflow gains or losses for the Little Arkansas River for the transient simulation is 5,610,089$ft^3$/day (64.9 $ft^3$/s) and the RMS error divided by the total range in streamflow (5,610,089/41,791,091 $ft^3$/day) is 13 percent. The RMS values are less than the maximum measurement errors and the RMS error divided by the total range in streamflow are less than 25 percent, indicating the acceptability of the simulated streamflow gains or losses in the transient calibrated model. The mean error between observed and simulated base flow gains or losses was 29,999 $ft^3$/day (0.34 $ft^3$/s) for the Arkansas River and -1,369,250 $ft^3$/day (-15.8 $ft^3$/s) for the Little Arkansas River. Cumulative streamflow gain and loss observations are similar to the cumulative simulated equivalents. Average percent mass balance difference for individual stress periods ranged from -0.46 to 0.51 percent. The cumulative mass balance for the transient calibration was 0.01 percent.

Composite scaled sensitivities are calculated by MODFLOW-2000 using dimensionless scaled sensitivities for all observations and indicate the total amount of information provided by the observations for the estimation of a parameter. The simulations are most sensitive to parameters with a large areal distribution. For the steady-state calibration, these include the recharge, hydraulic conductivity, and the vertical conductance. For the transient simulation, these include evapotranspiration, recharge, and hydraulic conductivity.

The ability of the calibrated model to account for the additional groundwater recharged to the *Equus* Beds aquifer as part of the ASR project was assessed using the USGS subregional water budget program ZONEBUDGET, and by comparing those results to metered recharge for 2007 and 2008 and previous estimates of artificial recharge. ZONEBUDGET used the cell by cell flow data from MODFLOW to calculate water-flow budgets for each index cell of the basin storage area (BSA) and the total model area. Initial conditions for the accounting simulations were obtained from the steady-state calibration simulation. For 1935 through 2006, the stress periods and stresses from the transient calibration simulation were used as model input. For 2007 and 2008, stress periods and stresses from the transient calibration simulation were used as model input for the artificial-recharge simulation, and stress periods and stresses from the transient calibration simulation, except for artificial-recharge well pumping, were used as model input for the no artificial-recharge simulation. The change in storage between simulations is the volume of water that estimates the recharge credit for the aquifer storage and recovery system.

The simulated change in storage between simulations for 2007 was 1,107 acre-ft and metered recharge was 963 acre-ft in the total model area. For 2008 the simulated change in storage was 684 acre-ft and metered recharge was 833 acre-ft. Total simulated change in storage was 1,790 acre-ft and total metered recharge was 1,796 acre-ft. The increased storage resulting from artificial recharge in the model was in the BSA where phase 1 artificial-recharge sites are located. The estimated increase in storage of 1,607 acre-ft in the BSA compared to metered recharge of 1,796 acre-ft indicates some loss of metered recharge. Increased storage outside of the BSA of 183 acre-ft accounts for all but 6 acre-ft or 0.33 percent of the total. Previously estimated recharge credits for 2007 and 2008 are 1,018 and 600 acre-ft, respectively, and a total estimated recharge credit of 1,618 acre-ft. Storage changes calculated for this study are 4.42 percent less for 2007 and 5.67 percent more for 2008 than previous estimates. Total storage change for 2007 and 2008 is 0.68 percent less than previous estimates. The small difference between the increase in storage from artificial recharge estimated with the groundwater-flow model and metered recharge indicates the groundwater model correctly accounts for the additional water recharged to the *Equus* Beds aquifer as part of the ASR project. Small percent differences between inflows and outflows for all stress periods and all index cells in the BSA, improved calibration compared to the previous model, and a reasonable match between simulated and measured long-term base flow indicates the groundwater model accurately simulates groundwater flow in the study area.

The change in groundwater level through recent years compared to the August 1940 groundwater level map has been documented and used to assess the change of storage volume of the *Equus* Beds aquifer in and near the Wichita well field. Storage volumes have been estimated for three different areas: the Central Wichita Wellfield, the pre-2012 Wichita Wellfield, and the current (2012) study area, the Wichita Wellfield. Two methods were used to estimated changes in storage from simulation results. The first method used the simulated change in groundwater levels in model layer 1 between stress periods, and the second method used ZONEBUDGET to calculate the change in storage within the area of interest in the same way the effects of artificial recharge were estimated for Index Cells within the BSA. The third method used measured groundwater levels.

The three methods used to estimate storage indicate similar trends although the magnitude of storage changes differ. The average ratio of storage changes estimated from simulated groundwater flow budgets to storage changes estimated from water level measurements is 0.63 and the average ratio of storage changes estimated from simulated groundwater levels to storage changes estimated from water level measurements is 0.45. Most of the difference can be explained by the use of a specific yield of 0.15 in the simulated estimates whereas the measured estimate uses a specific yield of 0.2. Thus, a 25 percent lower estimate for the simulated methods compared to the measured groundwater method is to be expected. In addition, estimates of storage from the measured groundwater level method used water levels from the part of the aquifer that is below the low vertical hydraulic conductivity deposits of the dune sand area. Using these water levels has the effect of increasing the estimate of storage change. Water levels are more variable from below the sand dune deposits than within them but changes in water level in this part of the aquifer most likely are not caused by dewatering, and the use of specific yield for unconfined aquifers likely overestimates changes in storage. In contrast, simulation methods include water levels from the dune sand area that are higher in altitude and water level changes (translated into changes in storage) that are much less variable because of the low vertical hydraulic conductivity of these deposits.

Information about the change in storage in response to hydrologic stresses is important for managing groundwater resources in the study area. The comparison between the three methods indicates similar storage change trends are estimated and each could be used to determine relative increases or decreases in storage. Use of groundwater level changes that do not include storage changes that occur in confined or semi-confined parts of the aquifer will slightly underestimate storage changes. However, use of specific yield and groundwater level changes to estimate storage change in confined or semi-confined parts of the aquifer will overestimate storage changes. Using only changes in shallow groundwater levels would provide more accurate storage change estimates for the measured groundwater levels method.

The value used for specific yield is also an important consideration when estimating storage. For the *Equus* Beds aquifer the reported specific yield ranges between 0.08 and 0.35 and the storage coefficient (for confined conditions) ranges between 0.0004 and 0.16. Considering the importance of the value of specific yield and storage coefficient to estimates of storage change over time, and the wide range and substantial overlap for the reported values for specific yield and storage coefficient in the study area, further information on the distribution of specific yield and storage coefficient within the *Equus* Beds aquifer in the study area would greatly enhance the accuracy of estimated storage changes using both simulated groundwater level, simulated groundwater budget, or measured groundwater level methods.

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# Tables

**Table 2.** Parameter name, hydraulic property, model layer, zone, and calibrated parameter value. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_2.xlsx*.

**Table 3.** Weather stations, periods of data, and average precipitation for each stress period. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_3.xlsx*.

**Table 6.** Corrections made to monthly pumpage values obtained from city of Wichita. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_6.xlsx*.

**Table 7.** Well number, date of observation, observed water level, and simulated water level of each well used in the steady-state calibration. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_7.xlsx*.

**Table 9.** Well number, date of observation, simulated water level, and observed water level of each well used in the transient calibration. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_9.xlsx*.

**Table 11.** Observed and simulated streamflow gains or losses for the Arkansas and Little Arkansas Rivers for each stress period of the transient simulation. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_11.xlsx*.

**Table 12.** Average flow rates and cumulative flows for each stress period of the transient calibration simulation. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_12.xlsx*.

**Table 15.** Groundwater flow budgets for the 2007 and 2008 artificial recharge and no artificial recharge simulations for each index cell of the basin storage area. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_15.xlsx*.

**Table 17.** Simulated changes in saturated volumes and storage volumes, for the Central Wichita well field, the pre-2012 Wichita well field, and the current (2012) Wichita well field study areas for each stress period and since December 31, 1939, December 31, 1992, and December 31, 2005. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_17.xlsx*.

**Table 18.** Cumulative simulated groundwater flow budget for the central Wichita well field area. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_18.xlsx*.

**Table 19.** Cumulative simulated groundwater flow budget for the pre-2012 Wichita well field area. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_19.xlsx*.

**Table 20.** Cumulative simulated groundwater flow budget for the Wichita well field area. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_20.xlsx*.

**Table 21.** Changes in storage calculated from simulated groundwater flow budgets for the Central Wichita well field, the pre-2012 Wichita well field, and the current (2012) Wichita well field study areas for each stress period and since December 31, 1939, December 31, 1992, and December 31, 2005. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_21.xlsx*.

**Table 22.** Storage changes calculated from simulated groundwater flow budgets, simulated groundwater levels, and measured groundwater levels (Hansen and Aucott, 2010) for the pre-2012 Wichita well field area from 1990 through 2008. The Excel file may be downloaded from *http://pubs.usgs.gov/sir/2013/5042/downloads/table_22.xlsx*.

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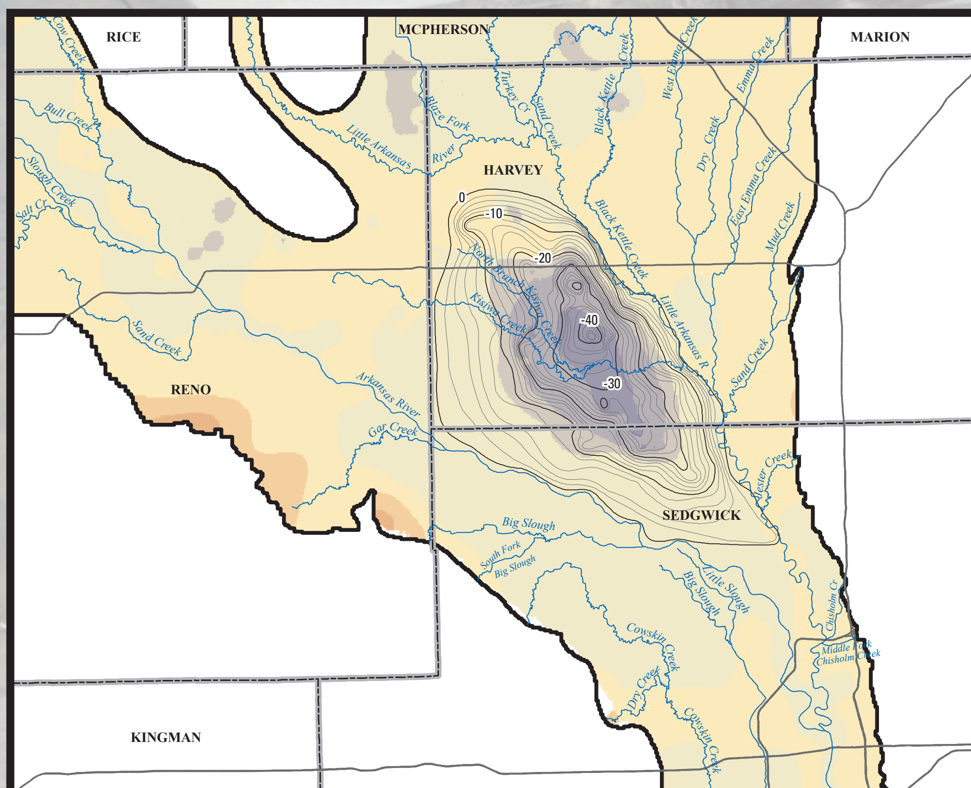


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