Prepared in cooperation with the Oklahoma Water Resources Board

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Scientific Investigations Report 2016–5180

Version 1.1, March 2017

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

Cover, Canadian River near Highway 283 in Roger Mills County, Oklahoma. Photograph by Kevin Smith, U.S. Geological Survey, Oklahoma, January 2013.

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By John H. Ellis, Shana L. Mashburn, Grant M. Graves, Steven M. Peterson, S. Jerrod Smith, Leland T. Fuhrig, Derrick L. Wagner, and Jon E. Sanford

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Conversion Factors

U.S. customary units to International System of Units

Multiply	By	To obtain
	Length	
inch (in.)	2.54	centimeter (cm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
acre	4,047	square meter (m^2)
acre	0.004047	square kilometer (km^2)
square mile (mi^2)	259.0	hectare (ha)
square mile (mi^2)	2.590	square kilometer (km^2)
	Volume	
acre-foot (acre-ft)	1,233	cubic meter (m^3)
	Flow rate	
foot per second (ft/s)	0.3048	meter per second (m/s)
foot per day (ft/d)	0.3048	meter per day (m/d)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)
	Hydraulic gradient	
foot per mile (ft/mi)	0.1894	meter per kilometer (m/km)

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

$$°F = (1.8 \times °C) + 32.$$

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

$$°C = (°F - 32) / 1.8.$$

Datum

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.

Abbreviations

BFI	base-flow index
DEM	digital elevation model
EPS	equal proportionate share
ET	evapotranspiration
GHB	General Head Boundary
HPT	Hydraulic Profiling Tool
IDW	Inverse Distance Weighting
MAY	maximum annual yield
NHDPlus	National Hydrography Dataset
NWIS	National Water Information System
OWRB	Oklahoma Water Resources Board
PEST	parameter estimation
RMSE	root-mean-square error
SFR	Streamflow-Routing Package version 2
SVDA	singular value decomposition-assist
SWB	soil-water balance
USGS	U.S. Geological Survey
WTF	water-table fluctuation

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By John H. Ellis, Shana L. Mashburn, Grant M. Graves, Steven M. Peterson, S. Jerrod Smith, Leland T. Fuhrig, Derrick L. Wagner, and Jon E. Sanford

Abstract

This report describes a study of the hydrogeology and simulation of groundwater flow for the Canadian River alluvial aquifer in western and central Oklahoma conducted by the U.S. Geological Survey in cooperation with the Oklahoma Water Resources Board. The report (1) quantifies the groundwater resources of the Canadian River alluvial aquifer by developing a conceptual model, (2) summarizes the general water quality of the Canadian River alluvial aquifer groundwater by using data collected during August and September 2013, (3) evaluates the effects of estimated equal proportionate share (EPS) on aquifer storage and streamflow for time periods of 20, 40, and 50 years into the future by using numerical groundwater-flow models, and (4) evaluates the effects of present-day groundwater pumping over a 50-year period and sustained hypothetical drought conditions over a 10-year period on stream base flow and groundwater in storage by using numerical flow models. The Canadian River alluvial aquifer is a Quaternary-age alluvial and terrace unit consisting of beds of clay, silt, sand, and fine gravel sediments unconformably overlying Tertiary-, Permian-, and Pennsylvanian-age sedimentary rocks. For groundwater-flow modeling purposes, the Canadian River was divided into Reach I, extending from the Texas border to the Canadian River at the Bridgeport, Okla., streamgage (07228500), and Reach II, extending downstream from the Canadian River at the Bridgeport, Okla., streamgage (07228500), to the confluence of the river with Eufaula Lake. The Canadian River alluvial aquifer spans multiple climate divisions, ranging from semiarid in the west to humid subtropical in the east. The average annual precipitation in the study area from 1896 to 2014 was 34.4 inches per year (in/yr).

A hydrogeologic framework of the Canadian River alluvial aquifer was developed that includes the areal and vertical extent of the aquifer and the distribution, texture variability, and hydraulic properties of aquifer materials. The aquifer areal extent ranged from less than 0.2 to 8.5 miles wide. The maximum aquifer thickness was 120 feet (ft), and the average aquifer thickness was 50 ft. Average horizontal hydraulic conductivity for the Canadian River alluvial aquifer was calculated to be 39 feet per day, and the maximum horizontal hydraulic conductivity was calculated to be 100 feet per day.

Recharge rates to the Canadian River alluvial aquifer were estimated by using a soil-water-balance code to estimate the spatial distribution of groundwater recharge and a water-table fluctuation method to estimate localized recharge rates. By using daily precipitation and temperature data from 39 climate stations, recharge was estimated to average 3.4 in/yr, which corresponds to 8.7 percent of precipitation as recharge for the Canadian River alluvial aquifer from 1981 to 2013. The water-table fluctuation method was used at one site where continuous water-level observation data were available to estimate the percentage of precipitation that becomes groundwater recharge. Estimated annual recharge at that site was 9.7 in/yr during 2014.

Groundwater flow in the Canadian River alluvial aquifer was identified and quantified by a conceptual flow model for the period 1981–2013. Inflows to the Canadian River alluvial aquifer include recharge to the water table from precipitation, lateral flow from the surrounding bedrock, and flow from the Canadian River, whereas outflows include flow to the Canadian River (base-flow gain), evapotranspiration, and groundwater use. Total annual recharge inflows estimated by the soil-water-balance code were multiplied by the area of each reach and then averaged over the simulated period to produce an annual average of 28,919 acre-feet per year (acre-ft/yr) for Reach I and 82,006 acre-ft/yr for Reach II. Stream base flow to the Canadian River was estimated to be the largest outflow of groundwater from the aquifer, measured at four streamgages, along with evapotranspiration and groundwater use, which were relatively minor discharge components.

Objectives for the numerical groundwater-flow models included simulating groundwater flow in the Canadian River alluvial aquifer from 1981 to 2013 to address groundwater use and drought scenarios, including calculation of the EPS pumping rates. The EPS for the alluvial and terrace aquifers is defined by the Oklahoma Water Resources Board as the amount of fresh water that each landowner is allowed per year per acre of owned land to maintain a saturated thickness of

at least 5 ft in at least 50 percent of the overlying land of the groundwater basin for a minimum of 20 years.

The groundwater-flow models were calibrated to water-table altitude observations, streamgage base flows, and base-flow gain to the Canadian River. The Reach I water-table altitude observation root-mean-square error was 6.1 ft, and 75 percent of residuals were within ±6.7 ft of observed measurements. The average simulated stream base-flow residual at the Bridgeport streamgage (07228500) was 8.8 cubic feet per second (ft^3/s), and 75 percent of residuals were within ±30 ft^3/s of observed measurements. Simulated base-flow gain in Reach I was 8.8 ft^3/s lower than estimated base-flow gain. The Reach II water-table altitude observation root-mean-square error was 4 ft, and 75 percent of residuals were within ±4.3 ft of the observations. The average simulated stream base-flow residual in Reach II was between 35 and 132 ft^3/s. The average simulated base-flow gain residual in Reach II was between 11.3 and 61.1 ft^3/s.

Several future predictive scenarios were run, including estimating the EPS pumping rate for 20-, 40-, and 50-year life of basin scenarios, determining the effects of current groundwater use over a 50-year period into the future, and evaluating the effects of a sustained drought on water availability for both reaches. The EPS pumping rate was determined to be 1.35 acre-feet per acre per year ([acre-ft/acre]/yr) in Reach I and 3.08 (acre-ft/acre)/yr in Reach II for a 20-year period. For the 40- and 50-year periods, the EPS rate was determined to be 1.34 (acre-ft/acre)/yr in Reach I and 3.08 (acre-ft/acre)/yr in Reach II. Storage changes decreased in tandem with simulated groundwater pumping and were minimal after the first 15 simulated years for Reach I and the first 8 simulated years for Reach II.

Groundwater pumping at year 2013 rates for a period of 50 years resulted in a 0.2-percent decrease in groundwater-storage volumes in Reach I and a 0.6-percent decrease in the groundwater-storage volumes in Reach II. The small changes in storage are due to groundwater use by pumping, which composes a small percentage of the total groundwater-flow model budgets for Reaches I and II.

A sustained drought scenario was used to evaluate the effects of a hypothetical 10-year drought on water availability. A 10-year period was chosen where the effects of drought conditions would be simulated by decreasing recharge by 75 percent. In Reach I, average simulated stream base flow at the Bridgeport streamgage (07228500) decreased by 58 percent during the hypothetical 10-year drought compared to average simulated stream base flow during the nondrought period. In Reach II, average simulated stream base flows at the Purcell streamgage (07229200) and Calvin streamgage (07231500) decreased by 64 percent and 54 percent, respectively. In Reach I, the groundwater-storage drought scenario resulted in a storage decline of 30 thousand acre-feet, or an average decline in the water table of 1.2 ft. In Reach II, the groundwater-storage drought scenario resulted in a storage decline of 71 thousand acre-feet, or an average decline in the water table of 2.0 ft.

Introduction

The Canadian River alluvial aquifer in western and central Oklahoma is an unconfined alluvial aquifer derived from erosion and redeposition of sediments by the river that are composed of clay, silt, sand, and fine gravel (pl. 1). The Canadian River alluvial aquifer underlies the Canadian River, which originates in Colorado, flows through New Mexico and Texas, and then flows into Oklahoma to Eufaula Lake in McIntosh County, Oklahoma (fig. 1). Groundwater in approximately 606 square miles (mi^2) of this aquifer is used for irrigation, municipal use, mining (oil and gas), livestock, and domestic supply. Groundwater from the Canadian River alluvial aquifer is used by several municipalities including Lexington, Noble, and Tuttle, Okla. (Oklahoma Water Resources Board, 2012a). Flow in the Canadian River has been regulated since 1964 by Sanford Dam (fig. 1), which impounds Lake Meredith in Hutchinson County, Tex.

The 1973 Oklahoma Water Law (82 OK Stat § 82-1020.5) requires the Oklahoma Water Resources Board (OWRB) to conduct hydrologic surveys of the State’s aquifers to determine the maximum annual yield (MAY) over the life of the basin. The MAY is the total amount of fresh groundwater that may be pumped from a groundwater basin for a minimum 20-year life of such basin. The equal-proportionate-share (EPS) is the MAY of groundwater allocated to each acre of land overlying the basin. The life of the basin is the period of time during which at least 50 percent of the total overlying land will retain a saturated thickness, allowing pumping of the MAY for a minimum 20-year life of the basin. Saturated thickness for alluvial and terrace aquifers shall remain at least 5 feet (ft) by law. To determine the MAY of a major groundwater basin, an investigation is conducted to obtain data and information related to the geology, hydrogeologic framework, and hydrogeology of an aquifer. A numerical groundwater-flow model can be developed by using these data and information to conceptualize the flow system and evaluate effects of water use on the aquifer. The study described in this report to evaluate the Canadian River alluvial aquifer was a cooperative effort between the U.S. Geological Survey (USGS) and the OWRB.

The objectives of the study were to (1) conduct a hydrologic survey to quantify the groundwater resources of the Canadian River alluvial aquifer by using groundwater-level and streamflow data collected during 2013–16 and 1981–2013, respectively; (2) summarize the general water quality of the Canadian River alluvial aquifer groundwater; (3) evaluate the effects of estimated EPS on aquifer storage and streamflow; and (4) evaluate the effects of present-day groundwater pumping and sustained hypothetical drought conditions on stream base flow and groundwater in storage. The scope of this study was the Canadian River alluvial aquifer and adjoining bedrock units.

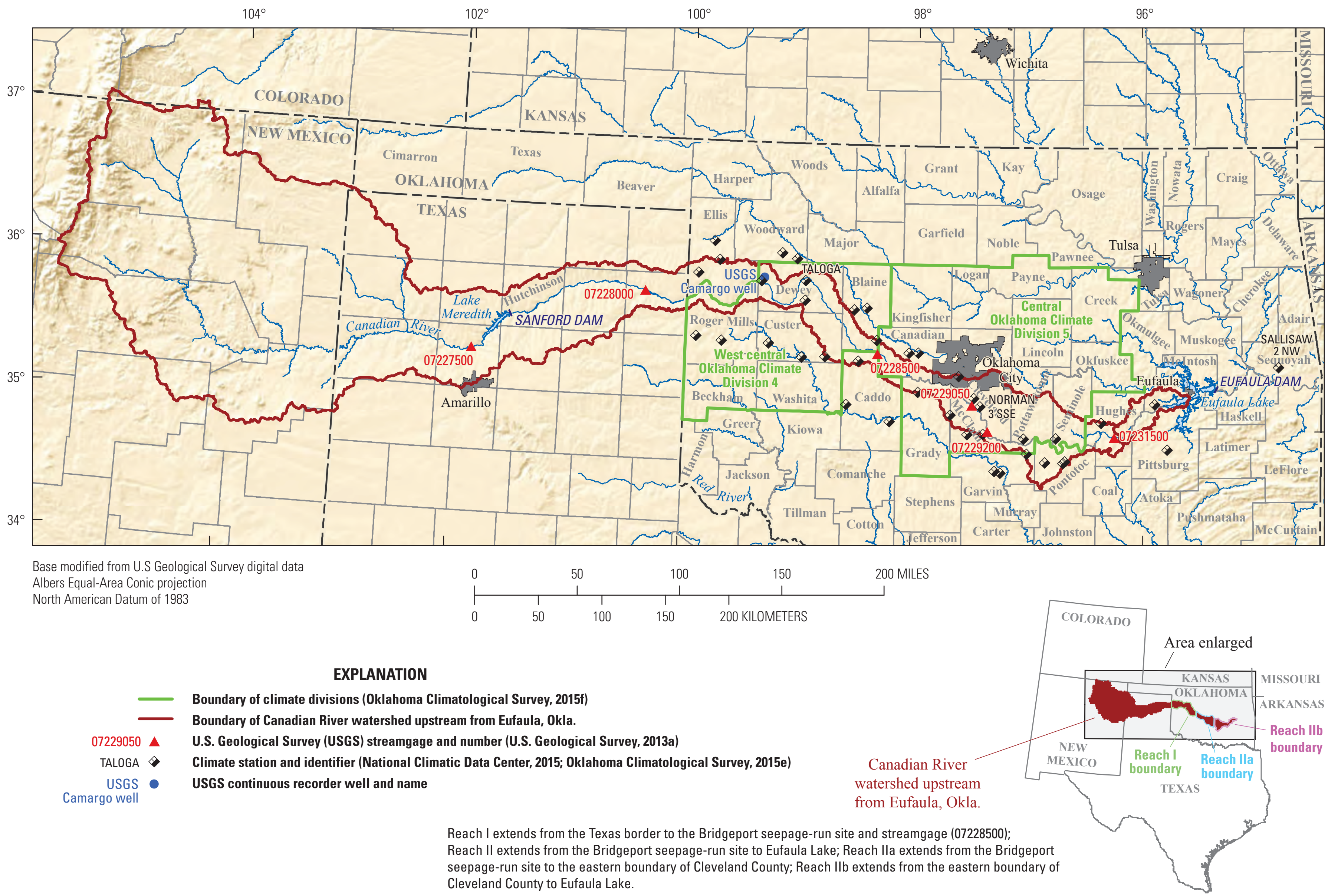


Figure 1. Extent of the Canadian River watershed upstream from Eufaula, Oklahoma, and climate divisions, climate stations, and streamgages.

Purpose and Scope

The purpose of this report is to describe the hydrogeology and simulation of groundwater flow for the Canadian River alluvial aquifer in western and central Oklahoma. The report (1) quantifies the groundwater resources of the Canadian River alluvial aquifer by developing a conceptual model; (2) summarizes the general water quality of the Canadian River alluvial aquifer groundwater by using data collected during August and September 2013; (3) evaluates the effects of estimated EPS on aquifer storage and streamflow for time periods of 20, 40, and 50 years into the future by using numerical groundwater-flow models; and (4) evaluates the effects of present-day groundwater pumping over a 50-year period and sustained hypothetical drought conditions over a 10-year period on stream base flow and groundwater in storage by using numerical flow models.

Study Area Description

The study area is the extent of the Canadian River alluvial aquifer in Oklahoma from the Texas border to Eufaula Lake, as well as the adjoining bedrock units as shown in plate 1. The Canadian River alluvial aquifer underlies approximately 606 mi^2 of land in 14 Oklahoma counties along the Canadian River from the Texas border to Eufaula Lake (fig. 1). The western part of the study area near the Texas border is characterized by gently sloping hills surrounding the bends of the Canadian River (Kitts and Black, 1959). The central part of the study area contains grass-covered plains and sandstone hills (Wood and Burton, 1968). The eastern part of the study area consists of grass-covered plains and northeast-trending cuestas cut by stream valleys (Weaver, 1954). The study area includes substantial changes in elevation from west to east, ranging from 2,150 ft above the North American Vertical Datum of 1988 (NAVD 88) in the west (Kitts and Black, 1959) to about 500 ft above NAVD 88 near Eufaula Lake (Oakes and Koontz, 1967).

For the study described in this report, the Canadian River alluvial aquifer was divided into two sections: Reach I, which is 242 mi^2 and extends from the Texas border to the Canadian River at the Bridgeport, Okla., streamgage (07228500); and Reach II, which is 364 mi^2 and extends from the Canadian River at the Bridgeport, Okla., streamgage (07228500) to Eufaula Lake (fig. 1).

Land Use

Land-use data for the Canadian River alluvial aquifer were obtained from the CropScape database (National Agricultural Statistics Service, 2015) (fig. 2). That database includes information about land-use characteristics at a 30-meter resolution. Land uses overlying the total reported alluvial aquifer area of approximately 388,000 acres (606 mi^2) were composed of grass/pasture (52.3 percent), crops (23.7 percent), and forest (13.3 percent) in 2015. The remaining acreage included developed land (4.2 percent), shrubs (1.9 percent), and other uses (4.7 percent). The crops grown in this area were mostly winter wheat (61.3 percent of the total crop area) and alfalfa or other hay (17.9 percent of the total crop area). Corn, soybeans, and cotton were each grown in less than 4 percent of the total crop area. About 4 percent of the total crop area was fallow or idle, and 7.1 percent of the total crop area was allocated to other crops.

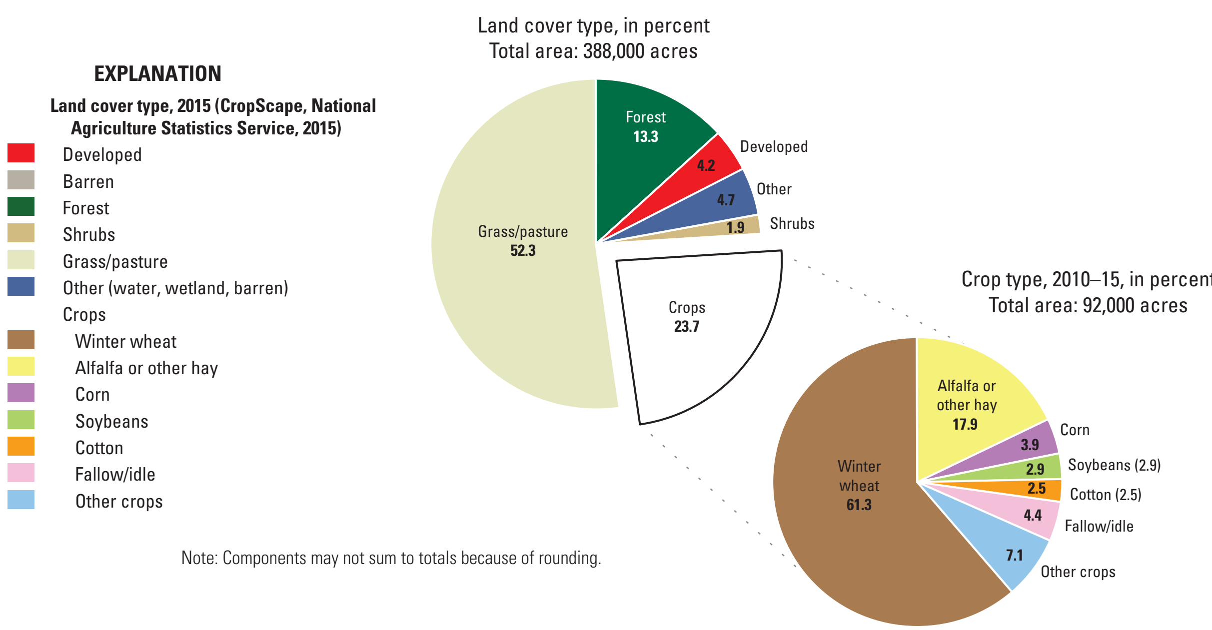


Figure 2. Land cover and crop types on the Canadian River alluvial aquifer, Oklahoma.

Climate

Climate in the area underlain by the Canadian River alluvial aquifer ranges from semiarid in the western part of the study area to humid subtropical in the eastern part of the study area (Oklahoma Climatological Survey, 2015a). Average annual precipitation and temperatures generally increase from west to east across the aquifer (Oklahoma Climatological Survey, 2015a). Daily air temperatures ranged from 44.4 to 71.6 degrees Fahrenheit (°F) in Roger Mills County, from 49.7 to 72.7 °F in Cleveland County, and from 52.2 to 72.8 °F in McIntosh County (Oklahoma Climatological Survey, 2015b, 2015c, and 2015d).

Climate data for three cooperative observer stations distributed across the extent of the Canadian River alluvial aquifer were obtained from the Oklahoma Climatological Survey's Web site (Oklahoma Climatological Survey, 2015e). These stations were chosen for data representing long-term climate conditions for the western, central, and eastern parts of the aquifer. The three stations used in this study were Taloga (western), Norman 3SSE (central), and Sallisaw 2 NW (eastern) (fig. 1, table 1). Some data were missing from the periods of record of these stations, and data from years containing less than 10 months of precipitation data were omitted from the calculation of average precipitation. Average annual precipitation increased by 2.5 inches per year (in/yr). during the period 1981–2014 compared to 1947–80 (table 1).

Table 1. Data collection time periods of average annual precipitation at selected cooperative observer climate stations used in the Canadian River alluvial aquifer study area.

[Data from Oklahoma Climatological Survey, 2015e. All units are in inches per year]

Station number	Station name	Period of record[1]	Number of years	1947–80 average	1981–2014 average
8708	TALOGA	1900–2015	66	25.8	27.3
6386	NORMAN 3SSE	1895–2015	101	34.7	37.4
7862	SALLISAW 2NW	1893–2015	81	45.0	48.4
			Average	**35.2**	**37.7**

[1]Not continuous.

Average annual precipitation in the study area from 1896 to 2014 was 34.4 in/yr. (fig. 3). Precipitation trends indicate (1) below-average precipitation from the late 1890s through the early 1920s, (2) above-average precipitation from the late 1920s to the late 1940s, (3) variable precipitation between 1950 and the early 1980s, (4) above-average precipitation from the mid-1980s to 2010, and (5) below-average precipitation from 2010 to 2014 (fig. 3). On average, the greatest amounts of precipitation occurred in May, and the least amounts occurred in January (fig. 4).

Reported Water Use

Water use in Oklahoma is regulated by the OWRB, and permitted water users are required to submit annual water-use reports to the OWRB. Water use was reported for 754 permits submitted to the OWRB by permitted groundwater users of the Canadian River alluvial aquifer for the 1967–2013 period (Oklahoma Water Resources Board, 2015a).

The average annual groundwater use from 1967 to 2013 was 11,887 acre-feet per year (acre-ft/yr) (table 2). The annual groundwater-use data show a period of greater use during the period of 1967–79, followed by a period of lower use during 1980–99 (fig. 5). During 2000–13, there was an increasing trend in groundwater use, with a peak of 23,380 acre-ft/yr in 2012 (fig. 5, table 2). The average annual groundwater use decreased from 12,919 acre-ft/yr during the period of 1967–79 to 9,392 acre-ft/yr during 1980–99 (fig. 5, table 2). That decrease may be a result of a change in the method of groundwater use reporting in 1980. Prior to 1980, irrigation use was based on crop type, acres, and frequency of application (Oklahoma Water Resources Board, 2002). In 1980, the method was changed to include inches applied to increase accuracy (Oklahoma Water Resources Board, 2002). Average annual groundwater use increased to 14,461 acre-ft/yr during the period of 2000–13 (fig. 5, table 2).

The greatest average groundwater use during the period of 1967–2013 was for irrigation, with 8,476 acre-ft/yr, or about 71.3 percent of average groundwater use (fig. 5). Public supply was the second largest use of groundwater during the period of 1967–2013, with 2,466 acre-ft/yr, or about 20.7 percent (fig. 5). The power, mining, and recreation, fish, wildlife use types combined accounted for 799 acre-ft/yr, or about 6.7 percent of groundwater use (fig. 5). All other use types, which includes self-supplied domestic, composed only 146 acre-ft/yr, or about 1.2 percent of groundwater use, during the period of 1967–2013.

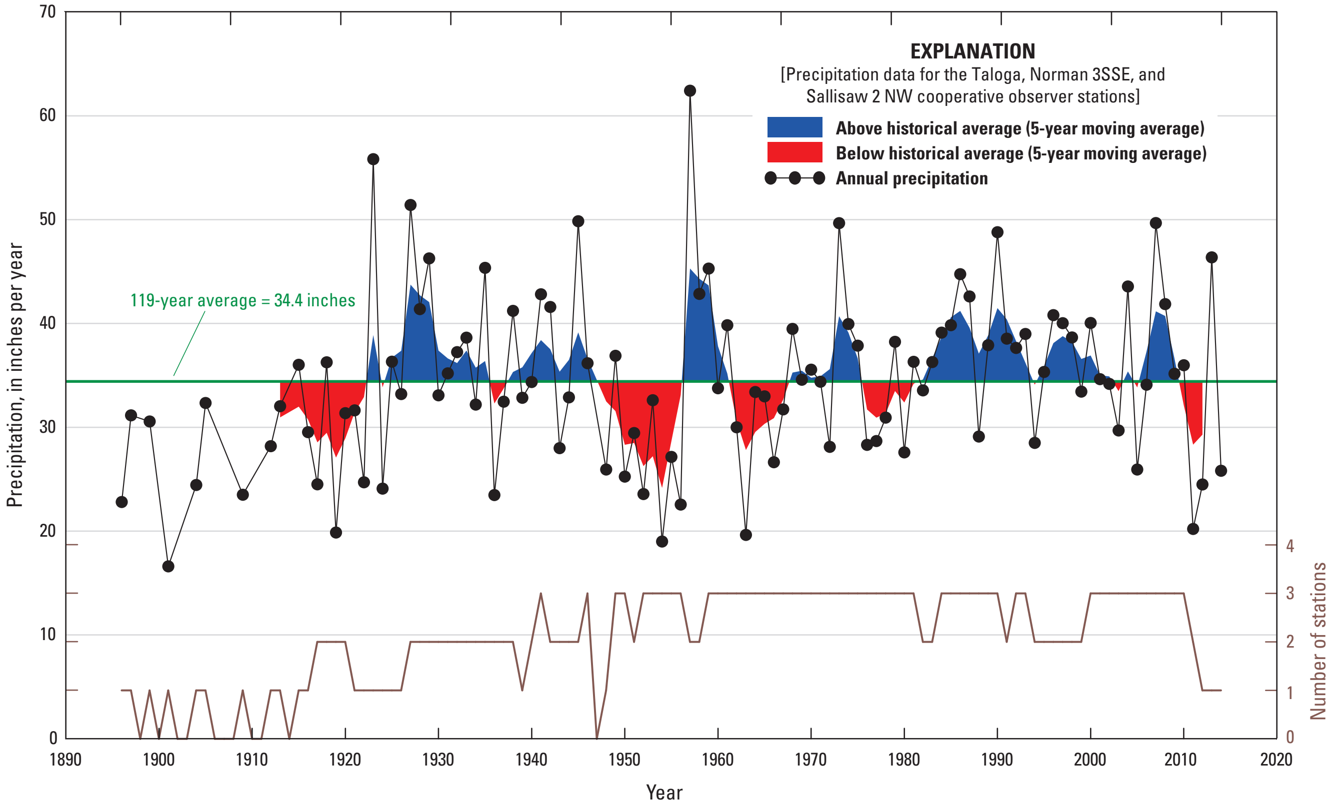


Figure 3. Average annual precipitation for the period 1896–2014, 5-year moving average, and the number of cooperative observer stations recording during each year in the Canadian River alluvial aquifer study area.

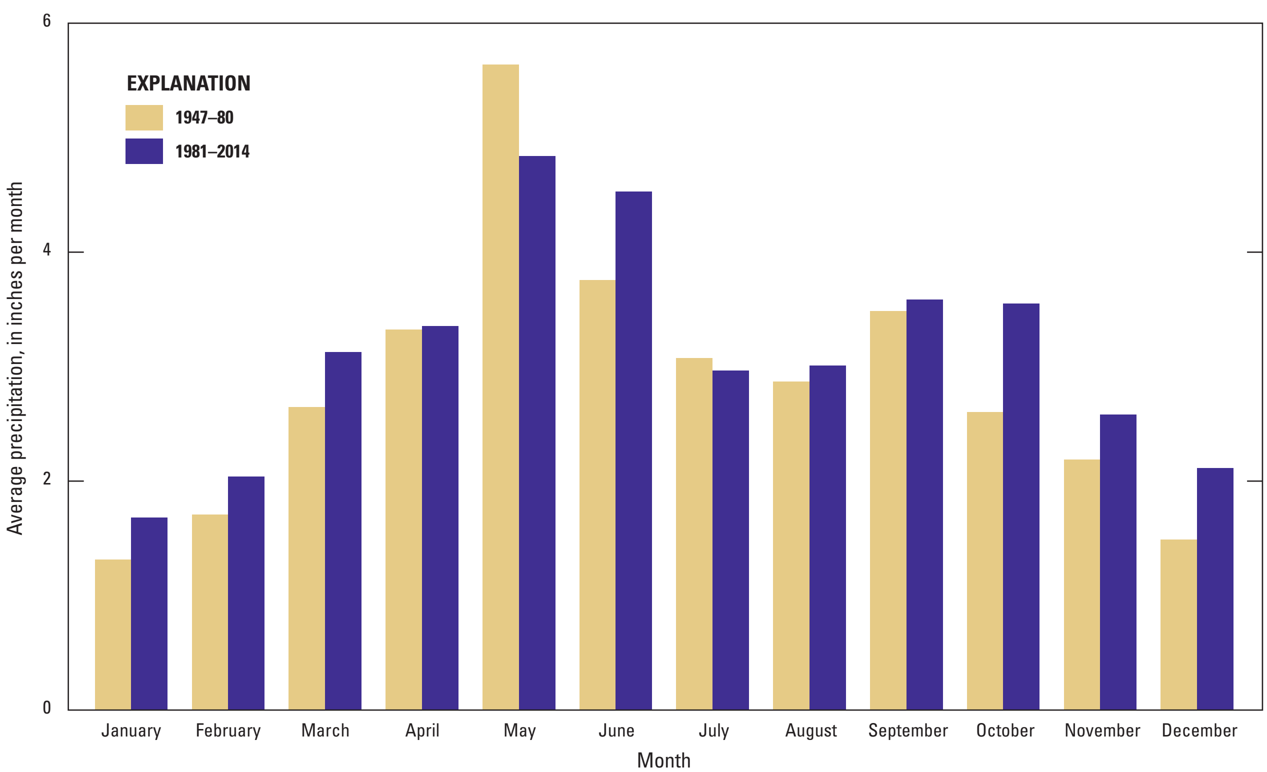


Figure 4. Average monthly precipitation during the time periods of 1947–80 and 1981–2014 for the Taloga, Norman 3SSE, and Sallisaw 2 NW cooperative observer stations in the Canadian River alluvial aquifer study area.

Table 2. Reported annual groundwater-use statistics for the Canadian River alluvial aquifer, Oklahoma, 1967–2013.

	Reported annual groundwater use (acre-feet per year)			
Statistic	**1967–2013**	**1967–79**	**1980–99**	**2000–13**
Average	11,887	12,919	9,392	14,461
Median	11,019	12,705	9,121	12,679
Minimum	6,763	8,088	6,763	9,562
Maximum	23,380	17,891	13,648	23,380

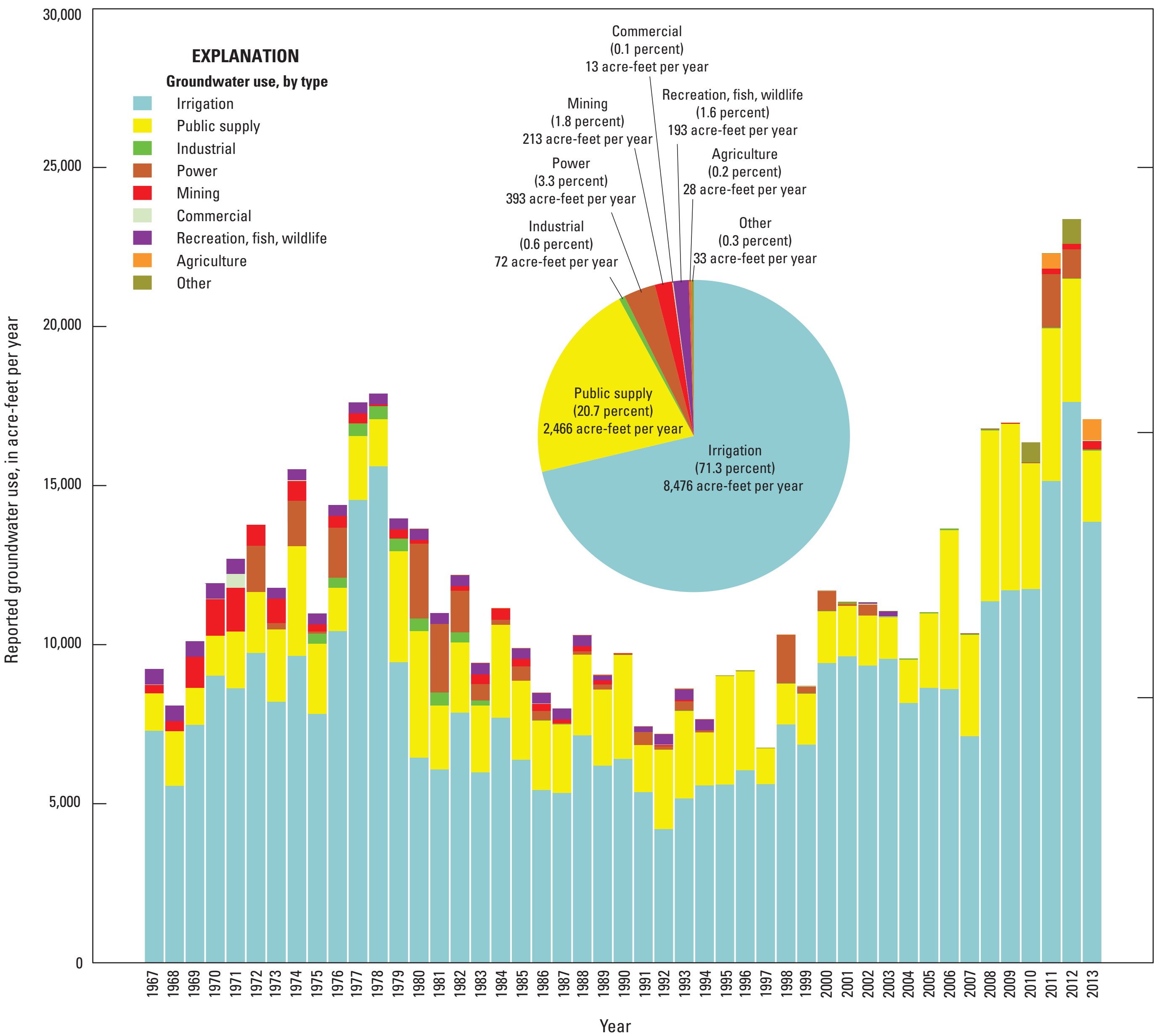


Figure 5. Groundwater use by type for the Canadian River alluvial aquifer, Oklahoma, 1967–2013.

Geology of the Canadian River Alluvial Aquifer

The geologic units in the study area include Quaternary-age alluvial and terrace deposits of the Canadian River alluvial aquifer and Tertiary-, Permian-, and Pennsylvanian-age sedimentary bedrock (pl. 1, fig. 6). The clay, silt, sand, and fine gravel sediment of the alluvial and terrace deposits unconformably overlie the bedrock units.

Quaternary-Age Alluvial and Terrace Deposits

The Canadian River alluvial aquifer originated as the Rocky Mountains were uplifted (Fay and others, 1962), with the river channel developing in outwash during subsequent continental glaciation and regression of the Pleistocene Epoch (Johnson and Luza, 2008). Downcutting of the river channel occurred, forming major valleys through the Permian-age bedrock units, possibly along the path of older stream channels (Fay and others, 1962). These valleys were gradually filled with sand and fine gravel eroded from the Rocky Mountains or the Tertiary-age Ogallala Formation (Hendricks, 1937). Multiple deposition and erosion cycles continued with streams cutting through and removing previously deposited material followed by infill from older terrace deposits and bedrock sediment (Davis, 1955). During years of slow glacial ice melt, older alluvial sediment remained to form various terrace shelves (Fay and others, 1962).

In the present-day alluvial flood plain, sediment with a total thickness up to 100 ft (Oakes and Koontz, 1967) constitutes the majority of aquifer material deposited during the most recent cycle of downcutting of the river (Davis, 1955). The alluvial deposits in western Oklahoma primarily consist of sand and fine gravel with traces of clay and caliche (Mogg and others, 1960). Substantial meanders in the alluvial deposits are present in western Oklahoma and form the boundaries for Roger Mills County and Dewey County. The sand and fine gravel are typically composed of quartz with varying amounts of quartzite, chert, flint, jasper, and mineralized wood (Davis, 1955). The Canadian River alluvial deposits in western Oklahoma overlie the Rush Springs and Marlow Formations and closely follow the southeastern strike of the Permian-age beds, a trend that continues through Oklahoma County (Wood and Burton, 1968). The stream-channel gradient in western Oklahoma is between 4 and 5 feet per mile (ft/mi) (Kitts and Black, 1959; Fay and others, 1962).

The alluvial deposits average 2 miles (mi) in width in central Oklahoma and widen to approximately 3 mi in width northwest of Norman (Wood and Burton, 1968). Alluvial deposits in central Oklahoma contain an abundance of quartz, quartzite, chert, flint, jasper, and silicified wood (Davis, 1955). The alluvial channel dissects sandstone hills with relief from 50 to 200 ft (Wood and Burton, 1968).

In east-central Oklahoma, incised meanders of the Canadian River are present, possibly because of low topographic relief (Tanner, 1956). This area has a poorly defined flood plain of less than a mile in width (Tanner, 1956). In contrast to western Oklahoma, the river and surrounding alluvial deposits in east-central Oklahoma cross the strike of the underlying bedrock at right angles, with the exception of an area near the confluence of the Canadian River and Little River (Tanner, 1956), which is located about 52 miles upstream from Eufaula Lake in Hughes County (pl. 1).

Three distinct terrace shelves (high, intermediate, and low) occur along the Canadian River from Roger Mills County in the west to Hughes County in the east, ranging from more than 220 ft to less than 15 ft above the alluvial flood plain (Weaver, 1954; Kitts and Black, 1959). In western Oklahoma, the high and intermediate terraces are eroded on the northern side of the river in Roger Mills County (Kitts and Black, 1959) but are still present towards the east through Dewey County and Blaine County. In central Oklahoma, two or three sets of terrace shelves occur along the river, for a total thickness of up to 100 ft (Wood and Burton, 1968). The high terrace units occur in well-exposed bluffs and are easily distinguishable from underlying Permian-age units by color and composition; the low terrace can be distinguished from nearby alluvial deposits, with difficulty in some areas (Mogg and others, 1960).

In eastern Oklahoma, the slow erosional rate that forms terrace shelves did not regularly occur, resulting in less-defined terrace shelves, although three distinct terraces occur in some areas (Weaver, 1954).

Eolian dune sands, which have been shifted by southerly winds, occur in the terrace units throughout the study area. Dune sand is typically above the water table and not a separate zone or unit of the aquifer, but because of the high permeability, dune sand can facilitate infiltration of precipitation and thus increase recharge to the aquifer (Wood and Burton, 1968).

Bedrock Units

The bedrock underlying and surrounding the Canadian River alluvial aquifer is composed of Tertiary-, Permian-, and Pennsylvanian-age units (pl. 1, fig. 6). Outcrops occur in the alluvial and terrace deposits where the bedrock surface has been exposed by weathering or other events.

System	Hydrogeologic unit	Group	Formation	Thickness (feet)	Description
Quaternary	Canadian River alluvial aquifer	Alluvial, terrace, and dune deposits		0–120	Silt, sand, and clay deposited by the Canadian River and tributaries
Tertiary	High Plains (Ogallala) aquifer	Ogallala Formation		[1]0–400	Brown to light tan, mostly unconsolidated clay, silt, sand, and gravel with zones of caliche near the surface
Permian		Cloud Chief Formation		[2]0–100	Reddish-brown to orange-brown shale interbedded with siltstone and sandstone, and dolomite
	Rush Springs aquifer	Whitehorse Group	Rush Springs Formation	[2]0–300	Red to pink massive very-fine-grained gypsiferous sandstone
			Marlow Formation		Siltstone and very fine-grained sandstone, clayey and gypsiferous
		El Reno Group	Dog Creek Shale	[2]0–190	Red, brown, and green gypsiferous shales with several beds of siltstone, sandstone, and dolomite
			Blaine Formation	[2]0–100	Beds of white gypsum and thin beds of gray medium-grained dolomite or dolomitic limestone
			Chickasha Formation	[3]150–200	Mudstone conglomerate and red-brown to orange-brown silty shale and siltstone
			Duncan Sandstone	[3]150–200	Red-brown to orange-brown fine-grained sandstone with mudstone conglomerate and shale
		Hennessey Group		[3]600–650	Mostly reddish-brown shale with reddish-brown to reddish-orange siltstone
	Central Oklahoma (Garber-Wellington) aquifer	Sumner Group	Garber Sandstone	[3]0–400	Cross-bedded fine-grained sandstone with interbedded shale and mudstone
			Wellington Formation	[3]0–700	Mostly reddish-brown shale with reddish-brown to reddish-orange siltstone
		Chase, Council Grove, and Admire Groups		[4]0–500+	Deep-red to reddish-orange massive cross-bedded fine-grained sandstone with shale and siltstone
Pennsylvanian		Vanoss Formation		[5]0–250	Alternating layers of limestone, shale, and fine-grained arkosic sandstone
	Vamoosa-Ada aquifer	Ada Formation		[5]0–100	Primarily shale with fine-grained sandstone
		Vamoosa Formation		[5]125	Alternating layers of shale, sandstone, and chert conglomerates
		Pre-Vamoosa bedrock units, undifferentiated		[5]36–680	Shale, with sandstone and limestone

Modified from Miser, 1954.

[1]Morton, 1980.

[2]Fay and others, 1962.

[3]Wood and Burton, 1968.

[4]Christenson and others, 1992.

[5]Tanner, 1956.

Figure 6. Geologic and hydrogeologic units of the Canadian River alluvial aquifer study area.

Tertiary-Age Bedrock Units

In the western part of Roger Mills County, the Canadian River alluvial aquifer unconformably overlies the Tertiary-age Ogallala Formation, the principal geologic unit of the High Plains (Ogallala) aquifer. The Ogallala Formation consists of fine- to medium-grained, well-sorted, yellow-brown quartz sands with calcium-carbonate cement present in small quantities (Kitts and Black, 1959). The average thickness of the Ogallala Formation in the study area is approximately 150 ft, and the maximum thickness is approximately 400 ft (Morton, 1980). The Ogallala Formation rests unconformably on Permian-age formations (Kitts and Black, 1959). In Roger Mills County, the Ogallala Formation and high terrace contact is 140–160 ft above the Canadian River alluvial flood plain (Kitts and Black, 1959).

Permian-Age Bedrock Units

The Permian-age bedrock units underlying the Canadian River alluvial aquifer in western and central Oklahoma are commonly referred to as “red beds” because of the distinctive red color that comes from sandstone containing iron-oxide minerals (Breit, 1998). Four bedrock units underlie the Canadian River alluvial aquifer between Roger Mills County and western Canadian County, including the Cloud Chief, Rush Springs, and Marlow Formations, and the Dog Creek Shale (pl. 1, fig. 6). The Cloud Chief Formation consists of gypsum-cemented shale and siltstone, and minor amounts of fine sandstone. In Dewey County, the Cloud Chief Formation forms the surface of about one-third of the county and forms heavy gypsum ledges (Six, 1930). The Rush Springs Formation consists of orange-brown fine-grained sandstone and siltstone, with interbedded red-brown shale, silty shale, and gypsum (Wood and Stacy, 1965). The Marlow Formation consists of siltstone and very fine-grained sandstone with clay and gypsum (Mogg and others, 1960). The Dog Creek Shale is a red-brown shale with discontinuous bands of silt and thin layers of dolomite, has very low permeability, and is not a predominant source of water (Mogg and others, 1960; Wood and Stacy, 1965).

The Blaine Formation, Chickasha Formation, Duncan Sandstone, Hennessey Group, Garber Sandstone, Wellington Formation, and Chase, Council Grove, and Admire Groups (pl. 1, fig. 6) underlie the Canadian River alluvial aquifer from the eastern extent of Canadian County to Pottawatomie County. The Blaine Formation consists mostly of thin gypsum beds in the study area with thin beds of dolomite below each gypsum layer, interbedded with red-brown shale (Fay and others, 1962; Bingham and Moore, 1975). The Chickasha Formation consists of mudstone conglomerate and red-brown to orange-brown silty shale and siltstone (Wood and Burton, 1968) (pl. 1, fig. 6). Some of the siltstone beds of the Chickasha Formation form ledges along the alluvial valleys (Davis, 1955) at the edge of terrace bluffs. The Duncan Sandstone, with a thickness of 150–200 ft (Wood and Burton, 1968), is a red-brown to orange-brown sandstone with shale and conglomerated mudstone (Bingham and Moore, 1975) that underlies the Canadian River alluvial aquifer. The Hennessey Group is characterized by red shales, which are thin and weather to a dark, rich loam. This group has a thickness of 600–650 ft in most places in Cleveland and Oklahoma Counties (Wood and Burton, 1968). The Garber Sandstone, Wellington Formation, and Chase, Council Grove, and Admire Groups (water-bearing units of the Central Oklahoma aquifer) are the oldest Permian units underlying the Canadian River. These Permian units are composed of deep-red to reddish-orange massive and cross-bedded fine-grained sandstone interbedded with shale and siltstone (Wood and Burton, 1968).

Pennsylvanian-Age Bedrock Units

In eastern Oklahoma, the bedrock units are predominantly Pennsylvanian in age (pl. 1, fig. 6) and do not share the same red color as the Permian-age units (Aurin, 1917). The Vanoss, Ada, and Vamoosa Formations underlie the Canadian River alluvial aquifer in western Seminole County, whereas the Belle City Limestone and Francis (also known as Coffeyville), Seminole, and Holdenville Formations underlie the aquifer in eastern Seminole County (pl. 1). Most of the information regarding Pennsylvanian-age bedrock units is from Tanner (1956).

The Vanoss Formation crosses the western part of Seminole County in a north-south strip and is approximately 4 mi wide at the intersection with the Canadian River. The Vanoss Formation is composed of shales, sandstones, conglomerates, and a few thin limestones. Arkosic sandstones and conglomerates are prominent in the southern part of the county near the Canadian River. The Vanoss Formation is about 250 ft thick in the vicinity of Konawa, just north of the Canadian River. In Seminole County, the Ada Formation includes shale, sandstone, and siltstone of various colors and conglomerated limestone. The shales can be discerned from the Vanoss Formation in most areas except the northern part of the county. The Vamoosa Formation (major water-bearing formation) primarily consists of shales and, to a lesser degree, sandstones, which thin southward. The Vamoosa Formation is 125 ft thick in the area where it is crossed by the Canadian River.

The pre-Vamoosa Formation bedrock units of the Pennsylvanian System in the Canadian River alluvial aquifer study area include the Belle City Limestone, Nellie Bly Formation, Francis Formation, Seminole Formation, and Holdenville Formation in Seminole County and the Wewoka Formation, Wetumka Shale, Calvin Sandstone, and Senora Formation in Hughes County (Weaver, 1954; Tanner, 1956; Oakes and Koontz, 1967). These formations typically consist of limestone, shale, and sandstone and range from 36 to 680 ft in thickness (Weaver, 1954; Tanner, 1956; Oakes and Koontz, 1967). These formations have outcrop widths of up to 2 mi (Weaver, 1954; Tanner, 1956; Oakes and Koontz, 1967).

Hydrogeologic Framework of the Canadian River Alluvial Aquifer

A hydrogeologic framework of the Canadian River alluvial aquifer was developed to provide a three-dimensional representation of the aquifer and of underlying and surrounding bedrock units. The hydrogeologic framework includes the aquifer areal and vertical extent, as well as the distribution, texture variability, and hydraulic properties of aquifer materials. The hydrogeologic framework was used to construct the numerical groundwater-flow models described in this report.

Lithologic-Log Standardization

Approximately 1,400 lithologic logs (Oklahoma Water Resources Board, 2015a) were used to characterize the alluvial and terrace deposits of the aquifer (pl. 2). These logs were recorded by well drillers and report the physical characteristics of geologic units encountered during drilling. The median depth of the lithologic logs was 60 ft, and 10 percent of those logs penetrated deeper than 100 ft below land surface (fig. 7). Most of the lithologic logs penetrated the base of the alluvial aquifer and the top of the underlying bedrock unit (fig. 8).

Techniques based on Mashburn and others (2013) were used to simplify the lithologic descriptions and to facilitate construction of the hydrogeologic framework of the aquifer. The lithologic logs were simplified by reducing the number of log interval descriptions from 3,300 unique terms to 237 standardized terms. These standardized terms were classified into five generalized lithologic categories, shown in figure 9 (clay/silt, fine sand, medium sand, coarse sand, fine gravel).

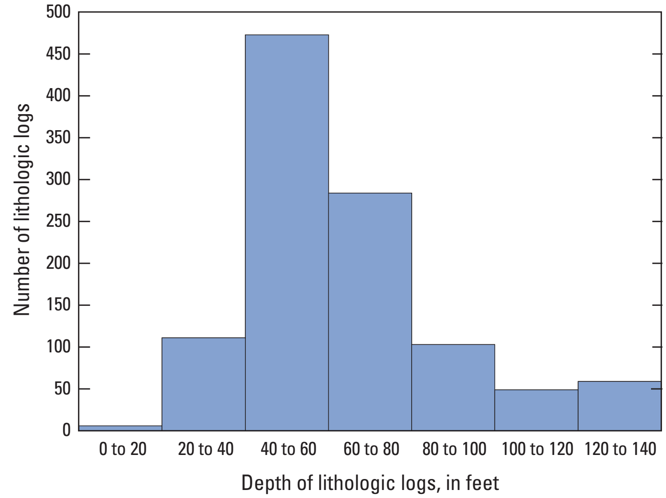


Figure 7. Frequency distribution of depth of lithologic logs used for the hydrogeologic framework in the Canadian River alluvial aquifer, Oklahoma.

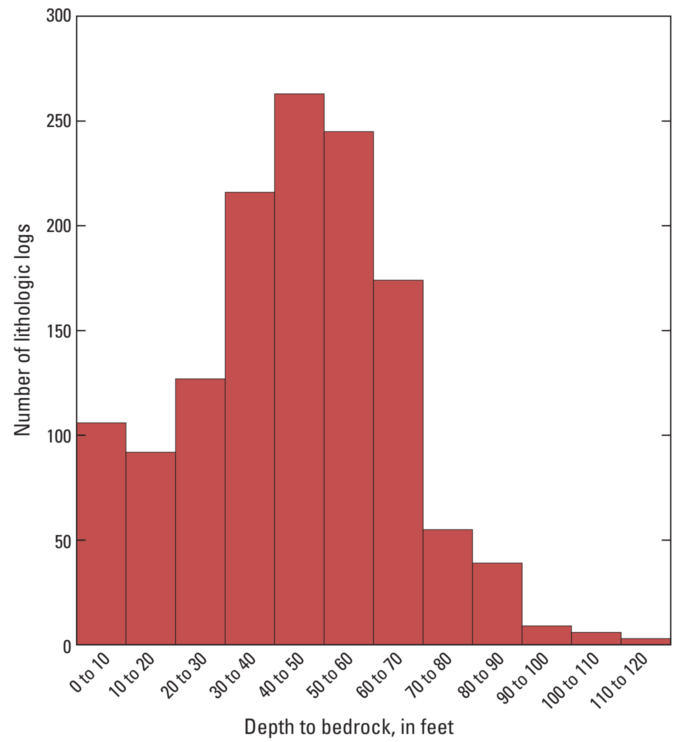


Figure 8. Frequency distribution of depth to bedrock (base of alluvial aquifer) determined from lithologic logs.

Percent coarse multiplier	10	30	50	70	90
Lithologic category	Clay/silt	Fine sand	Medium sand	Coarse sand	Fine gravel

Figure 9. Percent-coarse multiplier values for the standardized lithologic categories used to obtain horizontal hydraulic conductivity for the Canadian River alluvial aquifer, Oklahoma.

Aquifer Areal and Vertical Extent

The aquifer areal extent was determined by using 1:24,000-scale geologic maps (Stanley and Miller, 2001; Suneson and Stanley, 2001a; Suneson and Stanley, 2001b; Stanley and Miller, 2002), 1:100,000-scale geologic maps (Stanley, 2002; Miller and Stanley, 2004; Chang and Stanley, 2010; Fay, 2010a; Fay, 2010b), and 1:250,000-scale geologic maps (Heran and others, 2003). The aquifer areal extent ranged from less than 0.2 to 8.5 mi wide. Some areas designated as "alluvium" or "terrace" from the geologic maps were removed from the aquifer extent on the basis of an analysis of lithologic-log data. Other terrace deposits were excluded because they were either discontinuous or considered to be part of the groundwater-flow system of bedrock aquifers. The high terrace of the Canadian River alluvial aquifer was not included in the active area because of the lack of hydrologic connection with the lower terraces (Davis, 1955) and difficulty in differentiating between the high terrace and the Tertiary-age Ogallala Formation in western Oklahoma (pl. 1).

The bedrock surface of the Canadian River alluvial aquifer was constructed by using lithologic logs where the bedrock contact could be determined. Permian- and Pennsylvanian-age lithologic units were identified by terms denoting red or consolidated materials such as "red bed," "red clay," or "bedrock." The contact with bedrock units was defined where a continuous boundary was identified across a range of lithologic logs, regardless of the confining nature of the bedrock unit. Therefore, a red sand or an impervious red clay sequence adjoining an alluvial sand would define the aquifer boundary. Minor or interbedded intervals where bedrock was identified in alluvial or terrace deposits were discarded, and an approximate boundary determined where the contacts were gradational.

The lithologic-log bedrock contacts were then contoured by using professional judgment and geologic information from previous publications (pl. 2). The maximum aquifer thickness was 120 ft, and the average aquifer thickness was 50 ft. The bedrock contact is approximate at the western end of the Canadian River alluvial aquifer near the Texas-Oklahoma border, where there were fewer lithologic logs available to determine the bedrock contact.

Hydraulic Profiling and Sediment Coring

A direct-push Geoprobe Hydraulic Profiling Tool (HPT; Geoprobe Systems, 2007) as used to obtain 24 profiles of continuous horizontal hydraulic conductivity of the saturated zone of the Canadian River alluvial and terrace deposits (pl. 1). Locations of the HPT test holes were based on site access and proximity to locations of lithologic logs. Six of these locations were chosen for 2.25-in.-diameter sediment-core sampling to analyze and compare the described lithologic-core properties to the horizontal hydraulic conductivity profile values obtained from the HPT. The six sediment cores were described in 1-in. increments, noting grain size, sorting, and Munsell (1912) color. The lithologic descriptions from sediment cores were grouped into the five lithologic categories (fig. 9) from the lithologic-log standardization. Saturated horizontal hydraulic conductivity values from the HPT profiles were then matched to each described core depth interval to determine the relation between lithology and horizontal hydraulic conductivity. An average horizontal hydraulic conductivity was determined for the lithologic categories contained in each core.

Percent-Coarse Value and Hydraulic Properties

To determine the hydraulic conductivity of the Canadian River alluvial aquifer, each lithologic category determined from the lithologic logs was assigned a percent-coarse range. The percent-coarse ranges were 0–20 percent (clay/silt), 21–40 (fine sand), 41–60 (medium sand), 61–80 (coarse sand), and 81–100 (fine gravel). A percent coarse multiplier was based on the midpoint of each category (fig. 9), and was used to assign a percent-coarse value to each lithologic log depth interval. The percent-coarse value for each lithologic log was then computed as the thickness-weighted average of percent-coarse values assigned to the lithologic categories in each log.

A horizontal hydraulic conductivity of 0.1 feet per day (ft/d) was assigned to the clay/silt standardized category, and a horizontal hydraulic conductivity of 100 ft/d was assigned to the fine gravel standardized category on the basis of the weighted-average horizontal hydraulic conductivity values calculated for each lithologic category across each of the six sediment cores determined by HPT profiles. This range spans the expected grain sizes encountered in the Canadian River alluvial and terrace material. By assuming that a horizontal hydraulic conductivity of 0.1 ft/d and 10 percent-coarse value represent the clay/silt standardized category, that a horizontal hydraulic conductivity of 100 ft/d and 90 percent-coarse value represent the fine gravel standardized category, and that the relation between the two parameters is linear, average horizontal hydraulic conductivity for the aquifer material was assumed to be

$$K_h = (1.25 \times P_s) - 12.4 \quad (1)$$

where

K_h is the horizontal hydraulic conductivity in feet per day (ft/d), and

P_s is the percent-coarse value.

The frequency distribution of the horizontal hydraulic conductivity is shown in figure 10. The average, minimum, and maximum horizontal hydraulic conductivities for the Canadian River alluvial aquifer were calculated to be 39 ft/d, 0.1 ft/d, and 100 ft/d, respectively. These values are similar to those of Scholl and Christenson (1998) in which the average,

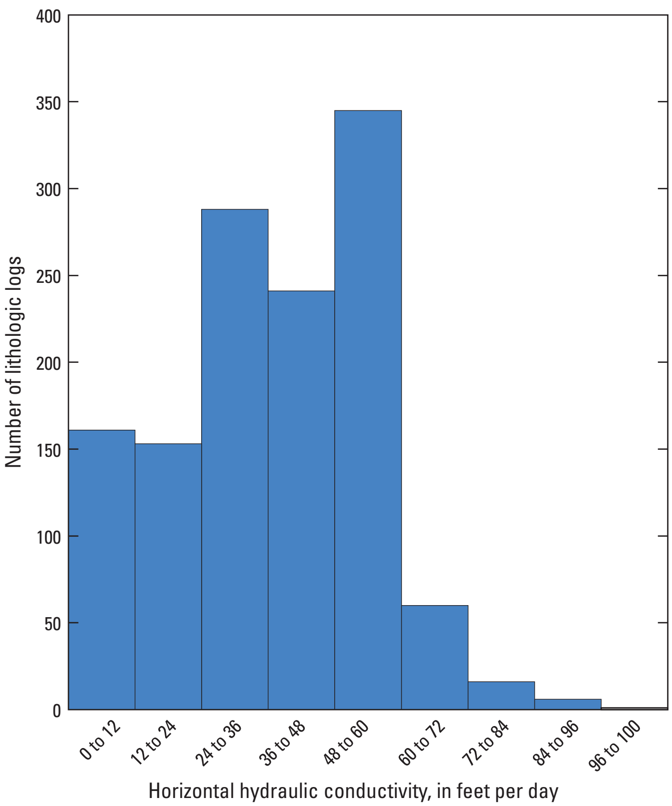


Figure 10. Frequency distribution of horizontal hydraulic conductivity determined from lithologic logs.

minimum, and maximum horizontal hydraulic conductivities for the Canadian River alluvial aquifer were determined to be 19 ft/d, 0.24 ft/d, and 79 ft/d. The Canadian River alluvial aquifer materials become finer grained with distance downstream to the east, possibly because of the longer average distance from the source of the sediment, or possibly due to a decline in the streambed elevation gradient. Local heterogeneities in the percent-coarse values of alluvial aquifer materials may be related to variations in the depositional environment and history or the variability in the lithologic-log descriptions provided by well drillers. Horizontal hydraulic conductivity for the underlying bedrock units was estimated by using values from Morris and Johnson (1967) to assign a uniform horizontal hydraulic conductivity of 0.2 ft/d for the bedrock units, which consisted predominantly of fine sandstone.

At the time this report was written (2016), no publications were available describing multiwell aquifer tests in the Canadian River alluvial aquifer; therefore, a specific storage value of 0.0001^{-1} was chosen from Domenico and Mifflin (1965), and a specific yield for each lithology category was chosen from Morris and Johnson (1967). The specific-yield values chosen were 6 percent for clay/silt deposits, 33 percent for fine sand, 32 percent for medium sand, 30 percent for coarse sand, and 28 percent for fine gravel. The specific yield for each lithologic log was then determined as the thickness-weighted average of specific yield assigned to the lithologic categories in each lithologic log. An average specific yield of 22 percent was calculated for the Canadian River alluvial aquifer. The specific-yield values used from Morris and Johnson (1967) were based on a laboratory determination of specific yield. Laboratory-determined specific yield is often much larger than values obtained from aquifer tests and may be more applicable for the evaluation of long-term aquifer yields than groundwater changes from pumping (Neuman, 1987).

Spatial interpolation of the percent-coarse values and specific-yield values was accomplished by using the Inverse Distance Weighting (IDW) method between lithologic-log locations, whereby values closest to the location to be interpolated are assigned the greatest weights (Esri, 2015). The IDW method assumes that nearby values provide more accurate data, which is appropriate for an aquifer where localized lithological trends may not continue for more than short distances. The IDW interpolation used a power function of two, which represents the exponent to which the inverse of the distance between the interpolated location and nearby values is raised. Additionally, an IDW search radius of 2 mi was applied to limit interpolations in areas where the distance between values was substantial, such as areas near the Texas border.

Hydrogeology of the Canadian River Alluvial Aquifer

This section describes the hydrogeology of the Canadian River alluvial aquifer. The information described in this section was used to create the conceptual flow model and forms the basis of the groundwater-flow models described in the “Simulation of Groundwater Flow in the Canadian River Alluvial Aquifer” section.

Groundwater

Groundwater in the Canadian River alluvial aquifer refers to water in the saturated zone between the water table and the underlying bedrock. The movement of groundwater in the saturated zone is described by Darcy’s Law (Darcy, 1856), by which groundwater flows from areas of high hydraulic head to areas of low hydraulic head through a porous medium. Groundwater inflows to the Canadian River alluvial aquifer include recharge to the water table from precipitation and groundwater lateral flow from the terrace contact with surrounding bedrock. Groundwater outflows include evapotranspiration (ET), groundwater use by wells, and some groundwater lateral flow from the terrace to the surrounding bedrock.

Temporal Water-Level Fluctuations

Depth to water was recorded continuously in eight wells (continuous recorder wells) in the Canadian River alluvial aquifer study area during the period 2014–16 (pl. 1, figs. 11 and 12). Water levels in the Canadian River alluvial aquifer can fluctuate 2–10 ft on a timescale of days, depending on distance from the river. Stressors affecting these water levels in the alluvial aquifer are precipitation, ET, groundwater use by wells, and streamflow. Water levels measured in wells in Reach I of the Canadian River alluvial aquifer near Crawford, Camargo, Taloga, and Bridgeport show a correlative response to precipitation data (fig. 11) from west-central Oklahoma Climate Division 4 (fig. 1). All continuous recorder wells, except for two, were completed in the alluvial deposits. Wells in Reach II were located near Purcell, Norman (Indian Hills), Norman (NLF TS5), and Konawa. These wells also show a correlative response to precipitation data (fig. 12) from central Oklahoma Climate Division 5 (fig. 1). The well near Konawa is completed in alluvial deposits but is also open below the alluvial deposits into the Pennsylvanian-age Vanoss Formation. The well near Norman (Indian Hills) is completed in the terrace deposits of the alluvial aquifer. The highest water levels in both reaches occurred in May 2015 (figs. 11 and 12) which corresponds to the greatest average monthly precipitation during the study period (fig. 4). Water levels also show a correlative response to below-average precipitation over the summer months and groundwater use by wells that are pumped during the crop-growing season (April to September). The lowest groundwater levels were measured in both reaches in October 2014, which was at the end of the 2014 growing season during this 2-year period (figs. 11 and 12). Groundwater levels were also low in both reaches in September 2015; however, more rain fell during the April–June 2015 period compared to the April–June 2014 period, buffering water levels from declining to the previous September–October levels. The range in water levels in the well near Konawa is less variable than the water levels in wells completed in the alluvial deposits. The well near Norman (Indian Hills) had less variable water levels than the other wells completed in the alluvial aquifer but showed similar trends as the groundwater levels measured in other wells in Reach II in response to precipitation (fig. 12).

Water-Table Surface in 2013

The water-table surface, which is usually interpolated or contoured from water-level altitude measurements at many individual wells, approximates the top of the zone of saturation. Water levels were measured between January and March 2013 in 140 wells completed in the Canadian River alluvial aquifer. Most of these wells were constructed with slotted polyvinyl chloride or steel casings with sand backfill on the outside of the well casings. The water-table surface in the Canadian River alluvial aquifer was shallow, typically less than 30 ft below land surface. The measured water levels were supplemented by additional water levels from OWRB well-completion reports (Oklahoma Water Resources Board, 2015a) which specified the depth interval at which water was first encountered during drilling.

Analysis of the 2013 water-table surface map (pl. 3) indicates that groundwater in the terrace deposits flows from the surrounding bedrock units and towards the alluvial deposits and main river channel. Higher water levels in the surrounding bedrock than in the alluvial deposits indicates that the Canadian River has incised down into the bedrock, with base-flow seepage from bedrock units through the alluvial deposits to the river being part of this hydrologic system. This incision was not observed in the Beaver-North Canadian River alluvial aquifer (Ryter and Correll, 2016). Groundwater in the alluvial deposits typically flows subparallel and downstream with streamflow. In areas where the water-table contours created a “V” more sharply upstream (the western part of the aquifer), groundwater discharges more readily to the river. The Canadian River is generally a gaining stream (deriving part of its flow from groundwater seepage) based on the water-table surface contours.

Groundwater Flow to and from Bedrock Units

Groundwater flow between the Canadian River alluvial aquifer and the adjoining bedrock units, or lateral flow, has been documented by previous studies. In central Oklahoma, Mogg and others (1960) reported lower groundwater levels in the Canadian River alluvial aquifer relative to the surrounding bedrock, particularly the Rush Springs Formation, where groundwater was determined to drain towards the Canadian River at an average water-table slope of 35 ft/mi. Seepage from topographically higher bedrock units generally located 20–100 ft or more above the alluvial deposits in central Oklahoma was also reported by Wood and Burton (1968).

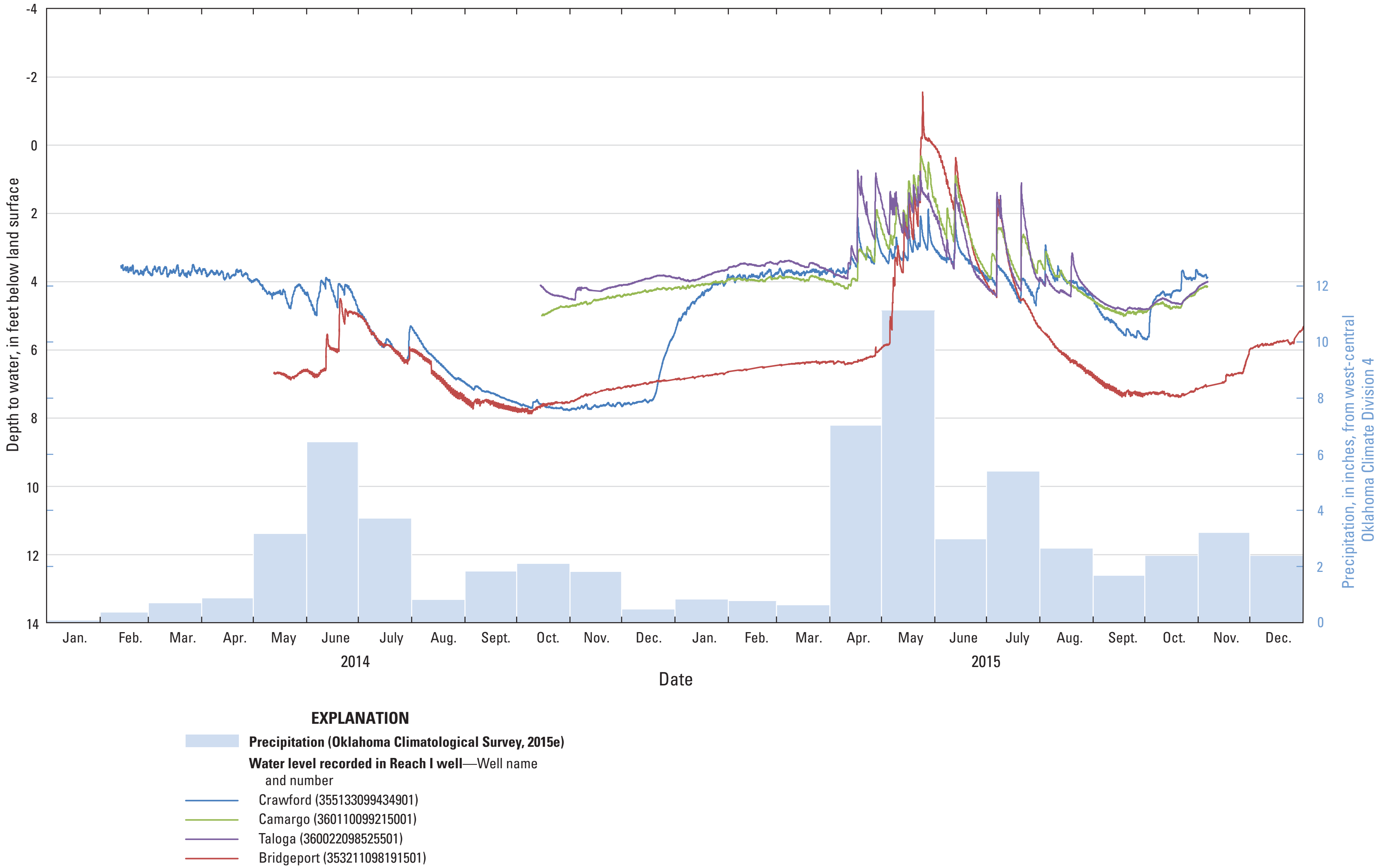


Figure 11. Continuous water levels recorded in wells and precipitation in Reach I of the Canadian River alluvial aquifer, Oklahoma.

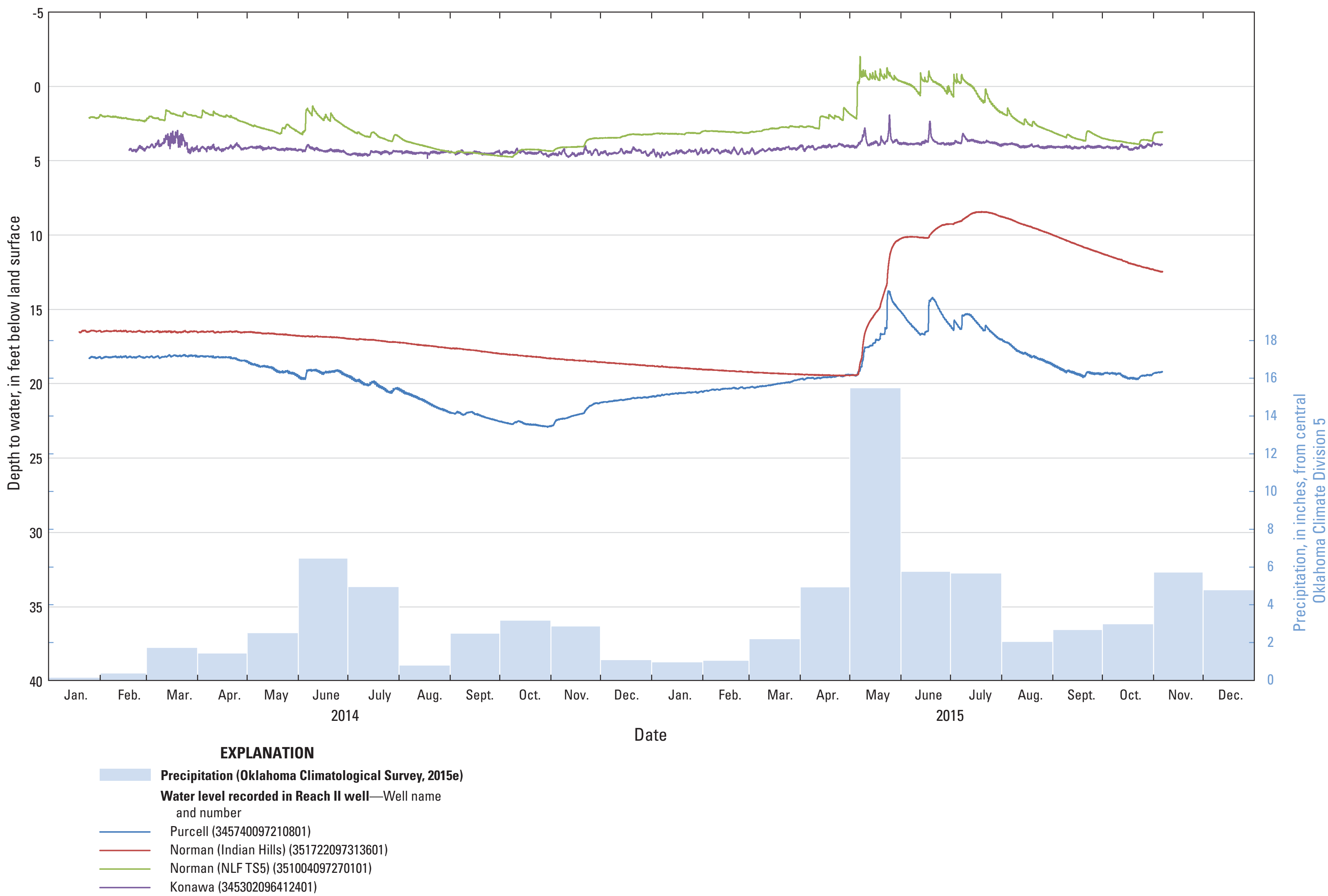


Figure 12. Continuous water levels recorded in wells and precipitation in Reach II of the Canadian River alluvial aquifer, Oklahoma.

Water Quality

Groundwater samples from 35 wells completed in the Canadian River alluvial aquifer were collected by the OWRB as part of the Groundwater Monitoring and Assessment Program (Oklahoma Water Resources Board, 2015b) in August and September 2013. Those groundwater samples were analyzed for physical properties (pH, temperature, dissolved oxygen, hardness, alkalinity, and specific conductance) in the field and for concentrations of major ions, trace metals, and nutrients in a laboratory (table 3). Data from 4 of the 35 samples were not included in the results in this report because of the greater than 10-percent discrepancies in cation-anion balances of those samples. Five of the remaining 31 groundwater samples contained nitrate-nitrogen concentrations (measured as nitrate plus nitrite concentrations) that exceeded the National Primary Drinking Water Regulation of 10 milligrams per liter (mg/L) (U.S. Environmental Protection Agency, 2016), with the maximum measured nitrate-nitrogen concentration for all of these samples being 16.1 mg/L. Twenty-one of 31 groundwater samples exceeded the National Secondary Drinking Water Regulation for dissolved solids concentration (500 mg/L), with the median dissolved solids concentration for all samples being 530 mg/L. Eleven of 31 groundwater samples exceeded the National Secondary Drinking Water Regulation for sulfate concentration (250 mg/L), with the median detected sulfate concentration for all samples being 103 mg/L. Twelve of 31 groundwater samples exceeded the National Secondary Drinking Water Regulation for iron concentration (300 micrograms per liter [µg/L]), with the median detected iron concentration for all samples being 1,429 µg/L. Fourteen of 31 groundwater samples exceeded the National Secondary Drinking Water Regulation for manganese concentrations (50 µg/L), with the median detected manganese concentration for all samples being 457.5 µg/L (table 3).

Concentrations of dissolved cations and anions were summarized by using Piper (1944) diagrams. Water type was determined by using methods from Back (1961). Water types in the Canadian River alluvial aquifer ranged from calcium-magnesium-sulfate-chloride in the western part, transitioning into calcium-magnesium-bicarbonate towards the central part of the aquifer (fig. 13). Towards the eastern part of the aquifer, the water type was calcium magnesium-bicarbonate but increased in sodium-chloride concentrations (fig. 13). This transition in water type from calcium-magnesium-sulfate-chloride to calcium-magnesium-bicarbonate type water to more sodium chloride type water in the Canadian River alluvial aquifer can be explained by the exchange of groundwater containing dissolved constituents between the bedrock units and the alluvial aquifer. Concentrations of calcium, magnesium, and sulfate in the groundwater in the Cloud Chief Formation come from the dissolution of gypsum, dolomite, and calcite (Becker and Runkle, 1998). The calcium-magnesium-sulfate chloride type water in the Canadian River alluvial aquifer in the western part of the study area was most likely caused by the inflow of groundwater from the Cloud Chief Formation. The central parts of the Canadian River alluvial aquifer in the study area overlie Permian sandstones, siltstones, and mudstones. Groundwater in the Rush Springs Formation is predominantly calcium-magnesium bicarbonate (Becker and Runkle, 1998). Groundwater in the shallow, unconfined part of the Central Oklahoma aquifer contained predominantly calcium, magnesium, and bicarbonate ions from the dissolution of dolomite. The calcium-magnesium-bicarbonate type water in the Canadian River alluvial aquifer in the central part of the study area is most likely from the inflow of groundwater from the Rush Springs Formation and Garber Sandstone (Parkhurst and others, 1996). In the eastern part of the Canadian River alluvial aquifer, the water type was calcium-magnesium-bicarbonate, which is associated with Pennsylvanian-age bedrock units (Hart, 1974). The eastern part of the Canadian River alluvial aquifer also contained greater concentrations of sodium and chloride, which may be due to seepage of naturally formed brines, disposal of industrial wastes, or solution of minerals (Hart, 1974).

Surface Water

Streamflow in the Canadian River is described in this section to quantify estimates of streamflow magnitude, gaining and losing reaches, and contributions to streamflow from tributaries and groundwater. To estimate gaining and losing reaches in the Canadian River, a seepage run was conducted. To quantify stream base flow for the study period (1981–2013), hydrograph-separation methods were performed on the basis of data from five USGS streamgages.

Streamflow Characteristics and Trends

Flow in the Canadian River has been regulated since 1964 by the Sanford Dam in Texas (fig. 1) and influenced by five wastewater-treatment plants in Oklahoma (pl. 1). Daily-averaged monthly streamflow data (U.S. Geological Survey, 2013a) recorded at streamgages (fig. 1) are summarized in table 4. Streamflow data from streamgages were recorded during the entire study period (1981–2013), with the exception of the Norman streamgage (07229050) for which continuous streamflow data began in January 2007 and the Purcell streamgage (07229200) for which continuous streamflow data began in January 1986. Average streamflow ranged from 75 cubic feet per second (ft^3/s) at the Canadian streamgage (07228000) to 1,937 ft^3/s at the Calvin streamgage (07231500). Flows between each streamgage increased in the downstream direction (table 4). In the study area, the minimum streamflow typically occurred during August, and the maximum streamflow typically occurred during May at the streamgages. Streamflows in the Canadian River are sustained by groundwater discharge during most of the year; however, there are occurrences of zero flow in the Canadian River, as demonstrated by the minimum stream base flows at the Canadian streamgage (07228000), Bridgeport streamgage (07228500), and Calvin streamgage (07231500) (table 4).

Table 3. Summary statistics of water-quality constituent values for groundwater samples collected August–September 2013 from the Canadian River alluvial aquifer, Oklahoma.

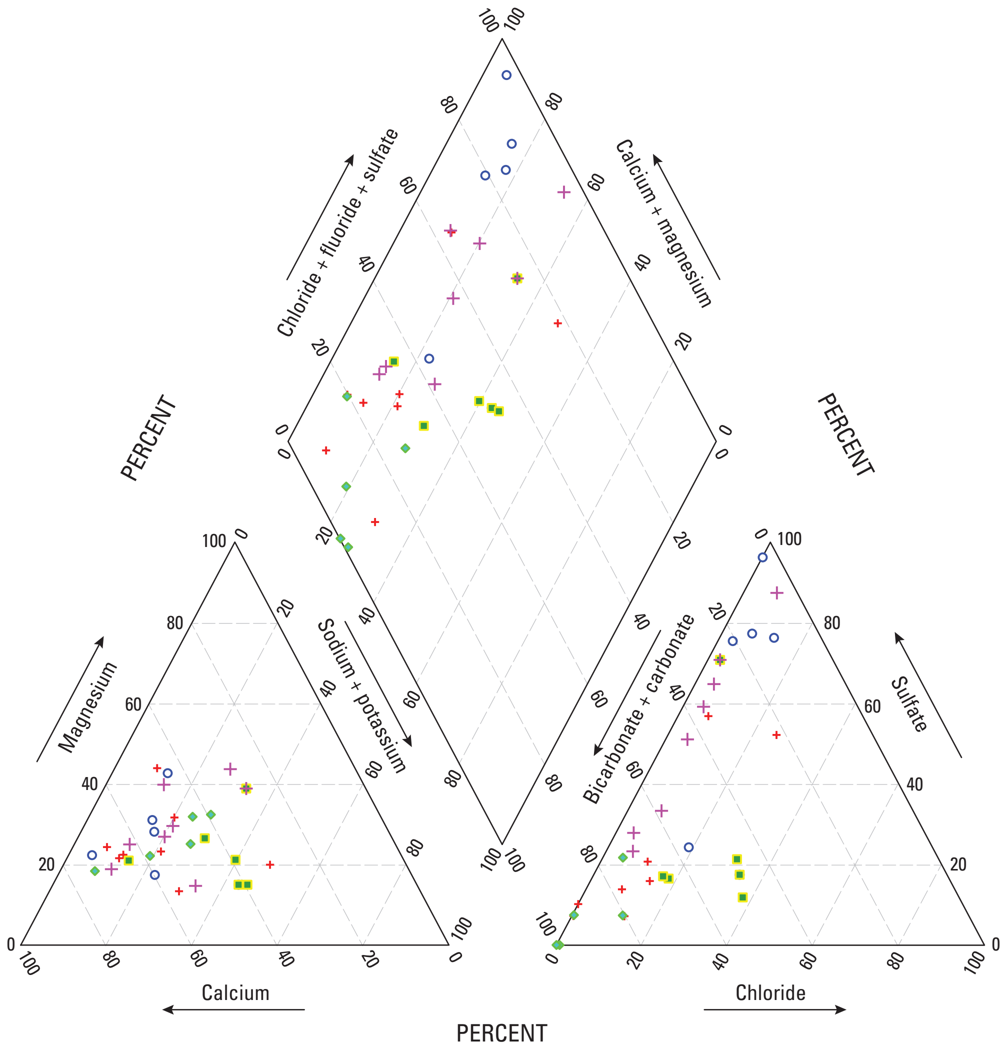
[mg/L, milligrams per liter; µg/L, micrograms per liter; --, not available]

Water-quality constituent	Detection limit	Number of samples analyzed	Number of samples less than detection limit	Minimum detected concentrations	Maximum detected concentrations	25th percentile of detected concentrations	50th percentile of detected concentrations	75th percentile of detected concentrations	Drinking water standard	Number of samples exceeding standard
Specific conductance	--	31	0	161.5	3,396.4	724.65	879.8	1,652.8	--	--
pH	--	31	0	6.25	7.45	6.865	7	7.105	6.5–8.5	2
Temperature	--	31	0	18.21	22.43	19.12	20.19	21.09	--	--
Dissolved oxygen, in mg/L	--	31	0	0.14	8.88	0.89	3.7	6.67	--	--
Hardness, in mg/L	--	31	0	46	1,484	293.5	386	811	--	--
Alkalinity, in mg/L	--	31	0	56	537	205.5	299	335	--	--
Dissolved solids, in mg/L	10	31	0	113	2,680	438	530	1,195	500	21
Calcium, in mg/L	5	31	0	16.7	445	80.6	111	207.5	--	--
Magnesium, in mg/L	5	31	0	5.3	180	16.75	39	65.55	--	--
Sodium, in mg/L	5	31	0	10.5	430	23	45.9	97.05	--	--
Potassium, in mg/L	0.5	31	3	0.5	5	1	2	2.725	--	--
Bicarbonate, in mg/L	12	31	0	68.8	661	253	368	412.5	--	--
Carbonate, in mg/L	6	31	31	--	--	--	--	--	--	--
Sulfate, in mg/L	10	31	2	16.6	1,750	74.7	103	650	250	11
Chloride, in mg/L	10	31	6	11.4	380	26.5	40.4	70.1	250	1
Fluoride, in mg/L	0.2	31	16	0.21	0.56	0.25	0.32	0.395	4	0
Bromide, in µg/L	100	31	3	127	966	241.75	331.5	459.25	--	--
Silica, in mg/L	50	31	0	11,000	54,200	20,050	21,700	24,750	--	--
Nitrate plus nitrite, in mg/L	0.05	31	8	0.07	16.1	0.925	3.79	6.92	10	5
Ammonia, in mg/L	0.1	31	24	0.2	0.97	0.235	0.26	0.405	--	--
Phosphorus, in mg/L	0.005	31	8	0.011	0.516	0.0285	0.042	0.0995	--	--
Aluminum, in µg/L	100	31	31	--	--	--	--	--	50–200	0
Antimony, in µg/L	1	31	31	--	--	--	--	--	6	0
Arsenic, in µg/L	10	31	29	12.2	19.9				10	2
Barium, in µg/L	10	31	1	10.5	987	26.05	92.15	230.75	2,000	0

Table 3. Summary statistics of water-quality constituent values for groundwater samples collected August–September 2013 from the Canadian River alluvial aquifer, Oklahoma.—Continued

[mg/L, milligrams per liter; µg/L, micrograms per liter; --, not available]

Water-quality constituent	Detection limit	Number of samples analyzed	Number of samples less than detection limit	Minimum detected concen-trations	Maximum detected concen-trations	25th percentile of detected concen-trations	50th percentile of detected concen-trations	75th percentile of detected concentrations	Drinking water standard	Number of samples exceeding standard
Beryllium, in µg/L	2	31	31	--	--	--	--	--	--	--
Boron, in µg/L	50	31	2	51.7	2,970	92.3	225	471	4	0
Cadmium, in µg/L	5	31	31	--	--	--	--	--	5	0
Chromium, in µg/L	5	31	29	6.1	6.4	--	--	--	100	0
Cobalt, in µg/L	10	31	31			--	--	--	--	--
Copper, in µg/L	5	31	29	6.7	13.1	--	--	--	1,300	0
Iron, in µg/L	50	31	15	59.6	4,940	507.5	1,429	2,237.5	300	12
Lead, in µg/L	10	31	31	--	--	--	--	--	15	0
Manganese, in µg/L	50	31	17	54	1,090	289	457.5	582.25	50	14
Mercury, in µg/L	0.05	31	30	0.73	0.73	--	--	--	2	0
Molybdenum, in µg/L	10	31	28	14.1	51.4	--	--	--	--	--
Nickel, in µg/L	10	31	31	--	--	--	--	--	--	--
Selenium, in µg/L	20	31	30	31.8	31.8	--	--	--	50	0
Silver, in µg/L	10	31	31	--	--	--	--	--	100	0
Thallium, in µg/L	1	31	31	--	--	--	--	--	2	0
Uranium, in µg/L	1	31	9	1.3	40.8	3	5.15	17.525	30	2
Vanadium, in µg/L	10	31	8	10.6	94.1	16.1	29.3	48.35	--	--
Zinc, in µg/L	10	31	13	10.2	424	11.675	25.45	67.9	500	0



EXPLANATION

Bedrock units adjoining or underlying the Canadian River alluvial aquifer and possibly contributing water types to the alluvial aquifer—Water type was determined by using methods from Back (1961)

West ↑ ↓ East

- Ogallala Formation
- Rush Springs Formation
- Dog Creek Shale and Duncan Sandstone
- Garber Sandstone
- Pennsylvanian units (undifferentiated)

Figure 13. Water types in wells completed in the Canadian River alluvial aquifer, Oklahoma, in 2013.

Table 4. Summary statistics of daily-averaged monthly streamflow and Base-Flow Index (BFI) stream base flow at streamgages in the Canadian River alluvial aquifer study area.

[All units in cubic feet per second. ft^3/s, cubic feet per second; TX, Texas; OK, Oklahoma]

Station number	Station name	Analyzed period of record	Average flow (ft^3/s)		Minimum flow (ft^3/s)		25th percentile flow (ft^3/s)	
			Stream	Base	Stream	Base	Stream	Base
07228000	Canadian River near Canadian, TX	1981–01–01 to 2013–12–31	75	46	0.0	0.0	31	21
07228500	Canadian River at Bridgeport, OK	1981–01–01 to 2013–12–31	326	142	0.3	0.0	88	30
07229050	Canadian River at Norman, OK	2007–01–01 to 2013–12–31	450	199	8.1	2.2	130	39
07229200	Canadian River at Purcell, OK	1986–01–01 to 2013–12–31	722	292	13.7	2.5	208	85
07231500	Canadian River at Calvin, OK	1981–01–01 to 2013–12–31	1,937	648	0.0	0.0	374	139

Station number	Station name	Analyzed period of record	Median flow (ft^3/s)		75th percentile flow (ft^3/s)		Maximum flow (ft^3/s)	
			Stream	Base	Stream	Base	Stream	Base
07228000	Canadian River near Canadian, TX	1981–01–01 to 2013–12–31	55	42	95	63	579	213
07228500	Canadian River at Bridgeport, OK	1981–01–01 to 2013–12–31	196	104	390	214	4,188	648
07229050	Canadian River at Norman, OK	2007–01–01 to 2013–12–31	319	155	517	294	2,988	1,060
07229200	Canadian River at Purcell, OK	1986–01–01 to 2013–12–31	426	229	825	424	7,717	1,329
07231500	Canadian River at Calvin, OK	1981–01–01 to 2013–12–31	970	393	2,421	778	20,643	4,941

Stream Base Flow

Stream base flow is defined for this report as the part of streamflow that is not from runoff and is discharged from aquifers adjoining a stream. Several methods can be used to quantify stream base flow. A seepage run consists of streamflow measurements at several places in a reach of a stream to determine if a section of the stream is receiving groundwater discharge or losing water to the underlying aquifer (Harvey and Wagner, 2000). A seepage run was done for this study in 2013. Base-flow separation techniques estimate the base-flow and runoff components of a streamflow record. The Base-Flow Index program (BFI) method (Wahl and Wahl, 2007) was used to estimate the average monthly and annual base-flow and runoff components of Canadian River daily streamflow records.

2013 Seepage Run

A seepage run was performed at 26 sites on the Canadian River during March 18–21, 2013, along a distance of 380 mi (fig. 14), beginning near the Texas-Oklahoma border. The seepage run was performed when ET and water use were considered to be minimal (during the winter and spring months in the study area) and when the runoff component of streamflow had dissipated. Streamflows measured in March 2013 were analyzed from upstream near Durham, Okla., referred to as the “Durham site,” to downstream near Calvin, Okla. Generally, the streamflow increased from upstream to downstream sites. Downstream increases in stream base flow could be caused by increased inflows to the stream and alluvial aquifer from adjoining bedrock aquifers. Average stream base flow from four perennial tributaries (Deer Creek, Buggy Creek, Walnut Creek, and Little River) to the Canadian River was estimated on the basis of available data. These streamflow estimates were subtracted from the net streamflow difference between the upstream and downstream sites from which they were located. All other intermittent streamflows were assumed to be zero.

Stream base flows remained similar from the Durham site to the Canadian River near Aledo, Okla., site, referred to as the “Aledo site.” Stream base flows steadily increased from 70 to 150 ft^3/s between the Aledo site and the Canadian River near Cogar, Okla., site, referred to as the “Cogar site” (fig. 14). Some of the increase in stream base flows between the Aledo and Cogar sites may have been caused from inflow from the local bedrock units (Rush Springs Formation). Stream base flows increased from 106 to 142 ft^3/s between the Canadian River near Thomas, Okla., site, referred to as the “Thomas site,” and the Bridgeport streamgage (07228500) (fig. 14). Deer Creek is a tributary between the Thomas site and the Bridgeport streamgage (07228500) that contributed an estimated 8.2 ft^3/s to the Canadian River in 2013, which would

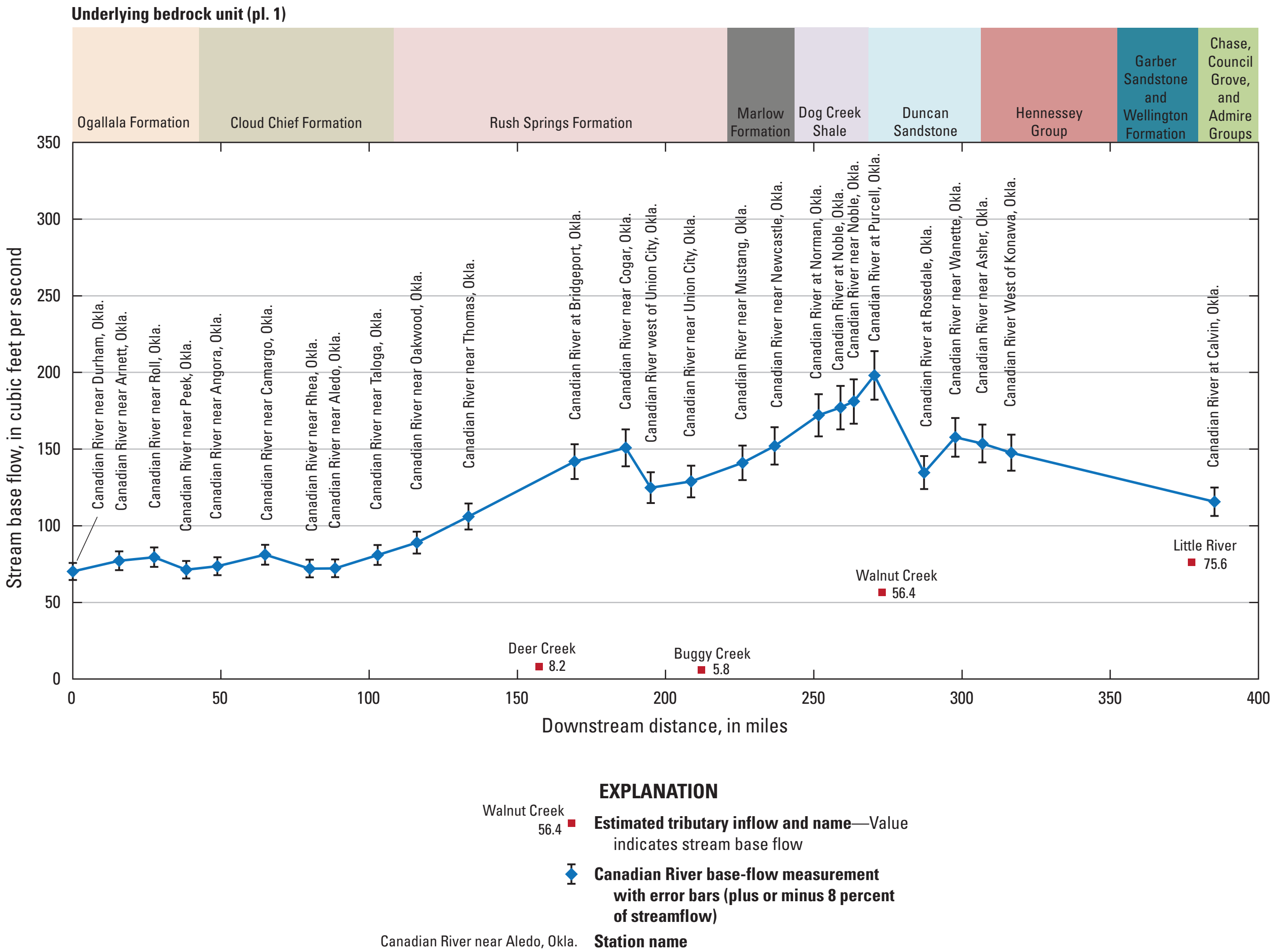


Figure 14. Stream base-flow measurements from the March 2013 seepage run conducted for the Canadian River alluvial aquifer and underlying bedrock units.

account for approximately 23 percent of the net flow increase measured between the Thomas and the Bridgeport streamgage (07228500). Stream base flows decreased from 151 to 125 ft^3/s between the Cogar site and the Canadian River west of Union City, Okla., referred to as the "Union City site" (fig. 14). No continuous water-table observations are available in this area; however, local differences in the water table between the Canadian River and the surrounding Rush Springs aquifer and Marlow Formation could result in a loss to the bedrock. Stream base flows increased from 125 to 197 ft^3/s between the Union City site and the Purcell streamgage (07229200) (fig. 14). Some of the increase in discharge between the Union City site and the Purcell streamgage (07229200) could be from inflow from the local bedrock unit (Rush Springs and Marlow Formations). The seepage run indicates that the Canadian River is generally a gaining stream between the Texas border and the Purcell streamgage (07229200) and a losing stream between the Purcell streamgage (07229200) and the Calvin streamgage (07231500) (fig. 14).

Base-Flow Separation

The BFI (Base Flow Index) method used to determine base flows is based on procedures developed by the Institute of Hydrology (1980). The BFI method divides the daily streamflow record into *n*-day increments and identifies the minimum streamflow during each *n*-day period. Minimum streamflows are then compared to adjacent minimums to determine turning points on a base-flow hydrograph. Straight lines drawn between turning points define the base-flow hydrograph; the area beneath the base-flow hydrograph is an estimate of the base-flow volume. The ratio of the base-flow volume to the total streamflow volume for the period of analysis is the base-flow index. Although these procedures may not always yield the true base-flow volume of the stream, tests in Great Britain (Institute of Hydrology, 1980), Canada (Swan and Condie, 1983), and the United States (Wahl and Wahl, 1995) indicated that the results of this base-flow separation procedure were consistent and indicative of the true base-flow volume (Esralew and Lewis, 2010).

For the five streamgages analyzed for this study along the Canadian River, the *n*-day partition length was tested over a range of values from zero to 10 days. A graph of the base-flow index compared to *n* was constructed to visually identify a slope change. The number of days where the slope no longer substantially changed, or the appropriate *n*-day value, was 5 days for streamflow data collected at the Canadian River streamgages.

The average annual BFI-computed base-flow index estimated by using the BFI method generally decreased from west to east at streamgages on the Canadian River (fig. 15); the average annual base-flow index was 63 percent at the Canadian streamgage (07228000) for 1981–2013, 45 percent at the Bridgeport streamgage (07228500) for 1981–2013, 47 percent at the Norman streamgage (07229050) for 2007–13, 41 percent at the Purcell streamgage (07229200) for 1986–2013, and 34 percent at the Calvin streamgage (07231500) for 1981–2013. This general decrease in base flow-index and increase in runoff may be due to the increased precipitation in Oklahoma from west to east (fig. 1, table 1), incised alluvial deposits in eastern Oklahoma (Tanner, 1956), and downstream decrease in horizontal hydraulic conductivity determined from this investigation. The base-flow index ranged from 33 percent in 1986 to 68 percent in 2003 (fig. 15). A higher than average base-flow index can be produced during extreme drought conditions (Institute of Hydrology, 1980). Year 2003 has the lowest recorded annual precipitation during the period 1981–2013; therefore, the large annual base-flow index in 2003 is probably due to the severe drought conditions.

Substantial increases in the average stream base flow were observed from upstream to downstream, particularly between the Purcell streamgage (07229200) and Calvin streamgage (07231500) (table 4). Stream base flows, determined from the seepage measurements at the Norman (07229050), Purcell (07229200), and Calvin (07231500) streamgages, including tributary inflows, were 7, 27, and 71 percent lower, respectively, than the average BFI-computed stream base flows for the period of record (1981–2013). This difference is likely due to the drought conditions during the 2013 stream base-flow measurements or from error in the BFI method. Because of less recharge to the alluvial aquifer and surrounding bedrock, the drought conditions reduced the altitude of the water table. As a result, lateral flow decreased from the bedrock to the alluvial and terrace deposits, as well as from the alluvial deposits to the Canadian River, particularly between the Purcell streamgage (07229200) and Calvin streamgage (07231500).

Streambed Hydraulic Properties

Estimated vertical saturated hydraulic conductivity of the streambed (K_{sb}) was measured by using falling-head permeameter tests with methods described in Fox and others (2011). These permeameter tests were completed at five streambed-hydraulic property sites near Konawa, Purcell, Crawford, Minco, and Thomas, Okla. (pl. 1). Each permeameter test was performed by driving a pipe into the streambed sediment to a depth of 20 centimeters (cm), adding a known amount of water in the pipe, and allowing the water level in the pipe to fall while measuring the head every 30 seconds for at least 5 minutes. Three permeameter tests were repeated to average results for a site to reduce potential human-induced error in each test. Calculations of K_{sb} were derived by using an applied version of Darcy's equation (Landon and others, 2001; Fox, 2004; Fox and others, 2011):

$$K_{sb} = \frac{d}{(t - t_0)} \ln\left(\frac{H_0}{H(t)}\right) \tag{2}$$

where

- $H(t)$ is the head above the streambed altitude at time (t),
- H_0 is the initial head in the pipe above the streambed altitude,
- d is the streambed sediment depth (20 cm), and
- $t\text{-}t_0$ is the elapsed time.

K_{sb} values calculated from these five falling-head permeameter tests ranged from 15 ft/d at the Thomas site to 54 ft/d at the Minco site, with an average value of 37 ft/d.

Recharge

Groundwater recharge was defined for this study as the portion of precipitation that enters the groundwater system at the zone of saturation and represents the primary means of water inflow to the Canadian River alluvial aquifer. Recharge rates are controlled by a number of factors including precipitation, land-surface gradient, soil and sediment hydraulic conductivity, ET, and vegetation cover type. Though recharge rates are difficult to quantify because of high spatial and temporal variability, methods using environmental tracers, physical measurements, streamflow-hydrograph techniques, and computer codes have been used to estimate recharge rates. For this study, a hydrograph-based water-table fluctuation method was used to estimate localized recharge rates for 2014, and a code-based water-balance estimation technique was used to estimate spatially distributed recharge rates for the period 1981–2013.

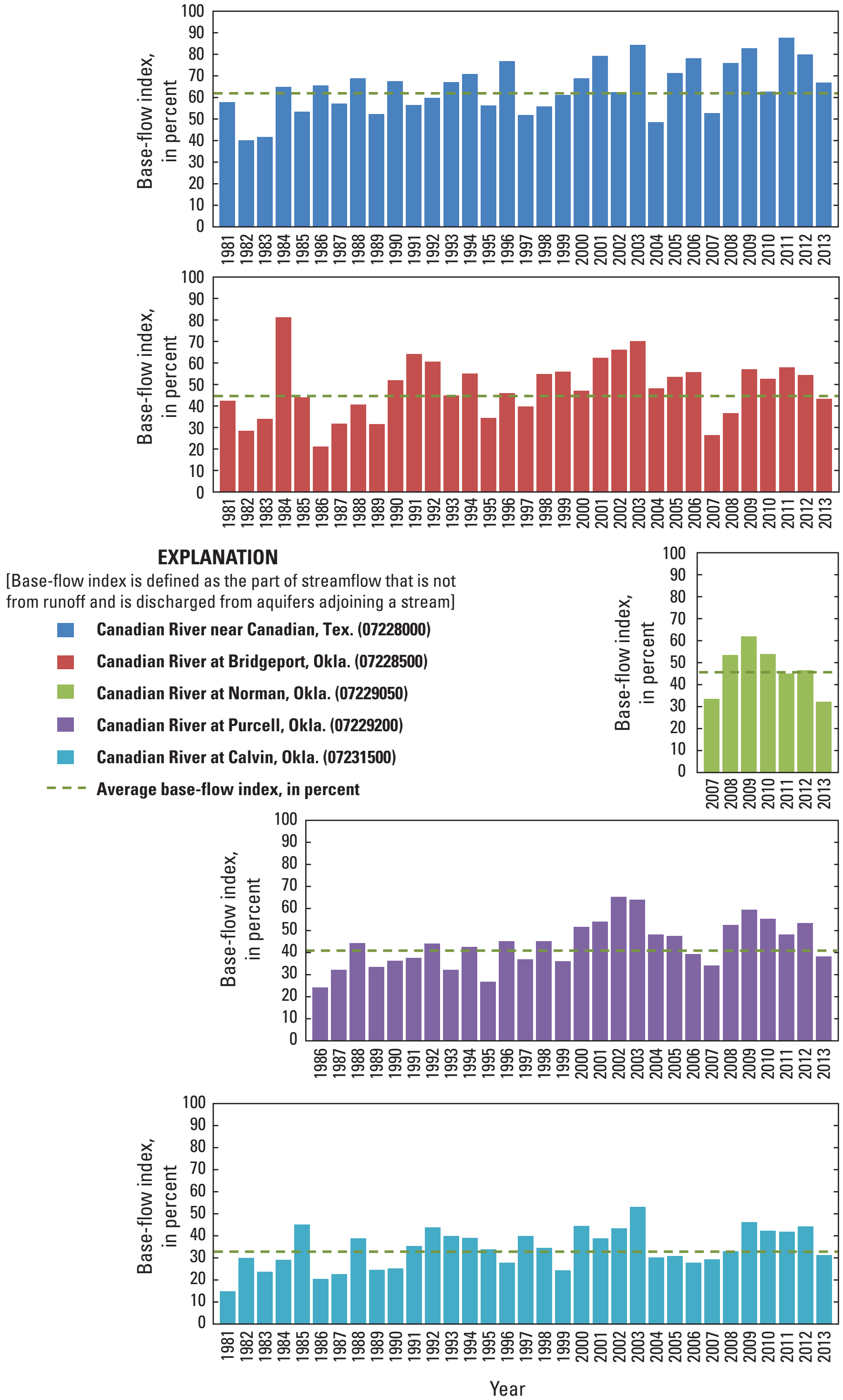


Figure 15. Average annual Base-Flow Index for streamgages along the Canadian River, Oklahoma, 1981–2013.

Water-Table Fluctuation Method

The Water-Table Fluctuation (WTF) method (Healy and Cook, 2002) was used to estimate recharge to the Canadian River alluvial aquifer. The WTF method is based on the premise that short-term (hours to a few days) rises in groundwater levels in unconfined aquifers are due to recharge arriving at the zone of saturation following a period of precipitation. This method is most appropriately applied to groundwater wells with shallow water tables and hydrographs that display sharp rises in water levels after precipitation. The WTF method cannot account for a steady rate of recharge or recharge from sources other than precipitation. Recharge is calculated by using the WTF method as follows:

$$R = Sy\Delta h / \Delta t \tag{3}$$

where

Sy	is the specific yield (dimensionless),
Δh	is the change in water table, in inches, and
Δt	is the change in time of the water-level rise, in inches.

A water-level hydrograph from the Crawford (355133099434901) well in Reach I (fig. 11) was analyzed by using the WTF method to estimate monthly recharge for 2014. The water-level hydrographs from other wells did not display enough sharp water-level rises, or the water-level rises were so great that they indicated that recharge was coming from a source other than precipitation. Annual precipitation during 2014 was 17.0 in., measured at the nearby Camargo climate station (Oklahoma Climatological Survey, 2015e). The minimum estimated monthly recharge at this well was 0.0 in. for the month of January because only 0.02 in. of precipitation fell during this month. The maximum estimated monthly recharge at this site was 6.2 in. during June, calculated by using the specific-yield value of 22 percent described in the “Percent-Coarse Value and Hydraulic Properties” section. Estimated annual recharge at this site was 9.7 in. (57 percent of annual precipitation) during 2014.

A portion of recharge estimated by using the WTF method at this alluvial site most likely flows through the sediments and discharges to the stream in days, so some of the estimated recharge is not accounted for on a monthly timescale. Additionally, a field-derived specific-yield value may be more appropriate for short-term changes in the water table (Scholl and others, 2005). As a result, the WTF method using the 22-percent specific yield may overstate recharge.

Soil-Water-Balance Code

The soil-water-balance (SWB) code (Westenbroek and others, 2010) also was used to estimate the spatial distribution of groundwater recharge to the Canadian River alluvial aquifer. The SWB code is based on a modified Thornthwaite-Mather method (Thornthwaite and Mather, 1957) and requires climatological and landscape characteristic data inputs including precipitation amount, temperature, soil-water storage capacity, hydrologic soil group, land-surface gradient, and land-cover type. These inputs are assigned to a user-specified grid (specified for this report as cell dimensions of 500 by 500 ft), and the SWB code uses a mass-balance approach to compute gridded recharge as the difference between sources and sinks, accounting for the cumulative effects of the change in soil moisture.

$$R = (P + S + Ri) - (Int + Ro + ET) - \Delta Sm \tag{4}$$

where

R	is recharge,
P	is precipitation,
S	is snowmelt,
Ri	is surface runoff into a cell,
Int	is plant interception,
Ro	is surface runoff out of a cell,
ET	is evapotranspiration, and
ΔSm	is the soil moisture change.

Daily climate data were obtained for 39 climate stations (fig. 1; National Climatic Data Center, 2015) and interpolated to a grid by using the IDW method (Esri, 2015). Of the 39 climate stations, 22 stations provided daily climate records throughout the study period of 1981–2013. In addition, these stations provided daily maximum and minimum air temperatures to allow for the SWB code to determine whether precipitation occurred as rain or snow.

Soil properties (soil-water storage capacity and hydrologic soil group) were derived from the Soil Survey Geographic database (Natural Resources Conservation Service, 2015). Surface runoff was calculated by using a flow-direction grid derived from a digital elevation model (DEM) (U.S. Geological Survey, 2013b), whereby values representing land-surface altitude were assigned to each SWB grid cell and runoff routed to downslope cells. In the SWB code, once water is routed to a closed surface, and the effects of ET from vegetation are accounted for, the only possible sink for the water is recharge, which can create situations where unreasonably high values are calculated. To ensure correct surface-water routing and to eliminate areas of internal drainage, depressions in the DEM were removed.

Evapotranspiration in the SWB code is calculated as potential ET by using methods from Hargreaves and Samani (1985), whereby spatially variable ET estimates are produced by using user-supplied daily minimum and maximum temperatures. This potential ET represents the maximum amount of ET possible when given no limitation to soil moisture. The change in soil moisture is based on Thornthwaite and Mather (1957), where the potential ET is subtracted from daily precipitation. The resulting positive values represent water surplus, and negative values represent a cumulative deficiency calculated as a running total. For positive values, typically periods of high precipitation, actual ET equals potential ET. The SWB code does not compute ET from the groundwater table and therefore underestimates ET in areas where groundwater occurs near land surface.

Scaling of vegetation root-zone values was necessary to allow for more recharge to the aquifer because stream base flow was otherwise substantially greater than inflows from recharge. Additionally, recharge obtained from the default root-zone values was 5.8 percent of precipitation, which was lower than the 7 to 11 percent range from literature describing recharge in Oklahoma (Barclay and Burton, 1953; Steele and Barclay, 1965) near the aquifer. Root-zone values represent the maximum depth to which various types of vegetation will grow and are classified on the basis of land use and soil type. Larger root-zone values result in the increased interception of soil moisture zone infiltration and thus decrease recharge, whereas smaller values increase recharge. For Reach I, the root-zone values for the forest and grass/pasture land-use types were scaled to 80 percent of the default root-zone values, and the crop land-use type was scaled to 50 percent of the default root-zone values. For Reach II, the root-zone values for the forest and grass/pasture land-use types were scaled to 50 percent of the default root-zone values, and the crop land-use type was scaled to 30 percent of the default root-zone values.

The SWB code produced areally distributed monthly and annual recharge grids for the Canadian River alluvial aquifer. A prolonged period of less than average recharge occurred for 8 of the 10 years during 2003–12 (fig. 16), and 2011 was the ninth driest year in Oklahoma since 1925 (Shivers and Andrews, 2013). Additionally, 1981 was the last year of the 1976–81 drought (Shivers and Andrews, 2013), and thus received less than average recharge. The SWB-estimated average annual recharge was 2.2 in/yr for Reach I (fig. 16*A*) and 4.2 in/yr for Reach II (fig. 16*B*). The SWB-estimated average annual recharge to the aquifer for both reaches was 3.4 in/yr, or 8.7 percent of the average annual precipitation for the period 1981–2013 (fig. 16*C*). The highest SWB-estimated annual recharge for both reaches was 5.9 in/yr (173 percent of the average annual recharge [1981–2013]) in 1985 (fig. 16*C*). The lowest SWB-estimated annual recharge for both reaches was 0.6 in/yr (18 percent of the average annual recharge [1981–2013]) in 2003 (fig. 16*C*).

A map of the spatially distributed average annual recharge in inches per year calculated by the SWB code is shown in plate 4. The decreased recharge in Reach I was expected to occur because of the lower precipitation from the semiarid climate versus the humid subtropical climate in Reach II. In Reach I, more precipitation occurred as recharge in the alluvial deposits than the terrace deposits because of a more sandy soil type that had a lower soil-water storage capacity. In Reach II, although the alluvial deposits were also of a more sandy soil type than the terrace deposits, less precipitation occurred as recharge in the alluvial deposits because of the lower infiltration rate from the assigned hydrogeologic soil group. Evapotranspiration was also higher in the Reach II alluvial deposits compared to the terrace deposits because of the predominance of the deeper-rooted grass/pasture land-cover type in the alluvial deposits, even after scaling of the root-zone values.

Compared to the WTF-estimated recharge, the SWB-estimated average annual recharge for Reach I is substantially lower. Although recharge in 2014 was not estimated by using SWB, the high percentage of precipitation as recharge from the WTF method results in a much larger annual recharge for any given year. It is expected that the SWB-estimated recharge is a better estimate of recharge in this area based on previous reports that determined recharge as 12 percent of precipitation for the North Canadian River alluvial aquifer (Mogg and others, 1960) and 10 percent for the Rush Springs aquifer (Tanaka and Davis, 1963).

Evapotranspiration from Groundwater

Evapotranspiration from groundwater was estimated by Scholl and others (2005) for the Canadian River alluvial aquifer in Norman, Okla., by using methods from White (1932). Uptake from plant transpiration was determined by using specific yield from soil columns and water-level measurements taken every hour from a continuous well for the following equation:

$$q = Sy\left(24r \pm s\right) \tag{5}$$

where

- q is the daily-averaged ET in inches,
- Sy is the specific yield (dimensionless),
- $24r$ is the average daily rise of the water table in inches per day (in/d), and
- s is the decline in the water table during a 24-hour period in inches.

By using this method, average, maximum, and minimum ET from groundwater values were estimated as 0.21, 0.51, and 0.10 inches per day, respectively.

The water table at the tested site was within 5.7 ft of land surface, which likely resulted in the relatively large contribution of evaporation and transpiration to water losses and may not characterize the alluvial deposits in other areas of the Canadian River alluvial aquifer where evaporative effects are minimal because of greater water-table depths. Additionally, the 22 percent specific yield may result in an overstated amount of ET from groundwater. As a result, the minimum ET value of 0.10 in/d was used in the conceptual model, which may provide a more realistic estimate of ET from groundwater.

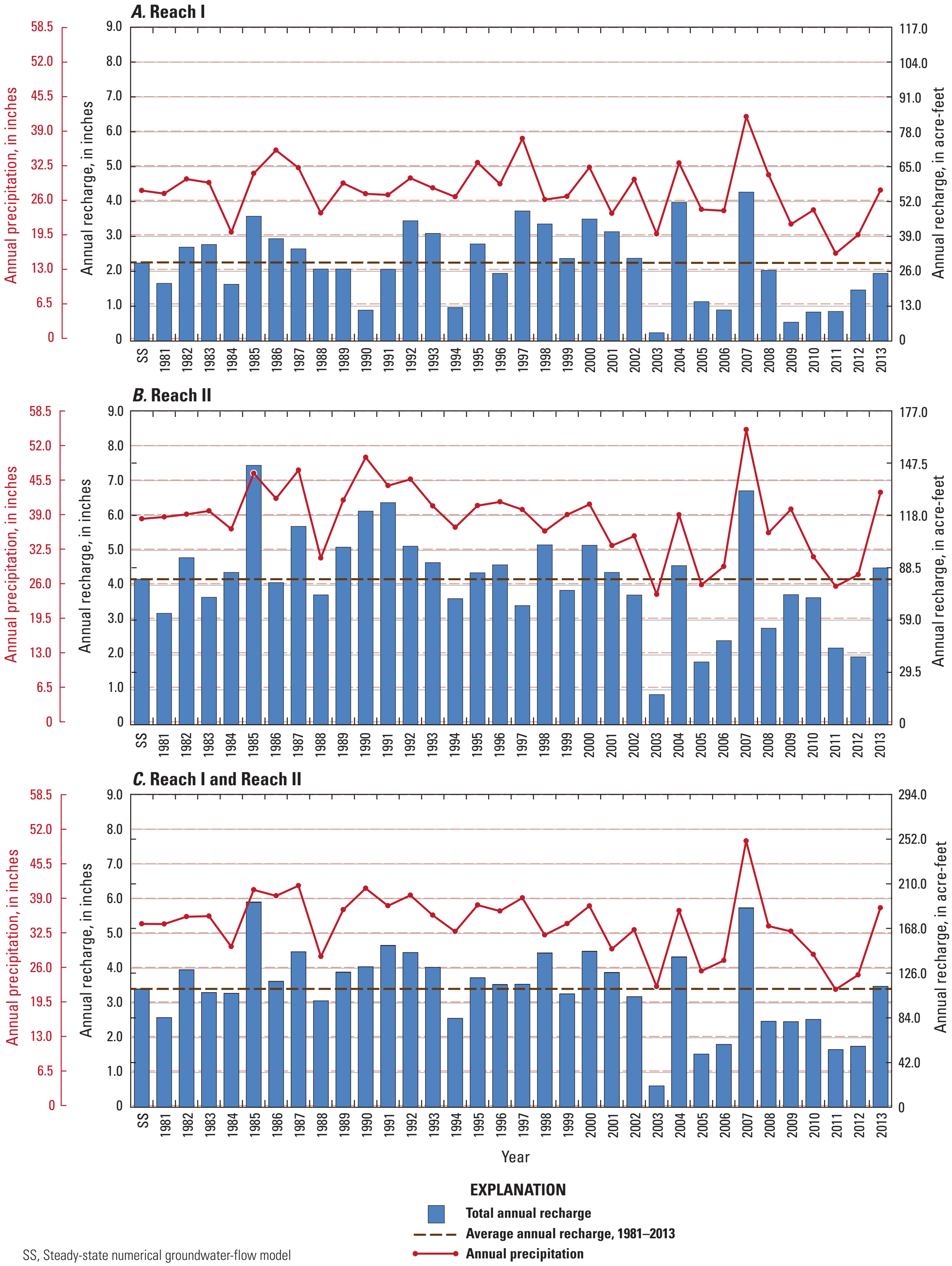


Figure 16. Annual recharge from 1981 to 2013 by using the soil-water-balance (SWB) code for *A*, Reach I; *B*, Reach II; and *C*, Reaches I and II of the Canadian River alluvial aquifer, Oklahoma.

Conceptual Flow Model of the Canadian River Alluvial Aquifer

A conceptual flow model is a simplified description of the movement of water in and out of an aquifer as defined by the hydrogeologic framework. Primary components of the conceptual flow model include the hydrogeologic boundaries, which describe the conditions for flow between the aquifer and surrounding units, and an aquifer conceptual water budget, which describes the amount of groundwater that moves between each hydrogeologic boundary.

Hydrogeologic Boundaries

The land surface of the Canadian River alluvial aquifer is the most extensive hydrogeologic boundary through which precipitation is either evapotranspired or recharges the water table. The base of the Canadian River alluvial aquifer is the hydrogeologic boundary through which water can move to or from the underlying bedrock units. Stream boundaries allow flow to be exchanged between the Canadian River alluvial aquifer and the Canadian River (or tributaries). The lateral extent of the Canadian River alluvial aquifer is a hydrogeologic boundary in most areas where bedrock units adjoining the Canadian River alluvial aquifer yield water to this aquifer. Lateral inflow is expected on the basis of previous investigations (Mogg and others, 1960; Wood and Burton, 1968) and on the distribution of water types described in the “Water Quality” section of this report.

Conceptual Water Budget

A conceptual water budget was constructed for the Canadian River alluvial aquifer to quantify the addition and removal of water from the boundaries of the aquifer, with estimated fluxes listed in table 5. The conceptual water budget is a generalized average of water fluxes for the model period 1981–2013. The aquifer alluvial and terrace deposits are considered a single unit; thus, water that flows from the Canadian River alluvial aquifer to the Canadian River is an aquifer outflow, whereas the reverse constitutes an inflow to the aquifer. The same is true for lateral flow from and to the surrounding bedrock. Inflows to the Canadian River alluvial aquifer include recharge to the water table from precipitation, lateral flow from adjoining bedrock units, and flow, or leakage, from the Canadian River, whereas outflows include lateral flow to bedrock, water-table ET, flow to the Canadian River, and groundwater use. Wells associated with water-table observations, or observation wells (fig. 17), do not show any substantial long-term water-table trends; therefore, the net change in groundwater storage is assumed to be negligible.

Table 5. Conceptual water budget for the Canadian River alluvial aquifer, Oklahoma, 1981–2013.

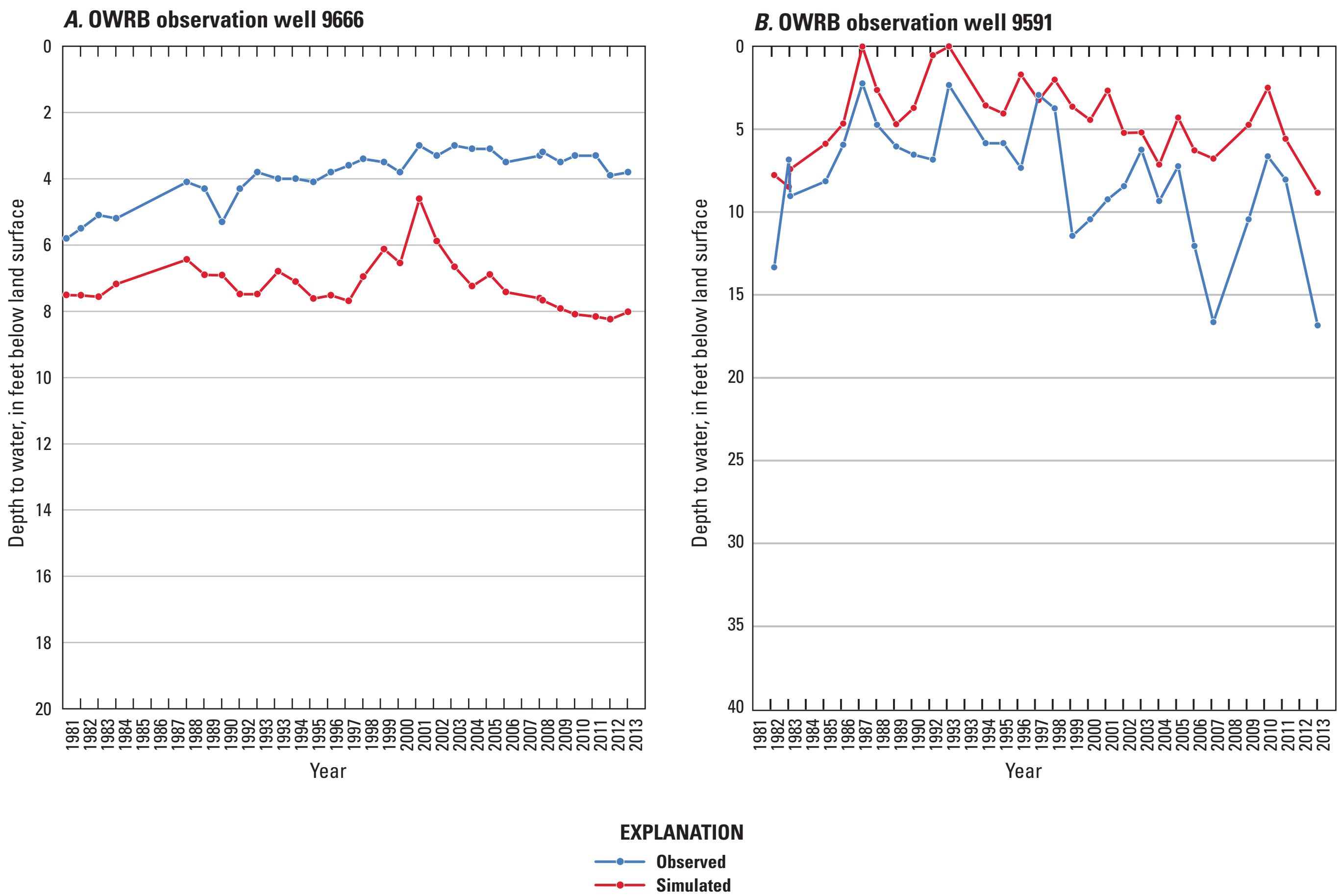
[All units are in acre-feet per year. See table 4 for station names. --, not applicable]

Groundwater budget category	Reach I	Reach II	Total
Inflows			
Recharge	28,919	82,006	110,925
Lateral flow[1]	43,803	164,849	208,652
Outflows			
Evapotranspiration	34,123	32,061	66,184
Base-flow gain[1]	38,006	205,377	243,383
Texas border to 07228500	38,006	--	--
07228500 to 07229050	--	44,849	--
07229050 to 07229200	--	29,768	--
07229200 to 07231500	--	130,759	--
Groundwater use	593	9,416	10,009

[1]These terms represent net flows. Base-flow gain was calculated as the difference between flow to and from the Canadian River. Lateral flow was calculated as the difference between the aquifer inflows and the outflows.

Total annual recharge inflows estimated by the SWB code, averaging 2.2 in for Reach I and 4.2 in for Reach II, were multiplied by the area of the watershed in each reach, then averaged over the simulated period to produce an annual average of 28,919 acre-ft/yr for Reach I and 82,006 acre-ft/yr for Reach II (fig. 16, table 5). A more detailed description of the SWB-derived recharge is available in the “Soil-Water-Balance Code” section of this report.

Net flow, or the difference between flow to and from the Canadian River, was predominantly towards the Canadian River from the aquifer. This flow is hereinafter referred to as “base-flow gain.” Flow from the Canadian River to the aquifer may occur during periods of low recharge, as shown by the 2013 seepage run measurements. However, the BFI analysis showed substantial downstream increases in average base flow for the period 1981–2013; therefore, the BFI-computed base flow (see “Base-Flow Separation” section of the report) was used in the conceptual model. Average annual base-flow gain in Reach I and Reach II was determined as the difference between stream base flow at the most upstream and downstream gages. The effects of surface-water flows were removed from the average annual base-flow gain estimation, which included tributary and wastewater treatment plant flows (Vikki Southerland, Oklahoma Department of Environmental Quality, written commun., 2015). The base-flow gain estimate is a simplification of a more complex system that includes



OWRB, Oklahoma Water Resources Board

Figure 17. Observed and simulated depth to water for water-table observations in Oklahoma Water Resources Board (OWRB) wells *A*, 9666, and *B*, 9591, in the Canadian River alluvial aquifer, Oklahoma.

other factors that influence the base-flow gain, such as regulation from upstream reservoirs; however, no releases from Sanford Dam in Texas were recorded during the period 1981–2013. Average annual base-flow gain during the period 1981–2013 was estimated to be 38,006 acre-ft/yr for Reach I, and 205,377 acre-ft/yr for Reach II (table 5). Most of the base-flow gain in Reach II occurred between the Purcell streamgage (07229200) and the Calvin streamgage (07231500) (table 5). The Canadian River alluvial aquifer was determined to contribute base flow of 0.3–2.3 ft^3/s per mile to the Canadian River.

Water-table ET was estimated by using methods from Scholl and others (2005) previously described in the “Evapotranspiration from Groundwater” section of this report. Evapotranspiration was estimated by using the 0.10-in/d rate applied to areas of the alluvial deposits with a water-table elevation near land surface. Because the ET rate from Scholl and others (2005) was based on a single site, the estimated ET for the aquifer may differ from the actual ET.

Net lateral flow was calculated as the difference between the aquifer inflows (recharge) and the outflows (base-flow gain, water-table ET, and groundwater use). The substantial lateral flow into the aquifer was expected based on an analysis of the 2013 water-table surface and documentation from previous studies. The large base-flow gain in Reach II between the Purcell streamgage (07229200) and the Calvin streamgage (07231500) indicates that most of the lateral inflow in Reach II may occur in this area of the aquifer. In other areas of the Canadian River alluvial aquifer, lateral flows are expected to produce smaller amounts of inflow. Groundwater use accounts for a minor percentage of outflows from the Canadian River alluvial aquifer, totaling 593 acre-ft/yr in Reach I and 9,416 acre-ft/yr in Reach II.

Simulation of Groundwater Flow in the Canadian River Alluvial Aquifer

Numerical groundwater-flow models were constructed on the basis of the hydrogeologic framework and conceptual flow model. The groundwater-flow models used the Newton solver (Niswonger and others, 2011), which is based on the MODFLOW code (Harbaugh, 2005), to simulate groundwater levels, interactions with surface-water features, and changes in streamflow and aquifer storage. These models were calibrated to better reproduce observed characteristics; improve the accuracy of future-predictive scenarios run to evaluate the effects of groundwater use over 20-, 40-, and 50-year periods; and observe the effects of a hypothetical 10-year drought and of groundwater-use rates in 2013 extended 50 years into the future.

Model Extents and Configuration

The Canadian River alluvial aquifer was split into two groundwater-flow models. Reach I extends from the Texas border to the Bridgeport streamgage (07228500); Reach II extends from the Bridgeport streamgage (07228500) to Eufaula Lake (pl. 1). This division was chosen because of the presence of the Bridgeport streamgage (07228500), which allows for surface-water inflow in the Reach II groundwater-flow model to be set to the stream base flow determined at this streamgage. The two models function independently and do not have linked flows. To aid groundwater-flow model assessment, individual water budgets are described in this report for Reaches I and II.

Layer 1 of the models represented the undifferentiated Quaternary-age alluvial and terrace deposits, and layer 2 represented undifferentiated bedrock units. The land-surface altitude of the models was represented by a DEM (U.S. Geological Survey, 2013b) and formed the top of layer 1, and the altitude of the bedrock surface formed the top of layer 2 for both reaches. The thickness of layer 1 included a maximum of 117 ft in Reach I near the Texas border and 139 ft in Reach II near Eufaula Lake. Layer 2 was assigned an arbitrary and uniform thickness of 100 ft in both models. The bottom of layer 2 was a no-flow boundary. Bedrock outcrops in the aquifer areal extent were also simulated as no-flow boundaries.

Initial values for horizontal hydraulic conductivity, specific yield, and specific storage for the Canadian River alluvial aquifer, represented by layer 1, are described in the “Percent-Coarse Value and Hydraulic Properties” section of this report. Vertical anisotropy was set to 3.0 for layer 1 and 5.0 for layer 2, as described by Todd (1980). Specific storage was set to 0.0001 ft^{-1} and specific yield was set to 20 percent for layer 2, as described by Domenico and Mifflin (1965).

Discretization

A grid of 500-by-500-ft cells was used to represent the groundwater-flow model domains (pl. 5). This cell size was chosen to best capture variations in hydrogeologic parameters and boundary conditions without unduly slowing simulation speeds that arise from having large numbers of small cells in the model area. The Reach I model grid consisted of 589 rows, 1,059 columns, and 2 layers, with 27,041 active cells in each layer. The Reach II model grid included 774 rows, 1,958 columns, and 2 layers, with 41,120 active cells in each layer. In Reach II, the top altitude of the model was raised by 0.4 percent to prevent cells from flooding and thus converting to confined-flow conditions. The model grids were oriented approximately parallel to the hydraulic gradient of the Canadian River alluvial aquifer. The groundwater-flow model boundaries were adjusted to ensure that each active cell was in connection with at least one other active cell.

There are few long-term water-table-altitude observation wells completed in the Canadian River alluvial aquifer; therefore, predevelopment steady-state conditions could not be determined, and the steady-state period for Reach I and Reach II was simulated as the average of model stresses for the period 1981–2013 and used a single 365-day stress period. The first stress period in both Reach I and Reach II was the steady-state period. The transient simulations for Reach I and Reach II simulated the period 1981–2013 and used 396 monthly stress periods, with 16 time steps per stress period to allow for temporal changes in the water table to be solved. The monthly stress periods implemented in the groundwater-flow models were used to capture seasonally variable processes, such as precipitation, water-table ET, and groundwater use.

Boundary Conditions

Boundary conditions for the Canadian River alluvial aquifer groundwater-flow models represent the locations where the inflow or outflow of water occurs (pl. 5). The selection of boundaries is important for development of an accurate groundwater-flow model (Franke and others, 1987). Boundary types are chosen that best represent hydrogeologic features on the basis of model objectives. Boundaries for the Canadian River alluvial aquifer were simulated by using two types of boundary conditions: specified-flux boundaries or head-dependent boundaries. For a specified-flux boundary, a user-specified discharge rate is set that allows for groundwater to move either in or out of the boundary. The no-flow boundaries established at the bottom of layer 2 and at some locations along the lateral boundary in both layer 1 and layer 2 where bedrock units do not yield water to the Canadian River alluvial aquifer are also specified-flux boundaries. A head-dependent boundary condition simulates flow on the basis of the difference between a user-specified altitude and the head in model cells. Unless otherwise specified, the boundary conditions described apply to both Reaches I and II.

Groundwater Recharge

A specified-flux boundary condition representing areally distributed recharge was applied by using the Recharge package (Harbaugh, 2005), in which recharge was applied to the highest active model cell. Recharge rates were obtained from the SWB code, in which the steady-state simulations used an average recharge value calculated for the period 1981–2013, and the transient simulations used recharge values calculated for each month. Time-varying adjustment of recharge during each stress period occurred during the calibration phase to obtain better matches between simulated and observed values.

Water-Table Evapotranspiration

A head-dependent boundary condition was applied through use of the Evapotranspiration package (Harbaugh, 2005). The Evapotranspiration package was used to simulate water-table ET processes that were not simulated by the SWB code, which only considers ET from vegetation on the basis of the soil moisture available. The maximum water-table ET rate for each stress period was set as the difference between the potential and actual ET computed from the SWB code. The water-table ET extinction depth, which is the depth at which water-table ET ceases to occur, was set equal to the SWB root-zone depth for each land-cover type. The water-table ET decreases linearly with decreases in the simulated water-table altitude, and no water-table ET occurs when the simulated water-table altitude is below the extinction depth. Water-table ET processes were limited to the alluvial deposits where the water table is most commonly near the land surface. The water-table ET rate was adjusted during groundwater-flow model calibration.

Streamflow

Stream base flow in the Canadian River alluvial aquifer was simulated by using the Streamflow-Routing package (SFR; Niswonger and Prudic, 2005) (plate 5), representing a head-dependent boundary condition. The SFR package routes flow between the aquifer and the stream by using Darcy's Law The flow between the aquifer and the stream is the product of the streambed conductance and the difference between the water-table altitude and the stream stage. The streambed conductance is the product of the hydraulic conductivity of the streambed sediments and the area of the stream channel, divided by the streambed thickness. Stream stage is calculated by using Manning's Equation, in which stream stage is a function of flow on the basis of the geometry of the stream channel. Where water-table altitudes are higher than stream stage, inflow to the stream occurs at a specified streambed conductance rate. Where water-table altitudes are lower than stream stage, streamflow discharge to the aquifer occurs.

A stream-water balance provides the amount of water available for exchange between the stream and the aquifer in groundwater-flow model simulations. Inflows include specified inflows (such as simulated inflows from outside the active area), tributary stream segments, and base-flow gain from the groundwater table. Outflows include water routed downstream, diversions, ET, and base-flow to the aquifer. Flows are calculated for each part of the stream contained in a model cell, known as reaches, until the end of a segment, or group of cells with uniform or linear hydraulic properties, is reached during each time step. Calculation of flows for the next downstream segment repeats this process by using flows from the upstream segment until water is routed out of the active area at a groundwater-flow model boundary.

The spatial location and extent of the river were derived from the National Hydrography Dataset (NHDPlus; McKay and others, 2012), a 1:100,000-scale geospatial dataset of surface-water features. The groundwater-flow model active area was not extended to include surface-water tributary channels, and inflow points were created to simulate monthly stream base-flow discharges from the perennial tributaries (Deer Creek, Buggy Creek, Walnut Creek, and Little River). Seven inflow points also were implemented that simulate smaller tributaries between the Norman streamgage (07229050) and the Purcell streamgage (07229200).

All other stream features were considered to be intermittent or ephemeral and were assumed to have no base flow. In Reach I, stream base flow from the Canadian River in Texas that crossed the border into Oklahoma was simulated with an inflow point. In Reach II, stream base flow for the Canadian River estimated at the Bridgeport streamgage (07228500) was simulated with an inflow point. Additionally, inflow points were created for five wastewater-treatment plants in Reach II that discharge to the Canadian River, and discharge data were used as inputs for each stress period.

The streambed hydraulic conductivity for Reach I and Reach II was set to 5 ft/d across all segments, representing a fine sand. This value increased the stability of the groundwater-flow models and is expected to reasonably represent the streambed hydraulic conductivity. Stream width was set to 150 ft in Reach I and 175 ft in Reach II on the basis of aerial photographs (U.S. Department of Agriculture, 2015) from 2015. The streambed altitude for each stream reach was set between 3 and 5 ft below land surface in each model cell. Differences in stream channel spatial location occurred between the DEM and the NHDPlus flowline because of the averaging of DEM cells and migration of the stream channel in the alluvial valley over time. To ensure correct surface-water routing in the stream, the stream channel location was adjusted to ensure that the altitude decreased in a downstream direction to avoid the uphill routing of water in the model.

Lateral Flow

Flow to and from bedrock aquifer units adjoining the Canadian River alluvial aquifer was represented by a head-dependent boundary condition by using the General Head Boundary (GHB) package (Harbaugh, 2005) (pl. 5). The GHB package uses a linear relation between flux and the water-table altitude (or water head), and a user-specified reference water-table altitude and conductance. Similar to the SFR package, conductance is defined as the product of hydraulic conductivity of the model cell and the cross-sectional area perpendicular to flow, divided by the distance between the general head condition and the model cell. When the simulated water-table altitude is lower than the reference water-table altitude, water flows into the groundwater-flow model at the specified conductance rate. When the simulated water table is higher than the reference water-table altitude, water is removed from the groundwater-flow model.

GHB cells were implemented along the boundary of the aquifer in layer 1 to simulate lateral inflow from bedrock. GHB cells were implemented in layer 2 to simulate lateral inflow from adjoining bedrock units. Conductance values were estimated on the basis of the local hydraulic properties of the aquifer for each GHB cell. Higher values were used for layer 1 when simulating the alluvial and terrace deposits, and lower values were used in layer 2 when simulating the bedrock. The reference water-table altitude for each GHB cell simulating flow from bedrock was determined from the nearest bedrock water-table measurement.

Flow to Eufaula Lake from the Canadian River alluvial aquifer was represented by a head-dependent boundary condition by using the Drain package (Harbaugh, 2005) (pl. 5). The function of these drain cells is similar to that of the GHB cells, except flow is only permitted in one direction.

Groundwater Use

Groundwater use representing a specified-flux boundary condition was simulated by using the Well package (Harbaugh, 2005). Annual groundwater-use rates reported by the OWRB were tabulated from 1981 to 2013, and monthly water demand, or the percent of annual water use per month, was obtained for the three planning regions encompassed by the Canadian River alluvial aquifer. The monthly water demand for Reach I was obtained from the West Central Watershed Planning Region Report (Oklahoma Water Resources Board, 2012b), whereas the monthly water demand for Reach II was obtained from the Central Watershed Planning Region Report (Oklahoma Water Resources Board, 2012c) and the Eufaula Watershed Planning Region Report (Oklahoma Water Resources Board, 2012d). This water demand quantifies the water use per month by each groundwater-well use type and watershed region. For the public-supply and irrigation well use types, annual groundwater-use rates were multiplied by the monthly public-supply water demand and the irrigation water demand, respectively. All other use types were multiplied by the monthly public-supply water demand. Groundwater use from self-supplied domestic wells was expected to be relatively small and was not simulated.

Groundwater-use rates for the steady-state simulation were determined from the average annual groundwater-use rates of 1981–2013, whereas transient simulation groundwater-use rates were simulated for each stress period based on the respective water-demand use type. Pumping wells and the respective observations that were located in cells with steep water-level gradients were moved to a nearby cell with a lower water-level gradient to lessen the reduction of pumping rates because of lack of saturation in the cell. All pumping wells were located in layer 1 of the groundwater-flow models.

Model Calibration

Model calibration is the process by which the initial values for model inputs are adjusted to improve the fit between observed and simulated data. During model calibration, the inputs to be adjusted are updated to new values that reduce the discrepancy, or residual, between the observed and simulated data. Calibration outcomes were evaluated on the basis of the reduction of this residual and the fit of the calibrated groundwater-flow model water-budget components to those of the conceptual flow model water budget.

The calibration process for the Reach I and Reach II groundwater-flow models included manual adjustment of model inputs, or parameters, by trial and error, followed by use of automated nonlinear regression techniques. Parameters were selected that represent important hydrologic processes, have a moderate uncertainty in their values, and have associated observations, or calibration targets, that are sensitive to adjustments in these parameters. Calibration targets included measurements of water-table altitude, estimated stream base flows, and base-flow gain to the Canadian River between streamgages. Manual parameter adjustment was conducted until minimal changes in the residuals occurred, and further adjustments did not yield improvements.

After the manual parameter adjustment, automated nonlinear regression techniques were performed by using parameter estimation (PEST) (Doherty, 2010), an open-source calibration and parameter uncertainty code. The PEST code uses the Gauss-Marquardt-Levenberg algorithm to adjust user-specified parameters to minimize the residual between the simulated data and the calibration targets (Doherty, 2010). PEST accomplishes this adjustment by running the model as many times as needed to determine the best parameter values (Doherty, 2010), while automatically updating the parameter values during each step of the process. This approach typically allows for the optimal values of a large number of parameters to be estimated together with greater speed than that of a traditional manual calibration approach.

During the PEST calibration, the calibration targets were weighted to account for error in each type of observation. Weights were assigned by using an inverse relation to the standard deviation (Doherty, 2010), and the weighted calibration targets were placed in observation groups on the basis of observation type. The sum of the squared weighted residuals represented the objective function, which measured the fit between observed and simulated data. Weighting for each observation group was then adjusted to balance the contribution from each observation group to the objective function. Table 6 shows the water-table altitude, stream base flow, and base-flow gain standard deviations, observation counts, and group contributions to the objective function.

The calibration approach implemented hybrid regularization (Tonkin and Doherty, 2005), also known as singular value decomposition-assist (SVDA; Doherty, 2010), to reduce the sustained run times associated with the calibration of a large number of parameters. SVDA computes a user-specified number of super parameters, or combinations of base parameters that are most responsive to the dataset (Doherty, 2010). The groundwater-flow model must be run one time for each adjustable parameter; thus, using fewer parameters greatly reduces the time required for calibration of the groundwater-flow models. Additionally, SVDA discards parameters with little to no sensitivity to the calibration targets, which produces a more stable calibration process.

Tikhonov regularization, whereby user-specified “soft” information pertaining to certain parameters is applied in the form of mathematical relations, was used to limit changes made to the calibrated values of certain parameters. This regularization is accomplished through use of a penalty applied to the objective function when the regularized parameters deviate from their original values (Doherty, 2010). Tikhonov regularization was applied to the horizontal hydraulic conductivity array multiplier for layer 1 of the groundwater-flow models to minimize discrepancies between the average horizontal hydraulic conductivities for Reaches I and II. Additionally, by limiting the automated calibration adjustments to these arrays, the changes applied by using PEST are constrained to more realistic values.

Calibration Parameters

Five groups of parameters, representing 311 parameters in Reach I and 483 parameters in Reach II, were used in the groundwater-flow model calibration. Those groups included parameters representing horizontal hydraulic conductivity, water-table ET rate, recharge rate, specific yield, and GHB conductance. The recharge group consisted of a time-varying recharge-rate parameter as a multiplier for recharge in each stress period for both reaches. The remaining parameter groups were set as fixed parameters that did not change

Table 6. Components of the objective function prior to calibration by using parameter estimation for Reach I and Reach II of the Canadian River alluvial aquifer, Oklahoma.

[USGS, U.S. Geological Survey; ft, feet; OWRB, Oklahoma Water Resources Board; ft^3/s, cubic feet per second; <, less than]

Observation group	Description	Standard deviation	Group contribution	Number of observations	Objective function components
Reach I of the Canadian River alluvial aquifer					
Water-table altitude	USGS water-table measurements	1.5 ft	24 percent	101	10,916
Water-table altitude	OWRB water-table measurements	1.2 ft	10 percent	82	4,621
Stream base flow	Bridgeport streamgage (07228500)	25.0 ft^3/s	56 percent	397	26,402
Base-flow gain	Reach I base-flow gain	16.1 ft^3/s	10 percent	1	4,218
			Total	**581**	**46,157**
Reach II of the Canadian River alluvial aquifer					
Water-table altitude	USGS water-table measurements	2.0 ft	25 percent	762	26,386
Water-table altitude	OWRB water-table measurements	4.0 ft	8 percent	244	8,779
Stream base flow	Norman streamgage (07229050)	34.4 ft^3/s	<1 percent	85	517
Stream base flow	Purcell streamgage (07229200)	55.3 ft^3/s	4 percent	336	4,102
Stream base flow	Calvin streamgage (07231500)	148.3 ft^3/s	55 percent	397	57,653
Base-flow gain	Bridgeport to Norman streamgages	23.7 ft^3/s	<1 percent	1	52
Base-flow gain	Norman to Purcell streamgages	32.1 ft^3/s	<1 percent	1	458
Base-flow gain	Purcell to Calvin streamgages	89.6 ft^3/s	8 percent	1	8,617
			Total	**1,827**	**106,564**

during the simulation. For areas where horizontal hydraulic conductivity was thought to be relatively uniform, horizontal hydraulic conductivity values were grouped into 25 parameter zones in Reach I, and 49 parameter zones in Reach II (pl. 6). The horizontal hydraulic conductivity was adjusted between 5 and 100 percent where mixed bedrock and alluvial deposits were thought to be present. This range in multipliers represents the uncertainty inherent in the lithologic-log reclassification and interpolation methods used to obtain horizontal hydraulic conductivity. Temporal recharge array multipliers were applied to the monthly recharge arrays. The water-table ET rates were adjusted by using an array multiplier until an optimal rate was determined and then applied uniformly to all temporal ET arrays for both reaches. A multiplier of up to 30 percent was applied uniformly to the specific-yield arrays in both reaches. GHB conductance was adjusted until the water table in the simulated terrace locations approached observed data.

Calibration Targets

Calibration targets included measurements of water-table altitude, estimated stream base flows, and base-flow gain to the Canadian River between streamgages, and were placed in an observation group of the same name (table 6). Water-table altitude observations were obtained from the National Water Information System (NWIS) database (U.S. Geological Survey, 2013a) and the OWRB database (Oklahoma Water Resources Board, 2015a). Screened intervals for observation wells were checked to ensure that the observations were made in the Canadian River alluvial aquifer. The land-surface altitude of the observations was compared to the altitude obtained from a DEM, and observation targets with large discrepancies were discarded. The water-table altitude observations in the alluvial and terrace deposits were spatially distributed, but only a few observations were available for each year of the model period. The Reach I groundwater-flow model used 64 observation wells. A total of 38 water-table altitude observations were used in the steady-state simulation, and 145 were used in the transient simulation. The Reach II groundwater-flow model used 363 observation wells. A total of 77 water-table altitude observations were used in the steady-state simulation, and 929 were used in the transient simulation.

Stream base flows at streamgages were used as primary calibration targets because of sparse water-table altitude observations in the simulated period. BFI-computed (estimated) stream base flows at the Bridgeport streamgage (07228500), the most downstream location in Reach I, were used as stream base-flow calibration targets in Reach I. BFI-computed stream base flows at the Norman streamgage (07229050), Purcell streamgage (07229200), and Calvin streamgage (07231500) were used as stream base-flow calibration targets in Reach II. Simulated stream base flows at streamgages were monitored by using the Gage package (Merritt and Konikow, 2000) in conjunction with the SFR package. For Reach I, 1 stream base-flow observation was used in the steady-state simulation, and 396 monthly stream base-flow observations (1 observation per stress period) were used in the transient simulation. For Reach II, 1 stream base-flow observation was used in the steady-state simulation at the Purcell streamgage (07229200), and Calvin streamgage (07231500), and 816 monthly stream base-flow observations were used in the transient simulation between all 3 streamgages.

Average annual base-flow gain to the Canadian River across the groundwater-flow model transient-simulation period was set as a calibration target (table 6). The simulated contribution was determined as the sum of the “Flow to Aquifer” column in the SFR flow file for the stream segments between each streamgage, divided by the 396 stress periods. In Reach I, the average annual base-flow gain to the Canadian River between the Texas border and the Bridgeport streamgage (07228500) was used as a single observation. In Reach II, each average annual base-flow gain to the Canadian River between successive streamgages was used as a single observation, for a total of three base-flow gain observations in Reach II.

Calibration-Target Uncertainty and Weighting

The water-table altitude, stream base flow, and base-flow gain calibration targets in each observation group were weighted by using error-based weighting (Hill and Tiedeman, 2007). Water-table weights were determined on the basis of three error components: (1) the location accuracy, or how precise the coordinates of the wells were known for observation wells; (2) the altitude accuracy, which is based on the method used to determine the land-surface altitude; and (3) the temporal variability of water levels at each observation well. Methods from Clark and Hart (2009) were used to determine the estimated uncertainty for the location accuracy. Stream base flow weights were determined on the basis of the accuracy of the streamflow data at each streamgage.

The location and altitude accuracy values recorded in the NWIS and OWRB databases are based on the method used to obtain each measurement, typically by using a Global Positioning System or a topographic map. The location accuracy of each observation well, as recorded in the NWIS and OWRB databases, was between 0.01 and 10 arc-seconds. A radius equal to the location accuracy was created for each observation well, and the standard deviation associated with the land-surface altitude in the location-accuracy radius was then calculated. The standard deviation of the altitude accuracy was calculated by dividing half of the altitude accuracy code (in feet) by the critical value of a 95-percent confidence interval (Hill and Tiedeman, 2007).

The temporal variability at each observation well was determined by using methods from Hill and Tiedeman (2007). The water-table range was determined by using either the seasonal water-table range from the continuous recorder wells or directly from the water-table altitude observations. Though the continuous recorder wells had measurement periods of less than 2 years, those measurements show the effects of partial drought conditions prevalent in 2014 and unusually high precipitation in May 2015 (fig. 11, fig. 12). The minimum, maximum, and average range in water-table altitude observations for the period 2014–15 at the continuous recorder wells were 0.9 ft, 11 ft, and 4.4 ft, respectively (fig. 11, fig. 12). The standard deviation for the temporal variability was calculated as the water-table range divided by four (Hill and Tiedeman, 2007).

By using methods from Hill and Tiedeman (2007), the standard deviations for the location accuracy, altitude accuracy, and temporal variability were converted to variances and summed for each observation well. The summed variance at each observation well was then converted to a standard deviation. The standard deviation of USGS water-table altitude observations was between 1.5 and 2.0 ft. The standard deviation for OWRB water-table altitude observations was between 1.2 and 4.0 ft (table 6).

Most of the field streamflow measurements collected during the transient period at each of the four streamgages were rated as "fair" in NWIS, which corresponds to a 95-percent confidence interval of ±15 percent of true streamflow (U.S. Geological Survey, 2013c). The standard deviation for the stream base-flow observations at each streamgage were then calculated as 15 percent of the stream base flow, divided by the critical value of a 95-percent confidence interval.

Weights for the base-flow gain were assigned by using methods from Hill and Tiedeman (2007) and were given a 95-percent confidence interval of ±15 percent of actual flow. The standard deviations and variance of the stream base-flow at each streamgage during the transient period were calculated, and the upstream and downstream streamgage variances were summed and converted to produce a standard deviation for the base-flow gain.

Prior to the automated calibration process using PEST, the observation weights for the calibration targets in each group were adjusted by using methods from Doherty and Hunt (2010). An observation group contribution to the objective function (table 6) was set through a series of PEST trial runs in which various combinations of observation-group weights were tested to determine the best weighting arrangement to reduce the objective function. Stream base-flow observations received the largest weighting in both Reach I and Reach II; at least one base-flow observation was available for each stress period. In Reach II, simulated stream base flows at the Norman streamgage (07229050) and Purcell streamgage (07229200) closely matched observed stream base flow; therefore, the group contribution for base-flow observations from those streamgages received minimal weighting (table 6).

Calibration Results

Calibration outcomes were evaluated on the basis of the reduction of the residual between the calibration targets and simulated values, as well as the fit of the calibrated groundwater-flow model water-budget components to those of the conceptual flow model water budget. Additionally, the calibrated parameters were evaluated for unrealistic parameter values from the PEST calibration.

Comparison of Simulated and Observed Values

The water-table altitude, stream base flow, and base-flow gain residuals were determined as the difference between observed and simulated values; therefore, residuals with positive values indicate lower simulated values than observed values, whereas negative values indicate higher simulated values than observed values. The root-mean-square error (RMSE) was calculated for the water-table altitude by using the following equation:

$$RMSE = \sqrt{\frac{1}{n}\sum_{i=1}^{n}\left(h_0 - h_s\right)^2} \qquad (6)$$

where

n is the number of water-table altitude observations,

h_o is the observed water-table altitude, in feet, and

h_s is the simulated water-table altitude, in feet.

Reach I

For Reach I, a good agreement between the observed and simulated water-table altitudes was obtained (fig. 18*A*, table 7). The combined Reach I water-table altitude RMSE for both steady-state and transient simulations was 6.1 ft, and 75 percent of residuals were within ±6.7 ft of observed measurements (table 7). The simulated water-table relief in Reach I was 826 ft, and maximum and minimum water-table altitudes were 2,215 ft and 1,389 ft, respectively. The RMSE as a percentage of this water-table relief was 0.7 percent for the steady-state simulation and 0.8 percent for the transient simulation (table 7), which relates the RMSE to water-table variability.

The spatial distribution of the residuals was not random; a slight bias occurred in the simulated water-table altitude below the observed water-table altitude (fig. 18*B*), particularly at altitudes near 1,700 ft, 1,850 ft, and 2,100 ft (fig. 18*A*). The underfit of the simulated water-table altitudes compared to the observed water-table altitudes occurred in the terrace where simulated water-table gradients between 50 and 120 ft/mi towards the Canadian River resulted in difficulties in maintaining a simulated saturated thickness. Most of the USGS observed water-table altitude data were from wells completed in the alluvial deposits and were above the 1:1 line, whereas many of the OWRB water-table altitudes were located in the terrace, and were below the 1:1 line (fig. 18*A*). The average simulated water-table altitude at OWRB observation well 9666 (fig. 17*A*) in the terrace was 3.3 ft below the average of the observed water-table altitudes. This observation well is located adjacent to multiple SFR stream cells, which may dampen fluctuations in the simulated water-table altitude.

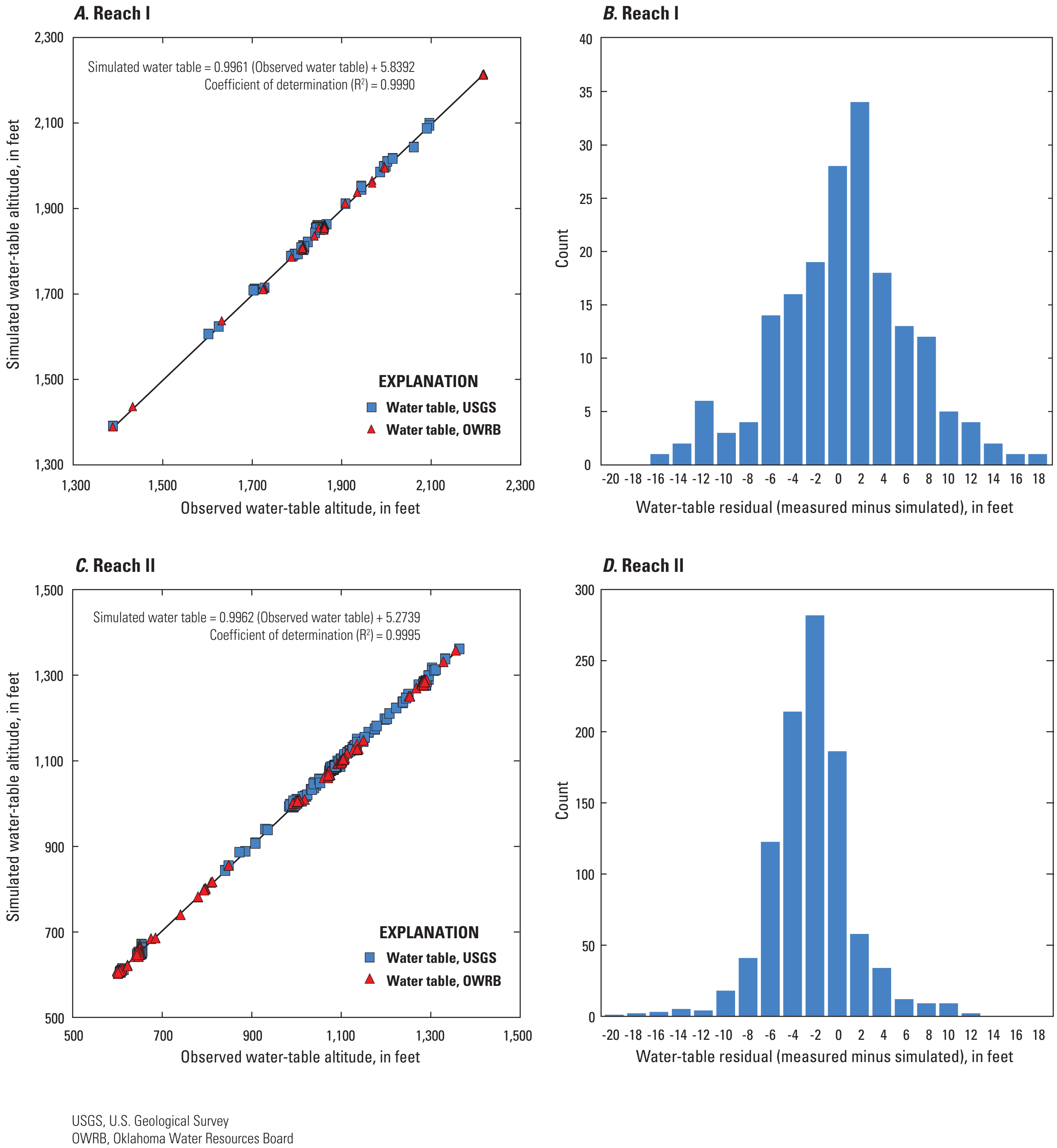


Figure 18. Observed and simulated water-table altitudes for *A*, Reach I, and *C*, Reach II, and histograms showing water-table residuals for *B*, Reach I, and *D*, Reach II, of the Canadian River alluvial aquifer, Oklahoma.

Table 7. Comparison of observed and simulated water-table altitudes for the Canadian River alluvial aquifer, Oklahoma, 1981–2013.

[Residual is calculated as the measured value minus the simulated value; thus, a minimum residual indicates that simulated values are smaller than observed values, and a maximum residual indicates that simulated values are larger than observed values. All units except percentage of head relief are in feet. RMSE, root-mean-square error; ±, plus or minus]

Statistic	Reach I			Reach II		
	Steady state	Transient	All	Steady state	Transient	All
Observation count	38	145	183	77	929	1,006
Maximum residual	-9.6	-15.2	-15.2	-14.4	-19.4	-19.4
Average residual	-0.04	1.9	1.5	-2.8	-1.2	-1.3
Minimum residual	18.0	18.8	18.8	8.6	13.6	13.6
Mean absolute residual	0.04	1.9	1.5	2.8	1.2	1.3
RMSE	5.6	6.2	6.1	5.3	3.9	4.0
75 percentile residual range	±5.2	±7.1	±6.7	±6.0	±4.0	±4.3
RMSE percentage of water-table relief	0.7	0.8	0.7	0.7	0.5	0.6

A total of 118 out of 145 observed water-table altitudes were measured between years 2001 and 2013; therefore, the temporal distribution of residuals was biased to this period. Fluctuations in the observed water-table altitude prior to this period were minimal, which simplified the water-table calibration, and resulted in a maximum water-table residual of less than 5 ft between 1980 and 2001. The largest residual of 18.8 ft (table 7) below the observed water table occurred in 2013 at an observation well with a single measurement, which may poorly represent longer-term conditions.

Observed (estimated) and simulated stream base flows at the Bridgeport streamgage (07228500) were comparable, though simulated average stream base flow was lower than the observed average stream base flow (fig. 19, table 8). This residual was due to stream base-flow differences of up to 466 ft^3/s that occurred between stress periods in the transient simulation, which also resulted in difficulties reproducing some stream base-flow peaks and valleys. Additionally, hydrograph separation techniques may not fully isolate base-flow conditions; therefore, it is possible that the peaks in the BFI-computed stream base flows represent a percentage of runoff. Moreover, bank storage effects (Chen, 2003), by which floodwater infiltrates the aquifer and is released days to weeks after floodwater recedes, may increase stream base flows during periods of no new recharge. The observed stream base-flow ranged from zero during summer months when water-table ET withdrawals were greatest to 648 ft^3/s during large precipitation events (table 4). The simulated average stream base flow in Reach I (table 8) was within the 95-percent confidence interval of the stream base-flow observations at the Bridgeport streamgage (07228500) (fig. 20). The variability in flow increases downstream at each successive streamgage because of the large increases in flows between streamgages. Streamflows at the Bridgeport streamgage (07228500) were typically the smallest and therefore had the narrowest confidence interval.

Simulated average base-flow gain from the border to the Bridgeport streamgage (07228500) was 8.8 ft^3/s (17 percent) lower than estimated average base-flow gain (fig. 20). This difference is largely due to the stream base-flow peaks that were not matched at each streamgage in the groundwater-flow model. The simulated average base-flow gain is within the 95-percent confidence interval of the estimated average base-flow gain (fig. 20).

Reach II

For Reach II, the observed and simulated water-table altitudes generally fit a 1:1 line (fig. 18*C*). The combined Reach II water-table altitude RMSE for both steady-state and transient simulations was 4.0 ft, and 75 percent of residuals were within ±4.3 ft of observed measurements (table 7). The water-table relief in Reach I was 759 ft, and maximum and minimum water-table altitudes were 1,362 ft and 602 ft, respectively. The RMSE as a percentage of this water-table relief was 0.7 percent for the steady-state simulation and 0.5 percent for the transient simulation (table 7), which indicates the relation between RMSE and water-table variability.

The residual distribution for Reach II was slightly above the 1:1 line, indicating a bias in the simulated water-table altitude above the observed water-table altitude (fig. 18*D*). Similar to Reach I, the large water-table gradients in the terrace—as large as 215 ft/mi towards the Canadian River in Reach II—resulted in lower simulated versus observed water-table altitudes compared to the alluvial deposits. This relation is shown on the 1:1 line for water-table altitudes below about 650 ft. However, this increased hydraulic gradient in Reach II compared to Reach I, combined with greater GHB inflow to the terrace, resulted in increased groundwater flow to the alluvial deposits. As a result, higher water-table altitudes occurred in the alluvial deposits, which contained 75 percent

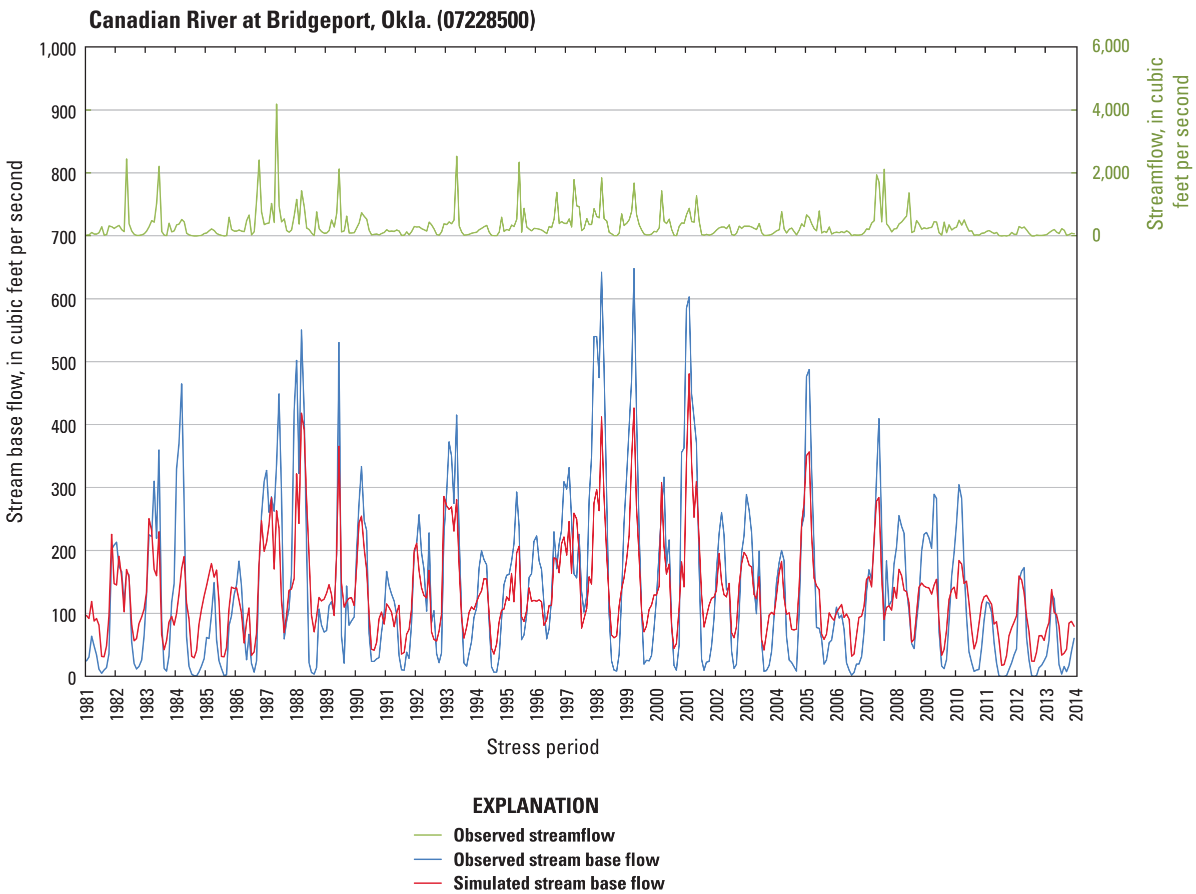


Figure 19. Observed streamflow, observed stream base flow, and simulated stream base flow at the Canadian River at Bridgeport, Oklahoma streamgage (07228500) for Reach I of the Canadian River alluvial aquifer, Oklahoma, 1981–2013.

Table 8. Comparison of observed and simulated average stream base flow for the Canadian River alluvial aquifer, Oklahoma, 1981–2013.

[Residual is calculated as the measured value minus the simulated value; thus, a minimum residual indicates that simulated values are smaller than observed values, and a maximum residual indicates that simulated values are larger than observed values. All units are in cubic feet per second. ±, plus or minus]

Statistic	Streamgage			
	Bridgeport (07228500)	Norman (07229050)	Purcell (07229200)	Calvin (07231500)
Average observed stream base flow	142	199	292	648
Average simulated stream base flow	133	164	238	517
Average residual	8.8	35	54	132
Minimum residual	293	460	781	3,729
Maximum residual	170	55	191	497
75 percentile residual range	±30	±52	±112	±235
Root-mean-square error	75	99	149	543

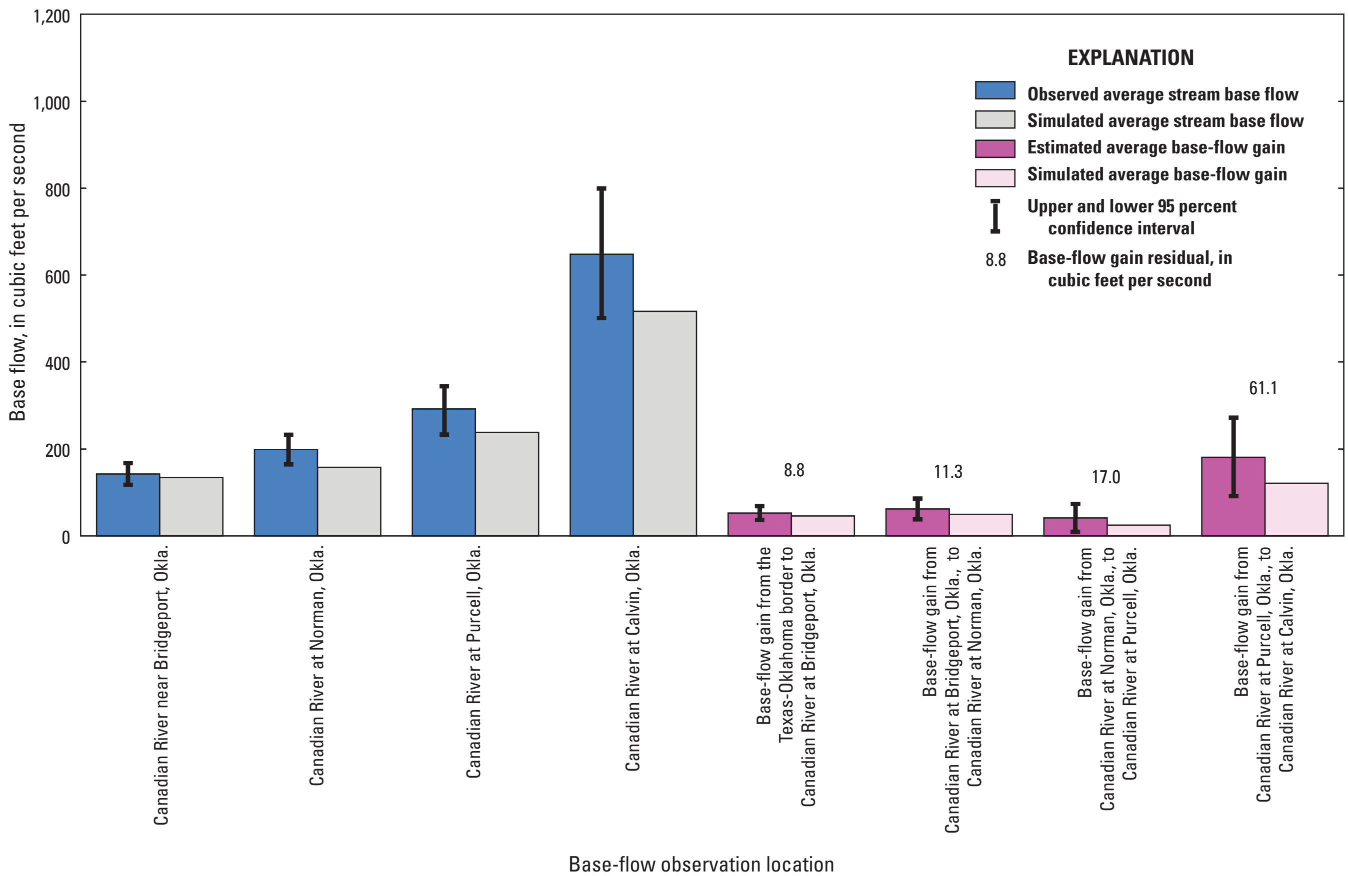


Figure 20. Observed and simulated average stream base flow, estimated average base-flow gain, simulated average base-flow gain, and confidence intervals for Reach I and Reach II of the Canadian River alluvial aquifer, Oklahoma, 1981–2013.

of observed water-table altitudes. Most of the USGS observed water-table altitudes were located in the alluvial deposits and were above the 1:1 line, while most of the OWRB observed water-table altitudes were located in the terrace, and were below the 1:1 line (fig. 18*C*).

The average water-table residual at OWRB observation well 9591 was 3.6 ft above the observed values (fig. 17*B*). This observation well was completed in alluvial deposits near several agricultural and irrigation wells which may account for some of the fluctuations in the observed water table. A trend of increasing recharge occurred from 1981 to 1991 for Reach II (fig. 16*B*) which resulted in the simulated water table reaching land surface in 1987 and 1992. The water-table altitude observations were generally reproduced by the simulation, though the magnitude of the water-table changes was not fully simulated, particularly in 2007 and 2011.

A total of 588 out of 929 observed transient water-table altitudes were measured between 1995 and 2000, although the magnitude of the residuals during this period are smaller than during the remaining model period. The largest residual of 19.4 ft above the observed water table (table 7) occurred in 1992, likely due to the generally increasing recharge between 1981 and 1991 (fig. 16*B*).

The simulated stream base flow in Reach II reproduced the majority of the observed stream base-flow trends at each of the three streamgages, though simulated average stream base flow was lower than the observed average stream base flow (fig. 21, table 8). The average simulated stream base-flow residual in Reach II was between 35 and 132 ft^3/s (table 8). The large fluctuations in stream base flow at the Calvin streamgage (07231500), ranging from nearly zero during summer months to as large as 4,941 ft^3/s (table 4) during precipitation events, were difficult to simulate and thus resulted in the largest residuals of base flow at this streamgage compared to the other streamgages (fig. 21, table 8). Additionally, this large stream base-flow variability resulted in a wide 95-percent confidence interval (fig. 20). As in Reach I, stream base-flow peaks may contain a percentage of runoff, which results in additional uncertainty in the observed stream base flows. Most of the simulated stream base flow at each Reach II streamgage (table 8) is within the 95-percent confidence interval of the stream base-flow observations (fig. 20). The stream base flow at the Norman streamgage (07229050) exceeded the 95-percent confidence interval by less than 1 ft^3/s.

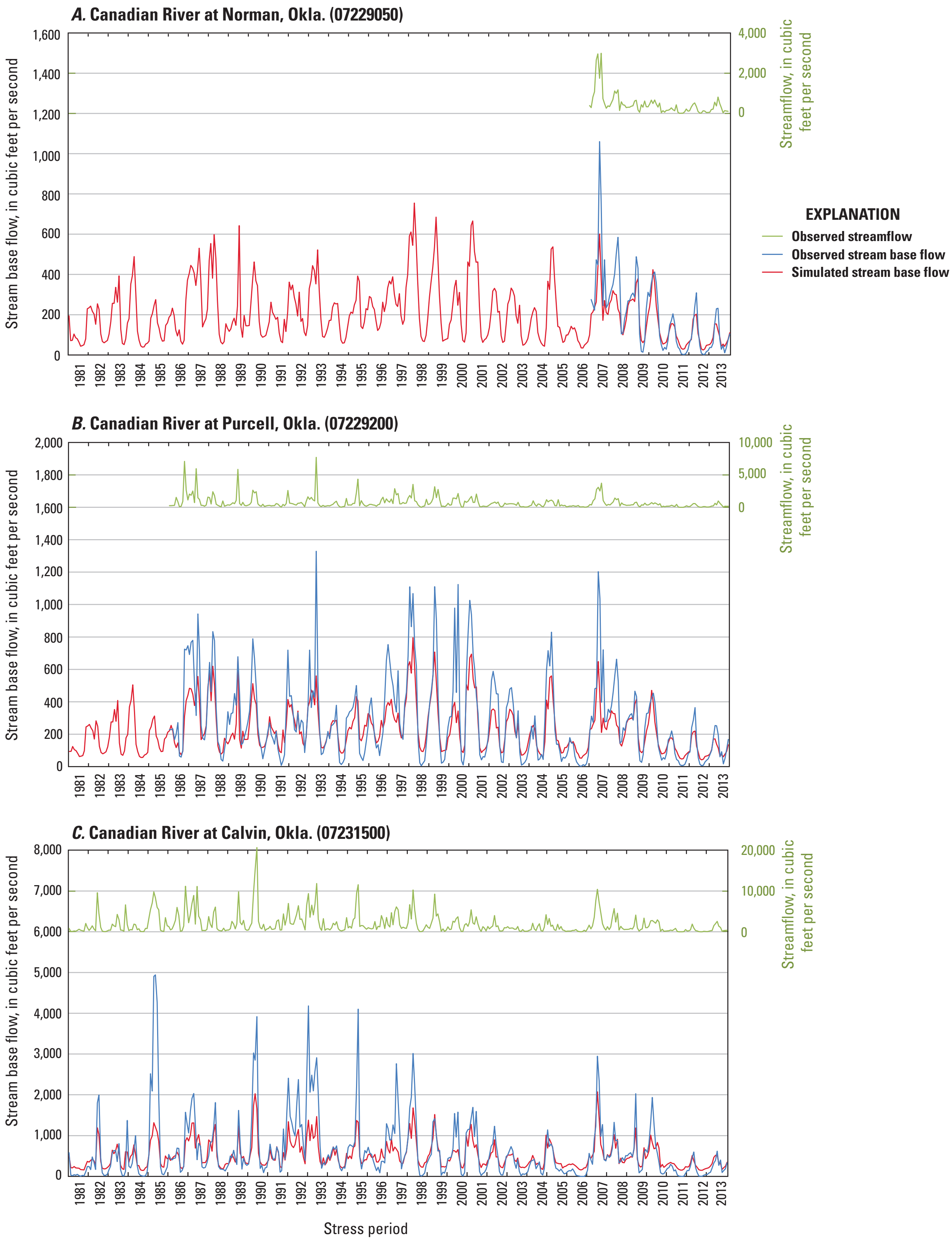


Figure 21. Observed streamflow, observed stream base flow, and simulated stream base flow at the Canadian River at *A*, Norman, Oklahoma (07229050); *B*, Canadian River at Purcell, Oklahoma (07229200); and *C*, Canadian River at Calvin, Oklahoma (07231500), streamgages for Reach II of the Canadian River alluvial aquifer, Oklahoma, 1981–2013.

The average simulated base-flow gain residual in Reach II was between 11.3 and 61.1 ft^3/s (fig. 20). Simulated average base-flow gain was lower than estimated average base-flow gain (fig. 20) because of the large stream base-flow peaks that were not matched at each streamgage in the groundwater-flow model. These peaks were substantially larger at the Calvin streamgage (07231500) (fig. 21) than at other streamgages, resulting in a wider 95-percent confidence interval for the base-flow gain at this streamgage. Although substantial tributaries were simulated in the groundwater-flow model, it is possible that other tributaries may contribute small amounts of continuous flow; thus, the lower simulated base-flow gain may be due in part to tributary flow that could not easily be quantified. The simulated average base-flow gain at each streamgage is within the 95-percent confidence interval of the estimated average base-flow gain (fig. 20).

Water Budget

Water budgets list average annual inflows to and outflows from the groundwater-flow models (table 9). Average annual recharge was the largest inflow in both models, which accounted for 56 percent of total inflows in Reach I and 49 percent of total inflows in Reach II. Average annual lateral inflow accounted for 26 percent of total inflows in Reach I and 41 percent of total inflows in Reach II. The increased percentage of lateral inflow in Reach II compared to Reach I may be due to the greater length of the aquifer, increased precipitation, or more incised alluvial deposits in eastern Oklahoma. Average annual flow to the Canadian River was the largest simulated outflow in both models, which accounted for 65 percent of total outflows in Reach I and 77 percent of total outflows in Reach II.

The net change in groundwater storage is shown in the simulated groundwater levels, which rise when water enters storage, or leaves the groundwater-flow model, and decline when water flows from storage, or flows into the groundwater-flow model. The changes in storage in both reaches are primarily caused by recharge and water-table ET. Major precipitation events such as Tropical Storm Erin in 2007 resulted in a large ratio of flows to storage compared to flows from storage, and a severe drought in 2003 resulted in a large ratio of flows from storage compared to flows to storage (tables 10–11). The changes in storage and water levels that result from these processes are shown in the changes in flows that rise after precipitation events and fall when water-table ET is greatest.

Table 9. Average annual water budget for the Reach I and Reach II numerical groundwater-flow models of the Canadian River alluvial aquifer, Oklahoma, 1981–2013.

[All units acre-feet per year. Under the net budget totals, base-flow gain was calculated as the difference between flow to and from the Canadian River. Lateral flow was calculated as the difference between lateral inflows and outflows]

Budget component	Reach I	Reach II	Reach I and Reach II combined
		Inflow	
Recharge	38,584	116,068	154,652
Flow from the Canadian River	12,694	25,299	37,993
Lateral inflow	17,561	96,323	113,884
Total inflows	**68,839**	**237,690**	**306,529**
		Outflow	
Evapotranspiration	22,372	45,416	67,788
Flow to the Canadian River	44,243	180,607	224,850
Groundwater use	542	8,640	9,182
Lateral outflow	1,235	793	2,028
Total outflows	**68,392**	**235,456**	**303,848**
		Net budget totals	
Base-flow gain	31,549	155,309	186,858
Lateral flow	16,326	95,530	111,856
Change in storage	447	2,234	2,681

Table 10. Numerical groundwater-flow model annual calibrated water budget for Reach I of the Canadian River alluvial aquifer, Oklahoma, 1981–2013.

[Annual fluxes in thousands of acre-feet per year. The average of each component, when converted to acre-feet per year, does not exactly match table 9 because of round-off errors after converting flow from model units. ET, water-table evapotranspiration; NA, not applicable]

Year	Inflow				Outflow					Total in	Total out
	From storage	Lateral flow	Recharge	Flow from the Canadian River	To storage	Wells	ET	Lateral flow	Flow to the Canadian River		
Steady state	NA	18.5	23.1	6.1	NA	0.6	9.7	0.9	36.5	47.7	47.7
1981	17.3	19.0	33.8	12.9	27.3	0.9	19.1	0.9	34.6	82.9	82.7
1982	26.0	17.6	31.4	11.9	20.1	0.9	21.7	1.0	43.0	86.9	86.8
1983	22.8	17.1	51.2	10.8	29.9	0.8	21.5	1.2	48.3	101.9	101.7
1984	23.5	18.2	28.0	13.8	19.6	1.3	23.0	1.0	38.4	83.5	83.2
1985	14.3	19.1	9.6	13.5	8.2	0.8	17.2	0.9	29.4	56.4	56.3
1986	12.7	18.7	52.7	13.3	40.8	0.6	18.1	1.0	36.7	97.3	97.2
1987	34.6	15.6	66.9	11.1	35.4	0.5	28.9	1.5	61.7	128.2	127.9
1988	39.1	15.6	57.6	14.6	37.7	0.8	29.7	1.8	56.8	127.0	126.7
1989	30.5	16.7	57.7	10.4	35.2	0.6	25.6	1.4	52.4	115.3	115.1
1990	26.8	17.2	28.2	12.1	12.7	0.7	22.5	1.4	46.7	84.2	84.0
1991	14.4	18.9	25.1	14.2	20.7	0.5	19.1	0.9	31.3	72.6	72.5
1992	19.3	18.0	57.1	12.4	42.5	0.4	20.0	1.1	42.6	106.7	106.6
1993	40.4	14.7	71.6	10.1	35.5	0.5	28.4	1.9	70.3	136.9	136.6
1994	24.9	17.9	15.0	12.4	7.1	0.4	22.2	1.0	39.3	70.2	70.0
1995	20.5	17.8	38.7	11.5	21.8	0.4	22.2	1.1	42.7	88.5	88.3
1996	13.5	18.6	28.1	13.4	16.2	0.4	19.6	0.9	36.3	73.6	73.4
1997	18.4	17.5	53.6	11.6	33.6	0.2	19.5	1.1	46.7	101.2	101.1
1998	47.9	14.8	77.1	11.7	51.9	0.7	26.6	1.9	70.1	151.5	151.3
1999	38.9	15.6	65.9	12.9	41.2	0.4	30.1	1.6	59.8	133.4	133.1
2000	36.4	15.9	67.1	11.9	45.6	0.6	25.5	1.6	57.8	131.4	131.1
2001	60.4	12.5	122.0	9.3	81.0	0.4	23.9	3.4	95.3	204.2	204.0
2002	30.1	17.0	17.8	12.1	9.7	0.3	22.5	1.2	43.0	76.9	76.8
2003	23.0	18.5	2.1	14.3	2.0	0.3	22.0	1.0	32.4	57.9	57.7
2004	18.9	17.9	43.0	12.8	29.4	0.3	20.5	1.0	41.3	92.6	92.5
2005	25.5	17.6	20.7	13.8	12.7	0.3	23.6	1.1	39.9	77.7	77.6
2006	14.5	19.6	2.3	16.1	2.9	0.6	22.1	0.8	25.8	52.5	52.3
2007	19.5	16.4	79.6	8.6	44.5	0.2	21.0	1.7	56.7	124.2	124.1
2008	22.1	17.2	20.7	12.3	8.8	0.2	21.2	1.4	40.6	72.4	72.3
2009	15.2	19.2	3.4	15.2	3.2	0.2	20.8	0.9	27.8	53.0	52.9
2010	13.1	19.4	9.7	14.2	7.4	0.2	19.1	0.8	28.8	56.5	56.4
2011	12.4	20.3	2.2	16.2	4.7	0.9	20.8	0.8	24.0	51.2	51.0
2012	16.2	19.5	25.7	14.1	20.0	0.8	21.1	0.8	32.7	75.6	75.4
2013	12.7	19.8	14.3	12.4	12.6	0.8	16.6	0.8	28.2	59.1	59.0
Average flux	**24.4**	**17.6**	**38.3**	**12.5**	**24.9**	**0.5**	**21.9**	**1.2**	**44.1**	**92.1**	**91.9**

Table 11. Numerical groundwater-flow model annual calibrated water budget for Reach II of the Canadian River alluvial aquifer, Oklahoma, 1981–2013.

[Annual fluxes in thousands of acre-feet per year. The average of each component, when converted to acre-feet per year, does not exactly match table 9 because of round-off errors after converting flow from model units. ET, water-table evapotranspiration; NA, not applicable]

Year	Inflow				Outflow						Total in	Total out
	From storage	Lateral flow	Recharge	Flow from the Canadian River	To storage	Wells	Drains	ET	Lateral flow	Flow to the Canadian River		
Steady state	NA	97.9	49.2	33.0	NA	9.5	0.1	30.6	0.4	139.3	180.1	180.1
1981	34.4	101.5	12.5	20.3	10.3	8.2	0.1	21.3	0.3	128.4	168.6	168.6
1982	44.6	98.8	88.9	22.9	63.8	8.3	0.1	32.0	0.5	150.6	255.2	255.3
1983	29.7	100.3	28.6	28.8	24.1	6.8	0.1	28.2	0.4	127.9	187.5	187.5
1984	31.7	101.2	23.3	22.4	19.2	7.1	0.1	23.7	0.4	128.2	178.6	178.7
1985	78.0	91.9	288.2	24.7	189.1	8.3	0.1	55.6	1.1	228.6	482.8	482.9
1986	50.9	96.8	137.0	24.2	96.0	6.6	0.1	33.5	0.7	172.1	308.9	308.8
1987	104.1	91.3	173.9	32.1	89.7	5.5	0.1	77.0	1.3	227.8	401.4	401.4
1988	64.6	98.5	27.8	27.2	14.3	8.1	0.1	40.8	0.6	154.2	218.0	218.0
1989	62.5	96.7	120.5	26.0	77.0	7.0	0.1	47.4	0.7	173.6	305.7	305.7
1990	96.4	90.0	294.8	24.0	158.5	7.4	0.1	72.3	1.6	265.3	505.2	505.2
1991	89.9	92.0	308.7	22.7	198.3	6.4	0.1	54.0	1.3	253.2	513.2	513.3
1992	106.4	90.2	178.9	24.4	64.1	4.8	0.1	73.8	1.5	255.8	399.9	400.1
1993	102.9	92.8	120.4	24.9	50.7	7.1	0.1	63.7	1.3	218.1	341.0	341.0
1994	69.8	96.8	66.7	21.6	25.0	6.3	0.1	45.9	0.9	176.6	254.9	254.7
1995	68.7	93.2	202.1	25.1	101.5	6.9	0.1	65.9	1.0	213.6	389.0	389.1
1996	68.7	96.0	107.8	26.5	68.8	6.8	0.1	40.5	0.8	181.9	299.0	299.0
1997	60.8	95.4	101.4	26.3	45.6	6.2	0.1	48.8	0.8	182.5	283.9	284.0
1998	96.3	92.9	144.4	27.8	84.7	8.9	0.1	57.8	1.1	208.9	361.5	361.6
1999	61.1	96.9	76.9	30.1	47.4	7.5	0.1	44.0	0.7	165.4	265.0	265.1
2000	54.2	95.1	210.6	23.7	122.5	9.3	0.1	45.3	0.9	205.8	383.7	383.8
2001	87.8	95.2	54.8	27.2	19.8	9.1	0.1	53.1	0.9	182.0	264.9	265.0
2002	60.1	95.8	142.7	23.7	82.1	9.3	0.1	44.6	0.8	185.6	322.5	322.5
2003	63.3	99.0	17.4	24.3	11.2	9.0	0.1	34.6	0.6	148.2	203.9	203.7
2004	36.0	98.5	136.1	25.6	106.7	8.1	0.1	24.7	0.6	155.9	296.1	296.2
2005	73.9	98.2	7.3	26.7	6.6	8.1	0.1	41.4	0.6	149.1	206.0	206.0
2006	36.5	102.2	9.4	22.0	9.8	9.8	0.1	25.9	0.4	124.1	170.0	170.1
2007	69.8	89.9	343.9	27.7	200.5	9.3	0.1	74.5	1.2	245.4	531.2	531.1
2008	80.0	95.9	75.9	27.8	30.5	11.8	0.1	57.2	0.8	179.2	279.6	279.6
2009	69.3	96.0	118.9	25.7	63.9	11.7	0.1	47.0	0.7	186.3	309.8	309.8
2010	83.6	95.1	131.1	26.4	77.5	11.3	0.1	49.0	0.9	197.5	336.2	336.3
2011	50.6	101.6	8.6	23.1	10.6	14.5	0.1	26.3	0.5	131.9	183.9	183.9
2012	40.0	102.2	7.6	24.5	12.6	16.4	0.1	23.3	0.4	121.6	174.3	174.4
2013	27.8	100.4	79.2	24.6	60.0	11.5	0.1	22.3	0.4	137.9	232.1	232.1
Average flux	**65.3**	**96.4**	**114.6**	**25.5**	**68.0**	**8.6**	**0.1**	**44.9**	**0.8**	**179.5**	**299.8**	**299.8**

Monthly inflows, including recharge and lateral flows, tended to be greatest between December and March in both reaches (fig. 22). Outflows, including flow to streams, water-table ET, and groundwater use, were greatest between March and August for both reaches (fig. 22). During this time, water-table ET was the largest outflow component and sometimes greatly exceeded the sum of recharge and lateral flow. Groundwater use, most of which is done by irrigation wells, peaked between June and September. The simulated groundwater use was 9 percent lower in Reach I and 8 percent lower than the specified input pumping rate in Reach II because of a lack of saturation in some areas of the terrace due to steep water-level gradients. Although the specified input pumping rates were checked for unrealistic values, it could not easily be determined if screened intervals extended into the bedrock units below the Canadian River alluvial aquifer for some wells with large pumping rates. Thus, the entire volume of groundwater pumped from each well occurred in this aquifer, resulting in a lack of saturation in some well locations regardless of any combination of parameter changes.

The temporal offset between maximum monthly inflows and outflows results in large storage changes that occur primarily in January–March, when inflows are greater than outflows, and July–October, when outflows are greater than inflows (fig. 22). Those large changes in groundwater storage limited the calibration that could be applied without causing numerical instability, despite using the Newton solver and 16 time steps per stress period. As a result, one monthly stress period exceeded a mass-balance error of 1 percent in Reach I. The maximum monthly stress period mass-balance error was 0.4 percent in Reach II.

Calibrated Parameters

The calibrated recharge was 33 percent greater in Reach I and 42 percent greater in Reach II than the recharge estimated in the conceptual flow model. The increase in calibrated recharge, particularly in January–April, and decrease in recharge during June–October (fig. 23) was primarily used to match the stream base-flow observations at each streamgage. The increased recharge also improved the simulated water-table observation residuals in the terrace, which were typically below the observed values. The average Reach I recharge temporal array multiplier was 1.05 (5 percent increase), and the average Reach II recharge temporal array multiplier was 1.22 (22 percent increase) (fig. 23).

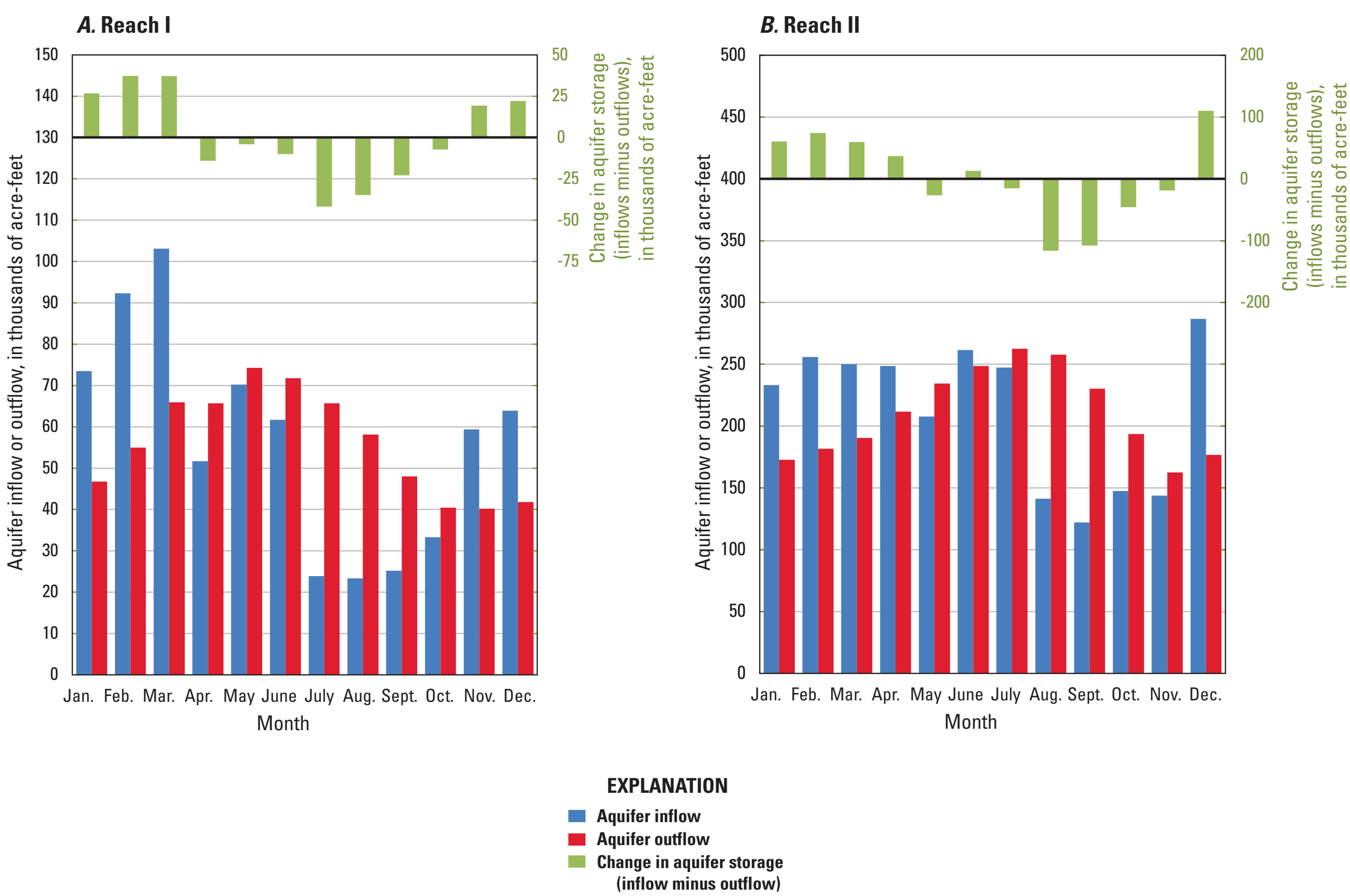


Figure 22. Average monthly aquifer inflow, outflow, and change in storage for *A*, Reach I, and *B*, Reach II, of the Canadian River alluvial aquifer, Oklahoma, 1981–2013.

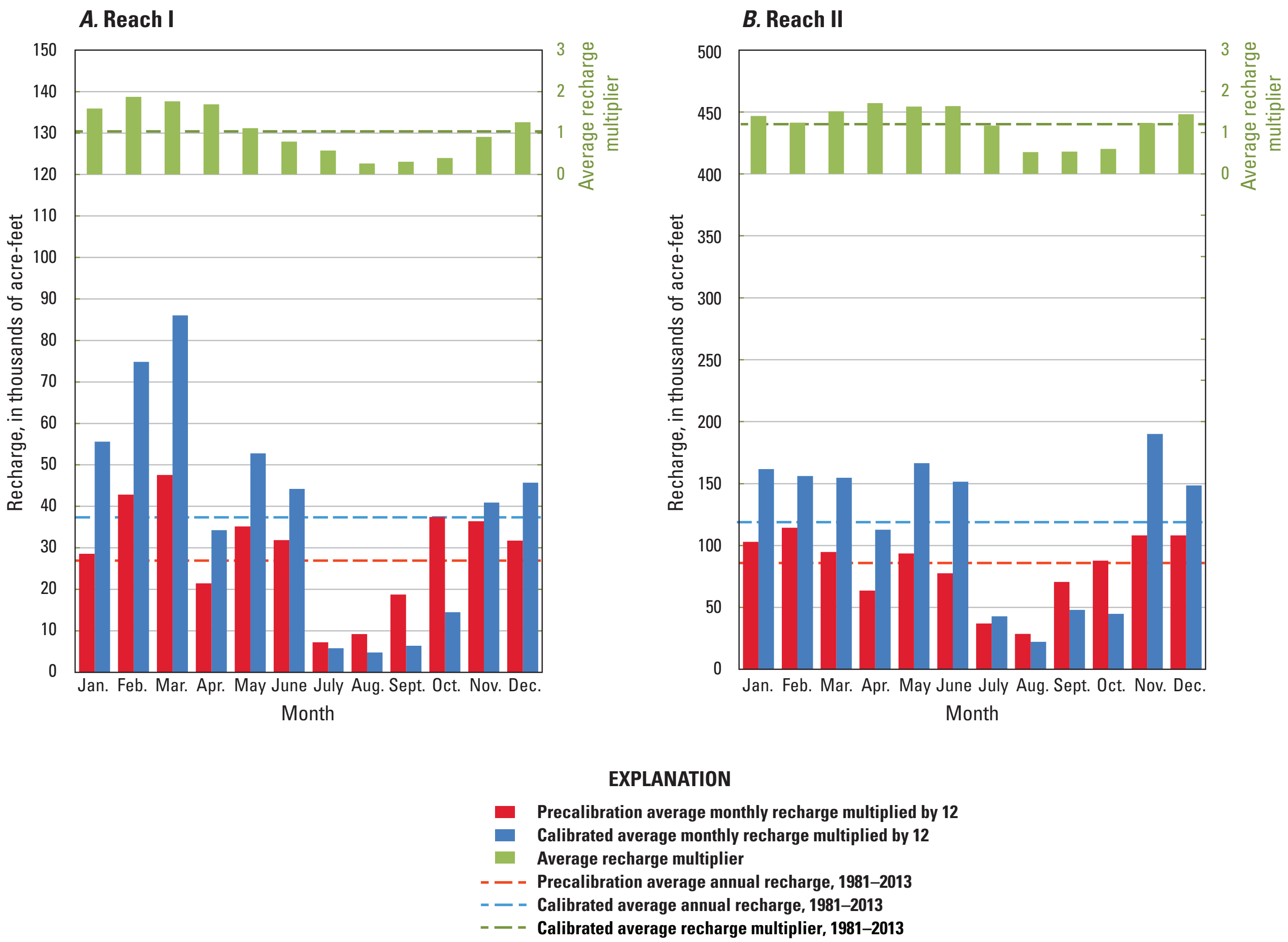


Figure 23. Precalibration recharge, calibrated recharge, and average recharge multipliers for *A*, Reach I, and *B*, Reach II, of the Canadian River alluvial aquifer, Oklahoma, 1981–2013.

In Reach I and Reach II, the calibrated water-table ET rate was increased by 60 percent in each stress period to match low base-flow (sometimes zero base-flow) conditions during summer months. Water-table ET in Reach I was 34 percent lower than the conceptual model, and water-table ET in Reach II was 42 percent greater. This change is due to the variability in the simulated water table, whereas the conceptual model ET withdrawals are based on a single site with a static depth to water.

The calibrated net lateral flow was 63 percent lower in Reach I and 42 percent lower in Reach II than the net lateral flow estimated in the conceptual flow model (table 5). In Reach I, average and maximum calibrated GHB conductances of 486 square feet per day (ft^2/d) and 1,320 ft^2/d, respectively, were used to simulate lateral inflow from bedrock. In Reach II, average and maximum calibrated GHB conductances of 260 ft^2/d and 924 ft^2/d, respectively, were used to simulate lateral inflow from bedrock. Increases in the lateral flow beyond the calibrated values resulted in smaller residuals for the base-flow gain. However, these increases also resulted in larger stream base-flow residuals during summer months, when the simulated base flow could not match the lower estimated base flow. As a result, no improvement in the model calibration occurred due to the greater group weight assigned to the stream base flow versus the base-flow gain (table 6). Increases in the water-table ET rate, combined with increases in the lateral flow, may have improved the model calibration; however, no increases in the water-table ET rate were possible, as they lead to model instability.

In Reach I, calibrated horizontal hydraulic conductivity ranged from 0.1 to 176 ft/d (pl. 6, fig. 24), with an average of 45 ft/d. A total of 95 percent of the calibrated horizontal hydraulic conductivity values in Reach I ranged from 0.1 to 93 ft/d; thus, the multiplier range resulted in the majority of these conductivities remaining near the originally defined range of 0.1–90 ft/d. In Reach II, calibrated horizontal hydraulic conductivity ranged from 0.15 to 158 ft/d (pl. 6, fig. 24), with an average of 75 ft/d. A total of 95 percent of the calibrated

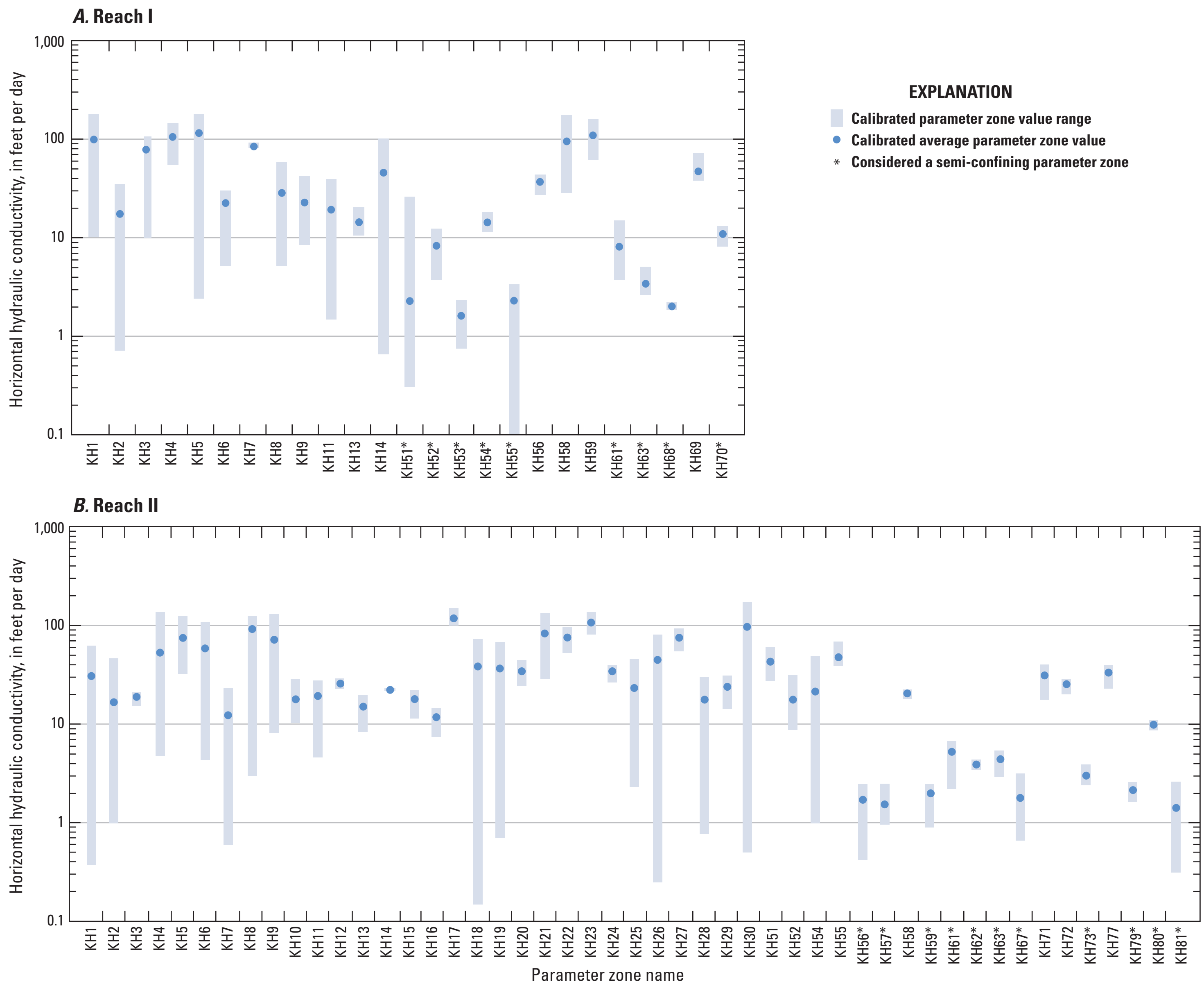


Figure 24. Calibrated average parameter zone values and ranges for *A*, Reach I, and *B*, Reach II of the Canadian River alluvial aquifer, 1981–2013.

horizontal hydraulic conductivity values in Reach II ranged from 0.1 to 115 ft/d. The increased horizontal hydraulic conductivity in Reach II compared to Reach I resulted in a lower water table, which improved the water-table observation residuals.

Specific yield was adjusted by an array multiplier to a final average value of 16 percent for Reach I and 15.2 percent for Reach II from an initial value of 22 percent for both reaches. Because specific yields from laboratory analyses are often larger than the specific yields obtained from well pumping tests in the field, laboratory-derived specific yields may be more useful for evaluating groundwater reserves over a long period of time versus the evaluation of more short-term groundwater-level responses to well pumping (Neuman, 1987). The determination of the aquifer response to well pumping and determination of a sustainable yield were investigated by this study; therefore, the average of 22 percent determined from previously published literature may overstate the specific yield of the Canadian River alluvial aquifer, thus the calibrated specific yield may provide a better estimation of specific yield for this groundwater study.

Sensitivity Analysis

A sensitivity analysis was performed by using the PEST sensitivity process (Doherty, 2010) to ensure that the parameters used during the calibration process were effective in reducing the objective function and the groundwater-flow model error. During calibration, PEST records the sensitivity of each calibration target to changes in parameters. These sensitivities are a measure of the change in residuals affected by adjustments to a parameter; thus, calibration target residuals are more easily reduced by larger sensitivities. Sensitivities were calculated by using the Jacobian matrix output from PEST and summed for each parameter group, listed in figure 25.

In both reaches, all three observation groups were most sensitive to changes in recharge, hydraulic conductivity, and GHB conductance. Spatially-distributed recharge to the water table was the largest inflow (table 9), which, when combined with the high hydraulic conductivity of the sandy alluvial and terrace deposits, directly affected the water table and stream base flows. In Reach II, the greater percentage of water-table observations located in the alluvial deposits versus the terrace may contribute to an increased water-table group sensitivity to recharge in Reach II compared to Reach I.

The observation groups were sensitive to changes in the GHB conductance because of the moderate to steep hydraulic gradient from the terrace to the alluvial deposits that resulted in a rapid decline of the water table in terrace areas without continuous lateral inflow. Additionally, lateral inflow provided flow downgradient from the alluvial and terrace deposits to the Canadian River, particularly in areas where the river channel was near areas of lateral inflow.

The evapotranspiration area of both reaches (pl. 5) included nearly all segments of the Canadian River alluvial deposits; therefore, the base-flow observation group was more sensitive to changes in evapotranspiration compared to the water-table observation group, which had observations upgradient in the terrace. The water-table observation group in Reach II was more sensitive to changes in evapotranspiration due to the greater number of observations in Reach II than in Reach I.

The observation groups were least sensitive to changes in specific yield. Decreases in specific yield magnified changes in stream base flow and the water-table altitude, particularly for stream base flow in Reach I, and for the water table in Reach II.

In both reaches, the smallest sensitivities were greater than 1 percent of the largest sensitivities, thus reducing the likelihood of convergence issues during the nonlinear regression using PEST (Hill and Tiedeman, 2007). The reduction in small sensitivities was accomplished by the removal of parameters that had no sensitivity.

Equal Proportionate Share Scenarios

The EPS for the alluvial and terrace aquifers is defined by the OWRB as the amount of fresh water that each landowner is allowed per year per acre of owned land to maintain a saturated thickness of at least 5 ft in at least 50 percent of the total overlying land of the groundwater basin for a minimum of 20 years (82 OK Stat § 82-1020.5). For the Canadian River alluvial aquifer, the EPS was computed for time periods of 20, 40, and 50 years into the future.

A hypothetical well was placed in each active cell in layer 1 of the calibrated groundwater-flow models, with all wells set to the same pumping rate for the duration of the EPS scenario. At the conclusion of each scenario, the number of cells with at least 5 ft of saturated thickness was compared to the number of active cells. If more than 50 percent of the cells had a saturated thickness of 5 ft or more, the pumping rate for each cell was increased by 5 cubic feet per day and the scenario was repeated until 50 percent of the cells had a saturated thickness of 5 ft or less. To provide a range of EPS pumping rates, the process was repeated with recharge increased and decreased by 10 percent. Additional 20-, 40-, and 50-year EPS scenarios were run without the GHB package simulating lateral flow. These scenarios simulated the effects of substantial reductions in the water-table altitude of the surrounding bedrock units on the EPS pumping rate for the Canadian River alluvial aquifer.

EPS estimates were provided for two subregions in Reach II, noted as Reach IIa and Reach IIb. Reach IIa includes the Canadian River alluvial aquifer extent between the Bridgeport streamgage (07228500) and the eastern border of Cleveland County (pl. 1). Reach IIb includes the remaining aquifer extent between the eastern border of Cleveland County and Eufaula Lake (pl. 1). For Reach IIa and Reach IIb, the EPS was calculated on the basis of the number of active cells in each subregion. The Reach I and Reach II groundwater-flow models are not coupled, and streamflow entering Reach II is the stream base flow observed at the Bridgeport streamgage

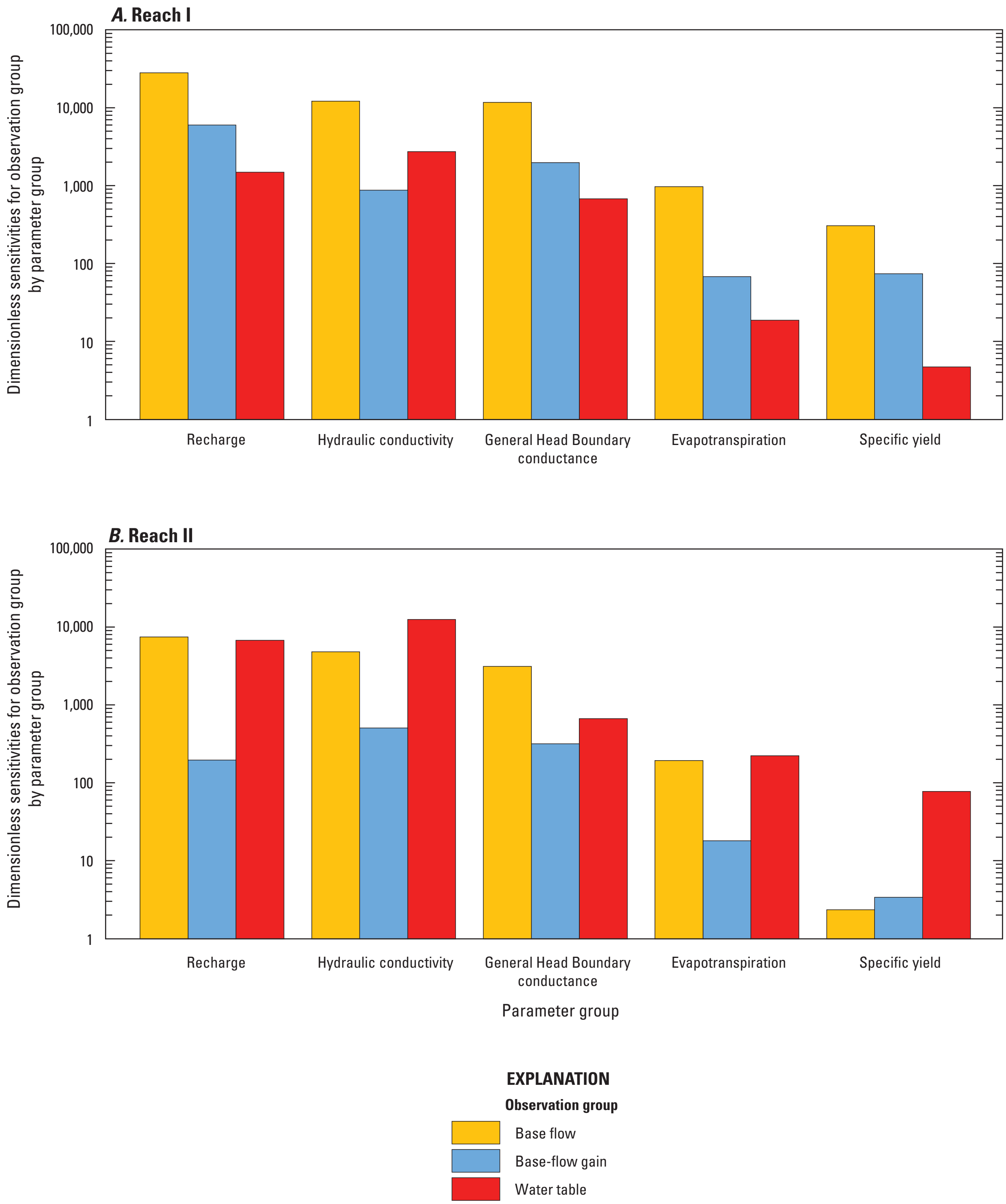


Figure 25. Observation group sensitivity by parameter group in the numerical groundwater-flow models for *A*, Reach I, and *B*, Reach II, of the Canadian River alluvial aquifer, Oklahoma.

(07228500); therefore, if stream base flow in Reach I at the Bridgeport streamgage (07228500) is decreased, a corresponding decrease in the Reach II EPS would be expected.

The EPS scenarios were configured to step backward through the transient simulation, starting with 2013, and the monthly stress periods were configured to step forward in each year. The EPS scenarios were stepped backward until 1981 and then stepped forward through the remaining scenario period; therefore, the 20-year EPS scenario included years 2013–1994, the 40-year EPS scenario included years 2013–1981 and 1982–88, and the 50-year EPS scenario included years 2013–1981 and 1982–98. The 2013 simulated water table was used as the starting water table in each EPS scenario. Model stresses for recharge, water-table ET, and tributary inflows were configured as the average of each stress period used in the calibrated groundwater-flow models. In some areas of the terrace, water levels are below the base of the Canadian River alluvial aquifer or are unsaturated through the calibrated transient simulation. As a result, only saturated cells were included in the total cell count for the EPS scenarios. The EPS rate was based on 5.74 acres in each 500-by-500-ft model cell.

During the EPS scenarios, the drawdown caused by the pumping wells placed in each cell would be expected to affect the water table in nearby areas. Lateral inflow of water from GHB cells reduced rebound effects at the boundary from pumping-well propagated stresses where the radius of influence may exceed the width of the active area of the model. Because of the linear relation between flux and water-table altitudes in the GHB package, continually increasing lateral inflows could result in an overestimated EPS pumping rate; however, the GHB conductance was limited in the calibrated models such that lateral inflows were not large enough to prevent dewatering from the pumping wells located in most areas of the alluvial and terrace deposits.

Substantial spatial variability in the aquifer saturation was present at the end of the EPS scenarios (pl. 7). During each scenario, water-table declines from groundwater pumping resulted in outflows from the Canadian River to the alluvial deposits in most of the aquifer. The majority of the terrace material upland of the alluvial deposits was unsaturated except for areas where a shallow hydraulic gradient occurred. For the EPS scenarios without lateral flow, this dewatering of the terrace was more rapid, which resulted in a larger reduction in groundwater flow from the terrace to the alluvial deposits. Flow in the Canadian River for the EPS scenarios was then sustained mostly from the inflows from the Canadian River in Texas. As a result, the western half of the Reach I alluvial deposits near the Canadian River remained saturated, whereas most other alluvial deposits had minimal or no saturation at the end of the EPS scenarios. The part of the alluvial deposits adjacent to the Rush Springs Formation also remained mostly saturated because of bedrock lateral inflow coupled with the narrow areal extent of the alluvial deposits in this area. Water availability under the maximum EPS pumping rates was primarily from the alluvial deposits.

The EPS pumping rates for the 40- and 50-year periods were similar for each scenario type (table 12). In the first 15 simulated years of each scenario, substantial amounts of water were removed from storage by the wells in each model cell in Reach I (fig. 27*A*). During this time, total groundwater pumping decreased as the thinner alluvial and terrace areas were dewatered and dropped below 5 ft of saturated thickness or went dry. Storage changes decreased in tandem with groundwater pumping, and approximate steady-state conditions were reached in as little as 15 simulated years; thus, storage changes were then only caused by the averaged stresses in each stress period. These stresses were averaged across each month (e.g. all model period stresses for January), and thus reflect seasonal stresses, which are shown by the oscillations in storage (fig. 27*A*).

Estimated Reach I Equal Proportionate Share

For Reach I, the 20-year EPS pumping rate was 1.35 (acre-ft/acre)/yr (fig. 26*A*, table 12). Decreasing recharge by 10 percent resulted in a 20-year EPS pumping rate of 1.32 (acre-ft/acre)/yr, and increasing recharge by 10 percent resulted in a 20-year EPS pumping rate of 1.38 (acre-ft/acre)/yr. For the 40-year EPS scenario, the 40-year EPS pumping rate was 1.34 (acre-ft/acre)/yr (table 12). Decreasing recharge by 10 percent resulted in a 40-year EPS pumping rate of 1.31 (acre-ft/acre)/yr, and increasing recharge by 10 percent resulted in a 40-year EPS pumping rate of 1.37 (acre-ft/acre)/yr. The results for the 50-year EPS scenario were the same as the 40-year EPS scenario. For the 20-year scenario without lateral flow, the EPS rate was 0.94 (acre-ft/acre)/yr. For the 40- and 50-year scenarios without lateral flow, the EPS rate was 0.91 (acre-ft/acre)/yr.

Estimated Reach II Equal Proportionate Share

For Reach II, the 20-year EPS pumping rate was 3.08 (acre-ft/acre)/yr (fig. 26*B*, table 12). Decreasing recharge by 10 percent resulted in a 20-year EPS pumping rate of 3.04 (acre-ft/acre)/yr, and increasing recharge by 10 percent resulted in a 20-year EPS pumping rate of 3.13 (acre-ft/acre)/yr (fig. 26*B*, table 12). The results for the 40- and 50-year EPS scenarios were the same as the 20-year EPS scenario (table 12).

For Reach IIa, the 20-year EPS pumping rate was 2.75 (acre-ft/acre)/yr (fig. 28*A*). Decreasing recharge by 10 percent resulted in a 20-year EPS pumping rate of 2.71 (acre-ft/acre)/yr, and increasing recharge by 10 percent resulted in a 20-year EPS pumping rate of 2.79 (acre-ft/acre)/yr (fig. 28*A*). For the 20-, 40-, and 50-year scenarios without lateral flow, the EPS rate was 2.28 (acre-ft/acre)/yr (table 12).

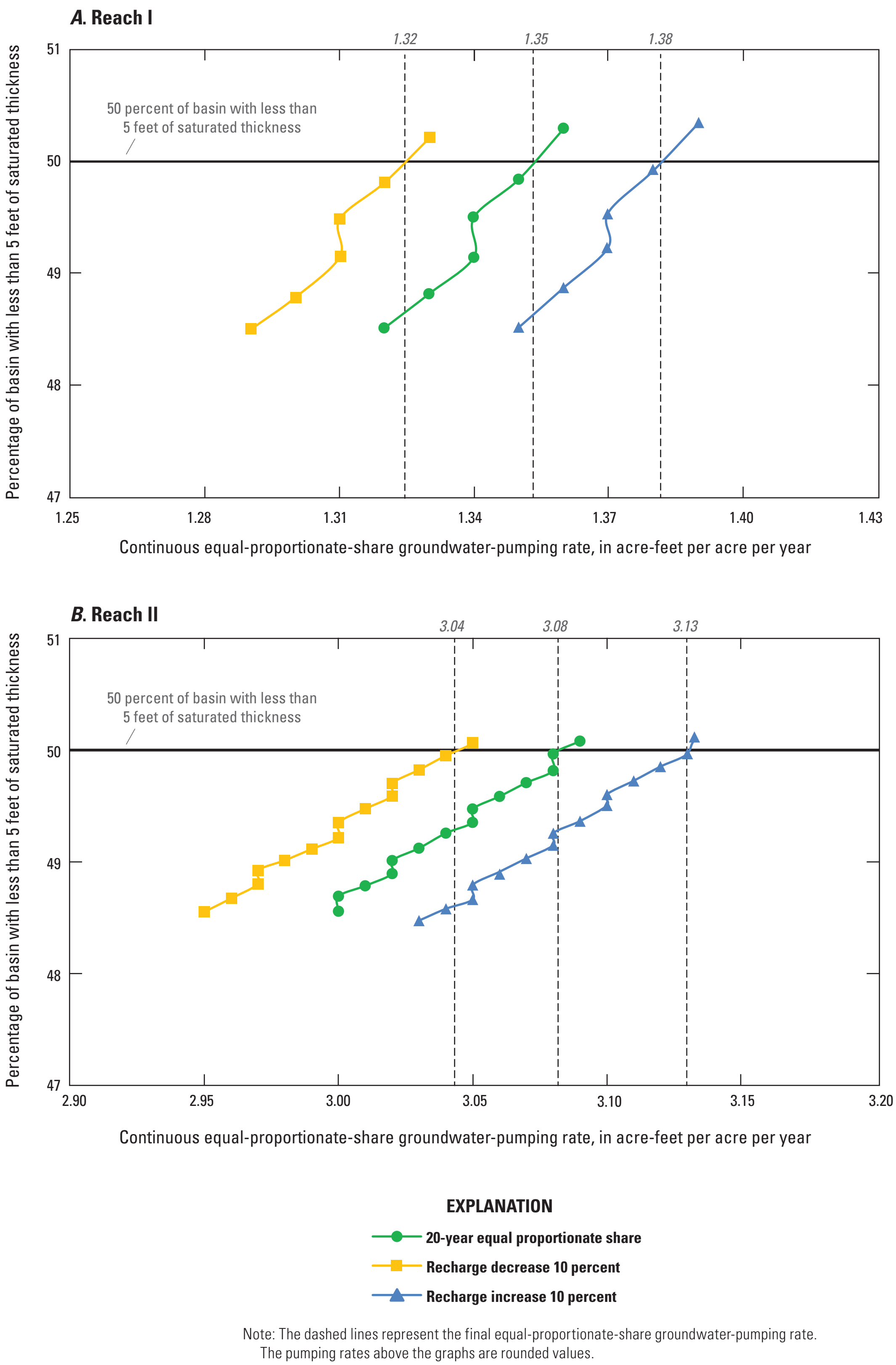


Figure 26. Percentage of *A*, Reach I, and *B*, Reach II, of the Canadian River alluvial aquifer, Oklahoma, with less than 5 feet of saturated thickness after 20 years of continuous equal-proportionate-share groundwater pumping.

Table 12. Equal-proportionate-share pumping for Reach I and Reach II of the Canadian River alluvial aquifer, Oklahoma.

[All units of pumping are in acre-feet per acre per year. no lateral flow, describes the results of the equal-proportionate-share scenario with no change in recharge and lateral flow removed]

Period in years	Reach I				Reach II			Reach IIa	Reach IIb
	Recharge reduced 10 percent	No change in recharge	Recharge increased 10 percent	No lateral flow	Recharge reduced 10 percent	No change in recharge	Recharge increased 10 percent	No lateral flow	No lateral flow
20	1.32	1.35	1.38	0.94	3.04	3.08	3.13	2.28	1.93
40	1.31	1.34	1.37	0.91	3.04	3.08	3.13	2.28	1.93
50	1.31	1.34	1.37	0.91	3.04	3.08	3.13	2.28	1.93

For Reach IIb, the 20-year EPS pumping rate was 4.30 (acre-ft/acre)/yr (fig. 28*B*). Decreasing recharge by 10 percent resulted in a 20-year EPS pumping rate of 4.23 (acre-ft/acre)/yr, and increasing recharge by 10 percent resulted in a 20-year EPS pumping rate of 4.36 (acre-ft/acre)/yr (fig. 28*B*). For the 20-, 40-, and 50-year scenarios without lateral flow, the EPS rate was 1.93 (acre-ft/acre)/yr (table 12).

Stream base flow from upstream and the multiple tributaries sustained saturation in the alluvial deposits for the majority of the groundwater-flow model area, particularly Reach IIa (pl. 7). Water-table declines from groundwater pumping resulted in outflows from the Canadian River and tributaries to the nearby alluvial deposits. As a result, stream base flow decreased downstream to nearly zero flow between the Union City site and the Purcell streamgage (07229200). Below the Purcell streamgage (07229200), inflow from Walnut Creek sustained streamflow in the most downstream section of Reach IIa and the upstream area of Reach IIb. Inflow from Little River sustained streamflow downstream to Eufaula Lake.

Lateral flow primarily sustained saturation in Reach IIb because of the large base-flow gain and narrow alluvial deposit extent. As a result, the alluvial deposits remained mostly saturated in Reach IIb even when stream base flow was minimal. When lateral flow was removed, the EPS pumping rate substantially decreased from 4.30 to 1.93 (acre-ft/acre)/yr in Reach IIb, compared to a decrease from 2.75 to 2.28 (acre-ft/acre)/yr in Reach IIa (fig. 28, table 12). The majority of the terrace material was unsaturated for each EPS scenario. Water availability under the maximum EPS pumping rates was primarily in the alluvial deposits.

The EPS pumping rate results for the 20-, 40-, and 50-year periods were largely the same for each scenario type. In the first 8 simulated years of each scenario, substantial amounts of water were removed from storage by the wells in each model cell in Reach II (fig. 27*B*) as the system approached steady-state conditions. During this time, groundwater pumping decreased (fig. 27*B*) as the thinner alluvial and terrace areas were dewatered and dropped below 5 ft of saturated thickness or went dry. This time period was shorter because of an average calibrated-model saturated thickness of 26.2 ft in Reach II versus 33.2 ft in Reach I, thus providing less available water from storage. Additionally, the increased stream base flow in Reach II compared to that in Reach I sustained more saturation for cells near the stream channel; thus, other cells upgradient to the stream channel were dewatered more rapidly, and the remaining cells remained saturated through the end of the scenarios.

Projected Water Use

The calibrated groundwater-flow models were used to determine the water available for groundwater use on the basis of the effects of current groundwater use projected over a 50-year period and a sustained drought. Future water availability was determined on the basis of the simulated amount of groundwater in storage and the simulated streamflow in the Canadian River. Groundwater storage is defined as the amount of water that can be pumped on the basis of the saturated thickness and specific yield.

Fifty-Year Water Use

A 50-year water-use scenario was used to evaluate the effects of water use on the groundwater resources of the Canadian River alluvial aquifer. The scenario used pumping rates from 2013, the last year of the calibrated transient groundwater-flow models, repeated over a 50-year period. The projected water-use scenario was configured the same as the 50-year EPS scenario, starting in year 2013, stepping backward to 1980, and then stepping forward to 1998. Model stresses for recharge, water-table ET, and tributary inflows were configured as the average of each stress period used in the calibrated groundwater-flow models. The effects of 2013 pumping rates were evaluated by comparing changes in groundwater storage between a 50-year scenario by using 2013 pumping rates and a 50-year scenario with no groundwater pumping. Groundwater storage at the end of the 50-year period was calculated by multiplying the calibrated specific yield by the saturated thickness for each active model cell in the groundwater-flow models.

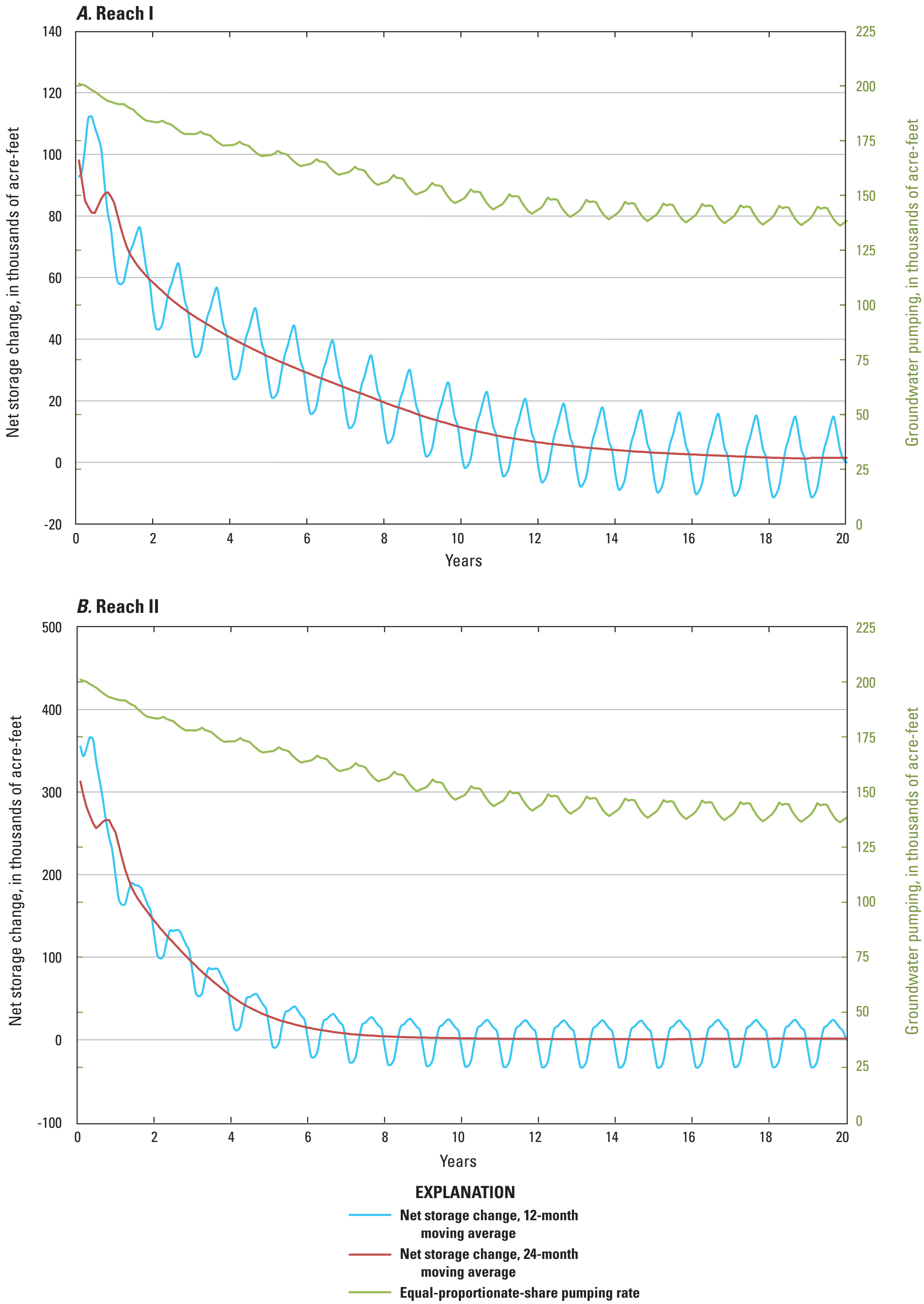


Figure 27. Net change in groundwater in storage and groundwater pumping for *A*, Reach I, and *B*, Reach II, of the Canadian River alluvial aquifer, Oklahoma, during 20 years of continuous equal-proportionate-share groundwater pumping.

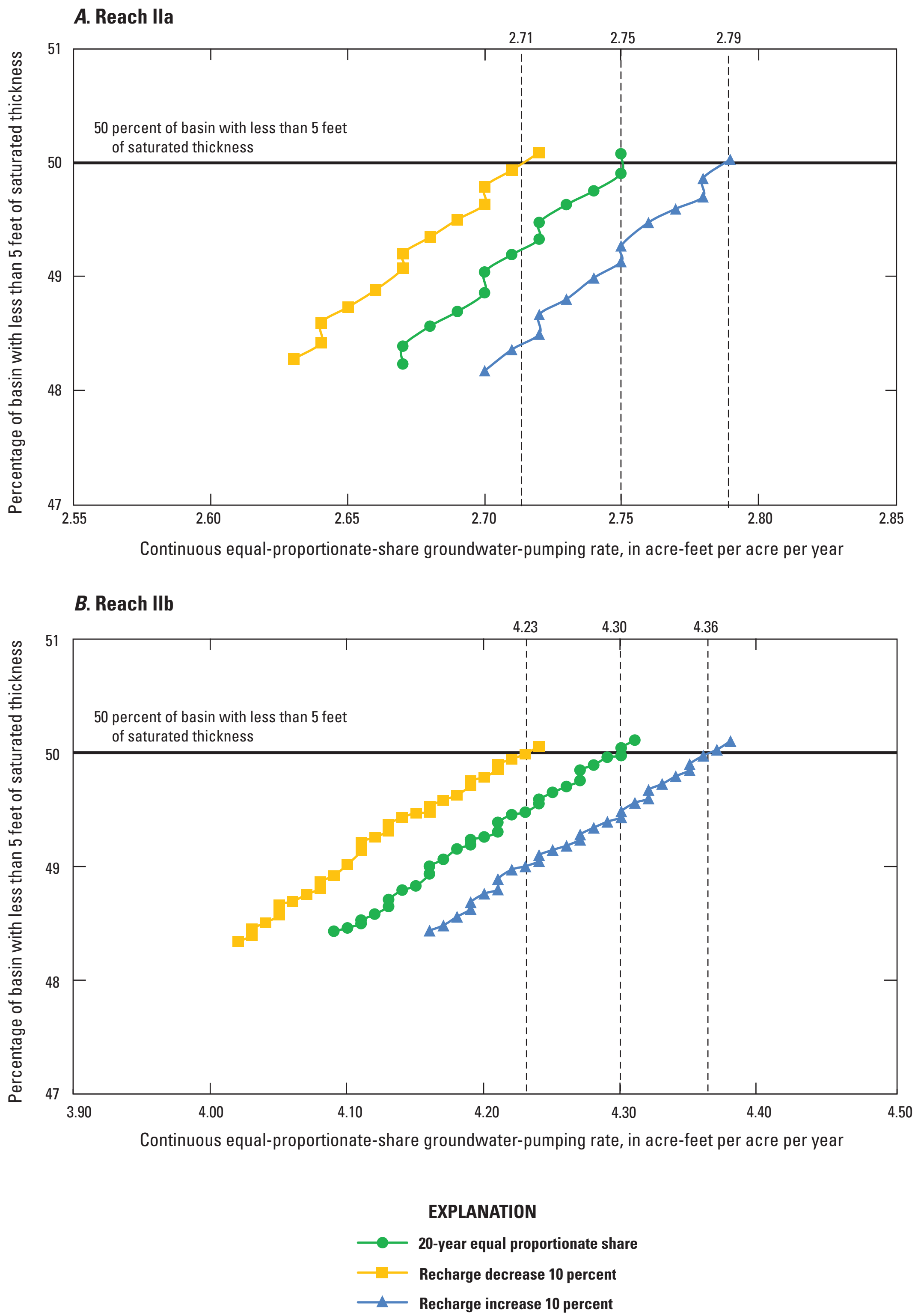


Figure 28. Percentage of *A*, Reach IIa, and *B*, Reach IIb, of the Canadian River alluvial aquifer, Oklahoma, with less than 5 feet of saturated thickness after 20 years of continuous equal-proportionate-share groundwater pumping.

Based on the results of the 50-year groundwater-use scenarios, the 2013 pumping rates can be sustained with only small decreases in groundwater storage. For Reach I, the groundwater-storage totals at the end of the 50-year period with and without groundwater pumping were 885 thousand acre-ft and 887 thousand acre-ft, respectively (table 13). The decrease in groundwater storage (0.2 percent) was small because groundwater pumping in Reach I was only a small percentage of the calibrated groundwater-flow model budget. For Reach II, the groundwater-storage totals at the end of the 50-year period with and without groundwater pumping were 1,343 thousand acre-ft and 1,351 thousand acre-ft, respectively (table 13). As in Reach I, the decrease in groundwater storage (0.6 percent) was small because groundwater pumping in Reach II was a relatively small percentage of the groundwater-flow model budget.

Some assumptions underlie these groundwater-storage projections. This scenario provides for no additional growth in the demand for groundwater use, which is primarily used for public supply and irrigation. Groundwater use has decreased 17 percent in Reach I but has increased 44 percent in Reach II over the model period (1981–2013). Additionally, the scenario assumes that future climate conditions are similar to those of the model period (1981–2013). Future changes in climate variability and duration, such as the drought conditions observed during years 2010–15, may affect the amount of available water and storage and the usefulness of these projected water-use scenarios.

Sustained Drought

Sustained drought scenarios were used to evaluate the effects of a hypothetical 10-year drought on water availability. A 10-year period starting in January 1984 and ending December 1993 was chosen where the effects of drought conditions would be simulated. This time period was chosen because of the similarity of stream base flows between these dates and the average annual stream base flow for the entire transient model period. One scenario was used to calculate the effects of a drought on stream base flow, and the second was used to calculate effects of drought on groundwater storage.

Table 13. Changes in groundwater storage after 50 years of groundwater pumping at the 2013 rate for the Canadian River alluvial aquifer, Oklahoma.

[All units are in thousands of acre-feet unless specified]

Reach	Water in storage with pumping at 2013 rate (percent decrease)	Water in storage without pumping	Water-table decline (in feet)
Reach I	885 (0.2)	887	0.1
Reach II	1,343 (0.6)	1,351	0.2

Both scenarios compared conditions during the hypothetical 10-year drought to the calibrated groundwater-flow models.

To simulate the effects of a sustained drought on stream base flow, recharge was reduced by 75 percent in the Reaches I and II groundwater-flow models for the duration of the simulated drought period. This reduction in average annual recharge for both reaches was estimated on the basis of the average recharge of the years with lowest recharge during the model period. To approximate stream base-flow conditions during an extended drought, tributary inflows were reduced by 60 percent in Reach I and by 77 percent in Reach II. This decrease represents the average decrease in annual stream base flow during the drought of years 1976–81 (Shivers and Andrews, 2013) at the Bridgeport streamgage (07228500) and Calvin streamgage (07231500) compared to the model period. The reduction in tributary inflows includes the estimated stream base flow that crosses the border into Oklahoma from Texas in Reach I and the Canadian River inflow into Reach II, as well as the Deer Creek, Buggy Creek, Walnut Creek, and Little River tributaries. Changes in the stream base flow were evaluated at the Bridgeport streamgage (07228500) in Reach I and at the Purcell streamgage (07229200) and Calvin streamgage (07231500) in Reach II.

Stream Base Flow

In Reach I, average simulated stream base flow at the Bridgeport streamgage (07228500) decreased by 58 percent during the 1984–93 hypothetical 10-year drought compared to average simulated stream base flow during the nondrought period (fig. 29). The decrease in stream base flows during these drought conditions was greatest during simulated periods of higher flow, in which stream base flow decreased by as much as 70 percent (fig. 29). In Reach II, average simulated stream base flows at the Purcell streamgage (07229200) and Calvin streamgage (07231500) decreased by 64 percent and 54 percent, respectively (fig. 30*A*, fig. 30*B*). During higher flow conditions in 1989, stream base flow at the Purcell streamgage (07229200) and Calvin streamgage (07231500) decreased by a maximum of 84 percent and 74 percent, respectively. Simulated low-flow conditions in some months may be underestimated during the drought scenario because periods of nearly zero flow periodically occur in the transient period for both reaches that could not be reproduced in the calibrated groundwater-flow model.

Groundwater in Storage

To simulate the effects of a sustained drought on groundwater in storage, drought conditions from 2003 were used for both reaches, which represent the year in which the most severe drought conditions were observed during the model period. The monthly recharge, ET, and stream base-flow values from that year were repeated over the hypothetical 10-year drought period, and the resulting changes on storage were then compared to initial water in storage. Changes in groundwater storage for both reaches were measured by

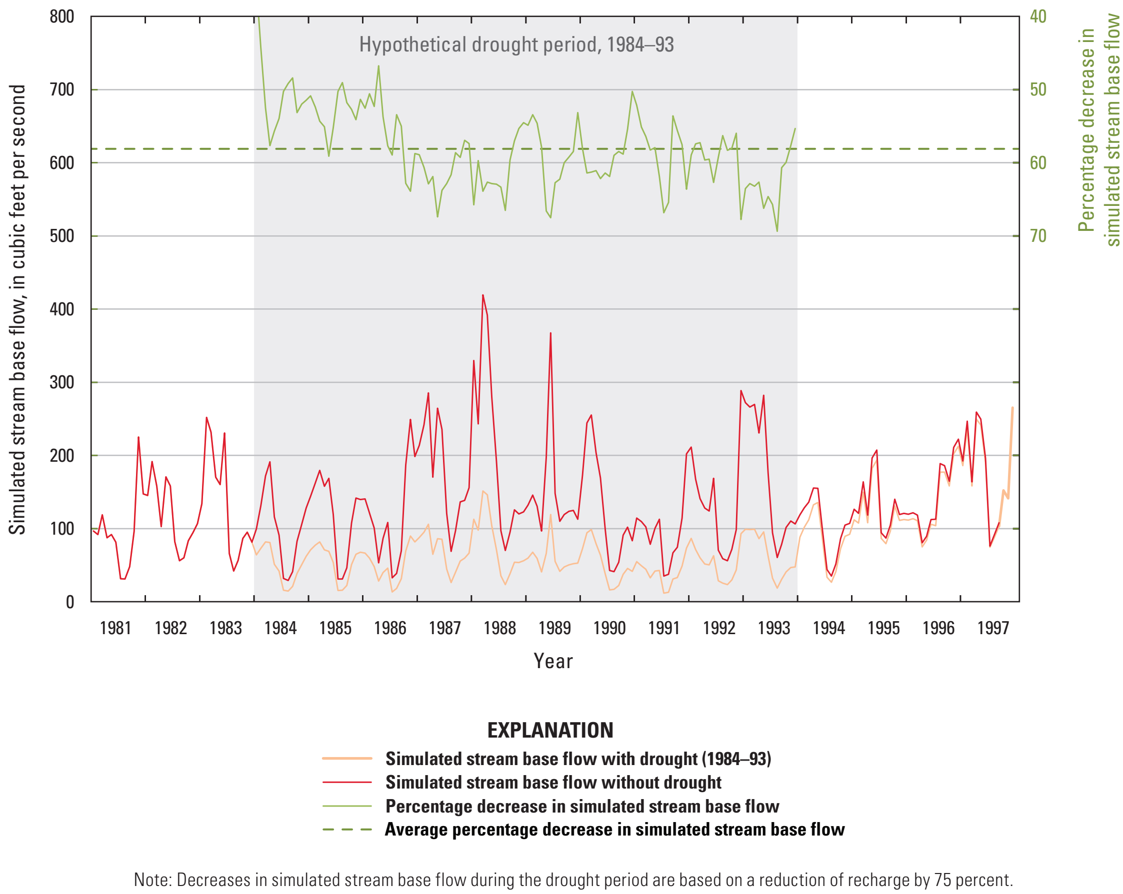


Figure 29. Changes in stream base flow in Reach I of the Canadian River at the Bridgeport, Oklahoma streamgage (07228500) during a hypothetical 10-year drought for the Canadian River alluvial aquifer, Oklahoma.

using an annual moving average of stress-period groundwater volumes. In Reach I, groundwater in storage declined from 872 thousand acre-ft to 842 thousand acre-ft, a total groundwater-storage decline of 30 thousand acre-ft (fig. 31), or an average decline in the water table of 1.2 ft. In Reach II, groundwater in storage declined from 1,260 thousand acre-ft to 1,189 thousand acre-ft, a total groundwater-storage decline of 71 thousand acre-ft (fig. 31), or an average decline in the water table of 2.0 ft. This scenario uses 1 drought year repeated for each of the 10 years of drought conditions, and therefore the results may understate the severity of a sustained drought on the groundwater in storage for the Canadian River; however, this scenario does provide a generalized representation of changes to groundwater in response to sustained drought conditions.

Assumptions and Limitations

Some assumptions and simplifications were necessary in the simulation of groundwater flow. The use of the MODFLOW code to simulate groundwater flow in an aquifer assumed that groundwater flows are governed by Darcy’s Law, water is incompressible and of uniform density, and the aquifer hydrogeology can be simulated appropriately by the cell size and number of layers present. The hydrogeologic characteristics of the aquifer vary at a scale smaller than that of the model cell size. Computing limitations prevented the use of cell sizes that could better represent these heterogeneities, assuming such variability could be reproduced. As a result, the response of the calibrated groundwater-flow models may vary compared to actual conditions, and results generated by the model may be more applicable to a regional, rather than local area.

Spatial and temporal data gaps in the water-table altitude observations occurred because of the uneven distribution of these observations. Although the simulated water table in these areas is in an expected water-table altitude range, more site-specific and local calibration target data would facilitate a more detailed characterization of water-table conditions. Additionally, base-flow gain to the Canadian River is based on the simulated water table and may not be well represented in locations where observation data were relatively sparse.

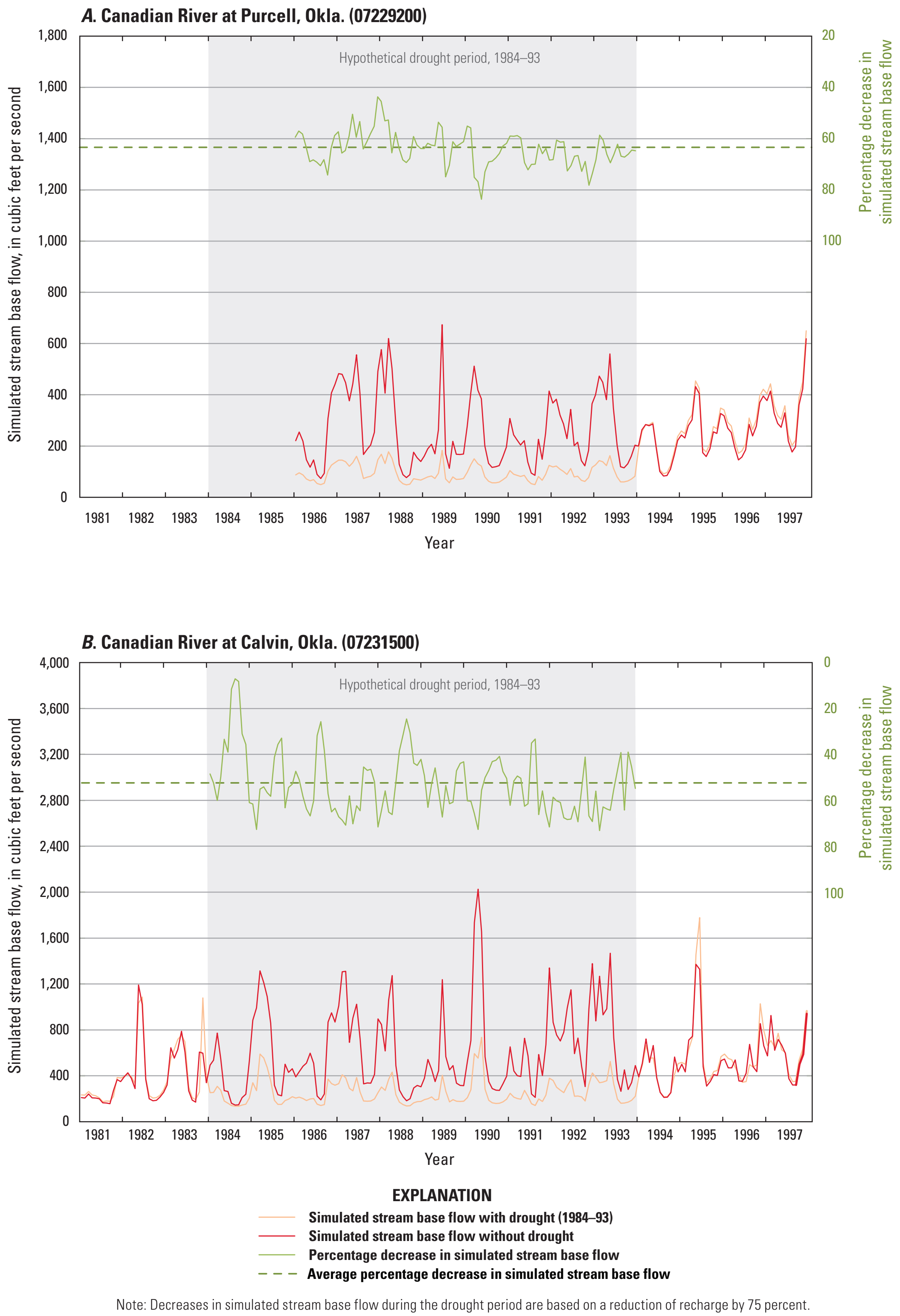


Figure 30. Changes in stream base flow in Reach II of the Canadian River at *A*, Purcell, Oklahoma (07229200), and *B*, Calvin, Oklahoma (07231500), streamgages during a hypothetical 10-year drought for the Canadian River alluvial aquifer, Oklahoma.

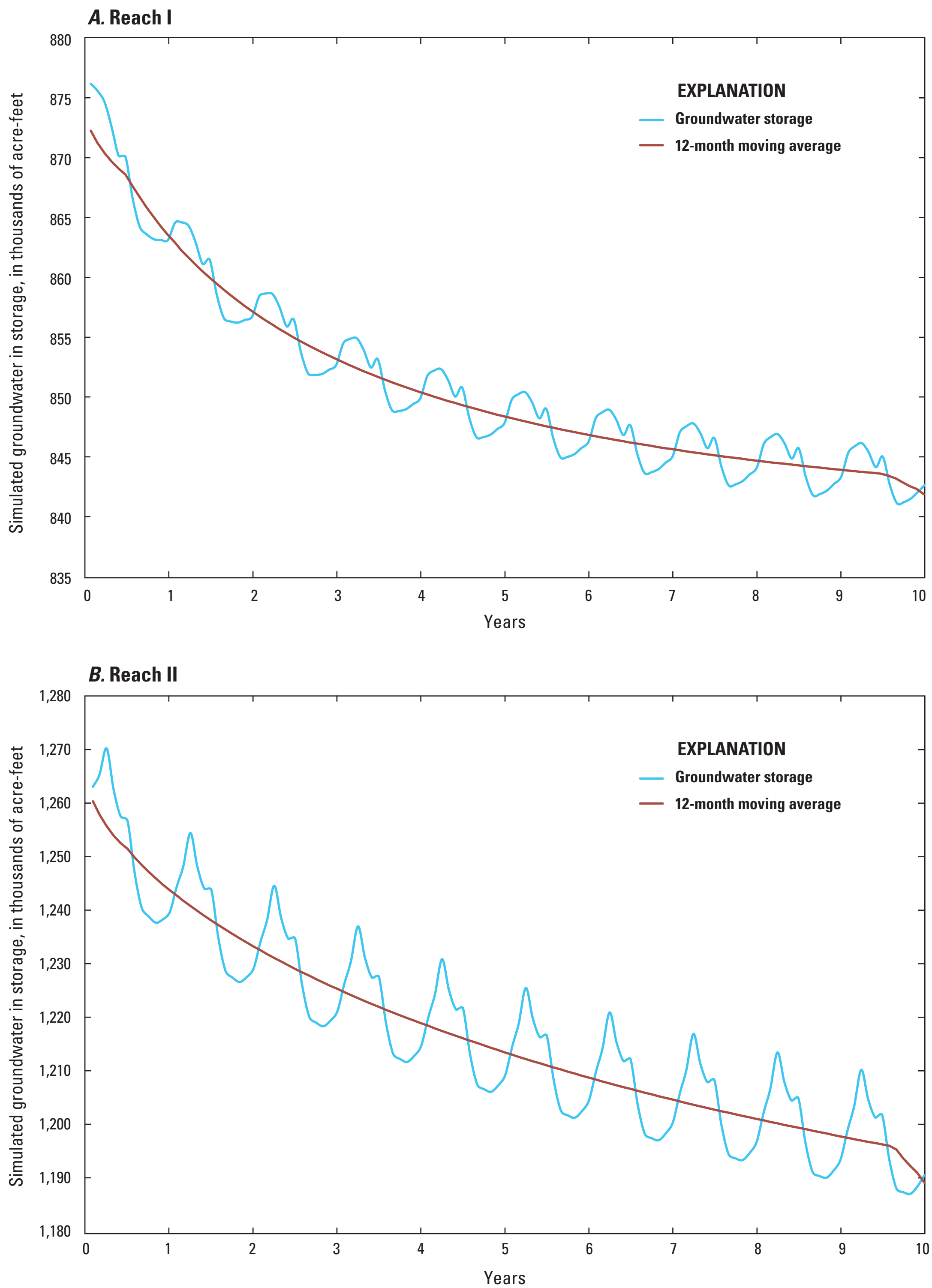


Figure 31. Change in groundwater storage during a hypothetical 10-year drought in *A*, Reach I, and *B*, Reach II, for the Canadian River alluvial aquifer, Oklahoma.

The stream network used in the groundwater-flow models is a simplification of the actual stream geometry and hydraulic properties. More work could be done to refine the stream channel width, streambed hydraulic conductance, and water exchange with the aquifer, all which can vary substantially at the local scale. The groundwater-flow models are calibrated primarily to stream base-flow estimates; therefore, collection of more streamflow data would further reduce uncertainty in local-scale simulation results.

Exact amounts of reported groundwater use are unknown because groundwater wells are not metered, and water-use data are based on self-reported estimates submitted to the OWRB by permit holders. Reported groundwater use declined substantially after 1980, which may be have been caused by changes in reporting requirements; thus, the actual groundwater available for permitted use may differ than was simulated in the groundwater-flow models. Additionally, domestic wells are not included in the permitted groundwater use, although they are not expected to have substantial effects on the water budget because of the relatively small amount of groundwater use by domestic wells in the study area.

Summary

This report describes a study of the hydrogeology and simulation of groundwater flow for the Canadian River alluvial aquifer in western and central Oklahoma. The report (1) quantifies the groundwater resources of the Canadian River alluvial aquifer by developing a conceptual model; (2) summarizes the general water quality of the Canadian River alluvial aquifer groundwater by using data collected during August and September 2013; (3) evaluates the effects of estimated equal proportionate share (EPS) on aquifer storage and streamflow for time periods of 20, 40, and 50 years into the future by using numerical groundwater-flow models; and (4) evaluates the effects of present-day groundwater pumping over a 50-year period and sustained hypothetical drought conditions over a 10-year period on stream base flow and groundwater in storage by using numerical flow models. The Canadian River alluvial aquifer is an unconfined alluvial aquifer consisting of beds of clay, silt, sand, and fine gravel sediments in western and central Oklahoma. Groundwater in approximately 606 square miles of the aquifer extent is used for irrigation, municipal use, mining (oil and gas), livestock, and domestic supply. For the study described in this report, the Canadian River alluvial aquifer was divided into two sections: Reach I, with an extent of 242 square miles from the Texas border to the Canadian River at the Bridgeport, Okla., streamgage (07228500); and Reach II, with an extent of 364 square miles from the Canadian River at the Bridgeport, Okla., streamgage (07228500) to Eufaula Lake.

Land cover on the Canadian River alluvial aquifer was composed of mostly grass/pasture, crops, and forest in 2015. The average annual precipitation in the study area from 1896 to 2014 was 34.4 inches per year (in/yr). On average, the greatest amounts of precipitation occur in May, and the least amounts occur in January. Precipitation trends indicate (1) below-average precipitation from the late 1890s through the early 1920s, (2) above-average precipitation from the late 1920s to the late 1940s, (3) variable precipitation between 1950 and the early 1980s, (4) above-average precipitation from the mid-1980s to 2010, and (5) below-average precipitation from 2010 to 2014.

The average annual groundwater use from this aquifer from 1967 to 2013 was 11,887 acre-feet per year (acre-ft/yr). The annual groundwater-use data indicate a period of greater use from 1967 to 1979, followed by a period of lower use from 1980 to 1999. From 2000 to 2013, groundwater use increased, with a peak of 23,380 acre-ft/yr in 2012. The greatest use of groundwater from 1967 to 2013 was for irrigation, 8,476 acre-ft/yr, or about 71.3 percent of groundwater use.

The aquifer extent was determined primarily by using 1:100,000-scale geologic maps, approximately 1,400 driller’s lithologic logs, and 6 sediment cores. The aquifer areal extent ranged from less than 0.2 to 8.5 miles wide. The maximum aquifer thickness was 120 feet (ft), and the average aquifer thickness was 50 ft. Average horizontal hydraulic conductivity for the aquifer was estimated at 39 feet per day, with a maximum of 100 feet per day and a minimum of 0.1 foot per day.

Water quality in the Canadian River alluvial aquifer was evaluated through summary of data from samples collected for the Oklahoma Water Resources Board Groundwater Monitoring and Assessment Program at 31 sites during 2013. Most of the samples met primary drinking-water standards established by the U.S. Environmental Protection Agency, but 5 of the 31 samples contained nitrate-nitrogen concentrations (measured as nitrate plus nitrite concentrations) that exceeded the standard of 10 milligrams per liter. Changes in water types from the western portion to the eastern portion of the Canadian River alluvial aquifer indicate interaction of groundwater between local bedrock units and the alluvial aquifer.

Stream base flow in both reaches was analyzed by using one seepage run and a base-flow separation method. Increases in the average stream base flow were observed from upstream to downstream, although drought conditions affected the 2013 seepage-run measurements. The average annual base-flow index (1981–2013), estimated by using the Base-Flow Index method, generally decreased from west to east at streamgages on the Canadian River. This general decrease in base flow-index and increase in runoff may be due to the increased precipitation in Oklahoma from west to east, incised alluvial deposits in eastern Oklahoma, and downstream decrease in horizontal hydraulic conductivity.

Recharge rates to the Canadian River alluvial aquifer were estimated by using a soil-water-balance code to estimate the spatial distribution of groundwater recharge and a water-table fluctuation method to estimate localized recharge

rates. By using daily precipitation and temperature data from 39 climate stations, recharge was estimated to average 3.4 in/yr, which corresponds to 8.7 percent of precipitation as recharge for the Canadian River alluvial aquifer from 1981 to 2013. The highest estimated annual recharge of 5.9 in/yr occurred in 1985, and the lowest estimated annual recharge of 0.6 in/yr occurred in 2003. The water-table fluctuation method was used at one site where continuous water-level observation data were available to estimate the percentage of precipitation that becomes groundwater recharge. Estimated annual recharge at that site was 9.7 in/yr during 2014.

Groundwater flow in the Canadian River alluvial aquifer was identified and quantified by a conceptual flow model for the model period 1981–2013. Groundwater inflows to the Canadian River alluvial aquifer included recharge to the water table from precipitation, lateral flow from the surrounding bedrock, and flow from the Canadian River. Total annual recharge inflows estimated by the soil-water-balance code were multiplied by the area of each reach and then averaged over the simulated period to produce an annual average of 28,919 acre-ft/yr for Reach I and 82,006 acre-ft/yr for Reach II. Average annual lateral inflow during the model period was 43,803 acre-ft/yr for Reach I and 164,849 acre-ft/yr for Reach II. Outflows include flow to the Canadian River (base-flow gain), evapotranspiration, and groundwater use. Average base-flow gain was determined to be the largest outflow at 38,006 acre-ft/yr for Reach I and 205,377 acre-ft/yr for Reach II. Base-flow gain rates ranged from 0.3–2.3 cubic feet per second (ft^3/s) per mile. Groundwater use accounts for a minor percentage of outflows from the Canadian River alluvial aquifer, totaling 593 acre-ft/yr in Reach I and 9,416 acre-ft/yr in Reach II.

Objectives for the numerical groundwater-flow models included simulating groundwater flow in the Canadian River alluvial aquifer from 1981 to 2013 to address groundwater use and drought scenarios, including calculation of the EPS pumping rates. The EPS for the alluvial and terrace aquifers is defined by the Oklahoma Water Resources Board as the amount of fresh water that each landowner is allowed per year per acre of owned land to maintain a saturated thickness of at least 5 ft in at least 50 percent of the overlying land of the groundwater basin for a minimum of 20 years.

The groundwater-flow models were calibrated to water-table altitude observations, streamgage base flows, and base-flow gain to the Canadian River. Both models were first calibrated manually by trial and error, followed by automated nonlinear regression techniques by using the parameter estimation code. The Reach I water-table altitude observation root-mean-square error was 6.1 ft, and 75 percent of water-table altitude residuals were within ±6.7 ft of observed measurements. The average simulated stream base-flow residual at the Bridgeport, Okla. streamgage (07228500) was 8.8 ft^3/s, and 75 percent of stream base-flow residuals were within ±30 ft^3/s of observed measurements. Simulated base-flow gain in Reach I was 8.8 ft^3/s lower than estimated base-flow gain. The Reach II water-table altitude observation root-mean-square error was 4 ft, and 75 percent of water-table altitude residuals were within ±4.3 ft of the observations. The average simulated stream base-flow residual in Reach II was between 35 and 132 ft^3/s. The average simulated base-flow gain residual in Reach II was between 11.3 and 61.1 ft^3/s.

Several future predictive scenarios were run, including estimating the EPS pumping rate, determining the effects of current groundwater use over a 50-year period into the future, and evaluating the effects of a sustained drought on water availability for both reaches. The EPS pumping rate was determined to be 1.35 acre-feet per acre per year ([acre-ft/acre]/yr) in Reach I and 3.08 (acre-ft/acre)/yr in Reach II for a 20-year period. For the 40- and 50-year periods, the EPS pumping rate was determined to be 1.34 (acre-ft/acre)/yr in Reach I and 3.08 (acre-ft/acre)/yr in Reach II.

A 50-year water-use scenario was used to evaluate the effects of recent groundwater use rates on future groundwater resources of the Canadian River alluvial aquifer. The scenario used groundwater-pumping rates from 2013, the last year of the calibrated groundwater-flow models, repeated over a 50-year period. For Reach I, the groundwater-storage totals at the end of the 50-year period with and without groundwater pumping were 885 thousand acre-feet (acre-ft) and 887 thousand acre-ft, respectively. For Reach II, the groundwater-storage totals at the end of the 50-year period with and without groundwater pumping were 1,343 thousand acre-ft, and 1,351 thousand acre-ft, respectively. The small changes in storage are due to groundwater use by pumping composing a small percentage of the total groundwater-flow model budgets for Reaches I and II.

Two scenarios were run to estimate the effects of a hypothetical 10-year drought on water availability. One scenario was used to calculate the effects of a drought on stream base flow by decreasing recharge by 75 percent, and the second was used to calculate effects of that drought on groundwater storage, which is the amount of water that can be extracted on the basis of the saturated thickness and specific yield. A 10-year period starting in January 1984 and ending December 1993 was chosen where the effects of drought conditions would be simulated.

In Reach I, average simulated stream base flow at the Bridgeport streamgage (07228500) decreased by 58 percent during the hypothetical 10-year drought compared to average simulated stream base flow during the nondrought period. In Reach II, average simulated stream base flows at the Purcell (07229200) and Calvin (07231500) streamgages decreased by 64 percent and 54 percent, respectively. In Reach I, the groundwater-storage drought scenario resulted in a storage decline from 872 thousand acre-ft to 842 thousand acre-ft, a total groundwater-storage decline of 30 thousand acre-ft, or an average decline in the water table of 1.2 ft. In Reach II, the groundwater-storage drought scenario resulted in a storage decline from 1,260 thousand acre-ft to 1,189 thousand acre-ft, a total groundwater-storage decline of 71 thousand acre-ft, or an average decline in the water table of 2.0 ft.

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