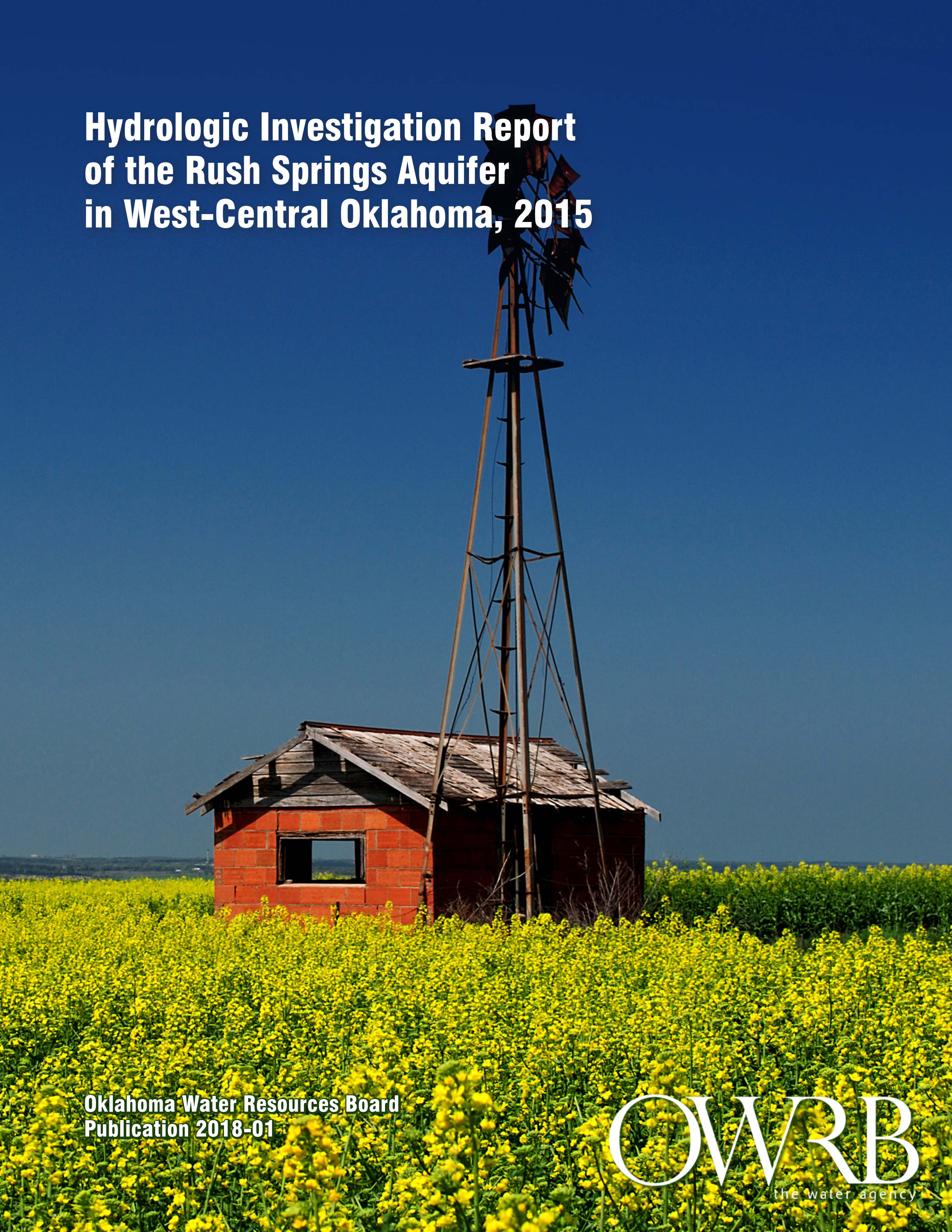
Hydrologic Investigation Report of the Rush Springs Aquifer in West-Central Oklahoma, 2015



Oklahoma Water Resources Board
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Cover. Near Thomas, Oklahoma, spring 2012. Photo by Christopher Neel

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For more information on the OWRB, Oklahoma's water agency, visit *www.owrb.ok.gov* or call 405-530-8800.

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Abstract

The Oklahoma Water Resources Board (OWRB) conducts hydrologic investigations and surveys of the state's groundwater basins as mandated by the State of Oklahoma to determine maximum annual yield (MAY) and equal-proportionate share (EPS). This report details the findings of the Rush Springs hydrologic investigation and provides information for constructing a groundwater-flow model to allow the OWRB to simulate various management scenarios.

The Rush Springs aquifer underlies 4,692 square miles of west-central Oklahoma, including portions of Blaine, Caddo, Canadian, Comanche, Custer, Grady, Stephens, and Washita Counties. The geographic boundaries of the Rush Springs previously determined by the U.S. Geological Survey (USGS) during a 1998 water resources investigation (Becker and Runkle, 1998) were expanded for this investigation to include portions of the Rush Springs and Marlow geologic formations that are part of the same groundwater-flow system.

The Permian-age Rush Springs Formation, the main water-bearing geologic formation in the aquifer, is predominantly a fine-grained sandstone with some dolomite and gypsum. The formation outcrops in the east and is overlain in the west by the Cloud Chief Formation, a siltstone with massive beds of gypsum. Below the Rush Springs Formation is the Marlow Formation, consisting predominately of siltstones and shales with some sandstones. Higher transmissive zones of the upper Marlow Formation, which are likely in hydrologic connection with the Rush Springs Formation, are considered part of the Rush Springs groundwater-flow system. Quaternary alluvium and terrace deposits from the Canadian River and Washita River, which are likely in hydrologic connection with the Rush Springs Formation, are also considered part of the aquifer's groundwater-flow system where they overlie the Rush Springs and Marlow Formations.

Average precipitation in the study area for 1905–2015 was 28.20 inches. A lengthy dry cycle occurred during 1936–1984 with an average annual precipitation of 26.90 inches, followed by a wet cycle during 1985–2008 with an average annual precipitation of 31.28 inches. Long-term annual water-level measurements for 1905–2015 typically correspond to the dry and wet cycles with lower elevations from 1970 to early 1980 followed by higher water-level elevations in the mid-1980s through the early 2000s.

The contours of the potentiometric surface, estimated using 2013 groundwater levels, bow in a "V" shape upstream along the Canadian River and Washita River as well as major tributaries, indicating that groundwater from the aquifer discharges as base flow to surface water features. Many streams, including Cobb Creek, Lake Creek, and Willow Creek, help sustain yields in Fort Cobb Reservoir during periods of below average precipitation; other streams discharge to the Washita River and Canadian River. Between the Washita River streamflow gauges at Carnegie (USGS 07325500) and near Clinton (USGS 07325000) for the periods 1964–1986 and 1990–2005, the Base Flow Index (BFI) base-flow separation technique was used to estimate a base flow increase of 158.5 cubic feet per second, which is 71 percent of the base flow at the Washita River streamflow gauge at Anadarko (07326500). From April 2013 to December 2015, Deer Creek discharged into the Canadian River at an average and median streamflow rate of 32.79 and 20.60 cubic feet per second, respectively. Average base flow for the same period was 18.10 cubic feet per second, which is about 20 percent of the base flow farther downstream on the Canadian River at the streamflow gauge at Bridgeport (USGS 07228500).

The BFI base-flow separation technique and RORA method were utilized on streamflow gauge data to determine subsurface watershed annual recharge, which ranged from 0.46 inches in 2006 at the Little Washita River streamflow gauge near Ninnekah (USGS 07327550) to 5.76 inches in 2007 at the Cobb Creek streamflow gauge near Eakly (USGS 07325800). For 1950–2015, annual recharge across the aquifer, calculated using the Soil-Water-Balance model, ranged from 0.03 inches in 1963 to 4.63 inches in 2007 with an average of 1.40 inches.

Reported average annual groundwater use for 1967–2015 was 68,719 acre-feet per year. Irrigation accounted for 91

percent, and public water supply accounted for 7.8 percent. The highest groundwater use reported for a single year was 132,904 acre-feet during 2014, a year with below normal precipitation totals. The lowest groundwater use reported for a single year was 38,405 acre-feet during 2007, a year with above average precipitation.

The base of the aquifer and base of the overlying Cloud Chief Formation were estimated using lithologic logs submitted to the OWRB by licensed water well drillers. Additional sources of information included geophysical logs, cores, and geologic maps. The most notable feature of the base of the aquifer is the axis of the Anadarko Basin that runs through central Caddo County and trends westward through Washita County. The base of the aquifer gradually rises in elevation to the north and northeast. There is also a sharp rise in elevation along the southwestern boundary of the study area. Saturated thickness was estimated by subtracting the base of the Rush Springs aquifer (including the Marlow formation) from the 2013 potentiometric surface, ranging from 0 to 432 feet with a mean value of 181 feet. The aquifer is thinnest in its southeastern portions where the Rush Springs Formation outcrops and has been eroded. Other thin areas of the aquifer are near the towns of Cyril and Cement and the area northeast of Sugar Creek. The thickest saturation is located along the Anadarko Basin axis, where the Cloud Chief Formation confines the Rush Springs aquifer. There is also a zone of thick saturation near the Town of Oakwood where a full section of Rush Springs Formation may be present between the Canadian River and North Canadian River.

Hydraulic properties for the Rush Springs aquifer were estimated using several methods. Drawdown data from 573 well completion reports were used to estimate hydraulic conductivity, which ranged from less than 0.01 feet per day to 90.90 feet per day with a mean and median of 3.30 feet per day and 1.60 feet per day, respectively. Slug tests were performed on 54 wells throughout the aquifer with estimates ranging from 0.13 feet per day to 7.64 feet per day and a mean and median of 1.71 feet per day and 1.46 feet per day, respectively. Three aquifer tests were analyzed with transmissivities ranging from 219 square feet per day to 4,129 square feet per day. Specific yield values ranged from 0.04 to 0.09. Analytical solutions for the aquifer-test data suggest that the Rush Springs Formation acts unconfined at the local scale.

A regional method was performed to determine specific yield, utilizing water-level changes and streamflow gauge data. Spatially distributed water-level changes were used to estimate the change in aquifer volume while streamflow gauges measured the volume of water that drained the aquifer during base flow conditions. Using this method, specific yield estimates for subsurface watersheds were 0.05 for Cobb Creek, 0.07 for Deer Creek, and 0.07 for Lake Creek.

Groundwater-quality data collected from 79 wells indicated a bimodal distribution of water types, which were primarily calcium bicarbonate and secondarily calcium sulfate. Mean and median total dissolved solids were 1,106 and 485 milligrams per liter, respectively. Magnesium and sodium anions were also present in groundwater samples (more prevalent in the calcium-bicarbonate water type), possibly caused by the dissolution of dolomite (magnesium) or clays (sodium). The spatial distribution of magnesium in the groundwater suggests that sulfate concentrations are likely caused by the dissolution of gypsum from the Cloud Chief Formation. Several samples contained concentrations of constituents that exceeded the maximum contaminant level (MCL) for primary drinking water regulations. Four samples exceeded the MCL for arsenic (10 micrograms per liter) with a high concentration of 16.5 micrograms per liter; 13 samples exceeded the MCL for nitrates (10 milligrams per liter) with a high concentration of 59.2 milligrams per liter.

Purpose and Scope

Oklahoma groundwater law requires the OWRB to conduct hydrologic investigations of Oklahoma's aquifers to determine the MAY and EPS. The MAY is defined as the total amount of fresh groundwater that can be produced from an aquifer allowing for a minimum 20-year life of the basin. Life of the basin is defined as the period of time during which the total overlying land of the basin will retain a saturated thickness of 15 feet for bedrock aquifers. The EPS is defined as the portion of the MAY allocated to each acre of land (Oklahoma Water Resources Board, 2014a). The objective of the Rush Springs Hydrologic Investigation is to provide the OWRB with information about the hydrogeology of the aquifer needed to determine the MAY based on various proposed management scenarios. Although a USGS study of the Rush Springs aquifer was completed in 1998 (Becker and Runkle, 1998) and a steady-state groundwater-flow model was completed by the USGS in cooperation with the OWRB and Oklahoma Geological Survey (OGS) (Becker, 1998), the MAY and EPS have not been determined. A future groundwater flow model based on updated parameters will test management scenarios and provide data for allocation decisions.

Introduction

The Rush Springs aquifer of Oklahoma is located in Blaine, Caddo, Canadian, Comanche, Custer, Grady, Stephens, and Washita Counties and includes the communities of Anadarko, Clinton, and Weatherford, among others (Figure 1). The study area for this investigation underlies 4,692 square miles. Groundwater is predominantly used for municipal and irrigation purposes, although other uses include agricultural (non-irrigation), industrial, commercial, and domestic water supply. The aquifer is one of the most utilized groundwater sources in the state. The USGS identified Caddo County as one of Oklahoma's largest groundwater consuming counties (Lurry and Tortorelli, 1995). Public water suppliers that use the aquifer include Caddo County Rural Water

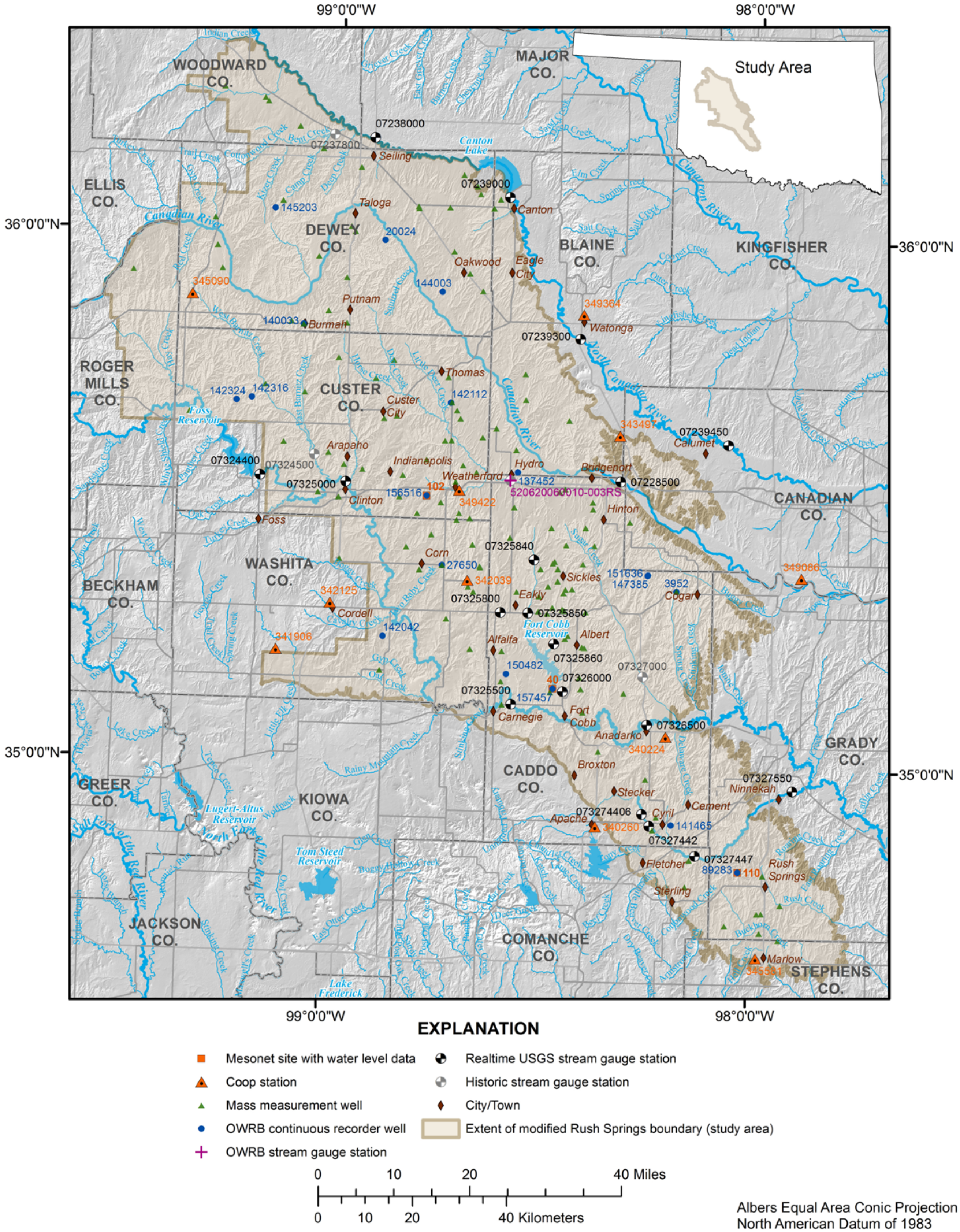

Figure 1. Map showing Rush Springs aquifer location, counties, cities, continuous-recorder wells, periodic water-level wells, cooperative observer stations, and Mesonet weather stations.

District (RWD) #3, Grady County RWD #6, and Washita County RWD #2, and the towns of Corn, Custer City, Cyril, Gracemont, Hinton, Marlow, Mountain View, Thomas, and Weatherford. The USGS estimated withdrawal from the aquifer in 1990 to be about 54.7 million gallons per day, or 61,272 acre-feet per year; 77.8 percent was estimated to be used for irrigation (Becker and Runkle, 1998).

The study area comprises the Western Sandstone Hills, Western Red-bed Plains, Weatherford Gypsum Hills, and the Western Sand-dune belts geomorphic provinces (Curtis and others, 2008). The study area can be characterized as slightly lithified, nearly flat-lying red Permian sandstones with gently rolling hills and occasional steep-walled canyons. The western portion of the aquifer contains the Weatherford Gypsum Hills and is described as gently rolling hills of massive gypsum beds with some sinkholes and caves. In some areas, fields of grass-covered sand dunes lay on top of the bedrock (Curtis and others, 2008).

The predominant geologic formation in the Rush Springs aquifer is the Permian-age Rush Springs Formation. The Rush Springs Formation has been described as an orange-brown, cross-bedded, fine-grained sandstone with some dolomite and gypsum beds, ranging in thickness from 186 to 300 feet (MacLaughlin, 1967; Carr and Bergman, 1976), and consisting of about 50 to 60 percent quartz sand (Allen, 1980). The depositional environment was described by the OGS as a nearshore marine environment with eolian deposits that experienced several marine transgressions (Ham and others, 1957; MacLaughlin, 1967) as evidenced by the presence of feldspar overgrowths that likely formed in marine environments. The Rush Springs Formation is underlain by the Marlow Formation, which was determined to be in hydrologic communication with the Rush Springs Formation, and below this the Dog Creek Shale serves as the confining unit for the aquifer (see Geology section). The western portion of the Rush Springs Formation is capped by the Cloud Chief Formation, which confines the aquifer and may minimize recharge in that area.

Groundwater is discharged as base flow from the aquifer into streams and rivers that flow into Fort Cobb Reservoir, which provides water supply to the communities of Bessie, Clinton, Cordell, and Hobart. Major streams emanate from the aquifer, including Barnitz Creek, Cobb Creek, Deer Creek, and Sugar Creek. The North Canadian River bounds the aquifer to the north where the river has completely eroded the Rush Springs Formation. Streams that discharge from the aquifer to the North Canadian River include Persimmon Creek and Bent Creek. The Canadian River gains base flow from the Rush Springs aquifer (Ellis and others, 2016) with the largest inflow coming from Deer Creek. The Washita River, after flowing into Foss Reservoir, gains flow from the reservoir and flows off of the aquifer near Anadarko (see Streamflow and Base Flow section).

The 2012 Oklahoma Comprehensive Water Plan (OCWP) (Oklahoma Water Resources Board, 2011) anticipates that several planning basins overlying the Rush Springs aquifer will experience significant groundwater depletion by 2060. One of these planning basins is located upstream from Fort Cobb Reservoir, where groundwater depletions could cause a decline of base flow into the reservoir. Additionally, a majority of the groundwater permits in the aquifer are concentrated around and upstream from Fort Cobb Reservoir. Concerns about future reservoir yield and storage have also been identified by the U.S. Bureau of Reclamation (USBR) (U.S. Bureau of Reclamation, 2006; Ferrari, 1994).

Several publications have delineated the aquifer boundary; however, the one most frequently referenced was created by the USGS using outcrop boundaries from hydrologic atlases covering west-central Oklahoma and creating an approximate western boundary where total dissolved solids begin to increase, indicating a change from fresh water conditions to more brackish conditions where the aquifer is confined by the Cloud Chief Formation (Becker and Runkle, 1998).

The study area for this investigation was expanded to include two additional areas where well yields have exceeded 50 gallons per minute, which by definition allows the classification of “major groundwater basin” by the OWRB (Oklahoma Statutes Title 82 Section 1020.1, 2011). Areas to the west were included in the study to delineate the increase in total dissolved solids. Areas north of the Canadian River and south of the North Canadian River extending to near Woodward were added to the study area as well. The eastern outcrop boundary has been updated based on recent geologic maps published by the OGS and includes the Marlow Formation (Chang and Stanley, 2010; Fay, 2010A; Fay, 2010B; Johnson and others, 2003; Stanley, 2002; Stanley and others, 2002; Stanley and Miller, 2004; and Stanley and Miller, 2005).

Climate

Oklahoma has nine climate divisions (Oklahoma Climatological Survey, 2014a). The Rush Springs aquifer is located within Climate Division 4 (west central) and Climate Division 7 (southwest). These climate divisions are classified as semi-arid according to the Koppen climate classification (Oklahoma Climatological Survey, 2014b). The average annual temperature ranges from 58 degrees Fahrenheit in the northern part of the aquifer to 61 degrees Fahrenheit in the southern region (Oklahoma Climatological Survey, 2014c). On average, most of the study area has more than 70 days of temperatures above 90 degrees Fahrenheit per year and fewer than 12 days per year with highs below 32 degrees Fahrenheit (Oklahoma Climatological Survey, 2014d; Oklahoma Climatological Survey, 2014e). The highest temperatures generally occur in July and August and the lowest temperatures generally occur in January. Average annual precipitation totals increase in the southern part of the aquifer.

Precipitation data are collected by Cooperative Observer (COOP) stations, a network of National Weather Service climate-observation volunteers who record observations in a variety of land-use settings (National Weather Service, 2014). Precipitation data were acquired from 13 COOP stations (Oklahoma Climatological Survey, 2016) in or near the study area to analyze long-term precipitation trends. The locations were selected based on the amount of data available and location relative to the study area (Figure 1). Precipitation data retrieved from the COOP observer stations were collected during 1895–2015; however, not all stations had available data for the entire period of record (Table 1). The number of stations in operation during a given year varied from 3 to 13. Years with fewer than 3 stations concurrently recording precipitation were not included in the analysis. For a given year, each station was required to have 10 months of data to be included in the analysis. Data collection methods differed between the Oklahoma Mesonet stations and the COOP stations, resulting in differing precipitation totals. The COOP stations, which began data collection in 1895, had a much longer period of record than the Mesonet stations, which began data collection in 1994. Therefore, precipitation data from the COOP stations alone were used for analysis in this report to maintain consistency.

The average annual precipitation derived from the COOP data was 28.2 inches for 1905–2015 (Figure 2). The data show numerous wet and dry patterns with two longer-term trends: 1936–1984 and 1985–2008. For 1936–1984, precipitation was predominantly below the 110-year average with several smaller patterns of above average precipitation (on the decade scale). During these 49 years, there were 33 years of below average annual precipitation and an overall average of 26.90 inches of annual precipitation, which is 1.2 inches below the 108-year average. For 1985–2008, there were 20 years with above average precipitation and an average annual precipitation, which is 3.18 inches above the long-term average. It should be noted that Tropical Storm Erin in the late summer of 2007 brought an unusually high amount of moisture to the region, which increased the average precipitation during this period (Arndt and others, 2009). Drier conditions have been prevalent from 2008–2015 (Figure 2).

A comparison of monthly data for the two periods of time shows higher monthly average precipitation for most months during the 1985–2008 period, with the exception of May and July (Figure 3). The months of March, June, and August show an increase of about an inch for each month during the wet period compared to the 1936–1984 period. The average monthly precipitation for the period of record was 2.4 inches; May had the highest monthly total at 4.4 inches, and January had the lowest at 1.0 inches. The increase in precipitation during the 1985–2008 period, and the timing of precipitation throughout the year, caused more recharge to the aquifer during months of low evapotranspiration and mitigated the effects of drier months by allowing more water to stay in the soils. Additionally, groundwater use during the wet period was lower than in dry periods (see Groundwater Use section). Increased recharge and decreased groundwater use may have allowed water levels in the aquifer to increase or rebound from stresses.

Table 1. Precipitation data collection time periods at the Cooperative Observer stations used in the Rush Springs aquifer study.

Station number	Station name	Period of analysis*	Number of years	Average annual precipitation, in inches	1936–1984 Average precipitation, in inches	1985–2008 Average precipitation, in inches
340224	Anadarko	1938-2015	78	30.29	25.60	31.29
340260	Apache	1909-2015	92	30.87	N/A	N/A
340332	Arnett 3NE	1911-2015	89	22.83	22.05	N/A
341906	Clinton-Sherman	1958-2015	29	24.08	N/A	N/A
342039	Colony	1983-2015	33	29.84	N/A	31.75
342125	Cordell	1936-2010	74	27.48	25.60	30.60
343497	Geary	1912-2015	104	28.15	27.19	32.06
345090	Leedey	1941-2015	72	24.21	N/A	26.05
345581	Marlow	1900-2013	113	33.87	32.71	38.29
349086	Union City	1914-2015	102	33.15	33.49	36.08
349364	Watonga	1902-2015	91	28.75	27.24	33.06
349422	Weatherford	1905-2015	111	28.63	27.27	31.48
349760	Woodward	1895-2015	114	24.67	24.18	25.79
				Total average, in inches	**27.26**	**31.64**

*Not continuous

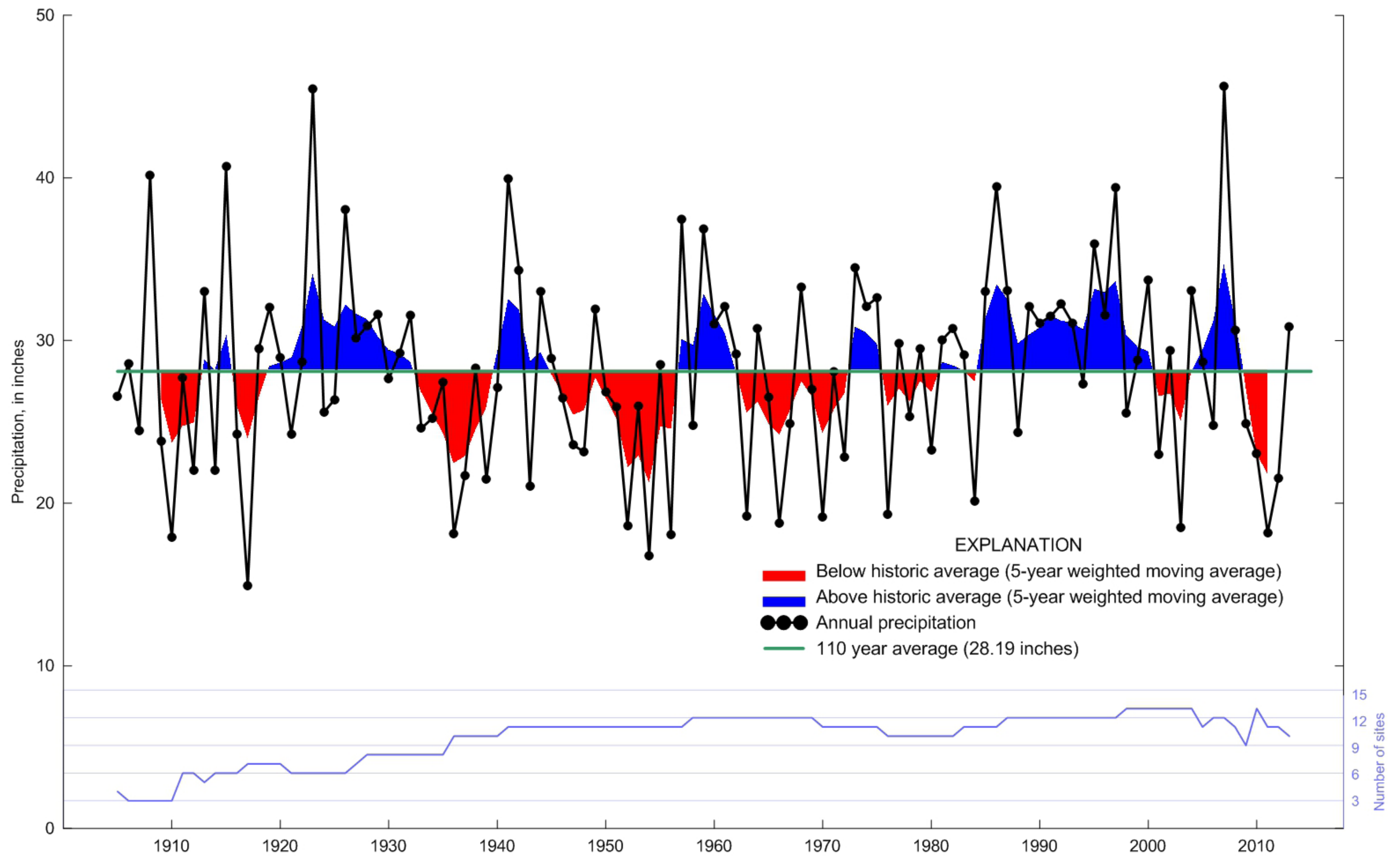

Figure 2. Graph showing annual precipitation and 5-year weighted average (1905–2015) at 13 Cooperative Observer stations.

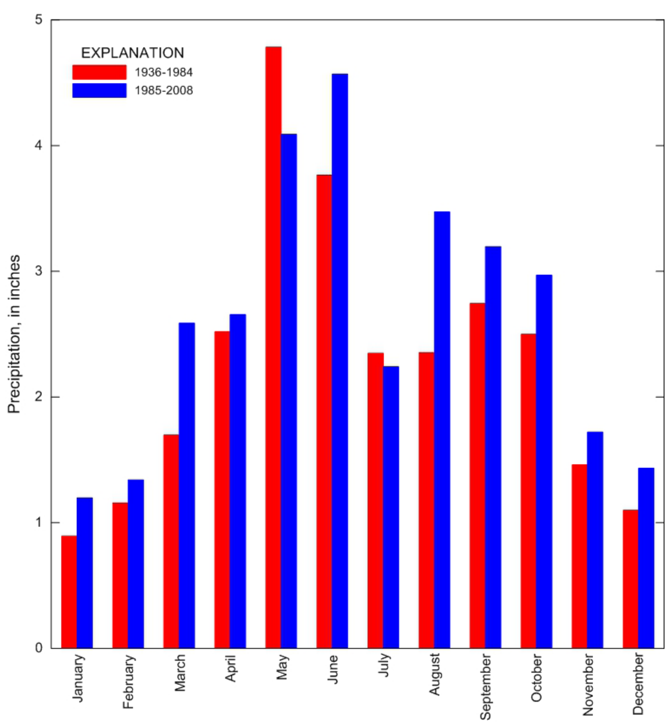

Figure 3. Graph showing average monthly precipitation for 1936–1984 and 1985–2008.

Geology

The Rush Springs aquifer consists of Permian-age Rush Springs Formation and Marlow Formation bedrock units and Quaternary-age alluvium and terrace deposits (Table 2 and Figure 4). The Rush Springs and Marlow Formations together make up the late Permian-age Whitehorse Group (Fay and Hart, 1978). Stratigraphically above the Rush Springs Formation is the Permian-age Cloud Chief Formation, which influences flow and chemistry of the groundwater in the study area. Below the Marlow Formation is the El Reno Group, which is defined as a minor aquifer by the OWRB (Belden, 2000). The upper unit in the El Reno Group is the Dog Creek Shale, which acts as an aquitard between the Rush Springs and El Reno aquifers.

Geologic History and Depositional Environments

Prior to the deposition of the Permian-age Rush Springs Formation, a continental collision in the Pennsylvanian Period between the Laurentia (North American craton) and Gondwana plates (Perry, 1989) caused a structural inversion (i.e., reactivation of older normal faults as reverse faults) of the Southern Oklahoma Aulocogen, creating the Wichita Mountains to the south with a deep foreland basin on the north flank. The foreland basin, called the Anadarko Basin, is the deepest Phanerozoic-age sedimentary basin within the North American craton (Perry, 1989). As the Anadarko Basin was forming, sediments of up to 40,000 feet were deposited in predominantly shallow water environments (Ham and Wilson, 1967). The thickest section (or axis) and northern extent of the Anadarko Basin extends to the southeast from Sherman County in the Texas Panhandle into Oklahoma north of the Wichita Mountains to its apex in south-central Oklahoma. Regional dip along the northern arm of the Anadarko Basin is approximately 20 feet per mile to the south-southwest; regional dip in the southern arm is approximately 50 feet per mile to the north-northeast (Becker and Runkle, 1998). In Oklahoma, the basin is bound by the Nemaha Uplift on the east, the Arbuckle Uplift to the southeast, and the Wichita-Criner Uplifts to the south (Poland, 2011). Within the Anadarko Basin, successively younger strata are exposed westward.

El Reno Group and Beckham Evaporites

The Permian-age El Reno Group in central Oklahoma consists of (from youngest to oldest) the Dog Creek Shale, Blaine Formation, Flowerpot Shale, Cedar Hills Sandstone, Chickasha Formation, and the basal Duncan Sandstone (Table 2). The thickness of the El Reno Group ranges from 700 feet in central Oklahoma to 250 feet in Kansas (Fay, 1962). The Chickasha Formation, Duncan Sandstone, and Cedar Hills Sandstone were deposited in a deltaic environment (Tussy delta) at the mouth of westward- and northwestward-flowing stream systems. The depositional environment shifted to a more restricted shallow sea, resulting in the formation of the Flowerpot Shale, Blaine Formation, and Dog Creek Shale (MacLaughlin, 1967); the Blaine Formation contained more gypsum and dolomite, indicative of an evaporitic environment. The Chickasha Formation, Duncan Sandstone, and Cedar Hills Sandstone have hydraulic properties that allow storage and flow of groundwater. Based on these factors, the OWRB has identified parts of the El Reno Group as a minor aquifer in Oklahoma (Belden, 2000).

During the Permian period, western Oklahoma was located near the equator and shifted between wet and dry climates (Ziegler, 1990). Within the Anadarko Basin, the El Reno Group transitioned from a deltaic system in central Oklahoma to an evaporitic environment in western Oklahoma. The Beckham Evaporites, deposited along the axis of the Anadarko Basin, show this transition and represent a facies change within the El Reno Group (Jordan and Vosburg, 1963; Johnson, 2008). The lower unit in the Beckham sequence is the Flowerpot Salt, which contains salts and shales and occupies the same stratigraphic position as the Flowerpot Shale. The middle unit is the Blaine Anhydrite, which is synonymous to the Blaine Formation with the exception of evaporitic anhydrite beds at the top. The upper unit of the Beckham Evaporites is the Yelton Salt, which represents a salt facies in the lower part of the Dog Creek Formation. The Yelton Salt is located directly west of the study area and ranges in thickness from 0 to 275 feet (Jordan and Vosburg, 1963). The El Reno Group is mentioned in this report to present observations of groundwater use, which is discussed in the Groundwater Use section.

Table 2. Stratigraphic column of geologic and hydrogeologic units in the Rush Springs Aquifer.

Period	Epoch	Group	Formation	Range of thickness, in feet	Aquifer
Upper Permian	Custerian	Foss	Cloud Chief	300[d,f]	
			Moccasin Creek Gypsum Bed	14[b,h]	
		Whitehorse	Weatherford Bed	0-60[c,f]	Rush Springs
			Rush Springs	90-417[a,b,j]	
			Emmanuel Bed	1[a]	
			Gracemont Shale	1[i]	
			Relay Creek Bed	5-10[h]	
			Verden Sandstone	2-10[d]	
			Marlow	100-135[b,h]	
Lower Permian	Cimarronian	El Reno	Dog Creek Shale	30-220[a,d]	El Reno Minor
			Yelton Salt	0-275[g]	
			Blaine	50-215[f,g]	
			Flowerpot Salt	0-250[g]	
			Flowerpot Shale	20-450[d,g]	
			Chickasha	30-600[a,d]	
			Cedar Hills Sandstone	180[a]	
			Duncan Sandstone	100-450[e]	

[a] Morton, 1980
[b] Becker and Runkle, 1998
[c] Hart, 1974
[d] Carr and Bergman, 1976
[e] Bingham and Moore, 1975
[f] Havens, 1977
[g] Jordan and Vosburg, 1963
[h] Fay, 1962
[i] Tanaka and Davis, 1963
[j] Poland, 2008
[k] Green, 1936

Marlow Formation

The Permian-age Marlow Formation, described as an orange-brown, cross bedded, fine grained sandstone and siltstone thinning northward, forms the lower portion of the Whitehorse Group (Carr and Bergman, 1976). Reported thicknesses range from 100 feet in Blaine County (Fay, 1962), 105 to 135 feet in Grady and Stephens Counties (Fay, 1962), 115 feet (Evans, 1928), and 120 feet (Sawyer, 1924). The formation outcrops on both limbs of the Anadarko Basin as a narrow band between half a mile to 5 miles wide, visible along creeks and streams flowing away from the aquifer (Figure 4). An unconformity has been reported to occur at the base of the Marlow Formation, separating it from the older Dog Creek Shale (Green, 1936). However, the Marlow Formation has also been reported as conformable with beds above and below with a conglomerate at the base in place in Grady County, which may be a sign of an erosional surface (Fay, 1962). Contact between the Dog Creek Shale and Marlow Formations has been found to be sharp and distinct (Evans, 1928). This distinction was also observed by OWRB staff in geophysical logs from unpublished work in the study area.

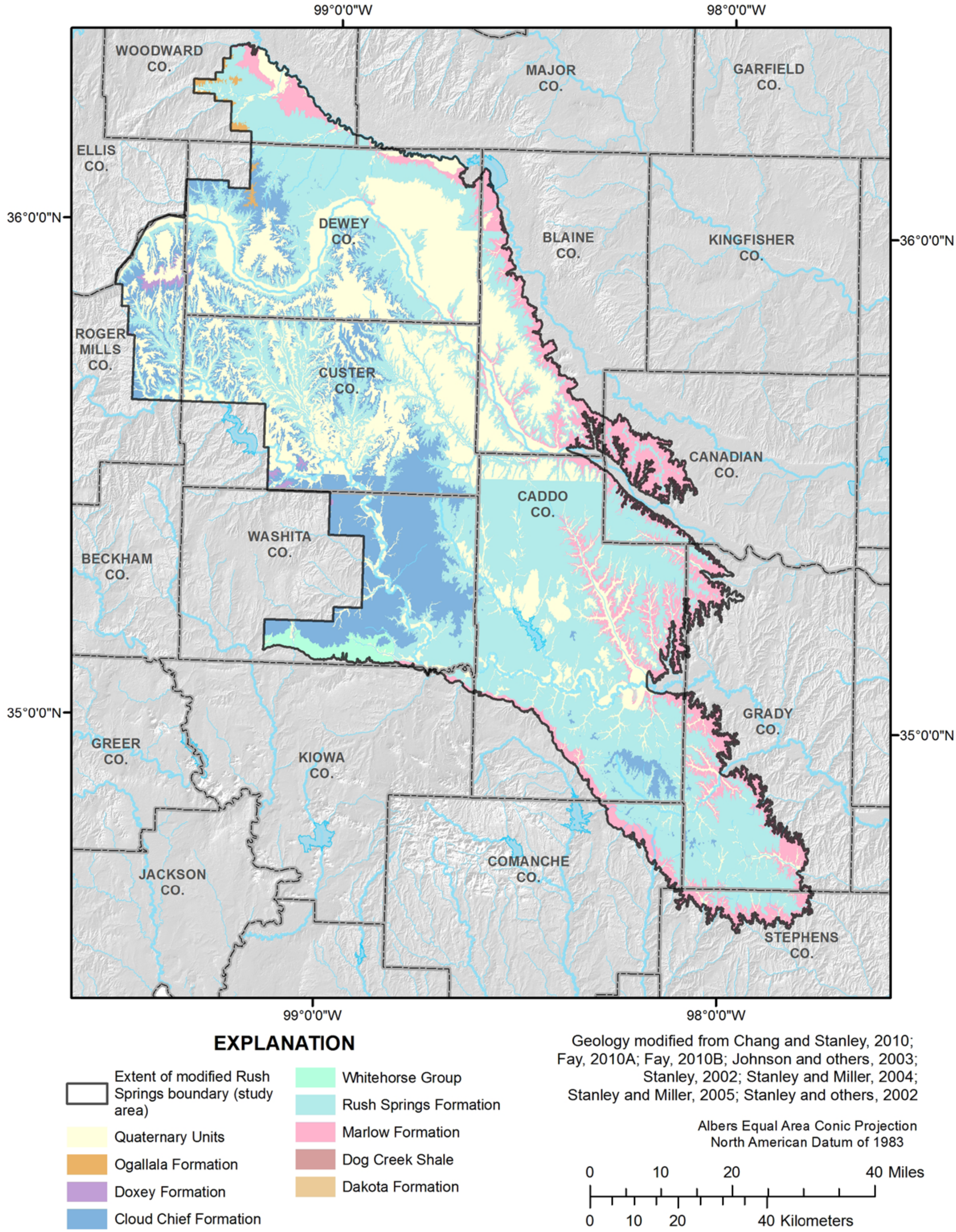

Figure 4. Surficial geologic units in the extent of the Rush Springs aquifer.

The Marlow Formation has many even-bedded and interbedded sandstone, siltstone, and mudstone layers with several gypsum-anhydrite, dolomite, and shale layers, namely the 1-foot thick Emmanuel Bed (gypsum) at the top of the Marlow Formation (Fay, 1962) and the Gracemont Shale directly below the Emmanuel Bed, about 1 foot below the top of the Marlow Formation (Brown, 1937). A pink shale, the result of an altered ash flow, has been described at 10 to 15 inches below the Emmanuel Bed (Tanaka and Davis, 1963); this is likely the Gracemont Shale. Another gypsum/dolomite layer, the Relay Creek Bed, also referred to as the Greenfield Dolomite (Evans, 1928), is situated about 20 to 28 feet below the top of the Marlow Formation (Fay, 1962). The Verden Sandstone, about 45 feet below the top and 85 to 105 feet above the base of the Marlow Formation, is a pinkish-brown, coarse-grained, calcareous, fossiliferous sandstone (Reed and Meland, 1924; Bass, 1939; Carr and Bergman, 1976). The Verden Sandstone ranges from 2 to 10 feet thick (Carr and Bergman, 1976) and is only about 1,000 feet at its widest surface exposure (Bass, 1939). The unit outcrops in Stephens County and trends northwestward into Canadian County (Bass, 1939). The Gracemont Shale and Verden Sandstone are not continuous across the Marlow Formation (Fay, 1962).

The predominant cement in the Marlow Formation is gypsum with small amounts of carbonate (Becker and Runkle, 1998) and iron oxide (Tanaka and Davis, 1963) with the unit typically being moderate to well-cemented and having low permeability. The USGS previously determined that the Marlow Formation acted as a confining unit that retards downward movement of water from the Rush Springs Formation (Becker and Runkle, 1998). However, northward from the town of Anadarko, shales in the Marlow Formation grade into sandstones, contain less gypsum (Green, 1936), and are more likely to store and transmit groundwater.

The presence of marine fossils in parts of the Marlow Formation has been interpreted as deposition in a lagoonal-marine environment that includes brackish water to nearshore-marine setting (Fay, 1962) or a tidal flat bordering an open marine environment (MacLaughlin, 1967). The Verden Sandstone has been described as a river channel that flowed northwestward from the Arbuckle Mountains (Reeves, 1921; Reed and Medland, 1924; Evans, 1949) and also as a barrier island in a broad shallow bay near the shore of a marine sea (Sawyer, 1924; Bass, 1939). The contact between the Marlow Formation and Rush Springs Formation grades from a marine deposition to eolian sand sheet deposition in the lower Rush Springs Formation. The sediment source for the Marlow Formation has been described as originating east-southeastward from the Ouachita Mountains and Ozark Uplift (Fay, 1962).

Rush Springs Formation

The Permian-age Rush Springs Formation, the primary water-bearing unit in the Rush Springs aquifer, is the upper portion of the Whitehorse Group. The term “Rush Springs” was first used in a 1929 publication by the OGS, where the formation was described as mostly red, cross-bedded sandstone located near the town of Rush Springs (Sawyer, 1929). More recent descriptions of the Rush Springs Formation depict orange-brown, coarse-bedded, fine-grained sandstone (Carr and Bergman, 1976) with a silt component (Davis, 1955; Fay, 1962; Tanaka and Davis, 1963), exhibiting predominantly medium to large-scale cross bedding (Reeves, 1921; Al-Shaieb, 1985). Rock cores show a composition primarily of very-fine to fine-grained quartz sand grains (Becker and Runkle, 1998). Quartz grains in the Rush Springs Formation are subround to subangular and moderately to poorly sorted (Davis, 1955; O’Brien, 1963; Tanaka and Davis, 1963; Allen, 1980). The upper portion of the Rush Springs Formation is a gypsum-bearing sandstone that abruptly changes to complete gypsum in the Moccasin Creek Gypsum Bed at the base of the Cloud Chief Formation (Poland, 2011).

Previous investigations considered the upper and lower contact of the Rush Springs Formation to be conformable (Fay, 1962; Tanaka and Davis, 1963; Al-Shaieb, 1985). However, others (Evans, 1928; Green, 1936; Donovan, 1974) found that the upper contact is unconformable or that both contacts are unconformable with 30 feet of relief with the Marlow Formation near Bridgeport, Oklahoma (Green, 1936).

The thickness of the Rush Springs Formation can vary depending on location. The USGS records the thickness as up to 300 feet. The OGS indicates a maximum thickness of 334 feet where there is a full section (Davis, 1950), a range of 200 feet in the south, and up to 330 feet to the north (Tanaka and Davis, 1963). A well log near Cordell indicates an approximate thickness of 350 feet (Green, 1936). Another source records the thickness (Upper Whitehorse Group) as 380 feet (Evans, 1928). A more recent analysis of a core shows 417 feet of Rush Springs Formation from a location near the axis of the Anadarko Basin (Poland, 2011). The OGS records the thickness as becoming greater westward along the axis of the Anadarko Basin (Tanaka and Davis, 1963) and indicates that the Rush Springs Formation thins to the north in the Eagle City area, eventually thinning to 90 feet in Kansas (Fay, 1962). New estimates of maximum thickness of the aquifer are discussed in the Hydrogeology section.

Gypsum is the most common cement within the Rush Spring Formation (Johnson and others, 1991), although other cements present include hematite, calcite, and dolomite (Suneson and Johnson, 1996). Thin-section analyses in the general locality of the Rush Springs Formation indicate that the unit is composed of 50 to 60 percent quartz, 8 to 12 percent orthoclase, 2 to 3 percent microcline and plagioclase, and less than 1 percent chert and other rock fragments (Allen, 1980). Additional thin-section analysis (Poland, 2011) confirms a high percentage of quartz in the Rush Springs, often with clay coating the grains. Samples from near the town of Cement showed a high degree of cementation, atypical for the Rush Springs Formation, which was caused by local alteration from oil and gas deposits below the formation (Allen, 1980; Kirkland and Rooney, 1995).

There are gypsum and dolomite beds within the Rush Springs Formation, most notably the Weatherford Gypsum Bed, which is a 30-foot layer of mainly carbonate and gypsum about 30 to 60 feet below the surface. Several other massive gypsum beds that are 2 to 5 feet thick are located below the Weatherford Bed in Dewey County and were named “Old Crow” at 30 feet below the Weatherford Bed and “One Horse” at 120 feet below the Weatherford Bed (Cragin, 1897). These beds are not continuous throughout the Rush Springs Formation. The thickness between the Weatherford Bed (called the Quartermaster Dolomite) and the younger Cloud Chief Formation decrease southward (Evans, 1928) with evidence suggesting that the Weatherford Bed grades out to the southeast (Green, 1936). Thin-section analysis of the Weatherford bed shows that it comprises as much as 40 percent carbonate (Poland, 2011). A section of outcrop identified in Section 35, Township 12N, Range 13W, located in northern Caddo County (Evans, 1928), was later determined to be Weatherford Bed with a variable thickness of a pinkish, conglomeritic, dolomitic bed containing geodes; about 5 feet of hard, light gray dolomite; and thinly laminated, reddish sandstone with somewhat irregular contacts with the underlying Whitehorse (Rush Springs Formation) and the Whitehorse Sandstone (Rush Springs Sandstone) (Moore and Snider, 1928).

A 1962 bulletin by the OGS identifies the Weatherford Bed as the top of the Rush Springs Formation (Fay, 1962). However, other studies (Hart, 1974; Carr and Bergman, 1976; Havens, 1977; Miller and Stanley, 2004; Stanley and Miller, 2005; Chang and Stanley, 2010) have identified strata above the Weatherford Bed and below the Moccasin Creek Bed of the Cloud Chief Formation as part of the Rush Springs Formation. The approximate 20 to 67 feet of strata between the Weatherford Bed and Moccasin Creek Gypsum Bed have been described by the OGS as silty shale (Fay, 1962).

Multiple theories regarding the depositional environment of the Permian-age Rush Springs Formation have been published. The historic view (Ham and others, 1957; O’Brien, 1963; MacLaughlin, 1967; Nelson, 1983; Al-Shaieb, 1985; and Johnson and others, 1991) indicates a shallow-marine or fluvial-deltaic environment based on the presence of eolian deposits with high porosity and permeability. The sandstone was thought to have been laid down along the eastern side of a shallow embayment that was occasionally restricted from the main Permian sea (to the west) as evidenced by the sandstone grading laterally into anhydrite and gypsum westward from Caddo County in what is interpreted as a desiccation basin (Tanaka and Davis, 1963).

A recent interpretation on Permian red beds in the southern midcontinent has challenged the shallow marine interpretation of the Permian-age Rush Springs Formation (Suneson and Johnson, 1996; Benison and others, 1998; Benison and Goldstein, 2002). One study interprets the Rush Springs Formation as having a terrestrial origin with fluvial and eolian influences with a facies assemblage ranging from eolian dune to interdune to extradune (Poland, 2011). The USGS identified drift sand deposits in the Rush Springs Formation that indicate eolian deposition (Becker and Runkle, 1998). This interpretation is based on the sedimentary structures indicative of eolian deposition (textures, surface hierarchy, paleocurrent data, and root casts) and a lack of any reported fossils within the Rush Springs Formation. Scanning electron microscopy also confirms grain surface textures characteristic of eolian transport and deposition (Poland, 2011). The prevalent direction of wind transport was to the south-southwest (Poland, 2011).

Eolian bedforms became larger and more organized through much of the time the Rush Springs Formation was being deposited until the deposition of the Weatherford Gypsum Bed in the upper portion of the Rush Springs Formation. Outcrops of fluvial deposits are occasionally present in the Rush Springs Formation, which suggest fluvial systems penetrated the Rush Springs dune system occasionally. Eolian conditions resumed after the deposition of the Weatherford Gypsum Bed but without the large-scale textures seen in the lower sections of the Rush Springs Formation (Poland, 2011).

The depositional system for the dolostone/gypsum Weatherford bed of the Rush Springs Formation has been interpreted as a restricted marine/saline lake with a rising water table and a reduced sand supply. Furthermore, the USGS identified recrystallized nodules in the Weatherford Bed that indicate a closed basin system with hypersaline conditions (Becker and Runkle, 1998). The source of the sediments in the Anadarko Basin likely came from multiple locations: from the Ozark Uplift and Ouachita Mountains (Fay, 1964; Suneson and Johnson, 1996); from the northwest, possibly from the ancestral Rocky Mountains (Davis, 1955); and from the south-southeast (Fay, 1962).

Cloud Chief Formation

The Permian-age Cloud Chief Formation, consisting of reddish-brown to orange-brown shale with interbedded sandstone and siltstone (Carr and Bergman, 1976), has been described as a widely distributed red bed unit in the central part of Oklahoma (Ham and Curtis, 1958). The USGS identified the maximum thickness of the Cloud Chief Formation in the study area as about 100 feet (Becker and Runkle, 1998). However, an earlier study found that the Cloud Chief can be as thick as 300 feet (Green, 1936). In the study area, much of the formation has been eroded off of the central and eastern portions of the aquifer. In areas where gypsum is near the surface, karst features, such as dissolution fissures, have been observed.

There are several gypsum layers in the Cloud Chief Formation, most notably the basal Moccasin Creek Gypsum Bed, which has also been called the Day Creek Dolomite (Fay, 1962). The Moccasin Creek Gypsum Bed is a triple gypsum sequence about 14 feet thick with shale and siltstone between the gypsum and dolomitic gypsum layers (Fay, 1965) and is the first of a series of desiccation periods during which the evaporites of the Cloud Chief Formation were deposited. The occasional presence of breccias in some areas within the clay and siltstones indicate deposition under turbulent conditions; rippled-marked, even-bedded, and fine-grained silty sandstones indicate less turbulent deposition. The Moccasin Creek Gypsum Bed is approximately 30 feet above the Weatherford Bed in the Rush Springs Formation.

Quaternary Deposits

Quaternary-age alluvium and terrace deposits lie unconformably on the Permian-age bedrock and range in age from Pleistocene to present time. They are described as wind-blown sand and stream-laid deposits of sand, silt, clay, gravel, and volcanic ash (Carr and Bergman, 1976). The alluvium and terrace deposits are considered one geologic unit in this report because they have similar hydrologic properties. They are considered to be in hydrologic connection with the Rush Springs Formation and are included as part of the same flow system as the Rush Springs Formation. Alluvium and terrace deposits in the study area are found in thicknesses of about 80 to 100 feet according to OWRB well driller logs.

The two largest stream systems with alluvium and terrace deposits in the study area are the Canadian River, flowing through the northern portion of the aquifer, and the Washita River, flowing through the southern portion. The alluvium and terrace deposits of the Canadian River and Washita River were a result of multiple cycles of deposition and erosion. The initial valleys were typically broad and were eroded into the bedrock. The sand and gravel deposited at the time were composed mostly of quartz and other siliceous rocks that were likely sourced in the Rocky Mountains or from the Tertiary deposits of the High Plains (Tanaka and Davis, 1963). Streams then degraded their channels and many older terrace deposits were transported away, allowing for the valleys to be refilled partly with material reworked from the older terrace deposits or with sand and silt sourced from the surrounding bedrock. Finally, valleys were cut into terrace deposits and partly filled with sand, silt, and clay, which comprise the alluvium of the Canadian and Washita Rivers (Tanaka and Davis, 1963). The Canadian River deposits are more deeply incised through the Rush Springs Formation while the Washita River deposits more directly overlie the Rush Springs Formation.

Characteristics of the Rush Springs Aquifer

Streamflow and Base Flow

Three large rivers flow over or adjacent to the Rush Springs aquifer: the Canadian, North Canadian, and Washita. All three rivers are impounded by surface water reservoirs at points along their flow path. Streamflow gauges maintained by the USGS on major rivers in the study area include the following: (Figure 1) Washita River at Anadarko (USGS 07326500), Washita River at Carnegie (USGS 07325500), Washita River near Clinton (USGS 07325000), Washita River near Foss (USGS 07324400), Canadian River near Bridgeport (USGS 07228500), North Canadian River near Seiling (USGS 07238000), North Canadian River at Canton (USGS 07239000), North Canadian River below Weavers Creek near Watonga (USGS 07239300), and North Canadian River near Calumet (USGS 07239450). Average annual streamflow of the 3 major rivers downstream of the aquifer for the common period of record (1984–2015) are 122,280 acre-feet (169 cubic feet per second) at the North Canadian River below Weavers Creek near Watonga (USGS 07239300), 236,523 acre-feet (327 cubic feet per second) at the Canadian River near Bridgeport (USGS 07228500), and 417,663 acre-feet (577 cubic feet per second) at the Washita River at Anadarko (USGS 07326500).

The Washita River has the highest average annual discharge of the 3 primary rivers. Groundwater discharges to perennial streams that drain into the Washita River. These include Barnitz Creek, Bear Creek, Beaver Creek, Cobb Creek, Sugar Creek, and Little Washita River. The confluence of the Washita River and Little Washita River is downstream of the Washita River streamflow gauge at Anadarko. Streamflow gauges maintained by the USGS that are located within the Washita River drainage basin in the study area include Cobb Creek near Eakly (USGS 07325800), Cobb Creek near Fort Cobb (USGS 07326000), Lake Creek near Sickles (USGS 07325840), Lake Creek near Eakly (USGS 07325850), Willow Creek near Albert (USGS 07325860), a historic streamflow gauge on Barnitz Creek near Arapaho (USGS 07324500), a historic streamflow gauge on Sugar Creek near Gracemont (07327000), and several on the Little Washita River that include Little Washita River above SCS Pond No. 26 near Cyril (USGS 073274406), Little Washita River near Cyril (USGS 07327442), and Little Washita River near Cement (USGS 07327447).

The Canadian and North Canadian Rivers do not have any active USGS streamflow gauges on tributaries in the study area. A historic site, Bent Creek near Seiling (USGS 07237800), is located in the North Canadian River drainage. In addition, a streamflow gauge on Deer Creek near Hydro (OWRB 520620060010-003RS) was installed in April 2013 as part of the study; water that flows through the site discharges to the Canadian River. This OWRB location corresponds to the historic USGS site on Deer Creek near Hydro (USGS 07228400).

The part of streamflow that is discharged from groundwater is referred to as base flow, defined for this report as the portion of streamflow that is not runoff. Base flow maintains streamflow in perennial streams within the study area. A base flow separation method was used to determine the volume of streamflow comprising base flow, which allows the streamflow hydrograph to be partitioned into either direct runoff or base flow. The base flow component of streamflow was computed using the BFI method, which analyzes the streamflow data from a gauge for days that fit a requirement of antecedent recession, designates base flow to be equal to streamflow on these days, and linearly interpolates the daily record of base flow for days that do not fit the requirement of antecedent recession (Rutledge, 1998).

Streamflow data from the Washita River near Foss (USGS 07324400), Washita River near Clinton (USGS 07325000), Washita River at Carnegie (USGS 07325500), and Washita River at Anadarko (USGS 07326500) (listed upstream to downstream) were analyzed and show a downstream increase in base flow discharged from the aquifer to the Washita River surface water basin. The common periods of record for the 4 streamflow gauges on the Washita River were 1964–1986 and 1990–2005. Foss Reservoir was actively storing water and regulating flow during the period analyzed. Water releases from the reservoir were obtained from the USBR and subtracted from the streamflow recorded from the gauges. Between the streamflow gauge near Foss and the streamflow gauge near Clinton, average base flow increased from 9.3 to 42.7 cubic feet per second during the common period of record (Table 3). From the streamflow gauge near Clinton to the streamflow gauge at Carnegie, the average base flow increased more than three times to 144.4 cubic feet per second, and from the streamflow gauge at Carnegie to the streamflow gauge at Anadarko, base flow increased 50.9 cubic feet per second to 195.3 cubic feet per second. The increase in average base flow between the streamflow gauges near Clinton and at Carnegie (107.0 cubic feet per second) indicates that the Washita River gains a

Table 3. Streamflows and base flows at streamflow gauging stations in the vicinity of the Rush Springs aquifer summarized through 2015.

Station number	Station name	Drainage area, in square miles	Period of analysis	Mean annual streamflow, in cubic feet per second	Median annual streamflow, in cubic feet per second	Mean annual base flow, in cubic feet per second	Median annual base flow, in cubic feet per second
07324500	Barnitz Creek near Arapaho, Okla.	243	1946-1963	14.4	0	1.9	0
07237800	Bent Creek near Seiling, Okla.	139	1967-1970	7.6	2.2	1.8	1.4
07325800	Cobb Creek near Eakly, Okla.	132	1968-2015	28.8	15	14.1	12.3
07228400*	Deer Creek at Hydro, Okla.	274	1961-1962, 1978-1979, 2014-2015	30.50	21.20	16.80	17.20
07325850	Lake Creek near Eakly, Okla.	52.5	1969-1978, 2005-2015	8.1	3.5	3	2.5
07327550	Little Washita East of Ninnekah, Okla.	232	1992-2015	52.20	24.00	24.70	16.20
07327000	Sugar Creek near Gracemont, Okla.	208	1956-1974	14.70	5.30	4.50	2.10
07325860	Willow Creek near Albert, Okla.	28.2	1970-1978, 2005-2015	4.10	1.90	1.60	1.40
07324400	Washita River near Foss, Okla.	1526	1956-1958, 1961-1987, 1989-2015	53.8	7.4	22.3	6.0
07325000	Washita River near Clinton, Okla.	1961	1935-2015	124.3	29.0	52.4	22.0
07325500	Washita River at Carnegie, Okla.	3116	1937-2006	361.5	116.0	148.4	84.1
07326500	Washita River at Anadarko, Okla.	3640	1903-1908, 1935-1937, 1963-2015	484.1	182.0	236.4	142.0
Washita River gauges common period of record							
07324400	Washita River near Foss, Okla.	1526	1964-1986, 1990-2005	22.5	12.4	9.3	4.7
07325000	Washita River near Clinton, Okla.	1961	1964-1986, 1990-2005	81.5	40.9	42.7	25.1
07325500	Washita River at Carnegie, Okla.	3116	1964-1986, 1990-2005	336.3	115.3	144.4	87.3
07326500	Washita River at Anadarko, Okla.	3640	1964-1986, 1990-2005	403.8	156.8	195.3	122.1

*OWRB stream gauging station number 520620060010-003RS significant amount of base flow between these streamflow gauges (approximately 48 percent of the base flow measured at Anadarko). Streams that drain into the Washita River between these two streamflow gauges include Bear, Boggy, Cavalry, Cedar, Gokey, Gyp, and Spring Creeks. Table 3 shows average annual and median stream flow and average base flow (estimated using BFI) from the streamflow gauges on the Washita River for the common periods of record.

Cobb Creek is a major tributary that drains into the Washita River between the Carnegie and Anadarko streamflow gauges. Flow contributions from Cobb Creek would be expected to significantly increase the total flow of the Washita River; however, flow from Cobb Creek is influenced by Fort Cobb Reservoir, about 7 miles upstream of the confluence of Cobb Creek and the Washita River, making an accurate assessment of the influence of Cobb Creek under natural conditions on the Washita River difficult. Average stream flow for the period of record (1939–2015) for the Cobb Creek near Fort Cobb (USGS 07326000) gauging station, which is downstream from the reservoir, was 37.2 cubic feet per second; however, for the common period of record of the streamflow gauges on the Washita River (1964–1986 and 1990–2005), average flow was 33.4 cubic feet per second. For the common period of record, stream-flow discharge to the Washita River from the Cobb Creek near Fort Cobb gauging station accounted for 49 percent of the stream flow increase between Carnegie and Anadarko gauging stations on the Washita River. The additional 7 miles of Cobb Creek between the gauging station and the Washita would provide an additional, but unknown, amount of stream water flow to the Washita River.

Annual base-flow volume and BFI were estimated for 8 streamflow gauge sites in the study area (Table 3): Barnitz Creek near Arapaho (USGS 07324500), Bent Creek near Seiling (USGS 07237800), Cobb Creek near Eakly (USGS 07325800), Lake Creek near Eakly (USGS 07325850), Little Washita River east of Ninnekah (USGS 07327550), Sugar Creek near Gracemont (USGS 07327000), Willow Creek near Albert (USGS 07325860), and Deer Creek near Hydro (USGS 07228400 and OWRB 520620060010-003RS). Annual base flow at the Cobb Creek near Eakly streamflow gauge (period of record 1968–2015) ranged between 3,499 acre-feet in 1972 and 21,002 in 2007, and the base-flow index was estimated to be 53 percent base flow, with a low of 27 percent in 1986, a wet year with over 39 inches of rain, and a high of 92 percent in 1984, a dry year with about 20 inches of rain (Figure 5). Base flow in Little Washita east of Ninnekah (period of record 1993–2015) ranged between 3,564 acre-feet in 2012 and 50,429 acre-feet in 1993 with a base-flow index of 49 percent between 1993 and 2013 and a range between 29 percent in 2013 and 67 percent in 2001 (Figure 6). Base flow in Lake Creek near Eakly for the periods of record (1969–1978 and 2005–2015) varied from 435 acre-feet in 1972 to 6,143 acre-feet in 2008 with a base-flow index of 42 percent, ranging from 12 percent in 1977 to 82 percent in 2010 (Figure 7). Base flow in Willow Creek near Albert for the periods of 1970–1978 and 2005–2015 ranged from 507 acre-feet in 1971 to 2,246 acre-feet in 2008 with a mean base-flow index of 46 percent, ranging from 18 percent in 1975 to 76 percent in 2006 (Figure 8). Base flow in Deer Creek near Hydro for the periods of 1960–1963, 1977–1980, and 2013–2015 ranged from 11,041 acre-feet in 2014 to 18,799 acre-feet in 1962 with a mean base-flow index of 45 percent, ranging from 25 percent in 1961 to 76 percent in 2014 (Figure 9). Base flow in Barnitz Creek near Arapaho ranged from 29 acre-feet in 1955 to 6,411 acre-feet in 1960 with a mean base-flow index of 13 percent, ranging from 0 percent in 1956, when the stream was dry for most of the year, to 52 percent in 1960. For the Bent Creek near Seiling streamflow gauge period of record (1967–1970) base flow averaged 1,302 acre-feet per year with a base-flow index of 23 percent over the period of record.

The period of record for the USGS/OWRB Deer Creek streamflow gauge near Hydro is October 1960 through December 1963, December 1977 through September 1980, and April 2013 through December 2015. Mean base flow from the gauge from April 5, 2013 through December 31, 2015 was 16.8 cubic feet per second. This accounted for approximately

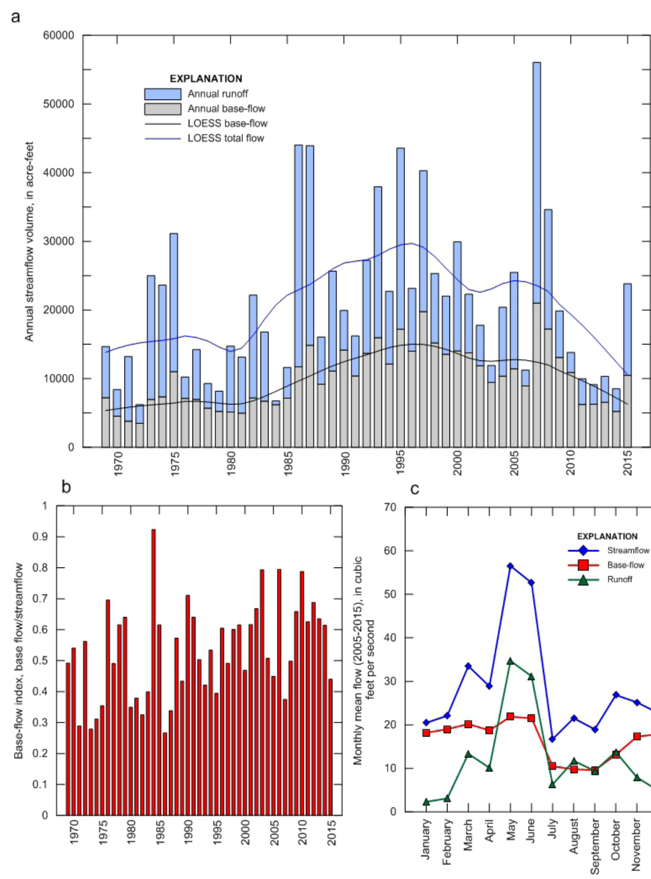

Figure 5. (A) Annual base flow and total flow volume with LOESS trend line; (B) base-flow index; and (C) monthly mean streamflow, base flow, and runoff for the USGS Cobb Creek streamflow gauge near Eakly (USGS 07325800), 1969–2015.

19 percent of the mean flow measured at the Bridgeport streamflow gauge on the Canadian River for the same period, indicating that the Rush Springs aquifer contributes a significant portion of flow to the Canadian River.

Precipitation Trends in Base Flow

LOESS (LOcally Estimated Scatterplot Smoothing) trend lines were incorporated in the study to show trends in base flow. LOESS is a nonparametric regression procedure that reduces the influence of outliers and displays a smooth trend line for the entire range of data (Cleveland and Devlin, 1988; Helsel and Hirsch, 2002). The LOESS trend line is derived from a LOESS regression (Helsel and Hirsch, 2002) and was created using a Microsoft Excel add-in application, LOESS Utility (Peltier Tech, 2009). The LOESS lines were used for trend visualization purposes only and were not used to determine the statistical significance of trends. LOESS plots were developed on an annual basis for base-flow volume, total-flow volume, and base-flow index.

Of the four streamflow gauges, only Cobb Creek near Eakly (07325800) and Little Washita River east of Ninnekah (07327550) had the periods of record to properly visualize trends. The base flow trend for Cobb Creek at Eakly shows an increase in base flow from the mid-1980s through the early 2000s (Figure 5). Base-flow data from the Little Washita River east of Ninnekah show the same base-flow trend beginning in 1993, the first full year in the period of record (Figure 6). The years 2007 and 2008 had higher base flow before decreases in base flow during 2010–2015. This base-flow trend coincides with the increase in precipitation observed over the same time period (see Climate section), which demonstrates the importance of precipitation and recharge to the flow of streams discharging the aquifer.

Water-level Fluctuations

Water-level observations can provide insight into aquifer response to stresses, including climate variations and groundwater pumping and recovery. Long-term periodic water-level observations provide information that can be used

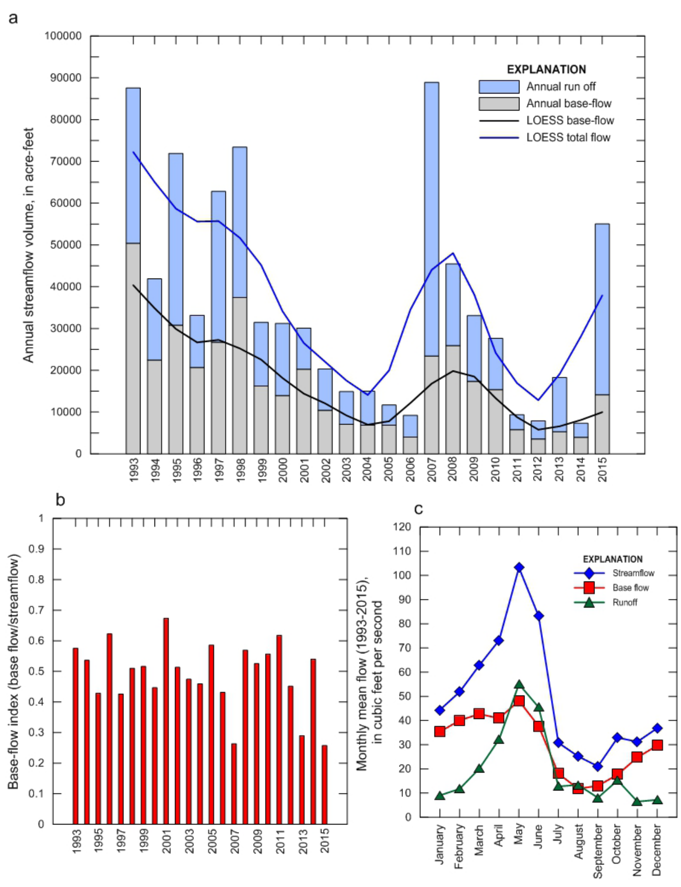

Figure 6. (A) Annual base flow and total flow volume with LOESS trend line; (B) base-flow index; and (C) monthly mean streamflow, base flow, and runoff for the USGS Little Washita River streamflow gauge east of Ninnekah (USGS 07327550), 1993–2015.

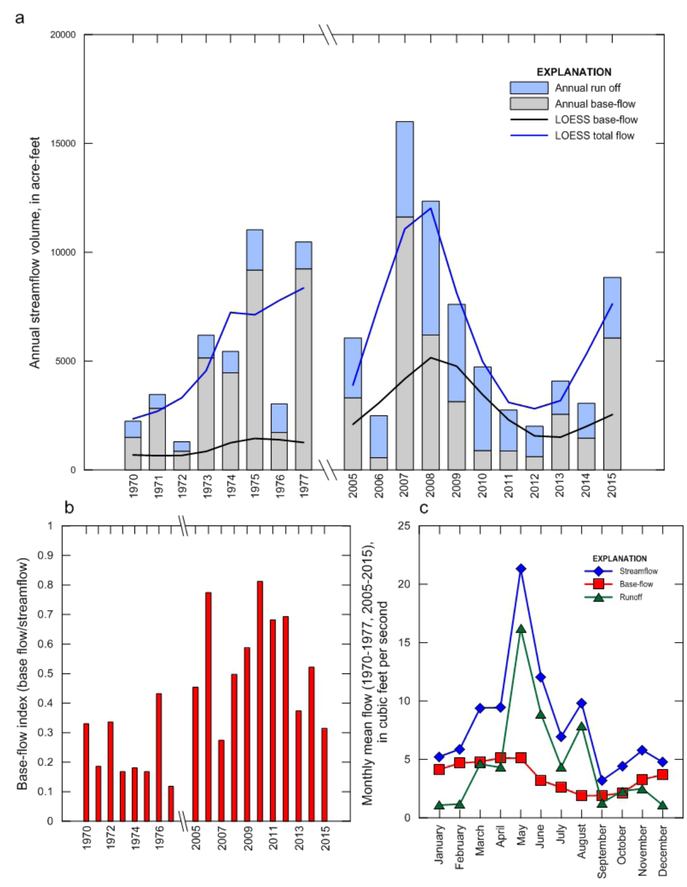

Figure 7. (A) Annual base flow and total flow volume with LOESS trend line; (B) base-flow index; and (C) monthly mean streamflow, base flow, and runoff for the USGS Lake Creek streamflow gauge near Eakly (USGS 7325850), 2005–2015.

to assess regional groundwater supply or calibrate groundwater-flow models. Continuous water-level observations, taken once per hour using a transducer installed in a well, can show aquifer response to climate variation and groundwater use on shorter time scales to help develop an understanding of recharge events, seasonal pumping demands, and interactions between surface water and groundwater. An aquifer's response to precipitation events or droughts can help characterize an aquifer as confined or unconfined at different locations and depths.

Historic

Long-term annual water-level measurements have been collected across the state by the OWRB since the 1950s. These data are also stored in the USGS National Water Information System (NWIS) database using unique USGS site numbers. Historic depth to water measurements are presented in the Appendix. Wells in this study were categorized based on trends of overall increasing water levels, overall decreasing water levels, water levels fluctuating along with changes in climate patterns, and indiscernible patterns. There were 139 wells with historic groundwater-level observations in the study area through 2016; 25 wells had 35 years or more of data and 70 wells had a period of record between 12 and 35 years. Some wells had been discontinued for various reasons while others had data gaps. For water level analysis, researchers looked at wells with a minimum of 12 years of data. Water-level graphs with fewer than 12 years of data often did not cover enough time to assess long-term trends.

Water-level data were normalized by calculating the Z-score (Standard Test) for all water levels. For water-level data at a specific site, the Z-score is a statistical measurement that compares the data to the mean; it is calculated by subtracting water levels from the mean and dividing by the standard deviation. A Z-score of 0 is equivalent to the mean water level in a well over the period of record. A positive Z-score indicates depth to water increased compared to the mean and water levels decreased, while a negative Z-score indicates depth to water decreased compared to the mean and water levels increased. Normalization provides a simple method to graphically compare water-level trends of many wells simultaneously.

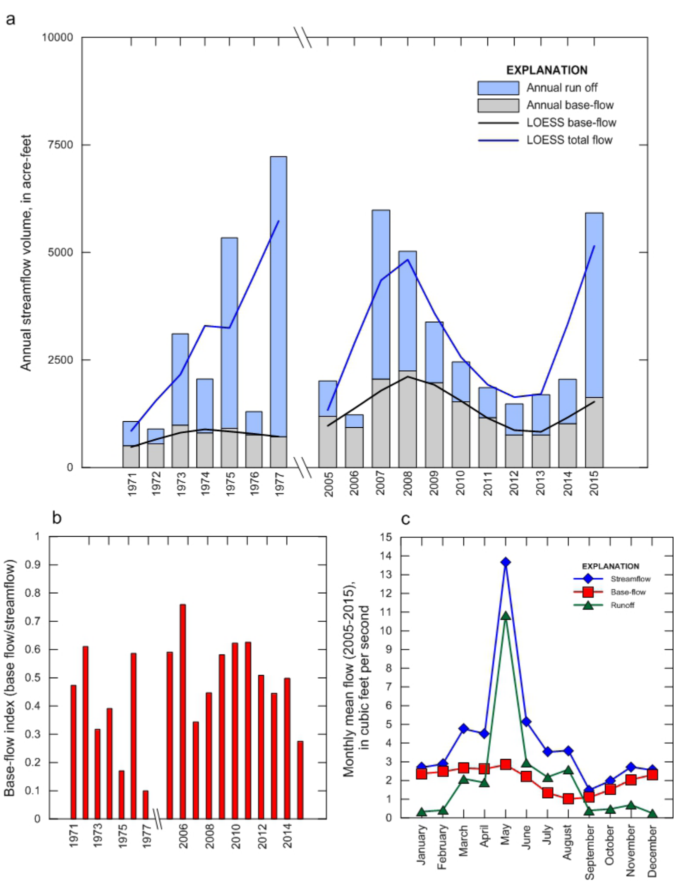

Figure 8. (A) Annual base flow and total flow volume with LOESS trend line; (B) base-flow index; and (C) monthly mean streamflow, base flow, and runoff for the USGS Willow Creek streamflow gauge near Albert (USGS 7325860), 2005–2015.

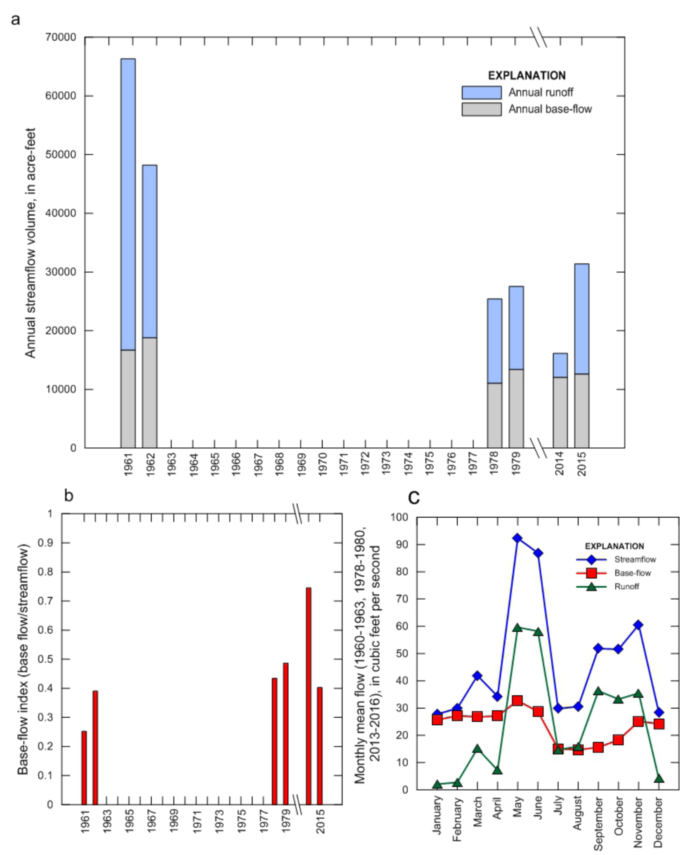

Figure 9. (A) Annual base flow and total flow volume with LOESS trend line; (B) base-flow index; and (C) monthly mean streamflow, base flow, and runoff for the USGS and OWRB Deer Creek streamflow gauge near Hydro (USGS 07228400 and OWRB 520620060010-003RS), 1960–1963, 1977–1980, and 2013–2015.

There were 9 wells with indiscernible water levels during the period of record characterized by intermittent increases and decreases in water levels, making the long-term changes difficult to assess.

Water levels in 54 wells were determined to primarily fluctuate along with climate. Water levels in these wells were above normal for the historical wet period in Oklahoma (mid-1980s through the early 2000s) and below normal during the historical dry period (1970s and 2010–2015). The wells were located predominately in the unconfined portion of the aquifer, which is exposed to atmospheric changes in pressure, temperature, and precipitation. Wells with water levels fluctuating with climate patterns are defined in this analysis as having rising groundwater levels during wet periods and declining levels during dry periods (see Climate section); declining and increasing trends are independent of climate variability. Groundwater wells with water levels fluctuating on a climate pattern. These wells are assumed to be outside the cone of depression for any nearby pumping wells or in an area where the effects of localized pumping are not noticeable and were located predominately in the unconfined portion of the aquifer, which is exposed to atmospheric changes in pressure, temperature, and precipitation. Figure 10 shows normalized water levels for selected wells identified with a climate trend in the Rush Springs aquifer. The solid black line is an average of all water-level Z-scores from wells showing a climate trend.

Groundwater wells showed overall increasing water levels at 15 sites and overall decreasing water levels at 17 sites. Most of the groundwater wells showing either increasing or decreasing water levels are located in the central portion of the aquifer where groundwater use is the highest. Both of these trends are thought to result from either increased local pumping (causing water levels to decline) or reduction in local pumping (causing lower water levels to show recovery).

As part of this investigation, 15 of the 143 groundwater wells measured by the USGS during 1986–1991 (Becker and Runkle, 1998) were measured again in 2013 (see 2013 Potentiometric Surface section). Water-level declines were observed at 12 well sites and water-level increases were observed at 3 well sites. Water-level change from the 1986–1991 period to 2013 ranged from a decline of 56.6 feet to an increase of 54.44 feet. The mean water level in the wells decreased by 11.0 feet and the median water level decreased by 16.6 feet. The majority of these wells are located in the heavily irrigated areas of Caddo County and western Washita County. The large water-level decline may be explained by the 1986–1991 measurements occurring during a wet period (see Climate section), which would have resulted in increased recharge to the aquifer and lower than normal irrigation, while the 2013 measurements were taken during a multi-year drought and significantly increased irrigation to compensate for the lack of rain.

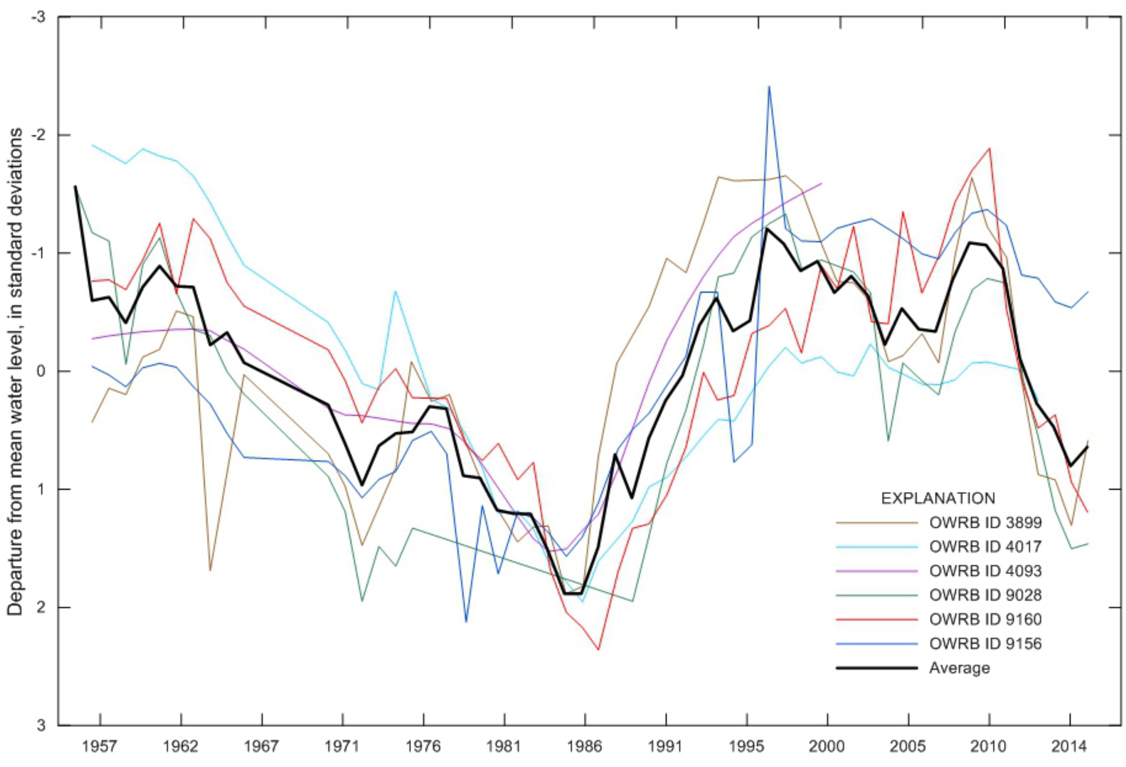

Figure 10. Graph showing normalized water levels for wells with a climate trend in the Rush Springs aquifer. A Z-score of 0 is equivalent to the mean water level in a well over the period of record. A positive Z-score indicates a decreasing water level and higher depth to water reading, and a negative Z-score indicates an increasing water level and lower depth to water reading.

Continuous

Continuous water levels were monitored at 15 groundwater well sites as part of this study to observe seasonal trends and regional stresses. Depth to water measurements were logged once per hour (Table 4). Three of the continuous sites are located at Oklahoma Climatological Survey (OCS) Mesonet stations, two of which were installed during the study. Groundwater wells are located at the Acme Mesonet site (OWRB 89283), the Fort Cobb Mesonet site (OWRB 157457), and the Weatherford Mesonet site (OWRB 156516).

Three of the continuous well sites (OWRB 137452, 142112, and 150482) were within 1 mile of irrigation wells, and hydrographs show continual or seasonal pumping signatures (Figure 11 D). The other wells (OWRB 27650, 20024, 140033, 141465, 142042, 142324, 144003, 145203, 147385, and 156516) showed possible seasonal water-level change related to precipitation patterns (Figure 11 A-C, E, F). One well (OWRB 157457) showed a pumping signature superimposed on a possible seasonal precipitation pattern. From the beginning of the study (January 2012) through the spring of 2015, the study area received below average precipitation, which is reflected in continual declines in water levels. Above average precipitation recorded in May and November 2015 likely accounts for water-level increases in the wells shown on Figure 12.

Figure 12 shows water-level data from the Acme Mesonet well (OWRB 89283), which has the longest active period of continuous record in the study area. High precipitation in 2007 is reflected by close to a 4.7 foot increase in water levels from June 2007 to January 2008. Between 2010 and 2014, the site received very little recharge due to the intensification of drought conditions across the state. The lowest groundwater levels at the site over the period of record occurred in 2014. Water levels showed recovery in 2015, when above average precipitation was recorded.

The USGS has monitored groundwater levels continuously at 5 sites in the Rush Springs aquifer since 2010. Figure 13 shows groundwater levels at 4 sites from October 2010 through the December 2015. Water-level decline is observed from late spring through early fall. Water levels stabilize during the winter months, presumably when evapotranspiration and pumping from the aquifer is at a minimum. In 2015, each site shows increasing water levels, likely caused by above average precipitation, which could be indicative of groundwater recharge.

Several of the USGS wells had manual measurements during the historic period of record. Measurements at USGS well 351308098341601 had the longest period of record in the study area and showed a decline of 37.52 feet from September 1948 to April 2015 (Figure 14).

Table 4. Groundwater well sites with continuous water-level recorders in the Rush Springs aquifer.

OWRB Well ID	Latitude	Longitude	Total well depth, in feet below land surface	Period of analysis
157457*	35.14886	-98.46619	205	2/3/2014 - present
156516*	35.50820	-98.77516	300	1/31/2014 - present
20024*	35.98978	-98.89006	270	3/27/2012 - present
137452*	35.55685	-98.56126	230	3/27/2012 - present
141465*	34.89508	-98.18182	92	4/11/2012 - present
142042*	35.24026	-98.86911	178	4/27/2012 - present
142112*	35.68481	-98.72428	290	4/20/2012 - present
142316	35.68559	-99.19504	376	4/27/2012 - 7/20/2012
142324*	35.67933	-99.23067	315	7/20/2012 - present
144003*	35.89506	-98.75095	184	7/20/2012 - present
151636	35.36647	-98.24883	170	12/13/2012 - 5/12/2015
27650	35.37765	-98.73433	238.5	6/11/2012 - 5/12/2015
140033	35.82682	-99.07784	272	12/3/2013 - 5/12/2015
145203*	36.04494	-99.15350	150	12/03/2013 - present
147385	35.36761	-98.24791	230	12/13/2012 - 11/12/2015
150482	35.17468	-98.57625	274	05/16/2013 - 09/04/2014
89283*	34.80833	-98.02318	50	10/9/2013 - present

*recording at time of publication

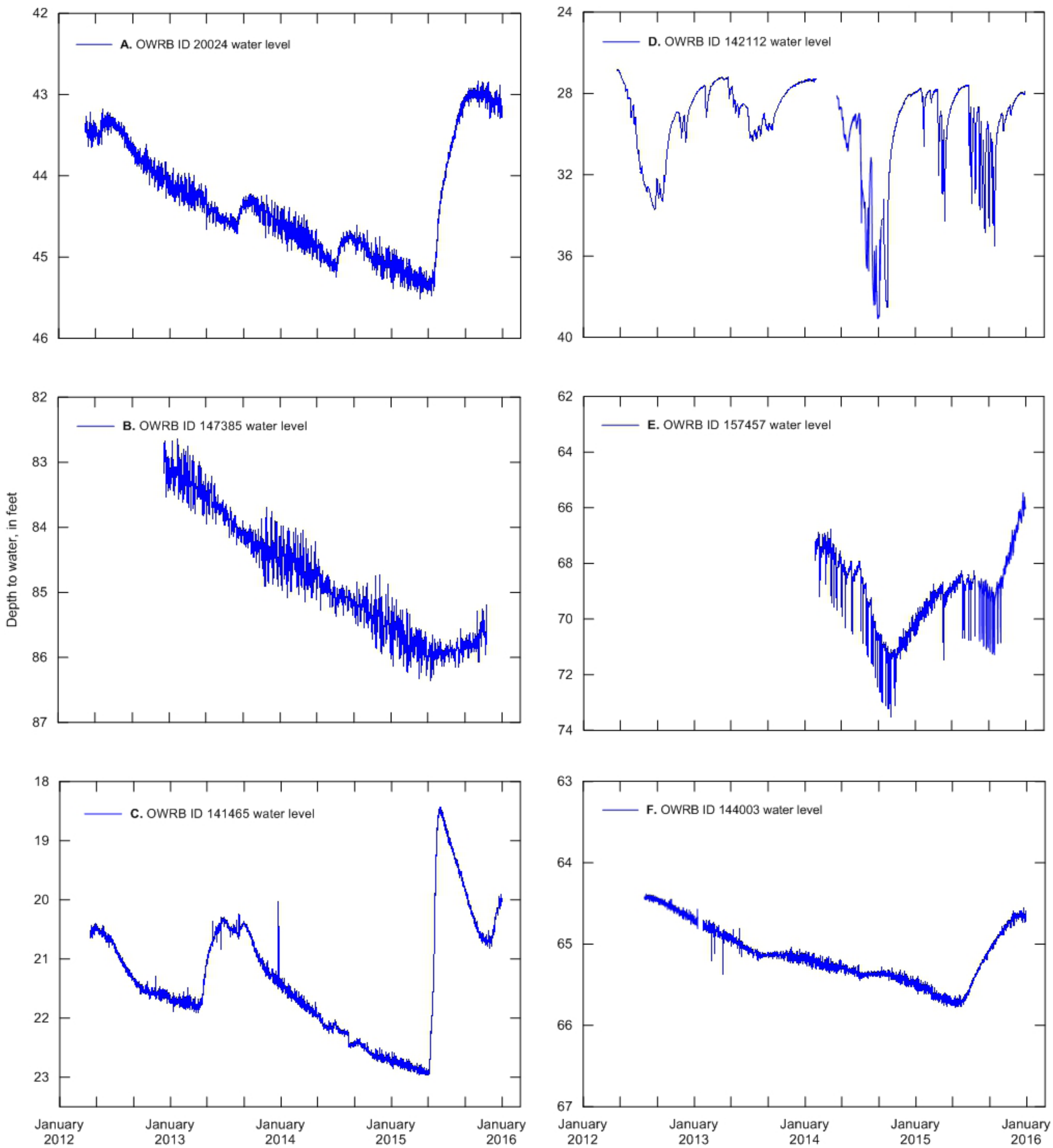

Figure 11. Water levels from OWRB continuous recorder wells from January 2012 through December 2015 showing possible responses from long-term precipitation response (A-C, F), localized groundwater pumping (D), or both (E) in the Rush Springs aquifer area.

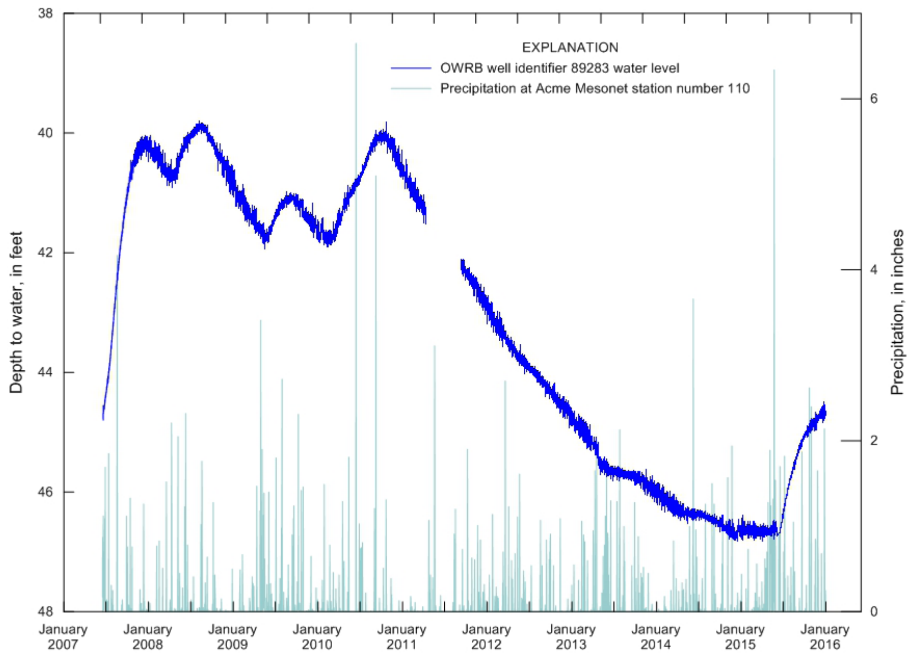

Figure 12. Long-term continuous water levels with precipitation for the Acme Mesonet well (OWRB 89283).

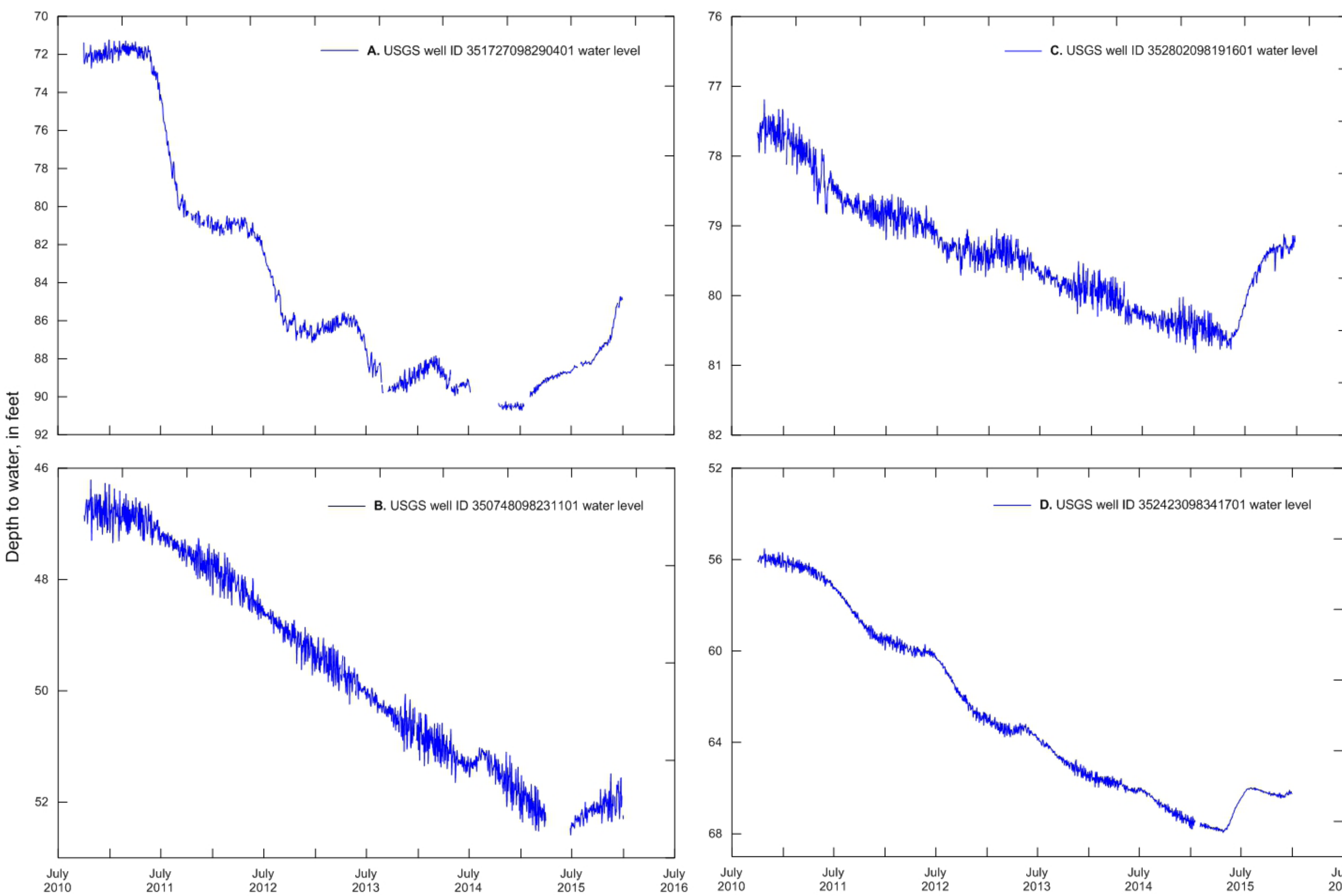

Figure 13. Groundwater levels measured in USGS wells 350748098231101, 351727098290401, 352423098341701, and 352802098191601 from October 2010 through December 2015.

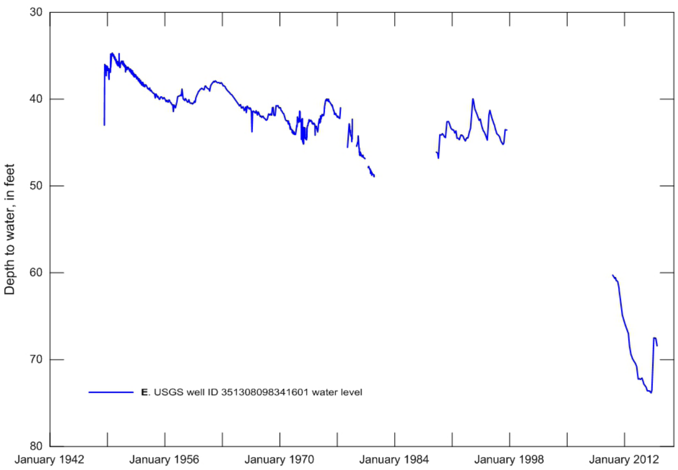

Figure 14. Groundwater levels measured in USGS well 351308098341601 from 1948 to April 2015.

Regional Groundwater Flow

Groundwater in the Rush Springs aquifer is under unconfined conditions in the eastern portion of the aquifer and confined conditions in the western portion where the Cloud Chief Formation overlaps the aquifer. In the unconfined area, groundwater flows toward streams that incise the bedrock to form perennial streams and will typically show water-level contours bending upstream. Where major rivers incise the bedrock, groundwater discharges directly to the alluvium aquifer systems. In confined portions of the aquifer, groundwater flow is generally slower because there is less recharge and there are few mechanisms to cause groundwater to drain. Regional groundwater flow is generally to the southeast, but locally groundwater flows toward the edges of the aquifer and toward streams and rivers. Age dating of groundwater has not been performed; however, groundwater in the confined portion of the aquifer is thought to be older than groundwater in the unconfined area.

In 1998, a potentiometric surface map was constructed for part of the Rush Springs aquifer using water levels measured in 143 wells from 1986–1991 (Becker and Runkle, 1998). However, data used to produce that map were limited in spatial extent, especially in the northern and western portions of the study area, where only 12 wells were measured north of the city of Weatherford. An older potentiometric surface map (Roles, 1976) and water-level measurements in the USGS NWIS database from 1905 provide data as well, but these are also concentrated in the more developed portions of the aquifer.

2013 Potentiometric Surface

The potentiometric surface of an aquifer is an estimated imaginary surface that reflects geographic variation in the fluid potential of the formation water in an aquifer. The potentiometric surface elevation at any point reflects the estimated height to which a column of water will rise in a cased well. A potentiometric surface map is constructed by contouring static water-level measurements in wells and can be used to determine the direction of groundwater flow. The potentiometric surface in an unconfined aquifer is the water table that is defined by the upper limit of the zone of saturation.

To create a potentiometric surface map for the Rush Springs aquifer, water levels were measured in 263 wells as part of this study between March 5, 2013, and March 15, 2013. The 2013 water levels, depth to water in feet below land surface, ranged from 0.0 to 182.9 with a median of 50.0 feet. The potentiometric surface elevation was estimated by subtracting depth-to-water measurements from land-surface altitude (Figure 15). The land-surface altitude of each well location was determined by using a differentially corrected Global Positioning System (GPS) receiver with a horizontal accuracy of 10 centimeters (3.9 inches) and a vertical accuracy of 15 to 50 centimeters (5.9 to 19.7 inches) and referenced to the North American Vertical Datum of 1983 (NAD 83). Elevation control points were also included at Fort Cobb Reservoir and USGS stream gauging stations on the Canadian and Washita Rivers within the study area. The potentiometric surface contours were generated in a geographic information system (GIS) and were adjusted manually to conform to basic topographic rules, especially within the Canadian and Washita River valleys. To further refine areas with poor well coverage, 74 groundwater levels in the Canadian River Valley in the north and northwestern portion of the aquifer that were measured during February and March 2013 by the USGS were used to interpolate the local potentiometric surface (Ellis and others, 2016). Four water levels collected during slug tests in 2014 were also used to check areas with no other water-level data within 5 miles.

Groundwater in the Rush Springs generally flows to the southeast with the Canadian and Washita rivers acting to drain the groundwater-flow system. Contours steeply bend upstream along both the Canadian River and Washita River, indicating that groundwater is discharging to the alluvial groundwater system. The increased base flow in the Canadian River is reported as attributable to groundwater discharged from the Rush Springs aquifer (Ellis and others, 2016).

Groundwater contours tend to bend around major streams, such as Cobb, Deer, Lake, Willow, Barnitz, and Sugar Creeks and the Little Washita River, indicating that groundwater is discharging in these areas. Field observations and streamflow gauge data for these streams confirmed that they were perennial streams with high base flow indices. Some smaller streams in the Rush Springs aquifer are intermittent and had no flow during the synoptic well measurement during the study as indicated in Figure 15 with contours cutting straight across.

Potentiometric surface contours are mostly parallel with the geologic boundary (eastern erosional boundary)

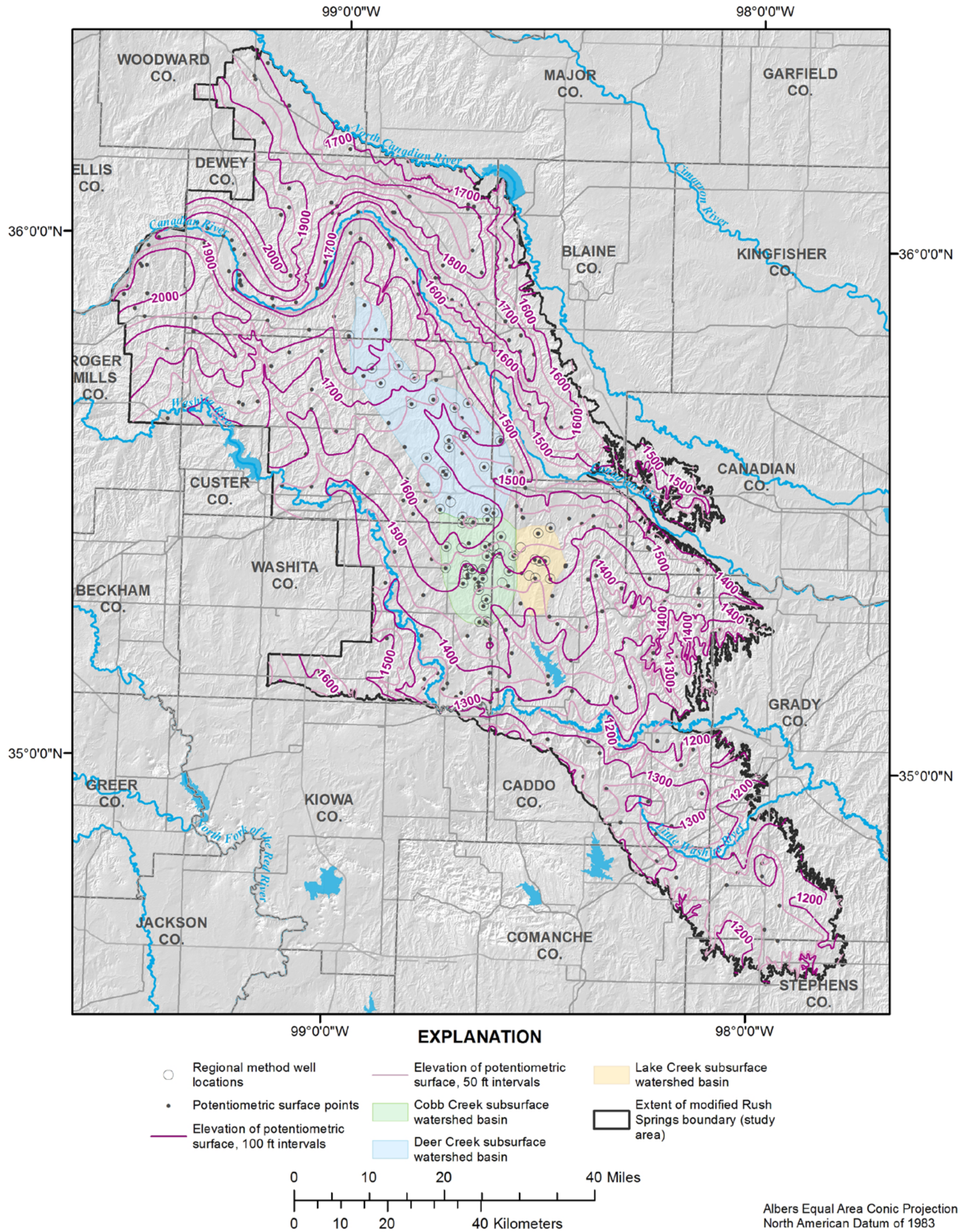

Figure 15. Potentiometric surface contour map of the Rush Springs aquifer, 2013.

of the Rush Springs Formation and underlying units with groundwater flowing toward the boundary from within the Rush Springs aquifer. Along the western edge of the aquifer, where the aquifer boundary is cut based on water use and water quality, rather than geologic contact, the potentiometric surface contours are mostly perpendicular to the boundary, indicating that there is little inflow from the west.

Potentiometric Surface Changes

The 2013 potentiometric surface (Figure 15) shows several differences from the 1986–1991 potentiometric surface (Becker and Runkle, 1998). The 2013 potentiometric surface displays fewer contour bends upstream in the minor streams draining the Rush Springs aquifer. This may be due to a denser data set in the 2013 map or to the fact that the 2013 measurements were taken during a drought, during which it would be expected that some streams lose base flow as groundwater levels decline. Many of the streams that did not show contour lines bent upstream were observed during the synoptic water-level measurement and found to be dry, indicating they were not connected to the aquifer at the time of measurement.

Eleven wells from the 2013 dataset had depth-to-water measurements that were less than 10 feet from the land surface, indicating some wells may flow under artesian pressure during periods of higher recharge. Three wells were located on a potentiometric high in Custer County, and two wells were located upstream from Fort Cobb Reservoir. Five of the wells were located in the southeast portion of the aquifer, which is a groundwater discharge area where the aquifer is thinner and the hydraulic gradient is low. This creates slower moving groundwater that is forced into a smaller aquifer volume, resulting in water levels closer to the surface compared to most wells in the central and northern portions of the aquifer. Several artesian wells were observed during the study, including one in the town of Marlow (OWRB 9427) and another (OWRB 2662) in Roger Mills County.

To create the 2013 potentiometric surface map, researchers utilized data from 19 groundwater wells that had previously been measured for the 1986–1991 potentiometric surface map created by the USGS. Of those wells, 15 had a water level decline and 4 had a water level increase. The mean water level change in these wells was a decline of 12.7 feet.

Groundwater Use

Groundwater use in the Rush Springs aquifer was documented as early as 1905 in a statewide assessment of groundwater resources (Gould, 1905), which identified the sandstone in Caddo County and surrounding areas as a major source of water. By the 1950s, there was an increasing reliance on groundwater in the study area, as described in an evaluation of the resources around the Weatherford area (Allen, 1953). Groundwater withdrawals from wells in Grady and Stephens Counties were estimated by OGS to be about 600 acre-feet per year by 1955 (Davis, 1955). However, this accounts for only a small portion of the study area. Groundwater use in the Caddo County area was estimated by OGS to be about 25,700 acre-feet per year from 1956–1959 (Tanaka and Davis, 1963). The OWRB provided a more comprehensive report in 1966 that showed groundwater use was about 28,000 acre-feet per year for irrigation, public water supply, industrial, domestic, and stock use. The number of groundwater wells in the study area increased from 12 in 1951 to more than 600 by 1964 (Oklahoma Water Resources Board, 1966). The OWRB began requiring groundwater users to submit annual water-use reports in the 1960s, and by 1967, a reasonable, annual estimation of groundwater use could be made.

Long-term Permitted Groundwater Use

Permit holders submit water use reports to the OWRB annually for long-term groundwater permits. There are 2,351 long-term temporary and prior right permits for groundwater use within the study area; the oldest prior-right permit dates back to 1929. Groundwater use from annual water-use reports typically indicate the type of beneficial use, including public water supply; irrigation; industrial; power; mining; and fish, recreation, and wildlife. The term “public water supply” is used to describe groundwater use by municipalities, rural water districts, housing additions, trailer parks, churches, and schools. For the study area, groundwater-use data were reviewed for outliers and inconsistencies to ensure accuracy. Some of the groundwater use along the southern boundary of the aquifer, near the town of Apache, Oklahoma, and southeastward through the towns of Fletcher, Sterling, and Marlow, Oklahoma, is from deep wells that penetrate through the thin portions of the aquifer and the underlying Dog Creek Shale into the El Reno Minor aquifer. Reported well yields in this area can be as high as 800 gallons per minute with well depths ranging from 250 to 973 feet. Water use from this area was excluded from the Rush Springs aquifer analysis.

Reported groundwater use from the Rush Springs aquifer for 1967–2015, shown in Figure 16, averaged about 69,900 acre-feet per year with a median of 62,154 acre-feet per year. The highest total reported annual groundwater use—about 115,016 acre-feet in 2014 and 133,113 acre-feet in 2015—correspond to drought conditions during these years. In 1992, only 37,210 acre-feet was reported, which was the lowest reported use for a single year; however, there is reason to believe data for the year 1992 may be incomplete. The second lowest total use for a single year occurred in 2007 at 40,418 acre-feet.

Four trends in reported use were identified by researchers: 1967–1980, 1981–1997, 1998–2009, and 2010–2015. For 1967–1980, reported use was relatively high, averaging 76,544 acre-feet per year, which may be attributable to below average precipitation during the period. Additionally, prior to the 1980s, users were required to report the number of acres

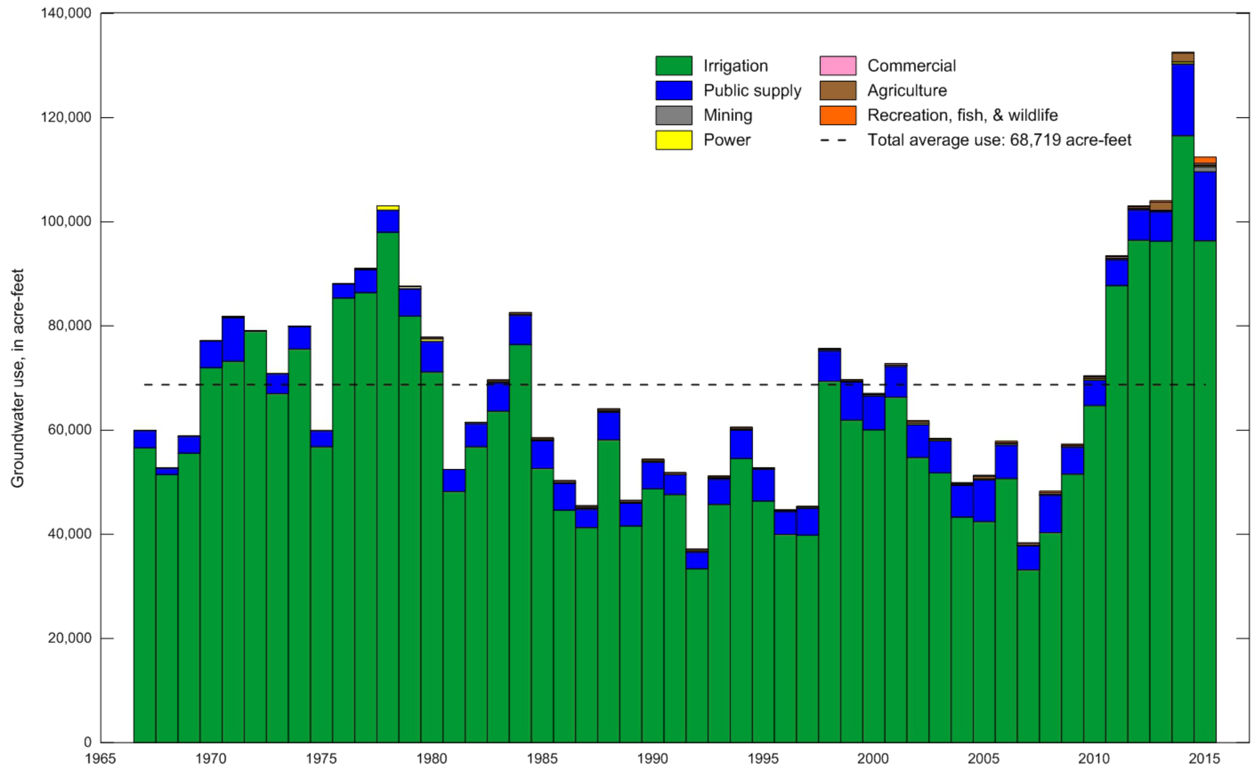

Figure 16. Graph showing annual reported groundwater use for the study area from 1967–2015.

irrigated and number of times irrigation occurred, but not the amount of water applied to the land; this means researchers had to make assumptions regarding the amount of water used. Groundwater use from the aquifer reached a peak in 1978 before declining throughout the 1980s and 1990s. The 1978 peak coincides with a period of below average precipitation in the late 1970s and early 1980s. During 1981–1997, the area received above average annual precipitation and annual groundwater use decreased to a mean of 55,875 acre-feet per year. The mean annual groundwater use increased to 59,247 acre-feet per year during 1998–2009. Within that time period, groundwater use averaged 73,349 acre-feet per year during 1998–2001. After 2001, groundwater use began to steadily decrease until a record low of 40,472 acre-feet was reported in 2007, which is likely attributable to record high precipitation. After 2007, groundwater use began to increase as the state reentered drought conditions (Oklahoma Climatological Survey, 2013). For 2010–2015, annual groundwater use increased to a mean of 103,656 acre-feet per year, with record high reported use in 2014 and 2015. Table 5 shows summary statistics of reported groundwater use in the Rush Springs since 1967.

During 1967–2015, 91.0 percent of reported groundwater use in the study area was for irrigation, 7.8 percent was for public water supply, and 1.2 percent was for other purposes. Table 6 shows the average annual reported groundwater use by type in the Rush Springs aquifer in three identified periods.

Table 5. Summary statistics of reported groundwater use in the study area from 1967–2015.

Statistic	Average annual reported water use, in acre-feet per year			
	Average	Median	Minimum	Maximum
1967–2015	68,719	62,179	38,485	132,904
1967–1980	76,544	78,958	52,766	103,112
1981–1997*	55,875	52,693	44,741	82,622
1998–2009	59,247	58,458	38,485	75,813
2010–2015	103,656	105,112	71,897	132,904

*Data do not include 1992.

Table 6. Table showing reported average annual groundwater use by type in the study area from 1967–2015.

Time span	Average annual reported water use by type, in acre-feet per year		
	Irrigation	Public water supply	Other
1967–2015	62,501	5,362	855
1967–1980	72,147	3,938	457
1981–1997*	50,382	4,903	584
1998–2009	52,139	6,306	807
2010–2015	93,030	8,023	2,600

*Data do not include 1992.

Provisional-temporary Groundwater Permits

For temporary use of water, the OWRB issues provisional-temporary groundwater permits that expire 90 days after issuance. Provisional-temporary permits were first issued in 1992 and continued to be utilized at the time of this investigation. The primary function of a provisional-temporary permit is to allow for a short-term water supply. These permits are typically issued for entities that need a short-term supply or long-term permit holders who have exceeded their allocation and need to supplement their supply. Unlike long-term permits where permit holders are required to submit annual use, provisional-temporary permits are issued for a reasonable volume and are not assumed to exceed the authorized amount. A more detailed description of provisional-temporary permits is available in OWRB Rules Chapter 30: Taking and Use of Groundwater (Oklahoma Water Resources Board, 2014a).

Figure 17 shows annual groundwater use from the study area for provisional-temporary permits. Authorized volumes for provisional-temporary permits for 1993–2015 averaged 668 acre-feet per year.

The highest volume of groundwater use authorized from provisional-temporary permits was 2,916 acre-feet in 2014. The lowest volume authorized was 48.4 acre-feet in 2009. Irrigation accounted for about 54 percent of the total authorized amount; oil, gas, and mining accounted for about 31 percent. Groundwater withdrawn from the aquifer utilizing provisional-temporary permits for 1993–2015 accounted for less than 1 percent of the total reported groundwater withdrawals (Table 7).

Hydrogeology

The primary water-bearing geologic unit of the Rush Springs aquifer is the Permian-age Rush Springs Formation. The aquifer is confined to the west by the Cloud Chief Formation and basally by impermeable shales and mudstones of the Dog Creek Shale. The USGS reported that the Marlow Formation acts as a confining unit that significantly retards downward movement of water from the Rush Springs aquifer to underlying units (Becker and Runkle, 1998). However, evidence suggests that the Marlow Formation contains water and is part of the groundwater-flow system (see Base of the Rush Springs Aquifer section). Quaternary-age alluvium and terrace deposits lay unconformably on top of the aquifer and are in hydrologic connection with the aquifer, transmitting water readily to the Rush Springs Formation. These alluvium and terrace deposits are considered part of the Rush Springs aquifer groundwater-flow system where these sediments directly overlie the Rush Springs Formation.

The term “Rush Springs aquifer” has been used synonymously with “Rush Springs Sandstone” and “Rush Springs Formation” in previous publications (Becker and Runkle, 1998). In this investigation, the “Rush Springs aquifer” is defined as the Permian-aged water-bearing rocks of the Whitehorse Group, which includes the Rush Springs and Marlow Formations. Generally, the Rush Springs Formation is a good substrate to store and transmit water because it consists of poorly cemented sandstone, whereas the Marlow Formation contains more siltstones and is less transmissive. The Dog Creek Shale below the Marlow

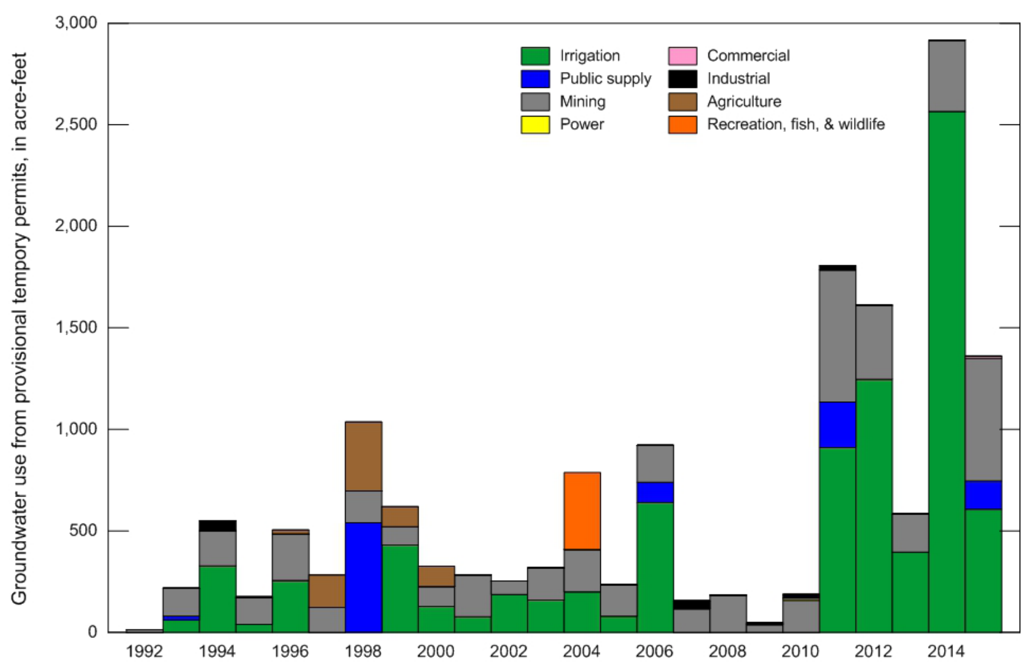

Figure 17. Graph showing annual authorized groundwater volume issued for provisional-temporary permits in the study area from 1993–2015.

Table 7. Summary statistics from provisional-temporary permits in the study area from 1993–2015.

Statistic	Reported annual water use, in acre-feet per year								
	Irrigation	Mining	Public water supply	Agriculture	Recreation	Industrial	Commercial	Power	Total
Average	361	207	45	31	17	7	1	0	668
Median	144	158	0	0	0	0	0	0	322
Minimum	0	38	0	0	0	0	0	0	48
Maximum	2,565	650	540	340	380	50	12	10	2,916
Percent of total use	54%	31%	7%	5%	2%	1%	0%	0%	

Formation consists of mudstones and is very limited for storage and transmission of water. The water-bearing rocks do not correspond to a conventional lithologic contact between geologic units. Therefore, for this investigation, the base of the Rush Springs aquifer is within the water-bearing portions of the Marlow Formation based on available lithologic descriptions from groundwater well logs (Oklahoma Water Resources Board, 2015) and in cores showing poorly cemented sandstone and siltstone within the Marlow Formation (Becker and Runkle, 1998).

Base of the Cloud Chief Formation

The western extent of the Rush Springs aquifer is partially overlain by the Cloud Chief Formation, which is composed of reddish-brown to orange-brown shale with interbedded siltstone and sandstone and has been reported to be up to 400 feet thick. The base of the Cloud Chief Formation is marked by the Moccasin Creek Gypsum Member, a 30 to 60 feet thick gypsum layer (Carr and Bergman, 1976).

The surficial geologic contact between the Cloud Chief and Rush Springs Formations was inferred where Quaternary-age deposits obscure the contact (Johnson and others, 2003; Miller and Stanley, 2004; Fay, 2010A; Fay, 2010B). Lithologic logs were examined to assess the depth to the base of the Cloud Chief Formation. Most depths were determined by using the base of the Moccasin Creek Gypsum Member as a marker bed in the lithologic description, which typically appears as a last gypsum layer under siltstone. About 30 to 60 feet below the Moccasin Creek Member is the Weatherford Gypsum Bed within the Rush Springs Formation, which is similar in lithology and thickness, according to well logs, to the Moccasin Creek Gypsum Member. The two gypsum layers are separated by red-brown sandstone.

About 350 lithologic logs were found to have adequate lithologic descriptions to determine the base of the Cloud Chief Formation. Base elevations were estimated by subtracting depth to base from a Digital Elevation Model (DEM) at each well site. Some areas of the Cloud Chief Formation in the study area have sparse well-log information, such as the area to the southeast. About 100 control points were added at the surficial Rush Springs Formation and Cloud Chief Formation contact, assuming the base of the Cloud Chief Formation was equal to surface elevation, to assist in interpretation.

Figure 18 is a map showing the base elevation of the Cloud Chief Formation from its highest in the northwest to its lowest in the southeast. The lowest elevation is along the axis of the Anadarko Basin, a dominant feature in the southeast. Subtracting the Base of the Rush Springs aquifer from the base of the Cloud Chief Formation indicates the maximum thickness of the Rush Springs aquifer ranges between 300 and 400 feet.

Base of the Rush Springs Aquifer

The USGS previously determined the base of the aquifer using lithologic logs, geophysical logs, and elevations from the Rush Springs Formation/Marlow Formation contact on 1:250,000 scale geologic maps (Becker and Runkle, 1998). However, the investigation only included the portion of the Rush Springs aquifer south of the Canadian River and approximately east of the Washita River. Since the USGS investigation, new sources of subsurface data have been collected, including geophysical logs, rock core logs, and thousands of new lithologic well logs, which have been submitted by well drillers to the OWRB (Oklahoma Water Resources Board, 2012). As a result, the base of the aquifer in the study area has been reanalyzed.

The USGS reported that the Rush Springs Formation/Marlow Formation contact can be gradational and difficult to establish in geophysical logs and lithologic logs (Becker and Runkle, 1998). Additionally, lithologic descriptions in OWRB well logs rarely differentiate the Rush Springs Formation and Marlow Formation. With the premise that well drillers typically stop drilling a well when the bottom of the aquifer is reached, two distinct features were noticed in the lithologic logs: (1) lithologic logs from fully-penetrating wells often described the last lithologic unit as either "red bed" or "dark red bed" and (2) although the variability in driller lithologic descriptions is high, the majority of logs describe the bottom of the boring as either "red bed," "dark red bed," "red shale," or "red siltstone." In most instances, this represents a change in bedrock texture from coarser-grained to finer-grained material and a change from water-bearing rock to dry rock.

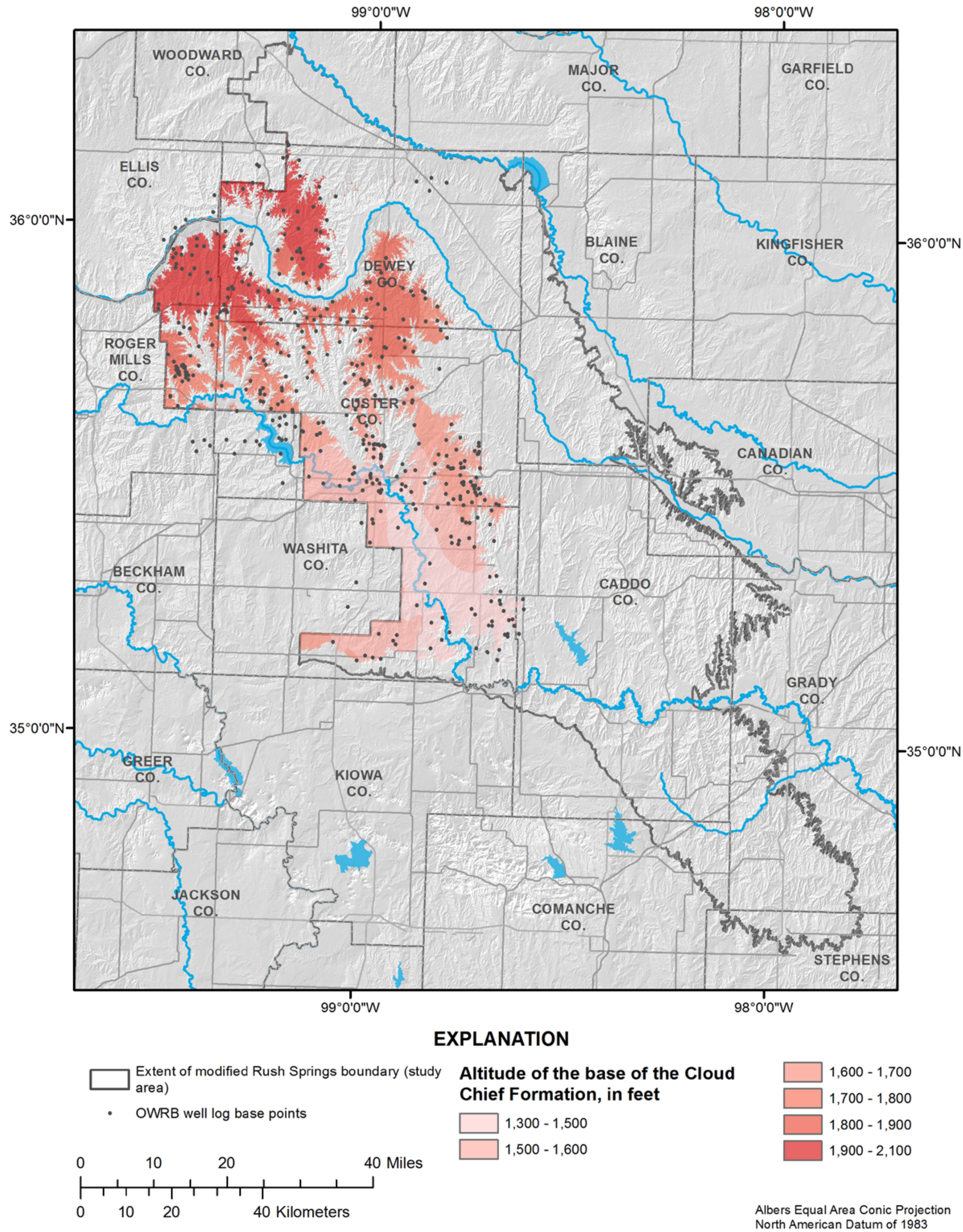

Figure 18. Raster map showing the elevation of the base of the Cloud Chief Formation derived from lithologic logs submitted to the OWRB.

Lithologic logs were also compared to geologic maps to estimate aquifer thickness for each well location. The base of the Rush Springs aquifer was estimated at each suitable well log site by subtracting the estimated depth to base from the land surface elevation provided by DEMs.

Geologic maps were analyzed alongside corresponding lithological logs to test the premise that the base of the aquifer was the Rush Springs Formation and Marlow Formation contact. Elevations of the contact were collected from a DEM or a geologic map and used as control points. When these points were used in conjunction with the lithologic logs, the edges of the base of the aquifer were at higher elevations than what was observed in the lithologic logs independently. Since geologic contacts were not able to provide useful information regarding the base of the aquifer, researchers determined that the geologic contact between the Rush Springs and Marlow Formations is not the base of the aquifer, and that the aquifer contains water-bearing portions of the Marlow Formation. The 1963 OGS report indicated the Marlow Formation had low yields in eastern Caddo County, but that in western Caddo County, the Marlow Formation contained many sandstone beds that could significantly increase well yield, suggesting that the Marlow Formation is likely in hydrologic connection with the Rush Springs Formation (Tanaka and Davis, 1963). The same process of adding control points was repeated for the current study using the Marlow Formation and Dog Creek Shale as a base of the aquifer, which showed a more gradational contact correlating to the regional dip of the bedrock.

Figure 19 shows the base elevation of the Rush Springs aquifer. The most notable feature is the axis of the Anadarko Basin that runs through central Caddo County and trends westward through Washita County. The base of the aquifer gradually rises in elevation to the north and northeast. There is also a rise in elevation near the towns of Cement and Cyril, indicating less aquifer thickness in these areas, which may be a result of diagenesis of the bedrock (Donovan, 1974; Allen, 1980; Al-Shaieb and Lilburn, 1988) or a fault zone (Marsh, 2016). Both of these features were observed in the 1998 USGS investigation (Becker and Runkle, 1998). An OGS report had previously noted the changing dip and dip direction of the Marlow Formation along the northern portion of the aquifer boundary near Greenfield in Blaine County, which averaged 17 to 18 feet per mile west-southwest in the northern part of the county, 7 feet per mile in the Greenfield area, and 16 feet per mile southwest to south in the southern and southwestern parts of the county (Fay, 1962). The OGS also noted the presence of a broad synclinal nose near the Canton area, where the strike is in a more westerly direction (Fay, 1962). These dips were also observed by researchers for the current analysis.

Aquifer Saturated Thickness

The saturated thickness of the Rush Springs aquifer was estimated by subtracting the base of the Rush Springs aquifer from the 2013 potentiometric surface (Figure 20). Saturated thickness ranged from zero to 432 feet, with a mean value of 181 feet. The aquifer is thinnest in the southeastern portions where the Rush Springs Formation outcrops and has been eroded. On a smaller scale, the thinnest portions of the aquifer correspond to where the Canadian and Washita Rivers have downcut (Figures 21–22). Other thin areas of the aquifer occur near the towns of Cyril and Cement where the base is at higher elevation. The area northeast of Sugar Creek also shows erosion of the Rush Springs Formation with thin saturated thicknesses. The thickest saturation is located along the Anadarko Basin axis where the Cloud Chief Formation confines the Rush Springs aquifer, allowing for a full section of the Rush Springs Formation. Between the Canadian and North Canadian Rivers, there is also a zone of thick saturation near the town of Oakwood where a full section of Rush Springs Formation may be present. Because the Cloud Chief Formation is not considered part of the groundwater-flow system, potentiometric heads were capped at the base of the formation to estimate saturated thickness.

Cross Sections

Four cross sections of the Rush Springs aquifer were created for the study (Figure 20) showing the base of the Cloud Chief Formation, base of the Rush Springs aquifer, and the 2013 potentiometric surface datasets (Figures 21–24). The cross sections trend from the north-northeast portion of the aquifer to the south-southwest, with approximate dip direction. The cross sections show numerous creeks intersecting the potentiometric surface that are likely draining the aquifer. In areas where the Cloud Chief Formation is present, streams have not downcut into the aquifer. In these areas, the aquifer is not draining, causing confining conditions and higher potentiometric head. These conditions can be observed in the southwest portions of cross sections A-A’ (Figure 21) and B-B’ (Figure 22). These areas coincide with the axis of the Anadarko Basin and show a fully-saturated aquifer.

Figures 21–23 are cross sections that show the Canadian River downcutting through the Rush Springs aquifer, which contributes base flow to the river (Ellis and others, 2017). Cross sections A-A’ and B-B’ show Barnitz Creek cutting into the aquifer to the west, which contributes base flow to the Washita River downstream from Foss Reservoir. Figure 23 shows the Deer Creek drainage basin, which drains into the Canadian River beginning about the 13-mile mark to about mile 23. Figure 23 illustrates the Sugar Creek drainage basin from about the 15-mile mark in the cross section to about mile 28, which drains a good portion of the aquifer and enters the Washita River east of Anadarko, Oklahoma.

Cross sections B-B’ and C-C’ also show the Washita River downcutting through the land surface. In B-B’, the Washita River is eroding the Cloud Chief Formation with a steep slope to the northeast. Downstream, in cross section

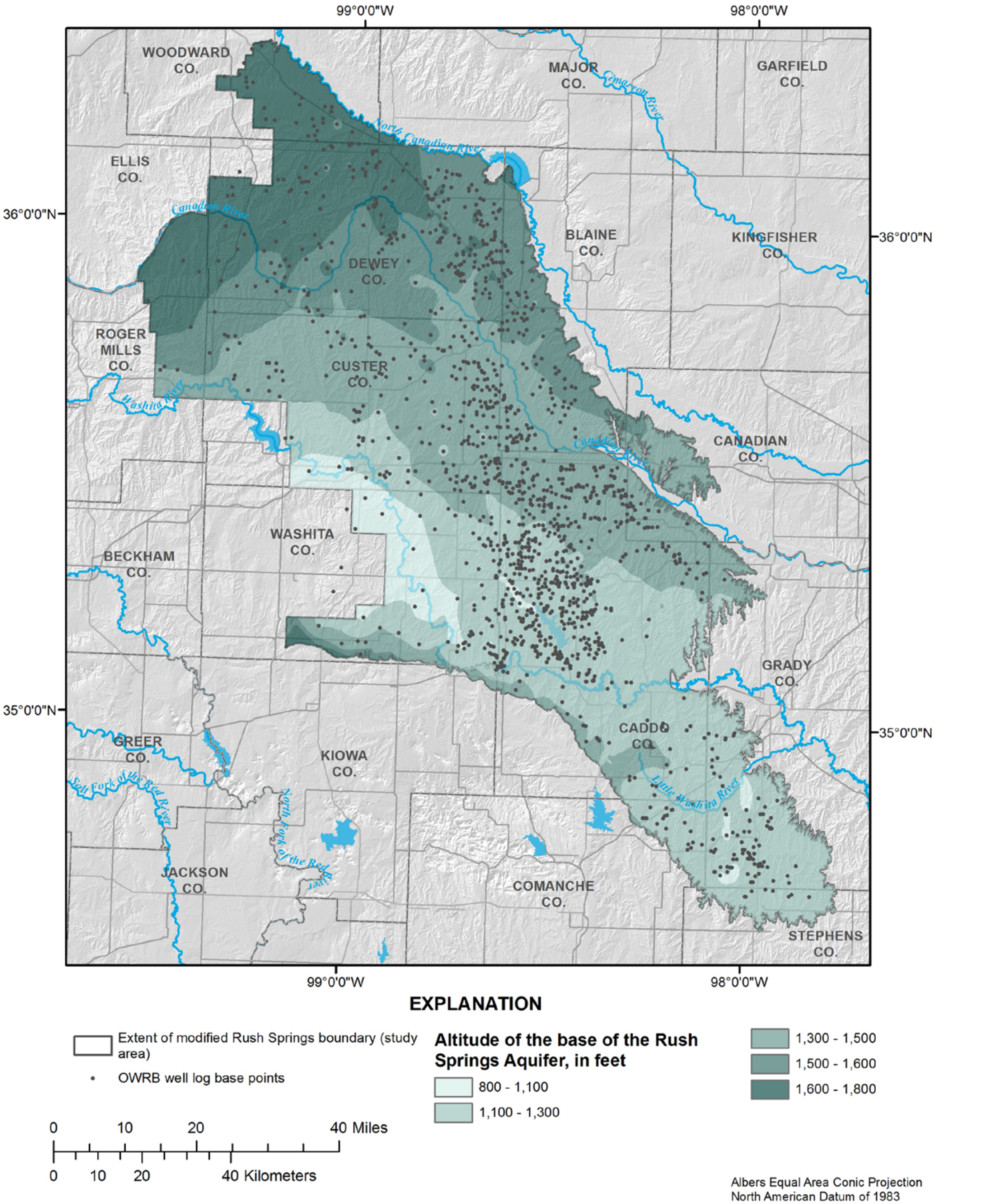

Figure 19. Raster map showing the elevation of the base of the Rush Springs aquifer derived using lithologic logs submitted to the OWRB.

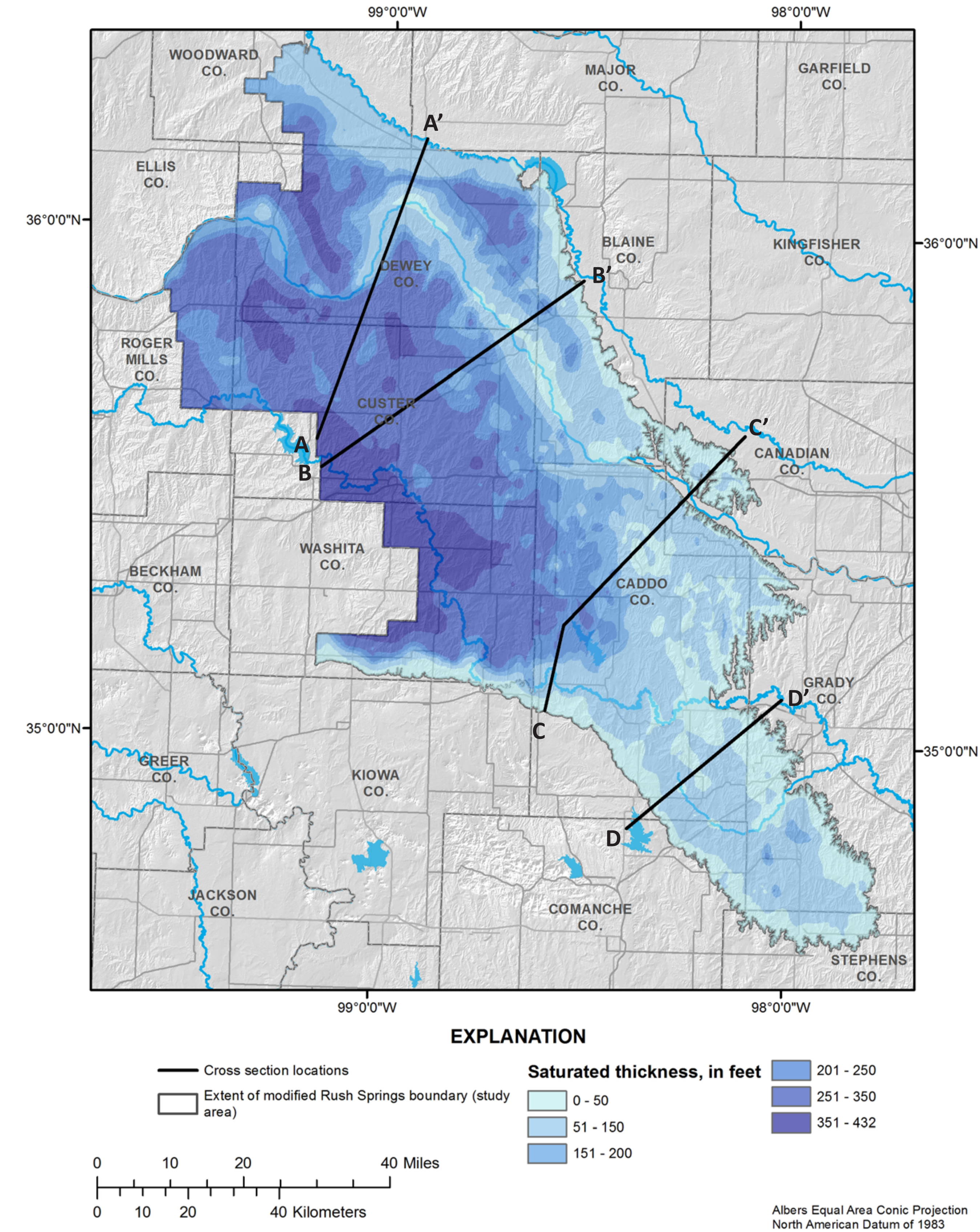

Figure 20. Map showing saturated thickness (2013) in the Rush Springs aquifer.

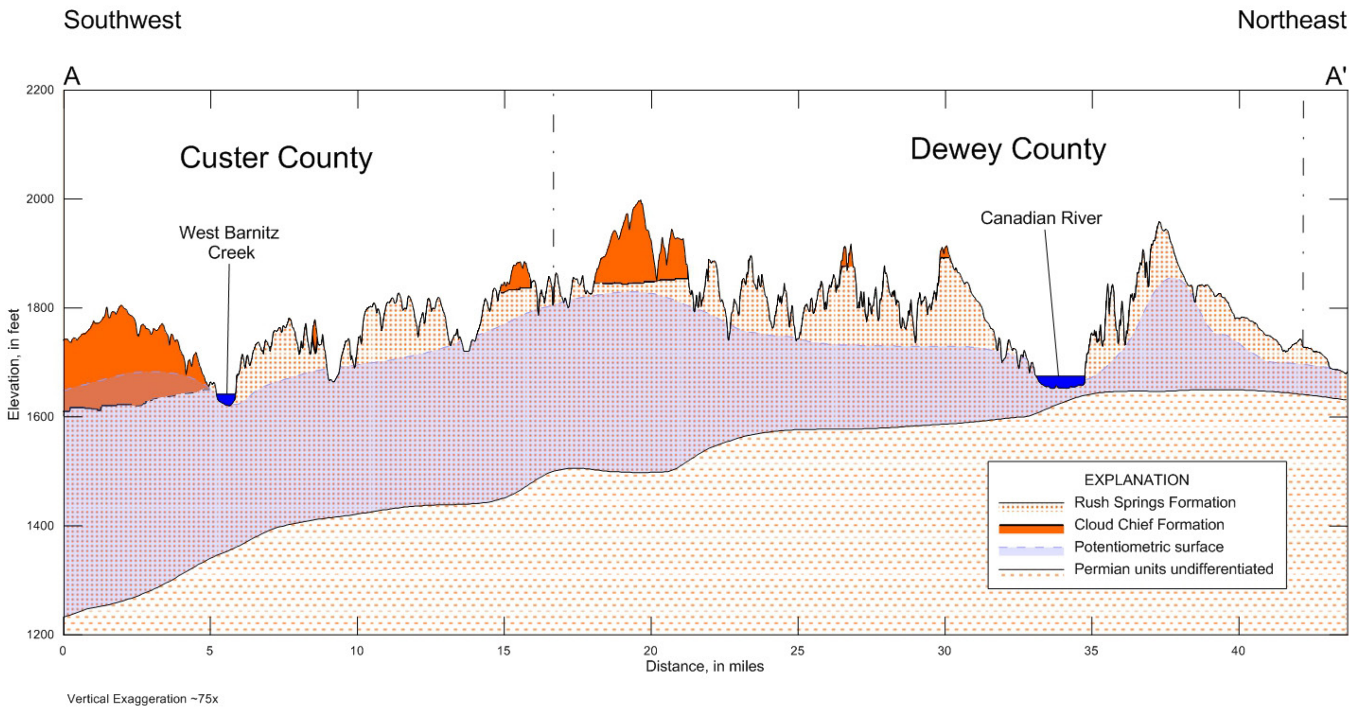

Figure 21. Cross section A-A’ from the southwest to the northeast showing geologic units, 2013 potentiometric surface, and saturated thickness.

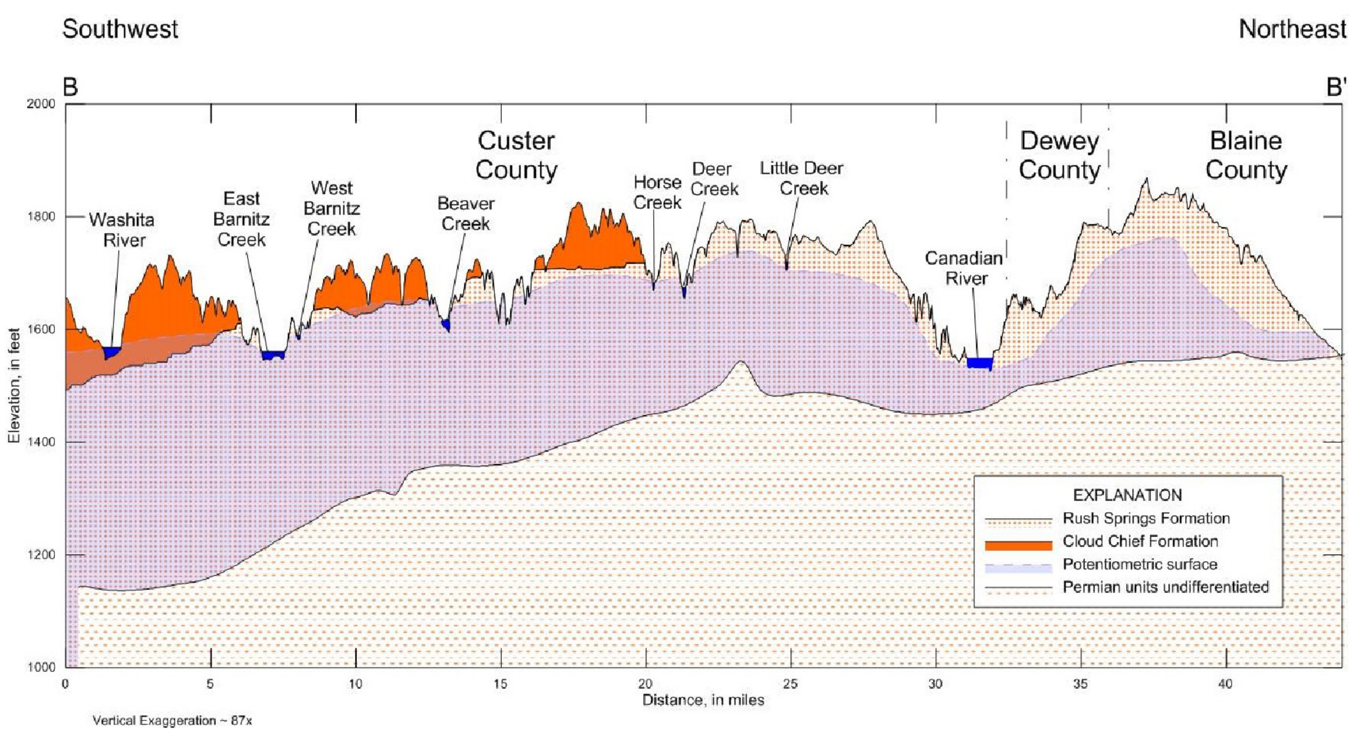

Figure 22. Cross section B-B’ from the southwest to the northeast showing geologic units, 2013 potentiometric surface, and saturated thickness.

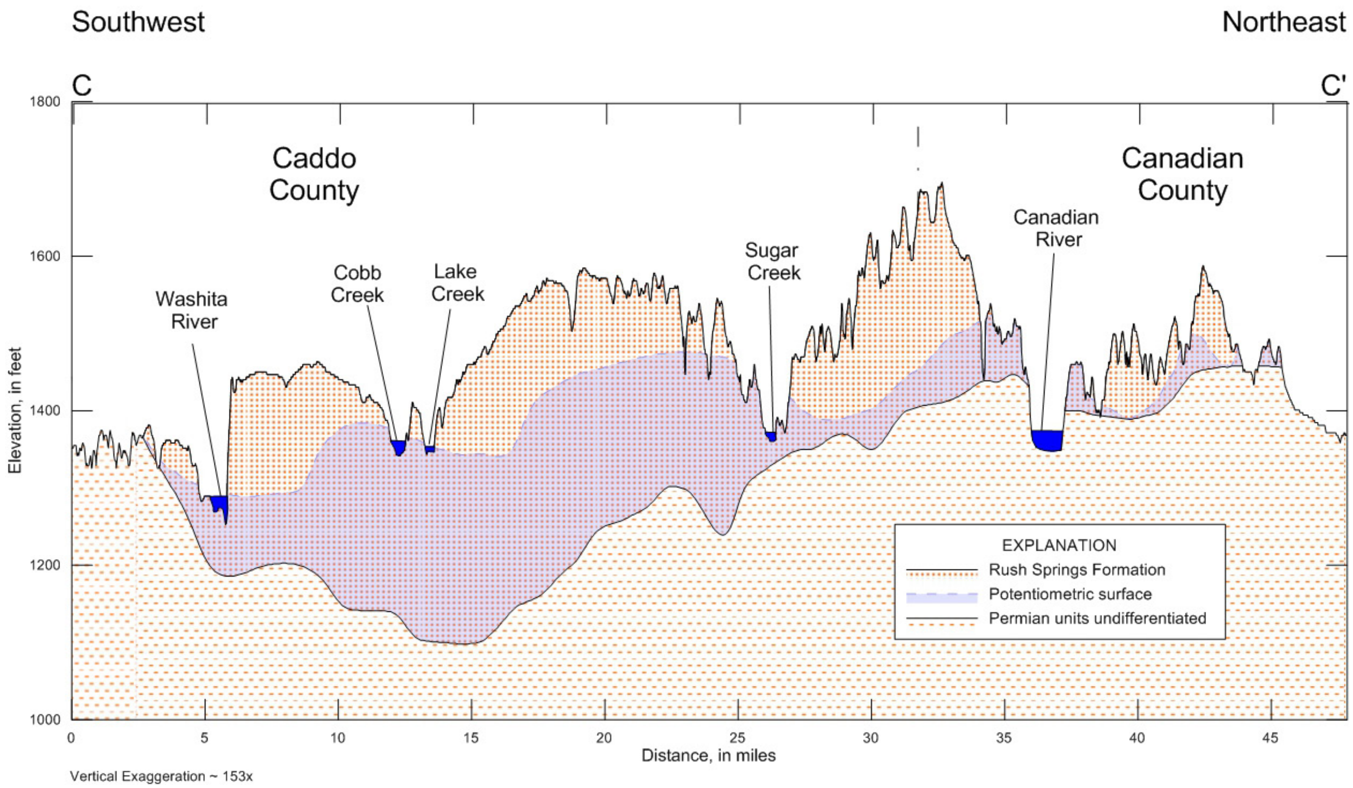

Figure 23. Cross section C-C' from the southwest to the northeast showing geological units, 2013 potentiometric surface, and saturated thickness.

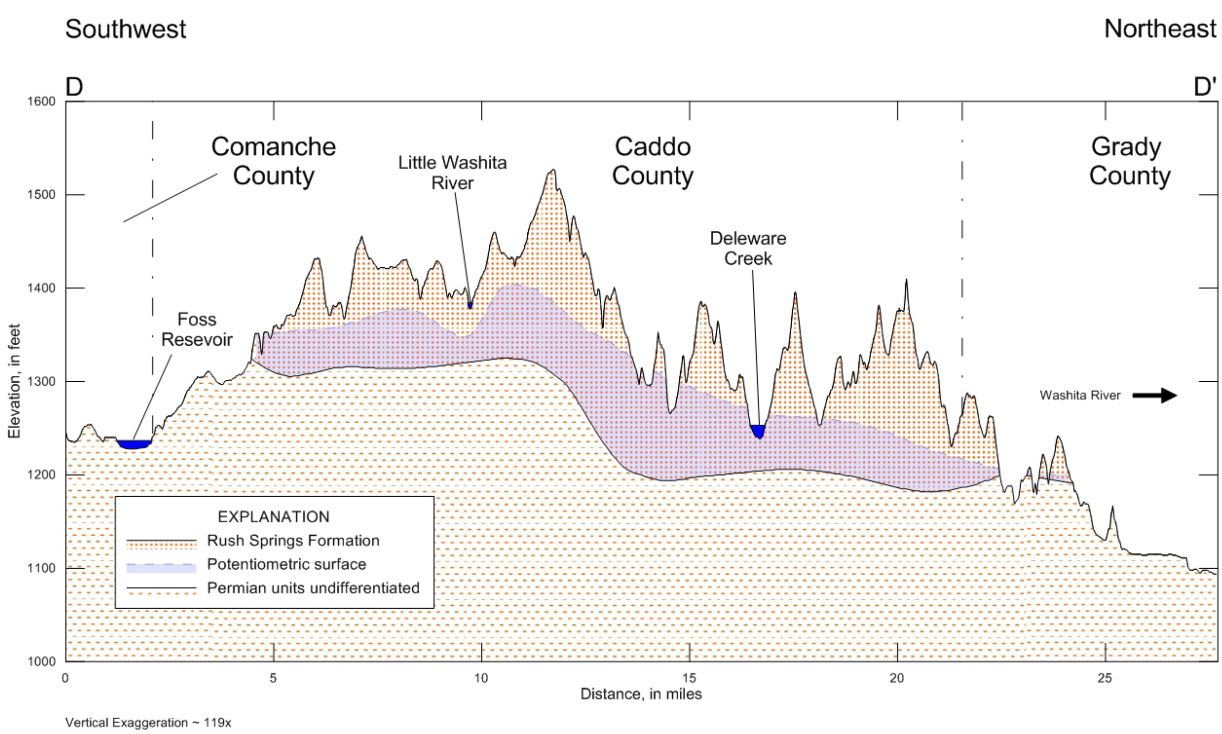

Figure 24. Cross section D-D' from the southwest to the northeast showing geological units, 2013 potentiometric surface, and saturated thickness.

C-C', the Washita River has down cut into the Rush Springs aquifer with a much steeper slope on the northeast side of the river in Section 21, Township 8N, Range 13W, just north of the town of Carnegie. The Washita River gains a large portion of base flow between the intersections of cross sections B-B' and C-C'.

Figure 24 shows that the southeastern portion of the aquifer is relatively thin compared to other areas, and there are no major rivers downcutting this part of the aquifer. The saturated thickness is thin in this area—less than 100 feet in most locations along the section. To the northeast, the Washita River has eroded the Rush Springs and Marlow formations and downcut into the underlying Permian-age bedrock.

Recharge

RORA Method

Groundwater recharge was estimated from streamflow hydrograph records using the computer program RORA, which utilizes a model developed by Rorabaugh (Rorabaugh, 1964). The Rorabaugh model is based on an ideal flow system in which the aquifer has uniform thickness, hydraulic conductivity, and storage coefficient, and where the stream fully penetrates the aquifer (Rutledge, 1998). The RORA program estimates the groundwater recharge in a basin based on the measurement of change in the total potential groundwater discharge. The program uses the recession-curve displacement method that is based on finding a critical time after a streamflow peak when recharge can be computed using the difference between the groundwater discharge from post-storm and pre-storm recessions, and the recession index (K), which is the time required for the groundwater discharge to decline on an e-log cycle after the recession curve becomes nearly linear on a semi-log hydrograph (Rorabaugh, 1964). RORA has often been used as a tool to estimate recharge for regional and local studies (Rutledge and Mesko, 1996; Flynn and Tasker, 2004; Risser and others, 2005; Mashburn and others, 2013). RORA produces a recharge rate expressed as inches per year for the subsurface drainage basin area.

RORA was used to calculate estimates of annual and monthly recharge from streamflow hydrographs at Barnitz Creek near Arapaho (USGS 07324500), Bent Creek near Seiling (USGS 07237800), Cobb Creek near Eakly (USGS 07325800), Deer Creek at Hydro (OWRB 520620060010-003RS and USGS 07228400), Lake Creek near Eakly (USGS 07325850), Willow Creek near Albert (USGS 07325860), Little Washita River near Ninnekah (USGS 07327550), and Sugar Creek near Gracemont (USGS 07327000). Basin size varies from stream to stream: Barnitz Creek is 243 square miles, Cobb Creek is 132 square miles, Deer Creek is 272 square miles, Lake Creek is 52.5 square miles, Little Washita River is 232 square miles, Sugar Creek is 208 square miles, and Willow Creek is 28.2 square miles. In order to calculate recharge for the aquifer using RORA, certain criteria should be met: drainage basins must be less than 500 square miles and completely within the aquifer, must not be affected by upstream regulation from reservoirs, and must not have major withdrawals of surface water or wastewater return flow (Rutledge, 1998). Most of the streams met the criteria except Cobb Creek near Eakly, which does have a small reservoir upstream, Crowder Lake, with a surface area of 158 acres and a capacity of 2,094 acre-feet. The period of record used for analysis of recharge from the streamflow gauges ranged from 7 years at the Deer Creek streamflow gauge near Eakly to 46 years at the Cobb Creek streamflow gauge near Eakly (Table 8).

Table 8. Average annual recharge estimated by the RORA program and recession index for stream gauging stations in the study area.

Station number	Station name	Drainage area, in square miles	Period of analysis	Minimum annual recharge, in inches	Maximum annual recharge, in inches	Mean annual recharge, in inches
07324500	Barnitz Creek near Arapaho, Okla.	243	1946-1963	0.00	0.74	0.24
07237800	Bent Creek near Seiling, Okla.	139	1967-1970	0.26	0.32	0.29
07325800	Cobb Creek near Eakly, Okla.	132	1968-2015	0.48	5.76	2.00
07228400*	Deer Creek at Hydro, Okla.	274	1961-1962, 1978-1979, 2014-2015	0.94	1.73	1.23
07325850	Lake Creek near Eakly, Okla.	52.5	1969-1978, 2005-2015	0.18	2.74	1.02
07327550	Little Washita East of Ninnekah, Okla.	232	1992-2015	0.41	5.63	2.26
07327000	Sugar Creek near Gracemont, Okla.	208	1956-1974	0.11	1.53	0.58
07325860	Willow Creek near Albert, Okla.	28.2	1970-1978, 2005-2015	0.08	2.15	1.04

*OWRB stream gauging station 520620060010-003RS

Annual recharge rate estimates were calculated for all streams and ranged from 0 inches in 1953 at the Barnitz Creek near Arapaho streamflow gauge to 5.76 inches in 2007 at the Cobb Creek near Eakly streamflow gauge (Figures 25–31). Mean annual recharge rates ranged from 0.24 inches per year at Barnitz Creek near Arapaho to 2.26 inches per year at Little Washita River (Table 8). A possible reason for the variation between the streamflow gauge stations is the variable period of record; most of the period of record for the Barnitz Creek station occurs prior to the

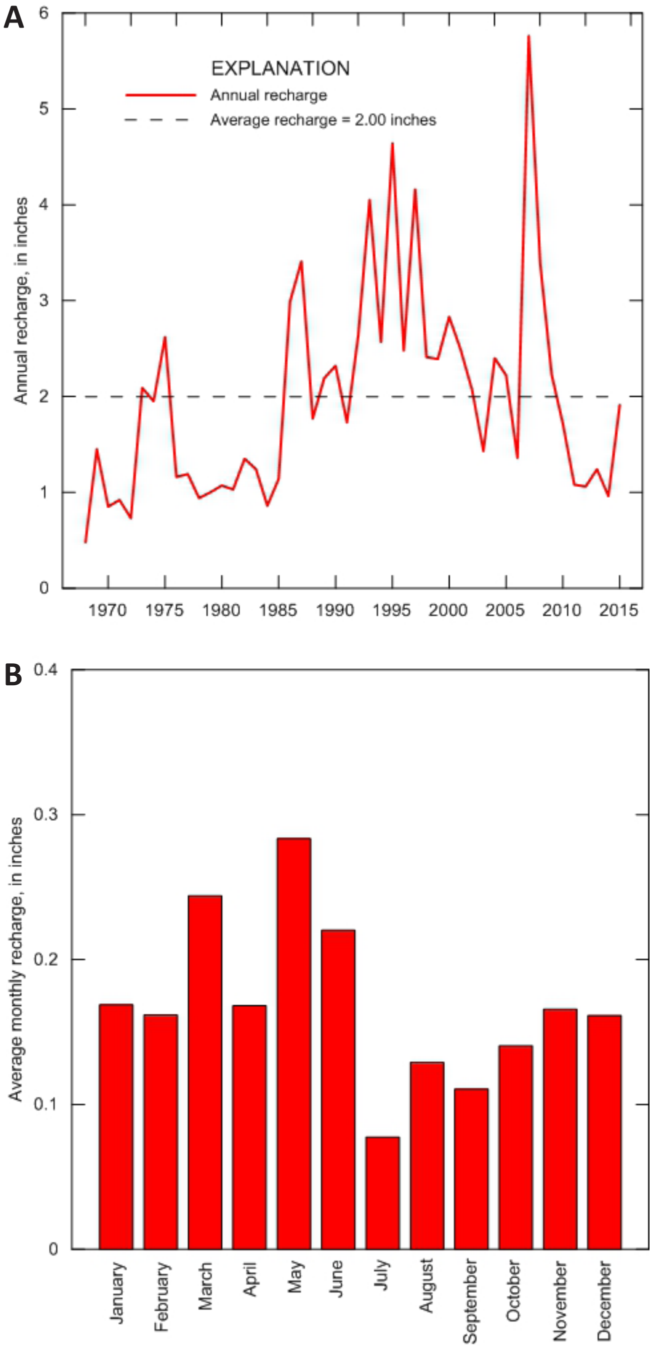

Figure 25. (A) Annual recharge, in inches, and (B) mean monthly recharge, in inches, estimated using the Rorabaugh method (Rorabaugh, 1964) or the USGS Cobb Creek streamflow gauge near Eakly, Oklahoma (USGS 07325800).

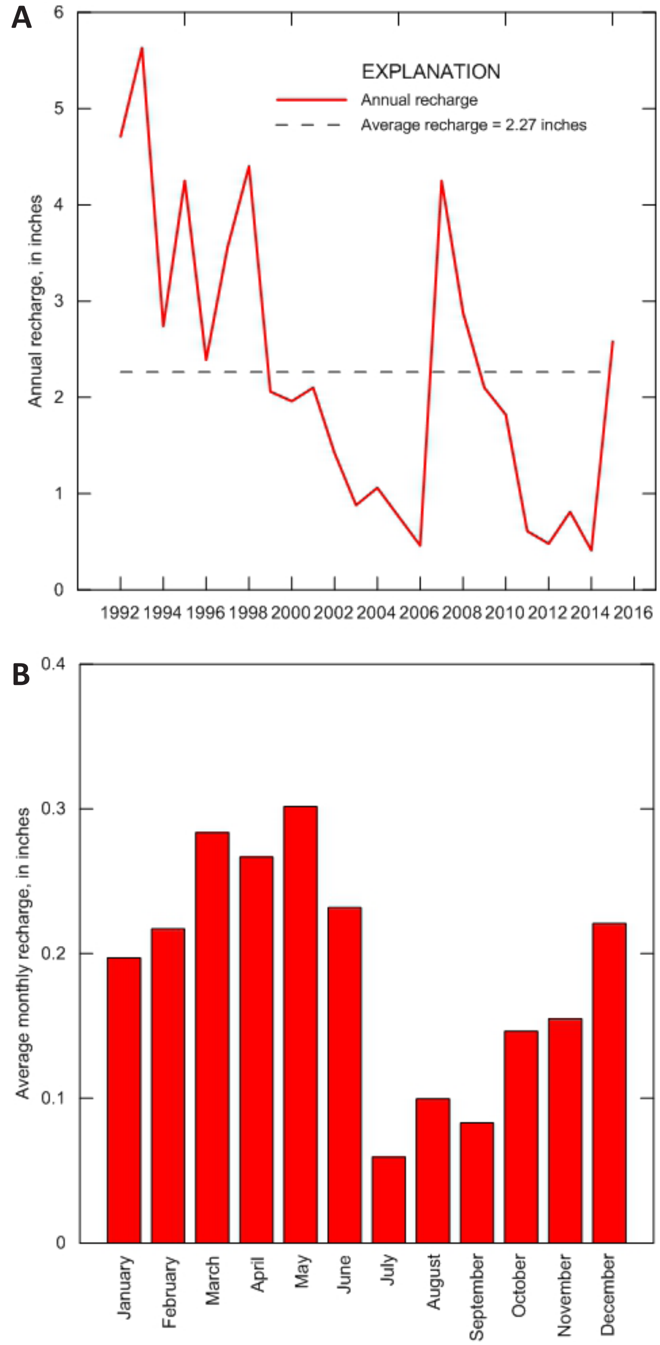

Figure 26. (A) Annual recharge, in inches, and (B) mean monthly recharge, in inches, estimated using the Rorabaugh method (Rorabaugh, 1964) or the USGS Little Washita River streamflow gauge near Ninnekah (USGS 07327550).

periods of record other stations. As noted in the Climate section, precipitation patterns during these time periods were quite different with a major drought occurring during the 1950s and above average precipitation from the mid-1980s through 2008. Also, the higher elevation recharge area of the Barnitz Creek watershed is partially confined by the Cloud Chief Formation, which may limit areal recharge. Additionally, both Lake Creek and Willow Creek have much lower flows than the other streams, which can cause error in recharge estimates; low flow rate data are

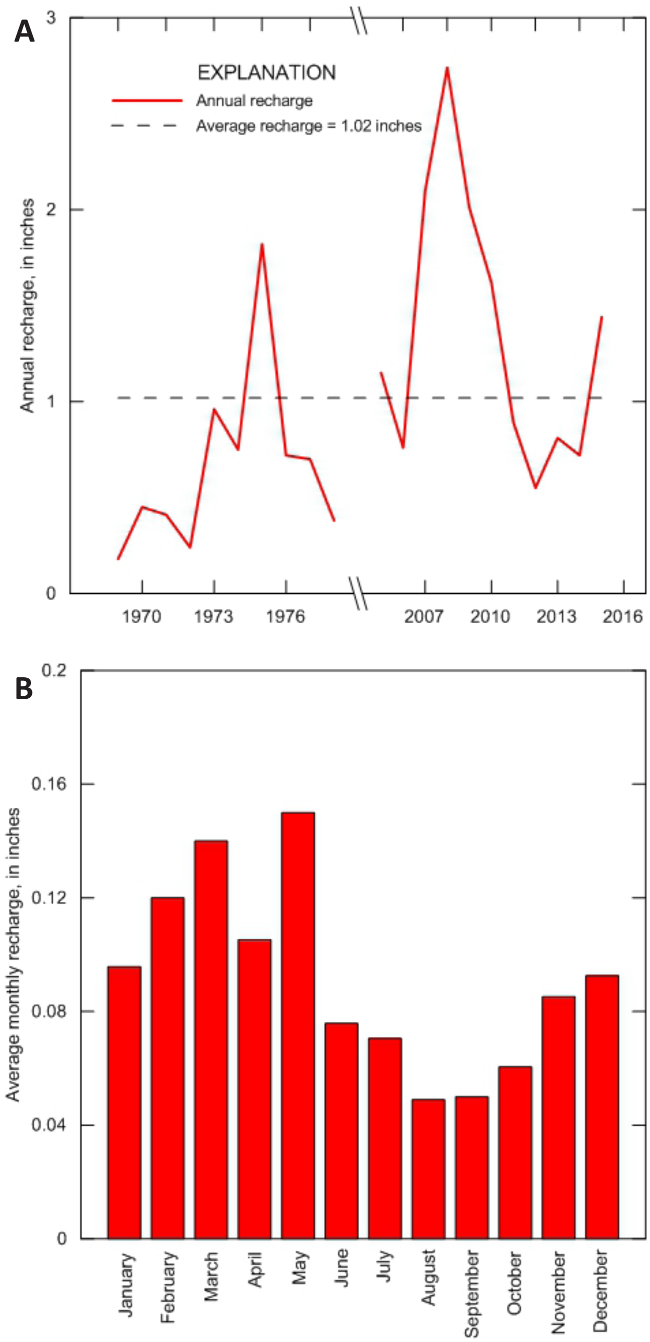

Figure 27. (A) Annual recharge, in inches, and (B) mean monthly recharge, in inches, estimated using the Rorabaugh method (Rorabaugh, 1964) or the USGS Lake Creek streamflow gauge near Eakly (USGS 07325850).

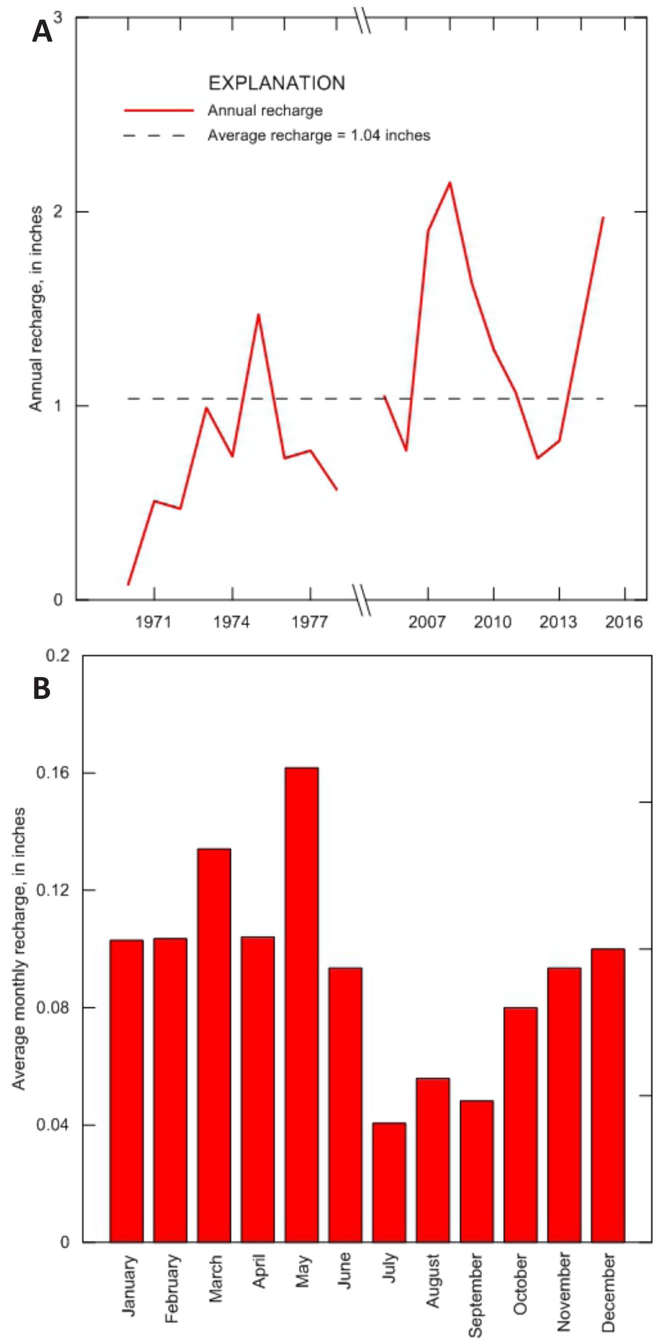

Figure 28. (A) Annual recharge, in inches, and (B) mean monthly recharge, in inches, estimated using the Rorabaugh method (Rorabaugh, 1964) or the USGS Willow Creek streamflow gauge near Albert (USGS 07325860).

subject to errors that may be considerable in proportion to the total (Rutledge, 1998). Mean monthly estimates of recharge (Figures 25–31) show recharge is typically highest during the months of March through May with a significant decrease in July, which had the lowest mean monthly recharge at every station, followed by increasing recharge into the late autumn months. The mean monthly recharge ranged from 0.01 inches per month at Barnitz Creek in January to 0.19 inches per month at the Little Washita River in May.

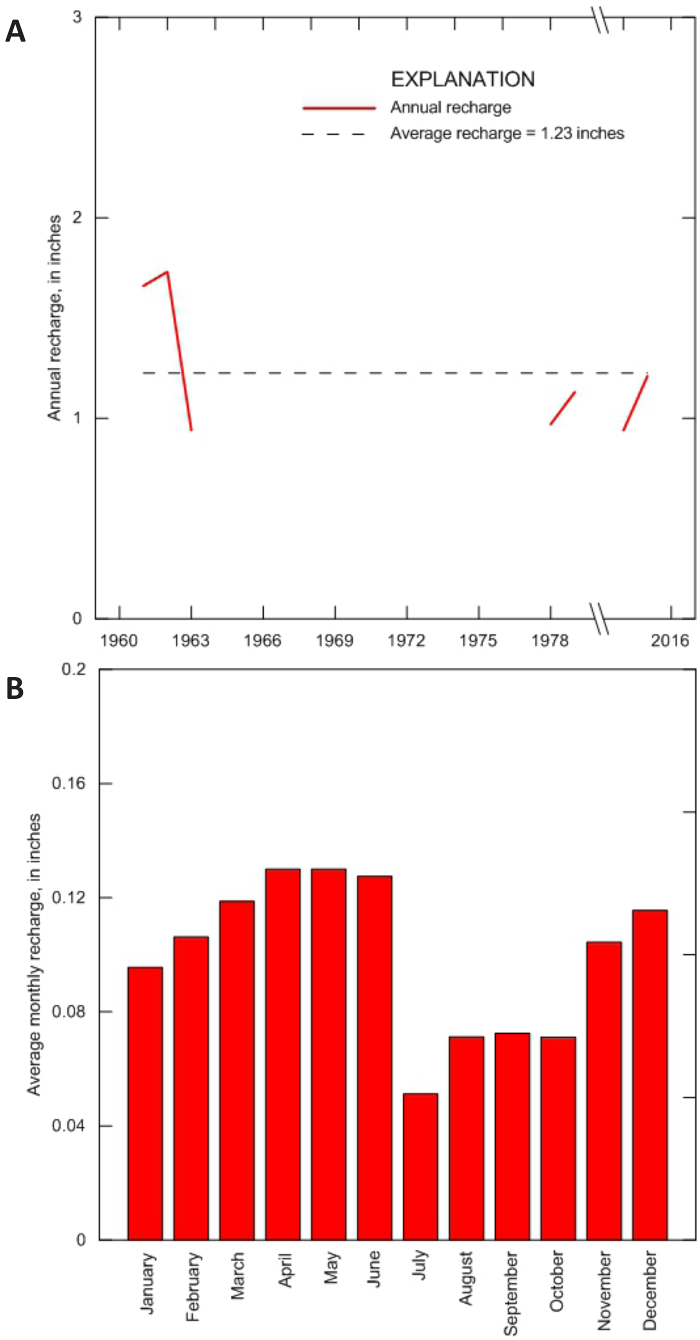

Figure 29. (A) Annual recharge, in inches, and (B) mean monthly recharge, in inches, estimated using the Rorabaugh method (Rorabaugh, 1964) or the USGS and OWRB Deer Creek streamflow gauge near Hydro (USGS 07228400 and OWRB 520620060010-003RS).

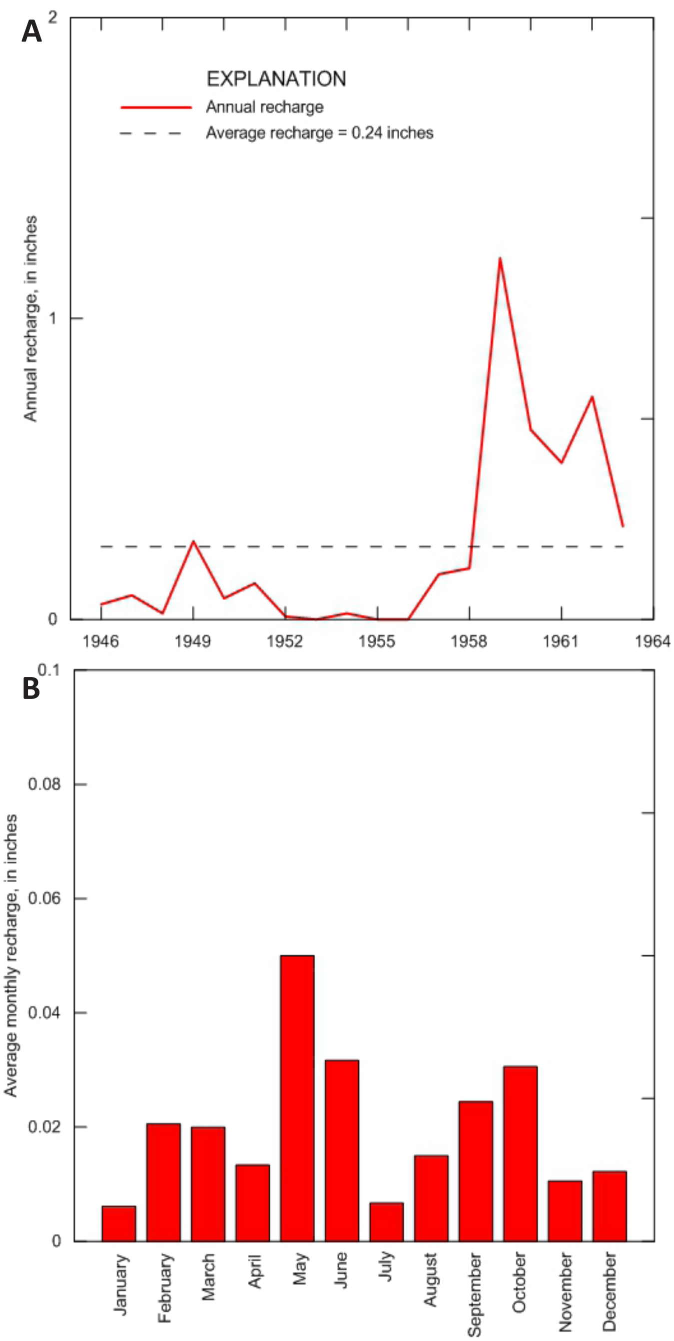

Figure 30. (A) Annual recharge, in inches, and (B) mean monthly recharge, in inches, estimated using the Rorabaugh method (Rorabaugh, 1964) or the USGS Barnitz Creek streamflow gauge near Arapaho (USGS 07324500).

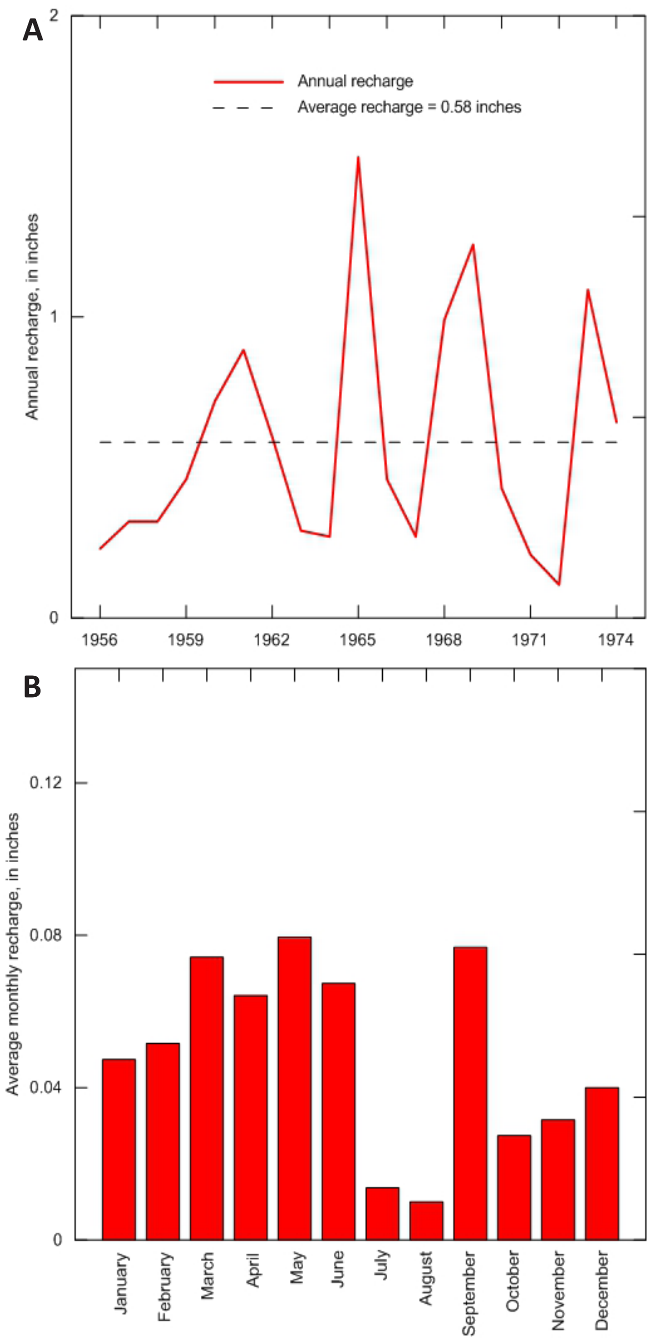

Figure 31. (A) Annual recharge, in inches, and (B) mean monthly recharge, in inches, estimated using the Rorabaugh method (Rorabaugh, 1964) or the USGS Sugar Creek streamflow gauge near Gracemont (USGS 07327000).

Soil-Water Balance

The soil-water balance (SWB) code provides a spatial and temporal estimation of groundwater recharge at a regional scale using a modified Thornthwaite-Mather soil-water balance approach in conjunction with landscape characteristics and climatological data (Westenbroek and others, 2010). Thornthwaite and Mather derived a non-linear relationship between soil moisture and water deficit; soils lose more water to evapotranspiration (ET) in the first few days of a water deficit and subsequently less as the deficit grows (Westenbroek and others, 2010). Using this relationship the SWB code calculates recharge as the difference between the change in soil moisture and the sources and sinks of water at each grid cell in the model domain at a daily time step (Westenbroek and others, 2010).

The SWB code estimates losses caused by interception, ET, and runoff at daily time steps and removes the volume from the estimated soil moisture. Interception is a user-defined amount of water utilized by vegetation that may be specified for each land-use type and season (growing or dormant). Spatially variable potential ET is estimated in the SWB code using climate data, such as air temperature, relative humidity, and wind speed. For this investigation, the Hargreaves-Samani method (Hargreaves, 1985) was used for two reasons: (1) this method utilizes data from multiple climate stations as spatially-gridded datasets and (2) this method estimates ET using the minimum and maximum air temperature in addition to daily precipitation. The potential ET represents the maximum amount of ET possible given no limitation to soil moisture. The change in soil moisture is calculated by the difference of potential ET and daily precipitation results in either positive or negative values, where actual ET equals potential ET, or negative values, indicating a cumulative deficiency.

The SWB code only considers water input in the form of precipitation and runoff entering the grid cell from up-gradient. Using temperature data, the code determines whether precipitation takes the form of rain or snow (Westenbroek and others, 2010). The daily precipitation value for a grid cell must exceed the interception and estimated potential evapotranspiration before water is assumed to contribute to soil moisture (Westenbroek and others, 2010). Once soil-moisture exceeds the maximum water capacity for the soil type and the grid cell is considered saturated, the excess is converted to recharge (Westenbroek and others, 2010). Any additional water applied to a grid cell is converted to runoff, which is either routed to an adjacent cell or out of the model domain completely. Runoff was estimated using the U.S. Department of Agriculture Natural Resources Conservation Service (NRCS) curve-number precipitation-runoff relation. Curve numbers are a baseline percentage of saturation that are modified at daily time steps using the precipitation history of the previous 5 days, vegetation dormancy, and, optionally, the frozen ground index (Westenbroek and others, 2010). As soils become saturated, there is less space for water to saturate and runoff increases. The slope of the land surface is used only to direct estimated runoff to adjacent cells (Westenbroek and others, 2010). Urban areas are typically paved and will have more runoff than pastures or irrigated croplands.

There are some limitations of the SWB code: (1) curve numbers, maximum soil recharge, interception, root zone depth, available water capacities, and infiltration rates are based on averages for land and soil types and were not directly measured for this study; (2) depth from the bottom of the root zone to the top of the water table are not factored in, resulting in recharge estimations that can be anomalously high in areas where the water table is close to the surface (Westenbroek and others, 2010); (3) ET from the groundwater table is not computed and can be underestimated in areas where groundwater occurs near land surface; and (4) soil type and maximum water capacity have the greatest impact on recharge estimation, most notably where surface water cuts through sandy soils. For this study, the root-zone values were scaled to 70 percent to ensure recharge values were not underestimated.

The SWB code was used to spatially estimate groundwater recharge over the Rush Springs study area using geospatial data sampled to a 500 square-foot grid including the following datasets: (1) Land Use (Multi-Resolution Land Characteristics Consortium, 2011, National Land Cover Database 2006 (NLCD 2006): USGS, accessed August 11, 2014, at http://www.mrlc.gov/nlcd2006.php); (2) Hydrologic soil group and soil-water capacity (NRCS, US Department of Agriculture, Web Soil Survey, accessed August 11, 2014, at http://websoilsurvey.nrcs.usda.gov/); (3) tabular climate data obtained from the COOP and Oklahoma Mesonet for a set of stations located in or near the study area consisting of daily precipitation, daily minimum temperature, and daily maximum temperature (Oklahoma Climatological Survey, 2014a; Oklahoma Climatological Survey, 2014b); and (4) additional values for soil properties based on land use such as interception and the available soil-water capacity (Westenbroek and others, 2010) in the form of a look-up table. The estimated recharge results were then clipped to the outline of the study area and statistics were tabulated.

Figure 32 is a map showing the spatial variability in estimated average annual recharge using the SWB code for 1950–2015. Estimated average annual recharge for 1950–2015 in the study area was 1.40 inches. The areas with the highest recharge estimates include the Cobb Creek, Lake Creek, and Willow Creek watersheds in Caddo county; the Little Washita River watershed in southeastern Caddo, southwestern Grady, northeastern Comanche, and northern Stephens Counties; and the area north of the Canadian River in eastern Dewey County and western Blaine County. The soils in these areas have lower available water capacities, meaning the grid cells become saturated quickly and lose less water from evapotranspiration allowing a greater percentage of precipitation to go to recharge. Recharge estimates are much lower in areas farther west due to lower precipitation totals and the presence of soils with higher available water capacity, most notably where the Cloud Chief Formation is present.

Figure 33 is a map showing estimated recharge for 2007, the wettest year on record in Oklahoma at the time of the study and the year with the highest estimated average annual recharge (4.63 inches). In the easternmost part of the study area, the basins that drain into Fort Cobb Reservoir have the highest estimated recharge. The northwestern portion of the study area was estimated to have the least amount of recharge at 0.50 inches or less.

Figure 34 shows a map of recharge estimates for 1980, a year with only 1.17 inches of estimated average annual recharge, one of the lowest estimates for the period of record. As with recharge estimates for 2007, much of the eastern area of the aquifer and areas north of the Fort Cobb Reservoir had the highest estimated annual recharge, and the northwest remained the driest. Estimated average annual recharge was the lowest for the years 1963 and 2003 (maps not shown in this report) with both years having only 0.03 inches and many cells with no estimated recharge.

Figure 35 shows estimated annual recharge for the study area for 1950–2015. The average annual recharge is estimated to be 1.40 inches with a median of 1.01 inches. Three time periods were selected that show trends of below or above average recharge estimates: (1) 1950–1984, (2) 1985–2001, and (3) 2002–2015.

The period 1950–1984 had an estimated average of 1.07 inches of recharge per year, which is 0.34 inches below the average annual recharge of 1.40 inches for the records analyzed. During this period, only 8 of the 34 years had above average recharge; median recharge was 0.88 inches.

Estimated recharge began to increase in 1985, the first year of the longest period of above average recharge for the period of record. Estimated mean annual recharge for 1985–2001 was 2.18 inches with a median of 2.00 inches. Over this 16-year period only three years were estimated to have had below average recharge.

During 2002–2015 there was a return to average conditions of 1.30 inches of annual recharge, which is close to the estimate of 1.40 inches for the period of record. However, with a median of 0.80 inches, the average is skewed by high precipitation amounts in 2007 and 2015, which have recharge estimates of 4.63 inches and 4.18 inches, respectively. Removal of these two outliers lowers the average annual recharge estimate for the period to 0.78 inches with a median of 0.73 inches. Table 9 shows the summary statistics for the SWB estimated recharge.

Figure 36 shows the average monthly recharge trends for four time periods: (1) 1950–2015, (2) 1950–1984, (3) 1985–2001, and (4) 2002–2015. During 1950–2015 the highest estimated recharge generally occurred in winter and spring and was likely caused by the combination of cool temperatures, dormant vegetation, and increasing precipitation. The sharp decline in April was likely caused by an increase in evapotranspiration. May had the highest average recharge estimate of the period of record analyzed with 0.23 inches. Recharge decreased throughout the summer months as evapotranspiration rates peaked; July had an estimated recharge of 0.02 inches, the lowest of all months.

For the 1950–1984 period, there was below average annual recharge, with 67 percent of the months receiving less than 0.10 inches. July was the only month that received more recharge for this period than for the period of record

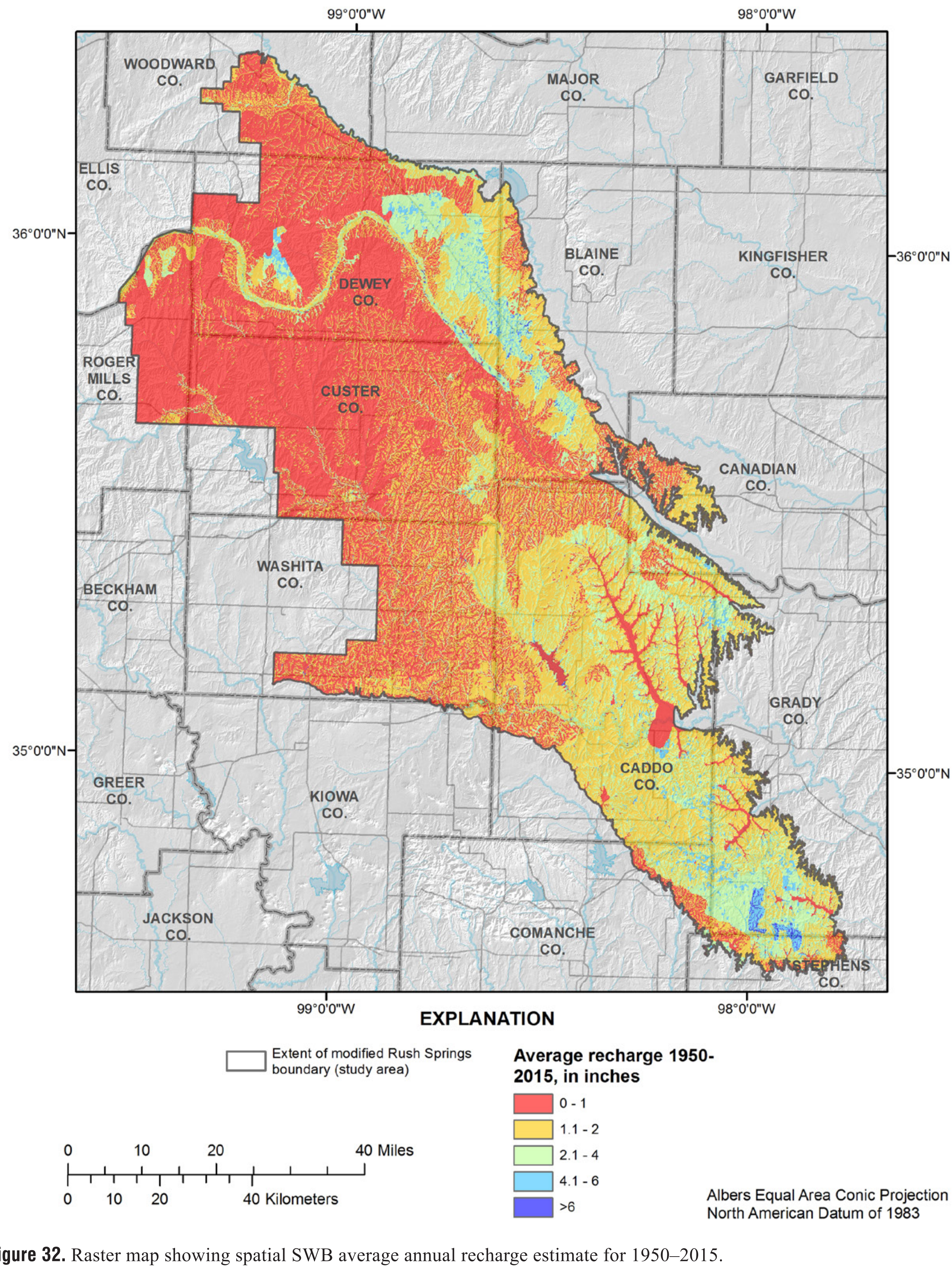

Figure 32. Raster map showing spatial SWB average annual recharge estimate for 1950–2015.

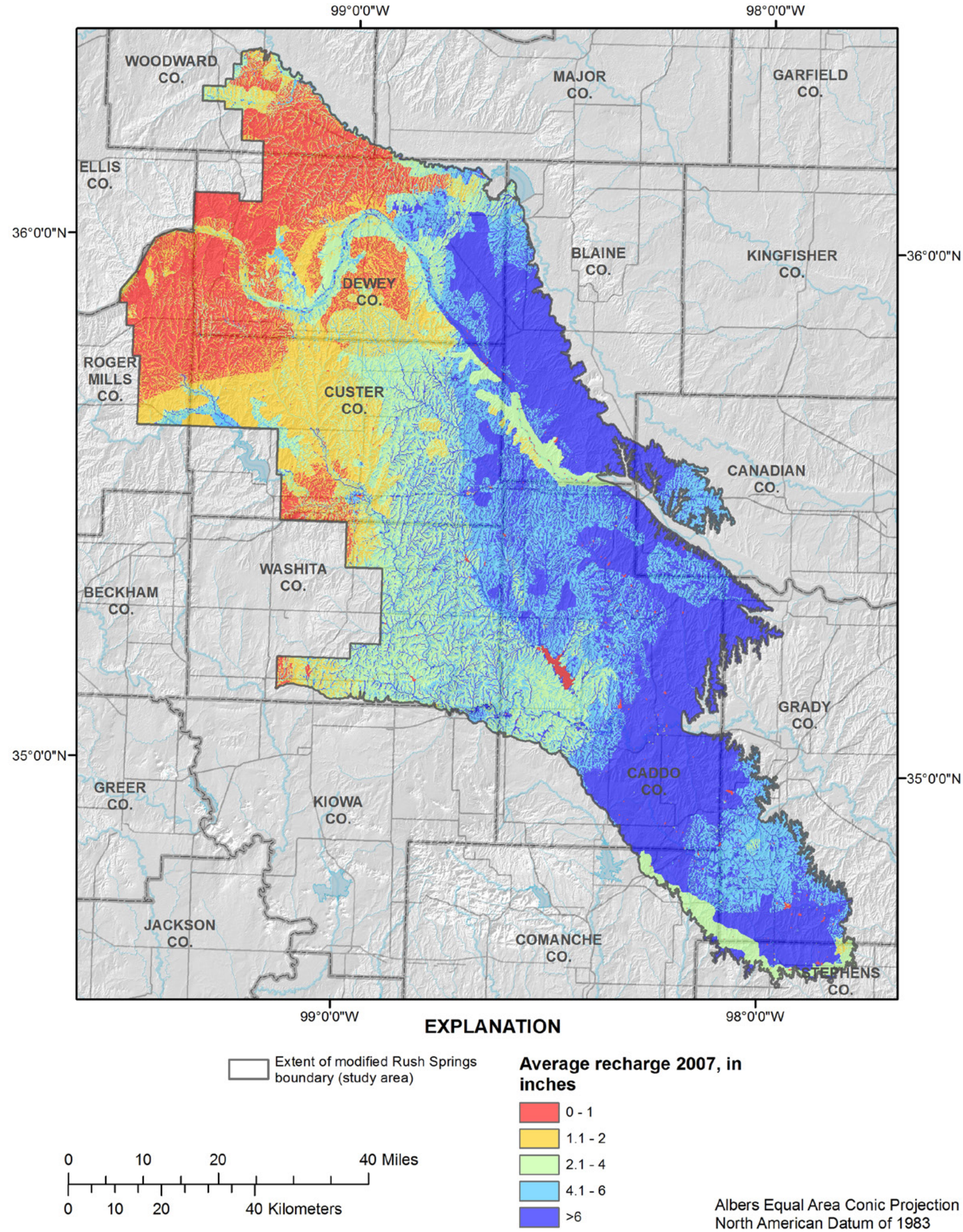

Figure 33. Raster map showing spatial recharge estimated by SWB for 2007, a year of high estimated recharge.

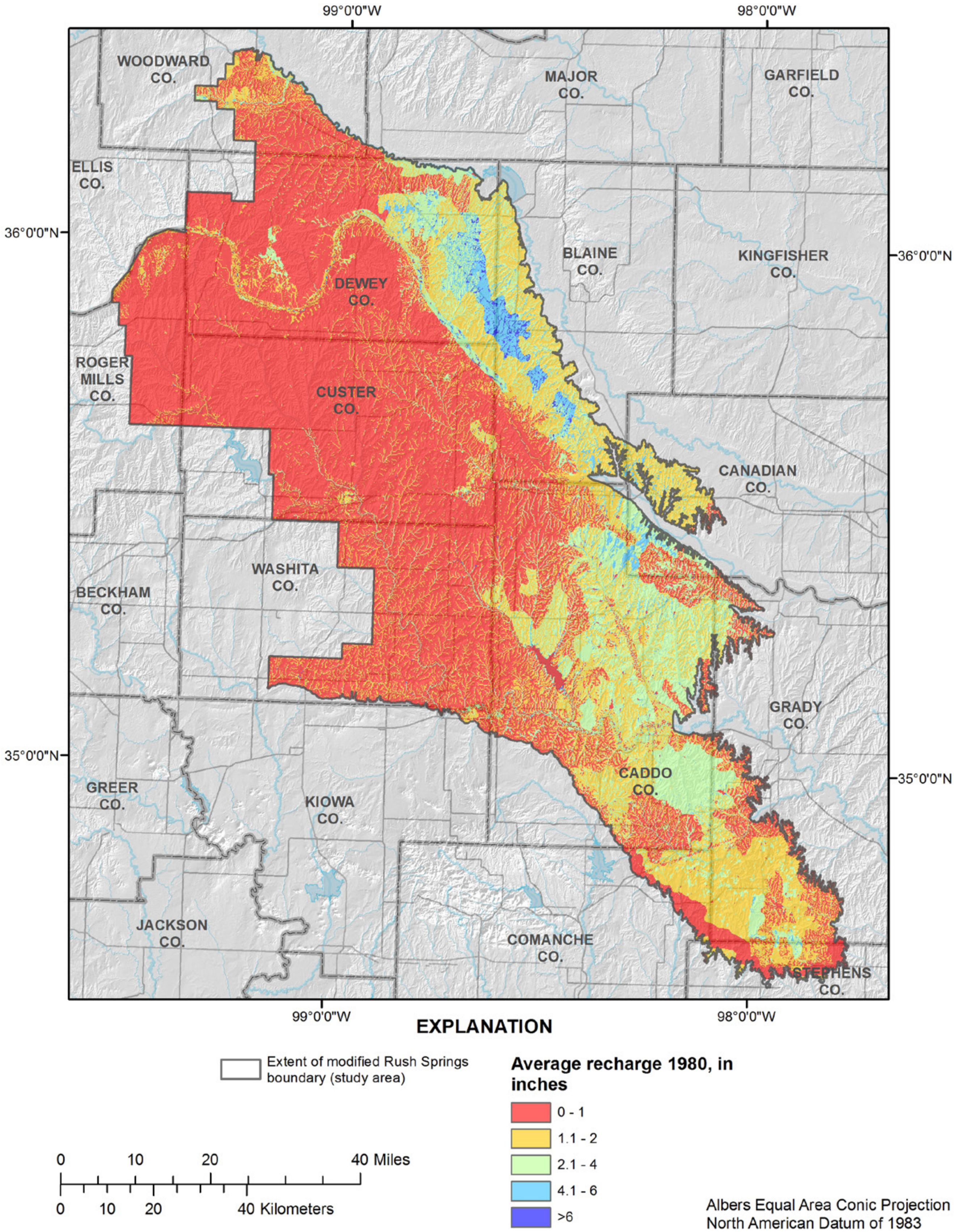

Figure 34. Raster map showing spatial recharge estimate by SWB for 1980, a year of below average recharge.

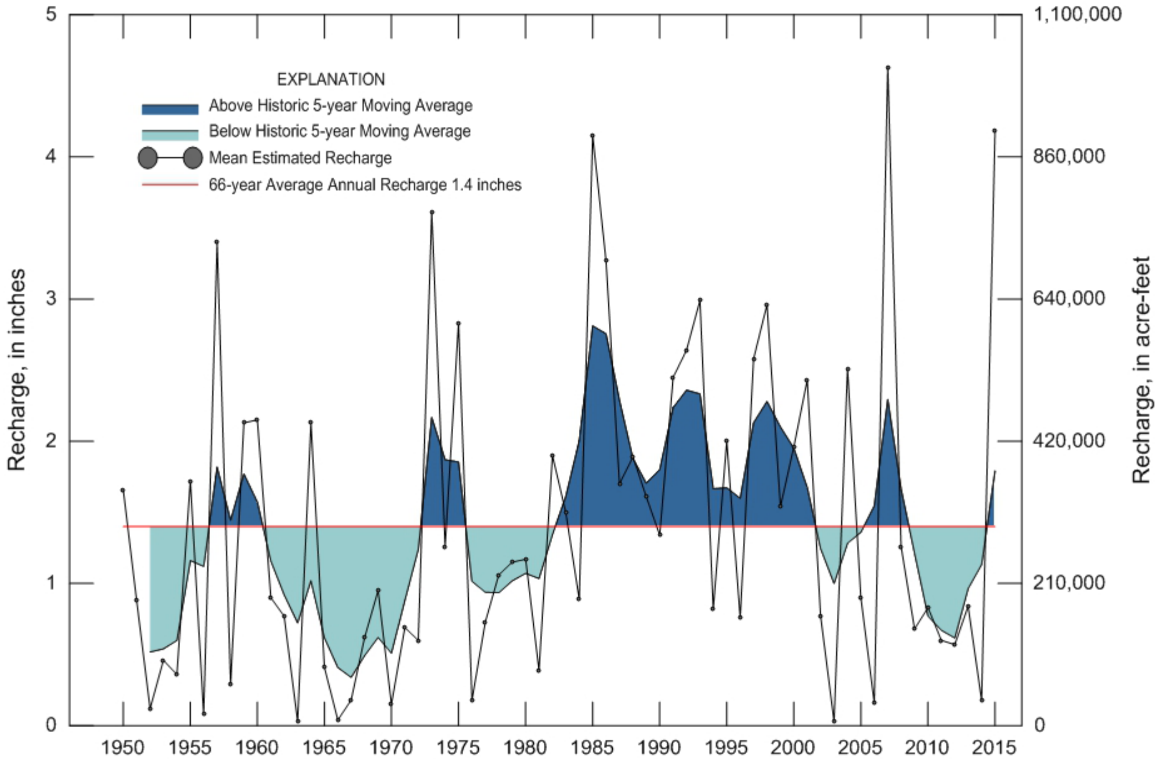

Figure 35. Graph showing annual recharge estimated by SWB for the study area for 1950–2015.

Table 9. Table of summary statistics for SWB estimated recharge for 1950–2015, 1950–1984, 1985–2001, and 2002–2015.

Statistic	Average annual SWB recharge, in inches			
	1950-2015	1950-1984	1985-2001	2002-2015
Minimum	.03	0.03	0.76	0.03
Maximum	4.63	3.61	4.15	4.63
Mean	1.40	1.07	2.18	1.30
Median	1.01	0.88	2.00	0.03

*Data do not include 1992.

analyzed. May had the highest estimated recharge with 0.23 inches, while August received the least at only 0.03 inches on average.

For the 1985–2001 period, there was above average recharge compared to the period of record. Estimated recharge was highest for January and March, with 0.37 and 0.31 inches, respectively, while the estimate for July was less than 0.01 inches, the lowest estimate for the period of record analyzed.

The estimated recharge values for 2002–2015 were slightly below average except during the spring and summer. Estimated recharge was lowest in September with only 0.02 inches; estimated recharge for April through August was above average for all months with May having the highest estimate at 0.29 inches.

Hydraulic Properties

Hydraulic properties of an aquifer are characteristics that describe groundwater flow and storage of water in an aquifer. For this investigation, hydraulic properties estimated for the aquifer include hydraulic conductivity, transmissivity, and storage (storage coefficient and specific yield). Hydraulic conductivity, expressed in units of length per time (feet per day in this report), is defined as a volume of water that is transmitted in a unit of time through a cross section of unit area (Lohman, 1972). Transmissivity, expressed in units of length squared per time (feet squared per day in this report), is defined as the rate at which water of the prevailing kinematic viscosity is transmitted through a unit width of the aquifer under a unit hydraulic gradient (Lohman, 1972). Storage refers to water held in the aquifer matrix that can be released from the aquifer under confined and unconfined conditions. Water released from the aquifer under confined conditions is caused by the compressibility of water and aquifer matrix through overburden pressure and is referred to as storage coefficient. Storage coefficient, dimensionless, is defined as the volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in the component of head normal to that surface (Sayre, 1955). The aquifer remains fully saturated under confined conditions. Under unconfined conditions, water is yielded from water-bearing material by gravity drainage and is the ratio of the volume of water that, after being saturated, is yielded by gravity to the volume of

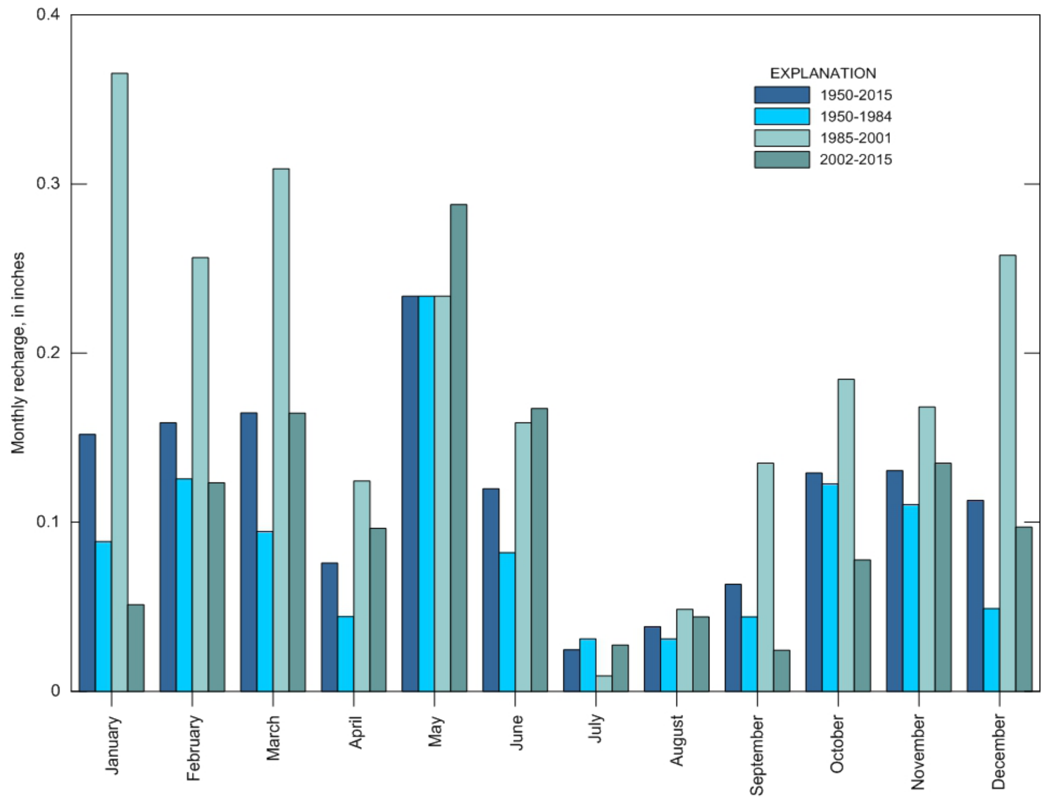

Figure 36. Graph showing average monthly recharge for 1950–2015, 1950–1984, 1985–2001, and 2002–2015.

aquifer (Lohman, 1972) and is referred to as specific yield. In unconfined aquifers, compressibility of the aquifer is negligible and the storage coefficient is considered equal to specific yield.

Hydraulic properties of the Rush Springs aquifer were estimated using several methods. Hydraulic conductivity and transmissivity were estimated using well-drawdown data, slug tests, aquifer tests, and a percent-sand method. Storage was estimated by conducting aquifer tests and using a regional method involving water-level measurements and data from streamflow gauges.

Slug Tests and Well Drawdown Data Analyses

Drawdown data from specific capacity tests of 750 municipal and irrigation wells submitted to the OWRB from well drillers were used to estimate hydraulic conductivity. These well driller logs included length of the drawdown of the water level due to constant pumping, pumping rate, pumping duration, and well radius. For each well location, the Cooper and Jacob (Cooper and Jacob, 1946) solution was applied, which was derived from the Theis nonequilibrium method (Theis, 1935) and utilized types of curves described by Jacob (1940), Wenzel and Fishel (1942), and Wenzel and Greenlee (1944). This solution is intended for analysis of wells in confined aquifers. Some of the wells in the Rush Springs aquifer are in a confined setting, but most are in unconfined settings. Since an equation for an unconfined solution was unavailable for this analysis, researchers assumed that the confined solution provided a reasonable estimate of transmissivity. This method was utilized in a previous investigation of the Rush Springs aquifer (Penderson, 1999). The Cooper and Jacob equation is:

$$\frac{Q}{S_w} = \frac{T}{0.183 \log\left(\frac{2.25Tt}{r_w^2 S}\right)}$$

where

- Q is discharge rate the well was pumped (cubic feet per day)
- Sw is the total length of equilibrated drawdown (feet)
- T is the aquifer's transmissivity near the well (square feet per day)
- t is time (days)
- S is the storativity of the aquifer (dimensionless)

The Cooper and Jacob equation can be written to solve for transmissivity:

$$T = 0.183\frac{Q}{S_w}\log\frac{2.25Tt}{r_w^2 S}$$

Because transmissivity is in the logarithm term of the equation, successive approximation may be used to solve for transmissivity. Transmissivity may be used to determine hydraulic conductivity:

$$T = Kb$$

where

K is the hydraulic conductivity of the aquifer near the well (feet per day)

b is the saturated thickness of the aquifer (feet), base of the aquifer minus static water level

The storativity value utilized for this method was derived from the regional method to determine storage coefficient (0.072).

Figure 37 is a map showing well sites with available drawdown data; 688 wells were completed in bedrock and 62 were completed in the alluvium and terrace on top of the bedrock. The northern and western portions of the aquifer lack drawdown data and results may be skewed to the areas with more data, such as the central portion of the aquifer in Caddo County. The minimum hydraulic conductivity for the Rush Springs aquifer estimated from the drawdown data was less than 0.01 feet per day and the maximum was 90.90 feet per day with a median of 1.63 feet per day and a mean of 3.27 feet per day. The mean is higher than the median, indicating that there are some higher hydraulic conductivity outliers in the dataset. For drawdown tests in the alluvium and terrace deposits within the study area, the minimum hydraulic conductivity was 0.16 feet per day, the maximum was 399.58 feet per day, and the mean and median were 26.02 and 6.19 feet per day, respectively. Figure 38 is a histogram showing the hydraulic conductivity values from all tested wells in the Rush Springs aquifer.

Slug tests are groundwater well assessments that are useful for determining the connectivity of a well with the aquifer and the hydraulic conductivity (K) of the aquifer near the well. A slug test can be conducted by observing the water-level response from an instantaneous change in head, which can be induced by adding or withdrawing water, increasing or decreasing air pressure within the well casing, or adding a solid mechanism of known volume, such as a solid PVC cylinder, to displace the water.

Fifty-four slug tests were conducted as part of the investigation to estimate hydraulic properties at well sites in the study area (Figure 38). Slug tests were performed according to published guidelines (Cunningham and Schalk, 2011) and data were analyzed with the AQTESOLV software package (Duffield, 2007). The Bouwer-Rice and Hvorslev solutions for unconfined aquifers were used to estimate hydraulic conductivity (Hvorslev, 1951; Bouwer and Rice, 1976). Both solutions are used for overdamped response in slug tests, which occurs in aquifers that have low to moderate hydraulic conductivities, such as non-karst bedrock aquifers (AQTESOLV, 2014). The majority of the slug tests were analyzed using the Bouwer-Rice solution because it provided the best match to the data. Furthermore, the Bouwer-Rice solution can be applied to wells in unconfined aquifers as well as confined aquifers that receive water from an upper confining layer (Bouwer and Rice, 1976). Hydraulic conductivities for the Rush Springs aquifer ranged from 0.13 to 7.60 feet per day. The median hydraulic conductivity was 1.40 feet per day, with a mean of 1.70 feet per day. Three slug tests performed in the alluvium of the Washita River yielded estimated hydraulic conductivities of 18.62, 37.63, and 52.19 feet per day.

The ranges of hydraulic conductivity identified in the drawdown data and slug tests indicate that some portions of the Rush Springs aquifer texturally consist of coarse-grained sandstone, poorly-cemented sandstone, or unconsolidated material.

Table 10 shows the descriptive statistics from the drawdown, slug test, and percent coarse (see Percent Coarse Analysis section) analyses. The comparison of drawdown and slug test datasets show estimated hydraulic conductivity within the range of published values for a sandstone aquifer, which is from 8.5E-05 to 1.70 feet per day, and the range for unconsolidated sand from 0.26 to 141 feet per day (Domenico and Schwartz, 1998).

Comparing the results from the drawdown and slug test datasets shows the drawdown data with a higher estimated median hydraulic conductivity than the slug tests. Considering that the data are from the same aquifer and both datasets have similar spatial coverage within the aquifer, the median value of each dataset may be expected to be similar. A reasonable explanation of the differences may be the well construction, specifically the efficiency of the well screen. Data from drawdown tests are typically collected from newly constructed wells, when groundwater flow into the well is the most efficient. However, slug test data are typically collected from existing wells that, over time, would have become less efficient. The efficiency of a well can be affected by silt or mineralization obstructing well screen. The hydraulic conductivity estimate from a pumping test is on average considerably larger than the estimate obtained from a series of slug tests in the same formation (Butler and Healey, 1998), where the differences are attributable to incomplete well development. An interpolated raster of hydraulic conductivity was created using the Inverse Distance Weighted method for both the slug test and drawdown analyses to determine the impact of any spatial clustering of the wells used for each analysis (Figure 39). This area-weighted technique of viewing the data helps determine if the arithmetic mean is influenced by many wells clustered together. If the arithmetic mean and area-weighted mean diverge significantly, poor well distribution may be indicated. The mean area-weighted hydraulic conductivity for slug tests in bedrock was 1.77 feet per day and 3.99 feet per day for wells included in the drawdown analysis. The arithmetic mean hydraulic conductivity for slug tests was 1.70 feet per day and 3.27 feet per day for the drawdown analysis, which indicated good well distribution for both datasets—the area-weighted mean and arithmetic mean were similar.

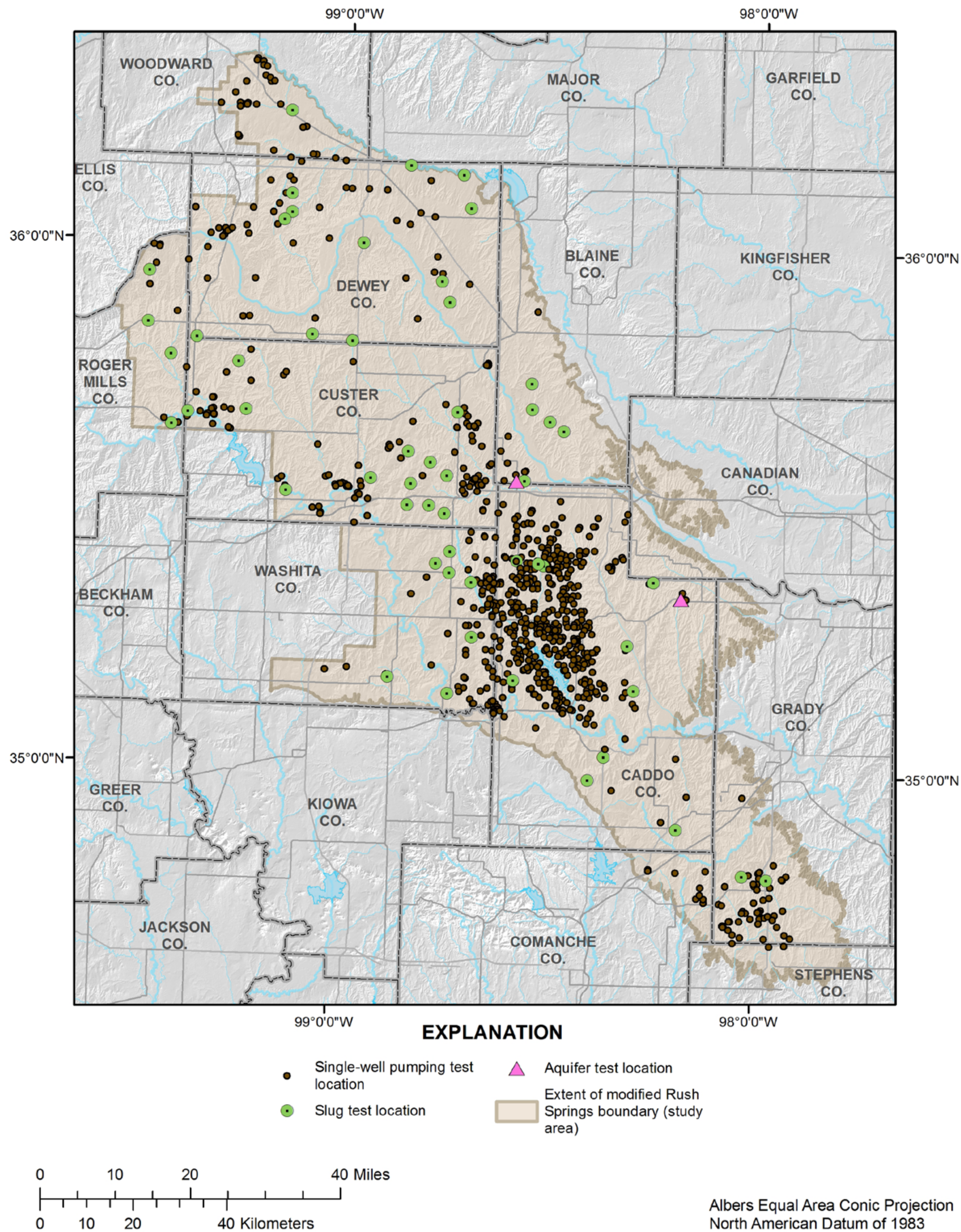

Figure 37. Map showing locations of wells with available drawdown data in the OWRB Drillers Database and where slug tests were performed in the Rush Springs aquifer as part of this study.

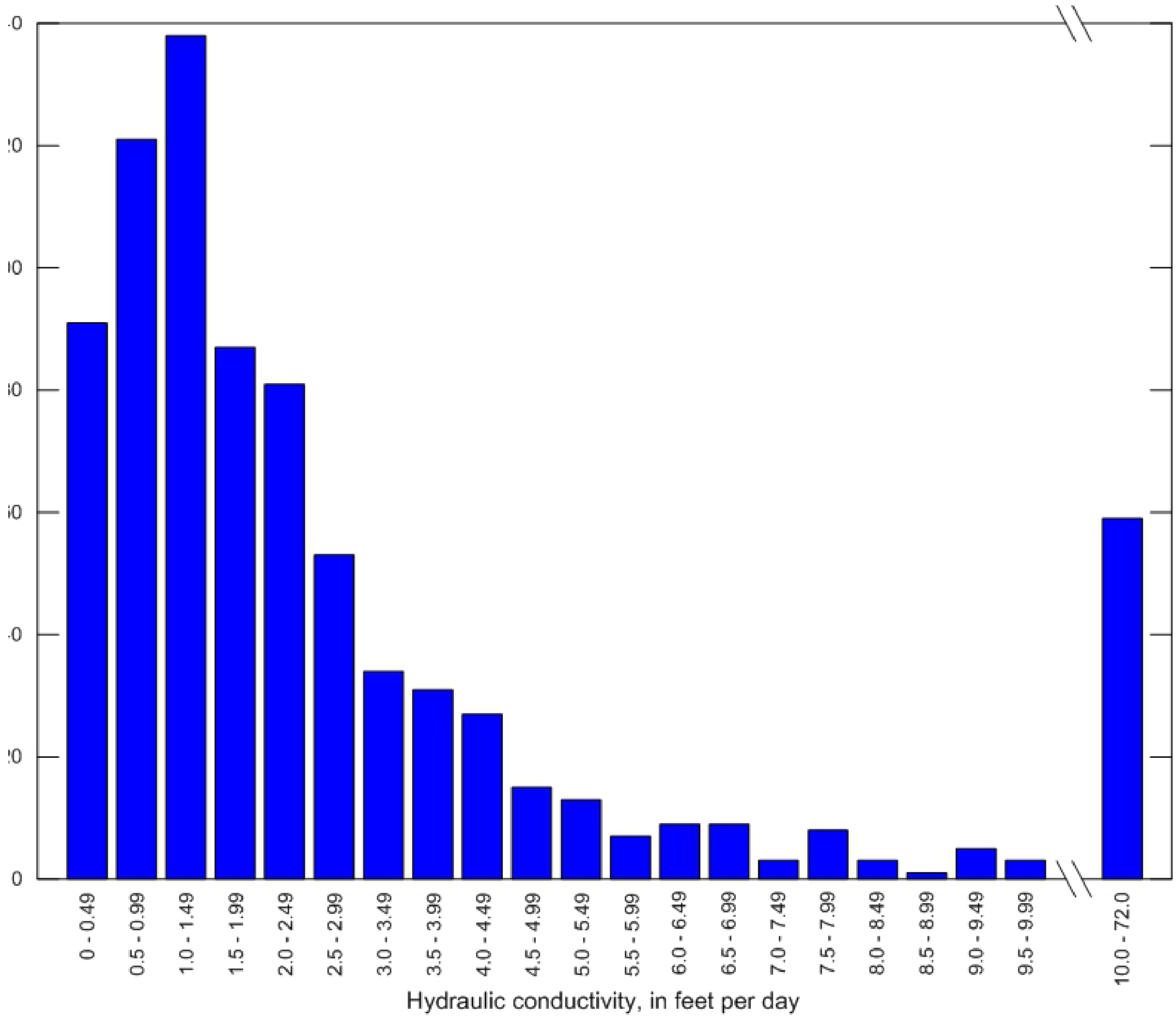

Figure 38. Histogram showing the hydraulic conductivity distribution of slug tests and drawdown data.

Table 10. Summary statistics show the count, minimum, maximum, mean, 25th percentile, 50th percentile, 75th percentile, and area-weighted mean values for hydraulic conductivity, in feet per day, derived from slug tests, drawdown analysis, and percent coarse analysis.

Statistic	Slug Tests	Drawdown Analysis	Percent Coarse
Count	52	688	4493
Minimum	0.13	< 0.01	< 0.01
Maximum	7.60	90.90	75.00
Mean	1.70	3.27	6.30
25th Percentile	0.66	0.90	2.75
50th Percentile	1.40	1.63	4.00
75th Percentile	2.18	3.01	7.95
Area-weighted	1.77	3.99	6.37

*Data do not include 1992 report.

Aquifer Tests

For this study, multi-well aquifer tests were performed on public water supply wells operated by Grady County Rural Water District #6 and the town of Hydro. The data produced from the aquifer tests were used to estimate transmissivity, hydraulic conductivity, and storage of the Rush Springs aquifer. All depth-to-water measurements were collected at 1 minute intervals.

Grady County Rural Water District #6

The Grady County Rural Water District #6 production well (OWRB 151469) was completed with a 12-inch casing to a depth of 180 feet below land surface and sealed to a depth of 130 feet. The production well construction information was not available from Grady County Rural Water District #6 and was presumed to be screened in the bottom 100 feet of the well, similar to other wells in the area.

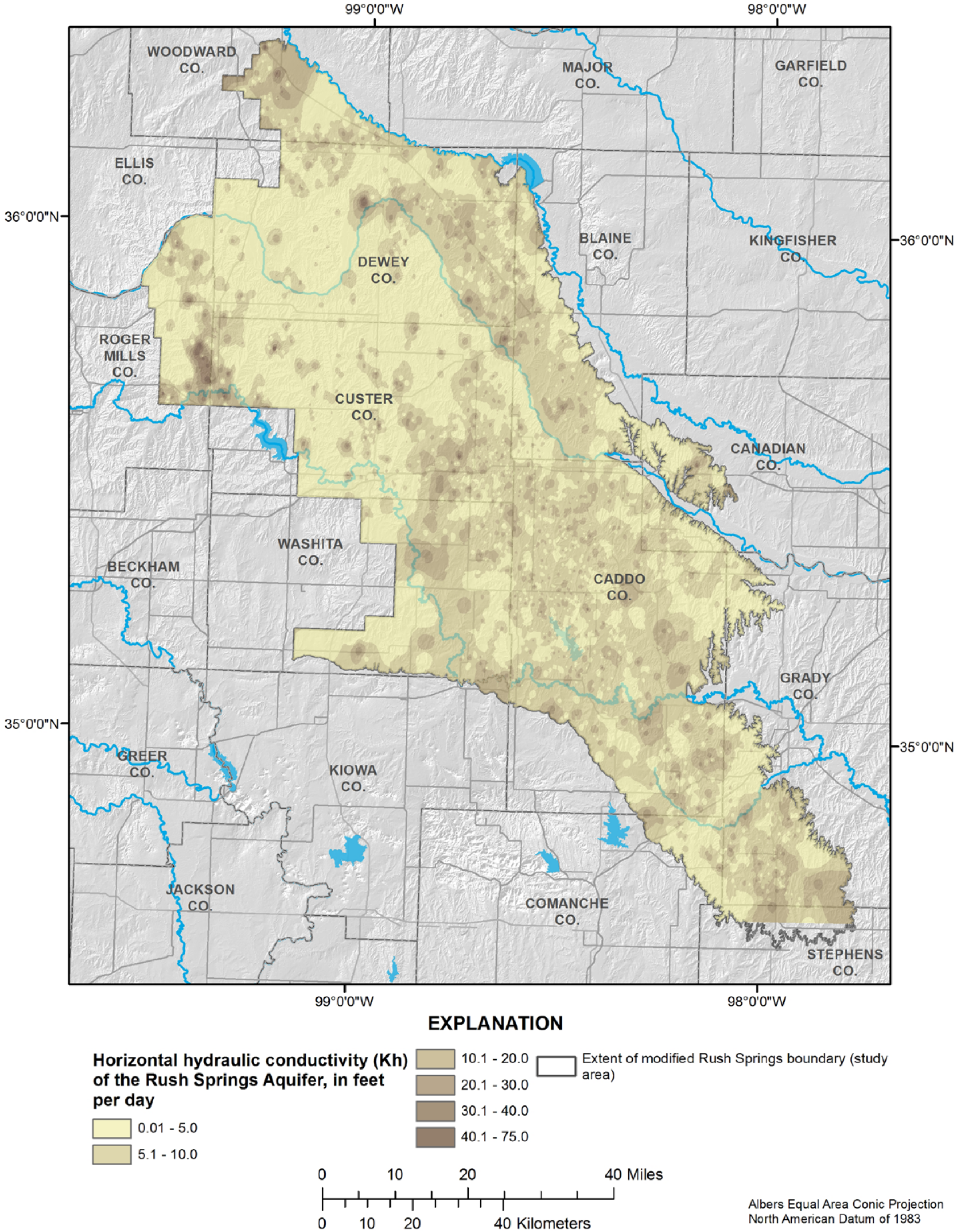

Figure 39. Horizontal hydraulic conductivity of the Rush Springs aquifer based on percent-coarse analysis of lithologic descriptions in over 4,700 well logs.

The observation well (OWRB 3952) had a 6-inch casing, was located 18.2 feet from the production well, and was completed as a 16-inch open hole to a depth of 185 feet below land surface. The production well was turned off prior to the test to allow for water to recover to static levels. The well began pumping at 12:19 PM on January 28, 2014, at an initial rate of 564 gallons per minute. During the test, the pumping rate gradually decreased from 564 gallons per minute to 525 gallons per minute. The pump ran for approximately 38 hours and was shut off at 2:14 AM on January 30, 2014. Maximum water-level displacement in the observation well during pumping was 22.20 feet. Water levels in the observation well were collected during recovery from January 30 through February 3, 2014 (Figure 40).

The aquifer test data were analyzed using the AQTESOLV software package (Duffield, 2007). Several unconfined and confined model curve matching solutions were tested for the pumping period. The curve matching solution with the best fit was the Moench for unconfined aquifer solution (Moench, 1997). The Moench solution for the pumping period estimated transmissivity to be 4,129 square feet per day, hydraulic conductivity was 44.9 feet per day, and the specific yield was 0.04 (Figure 41).

Town of Hydro, Oklahoma

The town of Hydro production well (OWRB 173538) was completed to a depth of 280 feet below land surface (oral communication with Hydro officials, 2016) with a 12-inch casing. Complete construction information for the production well was not available from Hydro officials and was presumed to be screened in the bottom 100 feet of the well. The observation well (OWRB 90884), located 395 feet from the production well, was completed to a depth of 255 feet, and sealed to a depth of 20 feet. The observation well was open hole from 20 to 255 feet with a casing diameter of 6 inches.

The production well was shut off at 7:42 AM on September 30, 2014, to allow water levels to return to static conditions. The production well began pumping at 8:34 AM on October 1, 2014, at a rate of 80 gallons per minute for about 30.3 hours until the pump was shut off at 2:55 PM on October 2, 2014. Maximum water-level displacement in the observation well during pumping was 13.00 feet. Water levels in the observation well were collected from October 2, 2014, through October 3, 2014, and 12.59 feet of recovery were recorded (Figure 42).

The aquifer test data were analyzed using the AQTESOLV software package (Duffield, 2007). Several unconfined and leaky-confined model curve-matching solutions were tested for the pumping and recovery period. The solution with the best visual curve match for the pumping period was the Moench solution for unconfined aquifers solution (Moench, 1997) (Figure 43). Estimated transmissivity from the Moench solution for the drawdown and recovery periods was 219 square feet per day, hydraulic conductivity was 1.60 feet per day, and specific yield was 0.09.

Re-analysis of 1955 Single Well Aquifer Test in Southern Grady County, Oklahoma

A single well pumping test was conducted by the OGS in 1955 in Section 3, Township 4N, Range 7W in southern Grady County, Oklahoma, in the southeastern portion of the Rush Springs Formation, about 2,000 feet from the contact between the Rush Springs and Marlow Formations (Davis, 1955). The test was analyzed using the Theis formula for estimating the drawdown at any place in the aquifer at any time for any rate of continuous pumping (Davis, 1955). The OGS report includes the drawdown and pumping information that allowed for the data to be imported into AQTESOLV and analyzed for transmissivity and specific yield.

The test was performed on a well owned by the Magnolia Petroleum Company (4N7W-3-1) that was drilled to 500 feet and plugged back to 122 feet below land surface. A 20-inch diameter casing was set to 72 feet and a 19-inch diameter hole was reamed to 122 feet. The well is perforated from 72 to 120 feet below land surface. The well fully penetrated the Rush Springs aquifer. The water level in the well was initially about 50 feet below land surface. The well was pumped continuously for 24 hours at an average rate of 163 gallons per minute. Water level declined by 96 feet during the test and then recovered for 24 hours to a level of 52.5 feet below land surface. The data were analyzed model using the Moench (1997) solution for an unconfined aquifer. Transmissivity was estimated at 956.1 square feet per day, hydraulic conductivity was 6.4 feet per day, and specific yield was 0.09.

Re-analysis of 1956 Multi-Well Aquifer Test Near Sickles, Oklahoma

The OGS conducted a 2-week multi-well aquifer test on the Rush Springs Formation in 1956 that included an irrigation well located in Section 23, Township 10N, Range 12W in Caddo County (Tanaka and Davis, 1963). The test involved 9 observation wells, of which 3 had accessible water levels to be analyzed. Observation wells were located 200, 600, and 1,085 feet from the pumped well. Water-level data from this test were digitized and re-analyzed using AQTESOLV to update the results.

The irrigation well was pumped at a constant rate of 730 gallons per minute from April 8 to 14, 1956, while depth-to-water measurements were taken with continuous automatic water-level recorders and steel tapes. The pumped well was drilled to 178 feet and had a diameter of 2 feet 8 inches. The well was perforated to the bottom 160 feet below land surface. The well partially penetrated the Rush Springs aquifer, which had an estimated base of 270 feet below land surface. The static depth to water in the pumped well was 56 feet below land surface and reached a maximum level of 142

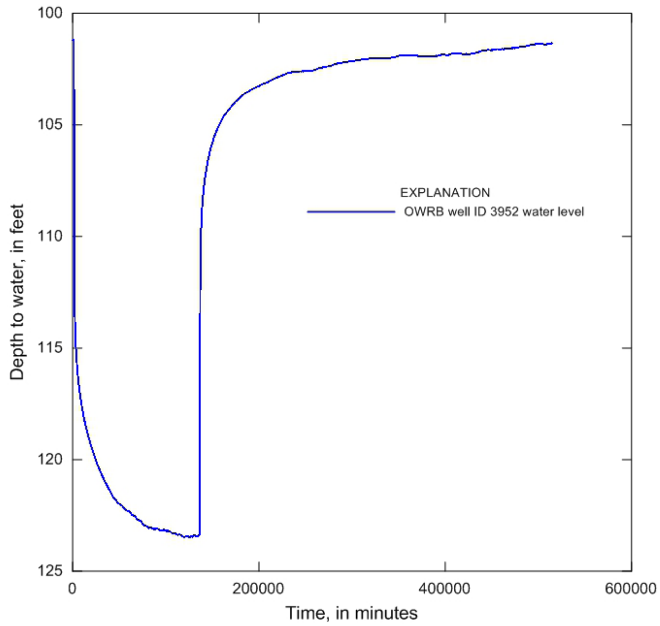

Figure 40. Graph showing water levels during the pumping (January 28–30, 2014) and recovery periods (January 30–February 3, 2014) of the Grady County Rural Water District #6 aquifer test in the Rush Springs aquifer.

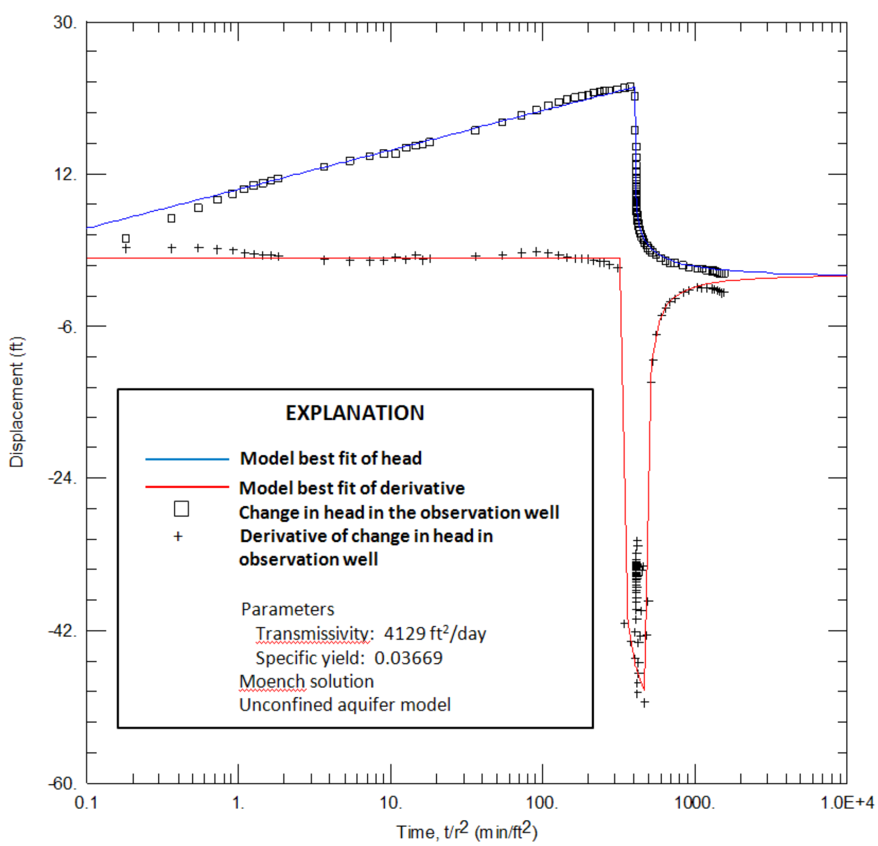

Figure 41. Pumping and recovery data curve and derivative of the Grady County Rural Water District #6 aquifer test with best-fit Moench solution for leaky confined aquifers (Moench, 1997).

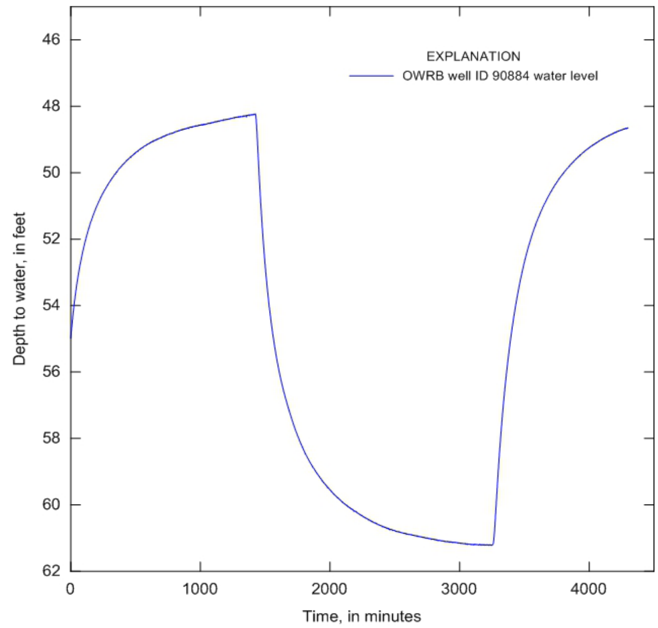

Figure 42. Graph showing water levels during the pumping (October 1–2, 2014) and recovery periods (October 2–3, 2014) of the Town of Hydro aquifer test in the Rush Springs aquifer.

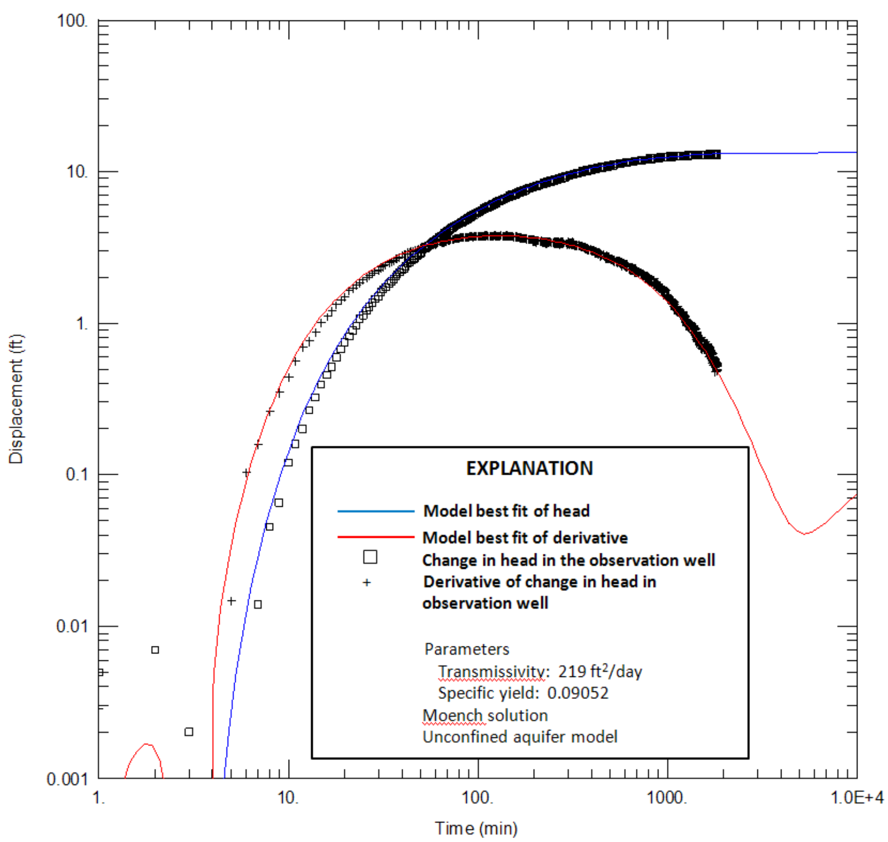

Figure 43. Pumping drawdown data curve and derivative of the town of Hydro aquifer test with best-fit Moench solution for unconfined aquifers (Moench, 1997).

feet below land surface. The data were analyzed using the Moench solution for an unconfined aquifer. Transmissivity was estimated at 1,225.4 square feet per day, hydraulic conductivity was 5.4 feet per day, and specific yield was 0.07. The OGS report estimated transmissivity between 1,470 and 1,870 square feet per day with a storage value of 0.01 to 0.03. Using modern estimation methods such as AQTESOLV allows for a more precise estimation of aquifer properties than the curve matching techniques available in the early 1960s.

Aquifer Test Discussion

For many of the existing production wells in the Rush Springs aquifer, screen length could not be determined. For these wells, researchers assumed a screen length of 100 feet, which is typical for other similar wells in the aquifer. Each solution was tested for sensitivity to screen length by changing the well construction in AQTESOLV to open borehole. Changing the screen length did not affect the solutions.

Regional Method to Determine Storage Coefficient

Regional methods can be used to characterize aquifers using hydrologic data at large scales. The regional method described in this report uses base flow discharge and monthly groundwater-level measurements between November 12, 2013, and March 24, 2014, to estimate storage in the Cobb Creek, Deer Creek, and Lake Creek watersheds within the Rush Springs aquifer. Storage coefficients estimated by this method are considered an average value for each subsurface watershed because the data are spatially distributed. The regional method to calculate specific yield assumes that if an aquifer is not being recharged during a specific time, but is only draining, the ratio of the volume of groundwater discharged to the volume of the aquifer drained is the storage coefficient for that volume of aquifer drained (Christenson and others, 2011). The limitation of this method is that it only estimates the storage in the portion of the aquifer that was drained. While multiple well pumping tests provide defensible estimates of storage coefficient and specific yield, those values are local to the area of influence around the well. The regional method has been used to provide an estimate of storage over an entire watershed within an aquifer (Schilling, 2009; Christenson and others, 2011).

An equation to calculate the storage value can be derived from the concept that base flow is often considered a proxy for diffuse recharge in watersheds with gaining streams (Scanlon and others, 2002; Risser and others, 2005). The water-table fluctuation method of calculating recharge (Healy and Cook, 2002), defines a relationship between specific yield, the change in the water table over time, and recharge. If base flow is substituted for recharge and little to no recharge occurs, then the base flow in a stream is the water being released from storage, and an equation can be written as follows:

$$Sy = Qb/\Delta DTW$$

where

Sy is the specific yield (dimensionless)

Qb is the amount of base flow during a set period of time

ΔDTW is the change in the depth to water over a set period of time

This method was used during the winter of 2013–2014, a period where major groundwater use had ended for the year and little precipitation had occurred. Precipitation on the central portion of the Rush Springs aquifer as measured by the Weatherford Mesonet weather station during the time between the first and last synoptic measurements was 3.1 inches (Oklahoma Mesonet, 2014), which is less than the average value of 6.9 inches for the time period of 1994–2014; a low of 0.03 inches was recorded in January 2014, and a high of 1.47 inches was recorded in March 2014. Groundwater levels were not influenced by precipitation during this period, as observed in nearby groundwater wells equipped with water-level recorders (Figure 44). Streamflow hydrographs for these 3 gauges indicated small increases in daily flow from precipitation, but the base flow index (the ratio of the groundwater to runoff in the stream discharge) computed by using the BFI program was 89 to 93 percent, indicating most of the streamflow was groundwater. Only the base flow portion of streamflow was used for this method.

Discharge was measured at streamflow gauges at Cobb Creek near Eakly (USGS 07325800), Deer Creek at Hydro (OWRB 520620060010-003RS), and Lake Creek near Eakly (USGS 07325850) during the time period of the synoptic water-level measurements. With the contributing groundwater area upgradient, these streamflow gauges defined the lower reaches for each subsurface watershed. Willow Creek and the Little Washita River were not utilized for this method because the low number of water-level measurements in those subsurface watersheds did not provide an adequate density of data to extrapolate water levels across the entire watershed. Monthly synoptic water-level measurements were collected between November 12, 2013, and March 24, 2014, and water-level maps were created for the Cobb Creek (127.3 square miles), Deer Creek (280.1 square miles), and Lake Creek (58.9 square miles) subsurface watersheds. To estimate the total volume of groundwater gained or lost from each subsurface watershed, the boundaries of each subsurface basin were determined using a combination of topographic maps and the potentiometric surface contours created from the March 2013 synoptic well measurements (Figure 15).

Groundwater levels rose slightly between November and December 2013, which is likely a result of aquifer recovery from late season irrigation and some recharge from

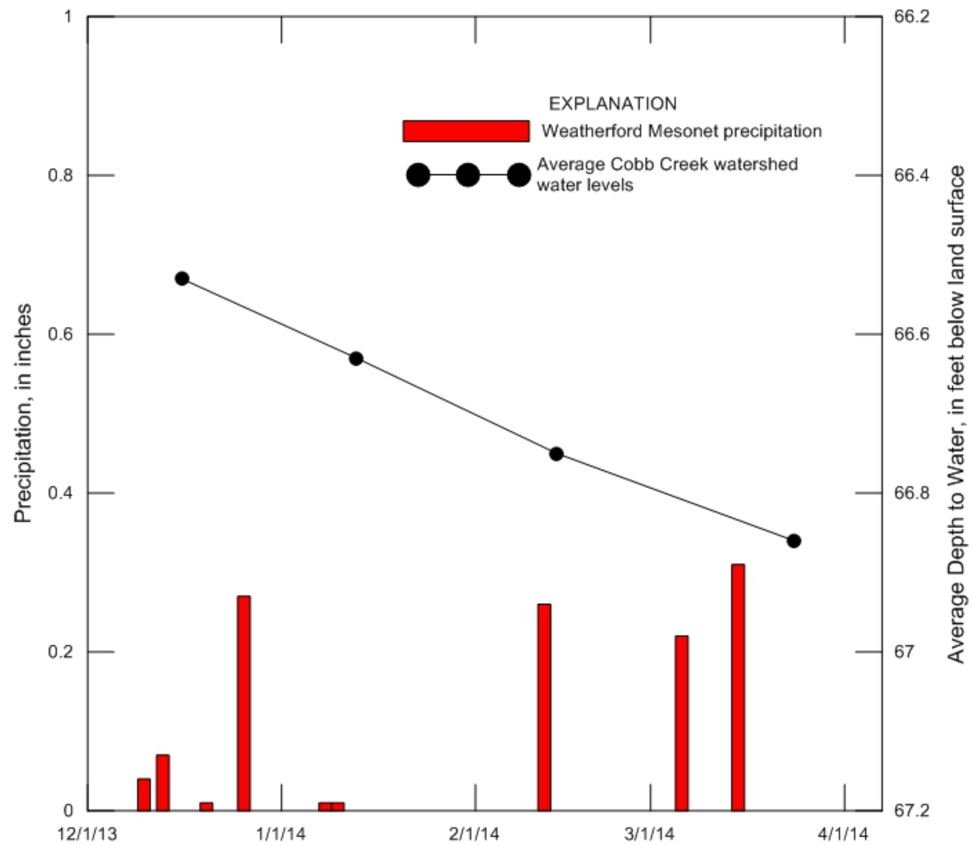

Figure 44. Graph of water levels in the Cobb Creek subsurface watershed showing no influence from precipitation from December 2013 through March 2014.

an early November rain event, and were not used for the storage calculation. Groundwater levels declined between the December 16, 2013, and March 24, 2014, synoptic measurements, which were utilized for storage estimates. During this period, a total of 1.09 inches of precipitation was recorded at the Weatherford Mesonet site, the nearest climate station to the groundwater basins used in this method. In those 99 days, there were 7 days with measurable precipitation of at least 0.01 inch, and 4 days with more than 0.01 inch. The largest event was 0.31 inch of precipitation on March 15, 2014. Water levels and base-flow hydrographs in the basins reflect little to no impact from these events and the assumption is that precipitation either exited the system as runoff or did not diffuse down to the aquifer as recharge.

The volume of aquifer drained in each subsurface watershed between each groundwater-level measurement was estimated using ESRI's ArcGIS desktop software. The volume of water drained was the base-flow component as computed by the PART program for each streamflow gauge. Dividing the volume of water drained by the volume of aquifer drained gave monthly storage coefficient estimates for each basin and ranged from 0.047 to 0.090 (Table 11). Using the total volume of aquifer drained and the total base flow discharged between December 2013 and March 2014, an average storage coefficient of 0.049 was estimated for the Cobb Creek subsurface watershed, 0.072 for the Deer Creek subsurface watershed, and 0.065 for the Lake Creek subsurface watershed. The lower storage estimated for the Cobb Creek subsurface watershed may be attributed to the upper watershed being regulated by Crowder and Worth Richmond Lakes, which would capture base flow from the upper reaches of the watershed and effectively lower the storage estimation. The range of storage values estimated by this method was of similar magnitude to the range of previous values calculated for the Rush Springs aquifer, with results from this study ranging between 0.04-0.09. Previous work had specific yield from core samples ranging between 0.13 and 0.34, and storage coefficients from pumping tests near Weatherford ranged between 0.0035 and 0.02 (Becker and Runkle, 1998).

Percent-Coarse Analysis

Another method used to determine the hydraulic conductivity of the Rush Springs aquifer was the percent-coarse analysis, which uses lithologic descriptions included in water well logs submitted by groundwater well drillers to the OWRB. This method has been utilized for bedrock and unconsolidated aquifers in Oklahoma (Mashburn and others, 2013; Ellis and others, 2017) and has given reasonable estimates of hydraulic conductivity and storage. The majority of the lithologic logs used in this analysis penetrated the entire depth of the aquifer, with some penetrating below the base of the aquifer.

Lithologic descriptions of about 5,700 well logs were standardized to 16 simplified categories (Table 12). The three categories that accounted for about 59 percent of all descriptions were medium sandstone, clay, and medium sand. Cumulatively, all sandstone categories accounted for

Table 11. Storage coefficients calculated from streamflows and change in water stored in subsurface watersheds, December 2013 through March 2014.

Subsurface watershed	Basin size (square miles)	Dates Measured	Total baseflow discharge based on daily gauged flow (acre-feet)	Volume of aquifer drained in subsurface watersheds (acre-feet)	Storage coefficient (dimensionless)	Combined storage coefficient (dimensionless)	Average water level decline (feet)
Cobb Creek	127.3	Dec. 2013 - Jan. 2014	477	12,932	0.037	0.05	0.16
		Jan. 2014 - Feb. 2014	519	11,042	0.047		0.14
		Feb. 2014 - Mar. 2014	802	12,816	0.063		0.16
Deer Creek	280.1	Dec. 2013 - Jan. 2014	1321	14,620	0.090	0.07	0.08
		Jan. 2014 - Feb. 2014	1537	28,048	0.055		0.16
		Feb. 2014 - Mar. 2014	1652	20,060	0.082		0.11
Lake Creek	58.9	Dec. 2013 - Jan. 2014	115	1,697	0.068	0.07	0.05
		Jan. 2014 - Feb. 2014	143	3,029	0.047		0.08
		Feb. 2014 - Mar. 2014	191	2,176	0.088		0.06

Table 12. Standardized lithologic categories and estimated hydraulic conductivity and storage from lithologic logs in the Rush Springs aquifer and Cloud Chief Formation.

Lithology	Lithologic distribution: Rush Springs Formation	Lithologic distribution: Cloud Chief Formation		Conductivity, in feet per day	Specific yield
Clay	14%	22%	Clay	*6x10^{-4}	[e]3%
Shale	5%	11%			
Siltstone	1%	4%			
Claystone	0.10%	1%			
Silt	5%	7%	Silt	*0.06	[e]5%
Fine sandstone	8%	8%			
Gypsum	3%	11%			
Anhydrite	0%	0.50%			
Limestone	0%	0.10%			
Fine sand	6%	4%	Fines	[+abc]4	[abce]8%
Medium sandstone	30%	12%			
Medium sand	16%	7%	Sand	*[d]30	[e]12%
Coarse sandstone	1%	0.20%			
Topsoil	9%	11%			
Coarse sand	1%	1%	Coarse	[f]60	[e]25%
Gravel	0%	1%	Gravel	[f]90	[e]25%

*Morrison and Johnson, 1967
[+] Becker and Runkle, 1998
[a] Aquifer test from Town of Hydro
[b] Tanaka and Davis, 1963
[c] Davis, 1955
[d] Aquifer test from Grady County Rural Water District #6
[e] Johnson, 1967
[f] Contained less than 3 percent of lithologic description and had minimal effect on conductivity for the aquifer

35 percent of the lithologic descriptions. The medium sand description sometimes occurred between sandstone lithologies at or below the water table, which may be caused by the dissolution of calcite and gypsum cement. Lithologic logs that included descriptions accounting for less than 30 percent of the total thickness of the Rush Springs aquifer at each wellhead were removed. Additionally, lithologic logs with only a single lithologic description, logs with inconsistencies in recorded depths, incorrect lithologies determined from comparisons with colocated wells, and unidentifiable lithologic descriptions were discarded. Lithologic logs completed in the alluvium and terrace deposits were removed from this analysis. About 4,900 lithologic logs were used to describe the Rush Springs Formation and 800 lithologic logs were used for the Cloud Chief Formation, which equates to about one log per 0.98 and 0.63 square miles, respectively. Each lithologic log was used to estimate the hydraulic properties where the well was located.

The 16 simplified lithologies were grouped into 6 categories that represented material of the Rush Springs aquifer and overlying Cloud Chief Formation (Table 12), which allowed comparison across the aquifer and the assignment of more streamlined hydraulic parameters. Hydraulic conductivity and specific yield values were assigned to each lithologic group based on results from aquifer tests, slug tests, and single-well drawdown tests, as well as values found in literature. This method was expected to provide estimations of hydraulic conductivity and specific yield that encompasses the majority of grain sizes encountered in the Rush Springs aquifer and Cloud Chief Formations. For the Rush Springs aquifer, the average and median hydraulic conductivity was 6.30 and 4 feet per day (Table 10), respectively, with higher hydraulic conductivity values in the eastern portion of the aquifer and lower values in the western portion (Figure 39). The average and median specific yield were 0.07 and 0.08, respectively. For the Cloud Chief Formation, the average and median hydraulic conductivity was 1.6 and 0.9 feet per day, respectively. Both the average and median specific yields were 0.06. The area-weighted average hydraulic conductivity of the Rush Springs aquifer was 6.37 (Table 10), based on an interpolated raster of all hydraulic conductivity values created using the IDW method in ArcGIS.

Groundwater Quality

The quality of the water from the Rush Springs aquifer has been described as fair to good, very hard, and moderately alkaline (Oklahoma Water Resources Board, 2014b) with the most common water types being calcium bicarbonate and calcium sulfate (Becker and Runkle, 1998). Groundwater in the high use areas of the aquifer (Caddo County), where the Rush Springs Formation is exposed at land surface, typically has the highest quality, which can be attributed to more precipitation recharge. To the west, where the Cloud Chief Formation overlies the Rush Springs Formation or has been partially eroded, total dissolved solid concentrations increase along with higher magnesium concentrations. Specific conductance measurements of the groundwater collected as part of this study using a Solinst model 107 Temperature-Level-Conductivity meter showed the same trend of increasing conductivity where the Cloud Chief Formation is present.

Groundwater Monitoring and Assessment Program (GMAP)

The OWRB's Groundwater Monitoring and Assessment Program (GMAP) staff collected water quality samples from 64 well sites in 2013 (Oklahoma Water Resources Board, 2014b). Another sample collected at that time was included in this analysis because the well penetrates the Rush Springs Formation. Additional samples were collected as part of this investigation from 14 well sites to achieve a better spatial distribution across the study area. Figure 45 shows a Piper diagram and Table 13 shows summary statistics of data collected from 79 well sites.

Mean total dissolved solids from all samples was 1,106 milligrams per liter and ranged from 178 to 4,680 milligrams per liter with a median of 485 milligrams per liter. Figure 46 is a map showing spatial distribution of total dissolved solids in the study area with a noticeable east-to-west trend of higher concentrations where the Cloud Chief Formation is present to lower concentrations where the Rush Springs Formation is exposed at land surface. There is also an area near the towns of Cyril and Cement and to the southeast toward the town of Rush Springs that has higher total dissolved solid concentrations that may be attributed to

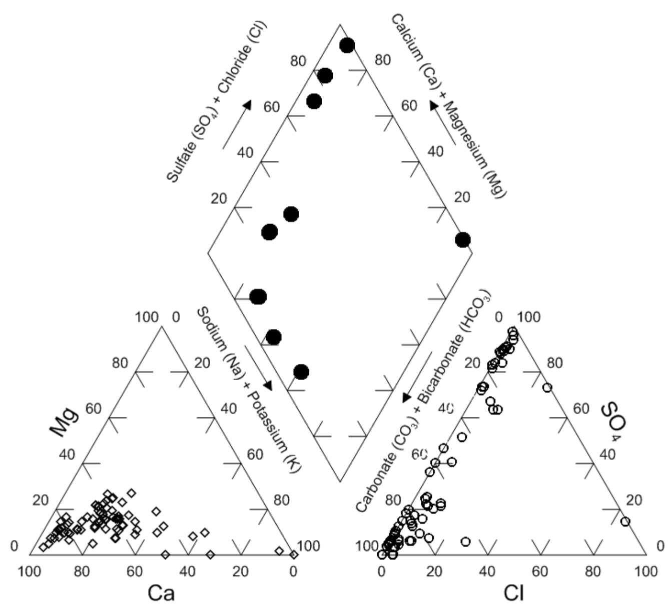

Figure 45. Piper diagram showing groundwater geochemistry from 79 samples collected in the study area.

Table 13. Summary statistics for groundwater-quality data for 79 samples collected from the Rush Springs aquifer.

Constituent	Mean	Minimum	Maximum	Number of samples below detection limit	Percentile		
					25	50	75
specific conductance	1,331	321	5,830	0	526	763	2365
Temperature	19.7	15.2	24.3	0	18.8	19.6	20.6
pH	7.17	6.41	7.82	0	7.04	7.18	7.28
total dissolved solids*	1106	178	4680	0	313	485	2155
Hardness*	632.4	144	2,050	0	198	297	936
Calcium*	214.4	32.2	556	0	61.2	94.9	454.5
Magnesium*	51.2	<5	1,114	2	13.75	21	55.9
Sodium*	47.7	8.4	890	0	20.3	26.6	43.8
Potassium*	1.7	<0.5	6	2	1	1.3	2.3
Bicarbonate*	218.6	30.5	473	0	170	216	267
Sulfate*	572	<10	2,300	6	21.5	73.3	1310
Chloride*	31.2	<10	812	26	<10	13	26.6
Fluoride*	0.21	<0.2	0.52	39	<0.2	0.21	0.52
Bromide**	279.4	121	1,200	0	197	248	319.5
Silica**	27.7	11.4	53	0	24.6	26.9	30.3
Nitrate as N*	6.6	0.24	59.2	0	1.5	3.9	7.3
Phosphorous**	0.02	<0.005	0.22	53	++	++	0.02
Aluminum**+	++	<100	++	79	++	++	++
Arsenic**+	++	<10	16.5	75	++	++	++
Barium**+	127.1	<10	859	16	12.4	81.6	180.3
Boron**+	169.2	<50	1,200	31	++	77.5	211.5
Cadmium**	++	<5	++	79	++	++	++
Chromium**	++	<5	23.7	64	++	++	++
Copper**	++	<5	15.5	72	++	++	++
Iron**+	++	<50	435	73	++	++	++
Lead**+	++	<10	19.7	78	++	++	++
Manganese**+	++	<50	60	78	++	++	++
Molybdenum**+	++	<10	26	77	++	++	++
Uranium**	6.6	<1	61.2	15	1.4	3.4	6.3
Vanadium**+	14.7	<10	40.2	22	++	13.7	18.5
Zinc**+	18.3	<10	299	53	++	++	15.33

++, analyses were below analytical detection limit and statistics could not be estimated

+includes analysis of samples with different detection limits

Specific conductance is in microseimens per centimeter at 25° C

*presented in milligrams per liter

**presented in micrograms per liter

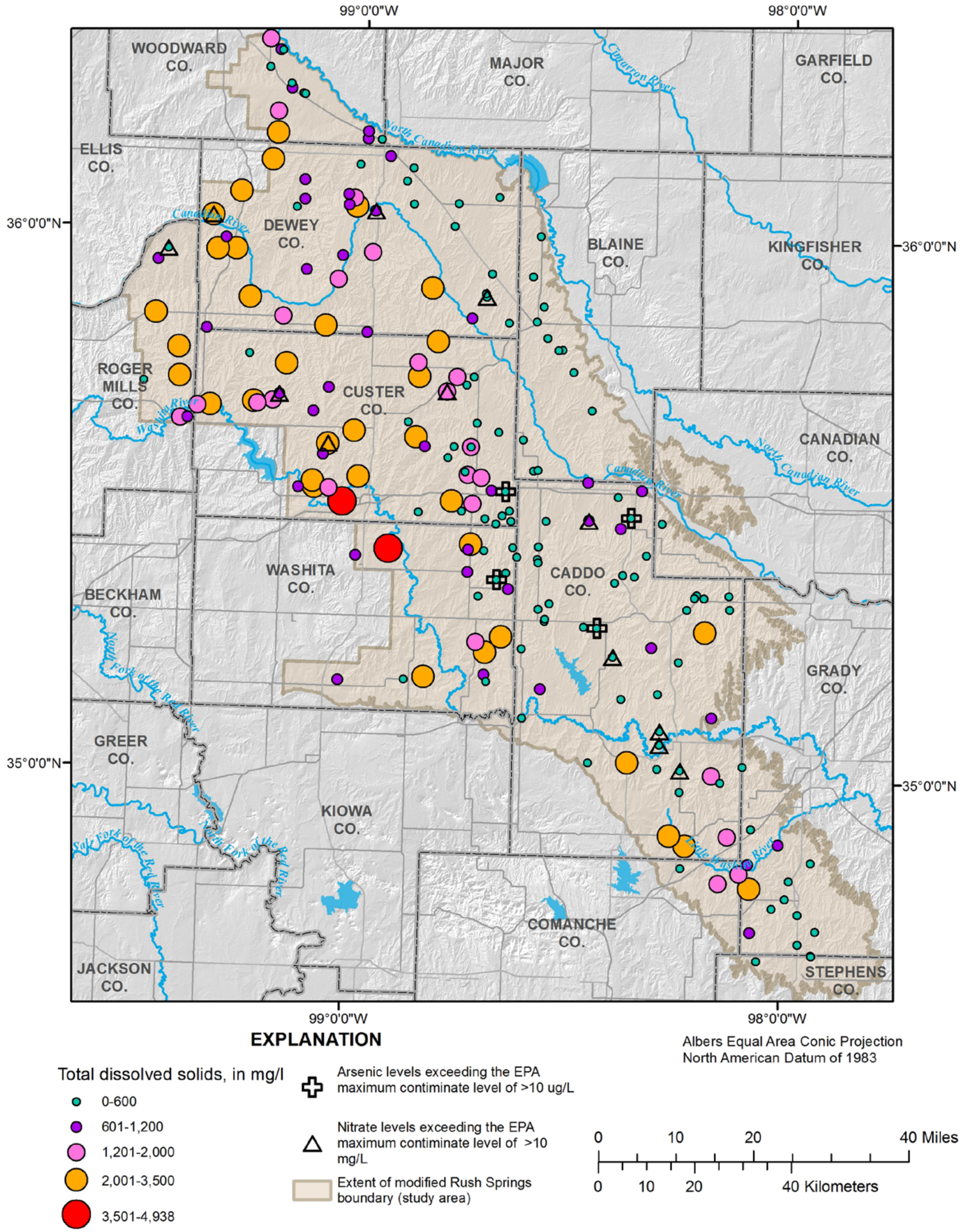

Figure 46. Map showing distribution of total dissolved solids and wells exceeding the EPA maximum contaminant levels for arsenic and nitrate in the study area.

the alteration of the cement in the Rush Springs Formation (Allen, 1980).

The dominant cation type was calcium; 74 samples had more than 60 percent calcium, which was likely attributable to the dissolution of calcite, dolomite, and gypsum from diffuse precipitation recharging the aquifer. Calcium concentrations ranged from 32 to 556 milligrams per liter with a mean of 214.4 and a median of 94.9 milligrams per liter. The anion distribution was bimodal with carbonate/bicarbonate as the dominant anion and a secondary population of sulfates. Mean bicarbonate content was 219 milligrams per liter, ranging from 30.5 to 473 milligrams per liter with a median of 216 milligrams per liter. Mean sulfate content was 572 milligrams per liter, ranging from <10.0 to 2,300 milligrams per liter with a median of 124 milligrams per liter. The low median sulfate concentration compared to the mean indicated that a majority of the samples had lower sulfate concentrations and that a small distribution of samples had higher sulfate content. The bicarbonate/sulfate distribution could be an indication that characteristics of the groundwater flow system control the groundwater chemistry of the aquifer. Since gypsum breaks down easily into calcium and sulfate ions, the expected trend is to see higher concentrations of calcium and sulfate where the Cloud Chief Formation overlies the aquifer, or also in areas where gypsum units are within the Rush Springs aquifer.

Some of the groundwater samples showed constituent concentrations that exceeded maximum contaminant levels for primary drinking water regulations (Figure 46). Four of the samples exceeded the maximum contaminant level (MCL) for arsenic of 10 micrograms per liter; the highest concentration of arsenic sampled was 16.5 micrograms per liter. Previous investigations have examined the chemical concentrations in the groundwater and sources of the arsenic (Becker and Runkle, 1998; Becker and others, 2010, Haggard and others, 2003; Magers, 2011). Water quality samples from the 1998 USGS study indicated a statistical mean arsenic concentration of 14.9 micrograms per liter in 64 samples, which were predominantly from Caddo and Grady Counties (Becker and Runkle, 1998). However, the statistical 75th percentile was 5.2 micrograms per liter, which is below the drinking water standard. The USGS reported arsenic concentrations in the Rush Springs Formation ranging from 7.1 to 18.2 micrograms per liter (Becker and others, 2010). An x-ray fluorescence analyzer was used in 2011 to determine average arsenic concentrations in core (8.20 ppm) and outcrop (7.62 ppm) samples, which was noted to fall within the range of background samples (Magers, 2011). The mobilization of arsenic was likely caused by competing ions; phosphorous was a potential constituent due to application of fertilizers (Magers, 2011). Three of the 4 samples that exceeded arsenic standards had phosphorous concentrations below the detection limit, which may indicate that phosphate sorption had taken place. The 4 samples that exceeded the arsenic MCL also had concentrations of magnesium and calcium that were near or below the 25th percentile for both constituents.

Nitrate concentrations exceeding maximum contaminant levels can cause health issues, most notably, shortness of breath, blue baby syndrome, and fatality (Environmental Protection Agency, 2015). The USGS reported that background nitrate concentrations in groundwater are generally less than 2 milligrams per liter (Mueller and Helsel, 1996). Three major sources of nitrates were identified in the Rush Springs, and inorganic fertilizer applied to cropland was determined to be the major contributor of nitrates to the groundwater flow system (Carrell and Murray, 2012). The USGS reported nitrate concentrations of 28.1 and 31.5 milligrams per liter in two groundwater wells upgradient of a wastewater lagoon near Fort Cobb Reservoir in Caddo County (Becker, 2001). Nitrate in the wells was determined to be sourced from commercial fertilizer. The USGS reported that mean nitrate concentrations from samples in the Rush Springs aquifer were 14.3 milligrams per liter, which exceeds the maximum contaminant level of 10 milligrams per liter (Becker and Runkle, 1998). The mean from samples reported in this investigation is 6.6 milligrams per liter with a maximum of 59.2 milligrams per liter. Thirteen of the samples reported concentrations exceeding the maximum contaminant level for nitrates (Figure 46).

Statistics reported for this investigation may differ from data previously reported. The scope of this investigation includes areas farther west than the 1998 USGS study where the Cloud Chief Formation overlies the Rush Springs Sandstone, which would change the overall characterization of the water quality from the aquifer.

Summary

The Rush Springs aquifer consists of the Permian-age Rush Springs and Marlow Formations, which are described as fine-grained sandstones and siltstones with some gypsum and dolomite. The study area includes 4,692 square miles in west-central Oklahoma, underlying portions of Blaine, Caddo, Canadian, Comanche, Custer, Grady, Stephens, and Washita counties. The study area for this investigation was expanded from a 1998 study by the US Geological Survey to include two additional areas where well yields are indicative of a “major groundwater basin” as defined by the OWRB. The western boundary for this investigation was expanded further westward from previous investigations based on increasing total dissolved solid content and decreasing reported groundwater use from the aquifer. The Rush Springs and Marlow formations north of the Canadian River and south of the North Canadian River were included based on similar geological and hydrological characteristics and well yields.

The study area received an annual average of 28.20 inches of precipitation from 1905–2015. Recharge occurred through diffuse precipitation and discharges through groundwater withdrawals and streams, including Barnitz, Cobb, Deer, and Lake Creeks. Groundwater also supplies baseflow to the Canadian and Washita rivers. Recharge was estimated using the SWB code and the RORA method. Estimates using SWB for the period 1950–2015 ranged from 0.03 inches in 1963 to 4.63 inches in 2007 and an average annual recharge of 1.4 inches. RORA, which utilizes a base-flow separation technique from streamflow gauging stations, ranged from 0.46 inches in 2006 on the Little Washita River streamflow gauge near Ninnekah to 5.76 inches in 2007 on Cobb Creek streamflow gauge near Eakly. From 1946–2015, at least one station from the study area had streamflow data to estimate recharge using RORA.

Reported groundwater use from the Rush Springs aquifer for 1967–2015 averaged 69,900 acre-feet per year with a median of 62,154 acre-feet per year. During this period, 91.0 percent of reported groundwater use in the study area was for irrigation, 7.8 percent was for public water supply, and 1.2 percent was for other purposes. The highest total reported annual groundwater use was about 115,016 acre-feet in 2014 and 133,113 acre-feet in 2015, which corresponded to drought conditions during these years. In 1992, only 37,210 acre-feet was reported, which was the lowest reported use for a single year; however, the data for that year may be incomplete. The second lowest reported total use for a single year occurred in 2007 at 40,418 acre-feet. Water use trends for the period of record correspond with changing precipitation patterns, with the highest groundwater use occurring during the 2010-2015 drought period and the lowest groundwater use during the wet period in the late 1980s and early 1990s.

Annual water-level measurements collected by the OWRB since the 1950s were analyzed for long-term trends. Water-level data from 95 wells with a period of record of greater than 12 years provided enough data to assess long-term trends. Water-level trends from 54 wells were determined to primarily fluctuate with climate trends, showing declining water levels during drought periods and increasing water levels during wet periods. Data from 15 sites showed overall increasing water levels and 17 sites showed decreasing water levels; nine sites had indiscernible water levels during the period of record. Measurements at the USGS well 351308098341601 had the longest period of record in the study area and showed a decline of 37.52 feet from September 1948 to April 2015.

Lithologic descriptions from groundwater wells were used to determine the base of the aquifer. Generally most of the descriptions indicated a “red bed,” “dark red bed,” or “red shale” at the bottom of the borehole, which was interpreted to be the base of the aquifer. The contact between the Rush Springs and Marlow formations on geologic maps was used to refine the edges of the aquifer where lithologic logs were sparse; however, this caused the edges of the base of the aquifer to be at higher elevations than what was observed on the lithologic logs independently. Rock cores collected in the study area also show the Marlow Formation consisting of some coarser-grained layers capable of transmitting water that can be considered part of the aquifer system. Therefore, for this study, the Marlow Formation was included as part of the aquifer. Average saturated thickness using the 2013 potentiometric map and base of aquifer is 181 feet with a maximum thickness of 432 feet. The area of greatest saturated thickness occurs along the axis of the Anadarko Basin where the Cloud Chief Formation confines the Rush Springs aquifer.

Hydraulic conductivity was estimated from drawdown analysis, slug tests, aquifer tests, and a percent-coarse analysis from lithologic logs. The minimum hydraulic conductivity for the Rush Springs aquifer estimated from drawdown data was less than 0.01 feet per day, and the maximum was 90.90 feet per day with a median of 1.63 feet per day and a mean of 3.27 feet per day. Hydraulic conductivity estimated from slug tests ranged from 0.13 feet per day to 7.60 feet per day, with a mean of 1.70 feet per day and median of 1.40 feet per day. Hydraulic conductivity estimates from three multi-well aquifer tests were 1.60, 6.40, and 44.9 feet per day. Using lithologic logs and assigning hydraulic conductivity to lithologic descriptions, mean and median hydraulic conductivity were estimated to be 6.3 and 4.0 feet per day, respectively. Transmissivity estimates for the three multi-well aquifer tests were 219, 956, and 4,129 feet squared per day.

Specific yield was estimated from regional methods and aquifer tests. Using base flow discharge and monthly groundwater-level measurements, specific yield was estimated in the Cobb Creek, Deer Creek, and Lake Creek subsurface watersheds. For this method, the ratio of the volume of groundwater discharged to the volume of the aquifer drained is the specific yield for the aquifer drained. The specific yield estimated for Cobb Creek, Deer Creek, and Lake Creek subsurface watersheds was 0.05, 0.07, and 0.07, respectively. Specific yield estimated from three multi-well aquifer tests was 0.04, 0.07, and 0.09, which correlates with the regional method.

The mean total dissolved solids concentration from 79 samples collected from the study area was 1,106 milligrams per liter. Concentrations ranged from 178 to 4,680 milligrams per liter with a median of 485 milligrams per liter. The dominant cation of the samples is calcium and the dominant anion is carbonate/bicarbonate with a secondary bimodal population of sulfates, which were predominantly collected in areas where the Cloud Chief Formation overlies the Rush Springs Formation. Four samples exceeded the maximum contaminant level for arsenic of 10 micrograms per liter; the highest concentration of arsenic sampled was 16.5 micrograms per liter. Thirteen samples reported concentrations exceeding the maximum contaminant level for nitrates of 10 milligrams per liter; the highest concentration of nitrate sampled was 59.2 milligrams per liter.

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