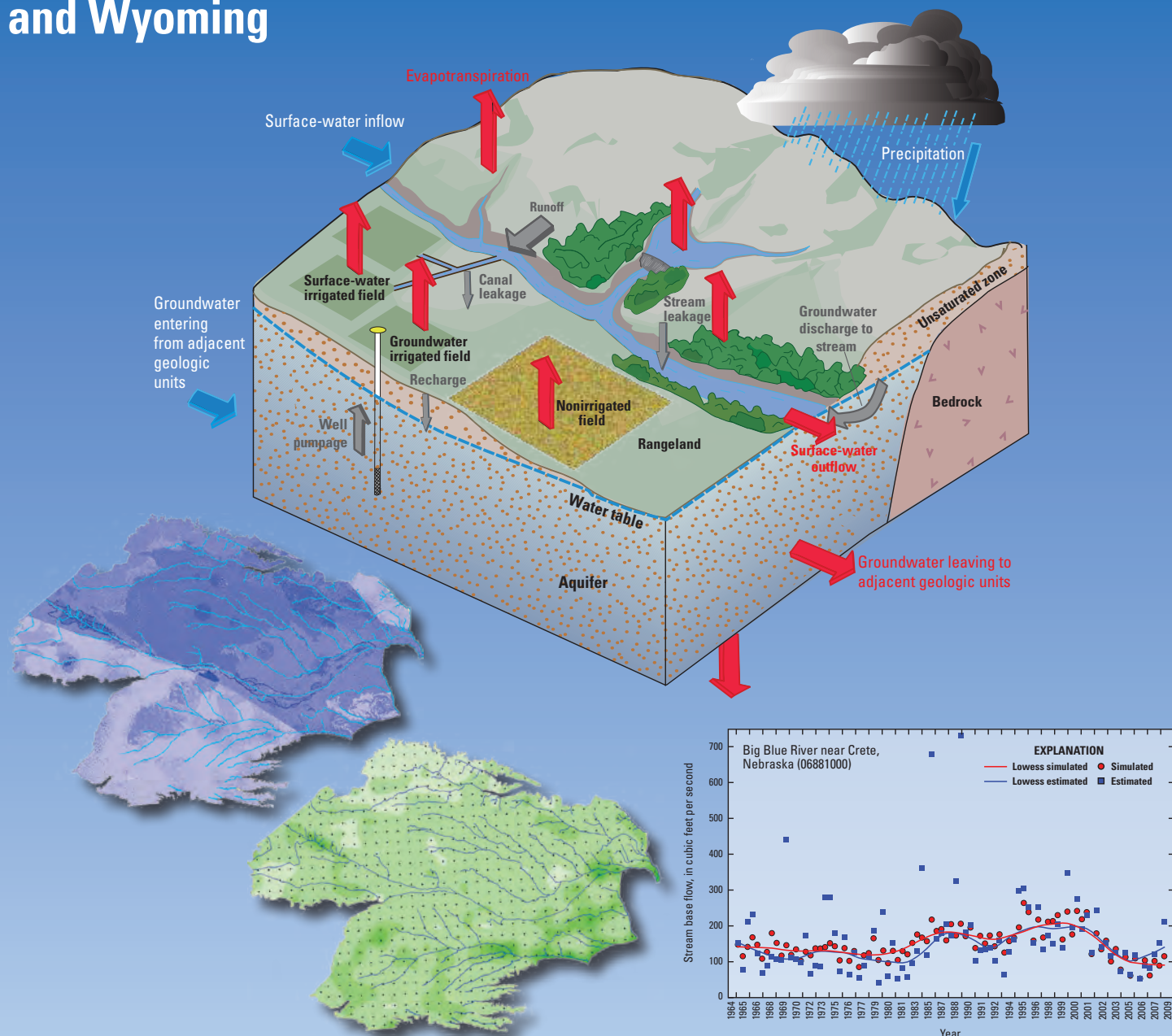


Water Availability and Use Science Program

Groundwater-Flow Model of the Northern High Plains Aquifer in Colorado, Kansas, Nebraska, South Dakota, and Wyoming



Time series of simulated and estimated base flow

Scientific Investigations Report 2016–5153

Cover. Upper illustration: block diagram showing water-budget components of the High Plains landscape and aquifer system.
Lower left illustration: Example of calibrated recharge to the Northern High Plains aquifer. Example from figure 18 of this report.
Lower center illustration: Example of calibrated horizontal hydraulic conductivity from the Northern High Plains aquifer. Example from figure 18 of this report.
Lower right illustration: Example of time series of simulated and estimated base flow. Example from figure 17*G* of this report.

Back cover. Map of the conterminous United States showing the High Plains aquifer.

Groundwater-Flow Model of the Northern High Plains Aquifer in Colorado, Kansas, Nebraska, South Dakota, and Wyoming

By Steven M. Peterson, Amanda T. Flynn, and Jonathan P. Traylor

Water Availability and Use Science Program

Scientific Investigations Report 2016–5153

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Conversion Factors

U.S. customary units to International System of Units

Multiply	By	To obtain
Length		
inch (in.)	2.54	centimeter (cm)
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
arc-second	30.87	meter (m)
arc-second	101.28	feet (ft)
Area		
acre	4,047	square meter (m ²)
acre	0.4047	hectare (ha)
acre	0.4047	square hectometer (hm ²)
acre	0.004047	square kilometer (km ²)
square mile (mi ²)	259.0	hectare (ha)
square mile (mi ²)	2.590	square kilometer (km ²)
Volume		
acre-foot (acre-ft)	1,233	cubic meter (m ³)
acre-foot (acre-ft)	0.001233	cubic hectometer (hm ³)
Flow rate		
cubic foot per second (ft ³ /s)	0.02832	cubic meter per second (m ³ /s)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
Acre-feet per year (acre-ft/yr)	0.00138	cubic foot per second (ft ³ /s)
Hydraulic conductivity		
foot per day (ft/d)	0.3048	meter per day (m/d)
Slope		
foot per mile (ft/mi)	0.3048	meter per mile (m/mi)

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

$$^{\circ}\text{C}=(^{\circ}\text{F}-32)/1.8$$

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

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

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

Abbreviations

CA	Census of Agriculture
CSFRK	calibrated streambed hydraulic conductivity
ET	evapotranspiration
HFB	Horizontal-Flow-Boundary Package
MODFLOW	modular three-dimensional finite-difference groundwater-flow model
MODFLOW–NWT	modular three-dimensional finite-difference groundwater-flow model with Newton-Rhapson solver
NWS–PET	National Weather Service potential evapotranspiration
PEST	parameter estimation suite of software
RMS	root-mean-square
SWB	Soil-Water-Balance
USGS	U.S. Geological Survey
UZF	Unsaturated-Zone-Flow Package

Groundwater-Flow Model of the Northern High Plains Aquifer in Colorado, Kansas, Nebraska, South Dakota, and Wyoming

By Steven M. Peterson, Amanda T. Flynn, and Jonathan P. Traylor

Abstract

The High Plains aquifer is a nationally important water resource underlying about 175,000 square miles in parts of eight states: Colorado, Kansas, Oklahoma, Nebraska, New Mexico, South Dakota, Texas, and Wyoming. Droughts across much of the Northern High Plains from 2001 to 2007 have combined with recent (2004) legislative mandates to elevate concerns regarding future availability of groundwater and the need for additional information to support science-based water-resource management. To address these needs, the U.S. Geological Survey began the High Plains Groundwater Availability Study to provide a tool for water-resource managers and other stakeholders to assess the status and availability of groundwater resources.

A transient groundwater-flow model was constructed using the U.S. Geological Survey modular three-dimensional finite-difference groundwater-flow model with Newton-Raphson solver (MODFLOW-NWT). The model uses an orthogonal grid of 565 rows and 795 columns, and each grid cell measures 3,281 feet per side, with one variably thick vertical layer, simulated as unconfined. Groundwater flow was simulated for two distinct periods: (1) the period before substantial groundwater withdrawals, or before about 1940, and (2) the period of increasing groundwater withdrawals from May 1940 through April 2009. A soil-water-balance model was used to estimate recharge from precipitation and groundwater withdrawals for irrigation. The soil-water-balance model uses spatially distributed soil and landscape properties with daily weather data and estimated historical land-cover maps to calculate spatial and temporal variations in potential recharge. Mean annual recharge estimated for 1940–49, early in the history of groundwater development, and 2000–2009, late in the history of groundwater development, was 3.3 and 3.5 inches per year, respectively.

Primary model calibration was completed using statistical techniques through parameter estimation using the parameter estimation suite of software with Tikhonov regularization. Calibration targets for the groundwater model included 343,067 groundwater levels measured in wells and 10,820 estimated monthly stream base flows at streamgages.

A total of 1,312 parameters were adjusted during calibration to improve the match between calibration targets and simulated equivalents. Comparison of calibration targets to simulated equivalents indicated that, at the regional scale, the model correctly reproduced groundwater levels and stream base flows for 1940–2009. This comparison indicates that the model can be used to examine the likely response of the aquifer system to potential future stresses.

Mean calibrated recharge for 1940–49 and 2000–2009 was smaller than that estimated with the soil-water-balance model. This indicated that although the general spatial patterns of recharge estimated with the soil-water-balance model were approximately correct at the regional scale of the Northern High Plains aquifer, the soil-water-balance model had overestimated recharge, and adjustments were needed to decrease recharge to improve the match of the groundwater model to calibration targets. The largest components of the simulated groundwater budgets were recharge from precipitation, recharge from canal seepage, outflows to evapotranspiration, and outflows to stream base flow. Simulated outflows to irrigation wells increased from 7 percent of total outflows in 1940–49 to 38 percent of 1970–79 total outflows and 49 percent of 2000–2009 total outflows.

Introduction

The High Plains aquifer (fig. 1) is a nationally important water resource underlying about 175,000 square miles (mi²) in parts of eight states: Colorado, Kansas, Nebraska, New Mexico, Oklahoma, South Dakota, Texas, and Wyoming (Qi, 2010). Irrigation, primarily using groundwater, has sustained agricultural production, resulting in nearly \$50 billion in sales in 2012 (U.S. Department of Agriculture, 2014). In 2010, the High Plains aquifer had the largest groundwater withdrawals of any major aquifer system in the United States (15.1 million acre-feet [acre-ft]; Maupin and others, 2014). The Northern High Plains aquifer (labeled study area on fig. 1) is a distinct region that has little hydrologic interaction with parts of the aquifer further south. Groundwater withdrawals for the Northern High Plains aquifer were about 42 percent of the

total groundwater withdrawals from the High Plains aquifer. Groundwater withdrawals for irrigation, public supply, live-stock, and other purposes were 95, 2, 2, and 1 percent, respectively, of the total groundwater withdrawals from the Northern High Plains aquifer (Maupin and others, 2014).

In 2008, about 8,379,000 acres (13,092 mi² or 13 percent of the area) of the Northern High Plains aquifer were developed for agriculture with irrigation (U.S. Department of Agriculture, 2009); of that area, about 11 percent was served by surface-water diversions and the remainder irrigated with groundwater. Droughts across much of the area from 2001 to 2007, combined with recent legislation (Nebraska Legislature, 2004), have heightened concerns regarding future groundwater availability and highlighted the need for additional information to support science-based water-resource management. For example, although in Nebraska surface-water development is managed at the State level and groundwater is managed at a regional level, shortages of surface water can trigger integrated management of surface water and groundwater (Nebraska Legislature, 2004). Tools with the capability to provide forecasts of groundwater availability and related stream base flows from the entire Northern High Plains aquifer have not been updated for decades (Weeks and others, 1988). To address these needs, in 2009 the U.S. Geological Survey (USGS) began the High Plains Groundwater Availability Study, one of a series of regional studies to evaluate the water availability and sustainability of major aquifers across the Nation. These studies are designed to assist State and local agencies that manage groundwater resources and assess the status of groundwater resources from a national perspective. One purpose of the study was to evaluate future groundwater availability in various potential future conditions, using a groundwater-flow model of the Northern High Plains aquifer. A properly constructed and calibrated groundwater-flow model can simulate regional patterns of groundwater occurrence and flow and related hydrology, and be used to evaluate the effects of potential future changes in climate or land use on the aquifer and related groundwater discharge. For example, in the Northern High Plains aquifer, many resource issues are driven by shortages in streamflows, and groundwater discharge to streams is an important component of streamflow. Therefore, using a model of the Northern High Plains aquifer to evaluate the effects of potential future climate or land cover changes on stream base flow, could provide important information about potential future streamflow.

Purpose and Scope

The purpose of this study is develop a tool to evaluate future groundwater availability in various potential future conditions. The purpose of the model is to simulate regional patterns of groundwater occurrence and flow and related hydrology, primarily groundwater discharge to streams. This report describes the conceptualization, construction, and calibration of a groundwater-flow model of the Northern High

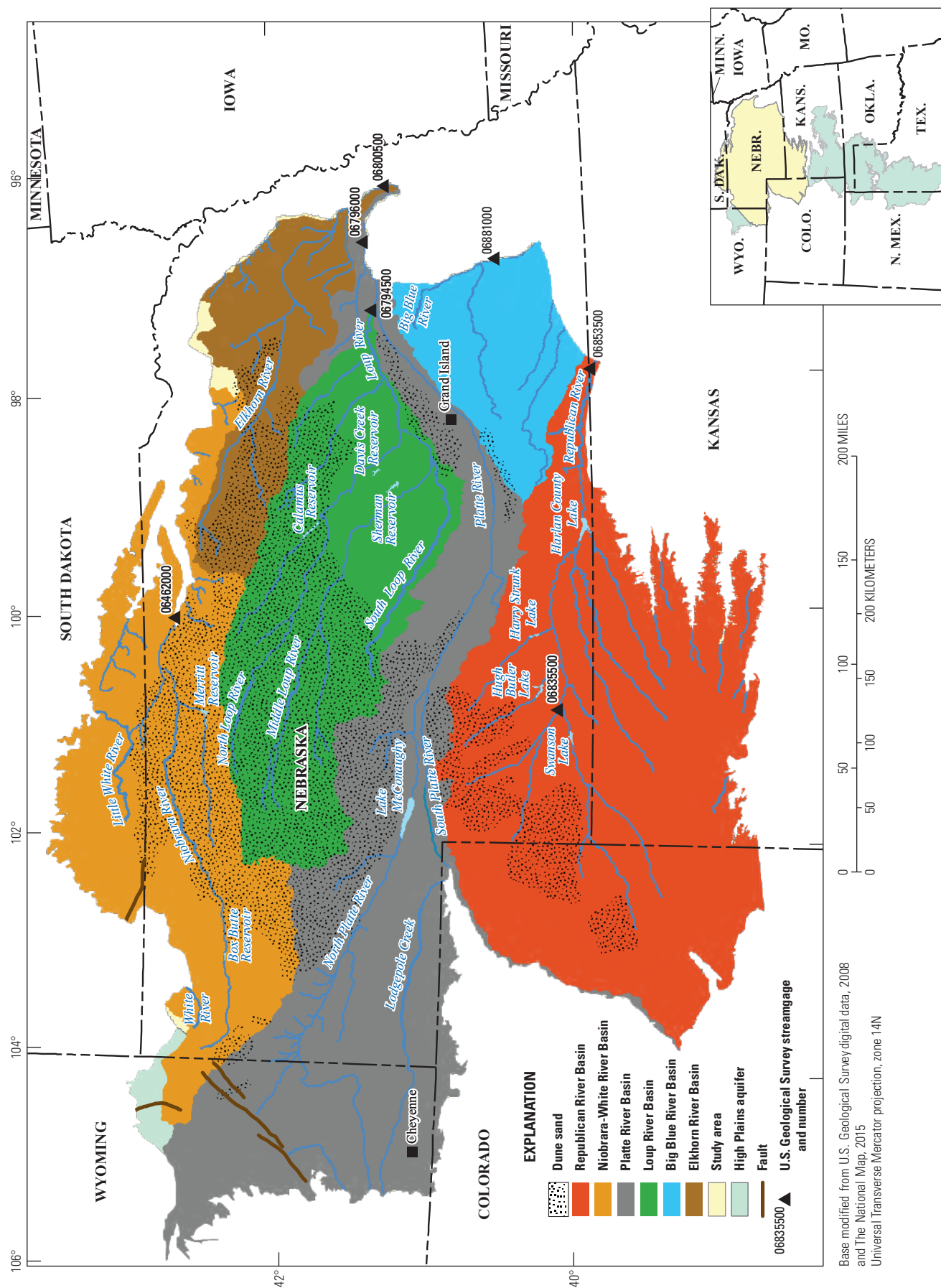
Plains aquifer and modifications to a previously published soil-water-balance (SWB) model. The conceptualization of the groundwater-flow system includes major groundwater inflows and outflows and the approach to simulating the aquifer system. The report describes how SWB model outputs were used as groundwater-flow model inputs and how certain groundwater model inputs were adjusted through calibration using the parameter estimation suite of software (PEST; Doherty, 2005) to reasonably reproduce calibration targets consisting of measured groundwater levels and estimated stream base flows. This report also discusses sensitivity of the groundwater-flow model to various parameter changes. The reported information will guide future users of the groundwater-flow model as a tool for regional evaluations of groundwater resources and of the interactions of groundwater with streams and other hydrologic features resulting from current or forecasted conditions.

Study Area Description

The Northern High Plains aquifer (study area, fig. 1) underlies about 62 million acres of the states of Colorado, Kansas, Nebraska, South Dakota, and Wyoming. Land-surface altitude ranges from more than 7,400 feet (ft) near the western edge to less than 1,100 ft near the easternmost point. Major streams primarily flow west to east and include the Big Blue River, Elkhorn River, Loup River, Niobrara River, Republican River, White River, and Platte River with its two forks—the North Platte River and South Platte River (fig. 1). Population is sparse with only two cities having a population greater than 30,000 (Cheyenne, Wyoming, and Grand Island, Nebraska; National Atlas of the United States, 2004).

Precipitation generally increases west to east and ranges from less than 16 inches (in.) in the western part of the Northern High Plains to almost 31 in. in the eastern part (High Plains Regional Climate Center, 2015a, 2015c). Mean 1981–2010 annual temperatures are generally highest in the southern part of the Northern High Plains (mean 54 degrees Fahrenheit [°F]; High Plains Regional Climate Center, 2015a) and decrease to the north (mean of 47 °F; High Plains Regional Climate Center, 2015b) and with increasing altitudes near the western edge (mean of 45 °F; High Plains Regional Climate Center, 2015c).

The economy of the Northern High Plains depends on agriculture, and the area of irrigated land has increased from about 1890 until today (2015). Early (circa 1890) irrigation was through diversion of surface water (State Board of Irrigation, 1899), and most surface water irrigation projects, covering 1.7 percent of the area or 13 percent of irrigated land in the study area, were substantially in place by 1955. Development of groundwater was limited before 1940 (Weeks and others, 1988), covering less than 0.5 percent of the study area. By 1949, about 827,000 acres of the Northern High Plains aquifer were irrigated (1.7 percent of the area; fig. 2), expanding to 8,379,000 acres by 2008 (U.S. Department of Agriculture, 1952, 1956, 1961, 1967, 1972, 1977, 1981, 1985, 1989, 1994,



1999, 2004, 2009, 2014). In 2008, most of the 1,085,000 acres irrigated with surface water used diversions from the Platte River system or reservoirs therein (fig. 1), whereas 7,294,000 acres of groundwater-irrigated areas (11.7 percent of the area) were dominantly in the eastern one-half of the Northern High Plains aquifer (fig. 3).

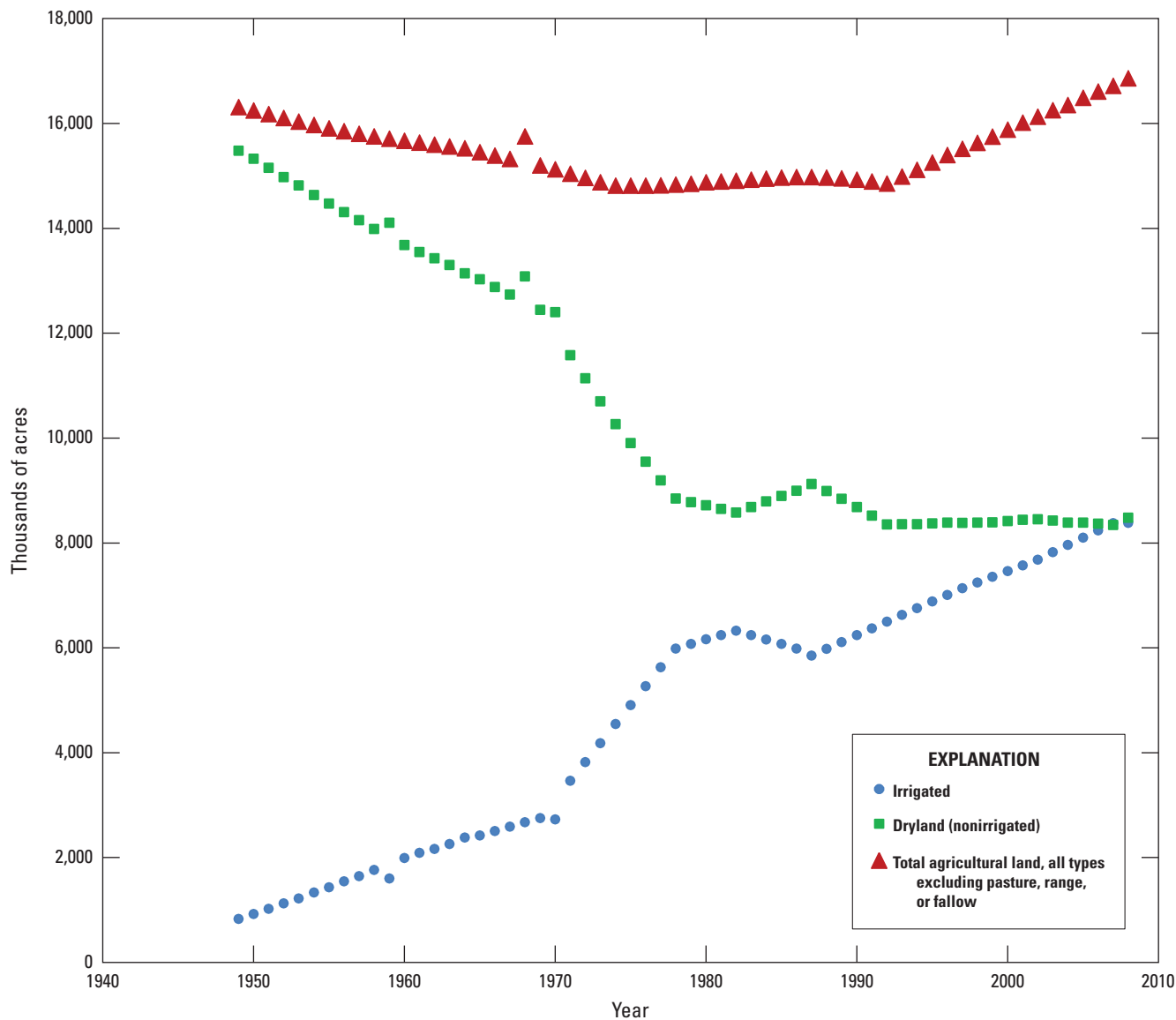
Hydrogeology

The general hydrogeology of the High Plains aquifer was well described in Gutentag and others (1984); a summary of that description is provided in the “Major Hydrogeologic Units” section as a convenience to the reader. That summary is

followed by a description of revisions to Gutentag and others’ (1984) aquifer-base altitude map as implemented for the current study of the Northern High Plains aquifer.

Major Hydrogeologic Units

The High Plains aquifer consists of hydraulically connected deposits of late Tertiary and Quaternary age (fig. 4.4; Gutentag and others, 1984). Late Tertiary-age deposits, from oldest to youngest, include the Chadron Formation of the White River Group (Bartos and others, 2014), Brule Formation of the White River Group, Arikaree Group, Ogallala Formation, and Broadwater Formation (Gutentag and



Data from the U.S. Department of Agriculture 1952, 1956, 1961, 1967, 1972, 1977, 1981, 1985, 1989, 1994, 1999, 2004, 2009, 2014

Figure 2. Agricultural land use (harvested) composed of dryland (nonirrigated) agriculture and irrigated agriculture for the Northern High Plains aquifer, 1949–2008.

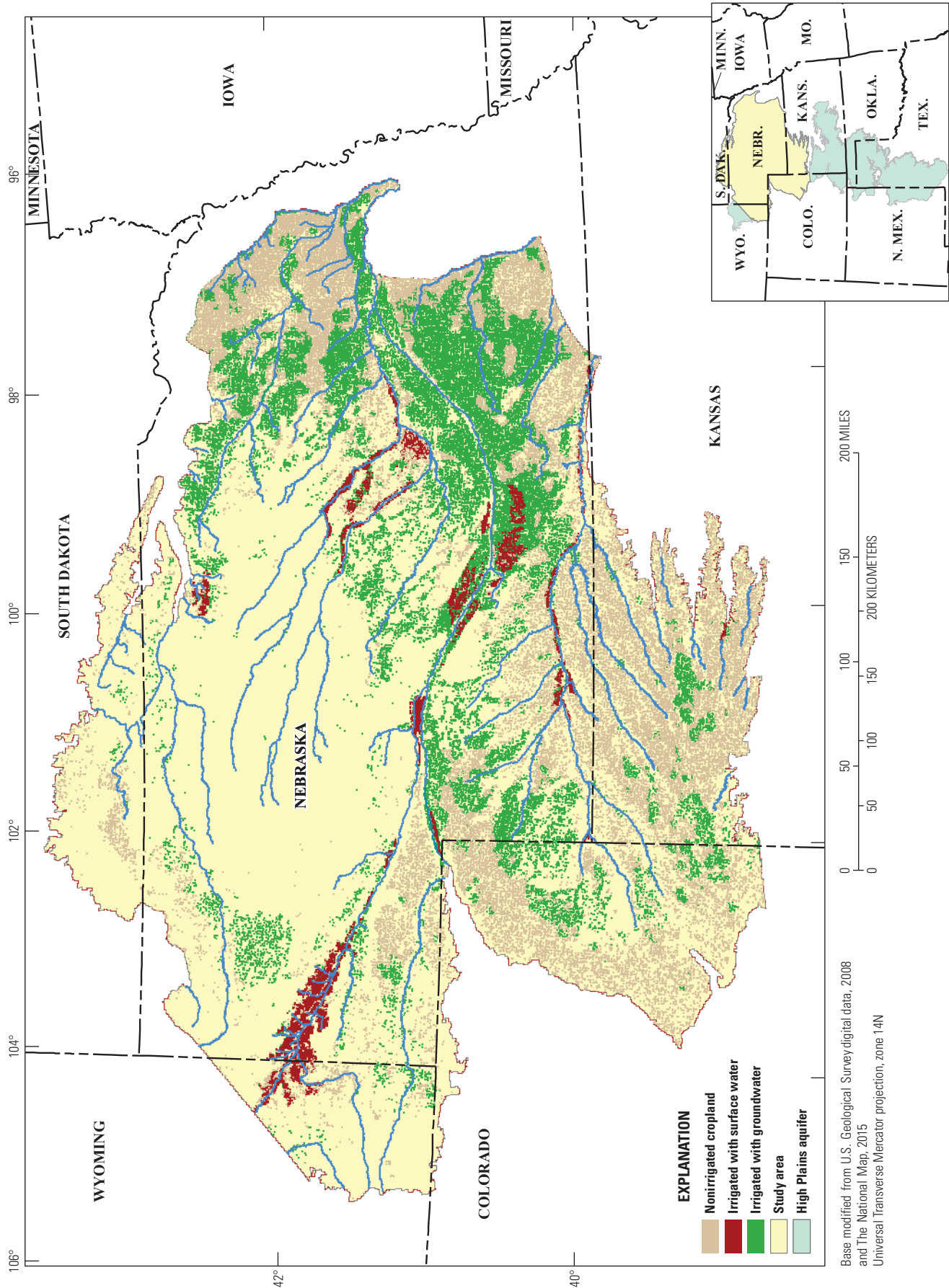


Figure 3. Locations of nonirrigated, groundwater and surface-water irrigated lands in 2008 for Northern High Plains aquifer, Colorado, Kansas, Nebraska, South Dakota, and Wyoming (from Houston and others, 2013).

others, 1984; Diffendal, 1995, Diffendal and others, 2008). Quaternary-age deposits include alluvial, valley-fill, dune sand (figs. 1 and 4), and glacial deposits. Although other geologic units are locally important, the Ogallala Formation composes most of the High Plains aquifer and underlies about 134,000 mi² of the area (Gutentag and others, 1984) and most of the Northern High Plains aquifer (fig. 4A). Generally, low-permeability geologic units of mid-Tertiary age or older underlie the High Plains aquifer (Weeks and Gutentag, 1981), forming an eastwardly dipping (5–7 feet per mile (ft/mi)) paleosurface upon which the High Plains aquifer units were deposited. Local variations in paleosurface altitude are present, forming buried valleys. Groundwater flow between the High Plains aquifer and the underlying units is minimal.

The Chadron and Brule Formations of the White River Group, together with the younger Arikaree Group, are generally fine-grained, low-permeability units except for limited areas of high permeability and areas where permeability has been increased by fractures. These are the oldest geologic units of the High Plains aquifer and are present along the northwestern margins of the Northern High Plains aquifer (fig. 4A). The Brule Formation is mainly a massive siltstone, though it locally contains coarser-grained deposits such as sandstone beds or channel deposits. The Brule Formation is considered part of the High Plains aquifer only where the permeability has been increased by secondary porosity such as joints, fractures, and solution openings (Gutentag and others, 1984). Areas containing coarser deposits, or where the permeability of the Brule Formation has been increased through secondary porosity, are difficult to map on a regional scale (Cannia and others, 2006). Where it has not been enhanced through secondary porosity, the top of the Brule Formation forms the base of the High Plains aquifer. In the western part of the Northern High Plains aquifer where both units are present, the Brule Formation is overlain by the younger Arikaree Group, mainly composed of very fine to fine-grained sandstone (fig. 4A). The Arikaree Group has a maximum thickness of about 1,000 ft in western Nebraska and eastern Wyoming.

The Ogallala Formation is generally coarser and more permeable than the older underlying units and extends over most of the study area (fig. 4A). The Ogallala Formation is a heterogeneous deposit of interlayered stream sediments, lakebeds, and eolian sand, silt, and clay. The Ogallala Formation varies greatly in particle size and physical character over short distances (Cannia and others, 2006). The maximum thickness of the Ogallala Formation is about 800 ft (Swinehart and others, 1988). In most places where it is present, sediments of the Ogallala Formation form the thickest part of the Northern High Plains aquifer. Sediments of the Ogallala Formation are less coarse than the overlying Quaternary alluvial and valley-fill deposits; gravel is not abundant within the Ogallala Formation (Lawton, 1984).

The Broadwater Formation, an alluvial sand and gravel deposit of late-Tertiary age, overlies the Ogallala Formation and underlies younger Quaternary-age deposits across the north-central part of the Northern High Plains aquifer

(Swinehart and others, 1985), though the extent of the Broadwater Formation has not been completely mapped. Sediments of the Broadwater Formation are generally coarser and more permeable than those of the underlying Ogallala Formation. The Broadwater Formation has a maximum thickness of 300 ft and contains more silt eastward, though generally it is only distinguished from overlying Quaternary-age alluvial deposits because of its age, whereas the grain size and other physical characteristics are similar. Though not necessarily called the Broadwater Formation in the eastern part of the study area, sand and gravel of equivalent age are present there as well.

Unconsolidated Quaternary-age sedimentary deposits overlie the older aquifer units. The oldest Quaternary alluvial deposits are largely to the east where the Ogallala Formation is absent (fig. 4A). Next oldest are the dune sands that overlie the central part of the study area (fig. 1), followed by the Quaternary glacial deposits to the northeast (fig. 4A). The youngest deposits are the alluvial deposits associated with the modern river basins (labeled Quaternary valley-fill deposits in fig. 4A).

Unconsolidated Quaternary-age alluvial gravel, sand, silt, and clay overlie and are in hydrologic connection with the Ogallala Formation in the eastern parts of the Northern High Plains aquifer (fig. 4A). Unconsolidated Quaternary alluvial deposits are generally coarser and more permeable than those of the Ogallala Formation and other older underlying units. Eastward, where the Ogallala Formation is absent, Quaternary alluvial and valley-fill deposits directly overlie poorly permeable bedrock. Where the aquifer consists mainly of Quaternary-age alluvial deposits, it is generally thinner than in areas dominated by the Ogallala Formation; maximum thicknesses of Quaternary alluvium are around 300 ft (Gutentag and others, 1984).

Eolian dune sand deposits of Quaternary age overlie the Ogallala Formation in the central part of the Northern High Plains aquifer (fig. 1). The largest contiguous area, known as the Nebraska Sand Hills, covers about 20,000 mi² of the Northern High Plains aquifer and was undergoing dune formation and migration as recently as about 700 years ago (Miao and others, 2007). The dune sands range from very fine to medium sand and, where saturated, are considered part of the High Plains aquifer (Gutentag and others, 1984). The dune sand deposits are as much as 300 ft thick, but on average their thickness is 100 to 150 ft, and usually are a thin veneer on top of the underlying deposits of the Ogallala Formation (Lawton, 1984). Though the dune sands compose only a minor part of the aquifer, they serve as an important surficial feature enhancing aquifer recharge. Ogallala Formation deposits underlie all Quaternary dune sands present in the High Plains aquifer (Muhs, 2007).

Though not always acknowledged in discussions regarding the High Plains aquifer, glacial deposits overlie the eastern end of the Northern High Plains aquifer (fig. 4A; Condra and others, 1950), forming a region of generally lower aquifer permeability but with poorly defined subsurface character and continuity. Whereas glacial deposits have been eroded in major stream valleys, glacial till remains in intervalley areas

of the Northern High Plains aquifer north of the Platte River (Soller and others, 2012). The glacial deposits consist of till and outwash overlain by eolian loess with possible buried valley-fill deposits of sand and gravel. The distribution and occurrence of buried valley-fill deposits within or underlying the till is not well known. Though the fine-grained till is only poorly permeable, groundwater may flow through local deposits of sand and gravel within the till and through underlying or intervening glacial valley-fill deposits. Eastern Nebraska glacial deposits have been a recent subject of study (Smith and others, 2008; Divine and others, 2009), but the interaction between groundwater within the glacial deposits and other aquifers is still poorly understood; however, groundwater-flow models of a subarea of the Northern High Plains aquifer (Peterson and others, 2008; Stanton and others, 2010) used the western edge of the glacial till as a no-flow boundary or simulated a very low rate of cross-boundary groundwater flow. Those models calibrated favorably with minimal groundwater discharge from the Quaternary alluvial deposits into the till deposits (west to east), supporting the concept that the Northern High Plains aquifer may not act as a continuous hydrogeologic unit through the area overlain by till.

Surficial deposits of eolian loess overlying parts of the Northern High Plains aquifer (Muhs and Bettis, 2000) are important because their fine texture limits the maximum infiltration rate and, therefore, the rate of groundwater recharge. Loess is defined as eolian deposits of primarily silt-sized particles (Pye, 1995) that can be as thick as 325 ft (Condon, 2006; Johnson and Brennan, 1960; Richmond and others, 1994).

Quaternary-age valley-fill deposits are similar in character and deposition to the Quaternary-age alluvial deposits and are distinguished because the valley-fill deposits are related to erosion and deposition by modern-day stream systems rather than ancient streams. These valley-fill deposits are as much as 60 ft thick and occupy most major river valleys that cross the Northern High Plains aquifer.

Aquifer Thickness and Base Altitude

Aquifer saturated thickness represents the vertical thickness in which pore spaces of the aquifer are filled with water (saturated) between the relatively impermeable aquifer base and water table. Aquifer saturated thickness was mapped for the High Plains aquifer by McGuire and others (2012). The saturated thickness of the Northern High Plains aquifer ranges from less than 50 ft to greater than 1,110 ft (fig. 4B). In the Northern High Plains aquifer, mean saturated thickness is 253 ft and is largest in the north-central part of the aquifer where it is more than 1,100 ft thick. Aquifer saturated thickness is less than 50 ft near most edges of the study area where

the aquifer has been eroded away and pinches out near the boundaries. Aquifer saturated thickness is less than 200 ft for the southern one-half of the Republican River Basin (figs. 1 and 4B).

An updated aquifer base altitude contour map was constructed as part of this study (fig. 4A) and documented in Peterson and Traylor (2016). Although the general distribution and thickness of the Northern High Plains aquifer remains as reported in Gutentag and others (1984), at the time of that work only limited data were available to constrain the interpretation of contours representing the altitude of the aquifer base surface (Weeks and Gutentag, 1981). As of 2010, 16,950 well logs in or near the Northern High Plains aquifer had been compiled (Houston and others, 2013); 8,082 of these wells had recorded drilling dates, and more than one-half (4,090) of these wells were drilled after 1980. This dataset represented a substantial amount of new information available to refine the aquifer base altitude contours.

In addition, aquifer altitude contours have been published by several other studies since Gutentag and others (1984). Cannia and others (2006) provided surface altitude contours for multiple hydrostratigraphic units for the Platte River Basin, mostly in Nebraska. Kansas Geological Survey (2005) provided updated aquifer base altitude contours for the section of the Northern High Plains aquifer in Kansas. McGuire and Peterson (2008) updated the aquifer base altitude contours for a large part of north-central Nebraska. Peterson and others (2015) provided highly detailed aquifer base altitude contours using test-hole logs and geophysical data for a small part of the alluvial aquifer near the North Platte River in western Nebraska.

The various altitude contours described in the previous paragraph were combined and reconciled. Reconciliation was necessary because of the various dates of publication and the differing data available at the time each map was compiled. Generally, the contours from the newest maps were followed most closely. For some parts of the Northern High Plains, no altitude contours had been published since Gutentag and others (1984). After all altitude contours had been combined and reconciled, they were compared to aquifer base altitudes estimated at test-hole locations, revised as needed, and published (Peterson and Traylor, 2016).

The aquifer base slopes from an altitude of greater than 6,800 ft in southeastern Wyoming to about 1,000 ft in eastern Nebraska (fig. 4A; Peterson and Traylor, 2016). The aquifer base slope is greater in the western one-half of the study area, averaging about 18 ft/mi, than in the eastern one-half of the study area, where it averages only about 5 ft/mi. Local aquifer base slope variations also exist, such as near the Nebraska-Wyoming border, as exhibited by the close spacing of the aquifer base altitude contour lines (fig. 4A).

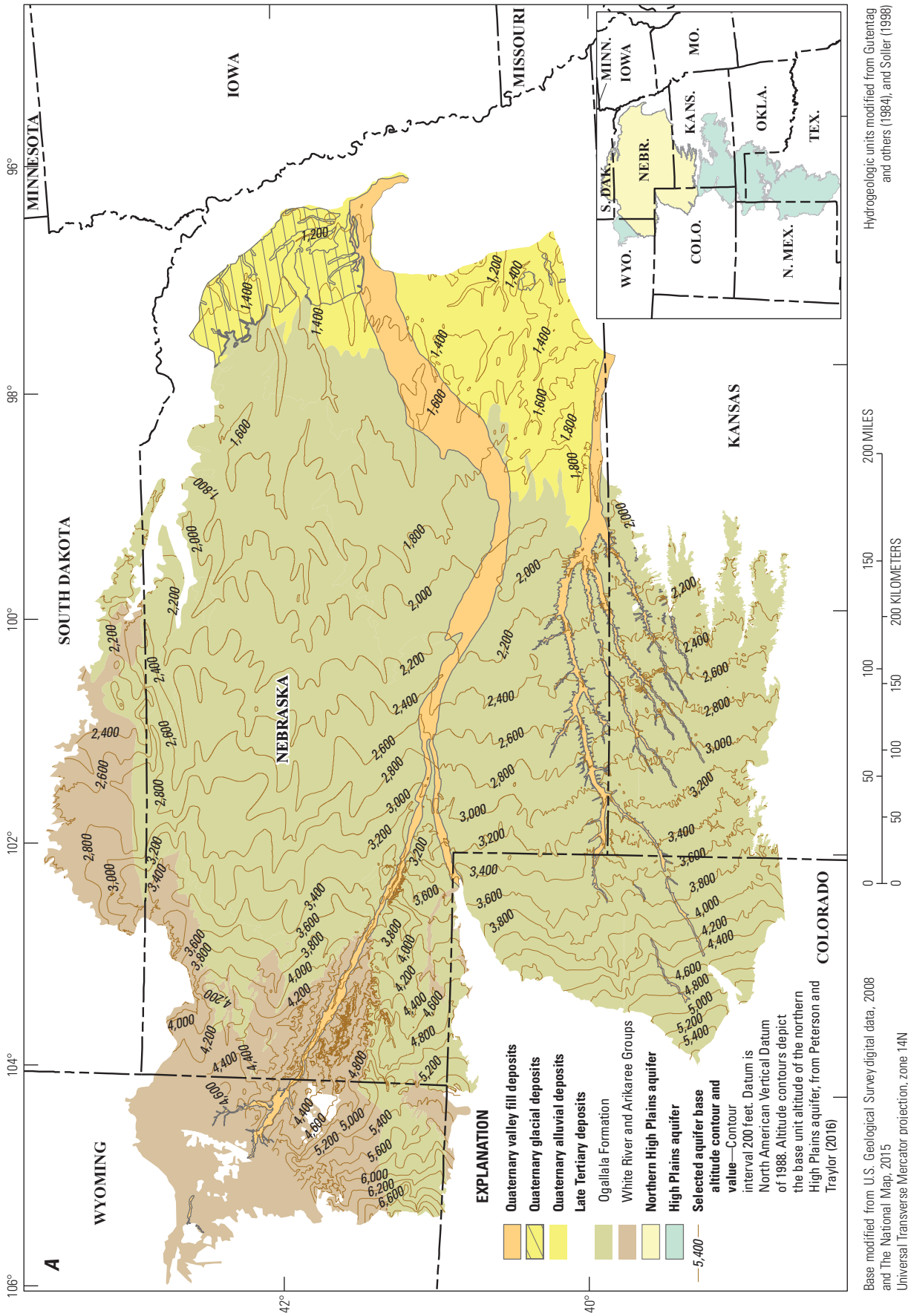
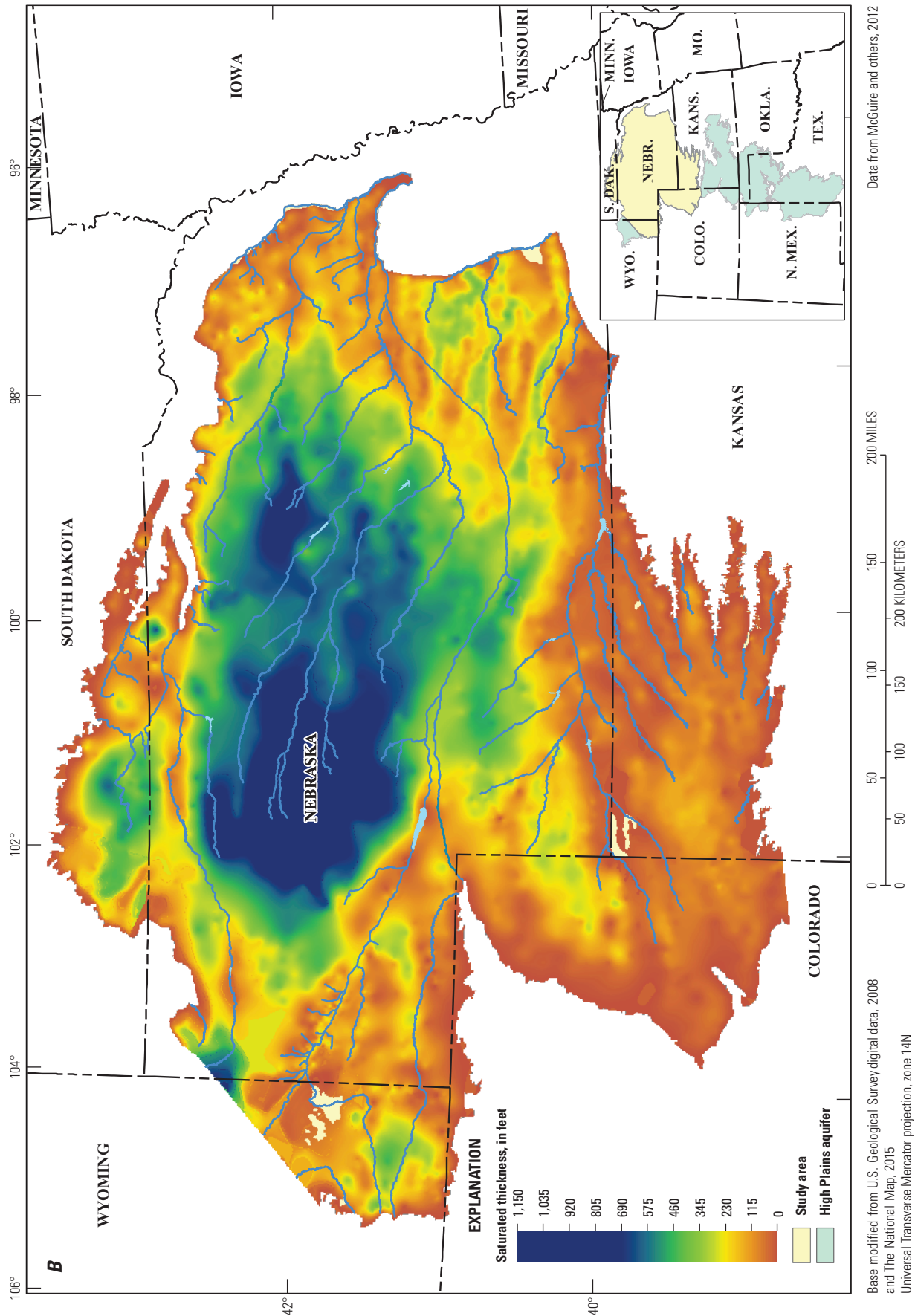


Figure 4. Hydrogeologic units, aquifer base altitude contours, and aquifer saturated thickness of the Northern High Plains aquifer. *A*, major hydrogeologic units and updated aquifer base altitude contours (Peterson and Traylor, 2016). *B*, aquifer saturated thickness.



Groundwater-Flow Model

This section of the report describes the groundwater-flow model of the Northern High Plains aquifer, including the conceptual model of groundwater flow, modular three-dimensional finite-difference groundwater-flow model (MODFLOW) construction, MODFLOW model calibration approach, MODFLOW model calibration results, and assumptions and limitations. Simulated groundwater budgets and composite parameter sensitivities are included in the section on calibration results. The data for this model are available in Peterson and others (2016).

Conceptual Model of Groundwater Flow

The conceptual model of groundwater flow is a narrative description of groundwater occurrence and movements in the study area, including physical characteristics such as saturated thickness, aquifer base elevation, and the maximum rate of groundwater flow. A conceptual model of groundwater flow also describes the largest inflows to and outflows from the aquifer and any changes in groundwater flows or storage through time. The conceptual model of groundwater flow was used to design a groundwater-flow model of the Northern High Plains aquifer, and the initial conceptual model evolved during the course of the study as different theories were tested to improve the match between simulated groundwater flow and estimated or measured hydrologic data. In the interest of brevity, only the final conceptual model is presented in this report.

For the Northern High Plains aquifer, groundwater flows generally from west to east, with most recharge coming from precipitation and most discharge to streams as base flow. Permeability of the aquifer generally increases as the geologic units get younger, with the generally least-permeable units being the oldest units exposed to the west and the Quaternary alluvial deposits being the most permeable. Although sediments of the Ogallala Formation are less permeable, most of the groundwater of the Northern High Plains aquifer is stored in this unit because of its considerable thickness and large areal extent. Precipitation and recharge from precipitation generally increase from west to east, but recharge is also larger in areas with coarse-grained soils, such as the Nebraska Sand Hills (dune sand north of the North Platte and Platte Rivers, fig. 1). In these areas, permeable sand deposits allow efficient transmission of recharge from land surface to the aquifer, minimizing losses to evapotranspiration (ET). Conversely, fine-grained soils, such as sandy clay loam or clay loam (fig. 5), limit infiltration rates and cause more runoff, so less water enters the soil profile and has the potential to become recharge (U.S. Department of Agriculture, 2006).

Aquifer Physical Characteristics

Aquifer boundaries mapped by Qi (2010) were used to define the study area, except along the northwestern part of the boundary where the modeled area was truncated by faults (fig. 1; Cederstrand and Becker, 1999a; Weeks and Gutentag, 1981) that act as barriers to groundwater flow. The last aquifer scale groundwater model of the Northern High Plains (Luckey and others, 1986) also omitted the area northwest of these faults. Aquifer thickness and base altitude are described in the “Aquifer Thickness and Base Altitude” section of this report. The water table is the upper surface of the saturated part of the unconfined aquifer and also slopes gently from west to east, though there are local variations (Gutentag and others, 1984).

Hydraulic conductivity (fig. 6) and specific yield (fig. 7) estimated at test holes in the Northern High Plains aquifer (Houston and others, 2013) were interpolated to assign respective values across the modeled domain using inverse-distance weighting. Mean estimated horizontal hydraulic conductivity of the aquifer was 30 feet per day (ft/d) with a range from 2 to greater than 178 ft/d, though the estimation technique (Houston and others, 2013) precluded estimation of values greater than 178 ft/d that are possible locally (Gutentag and others, 1984). Mean estimated specific yield was 0.15 (dimensionless) with a range from 0.04 to 0.24.

Groundwater Inflows, Outflows, and Changes in Storage

The exact proportions of groundwater inflows to and outflows from the Northern High Plains aquifer (fig. 1) were not known at the outset and, therefore, were subjected to further analysis during the course of this study; however, the general proportions of inflows and outflows were expected to be similar to those simulated by the Elkhorn-Loup Models (Peterson and others, 2008; Stanton and others, 2010) or of the Cooperative Hydrology Study models (Carney, 2008; Luckey and Cannia, 2006; Peterson, 2009). Those models featured recharge from precipitation as the largest groundwater inflow, and groundwater discharge to streams, ET, and groundwater withdrawals for irrigation as the largest outflows.

The main source of inflows to groundwater in the Northern High Plains aquifer is recharge from precipitation that infiltrated past the root zone and was transmitted to the water table. Recharge from precipitation can vary substantially from day to day, month to month, or decade to decade depending on the amount and timing of rainfall and the amount of ET within the root zone. Estimates and measurements of climatic variables, such as precipitation and ET, can vary considerably depending on the estimation technique, as can estimates of recharge (Stanton and others, 2011); however, mean annual recharge for the Northern High Plains aquifer is generally reported to be no more than a few inches per year (Carney, 2008; McMahon and others, 2011; Luckey and Cannia, 2006; Peterson, 2009; Stanton and others, 2010).

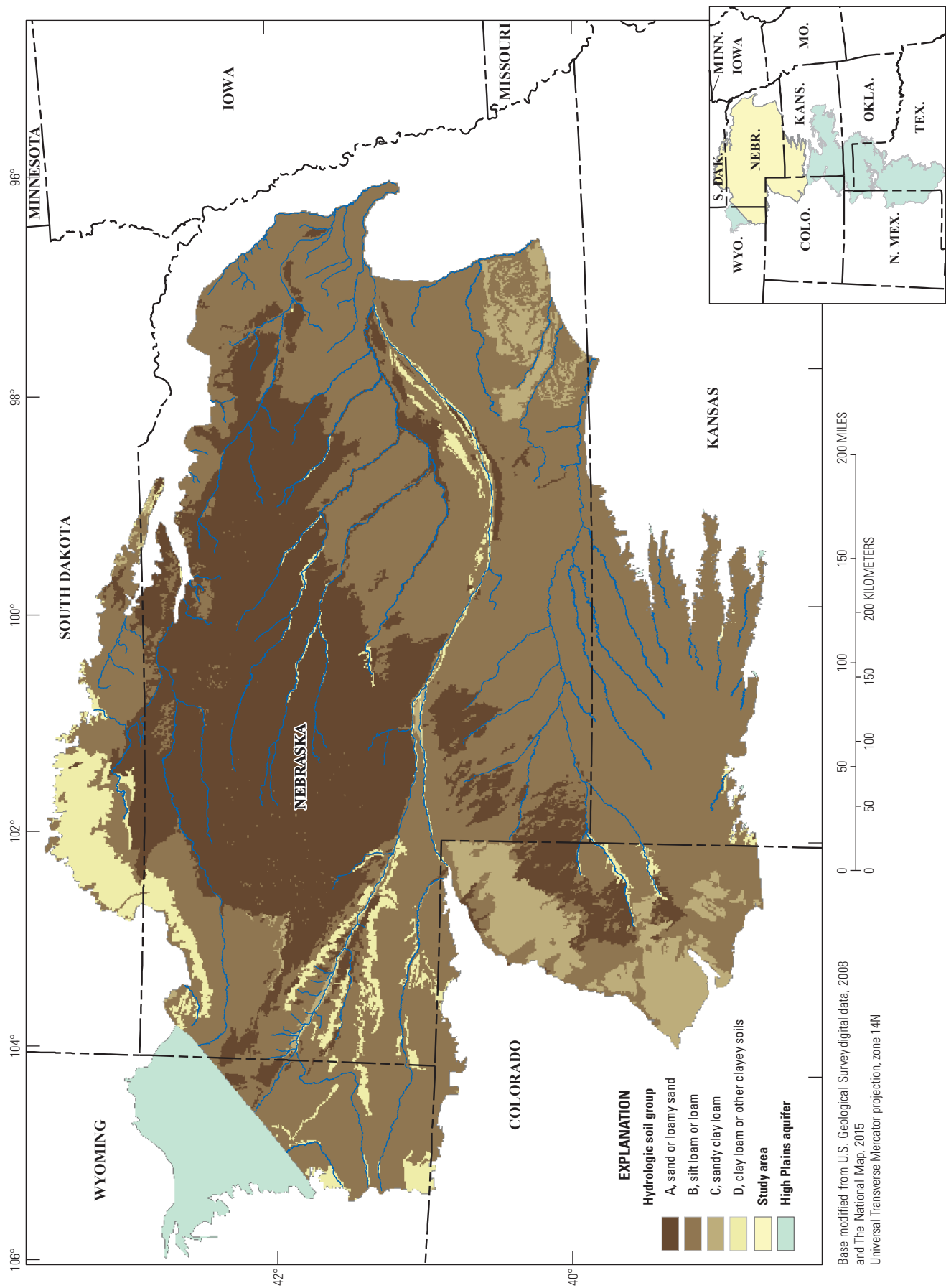


Figure 5. Hydrologic soil groups for the Northern High Plains aquifer.

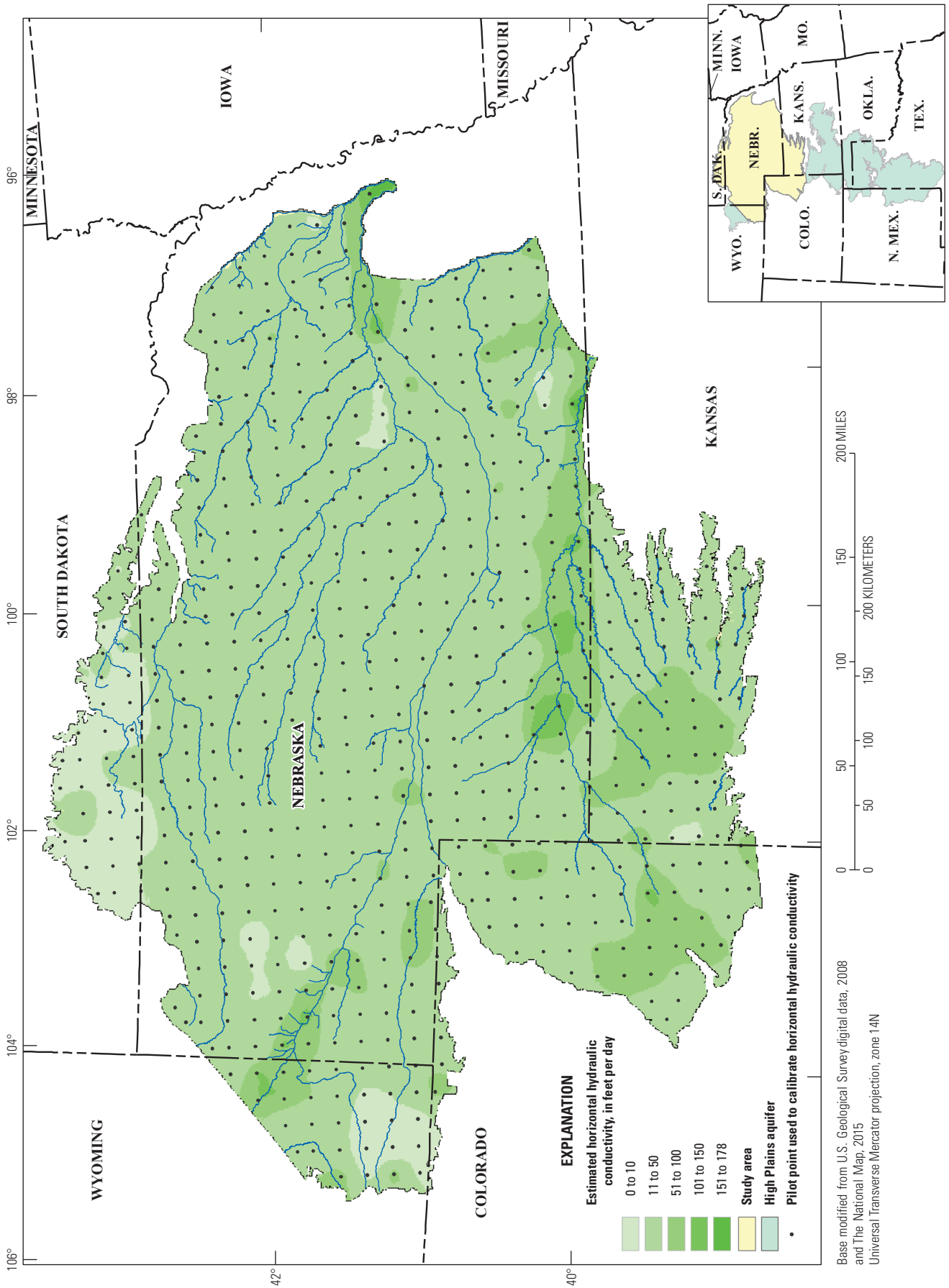


Figure 6. Interpolated estimated horizontal hydraulic conductivity and pilot points used for calibration of horizontal hydraulic conductivity for the Northern High Plains aquifer.

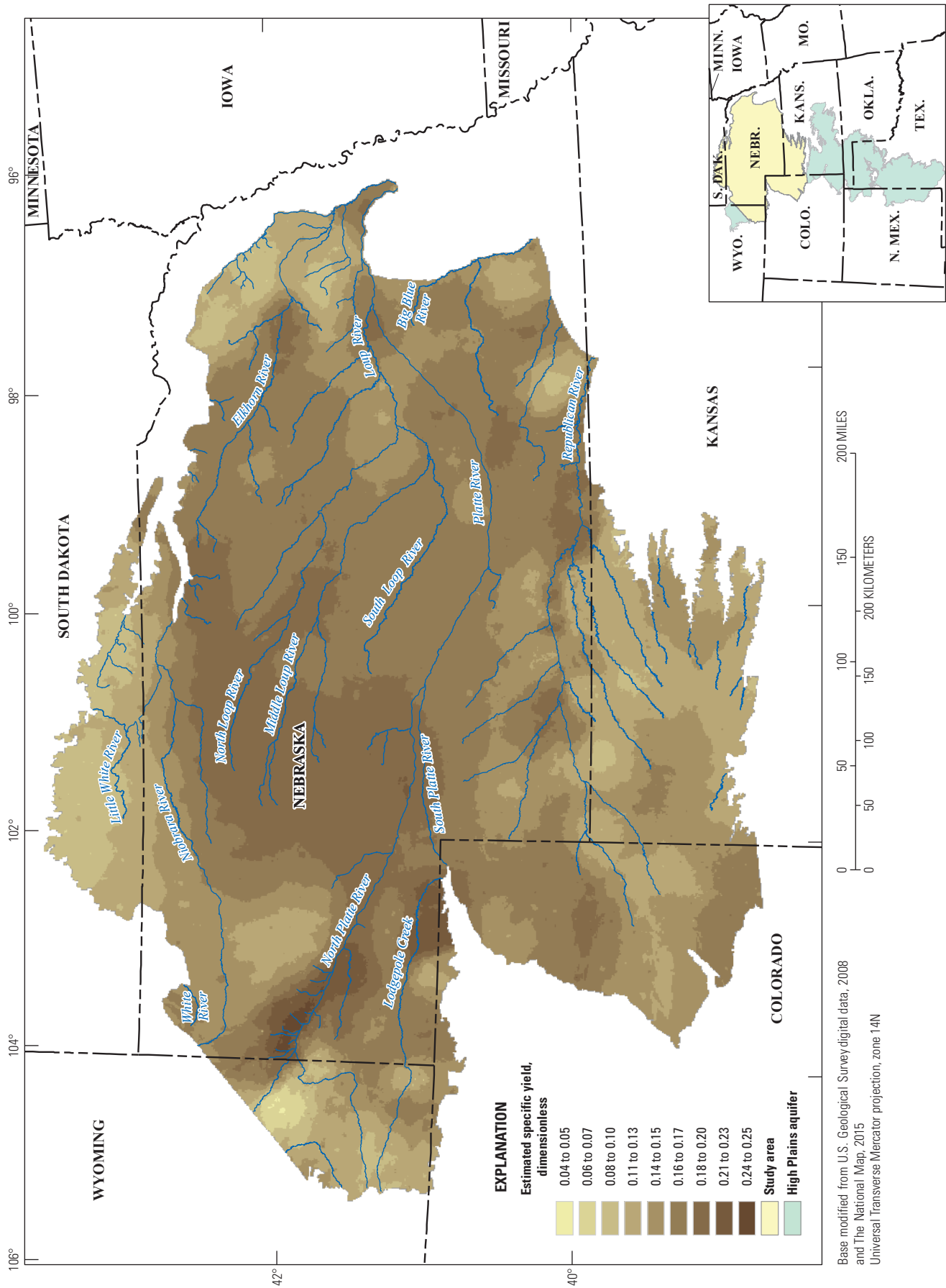


Figure 7. Interpolated estimated specific yield for the Northern High Plains aquifer.

Smaller but locally important sources of inflows to the Northern High Plains aquifer include recharge from canal and reservoir seepage and cross-boundary groundwater flows primarily happening near where the western boundary of the study area was imperfectly truncated by faults. Of these, only recharge from canal seepage constitutes a substantial part of inflows, and even this recharge is far smaller at the scale of the study area than recharge from precipitation. Recharge from canal seepage constitutes an important source of inflows at the local scale, within miles or tens of miles of canals, depending on the rate that water leaks from the canals and becomes recharge to groundwater. Recharge from canal seepage increased as surface-water irrigation was developed beginning in the 1890s (State Board of Irrigation, 1899); development was largely complete by 1955, though surface water was developed in a few additional areas as late as 1993 (Rucker, 2011). Since 1955, only a few canals have ceased being used or have been improved to limit seepage (C. Steinke, Central Nebraska Public Power and Irrigation District, oral commun., 2012), causing reductions in recharge from canal seepage. Similarly, seepage from reservoirs primarily affects groundwater within several miles of those reservoirs, was probably largest for a number of years after reservoir construction, and likely declined after that time as local groundwater levels and flows reached equilibrium with the altitude of the water in the reservoir (Peterson and others, 2008). Seepage from reservoirs into groundwater is not a major inflow to the Northern High Plains aquifer. Cross-boundary groundwater flows are limited to a few areas near the western edge of the model in Wyoming and are only important in those local areas.

Stream base flow is a major groundwater outflow of the Northern High Plains aquifer. Stream base flow, the component of stream flow contributed by groundwater discharge, can constitute more than 90 percent of total stream flow in north-central Nebraska (Stanton and others, 2010). The quick-flow runoff component of stream flow (that is, precipitation onto land or water surfaces that then travels by overland flow or other quickflow paths to a stream) is a component of the surface water system and does not contribute to groundwater. Average stream base flow is generally constant through time at the scale of the Northern High Plains aquifer, except at some locations where it increased after 1970, similar to what was reported by McCabe and Wolock (2002).

A second major outflow of groundwater of the Northern High Plains aquifer is groundwater discharge to ET. Groundwater outflows to ET are limited to areas where the water table is shallow (within several feet of land surface), such as near streams, lakes, and wetland areas (Stanton and others, 2011), and is dependent on climate. However, ET rates can be large, potentially as much as recharge from precipitation (Stanton and others, 2011); hence, ET is an important component of total groundwater outflows at the scale of the Northern High Plains aquifer. The ET outflow is approximately constant through time at multi-decadal and centurial time scales such as considered by this study (pre 1897–2009), though it changes from day to day, month to month, and year to year depending on the climate.

Groundwater withdrawals for irrigation are another major outflow of groundwater from the Northern High Plains aquifer, though they have changed substantially through time. Groundwater was largely undeveloped for irrigation in 1940 but increased during 1940–2009. For additional description on the historical development of surface water and groundwater for irrigation, see the “Study Area Description” section. Groundwater withdrawals for irrigation were not measured historically across the Northern High Plains aquifer but have been estimated as a large outflow of the Northern High Plains aquifer from 2000 to 2009 (Stanton and others, 2011).

Smaller outflows of the Northern High Plains aquifer include municipal well withdrawals and discharge to seeps and springs near the edges of the aquifer, though these are quite small compared with other aquifer outflows. Municipal well withdrawals from the High Plains aquifer are only a few percent of the total well withdrawals (Maupin and others, 2014). Outflows to seeps and springs near the edges of the study area represent water leaving the Northern High Plains aquifer as seeps or springs at outcrops, which only happen near the eastern edge of the aquifer in Kansas and along the edge in the northeastern part of the aquifer near the Nebraska-South Dakota State line.

Changes in groundwater storage, or groundwater flows to and from storage, are represented by groundwater levels either rising or declining in the Northern High Plains aquifer. Groundwater levels in the Northern High Plains aquifer have been generally stable since groundwater withdrawals for irrigation began, but water levels have risen or declined locally (McGuire, 2014). Most rises were in areas where surface water is used for irrigation and are associated with canal leakage, surface water impoundments, or irrigation inefficiency. Declines were primarily in small areas of groundwater irrigation in the arid southwestern part of the Northern High Plains, and the overall volume of water in storage in the Northern High Plains aquifer has changed very little since the onset of groundwater withdrawals for irrigation.

Groundwater-Flow Model Construction

This section of the report describes how the Northern High Plains aquifer conceptual model of groundwater flow was represented in the USGS modular three-dimensional groundwater model with Newton-Raphson solver (MODFLOW-NWT; Niswonger and others, 2011) simulation, including spatial and temporal discretization of the simulation. This section also describes boundary conditions; how selected inputs of the simulation were generated, including estimation of recharge from canal seepage; how recharge and groundwater withdrawals were estimated with a SWB model; and the remainder of the groundwater model inputs. MODFLOW-NWT includes backward compatibility to many packages developed for earlier versions of MODFLOW, therefore references to original package documentation are included with package descriptions in the “MODFLOW-NWT Inputs and Configuration” section.

Spatial and Temporal Discretization

In order to simulate groundwater flow using MODFLOW-NWT, the study area must be broken into an orthogonal grid of blocks, called cells; within each cell, uniform properties are assigned. Similarly, the time period of the simulation must be broken into a series of intervals called stress periods; stress periods can be of different lengths, but sources and sinks of water must be constant within each stress period. Time steps are a further subdivision of stress periods, used to facilitate numerical computations.

To spatially discretize this MODFLOW-NWT model, an orthogonal grid was built to cover the Northern High Plains aquifer, with active cells modified from Qi (2010) in that the northwest region of the study area was omitted where the aquifer is hydraulically separated by faults (fig. 1; Cederstrand and Becker, 1999a; Weeks and Gutentag, 1981). The orthogonal grid consisted of 565 rows and 795 columns and covered about 59.5 million acres, and each grid cell measured 3,281 ft (1,000 m) per side for a total of 449,175 model cells with 240,819 active cells. The model consisted of one layer vertically, simulated as unconfined. The top altitude of the model was set to 1.1 times the altitude of each model cell, with cell altitudes defined as the mean digital elevation model altitude (U.S. Geological Survey, 2015). A multiplier of 1.1 was used to ensure that the simulated water table would not encounter the simulated land surface, except where boundary conditions were represented by MODFLOW-NWT boundaries (for example, streams). This altitude increase prevented cells from inadvertently being simulated as confined-flow conditions and thereby erroneously using a confined-aquifer storage coefficient. This increase was particularly important during early

stages of model development and during computer-assisted parameter estimation.

To temporally discretize this model, two periods of groundwater development in the area were simulated using sequentially linked MODFLOW-NWT models (table 1): the period before substantial groundwater withdrawals (the pre-1940 model) and the period of increasing groundwater withdrawals from May 1940 through April 2009 (the 1940–2009 model). The pre-1940 model, used for calibration, began with a 1,000-year long stress period to represent steady-state equilibrium ending around 1892, the approximate year when surface-water diversions for irrigation began. Use of a 1,000-year long transient simulation to approximate the steady state provided a robust approach and allowed the model to run reliably no matter how parameters were perturbed during automated calibration. The final pre-1940 model was modified by converting the 1,000-year long initial stress period from a transient simulation to steady-state simulation to verify that the transient effects of starting water-level conditions did not affect the simulated water levels in the final pre-1940 model. The initial stress period of the pre-1940 model was followed by seven transient stress periods of 3 to 11 years in length, corresponding to years when various surface-water diversions began (table 1). Each transient-stress period consisted of five to eight time steps, and individual time steps were 292 to 670 days in length (table 1). The 1940–2009 model spanned 138 stress periods, representing the irrigation season (May–September) and nonirrigation season (October–April) for each year. The seasonal stress periods allowed for simulation of the seasonal differences in recharge, ET, and groundwater withdrawals for irrigation.

Table 1. MODFLOW-NWT temporal discretization for groundwater models of Northern High Plains aquifer in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.

[--, not applicable]

Simulation period	Stress period		Start date	End date	Stress-period length (years)	Number of time steps
	Pre-1940 model	1940–2009 model				
Pre-1940	1, during calibration	--	4/30/1892	4/30/1892	1,000	50
Pre-1940	1, final calibrated model	--	4/30/1892	4/30/1892	Steady state	1
Pre-1940	2	--	5/1/1892	4/30/1897	5	5
Pre-1940	3	--	5/1/1897	4/30/1902	5	5
Pre-1940	4	--	5/1/1902	4/30/1906	4	5
Pre-1940	5	--	5/1/1906	4/30/1917	11	8
Pre-1940	6	--	5/1/1917	4/30/1926	9	7
Pre-1940	7	--	5/1/1926	4/30/1929	3	5
Pre-1940	8	--	5/1/1929	4/30/1940	11	6
1940–2008	--	Odd numbers from 1–137	May 1st	September 30th	0.42	11
1940–2009	--	Even numbers from 2–138	October 1st	April 30th	0.58	15

Soil-Water-Balance Model

The SWB model (Westenbroek and others, 2010) was selected to estimate recharge from precipitation and groundwater withdrawals for irrigation, based on comparisons to published recharge and groundwater withdrawal estimates and evaluation of other estimation techniques (Stanton and others, 2011). The SWB model is a gridded SWB model that uses spatially distributed soil and landscape properties with daily weather data to calculate spatial and temporal variations in potential recharge. Each cell is assigned soil properties and daily climate data, and the SWB model calculates the fractions of precipitation and snowmelt that become surface runoff, actual ET, and recharge using a modified Thornthwaite-Mather soil-water accounting method (Stanton and others, 2011; Westenbroek and others, 2010). Surplus water in the soil column between land surface and the bottom of the root zone becomes potential groundwater recharge, which is calculated by subtracting the sum of the cell losses or outputs (actual ET, plant interception, and surface runoff) from the cell inputs (precipitation, snowmelt, and surface runoff). The SWB model was originally designed to estimate recharge (Westenbroek and others, 2010), and irrigation requirement estimation was added later (S. Westenbroek, written commun., 2013). The SWB model irrigation requirement estimation is based on a simple approach using crop coefficients for early, peak, and late stages of growth (Allen and others, 1998). Few measured data for groundwater withdrawals exist to calibrate estimated withdrawals. Nonetheless, review of the SWB-estimated groundwater withdrawals for irrigation showed the rates to be comparable to published crop water-use data (Klocke and others, 1990; Kranz and others, 2008; Yonts, 2002) minus effective growing-season precipitation (National Climatic Data Center, 2010). Additional details about this approach are provided in Stanton and others (2010) and Westenbroek and others (2010).

Stanton and others (2011) presented recharge from precipitation and groundwater withdrawals for irrigation for the High Plains aquifer estimated using the SWB model (Westenbroek and others, 2010). That High Plains SWB model used a 5,000-ft uniform grid spacing and simulated two periods: 1940–49 and 2000–2009. For this study, the SWB model and inputs from Stanton and others (2011) were rediscritized to match the current model grid with a 3,281-ft spacing and expanded to cover 1940–2009, but using the original data sources for climate data and physical properties. The SWB executable (Westenbroek and others, 2010), used by Stanton and others (2011), was further refined to better represent changes in soil moisture related to irrigation practices and the increase in recharge associated with inefficient application of irrigation water.

The SWB model is capable of using either dynamic or static land-cover data. For this study, dynamic land-use data were used, requiring historical land-cover data. Land-cover data were first defined using the FOREcasting SCEnarios of Land-use Change historical reconstruction of land use (Sohl

and others, 2007) for 1949–2008 (Houston and others, 2013). Land-cover data from Houston and others (2013) did not differentiate irrigated crops from dryland (nonirrigated) crops; therefore, grid cells that were irrigated with surface water in 2008 (fig. 3) were classified as irrigated for 1949–2008, and grid cells containing irrigation wells (Nebraska Department of Natural Resources, 2008) were assumed to be irrigated for all growing seasons subsequent to the well-construction date. The remainder of the cropped cells were assumed to be dryland crops. These parcels were summarized by county and checked against Census of Agriculture (CA) county-level statistics (U.S. Department of Agriculture, 1952, 1956, 1961, 1967, 1972, 1977, 1981, 1985, 1989, 1994, 1999, 2004, 2009, 2014) to ensure that the proper amount of land was designated as irrigated. The CA data are published about every 5 years, and intervening years were linearly interpolated from published data.

Where the CA indicated less irrigated land than the county totals from mapped irrigated parcels, the smallest parcels were reclassified from irrigated to dryland for each county until the total area of mapped irrigated parcels agreed with the number of irrigated acres in the CA to within the area of one model grid cell (247.1 acres). Where the CA indicated more irrigated land in a county and year than did mapped irrigated parcels, parcels were added in each county until the total area of mapped irrigated parcels agreed with the number of irrigated acres in the CA to within the area of one cell (247.1 acres). Irrigated parcels were added using an irrigation density map for guidance so that parcels were added closest to known irrigation acres, commonly in neighboring cells. Areas designated as groundwater or surface water irrigated for 2008 are shown in figure 3; surface water irrigated areas were about constant from 1949 to 2008 and held constant in this approach; therefore, increases in irrigated land from 1949 to 2008 shown in figure 2 are attributed to increases in groundwater-irrigated lands. Because the land-cover data from Houston and others (2013) did not cover 1940–48 and 2009, land cover for 1940–48 was assigned the pattern from 1949, and land cover for 2009 was assigned the pattern from 2008.

Recharge from precipitation estimated with the SWB model had the same general distribution for early groundwater development years (1940–49; fig. 8A) as for years representing later groundwater development (2000–2009; fig. 8B), with less recharge to the south and west and more recharge to the north and east. Recharge also tended to be larger for sandy soils than for fine-grained soils (fig. 5). For the Northern High Plains aquifer, mean annual recharge estimated for 1940–49 and 2000–2009 was 3.3 and 3.5 inches per year (in/yr), respectively. A map of the difference in average annual recharge from precipitation between 1940–49 and 2000–2009 (fig. 8C) shows that for the 2000–2009 period, the recharge difference was within 1 in/yr for much of the study area, recharge was less for 1940–49 for some areas tending to be in western part of the study area, and recharge was more for 1940–49 for much of the eastern part of the study area.

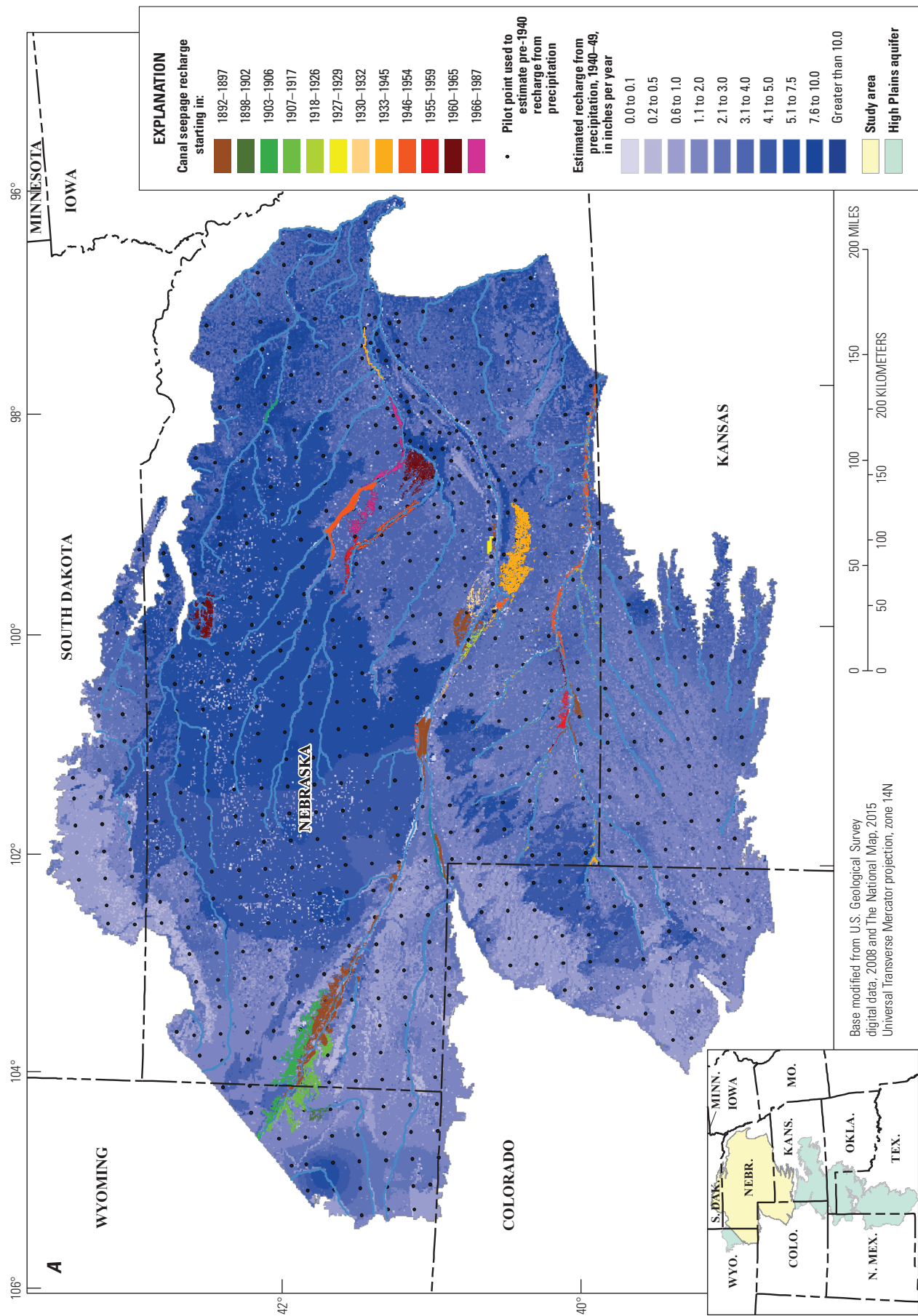
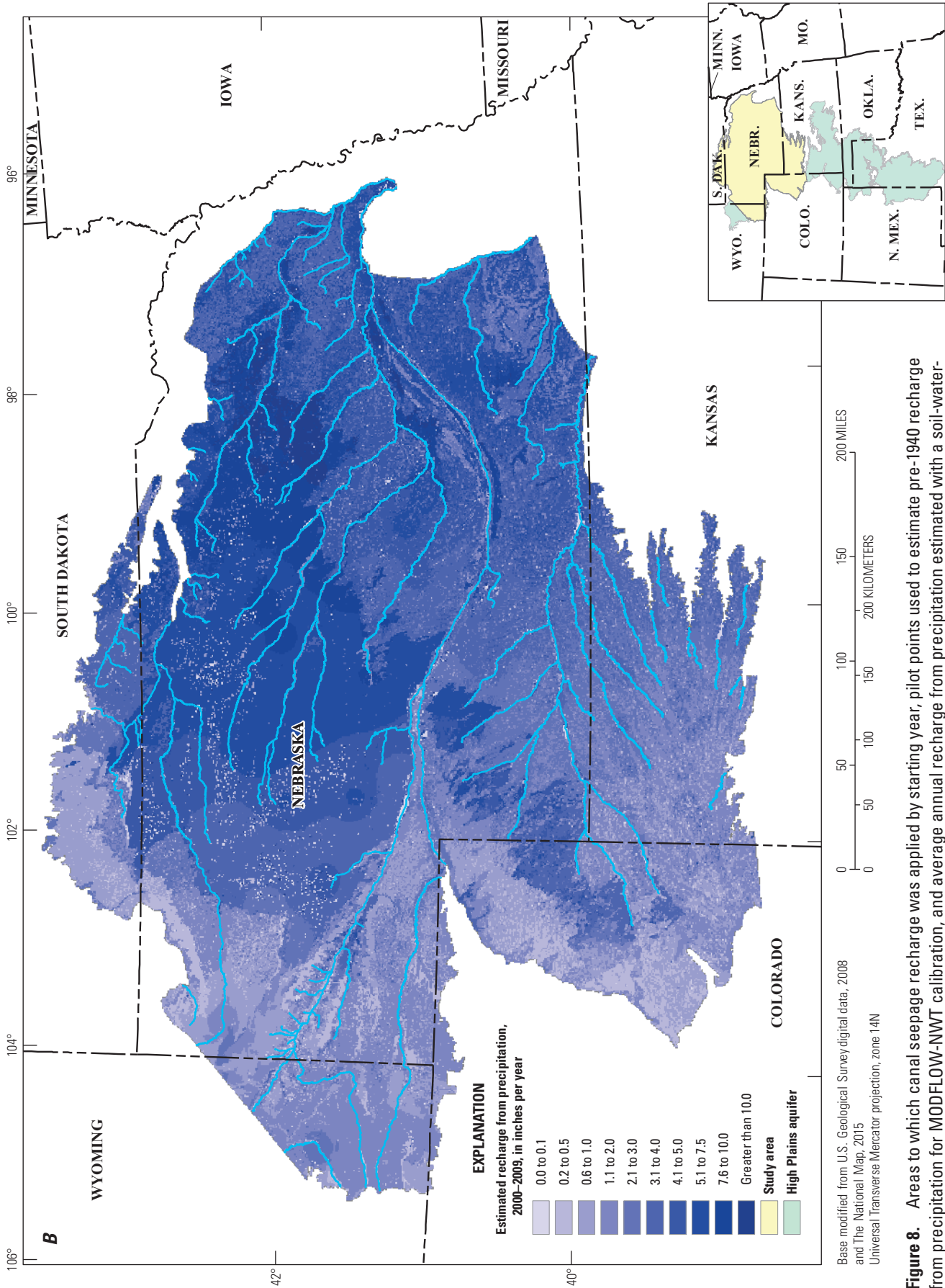


Figure 8. Areas to which canal seepage recharge was applied by starting year, pilot points used to estimate pre-1940 recharge from precipitation for MODFLOW-NWT calibration, and average annual recharge from precipitation estimated with a soil-water-balance model for the Northern High Plains aquifer in Colorado, Kansas, Nebraska, South Dakota, and Wyoming. *A*, 1940–49, representative of early groundwater development. *B*, estimated average annual recharge from precipitation for 2000–2009, representative of later groundwater development. *C*, the difference in average annual recharge from precipitation between 1940–49 and 2000–2009.



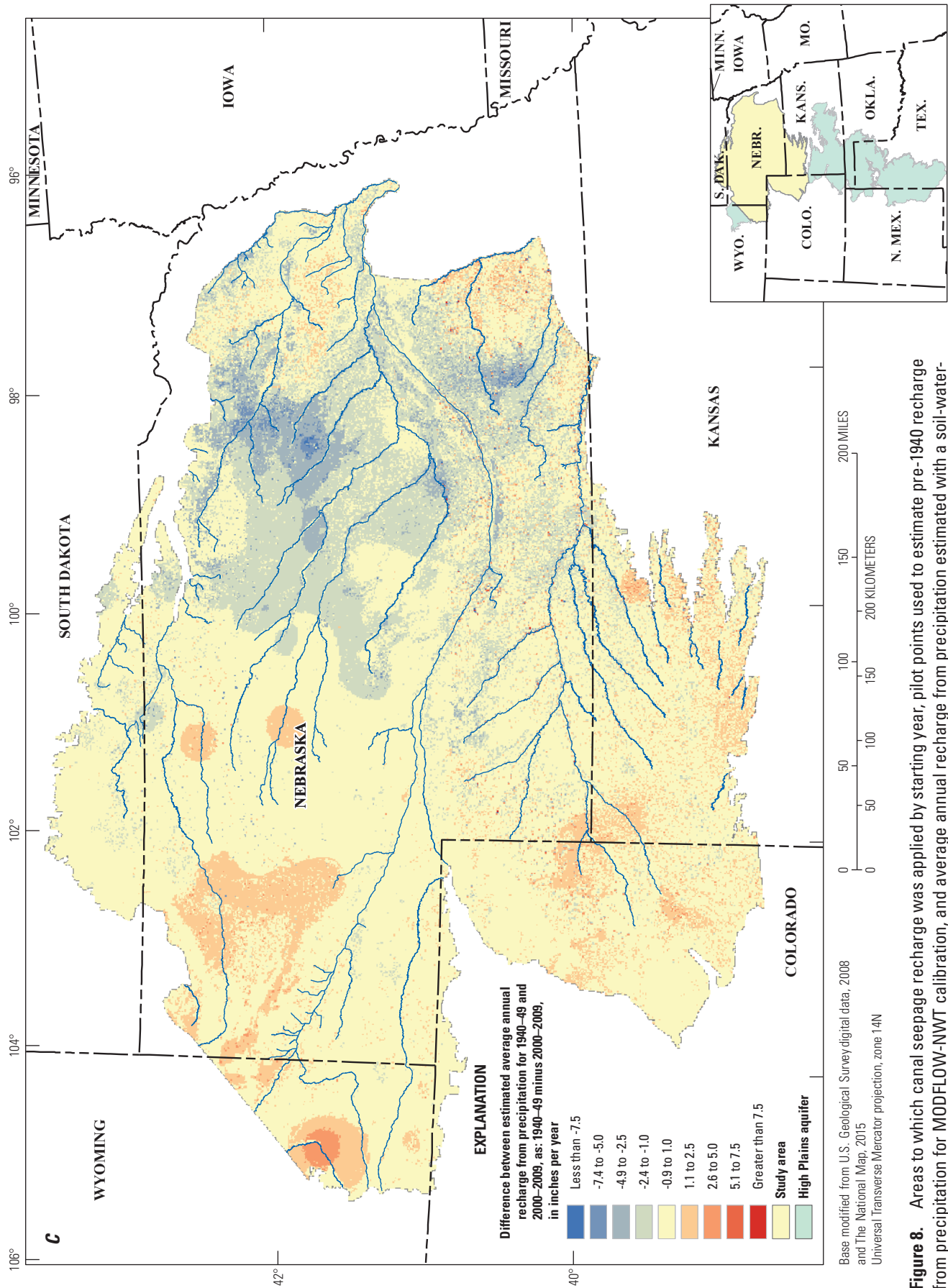


Figure 8. Areas to which canal seepage recharge was applied by starting year, pilot points used to estimate pre-1940 recharge from precipitation for MODFLOW-NWT calibration, and average annual recharge from precipitation estimated with a soil-water-balance model for the Northern High Plains aquifer in Colorado, Kansas, Nebraska, South Dakota, and Wyoming. A, 1940–49, representative of early groundwater development. B, estimated average annual recharge from precipitation for 2000–2009, representative of later groundwater development. C, the difference in average annual recharge from precipitation between 1940–49 and 2000–2009.—Continued

Estimated groundwater irrigation was extracted from the total irrigation amount estimated with the SWB model for this study by setting the irrigation amount to zero in areas with available surface-water irrigation (fig. 3). Most irrigators preferentially use surface-water irrigation before groundwater irrigation, primarily because of lower costs. In limited areas in a few years of the historical record, irrigators used supplemental groundwater supplies in times of surface-water shortage, but estimating supplemental groundwater usage was beyond the scope of this study and was assumed to be negligible compared to withdrawals for irrigation of parcels irrigated with groundwater only.

MODFLOW-NWT Inputs and Configuration

A formulation of the 2005 version of the USGS modular three-dimensional groundwater model, MODFLOW, that uses an alternate method for solving groundwater flow (MODFLOW-NWT, ver. 1.0.9; Niswonger and others, 2011) was selected as the groundwater-flow modeling code for this study. The MODFLOW-NWT incorporated an improved ability to solve nonlinear unconfined aquifer simulations with wetting and drying of cells by applying the Newton-Raphson linearization approach to solving the flow equations (Niswonger and others, 2011). Though most cells in the model were not expected to desaturate during the simulation, the aquifer is thin in some areas, and use of MODFLOW-NWT prevented removal of desaturated cells from the simulation.

Standard MODFLOW components of the Groundwater-Flow Process that were used in the model included the Basic Package and the Discretization File (Harbaugh and others, 2000). The Basic Package was used to specify which cells of the orthogonal grid were active in the simulation, which included 240,819 of the 449,175 total cells in the grid that had 565 rows and 795 columns. The Basic Package was also used to identify which cells were assigned specified groundwater levels (fig. 9), which included 60 cells in this simulation mostly along the western boundary in Wyoming or northwestern Nebraska but also a few in northeastern Nebraska, where interpretation of published contour maps of groundwater-level altitude (Cederstrand and Becker, 1999b) indicated the possibility of groundwater inflows from outside the model area. Altitudes assigned as specified groundwater levels were those representing 2000 conditions from an available continuous groundwater altitude surface (V. McGuire, U.S. Geological Survey, written commun., 2010). Groundwater levels from 2000 may or may not represent multi-decadal average groundwater levels, but groundwater levels had changed little with time in the area of the specified groundwater levels (McGuire and others, 2012), so it was assumed that specified groundwater level altitudes for 2000 were about the same as those for other years. For the remainder of the active model area, starting groundwater levels for the pre-1940 model were from preliminary model outputs, but because the first simulated stress period was steady state, the MODFLOW-NWT simulated groundwater levels were not affected by

the input starting groundwater levels. Because the pre-1940 and 1940–2009 models were sequentially-linked, simulated groundwater levels from the pre-1940 model were used as starting groundwater levels for the 1940–2009 model. The Discretization File specified the grid spacing and temporal discretization described in the “Spatial and Temporal Discretization” section.

The Upstream Weighting Package (Niswonger and others, 2011) specifies properties controlling flow between cells, such as hydraulic conductivity, anisotropy, specific storage, and specific yield. Initial hydraulic conductivity and specific yield were set to the estimated values described in the “Aquifer Physical Characteristics” section. Hydraulic conductivity was assumed to be isotropic and adjusted during calibration. Specific yield was not adjusted during calibration. Horizontal anisotropy was set to 1.0 (no anisotropy). Similarly, specific storage was specified as 10×10^{-5} per foot (Fetter, 1994) and was not adjusted during calibration.

Total recharge simulated with the Recharge Package (Harbaugh and others, 2000) was the sum of canal seepage recharge and recharge from precipitation (described in the “Soil-Water-Balance Model” section). Every simulated stress period of the model was assigned a different recharge input because of variations in canal seepage recharge in the pre-1940 model and stress period to stress period differences in canal seepage recharge and recharge from precipitation in the 1940–2009 model.

Recharge from canal seepage was estimated using available data (T. Naprstek, Lower Loup Natural Resources District, written commun., 2012; Nebraska Department of Natural Resources, 2015; C. Steinke, Central Nebraska Public Power and Irrigation District, written commun., 2012; U.S. Bureau of Reclamation, 2015). Monthly water-distribution reports for 1940–2009 contained, at a minimum, the amount diverted into the canal, and in some cases, also included flows to wasteways, estimated losses, and the amount of water delivered to farms in acre-feet per year. Although direct measurements of evaporation from canals do not exist, it was assumed to be negligible (no more than a few percent); for example, a canal 50 miles (mi) long and 60 ft wide, flowing at 1,000 cubic feet per second (ft^3/s), that evaporates 50 in. of water from May through September, would lose only 0.2 percent of flow to evaporation. It was considered reasonable to assume, therefore, that losses from canals were equivalent to recharge from canal seepage reaching the water table of the Northern High Plains aquifer. For canals with diversion and delivery data, seepage loss was calculated as the amount diverted minus the amount delivered for irrigation and minus waste (if available). For canals without delivery data, seepage loss was assumed to be 40 to 50 percent of the diversion (Carney, 2008; Luckey and Cannia, 2006; Peterson, 2009; Peterson and others, 2015; Stanton and others, 2010). Few data existed for pre-1940 canal operations with which to estimate seepage losses. In cases where the canals were known to have been in operation for an early time period before recorded operational data, the earliest 5 years of operational data were averaged and assumed to be

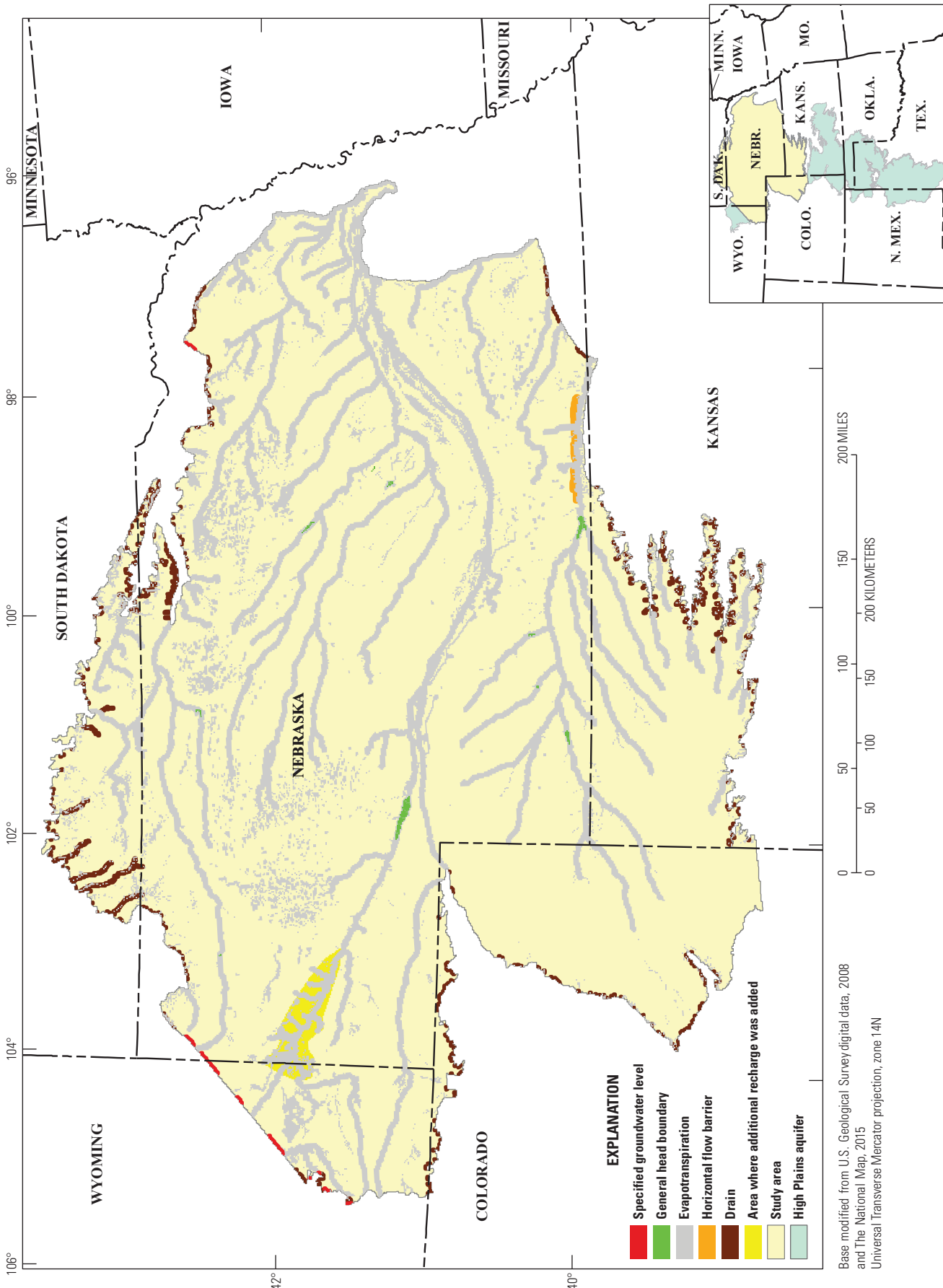


Figure 9. Boundary conditions for the Northern High Plains aquifer in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.

similar to the canal seepage loss that happened earlier but was not recorded. Time-varying canal seepage recharge used in the groundwater model was determined by using the amount of loss and dividing the value by the area of the cells crossed by each canal in that year, resulting in a seepage rate per canal per unit area per year. Canal seepage recharge was applied for surface water irrigated areas depending upon the year each surface water irrigation district started operations (fig. 8A).

Because the SWB model did not simulate pre-1940 conditions, the 1940–49 average annual recharge from precipitation was assigned to the pre-1940 model and similarly summed with canal seepage recharge. Recharge assigned to each stress period of the 1940–2009 model corresponded to the recharge from precipitation estimated with the SWB model for the months that fell within the stress period, summed with the corresponding canal-seepage recharge for that stress period. In addition, preliminary calibration efforts (see “Groundwater-Flow Model Calibration Approach” section) indicated that simulated groundwater levels and stream base flows in the North Platte River valley in western Nebraska and eastern Wyoming were lower than groundwater-level and stream base flow targets for 1940–2009; therefore, an additional 3.0 in/yr of recharge was added to this area for 1940–2009 (fig. 9), which caused the simulated groundwater levels and stream base flows to better match calibration targets.

The MODFLOW Evapotranspiration Package (Harbaugh and others, 2000) was used to simulate groundwater discharge to ET in areas of active model cells but only where the water table is shallow (fig. 9). The ET extinction depth was set to 7 ft. The National Weather Service potential evapotranspiration (NWS–PET) rate (Stanton and others, 2011) was selected for the maximum ET rate. Natural total ET for a given location consists of ET of soil moisture and available groundwater; hence, actual ET of groundwater will be much less than potential ET. The MODFLOW initial maximum ET rate, therefore, was set to 60 percent of the average estimated 2000–2009 NWS–PET for the pre-1940 model and irrigation seasons of the 1940–2009 model, and 50 percent of the 2000–2009 maximum ET rate from Stanton and others (2011) for nonirrigation seasons of the 1940–2009 model. During model calibration, the ET extinction depth was held constant, but the maximum ET rate was calibrated as described in the “Maximum Evapotranspiration Rate Multipliers” section.

The MODFLOW General Head Boundary Package (Harbaugh and others, 2000) was used to simulate interaction of groundwater with several reservoirs in the Northern High Plains aquifer model (figs. 1 and 9), including Lake McConaughy in the North Platte River Basin; Calamus Reservoir, Davis Creek Reservoir, and Sherman Reservoir in the Loup River Basin; Box Butte Reservoir and Merritt Reservoir in the Niobrara River Basin; and Harlan County Lake, Harry Strunk Lake, Hugh Butler Lake, and Swanson Lake in the Republican River Basin. Altitudes were assigned to general head boundaries to represent average lake stage as sampled from a 1 arc-second resolution digital elevation model (U.S. Geological Survey, 2015). Interaction from other reservoirs that

were part of canal systems was assumed to be predominantly seepage and was not simulated explicitly; rather, seepage was estimated as part of recharge from canal seepage.

The Streamflow-Routing Package (Niswonger and Prudic, 2005) was used to simulate streams, including all major streams and selected tributaries (fig. 10). Whereas a comprehensive list of every simulated stream name is too lengthy to provide here, the major streams and river basins (fig. 1) represented in the Northern High Plains aquifer model are the Big Blue River, Elkhorn River, Loup River, Niobrara River, Platte River, Republican River, and larger tributaries associated with these rivers. An initial group of streams was selected for model inclusion if the mean annual estimated base flow was greater than 10 ft³/s at streamgages with at least 5 years of record (see Houston and others, 2013, for streamgage locations). Additionally, streams were added to this group where previous modeling efforts showed the importance of representation of streams in particular regions (Peterson and others, 2015; Stanton and others, 2010; Republican River Compact Administration, 2003) or based on preliminary model results. The headwaters of simulated streams above the streamgages with 10 ft³/s of estimated base flow were selected to represent the stream reaches coded as decadal or longer average perennial reaches in the National Hydrography Dataset (U.S. Geological Survey, 2008). The North Platte and South Platte Rivers are the only streams that originate outside the study area; base flows in these two streams at the model boundary were specified based on monthly average base flow for April (for irrigation seasons) and October (for nonirrigation seasons; Houston and others, 2013). Base flow was also specified for the North Platte River where it leaves Lake McConaughy, which was simulated using a general head boundary in the model and, therefore, did not directly interact with the Streamflow-Routing Package in the model. For the North Platte River downstream from Lake McConaughy, flows were estimated using the base-flow index (Wahl and Wahl, 1995) for the streamgage upstream from Lake McConaughy (the North Platte River near Lewellen, Nebr., streamgage number 06687500) applied to the recorded releases to reconstruct base flow leaving Lake McConaughy. This process likely introduced uncertainty into the downstream simulated flows but was appropriate given the limitations of the approach wherein only stream base flow was simulated, rather than total flow or runoff plus base flow.

Stream widths input to MODFLOW-NWT were estimated using aerial photographs (Dollison, 2010) and ranged from 1 to 1,148 ft with a mean of 104 ft. The estimation approach resulted in largest widths being assigned for large streams, such as the downstream reaches of the Platte River and Loup River, whereas the smallest widths were assigned to the upstream-most sections of rivers and tributaries, such as the South Loup River, Plum Creek, and Frenchman Creek. It was conceptualized that stream bed materials would have different properties for larger streams as opposed to smaller streams because of differences in velocity and discharge causing deposition of different-sized particles; therefore,

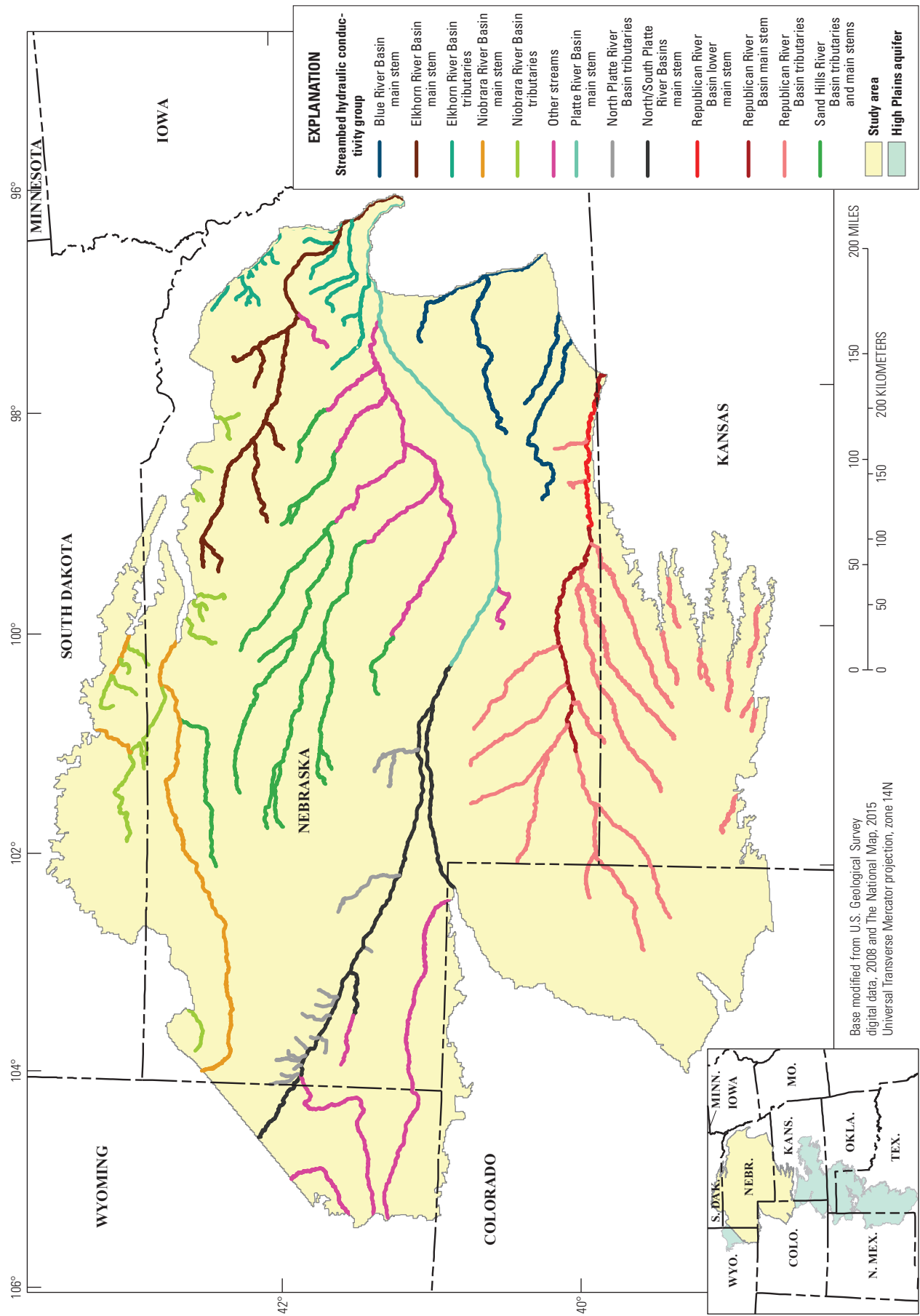


Figure 10. Simulated streams and streambed hydraulic conductivity groups for the Northern High Plains aquifer model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.

initial streambed hydraulic conductivity was assigned using 13 groups (fig. 10) based on the size of stream (tributary or main stem river) and general geographic area (Big Blue River Basin, Elkhorn River Basin, Niobrara-White River Basin, Loup River Basin, Platte River Basin including the North and South Platte Rivers, and Republican River Basin; fig. 1). In the downstream part of the Republican River, the regional Northern High Plains aquifer is only poorly connected to the river (Peterson, 2009), so an additional group was added for the decreased streambed hydraulic conductivity in the reaches corresponding to the poor connection. Streambed hydraulic conductivity assigned to each group ranged from 0.1 to 5 ft/d and was further adjusted during calibration. In addition, the Horizontal-Flow-Boundary Package (HFB; Harbaugh and others, 2000) was used to further limit the connection of the High Plains aquifer with the river and nearby alluvial deposits in downstream reaches of the Republican River (fig. 9) after Peterson (2009). The HFB Package limits the simulated flow of groundwater between adjacent MODFLOW-NWT cells by adjusting the inter-cell conductance using an assigned hydraulic characteristic. The hydraulic characteristic assigned to the three simulated HFB boundaries in the Northern High Plains aquifer model was equivalent to a 1,000-ft wide barrier with a hydraulic conductivity of 1 ft/d.

The Well Package (Harbaugh and others, 2000) was used to simulate groundwater withdrawals for irrigation. Calculation of groundwater withdrawal rates is described in the “Soil-Water-Balance Model” section. Recent (2010) groundwater withdrawals for municipal supply and well locations were compiled from available data (P. Bonebright, Nebraska Department of Natural Resources, written commun., 2012). Municipal groundwater withdrawals were simulated using the MODFLOW Multi-Node-Well Package (Konikow and others, 2009). Use of the Multi-Node-Well Package facilitated tracking simulated municipal withdrawals separately from irrigation withdrawals by representing each component in a separate MODFLOW-NWT package. Historical municipal withdrawals were estimated using population data (Forstall, 1996; United States Census Bureau, 2012) by calculating the proportions of historical population to current population and then applying the same proportions to calculate municipal withdrawals for the given time period relative to current withdrawals. This approach introduced some uncertainty to the municipal withdrawal rates, but data were not available to do otherwise, and municipal groundwater withdrawals only affected a small area near the wells. None of the municipalities involved in this component are large cities with large rates of groundwater withdrawal; therefore, the introduced uncertainty was assumed unlikely to have a measurable effect on simulation results in all other areas. Groundwater withdrawals for livestock, and for any other unmentioned categories, were assumed to be negligible.

Simulated stream base flows and simulated groundwater levels were output to separate files for analysis. The MODFLOW Gage Package (Niswonger and Prudic, 2005) was used to output simulated stream base flows at target locations

into separate formatted files. Though there is a MODFLOW input package that will extract simulated groundwater levels at specified times and locations, a PEST utility, *mod2obs.exe* (Doherty, 2009), was used to extract simulated water levels for comparison with calibration targets. *Mod2obs* was preferred because of the simpler method of inputting lateral and longitudinal coordinates of each measurement location, rather than having to input cell identifiers and having to calculate and input the offsets of the measurement location from the edge of the containing cell.

Groundwater-Flow Model Calibration Approach

This section of the report describes the approach to model calibration through use of PEST (Doherty, 2005) and the parameters that were adjusted to improve model calibration. In general, when groundwater-flow models are constructed, model inputs such as hydraulic conductivity, recharge, and aquifer base altitudes are either unknown and must be estimated or are partially known and must be interpreted spatially and temporally across the simulation area and through time. Outputs of the model, such as simulated groundwater levels and groundwater discharge to streams (base flow), are then compared against calibration targets—in this case, measured groundwater levels and estimated stream base flows. These data are described in more detail in the “Calibration Targets” section of this report. Usually, the initial model outputs do not closely reproduce all the calibration targets and, therefore, the model inputs must be adjusted so that the outputs more closely match the calibration targets. This process is known as “calibration.” Commonly, little information is available for the adjustable model inputs or the data are poorly constrained. Conversely, model inputs reflecting large amounts of information are usually fixed after model construction and are not adjusted during calibration. For this study, model parameters were explored systematically, and the final calibrated model, based on available information and best professional judgment, is presented in this report. Initial (uncalibrated) estimates of model inputs are presented for comparison with the final calibrated model.

The term “parameters” has a special connotation in this report and is defined as model inputs that were potentially adjusted by PEST (Doherty, 2005). Many thousands of model inputs exist, but only a small number (1,312) were selected as parameters for PEST. In general, these parameters consist of the following five groups: horizontal hydraulic conductivity estimated at pilot points, pre-1940 recharge from precipitation estimated at pilot points, maximum ET rate multipliers, streambed hydraulic conductivity, and spatial and temporal adjustments to 1940–2009 recharge. Other model inputs were set during construction, were not adjusted during calibration, and are not referred to as parameters.

Primary model calibration was completed using statistical techniques through parameter estimation using PEST (Doherty, 2005). Because the pre-1940 model and 1940–2009

model were sequentially linked, both models were calibrated simultaneously. This process generally followed Stanton and others (2010, appendix 2); parameters are discussed in the “Calibration Parameters” section, and targets are discussed in the “Calibration Targets” section. All parameters were subjected to Tikhonov regularization in the parameter estimation process. Regularization prevents overfitting and spurious parameter estimates by imposing a numerical penalty on the objective function for deviations from the initial estimates (Doherty, 2010). Additional information on parameter estimation and regularization using PEST are available from Doherty (2003), Doherty (2005), Hunt and others (2007), and Fienen and others (2009).

Calibration Targets

The Northern High Plains aquifer model was calibrated to minimize the sum of squared differences between calibration targets and simulated equivalents representing hydrologic quantities, in particular, groundwater levels and base flow to streams. Calibration targets are measured or estimated hydrologic data assumed to represent the actual hydrologic behavior of the groundwater system in the study area. For this study, the only widely available data were groundwater levels measured in wells and estimated stream base flows at streamgages (see Houston and others, 2013, for streamgage locations). For parameter estimation, weights are assigned to each target, to adjust for different magnitudes of the different kinds of measurements, and for greater or lesser certainty of the observations or estimated data used as targets; for example, an estimated stream base flow of 10 ft³/s has a target value of 864,000 cubic feet per day (ft³/d) because the MODFLOW-NWT model used units of feet for length and days for time. A groundwater level might have an altitude of around 3,000 ft, which is more than two orders of magnitude smaller than the target for estimated stream base flow; hence, if the weight assigned to the stream base flow target was two orders of magnitude smaller than the target value, both targets would have about the same influence on the calibration process.

For this study, the weights assigned to each target were based on error-based weighting (Hill and Tiedeman, 2007), where the target weight was assigned as the inverse of 1.96 (related to the 95-percent confidence interval) multiplied by the expected uncertainty associated with the particular type of measurement. For groundwater levels, the expected uncertainty was assigned as 10 ft for groundwater level targets for the pre-1940 model and 5 ft for groundwater level targets for the 1940–2009 model. The expected uncertainty assigned to groundwater level targets for the pre-1940 model was larger because of the wide range of dates associated with those targets, which were assumed to represent predevelopment groundwater conditions, defined for this study as being around 1940. Some of the targets were measured years after 1940 (Houston and others, 2013) but were thought to be in undeveloped areas and, hence, were still representative of pre-1940 conditions. Measurement error at streamgages

commonly ranges from 5 to 8 percent or greater (Rantz and others, 1982) and base flow separation probably has at least that much uncertainty; therefore, for estimated stream base flows, the expected uncertainty was estimated to be 10 percent of the streamgage mean estimated stream base flow except where otherwise noted in the “Estimated Stream Base Flows” section.

The objective function that PEST attempts to minimize during calibration is influenced by the number of targets in each group, the residual of each target in the group, measurement units of the hydrologic data, and the weight assigned to each target. Because of the far larger number of targets in the 1940–2009 groundwater level targets group, that group had an overwhelming influence on the objective function for PEST, obscuring important hydrologic signals provided by the other target groups, and preventing PEST from adequately estimating parameters to decrease the difference between calibration targets and simulated equivalents. Therefore, observation weights were modified during the calibration process to reduce the influence of 1940–2009 groundwater level targets on the calibration process so that the large number of those targets did not preclude accurate calibration to other target groups. Also, observation weights were modified to increase the influence of 1940–2009 stream base flow residuals on the calibration process overall, as well as for the Republican River near Orleans (USGS streamgage number 06844500) and the Republican River near Hardy (USGS streamgage number 06853500), in particular.

Groundwater Levels

Groundwater levels measured in wells used as calibration targets (Houston and others, 2013) totaled 8,149 groundwater level targets for the pre-1940 model and 334,918 groundwater level targets for the 1940–2009 model (fig. 11). Only a few groundwater-level measurements were made in the study area before 1940, and those measurements were not evenly distributed across the study area but tended to be only near streams; therefore, for the purposes of calibration, some groundwater levels used for pre-1940 model calibration (Houston and others, 2013) were measured after 1970 but were thought to be in undeveloped areas and hence were still representative of pre-1940 conditions.

Groundwater levels selected for use for 1940–2009 were those closest to April 30 or September 30 of each year, corresponding to the end of the seasonal stress periods for the 1940–2009 model. The number of groundwater level targets available for calibration varied considerably by year during 1940–2009, with a minimum of 464 groundwater levels in 1940 to a maximum of 8,118 groundwater levels in 1996 (fig. 11). Generally, more groundwater level targets were available for later years. During quality-assurance review, a number of measured groundwater levels seemed questionable (groundwater levels above land surface or below the aquifer base). These targets were assigned a weight of zero in the parameter estimation process so that they would not affect the

outcome. Other target groundwater levels were far different from all other nearby groundwater levels and, if they could not be verified, were also assigned a weight of zero. These weight reductions resulted in 675 of the 334,918 groundwater levels having a zero weight for parameter estimation. It is likely that additional more subtle erroneous values are still present in the groundwater levels used as calibration targets, but it was assumed that erroneous values were a small fraction of the total population and, hence, would have a negligible effect on the calibration results.

To evaluate groundwater level calibration results, the mean residual, the mean absolute residual, and the root-mean-squared (RMS) residual were calculated for all groundwater-level targets and also for subsets representing specific periods of the pre-1940 model and by decade for the 1940–2009 model. The mean residual was calculated as the sum of residuals of each group divided by the number of residuals in the group. The mean absolute residual was calculated as the sum of the absolute value of the residuals of each group divided by the number of residuals in the group. The RMS residual was calculated as:

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

where

- n is the number of residuals,
- h_m is the measured groundwater level at location i , and
- h_s is the simulated groundwater level at location i (Anderson and Woessner, 1992).

Estimated Stream Base Flows

Stream base flows used as targets for the pre-1940 model were estimated from flows recorded at a total of 26 streamgages (Houston and others, 2013; table 2). Whenever data were available for 1940, the target was estimated as the average estimated base flow for 1940 (10 targets); however, not all streamgages had data from 1940, so in those cases average estimated stream base flows from other years were used as calibration targets. For those 14 targets, estimated base flows were averaged from the beginning of the record until such time as shifts or trends appeared in the data—for instance, from 1946–70 for the Niobrara River near Gordon, Nebr. (streamgage number 06457500). It was assumed that averaging for longer periods would yield targets representative of steady-state conditions and, thus, be most representative of conditions before 1940. Although use of estimated stream base flows from other periods for calibration of the

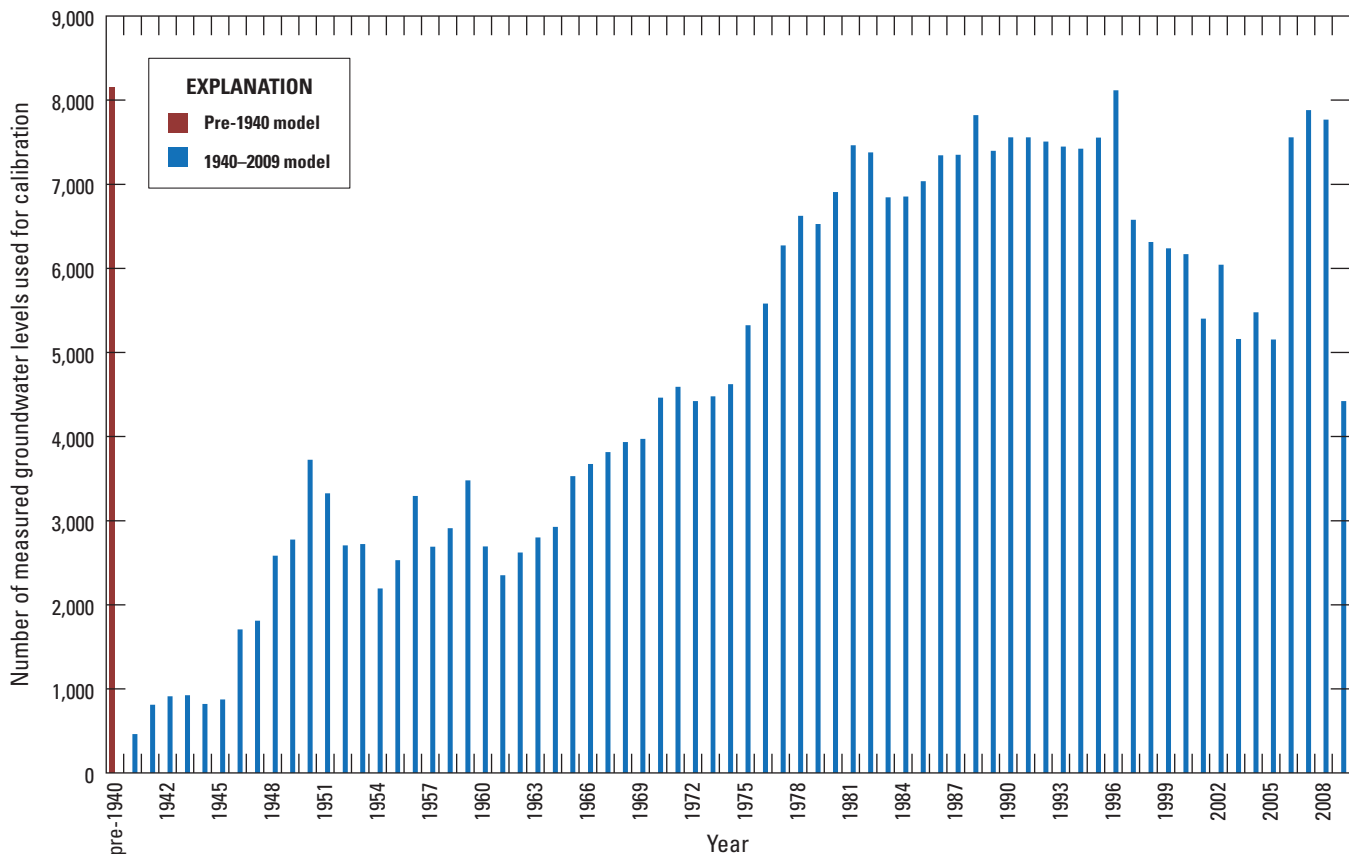


Figure 11. Count by year of groundwater-level measurements used as calibration targets for the Northern High Plains aquifer groundwater-flow model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.

pre-1940 model undoubtedly introduced uncertainty, data were not available to do otherwise, and the use of a target from a later period was considered superior to having no targets at all for those streamgages. In two cases, estimated base flows from years before 1940 were used as targets because all the post-1940 base flows had been affected by canal diversions or reservoir operations (North Platte River at North Platte, Nebr. [streamgage number 06693000]; Loup River near Genoa, Nebr. [streamgage number 06793000]). Only a few years of data exist for the Loup River near Genoa (table 2) before flows of that river were changed by diversions into the Loup Power Canal. Therefore, manual regression was performed by removing the largest and smallest 10 percent of the daily flows, and a conservative estimate of base flow was selected as the 15th percentile of the remaining daily flows.

Estimated stream base flow targets for the 1940–2009 model totaled 10,820 April or October monthly average values (Houston and others, 2013) to match the seasonal stress periods of the 1940–2009 model (May 1 through September 30, and October 1 through April 30). The seasonal stress periods allowed for simulation of the seasonal differences in recharge, ET, and groundwater withdrawals for irrigation, as described in the “Spatial and Temporal Discretization” section, and targets were estimated to match the end of each of these seasons. Data from 91 streamgages in total were used for calibration of the 1940–2009 model. Base flow estimates for some streamgages were reconstructed to account for the effects of canal diversions. Canal diversions upstream from a streamgage decreased the total flow of the stream and, therefore, decreased the estimated base flow; however, canal diversions were not explicitly simulated in this model and recharge from canal seepage was estimated separately and added to the simulation as recharge. Because it was assumed that water diverted upstream from a streamgage would have been base flow at the streamgage, the average daily diversion for the target period was added to the monthly estimated base flow target at the streamgage. This affected estimated base flows of the North Platte River at North Platte, Nebr. (streamgage number 06693000), the Platte River at Brady, Nebr. (streamgage number 06766000), the Platte River at Cozad, Nebr. (streamgage number 06766500), the Platte River at Odessa, Nebr. (streamgage number 06770000), the Platte River at Grand Island, Nebr. (streamgage number 06770500), the Loup River at Genoa, Nebr. (streamgage number 06793000), and the Loup River at Columbus, Nebr. (streamgage number 06794500). Targets associated with these streamgages were assigned weights corresponding to an increased uncertainty of 20 percent of the mean estimated base flow used as targets (as opposed to the default 10 percent uncertainty assigned to most base flow targets, as noted in the beginning of the “Groundwater-Flow Model Calibration Approach” section). The larger uncertainty was considered appropriate because of the addition of average, rather than

monthly, values for upstream canal diversions when reconstructing the monthly series of base flows for use as a target. Some estimated stream base flows were affected by reservoir operations, and data were not available to reconstruct base flow. In other cases, occasional extreme high flow events, such as sustained high runoff for much of April during spring snowmelt, affected the monthly average base-flow estimate so that it most likely did not represent groundwater discharge to the stream. These records were removed from the dataset, resulting in 10,154 estimates of monthly-average stream base flows for April or October that were used as calibration targets. A summary of all 750 monthly average estimated stream base flows for the major stream basins (and one smaller stream) of the Northern High Plains aquifer (fig. 1) is provided in table 3. For readers interested in additional detail, a summary of monthly estimated stream base flow for each streamgage of the Northern High Plains aquifer is provided in appendix 1.

To evaluate stream base flow calibration results, percent difference of each simulated base flow from each base flow target was computed as:

$$\text{Percent difference} = \ln \left(\frac{\text{estimated stream base flow}}{\text{simulated stream base flow}} \right) * 100, \quad (2)$$

the natural logarithm of the estimated base flow divided by the simulated base flow multiplied by 100, averaged by streamgage (table 3).

Calibration Parameters

A total of 1,312 parameters in 5 parameter groups (table 4) were adjusted during calibration to improve the match between calibration targets and simulated equivalents. Parameter groups included hydraulic conductivity estimated at pilot points (604 parameters); pre-1940 recharge from precipitation estimated at pilot points (665 parameters); adjustments to 1940–2009 recharge based on soil type, land-cover group, and decade (27 parameters); adjustments to maximum ET rate of shallow groundwater (3 parameters); and adjustments to streambed hydraulic conductivity based on stream size and physiography (13 parameters). Only adjustable model inputs, called parameters, were modified during calibration, as described in this section of the report. Other inputs specified in the “Groundwater-Flow Model Construction” section were not modified during calibration. The PEST process requires upper and lower bounds for each parameter. The parameter bounds used for this study were set to a realistic range by estimating the 95-percent confidence interval around the initial estimated value (Anderson and others, 2015), except for the limits for maximum ET rate multipliers. For additional information about the restricted upper bounds, see the “Maximum Evapotranspiration Rate Multipliers” section.

Table 2. Summary of estimated base flows used as calibration targets, by streamage number, and corresponding simulated pre-1940 base flows from the calibrated Northern High Plains aquifer model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.

[Nebr., Nebraska; residual signs: positive values are underestimates, negative values are overestimates]

Streamgage	Streamgage number	Latitude (degrees, minutes, seconds)	Longitude (degrees, minutes, seconds)	Estimated base flow, in cubic feet per second	Time period of averaged estimated values used for pre-1940 model target	Simulated base flow, cubic feet per second	Residual, cubic feet per second
Niobrara River near Gordon, Nebr.	06457500	42°38'23"	-102°12'38"	88	1946–70	139	-51
Niobrara River near Norden, Nebr.	06462000	42°47'13"	-100°02'06"	690	1953–80	720	-30
North Platte River at Bridgeport, Nebr.	06684500	41°40'38"	-103°05'47"	403	1940	652	-249
North Platte River at Oshkosh, Nebr.	06686500	41°22'56"	-102°20'53"	848	1940	701	147
North Platte River at North Platte, Nebr.	06693000	41°09'14"	-100°45'32"	857	March, 1931–39	939	-83
Lodgepole Creek at Bushnell, Nebr.	06762500	41°13'40"	-103°53'35"	9.0	1940	4.0	5.0
Lodgepole Creek at Ralton, Nebr.	06763500	41°02'00"	-102°24'18"	7.2	1952–63	2.3	4.9
South Platte River at North Platte, Nebr.	06765500	41°07'05"	-100°46'22"	86	1940	80	6.1
Platte River near Grand Island, Nebr.	06770500	40°52'28"	-98°16'54"	61	1940	935	-874
Middle Loup River at Dunning, Nebr.	06775500	41°49'51.3"	-100°05'58.3"	356	1946–70	225	131
Dismal River at Dunning, Nebr.	06776500	41°49'21"	-100°06'01"	301	1946–80	210	91
Middle Loup River at Arcadia, Nebr.	06779000	41°25'19"	-99°07'54"	562	1940	470	92
South Loup River at St. Michael, Nebr.	06784000	41°01'57"	-98°44'26"	147	1944–80	46	101
North Loup River at Taylor, Nebr.	06786000	41°46'37"	-99°22'45"	259	1940	290	-31
Calamus River near Burwell, Nebr.	06787500	41°48'37"	-99°10'59"	236	October–December, 1940	153	83
North Loup River near St. Paul, Nebr.	06790500	41°15'48"	-98°26'56"	452	1940	507	-55
Loup River near Genoa, Nebr.	06793000	41°25'07"	-97°43'25"	1,727	Manual regression of data from April, 1929–June, 1932	1,132	595
Platte River at North Bend, Nebr.	06796000	41°27'10"	-96°46'32"	2,086	1949–70	1,991	95
Elkhorn River at Norfolk, Nebr.	06799000	42°00'13.6"	-97°25'33.7"	262	1946–80	254	7.6
Elkhorn River at Waterloo, Nebr.	06800500	41°17'36"	-96°17'02"	524	October–November, 1950–59	300	224
Republican River near Orleans, Nebr.	06844500	40°07'54"	-99°30'09"	177	1948–65	178	-1.0
Republican River near Hardy, Nebr.	06853500	39°59'33"	-97°55'56"	171	1940	171	-0.4
Big Blue River at Seward, Nebr.	06880500	40°54'11"	-97°06'42"	22	1955–80	0.5	22
West Fork Big Blue River near Dorchester, Nebr.	06880800	40°43'52"	-97°10'38"	70	1965–80	52	18
Big Blue River near Crete, Nebr.	06881000	40°35'48"	-96°57'38"	119	1965–80	69	50
Little Blue River near Deweese, Nebr.	06883000	40°19'57"	-98°04'01"	68	1954–70	29	39
				SUM	SUM	10,250	337
				MEAN	MEAN	394	13

Table 3. Summary of estimated base flows used as calibration targets, by streamgauge number, and corresponding simulated 1940–2009 base flows for selected streamgages near the downstream end of major stream basins, and for Frenchman Creek at Culbertson, Nebraska, for the calibrated Northern High Plains aquifer model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.

[ft³/s, cubic foot per second; Nebr., Nebraska; residual signs: positive values are underestimates, negative values are overestimates]

Streamgauge	Streamgauge number	Estimated base flow (cubic foot per second)			Number of monthly averages	Simulated base flow at stress period end (cubic foot per second)			Comparative statistics		
		Average monthly	Minimum monthly average	Maximum monthly average		Average	Minimum	Maximum	Mean residual (ft ³ /s)	Median residual (ft ³ /s)	Average percent difference
Niobrara River near Norden, Nebr.	06462000	865	541	1,248	61	842	723	1,038	23	-0.1	1.1
Loup River at Columbus, Nebr.	06794500	2,273	1,433	3,519	76	2,223	1,199	2,700	49	75	1.4
Platte River at North Bend, Nebr.	06796000	3,026	923	7,026	121	3,777	2,164	9,267	-751	-697	-28
Elkhorn River at Waterloo, Nebr.	06800500	1,008	139	3,481	137	478	211	771	531	305	57
Frenchman Creek at Culbertson, Nebr.	06835500	67	0	207	138	71	13	141	-4.4	0.4	-21
Republican River near Hardy, Nebr.	06853500	308	31	1,235	128	225	53	405	83	41	12
Big Blue River near Crete, Nebr.	06881000	165	12	81	89	149	53	264	16	-6.6	-1.9

Table 4. Summary of the number of parameters by group used to calibrate the pre-1940 and 1940–2009 groundwater-flow models of the Northern High Plains aquifer in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.

[--, not applicable]

Parameter group	Number of parameters affecting each model	
	Pre-1940	1940–2009
Horizontal hydraulic conductivity estimated at pilot points	¹ 604	¹ 604
Streambed hydraulic conductivity	² 13	² 13
Maximum evapotranspiration rate multipliers	1	2
Pre-1940 recharge from precipitation	665	--
Spatial and temporal adjustments to 1940–2009 recharge	--	27

¹The same 604 parameters affected both models.

²The same 13 parameters affected both models.

Horizontal Hydraulic Conductivity Estimated at Pilot Points

Estimated hydraulic conductivity was sampled using 604 pilot points, distributed across the model area (fig. 6). Hydraulic conductivity was estimated for each pilot point individually, resulting in 604 parameters for calibration. The use of 604 points allowed a reasonable representation of the regional heterogeneity present in interpolated estimated hydraulic conductivity (fig. 6), yet the number of parameters was manageable, precluding overly long parameter-estimation execution time. The same estimated hydraulic conductivity map, generated from the 604 pilot points, was used for the pre-1940 and 1940–2009 models. Hydraulic conductivity was interpolated from pilot points across the model grid using kriging through the PEST utility *fac2real* (Doherty, 2009).

Maximum Evapotranspiration Rate Multipliers

Maximum ET rates were estimated for three time periods using a multiplier of the initial estimated ET: for the pre-1940 model (1 parameter) and for the 1940–2009 model (the irrigation and nonirrigation season stress periods each had a multiplier; 2 parameters). Because the initial ET rate was set to the NWS–PET rate (Stanton and others, 2011), and some of this ET demand would be met by soil moisture and replenished by precipitation rather than groundwater (Hall and Rus, 2013), it was assumed that groundwater discharge to ET was unlikely to constitute much more than 50 percent of the total ET in areas of shallow groundwater; therefore, the initial multiplier was specified as 0.50 for all three parameters, and an upper limit of 0.50 was also supplied to PEST so that larger values would not be accepted.

Streambed Hydraulic Conductivity

Streambed hydraulic conductivity was estimated for each of the 13 groups described in the “MODFLOW-NWT Inputs and Configuration” section, corresponding primarily to stream size (main stem or tributary) and major stream basins, plus one additional group representing the downstream section of

the Republican River (fig. 10). For the downstream section of the Republican River, streambed hydraulic conductivity was expected to be low, representing the poor connection of that river to regional groundwater flows of the Northern High Plains aquifer. Initial values were set based on preliminary testing using MODFLOW, and ranged from 5.1 ft/d for the Big Blue River to 0.1 ft/d for the poorly connected area of the downstream reaches of the Republican River with an average for all groups of 2.0 ft/d. The same streambed hydraulic conductivity parameters were used for the pre-1940 and 1940–2009 models, so adjustment to each of these parameters had the potential to affect outputs of both models.

Pre-1940 Recharge from Precipitation Estimated at Pilot Points

Pre-1940 recharge from precipitation was resampled from the average recharge for 1940–49 estimated with the SWB model (fig. 8A) using 665 pilot points. The resampling captured only the regional patterns of recharge (fig. 12) and not the small-scale detail of the estimated recharge map (fig. 8A), because the actual data input to MODFLOW-NWT are reproduced from interpolation of the pilot points to the model using kriging through the PEST utility *fac2real* (Doherty, 2009). However, the regional patterns of recharge were most important to capture, and it was assumed that the small-scale details would not have greatly benefitted the regional simulation of groundwater flow.

The 665 pilot points for pre-1940 recharge from precipitation were calibrated individually, resulting in 665 parameters for calibration. Pilot point values were then modified, based on preliminary testing using MODFLOW, until simulated pre-1940 stream base flow was about the same as the pre-1940 stream base flow targets. This modification resulted in a mean initial recharge of 0.8 in/yr with a range from 0.1 to 3.3 in/yr. Similar to the approach for horizontal hydraulic conductivity, the use of 665 pilot points produced a reasonable representation (fig. 12) of the regional heterogeneity in the estimated recharge from precipitation (fig. 8A) and provided

a small-enough number of parameters to prevent excessively long parameter-estimation execution time. Adjustment of pre-1940 recharge from precipitation allowed the calibration process to compensate for 1940–49 recharge potentially being different than multi-decadal recharge before 1940, as simulated in the pre-1940 simulation. The use of pilot points also allowed the calibration process to adjust for the effect of regional bias in the climate data on the estimated recharge from precipitation.

Spatial and Temporal Adjustments to 1940–2009 Recharge

Recharge for 1940–2009 was adjusted using 27 parameters: 5 parameters representing soil type; 21 parameters describing land-cover type, which changed for each decade; and 1 parameter to adjust irrigation canal seepage. Soil-type parameters were multipliers applied to 1940–2009 recharge from precipitation (5 parameters applied across all 138 stress periods; table 4) and were based on hydrologic soil groups (fig. 5; U.S. Department of Agriculture, 2006). From the initial four classes of soils, ranging from sand or loamy sand to clay loam or other clayey soils, a fifth type was defined by splitting soil hydrologic group A so that sandy soils south of the South Platte River (in the upper Republican River Basin, fig. 1) had a separate multiplier. Initial values of these multipliers were all set to 1.0.

Addends were used to increase or decrease 1940–2009 recharge from precipitation based on land-cover type (irrigated agriculture, dryland agriculture, or rangeland) and decade (21 parameters). Agricultural land-cover type could vary for every year of the simulation (fig. 2); however, the land-cover addends used to calibrate recharge from precipitation were adjusted by decade, so each land-cover addend was applied to the 20 stress periods within the respective decade. Initial values of these addends were set to increase recharge for irrigated agricultural cells by 2.0 in/yr for 1940–49, 1950–59, and 1960–69, and by 3.0 in/yr for 1970–79, 1980–89, 1990–99, and 2000–2009. Initial addends for dryland agriculture were set to increase recharge from precipitation by 0.5 in/yr from 1940 to 2009. Initial addends for rangeland were set to reduce recharge from precipitation by 1.0 in/yr from 1940 to 2009. Each time these adjustments were applied to the recharge, soil-type multipliers were applied to the estimated 1940–2009 recharge first, followed by the land-cover addends and the addition of canal seepage recharge; therefore, soil-type multipliers only affected the initial 1940–2009 recharge from precipitation, not the land-cover addends or canal seepage recharge. Multipliers and addends applied to different geographic areas (soil type versus land-cover type) and different times (addends were applied by decade, whereas multipliers applied to all periods); therefore, the MODFLOW-NWT 1940–2009 recharge was adjusted by a combination of

addends and multipliers. Additionally, this group of parameters included a multiplier to allow adjustment of the 1940–2009 canal seepage recharge rate (1 parameter) that was set to an initial value of 1.0.

The modification of 1940–2009 recharge using the parameters described (multipliers and addends) did not allow for subregional adjustments to the magnitude of recharge to account for regional bias in the initial estimate, as did the pilot-point parameters used to calibrate the pre-1940 recharge from precipitation. Therefore, whereas preliminary calibration for the pre-1940 recharge from precipitation resulted in about a 50-percent reduction in MODFLOW-NWT recharge (as compared to estimated) for most areas other than the Nebraska Sand Hills (fig. 1), the 1940–2009 MODFLOW-NWT recharge from precipitation in the driest parts of the study area produced too much simulated stream base flow. This was not surprising, given that the magnitude of the rate of recharge in these areas, about 0.2 in/yr or less (McMahon and others, 2011; Luckey and Cannia, 2006), is less than the uncertainty in the precipitation data input to the SWB model, and given that (preliminary) PEST-estimated pre-1940 recharge was much lower than 1940–2009 recharge, especially in the driest parts of the study area. Mapped precipitation can vary on the order of inches, depending on mapping methods and interpolation approaches (Stanton and others, 2011). Furthermore, as the parameters were defined, pre-1940 recharge from precipitation was estimated through calibration using PEST with pilot points as parameters, whereas 1940–2009 recharge from precipitation estimated by the SWB model was calibrated using soil-zone multipliers and either adding or subtracting recharge on different land-cover types by decade. The pre-1940 recharge from precipitation, calibrated using pilot points as parameters, provided more freedom to PEST to account for regional bias than did the calibration parameters (multiplication or addends) for adjusting recharge from precipitation for 1940–2009.

It was hypothesized, therefore, that adjustments to pre-1940 recharge from precipitation were compensating for uncertainties in precipitation in dry areas before 1940 and that similar changes would be appropriate for 1940–2009 in the areas where estimated 1940–2009 recharge was much larger than stream base flow (fig. 13). Pre-1940 adjustments were applied to 1940–2009 recharge by analyzing the difference between initial pre-1940 recharge from precipitation (fig. 8A) and the calibrated pre-1940 recharge from precipitation interpolated from pilot points and classifying the differences into seven groups based on magnitude. Average differences from the seven groups were then applied to the 1940–2009 recharge from precipitation for part of the model area (fig. 13). Differences were applied using a custom-written computer program so that the applied differences could change every time the calibrated pre-1940 recharge from precipitation changed.

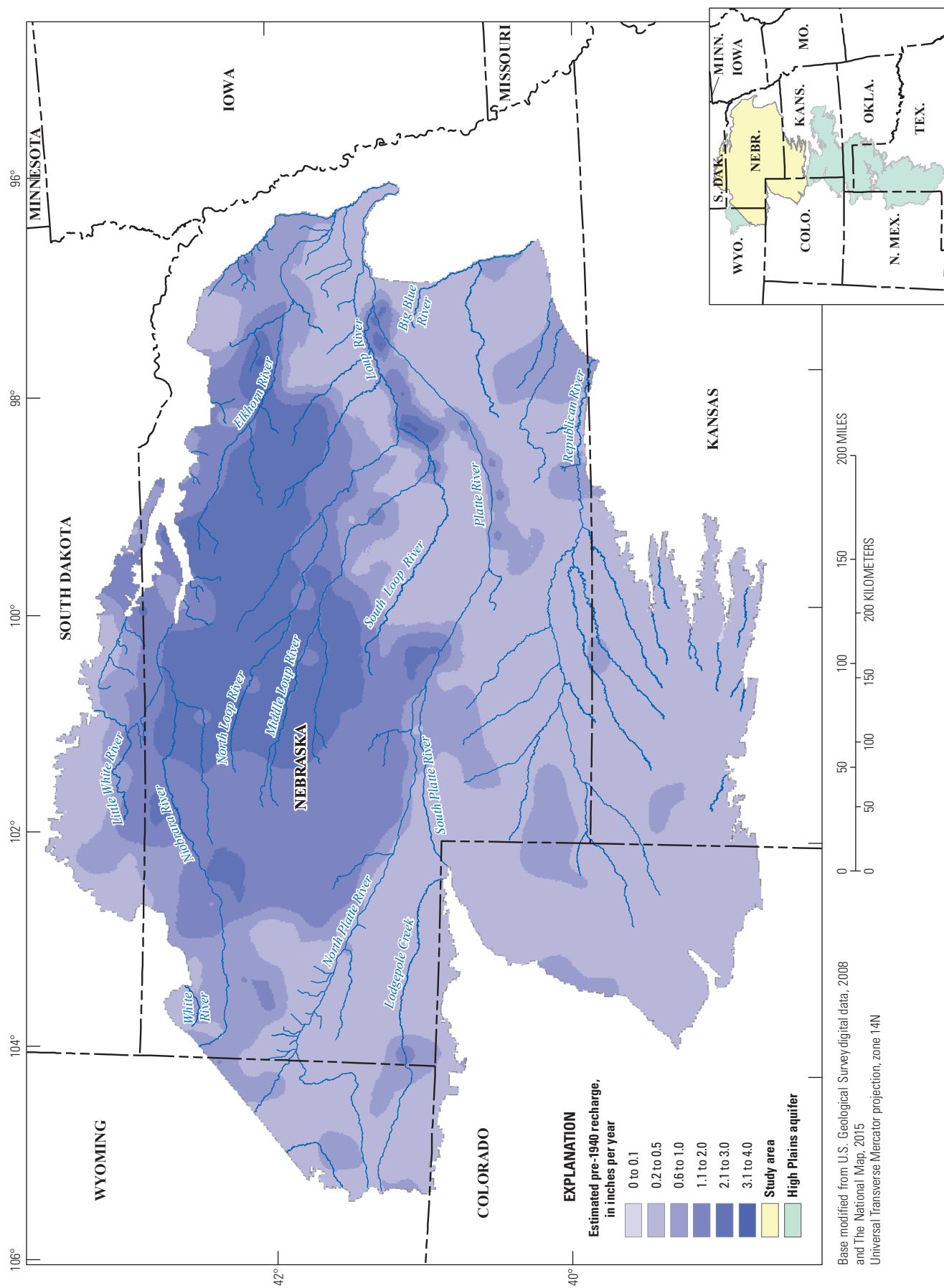


Figure 12. Initial recharge from precipitation, as interpolated from soil-water-balance estimates resampled at pilot points only for the pre-1940 groundwater-flow model of the Northern High Plains aquifer.

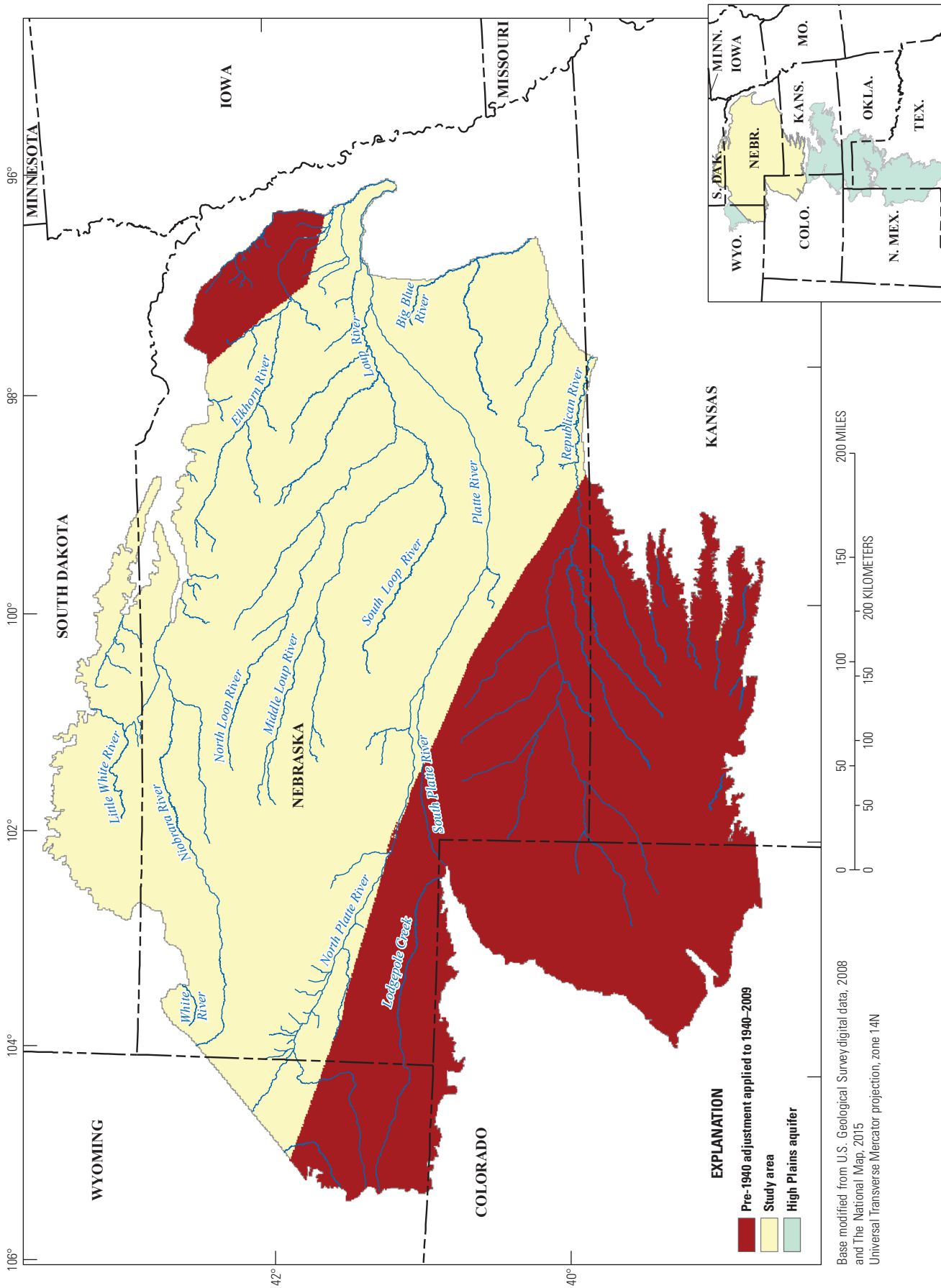


Figure 13. Areas where 1940–2009 recharge from precipitation was adjusted based on calibration results from the pre-1940 model of the Northern High Plains aquifer in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.

The application of these seven differences only for part of the model area (fig. 13) created an artifactual abrupt change in recharge where average differences were applied. Such abrupt changes are thought to be unnatural conditions, and natural transitions are assumed to be more gradual, but the approach did allow for correction of regional bias in the estimated 1940–2009 recharge from precipitation. Application of addends and factors to adjust MODFLOW-NWT recharge, and through categorized adjustments based on calibration of the pre-1940 recharge from precipitation together, caused the initial MODFLOW-NWT recharge from precipitation for 1940–49 and 2000–2009 to be 1.77 and 1.87 in/yr, respectively, which is a reduction of about 1.5 and 1.6 in/yr from their respective unadjusted estimates (3.3 and 3.5 in/yr).

Groundwater-Flow Model Calibration Results

This section of the report describes the results of the groundwater-flow model calibration through comparison of calibration targets to simulated equivalents, review of calibrated parameters, and presentation of simulated groundwater budgets of the calibrated model. This section of the report concludes with a discussion of composite parameter sensitivities.

Comparison of Calibration Targets to Simulated Equivalents

In all instances in this report, the term residual is defined as the calibration target minus the simulated equivalent for either groundwater levels or stream base flows; therefore, a positive residual indicates an underestimated simulated groundwater level or one that is less than the measured groundwater level, and a negative residual indicates an overestimated simulated groundwater level or one that is larger than the measured groundwater level. The same convention applies to residuals calculated from comparing stream base-flow calibration targets and simulated stream base flows.

Groundwater-Level Residuals

Groundwater-level residuals for the pre-1940 model and by decade for the 1940–2009 model are shown in figure 14. Most of the absolute values of the residuals for pre-1940 (represented by April 30, 1940) and by decade are less than 25 ft (fig. 14; table 5), which indicates that the models reasonably reproduce the regional and temporal trends of groundwater level targets. Local variations exist, with simulated groundwater levels above or below measured groundwater levels, but at the scale of the Northern High Plains aquifer, the residuals spatial and statistical distributions of calibration results demonstrate the lack of spatial bias of the model with regard to groundwater levels. Parts of the Northern High Plains aquifer in South Dakota and Wyoming seem to have consistently larger-magnitude residuals than other areas though they are not consistently biased in the same direction (that is, high or low).

These are also areas of relatively higher gradient, or slope of the water-table, than other parts of the model area (fig. 15). Evaluation of chi-squared statistics used to test for independence (Helsel and Hirsch, 2002) for average residuals for 2000–2009 indicated that there is likely a correlation between residuals being larger than 50 ft and being within 30 mi of the study area boundary; however, when the residuals from South Dakota and Wyoming were omitted, the chi-squared statistics indicated that there is likely not a correlation between residuals larger than 50 ft being within 30 mi of the study area boundary. For the study area other than South Dakota and Wyoming, wells with extremely high or extremely low residuals were not spatially concentrated (fig. 14); that is, simulated water levels were not biased particularly high or low in any region of the study area. Large residuals in South Dakota and Wyoming may have been because the regional groundwater model did not closely simulate the locally steep water-table gradients in these areas. Inaccurate simulation of the steep gradients in these areas and the resulting large residuals may have been influenced by a lack of process representation in the model; in some places, simulated groundwater levels were at altitudes greater than that of land surface (suggesting groundwater should have been discharging to land surface or recharge should have been rejected because groundwater was near or above land surface altitude). For actual natural conditions, either of these cases would not have resulted in increases in groundwater levels but are likely to cause seeps, springs, or increased overland flow towards the nearest surface water feature. These processes, however, were not represented within the Northern High Plains aquifer model; consequently, the groundwater levels unrealistically increased instead. Preliminary testing with the Unsaturated-Zone-Flow Package (UZP; Niswonger and others, 2006) for MODFLOW–2005 indicated that including rejected recharge and groundwater discharge to land surface in the models would have improved simulated water levels in South Dakota and Wyoming, but this inclusion also caused other problems in simulated base flow and water budgets that were beyond the scope of this study to resolve.

Figure 14. Groundwater-level residuals for the Northern High Plains aquifer groundwater-flow model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming. *A*, April 30, 1940. *B*, average 1940–49. *C*, average 1950–59. *D*, average 1960–69. *E*, average 1970–79. *F*, average 1980–89. *G*, average 1990–99. *H*, average 2000–2009. Available online at <https://doi.org/10.3133/sir20165153>.

Figure 15. Simulated groundwater levels for the Northern High Plains aquifer groundwater-flow model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming. *A*, April 30, 1940. *B*, average 1940–49. *C*, average 1950–59. *D*, average 1960–69. *E*, average 1970–79. *F*, average 1980–89. *G*, average 1990–99. *H*, average 2000–2009. Available online at <https://doi.org/10.3133/sir20165153>.

Table 5. Statistical summary of measured groundwater levels and simulated groundwater-level residuals for the Northern High Plains aquifer model for April 30, 1940, and by decade for 1940–2009.

[ft, foot; root-mean-squared residual is analogous to standard deviation]

Period	Number of calibration targets	Minimum measured groundwater level (ft)	Maximum measured groundwater level (ft)	Mean residual (ft)	Median residual (ft)	Minimum residual (ft)	Maximum residual (ft)	Mean absolute residual (ft)	Median absolute residual (ft)	Root-mean-squared (ft)	Percent of absolute residuals less than 25 ft
April 30, 1940	8,119	1,114	6,491	0.0	-0.6	-168	178	13	8.8	18	88
1940–49	13,690	1,158	7,003	-4.1	-5.8	-198	256	18	9.2	31	81
1950–59	29,581	1,178	7,005	-4.9	-6.5	-146	274	13	8.5	22	89
1960–69	32,321	1,135	6,803	-1.5	-2.3	-154	510	13	8.6	21	88
1970–79	52,912	1,114	6,735	-1.7	-2.3	-195	347	14	9.7	21	85
1980–89	72,403	1,108	6,722	-3.9	-3.8	-169	241	15	11.2	22	83
1990–1999	72,301	1,100	6,180	-5.2	-4.6	-225	193	16	12.3	23	80
2000–2009	61,035	1,096	6,793	-4.0	-4.1	-197	381	18	13.1	26	78

Summary statistics for groundwater level residuals for the pre-1940 model and by decade for the 1940–2009 model are shown in table 5. If simulated groundwater levels were unbiased, the mean residual would be expected to be near zero. The mean residual for pre-1940 and for each decade ranged from 0.0 to -5.2 ft, which indicates that although simulated groundwater levels tended to be on average slightly higher than groundwater level targets, the average difference was less than 6 ft. It is likely that the pre-1940 residuals were closest to zero because the use of pilot points to estimate pre-1940 recharge from precipitation allowed for a more locally specific calibration of recharge and, hence, smaller residuals than for the 1940–2009 model.

Minimum and maximum residuals by period in table 5 mainly ranged from -225 to 381 ft, except for the maximum residual for 1960–69 (510 ft). That particularly large residual related to a single measurement taken in 1964 at a well in Wyoming (U.S. Geological Survey, 2012, National Water Information System site 412341105023101) that was not matched by the simulated groundwater levels. No other measured groundwater levels are recorded for this site, and it is possible that this measured groundwater level is erroneous, that it represents some local condition not representative of regional groundwater flow, or that the simulation is inaccurate for that location. No nearby measured groundwater level data exist to either support or refute the single measurement.

The median absolute residual is resistant to the influence of outliers and provides information about central tendencies (Helsel and Hirsch, 2002). Median absolute residuals in table 5 ranged from 8.5 to 13.1 ft. Mean absolute residuals ranged from 13 to 18 ft, and RMS residuals, which are more heavily influenced by outliers, ranged from 18 to 31 ft (table 5). These measures of model fit indicate that errors are small compared with the range of measured groundwater levels used as targets, and that although local variations and large residuals may exist, the simulation appropriately represents the regional patterns of groundwater levels of the Northern High Plains aquifer, which is consistent with the purpose of the regional groundwater-flow model.

Graphical comparison of calibration targets against simulated equivalents provides another visual analysis of model fit. Measured groundwater level targets and simulated groundwater levels for the pre-1940 and 1940–2009 models are compared in figures 16A and 16B, respectively. Also, figures 16A and 16B show the 1:1 line for reference; if all simulated groundwater levels perfectly matched groundwater-level targets, all points would lie on the 1:1 line. Generally, the scatter of points in figures 16A and 16B is spread equally small distances above and below the 1:1 line, demonstrating that although the groundwater levels did not perfectly match, simulated groundwater levels were equally balanced above and below groundwater-level targets. Only for groundwater-level targets above about 6,400 ft altitude, simulated water levels tended to be near to but slightly underestimated of the groundwater-level targets, as evidenced by their position slightly below the 1:1 line.

Stream Base-Flow Residuals

The Northern High Plains aquifer groundwater-flow models simulated about the same amount of stream base flow as was estimated overall for pre-1940 (table 2) and 1940–2009 (tables 3 and 6; fig. 16C). Estimated stream base flows tended to be more variable during short periods, but simulated stream base flows were about correct as to magnitude and multiyear trend (figs. 17A–17G; see Houston and others, 2013, for streamgage locations). For three of six major river basins, simulated base flow was less than 2 percent different from estimated base flow, and for a fourth, simulated base flow was about 12 percent different from estimated base flow. Small percent differences indicate that the models simulated about the right amount of regional groundwater discharge to streams (stream base flow), consistent with the purpose of a regional groundwater model. Only results for selected streamgages and summary statistics are presented in the main body of the report, as an indication of the overall fit of simulated stream base flows to estimated stream base flows; simulated and estimated base flows for the remainder of the streamgages are summarized statistically and graphically in appendix 1 for readers interested in comparisons at additional locations.

The mean base-flow residual for the pre-1940 model was 13 ft³/s (table 2). The positive sign of the mean residual indicated that, on average, simulated base flows were less than estimated base flows. The largest residual (–874 ft³/s) was for the Platte River at Grand Island, Nebr. (streamgage 06770500), where the calibration target flow was estimated from the average base flows of 1940; however, the 1940 average base flow for that streamgage was unusually low (61 ft³/s) compared with the 1933–2009 average for that streamgage (888 ft³/s). It is possible that the 1940 estimated base flows were caused by multi-year drier climatic conditions rather than being indicative of the decadal or longer equilibrium preceding 1940.

A graphical comparison was done for 1940–2009 simulated stream base flows against calibration targets (fig. 16C). Generally, the points in figure 16C are evenly distributed above and below the 1:1 line, indicating that simulated stream

base flows were larger and smaller than estimated stream base flows used as calibration targets and were not biased in either direction. Only for points above about 6,000 ft³/s, simulated stream base flows tended to be smaller than estimated stream base flows, as evidenced by their position below the 1:1 line.

The mean base-flow residual for the 1940–2009 model was 30.4 ft³/s; however, a mean base-flow residual is difficult to interpret because of the large range of estimated seasonal average stream base flows (from 0 to greater than 3,000 ft³/s) and the large amount of data (10,154 stream base flow estimates for the 89 streamgages). When grouped by size of estimated base flow, however, for 12 streamgages with mean estimated base flows within the range from 0 to 10 ft³/s, the mean residual was 2.7 ft³/s and the median was 0.7 ft³/s; for 10 streamgages with mean estimated base flows within the range from 1,000 to 10,000 ft³/s, the mean residual was 180 ft³/s and the median was 170 ft³/s (table 6). Median absolute residuals provide information about the total difference between simulated and estimated base flows because differences in sign do not cancel out. The mean of median absolute residuals for the streamgages in each size range does not increase as quickly as the maximum base flow range for each class (table 6), indicating that larger simulated stream flows match base flow estimates closer than simulated flows for smaller streams.

Overall, the proportionate size of the residual relative to the size of the mean estimated base flow decreased as the mean estimated base flow increased, indicating that the model simulated larger streams (especially those with mean estimated base flow larger than 50 ft³/s) more accurately than smaller streams. This is consistent with the purpose of the model in representing regional patterns of groundwater flow and discharge to streams. Smaller estimated base flows, below 50 ft³/s, may have been less well represented because of the regional nature of the simulation, whereas larger estimated base flows included some estimates that were much larger than neighboring seasons or years, and which may not have been representative of base flow from regional groundwater discharge.

Table 6. Summary of simulated stream base-flow residuals for the 1940–2009 Northern High Plains aquifer model by range of estimated base flow.

[Residual sign: positive values are underestimated, negative values are overestimated. ft³/s, cubic foot per second]

Mean estimated base flow range	Number of stations	Mean residual (ft ³ /s)	Median residual (ft ³ /s)	Mean of median absolute residuals (ft ³ /s)
0–10	10	2.7	0.7	2.1
10–50	20	7.3	4.9	11
50–100	17	11	5.1	28
100–500	21	33	16	54
500–1,000	11	–28	–20	220
1,000–10,000	10	180	170	410

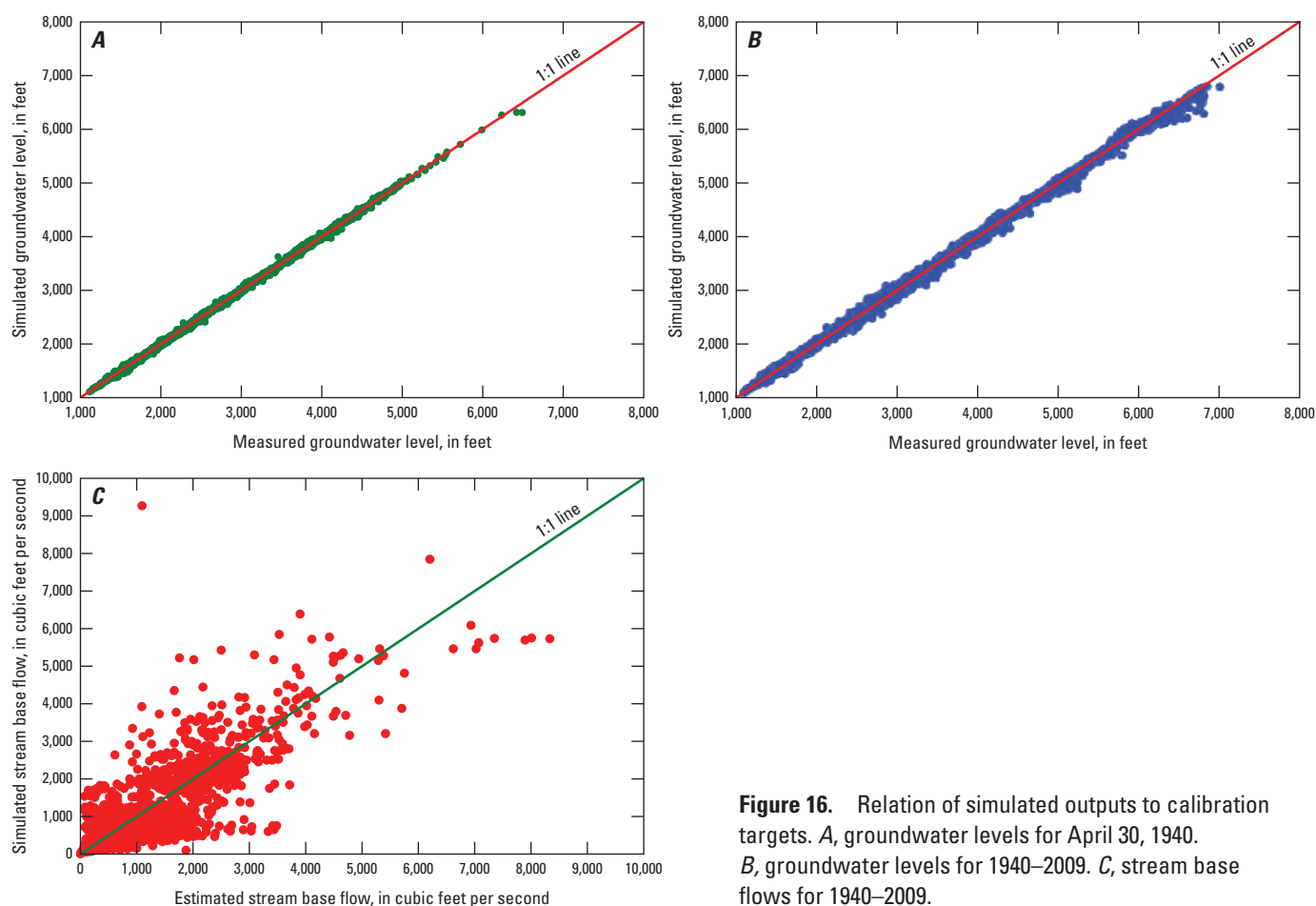


Figure 16. Relation of simulated outputs to calibration targets. *A*, groundwater levels for April 30, 1940. *B*, groundwater levels for 1940–2009. *C*, stream base flows for 1940–2009.

Summary statistics for simulated and estimated 1940–2009 stream base flows at the lower end of major stream basins within the study area (fig. 1) provide additional insight on the overall fit of the calibrated model to targets (table 3). In addition, a graphical comparison of simulated base flow to base flow targets (figs. 17A–17G) furnishes additional context to illustrate how the Northern High Plains aquifer model adequately simulated regional groundwater discharge to streams from 1940 to 2009 in a general sense, if not its seasonal or annual variations. To reduce the influence of seasonal variation in flows that happened at some locations, lowess trend lines (Cleveland, 1979; Cleveland and Devlin, 1988) were used as a graphical technique to evaluate whether trends of simulated and estimated base flows were similar, with similarity taken to indicate efficacy of the groundwater model in matching regional groundwater discharge to streams (fig. 17).

Simulated base flow for the Niobrara River near Norden, Nebr. (streamgage number 06462000), averaged 1.1 percent less than estimated base flow (table 3); however, estimated base flow from 1965 to 1987 (end of record) was affected by operations of Merritt Reservoir on the Snake River (fig. 1), a tributary to the Niobrara River about 50 miles upstream from the Norden streamgage. Estimated stream base flows after 1965 have larger variability than before 1965, whereas

simulated base flow after 1965, at around 800 ft³/s, is reasonably similar to estimated and simulated base flow before 1965 (800–1,000 ft³/s; fig. 17A). The Northern High Plains aquifer model did not explicitly simulate reservoir operations, so a disparity in the simulated and estimated base flows during periods affected by reservoir operations was expected. Approaches exist to simulate runoff, base flow, and reservoir and canal operations, but these were beyond the scope of this study. Lowess trend lines for simulated and estimated flows were similar for most of the record.

The base-flow estimates for Loup River at Columbus, Nebr. (streamgage number 06794500), were reconstructed estimates made necessary because of year-round diversions from the Loup River into a canal upstream from this streamgage; hence, the period of calibration target flows was affected by canal operations (fig. 17B). Simulated base flow, however, was only 1.4 percent smaller than estimated base flow on average (table 3), and lowess trend lines indicate that the simulation generally represented the overall pattern of multiyear increases and decreases in estimated base flow from about 1943 to 1978 (end of record). In contrast, the simulated base flow of the Platte River at North Bend, Nebr. (streamgage number 06796000), was 28 percent larger than estimated base flow on average (table 3), but all estimated base flows

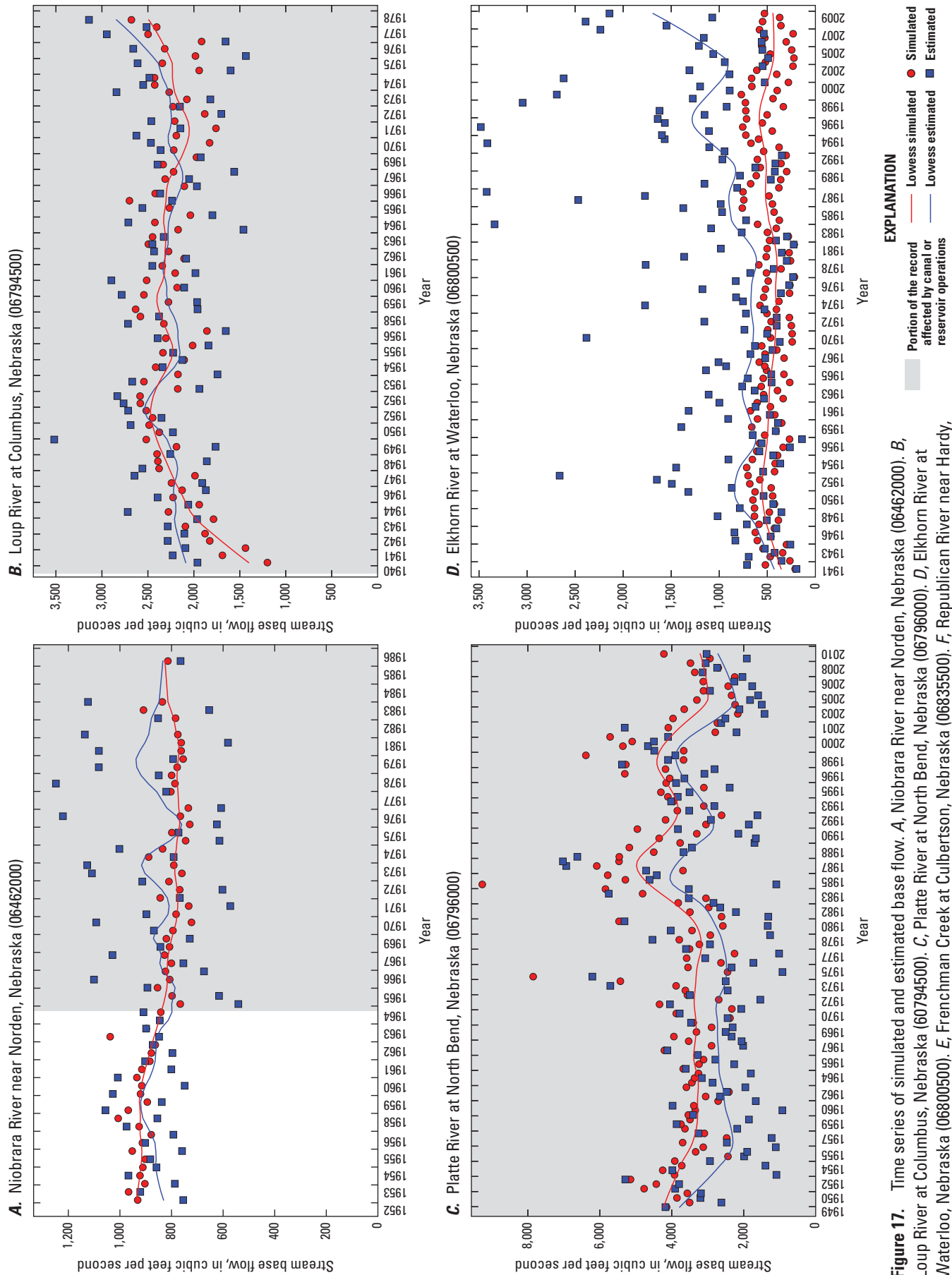


Figure 17. Time series of simulated and estimated base flow. *A.* Niobrara River near Norden, Nebraska (06462000). *B.* Loup River at Columbus, Nebraska (06794500). *C.* Platte River at North Bend, Nebraska (06796000). *D.* Elkhorn River at Waterloo, Nebraska (06800500). *E.* Frenchman Creek at Culbertson, Nebraska (06835500). *F.* Republican River near Hardy, Nebraska (06853500). *G.* Big Blue River near Crete, Nebraska (06881000).

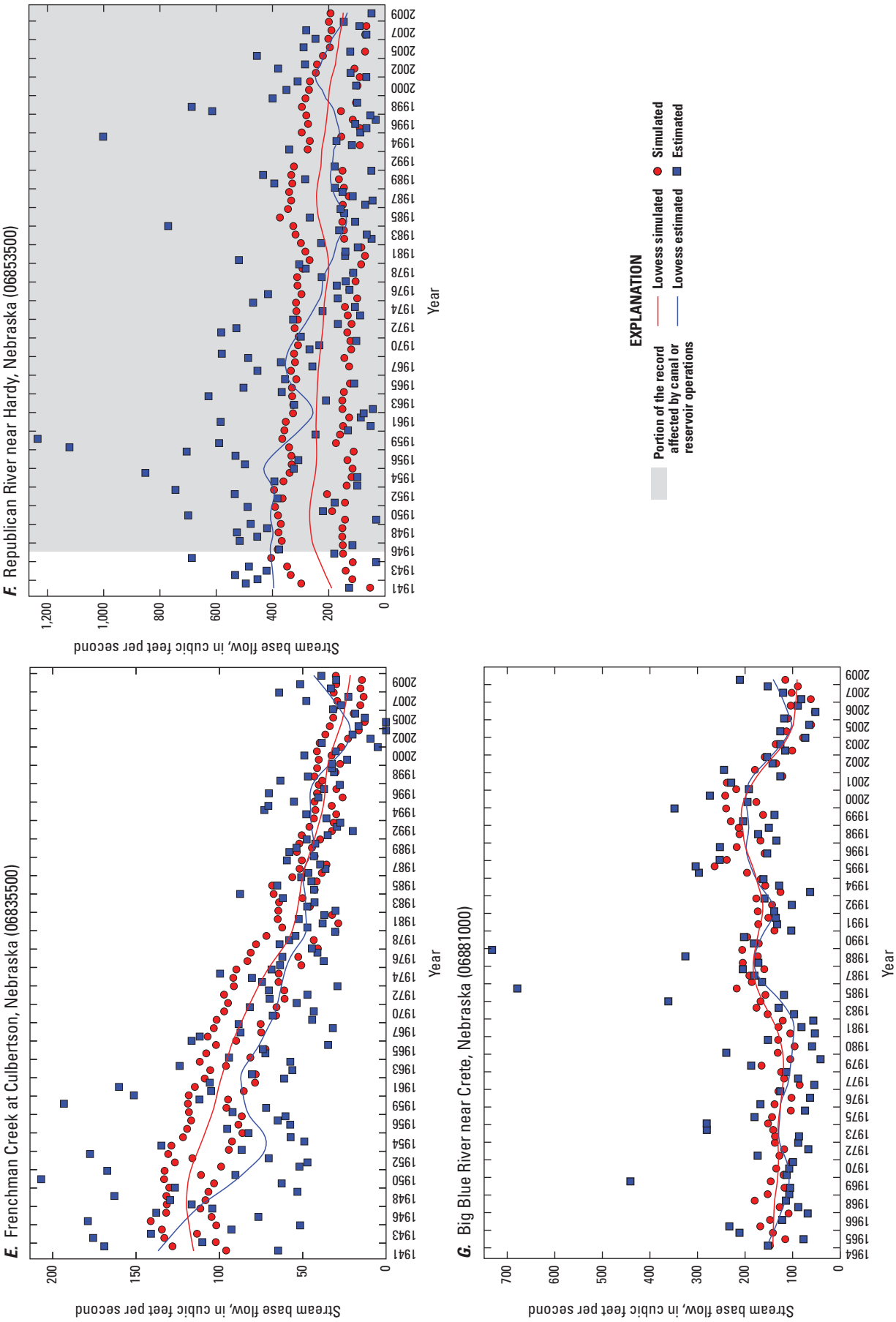


Figure 17. Time series of simulated and estimated base flow. *A*, Niobrara River near Norden, Nebraska (06462000). *B*, Loup River at Columbus, Nebraska (60794500). *C*, Platte River at North Bend, Nebraska (06796000). *D*, Elkhorn River at Waterloo, Nebraska (06800500). *E*, Frenchman Creek at Culbertson, Nebraska (06835500). *F*, Republican River near Hardy, Nebraska (06853500). *G*, Big Blue River near Crete, Nebraska (06881000).—Continued

were affected by reservoir operations (Lake McConaughy); moreover, estimated base flow is highly variable from season to season and year to year (fig. 17C). Lowess trend lines, however, show that the overall trends of simulated and estimated base flow were similar from 1949 to 2009, and simulated base flows track the multiyear trends in the estimated stream base flows, though simulated base flow was larger than estimated base flow.

The Elkhorn River at Waterloo, Nebr. (streamgage number 06800500), had the largest average percent difference of any of the streamgages reported in table 3 (57 percent), but for most values less than 1,000 ft³/s, simulated and estimated flows reasonably agree (fig. 17D). Many estimated base flows at this streamgage were in the range from 1,000 to 2,000 ft³/s, and several more were from 2,000 ft³/s to greater than 3,000 ft³/s; but base flows exceeding 1,000 ft³/s were not represented in the simulation results. Residuals for base flows greater than 1,000 ft³/s certainly skew the estimated mean residuals and percent differences, but those estimated base flows may not be indicative of regional groundwater discharge. Anomalously large residuals may be the result of short-term (seasonal or annual) conditions or reflect failure of the base-flow estimation technique to distinguish gradually varied seasonal high-flow conditions from steady groundwater discharge to streams (base flow). Lowess trend lines show that simulated stream base flow for the Elkhorn River at Waterloo was about 500 ft³/s for most of the simulation period, whereas estimated base flow started near 500 ft³/s and increased later in the record, especially after 1981.

Though a main purpose of the models was to capture regional patterns of groundwater occurrence and flow, and the models better represent larger streams than smaller streams (table 6), the simulated base flows also represented the distinct hydrologic signals present in some smaller tributary streams, such as Frenchman Creek at Culbertson, Nebr. (streamgage number 06835500). In 2009, Frenchman Creek at Culbertson, Nebr., had much less base flow than the rest of the streamgages presented in the main body of this report but was included in this section to exemplify a tributary stream that underwent an important and distinct hydrologic decline in estimated stream base flow from 1941 to 2009. In 1960, estimated base flow of Frenchman Creek at Culbertson, Nebr., was about 80 ft³/s, similar to the estimated flow rate of the Big Blue River near Crete, Nebr. (streamgage number 06881000; fig. 17G). In contrast, in 2009, estimated stream base flow was less than 30 ft³/s. Overall, simulated base flow of Frenchman Creek at Culbertson, Nebr., was 21 percent larger than estimated base flow on average (table 3), but the Northern High Plains aquifer 1940–2009 model also reasonably captured the decline in stream base flow from 1941 to 2009 (fig. 17E). Lowess trend lines of estimated and simulated base flow show the same declining trends and are mostly close to each other, especially from 1963 to 2008.

Simulated base flow of Republican River near Hardy, Nebr. (streamgage number 06853500), was 12 percent smaller than estimated base flow on average (table 3), but lowess trend

lines indicate that simulated base flows are generally similar to estimated base flows, especially for 1972–2009 (fig. 17F). Before 1972, simulated base flows are consistently smaller than estimated base flows. At this streamgage, estimated base flows greater than 600 ft³/s may not be indicative of regional groundwater discharge to streams (base flow) or could represent misidentification of runoff as base flow. In addition, reservoir operations at Harlan County Lake (fig. 1) from 1952 to 2009 potentially affected estimated stream base flow. The series of smaller simulated base flows at this streamgage (around 100–200 ft³/s) were simulated during irrigation seasons, whereas the larger simulated base flows (200–400 ft³/s) are for nonirrigation seasons. Estimated stream base flows contained similar smaller values for irrigation seasons, but nonirrigation season estimated base flows contained much more variability than did simulated stream base flows. The mean percent difference between simulated and estimated stream base flows for irrigation seasons was 6.5 percent.

Simulated base flow of Big Blue River near Crete, Nebr. (streamgage number 06881000), was just 2 percent larger than estimated base flow on average (table 3). Simulated base flow matched the magnitude of estimated base flow for most of the record, with the exception of several seasons (such as in 1969, 1984, and 1988) when estimated base flow was much greater than the rest of the seasons (fig. 17G). Lowess trends of multi-year increases or decreases in estimated base flow were similar from 1964 to 2009.

Calibrated Parameters

This section of the report describes the calibrated values of the 1,312 parameters across 5 groups that produced model outputs that most closely matched the calibration targets. This section also describes comparison of calibrated parameter values to initial parameters. Generally, the calibrated parameters represent refinements of the initial parameter estimates; for instance, calibrated horizontal hydraulic conductivity estimated at pilot points only changed in selected sub-regional areas, but the regional patterns present in the initial estimates were preserved in the calibrated parameters, and the regional average horizontal hydraulic conductivity only increased slightly. Perhaps the largest calibration changes were for adjustments to pre-1940 recharge from precipitation and for spatial and temporal adjustments to 1940–2009 recharge from precipitation, which tended to decrease recharge in the driest parts of the study area, and increase recharge on cells containing dryland or irrigated agriculture.

Calibrated Horizontal Hydraulic Conductivity Estimated at Pilot Points

Calibrated hydraulic conductivity estimated at pilot points and interpolated to the model grid (fig. 18) maintained the general trends and magnitude of the initially estimated hydraulic conductivity (fig. 6). Calibrated horizontal hydraulic conductivity of the aquifer averaged 35 ft/d (or about 5 ft/d

greater than initial values) and ranged from 2 to 285 ft/d. Regional patterns in the estimated hydraulic conductivity (fig. 6) are preserved in the calibrated values (fig. 18), such as the larger values of hydraulic conductivity adjacent to the Republican River and near the confluence of the Loup and Platte Rivers. Calibrated hydraulic conductivity tended to expand the areas of low estimated hydraulic conductivities, but low-to-moderate values remained typical of much of the north-central part of the study area. Many of the differences between the calibrated horizontal hydraulic conductivity (fig. 18) and the initial values (fig. 6) are local increases of conductivity, such as in parts of the Big Blue River Basin, the upper Republican River Basin in Colorado, areas adjacent to the upper Elkhorn River, and a small area adjacent to the North Platte River. The largest values of calibrated horizontal hydraulic conductivity (fig. 18), adjacent to the Republican River and near the confluence of the Loup and Platte Rivers, are larger than their initial estimates (fig. 6), in which the estimation technique (Houston and others, 2013; see “Aquifer Physical Characteristics” section) had limited the maximum value. The calibrated values also are consistent with other modeling studies completed for parts of this area (Carney, 2008; Luckey and Cannia, 2006; Peterson, 2009; Peterson and others, 2015; Stanton and others, 2010). In addition, because of regularization, the calibrated horizontal hydraulic conductivity was not “over-fit” to improve model calibration, in which case figure 18 would have exhibited local “bulls-eyes,” where calibration at individual pilot points caused values much larger than surrounding values, contrary to other indications (Doherty and Hunt, 2010).

Calibrated Maximum Evapotranspiration Rate Multipliers

Multipliers on the maximum ET rate remained at the initial value of 0.50 for all three instances; that is, the single value for the pre-1940 model (1 parameter) and separate values for irrigation season and nonirrigation season stress periods of the 1940–2009 model (2 parameters). This initial value was also the upper limit for the parameter because larger values were not considered defensible; however, because these parameters remained at the upper limit, it is possible that further increases in the maximum ET rate might have improved model calibration, perhaps by mitigating the potential overestimation of recharge from SWB modeling in areas of shallow groundwater.

Calibrated Streambed Hydraulic Conductivity

Calibrated streambed hydraulic conductivity (*CSFRK*) for the 13 groups established as parameters (fig. 10) ranged from 0.1 ft/d for the area of poor connection between the river and the aquifer along the downstream reaches of the Republican River to 10 ft/d (the upper limit for these parameters) for the Elkhorn River and its tributaries. The *CSFRK* was larger than initial streambed hydraulic conductivity for 6 of the 13 groups, was smaller for 5 of the 13 groups, and about the same for the remaining 2 groups (Loup River Basin and principal streams

of the Niobrara River Basin). All *CSFRK* values were within defensible ranges. Because two of the *CSFRK* values were at upper parameter limits, it is possible that further increases above *CSFRK* for those groups might have further improved model calibration; however, further increases might have been only mitigating other errors or omissions in the conceptual model of groundwater flow. For instance, further increases might have been mitigating errors in the recharge from precipitation estimated with SWB because SWB does not account for shallow water tables, such as those that commonly exist near the Elkhorn River; therefore, no additional increases of *CSFRK* were allowed beyond the initial limits.

Calibrated Pre-1940 Recharge from Precipitation Estimated at Pilot Points

Calibrated pre-1940 recharge from precipitation (fig. 19) ranged from 0.1 to 3.9 in/yr with a mean of 0.9 in/yr, which is an increase of about 0.1 in/yr above the initial mean values (fig. 12). Changes from initial values ranged from an increase of 1.9 in/yr to a decrease of 0.5 in/yr, and 87 percent of the calibrated values (fig. 19) were within 0.25 in/yr from the initial values (fig. 12). Although a parameter change of 0.25 in/yr might seem small, for perhaps 40 percent of the area, estimated and calibrated recharge were less than 0.1 in/yr (figs. 12 and 19). Recharge increased for a few areas near the Republican River in southern Nebraska near the Kansas state line, in the upper Republican River Basin, and at the westernmost part of the Republican River Basin and in easternmost Colorado.

Calibrated Spatial and Temporal Adjustments to 1940–2009 Recharge

Calibrated spatial and temporal adjustments to 1940–2009 recharge for soils (fig. 5) generally decreased recharge from precipitation, whereas addends related to land cover either increased or did not change recharge. While this may appear contrary, the effects of these changes did not counteract each other because they were applied to different areas. Multipliers applied by soil type to adjust the 1940–2009 recharge from precipitation were on average 0.87 and ranged from 0.05 (for clay loam soils) to 2.0 (for clay soils and sandy soils south of the South Platte and Platte Rivers). The calibrated multiplier for the rest of the sandy soils was 0.88, and the multiplier for silty soils was 0.78.

Addends used to increase or decrease 1940–2009 recharge from precipitation based on combinations of land-cover type (for example, irrigated agriculture, dryland agriculture, and rangeland; fig. 3) and decade ranged from -2.0 (estimated for nonagricultural cells from 1960 to 1969) to 6.8 (applied to irrigated cells from 1980 to 1989). The 1940–2009 mean values of addends were 4.7 in/yr for irrigated agriculture cells, 1.1 in/yr for dryland agriculture cells, and 0.0 in/yr for nonagricultural cells. Addends applied to irrigated agriculture cells were smaller for 1940–69 (mean 3.2 in/yr) than for 1970–2009 (mean 5.8 in/yr). Addends applied to dryland agriculture cells were smaller for 1940–69 and 2000–2009

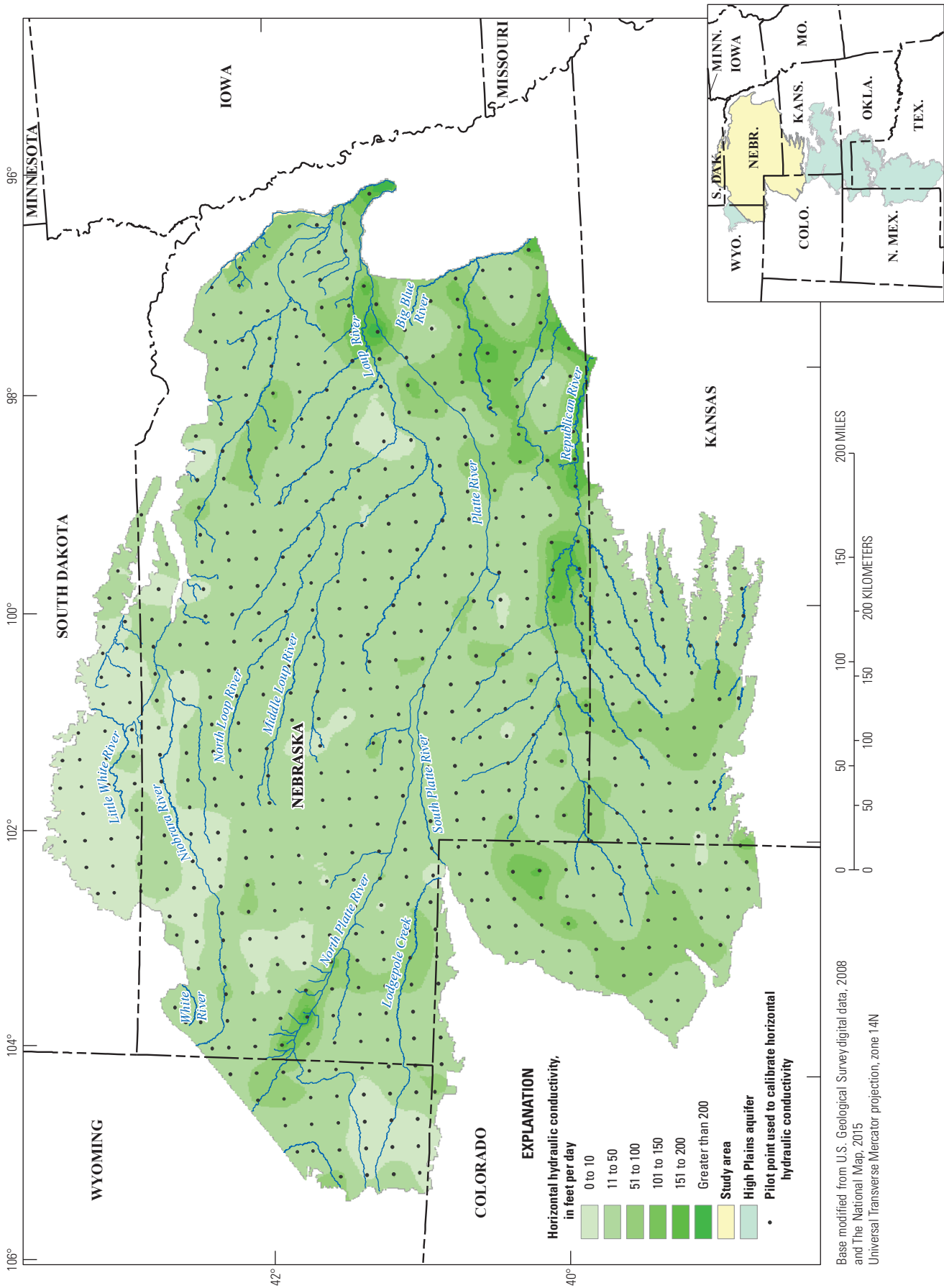


Figure 18. Calibrated horizontal hydraulic conductivity from the Northern High Plains aquifer groundwater-flow models (pre-1940 and 1940–2009) in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.

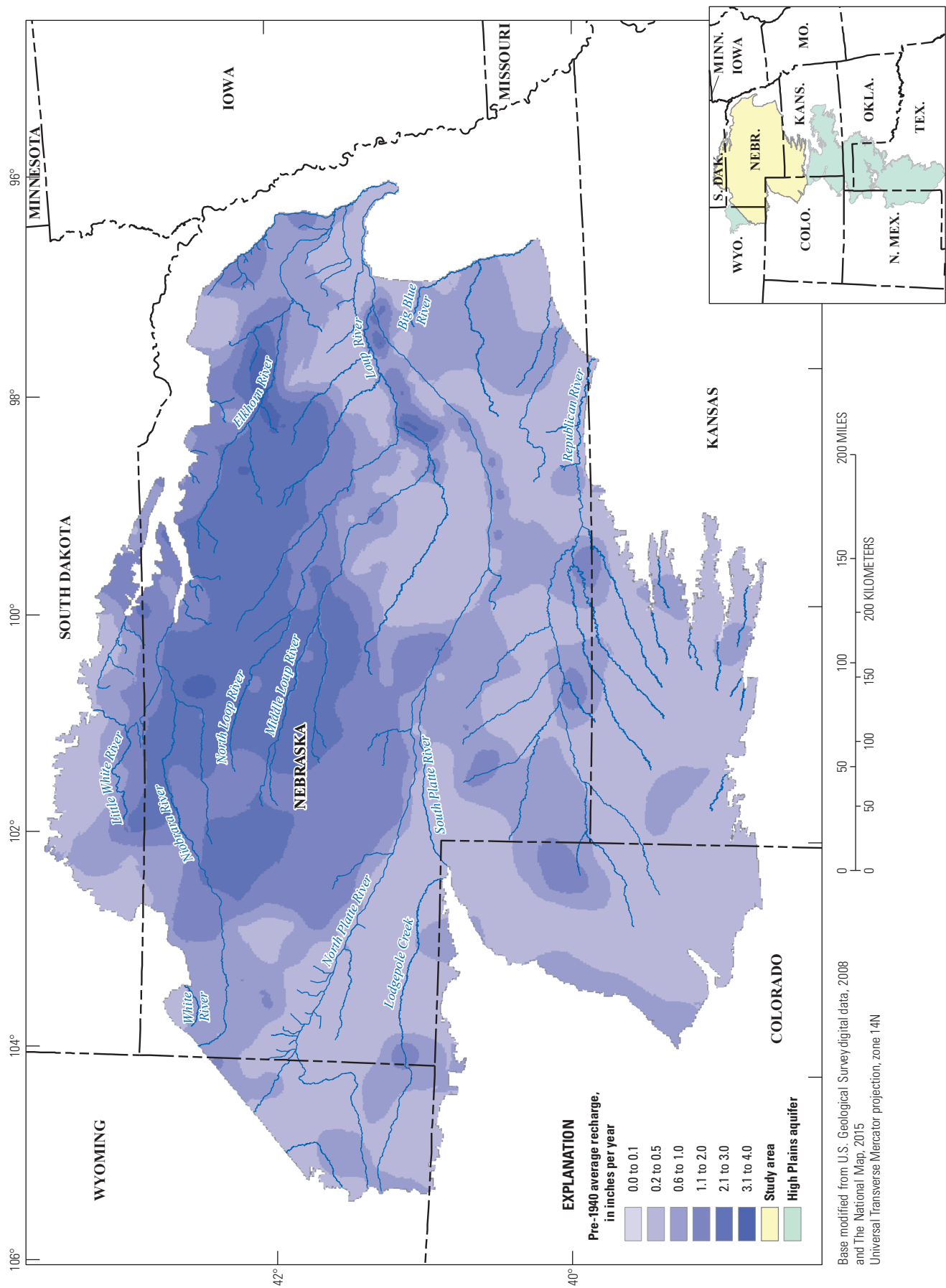


Figure 19. Calibrated pre-1940 recharge from precipitation for the Northern High Plains aquifer groundwater-flow model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.

(mean 0.6 in/yr) than for 1970–99 (mean 1.9 in/yr). Addends applied to nonagricultural cells did not exhibit an increasing or decreasing temporal trend for 1940–2009 but were the largest for 1940–49 (2.0 in/yr) and smallest for 1960–69 (–2.0 in/yr). Additionally, the calibrated multiplier estimated for the 1940–2009 canal-seepage recharge rate was 0.82 and was probably within the unknown range of uncertainty of the estimated canal-seepage recharge, which is strongly influenced by the inherent uncertainty in canal diversions and delivery data.

Effects of application of pre-1940 recharge adjustments (fig. 13) are visible in calibrated recharge for 1940–49 and 2000–2009 where recharge changes along the northwest to southeast line running from south of the North Platte River in Wyoming to intersect the Republican River in south-central Nebraska, and also in the northeast part of the study area including parts of the lower Elkhorn River Basin (figs. 20A and 20B). The abrupt changes in recharge along these lines correspond to where the pre-1940 recharge adjustments were applied (fig. 13); therefore, they are an artifact of the approach (as described in the “Spatial and Temporal Adjustments to 1940–2009 Recharge” section). An abrupt change in recharge along an arbitrary boundary is not expected under natural conditions; however, the reduction in recharge applied in those areas allows the groundwater-flow model to simulate more closely the groundwater levels and estimated stream base flows used as calibration targets. Natural regional patterns of recharge that likely transition more smoothly are a desirable, but not essential, feature of a useful regional model.

Calibrated recharge for 1940–49 (fig. 20A) was reduced in the vicinity of clay loam soils (fig. 5) in the Big Blue River Basin (fig. 1). Calibrated recharge was largest under canals and in the north-central part of the study area for 1940–49 and 2000–2009 (figs. 20A and 20B, respectively). The general, regional relative distribution of calibrated recharge was about the same for calibrated recharge as for uncalibrated recharge (figs. 8A and 8B), though the magnitude of calibrated recharge was generally less. Calibrated recharge was smaller than uncalibrated recharge in the areas where the predevelopment modification factor had been applied (fig. 13). Calibrated recharge for 2000–2009 was also larger in areas of irrigated agriculture (fig. 3) in the Big Blue River Basin, the downstream end of the Platte River Basin, the downstream end of the Loup River Basin, and near the Elkhorn River. Mean calibrated recharge for 1940–49 was 1.53 in/yr, which was 14 percent less than the initial MODFLOW-NWT recharge (1.77 in/yr) and 54 percent less than the estimated SWB recharge (3.3 in/yr). Mean calibrated recharge for 2000–2009 was 1.14 in/yr, which was 39 percent less than the initial MODFLOW-NWT recharge (1.87 in/yr) and 67 percent less than the estimated SWB recharge (3.5 in/yr).

Composite Parameter Sensitivities

The sensitivity of a model parameter is generally characterized by the effect that a change in that parameter has on the residuals. (As previously noted, residuals were calculated

as the calibration targets minus the simulated equivalents for stream base flows and groundwater levels.) The responses of the model to incremental changes in parameters were extracted from the Jacobian matrix generated by using the PEST automated calibration process (Doherty, 2005) and then multiplying each residual by weight of the corresponding observation (weight assignment is described in the “Calibration Targets” section). Sensitivities were summed for the same parameter groups used for calibration, including horizontal hydraulic conductivity estimated at pilot points, maximum ET rate multipliers, streambed hydraulic conductivity, pre-1940 recharge from precipitation estimated at pilot points, and spatial and temporal adjustments to 1940–2009 recharge. Larger composite sensitivities indicate that changes in those parameter groups had a larger effect on the magnitude of residuals. The pre-1940 model and 1940–2009 model were run in succession, and the pre-1940 final simulated groundwater levels were used as starting water levels for the 1940–2009 model. Therefore, and also because some parameters were used for both models (table 4), changes in some parameters caused changes in the outputs of both models.

Calibration targets for which sensitivities were analyzed included the following four groups: groundwater-level targets for the pre-1940 model (8,149 targets), estimated stream base flow for the pre-1940 model (26 targets), groundwater-level targets for the 1940–2009 model (334,918 targets), and estimated stream base flow for the 1940–2009 model (10,154 targets). There were many more targets for the 1940–2009 model than for the pre-1940 model, and there were many more targets corresponding to groundwater levels for 1990–2009 than for other groups. Composite sensitivities were also affected by the observation weights; however, composite sensitivities reflected what influenced PEST during the calibration process. Given the large number of targets corresponding to measured groundwater levels from 1940–2009 and the extensive distribution of these targets, especially for 1990–2009, it is not surprising that these targets had the largest composite sensitivity of any observation group, across all parameter groups (fig. 21); that is, as parameters were changed for each group, the largest part of composite change in residuals was contributed by the residuals from measured groundwater levels for 1940–2009. The smallest part of the composite response was contributed by estimated stream base flow for 1940, a group consisting of only 26 targets and corresponding to one point in time (April 30, 1940). This target group had the fewest targets and covered the shortest part of the simulation period.

Composite sensitivity by parameter group was most sensitive to modifications to pre-1940 recharge estimated at pilot points (fig. 21), likely because these have a strong control on pre-1940 groundwater levels and stream base flows; furthermore, pre-1940 recharge estimated at pilot points established the starting condition for the 1940–2009 model and, hence, indirectly affected all the calibration targets for 1940–2009. In addition, modifications to pre-1940 recharge estimated at pilot points have an additional effect on the 1940–2009 model through the application of the recharge adjustments

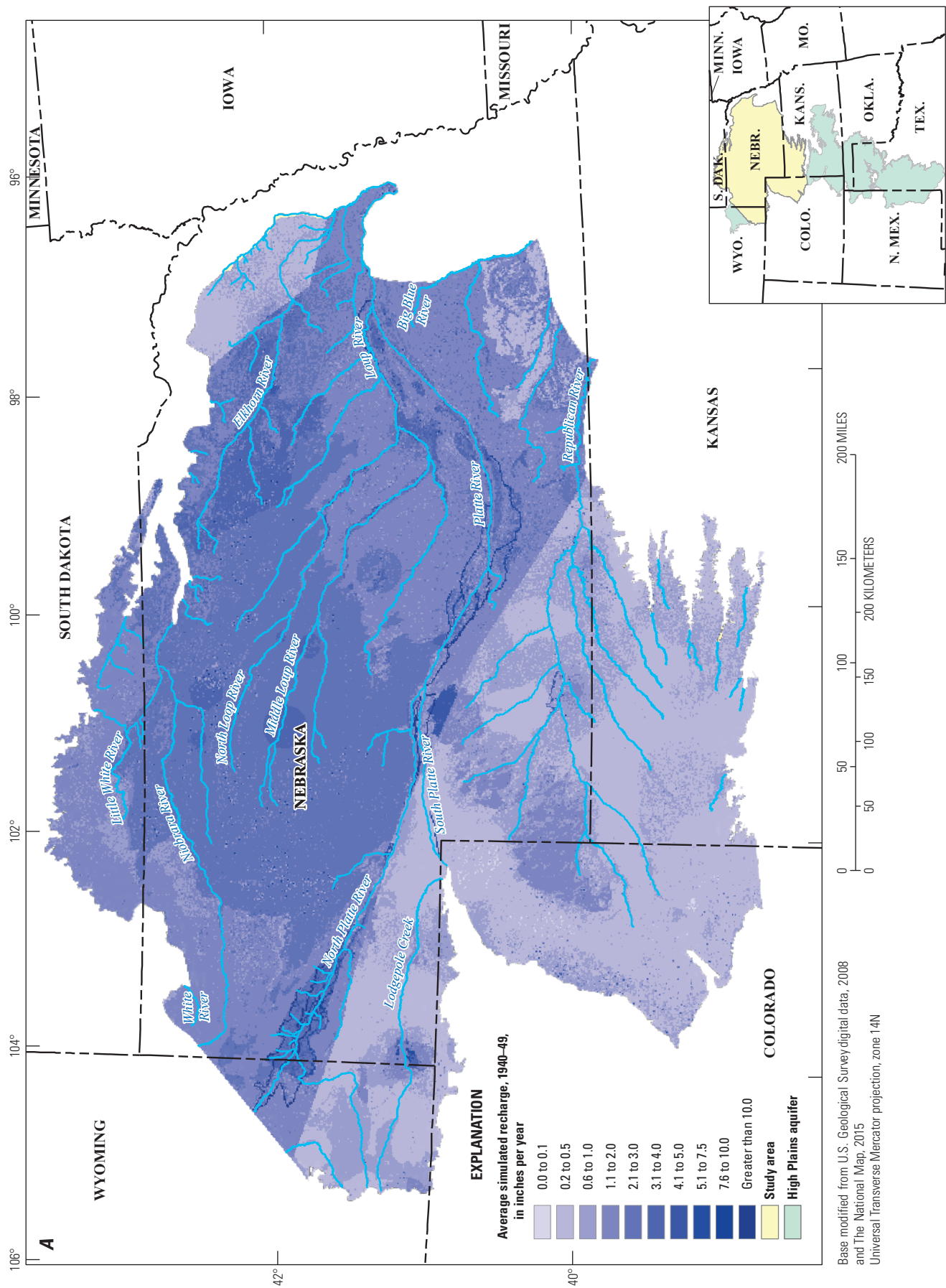


Figure 20. Calibrated recharge to the Northern High Plains aquifer in Colorado, Kansas, Nebraska, South Dakota, and Wyoming simulation results. *A*, 1940–49. *B*, 2000–2009.

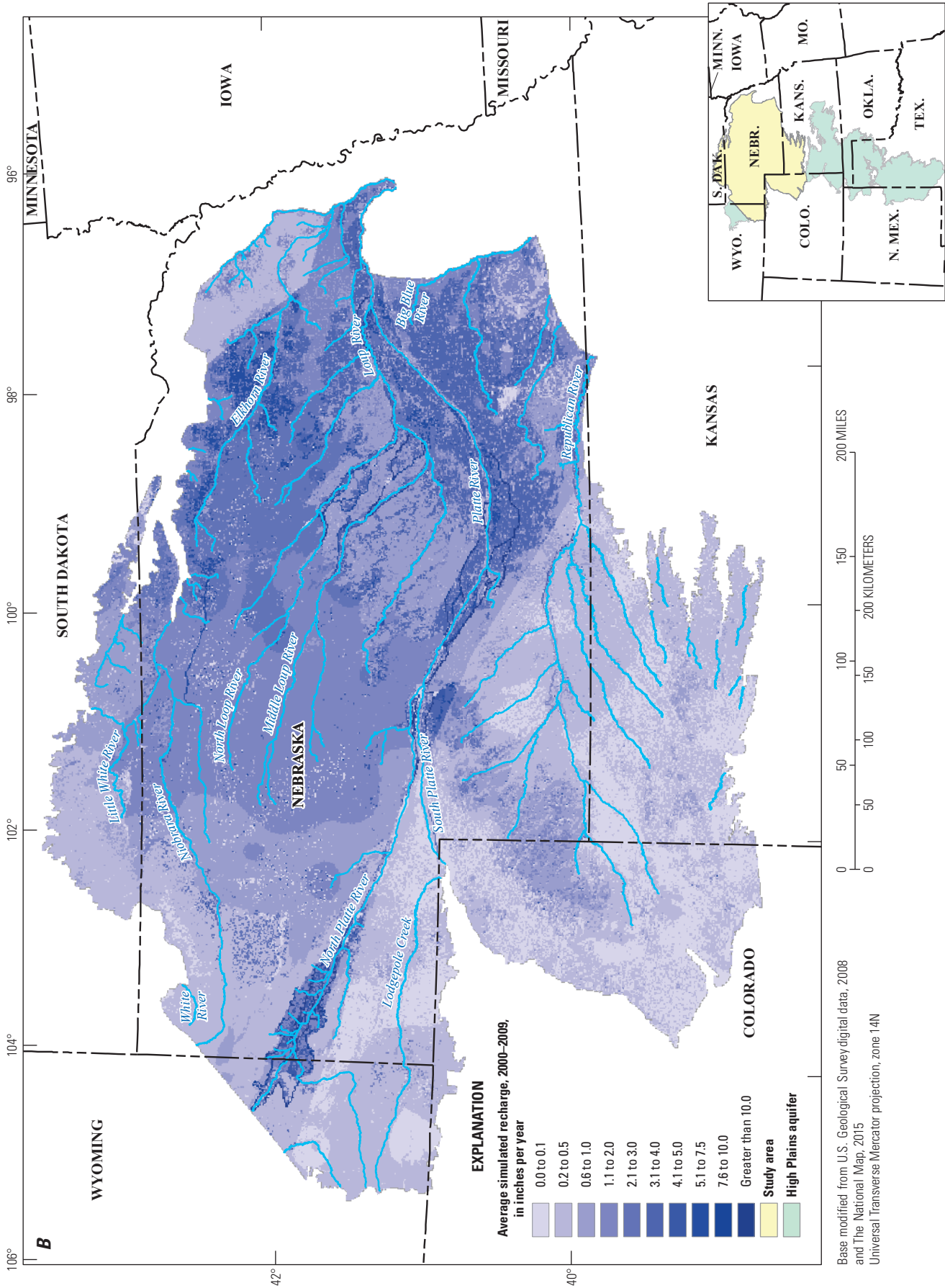


Figure 20. Calibrated recharge to the Northern High Plains aquifer in Colorado, Kansas, Nebraska, South Dakota, and Wyoming simulation results. *A*, 1940–49. *B*, 2000–2009.—Continued

that directly modifies 1940–2009 recharge in selected areas (fig. 13), as described in the “Spatial and Temporal Adjustments to 1940–2009 Recharge” section. In short, for certain areas (fig. 13), any changes to pre-1940 recharge estimated at pilot points caused similar changes to be applied to 1940–2009 recharge from precipitation.

Composite sensitivity was about the same for the remainder of the parameter groups (fig. 21). The models were moderately sensitive to modifications to horizontal hydraulic conductivity estimated at pilot points and spatial and temporal adjustments to 1940–2009 recharge and less sensitive to modifications to the maximum ET rates and streambed hydraulic conductivity.

Simulated Groundwater Budgets

Overall, the dominant inflow component of the average simulated groundwater budgets was recharge from precipitation. Outflow components were predominantly ET and stream base flow, though outflows to irrigation wells increased considerably from 1940 to 2009, and are 49 percent of the

2000–2009 outflow budget (table 7). The simulated budget changes considerably on an annual basis in relation to year to year changes in precipitation and the attendant changes in recharge from precipitation. Evaluation of decadal average budgets from the beginning, middle, and end of 1940–2009 illustrate the effects of changes in budget terms with time caused by anthropogenic activities, such as withdrawals by irrigation wells. Differences in budget terms caused by climatic variations during these decades are superimposed upon the effects caused by anthropogenic changes.

Recharge from precipitation was the largest inflow for all time periods but was larger for 1940–49 than for 1970–79 and 2000–2009. This is in contrast to the initial estimate of recharge from precipitation, which for 1940–49 was slightly smaller than that for 2000–2009, as discussed in the “Soil-Water-Balance Model” section. Inflows of recharge from canal seepage were relatively consistent from 1897 to 2009, ranging from 1,015 to 1,787 ft³/s and from 7 to 14 percent of the total inflow budget (table 7). Inflows from storage, representing groundwater-level declines, were 6 percent of the simulated inflows for 1970–79 and 25 percent of the inflows for the

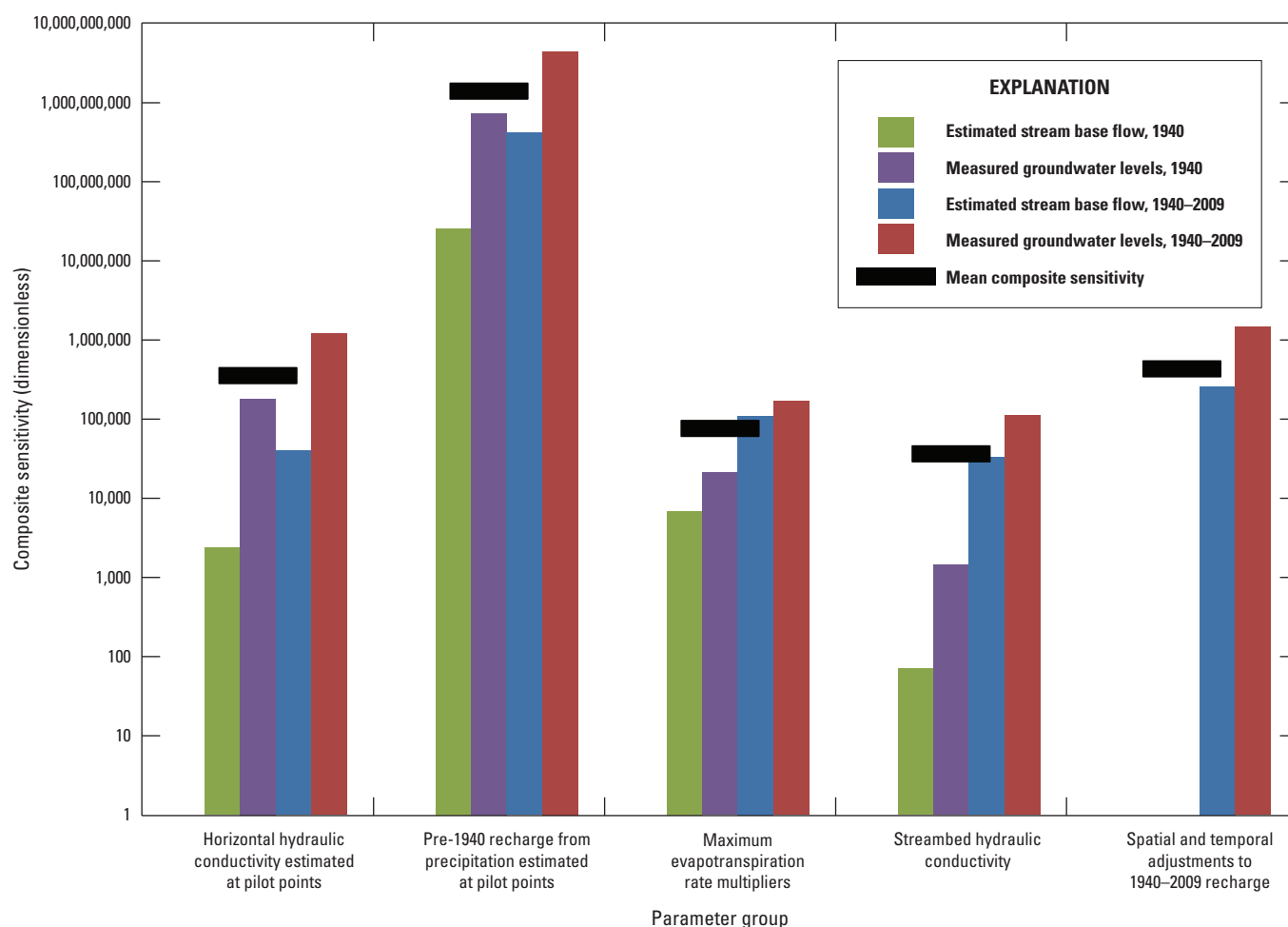


Figure 21. Composite sensitivity by parameter group used for calibration of the Northern High Plains aquifer groundwater-flow model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.

Table 7. Summary of average simulated annual groundwater budgets for pre-1897, 1897–1939, 1940–49, 1970–79, and 2000–2009 simulated by the Northern High Plains aquifer groundwater-flow model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.

[Parenthetical entries are MODFLOW package codes from Niswonger and others, 2011; --, zero; <, less than]

Water budget component (MODFLOW package in parentheses)	Pre-1897 steady state			1897–1939			1940–49			1970–79			2000–2009		
	Quantity average (cubic foot per second)	Relative quantity (percent- age of total budget)		Quantity average (cubic foot per second)	Relative quantity (percent- age of total budget)		Quantity average (cubic foot per second)	Relative quantity (percent- age of total budget)		Quantity average (cubic foot per second)	Relative quantity (percent- age of total budget)		Quantity average (cubic foot per second)	Relative quantity (percent- age of total budget)	
Inflows															
Recharge (RCH)	--	--		7,031	99		20,960	100		15,272	93		15,638	74	
–from canal seepage	--	--		1,015	14		1,489	7		1,787	11		1,641	8	
–from precipitation	6,014	99		6,014	85		19,472	93		13,485	82		13,997	66	
Groundwater inflows simulated as specified water levels (BAS)	57	<1		56	<1		47	<1		55	<1		54	<1	
Head dependent bounds repre- senting reservoirs (GHB)	--	--		--	--		17	<1		--	--		--	--	
Change in storage: groundwa- ter-level decreases (UPW)	--	--		--	--		--	--		1,060	6		5,366	25	
Outflows															
Change in storage: groundwa- ter-level increases (UPW)	--	--		225	3		9,371	45		--	--		--	--	
Evapotranspiration (EVT)	3,206	53		3,572	50		4,646	22		4,669	28		4,873	23	
Irrigation wells (WEL)	--	--		--	--		1,416	7		6,276	38		10,288	49	
Drains (DRN)	177	3		177	2		254	1		257	2		301	1	
Head dependent bounds repre- senting reservoirs (GHB)	--	--		--	--		--	--		6	<1		71	<1	
Municipal wells (MNW2)	--	--		--	--		65	<1		101	<1		95	<1	
Stream base flow (SFR)	2,687	44		3,108	44		5,273	25		5,080	31		5,437	26	
Total inflows	6,070	100		7,085	100		21,025	100		16,387	100		21,058	100	
Total outflows	6,070	100		7,082	100		21,025	100		16,389	100		21,065	100	

relatively dry period of 2000–2009. The inflows from storage for the later decades are likely a response to relatively less recharge from precipitation and increased outflows to irrigation wells.

Simulated outflows to stream base flow were about the same for 1940–49, 1970–79, and 2000–2009, varying within the range from 5,080 to 5,437 ft³/s and from 25 to 31 percent of total outflows (table 7). Simulated outflows to stream base flow were constrained by the calibration targets. Simulated outflows to ET were similarly consistent for 1940–49, 1970–79, and 2000–2009, varying within the range from 4,646 to 4,873 ft³/s and from 22 to 28 percent of total outflows. Simulated outflows to ET were only indirectly constrained by calibration targets but also were only about one-third of the maximum ET from groundwater estimated by Stanton and others (2011); furthermore, simulated ET composed a similar percentage of the outflows budget as in previously simulated parts of the study area (Luckey and Cannia, 2006; Carney, 2008; Peterson and others, 2008; Peterson, 2009; Stanton and others, 2011).

Simulated outflow to irrigation wells increased from 1940 to 2009. Outflow to irrigation wells increased from only 7 percent of total outflows for 1940–49 to about 49 percent of total outflows for 2000–2009, with corresponding average annual quantities withdrawn duplicating the seven-fold increase (table 7). The 2000–2009 irrigation-well withdrawal rate was about equal to the next two largest outflow terms combined, yet it was about 13 percent smaller than that estimated with the previous version of the SWB model for this area (Stanton and others, 2011) and 24 percent smaller than the groundwater withdrawals for irrigation estimated by the USGS Water Use Program for 2005 (Kenny and others, 2009). Increases in outflows to irrigation wells from 1940–49 to 1970–79 and 2000–2009 seem to have affected inflows from storage, corresponding to groundwater-level declines that composed 25 percent of the 2000–2009 total inflows budget.

The magnitude of the simulated annual groundwater budgets from the calibrated model changed considerably from the pre-1940 model (maximum total groundwater budget of 7,085 ft³/s) to the 1940–2009 model (maximum total groundwater budget of 21,065 ft³/s; table 7). Also, the simulated water budget for 1897–1939 featured smaller inflows from recharge and outflows to stream base flow and ET than did budgets for subsequent periods (table 7); however, fewer data existed to calibrate that simulation period and, thus, the simulated water budget also has a larger degree of uncertainty than those for subsequent periods. Additional data do not exist for earlier simulation periods that otherwise might allow the uncertainty to be decreased.

As a postaudit of model efficacy, the mass balance of the simulated budget components should be reviewed (Reilly and Harbaugh, 2004), and model inflows and outflows should be about equal. For this study, the mass balance of inflows and outflows was appropriately maintained because inflows balanced outflows (to within 10 ft³/s, less than 0.05 percent of the budget) for all simulated periods. The largest difference

between inflows and outflows was 7 ft³/s for 2000–2009, but this difference was less than 0.05 percent of the total budget.

Assumptions and Limitations

The model documented in this report is a regional groundwater-flow model meant to capture regional patterns of groundwater flow. It is assumed that the spatial, temporal, and hydrological-process resolution selected for this study are adequate to characterize the regional groundwater-flow patterns and variations that happened throughout the simulation period. Local hydrology may not be represented or may be lumped with regional features. The 1940–2009 model was constructed with seasonal stress periods and calibrated to seasonal calibration targets, so the model documented in this report would not be an appropriate tool for investigation of aquifer responses within seasons. Use of a regional model, such as the one documented in this report, to answer a local-scale question is most often inappropriate.

For one example, small tributary streams with less than 10 ft³/s mean estimated base flow generally were not represented in the model, but larger streams and rivers were represented. The larger streams represented in the model were expected to have generally correct simulated stream base flow for multiyear simulation periods and, hence, represented the correct groundwater discharge to streams for the regions also containing the unrepresented tributaries; therefore, the groundwater discharge to the unrepresented tributaries was represented by or lumped together with groundwater discharge to the larger regional streams represented in the model. Users of the model, however, must be aware that any analysis using the model might be adversely affected by the lack of local detail if the analysis is focused on a small area, especially an area near unrepresented streams. Alternatively, these smaller tributary streams may not be receiving regional groundwater discharge and, therefore, did not belong in the regional groundwater-flow model.

As additional examples, the model documented in this report could be an appropriate tool for analyzing the effect on stream base flow from management scenarios in which all irrigation wells in an area of several counties increased or reduced groundwater withdrawals by some amount for a number of growing seasons. But this model might be a poor tool to investigate what happens to stream base flow near some small ungaged tributary streams in the Nebraska Sand Hills if a well is added to supply irrigation to 40 acres of alfalfa at some future date. The simulated features (stream, well, and alfalfa field), as well as the scenario to be analyzed, are small and either unrepresented or only approximately represented in the regional groundwater-flow model; hence, there would be a large degree of uncertainty in any effects of the changes forecasted by the simulation.

Additional important assumptions and limitations that affected construction of the regional model and, therefore, the results of the simulations, are included in the following list:

1. Few pre-1940 data exist to calibrate the pre-1940 model. Better accuracy of results was assumed to be achievable by using calibration data from later periods for the pre-1940 model than to have only limited calibration data. Many streamgages did not exist before 1940; therefore, additional stream base flow targets were generated using data from other periods to produce the 26 stream base flow targets. In addition, only a few groundwater level measurements were made in the study area before 1940, and those that were made were not evenly distributed across the study area, but were most often near streams. Therefore, for the purpose of calibration, some groundwater levels used for pre-1940 model calibration (Houston and others, 2013) were measured after 1970 but were thought to be in undeveloped areas and, hence, representative of pre-1940 conditions.
2. Measurement errors or erroneous data in 1940–2009 groundwater level data were assumed to affect only a small fraction of the total population of targets and, therefore, to have a negligible effect on model calibration. Hundreds of suspect groundwater-level measurements were identified and removed from the 1940–2009 groundwater level targets, but because of the large number of measurements (334,918), it is likely that additional erroneous data are still present among the targets. In addition, in some areas, groundwater-level data consisted of individual isolated measurements; for example, a groundwater level measured at a well one time during 1940–2009 with no other nearby measurements to either confirm or refute the single measurement. It was assumed that these single measurements were valid data, and they were used as targets. It was also assumed that, although errors likely exist in the data used as groundwater level targets, the erroneous data are relatively few and that remaining errors have a negligible effect on calibration results.
3. In areas with available surface water irrigation (fig. 3), groundwater withdrawals for irrigation were assumed to be negligible. In actuality, in some years of the historical record (especially from 2000 to 2009) and in limited locations (such as near the Republican River, in the North Platte River valley in western Nebraska, and in the Central Nebraska Public Power and Irrigation District (between Harlan County Lake and the Platte River, fig. 1), surface water shortages probably caused additional groundwater withdrawals for supplemental supply. At such times and locations, the simulation may have underrepresented groundwater withdrawals for irrigation; however, it was assumed that supplemental irrigation withdrawals were negligible in the overall water balance of the Northern High Plains aquifer; moreover, omitting simulation of the supplemental withdrawals was assumed to be inconsequential for the simulation of the regional aquifer system.
4. Canal leakage that eventually reaches the water table was simulated as constant recharge throughout seasonal stress periods. This assumption may not be true for short periods (scales of weeks) because the actual movement of recharge from the root zone to the water table is affected by the thickness and other physical properties of the unsaturated zone, which can cause substantial delays in the arrival of recharge at the water table; however, groundwater-level rises near canal systems observed by the authors (Central Nebraska Public Power and Irrigation District, unpub. data, 2005) suggest that canal seepage recharge can affect groundwater levels within several days in certain areas. Even if there is a small delay from the onset of canal seepage until recharge reaches the water table, it was assumed that simulating constant canal seepage recharge during each irrigation season would result in about the same seasonal total recharge to the water table by the end of the stress period (season); therefore, the assumption is appropriate for the stress-period lengths used in this model and for simulating regional groundwater flow.
5. Monthly average stream base flows were estimated for April and October of each simulated year to match seasonal stress periods of the 1940–2009 model, as explained in the “Estimated Stream Base Flows” section. Estimates using base flow separation methods with streamgage data were assumed to (a) represent the seasonal average amount of groundwater discharge in the stream and (b) be comparable to simulated stream base flows at the end of seasonal stress periods. Base flow separation is largely a signal-processing technique, but the base flow separation results are commonly used as though indicative of base flow (Santhi and others, 2007). Upstream regulation of streamflow, such as reservoir releases or diversions that are between streamgages, can obscure the actual stream base flow. Changes in channel bedforms, debris, beaver activity, ice effects, or freezing conditions can produce inaccurate or estimated streamflow records; and ET could directly or indirectly remove base flow from the stream before the base flow is measured by a streamgage. The months of April and October, however, are usually free of ice or substantial ET and commonly include little or no irrigation canal diversions; however, any streamflow records affected by reservoir or power canal operations introduce uncertainty to base flow estimates and to the assumption that the targets actually represented groundwater discharge to the stream. In selected instances, stream base flows were reconstructed for some streamgages and times using power canal or reservoir operational data, but this method introduced much uncertainty into those stream base flow targets.
6. It was assumed that groundwater withdrawals for irrigation can be accurately estimated using a SWB model. The SWB model was originally designed to estimate

recharge (Westenbroek and others, 2010), and code to estimate irrigation requirements was added later; furthermore, the SWB approach to irrigation-requirement estimation is a simple one, using crop coefficients for early, peak, and late stages of growth (Allen and others, 1998). Few groundwater withdrawal measurements exist to calibrate estimated withdrawals. Nonetheless, review of estimated groundwater withdrawals for irrigation showed the rates to be comparable to published crop water-use data (Klocke and others, 1990; Kranz and others, 2008; Yonts, 2002) minus effective growing season precipitation (National Climatic Data Center, 2010).

7. Land-cover types (Houston and others, 2013) subsequently classified as irrigated and nonirrigated for 1949–2008 were assumed to reasonably estimate the actual distribution of land cover at the scale of the study area. Land cover is a determining factor in estimation of irrigation requirements (Westenbroek and others, 2010); thus, errors in land-cover classification produced errors in estimated irrigation requirement and potentially resulted in faulty estimates of recharge from precipitation. Subsequent classification into irrigated and nonirrigated agricultural land-cover classes was based on the presence of nearby irrigation wells (Nebraska Department of Natural Resources, 2008) and county-level crop statistics (U.S. Department of Agriculture, 1952, 1956, 1961, 1967, 1972, 1977, 1981, 1985, 1989, 1994, 1999, 2004, 2009, 2014), but no maps showing these classifications have previously been published for this study area for 1940–2009. Thus, the accuracy of the composite land-cover and irrigation classification could not be verified; therefore, it is assumed that the classification is reasonably accurate for use at the regional scale of the model.
8. Groundwater discharge to ET was assumed to happen primarily in areas of shallow groundwater and at a rate far less than potential ET. Simulated ET from this study compared favorably to that from other groundwater-flow models of parts of the area (Luckey and Cannia, 2006; Carney, 2008; Peterson and others, 2008; Peterson, 2009) and also was only about one-third of estimated maximum ET of shallow groundwater reported by Stanton and others (2011); however, field measurements of groundwater discharge to ET did not exist to allow direct calibration of ET rates or discharge. Groundwater discharge to ET is an appreciable, yet poorly constrained, component of the simulated groundwater budget.
9. The SWB model used to estimate recharge from precipitation and groundwater withdrawals for irrigation does not estimate rejected recharge in areas where the water table is near, at, or above land surface—and neither does the MODFLOW-NWT model constructed for this study. This limitation may have been fully mitigated by adjustment of MODFLOW-NWT inputs through calibration

until the approximately correct magnitude of recharge was achieved across the region of the Northern High Plains aquifer, but this may not be true at a local scale. The SWB and MODFLOW-NWT models documented in this report were limited in that each simulated only partially the set of relevant processes, rather than solving all related processes simultaneously in a coupled fashion.

10. Groundwater-flow model inputs are nonunique, and multiple combinations of parameters exist that can provide similar fits to calibration targets (Anderson and others, 2015; Konikow and Bredehoeft, 1992). For example, if calibration targets were limited to groundwater levels, a unique, best-fit estimate of recharge and aquifer hydraulic conductivity could not be estimated because different combinations of recharge and aquifer hydraulic conductivity could produce the same groundwater level. Nonuniqueness in this study was more limited through the inclusion of estimated stream base flows as calibration targets. Simulated stream base flows can only estimate these targets if the magnitude of regional recharge and groundwater discharge causing the stream base flow is nearly correct. This does not ensure that local errors and nonuniqueness are not a part of the calibrated model result but only that, at the scale of the groundwater basin supplying stream base flow, an approximately correct amount of groundwater discharge (originating as recharge) was simulated as reaching the stream.

Potential Topics for Additional Study

Application of calibration changes from pre-1940 recharge from precipitation to 1940–2009 recharge from precipitation in only part of the model area (fig. 13) created an unnaturally abrupt change in the simulated regional patterns of 1940–2009 recharge (figs. 20A and 20B). Use of this method for correcting regional bias in the recharge from precipitation estimated with the SWB model implicitly acknowledges that additional study is needed on adjustment of SWB inputs for individual climate stations or for certain crop types (for instance, to adjust crop characteristics) to produce recharge estimates close to the calibrated recharge. In addition, recharge estimated with SWB was much larger than the final calibrated MODFLOW-NWT recharge, which suggests another profitable area of study, to refine the SWB inputs, so that SWB produces a more unbiased estimate of recharge. If SWB results were closer to the final calibrated recharge, it would require fewer adjustments to produce suitable MODFLOW-NWT outputs.

Several test holes in South Dakota showed the aquifer base altitude to be much deeper than what could be supported using surrounding data. There are few test holes in that area of the Northern High Plains aquifer, and additional test hole drilling or other geophysical data collection is needed to improve

the understanding of the configuration of the aquifer base in that area. Such studies might lead to improved accuracy of simulated groundwater levels in that area.

Parts of the Northern High Plains aquifer in South Dakota and Wyoming seemed to have consistently larger groundwater level residuals than other areas, though they were not consistently biased either high or low. Large residuals in these areas may have been caused by the lack of representation of the land surface in the model. Preliminary testing with the UZF Package for MODFLOW-2005 indicated that including a representation of land surface in the models would have improved simulated water levels in South Dakota and Wyoming. Further study using the UZF package is needed to resolve the complications in the MODFLOW-NWT model associated with the preliminary testing of UZF simulations.

This study used a sequentially coupled approach, where a SWB model was used to estimate recharge from precipitation and groundwater withdrawals for irrigation, and the SWB outputs were used as inputs for the groundwater-flow model. The SWB model, however, does not have any method for reducing groundwater withdrawals for conditions under which crops would be subirrigated (that is, part of the crop water demand is satisfied by uptake of shallow groundwater) because it does not track depth to groundwater; consequently, the SWB model also does not have a method for rejecting recharge in areas where the water table is near land surface. Improvements to SWB code to either allow water-table tracking or to accept an equivalent input signal are invited.

The groundwater model simulated only partially certain hydrologic processes; for example, regional groundwater discharge to streams was simulated, assumed to be equivalent to base flow, and was calibrated to stream base flows estimated from streamgauge data that in some cases were affected by canal diversions or reservoir operations. As an example of a regionally important process only indirectly simulated, recharge from canal seepage was estimated and added to the recharge from precipitation before using the sum as a model input. More work using detailed process-representation approaches should be considered; for example, modeling studies using the One-Water Hydrologic Flow Model (Hanson and others, 2014) can (1) allow for a comprehensive simulation of landscape (surface) hydrology fully coupled with groundwater hydrology, (2) provide for simulation of base flow and runoff, (3) allow for more realistic simulation of uptake of shallow groundwater by crops and the apportionment of irrigation water through surface water delivery and supplemental groundwater withdrawal, (4) allow groundwater and surface water irrigation to be driven by supply and demand concepts, (5) allow for an improved calibration to standard streamflow records from streamgages as opposed to processed stream base flow data, and (6) provide an opportunity to calibrate the simulation using metered groundwater withdrawal data. Together, these improvements in data and methodology could provide additional accuracy and insight into the estimation of groundwater withdrawals in surface water irrigated areas in

times of shortage and the interaction of conjunctive water supplies to irrigated fields.

Summary

The High Plains aquifer is a nationally important water resource underlying about 175,000 square miles in parts of eight states: Colorado, Kansas, Oklahoma, Nebraska, New Mexico, South Dakota, Texas, and Wyoming. Recent droughts across much of the Northern High Plains have combined with interstate conflicts and legislative mandates to elevate concerns regarding future availability of groundwater and the need for additional information to support science-based water-resources management; moreover, tools capable of providing forecasts of groundwater availability for the Northern High Plains aquifer have become dated. To address these needs, the U.S. Geological Survey's High Plains Groundwater Availability Study, one of a series of regional studies to evaluate the water availability and sustainability of major aquifers across the Nation, was begun in 2009. This report documents the conceptualization, construction, and calibration of a groundwater-flow model suitable for use as a tool for water-resource managers and other stakeholders to assess the status and availability of groundwater resources in the Northern High Plains aquifer.

Development of groundwater was limited before 1940, covering less than 0.5 percent of the study area. By 2008, the predominantly agricultural land above the Northern High Plains aquifer included about 7,294,000 acres (11.7 percent) of groundwater-irrigated area. The relative magnitudes of groundwater inflows and outflows to and from the Northern High Plains aquifer were not known exactly at the outset of study, but based on previous studies of the area, the general expectations were that (1) recharge from precipitation is the largest groundwater inflow, and (2) groundwater discharge to streams, evapotranspiration, and groundwater withdrawals for irrigation are the largest outflows. Of the smaller sources of inflow, only recharge from canal seepage constitutes a substantial part of inflows. Changes in groundwater storage, or groundwater flows to and from storage, are represented by groundwater levels either rising or declining in the Northern High Plains aquifer. Groundwater levels in the Northern High Plains aquifer have been generally stable since groundwater withdrawals for irrigation began.

The Northern High Plains aquifer groundwater-flow model was constructed using a Newton formulation of the U.S. Geological Survey modular three-dimensional finite-difference groundwater-flow model (MODFLOW-NWT). The groundwater-flow model covers an area of about 59.5 million acres of the states of Colorado, Kansas, Nebraska, South Dakota, and Wyoming. The model uses an orthogonal grid of 565 rows and 795 columns, and each grid cell measures 3,281 feet per side, with one variably-thick vertical layer simulated as unconfined. Groundwater flow was simulated

for two distinct periods: the period before substantial groundwater withdrawals, or before about 1940, and the period of increasing groundwater withdrawals from May 1940 through April 2009. A soil-water-balance model was used to estimate recharge from precipitation and groundwater withdrawals for irrigation. The soil-water-balance model uses spatially distributed soil and landscape properties with daily weather data and estimated historical land-cover maps to calculate spatial and temporal variations in potential recharge. Mean annual recharge estimated for 1940–49, early in the history of groundwater development, and 2000–2009, late in the history of groundwater development, was 3.3 and 3.5 inches per year, respectively.

Primary model calibration was completed using statistical techniques through parameter estimation using PEST with Tikhonov regularization. Calibration targets for the 1940–2009 model included 334,918 groundwater levels measured in wells and 10,154 estimated monthly mean stream base flows at streamgages. A total of 1,312 parameters were adjusted during calibration to improve the match between calibration targets and simulated equivalents. The largest parameter adjustments were to reduce 1940–2009 estimated recharge from precipitation that, for the calibrated MODFLOW-NWT model, was 54 percent less for 1940–49 and 67 percent less for 2000–2009 than the recharge from precipitation estimated with the soil-water-balance model.

Comparison of calibration targets to simulated equivalents indicated that, at the regional scale, the model correctly reproduced groundwater levels and stream base flows. Simulated groundwater levels for pre-1940 and by decade for 1940–2009 differed on average by less than 6 feet from measured groundwater levels used as calibration targets. The groundwater-flow models simulated about the same amount of stream base flow as was estimated overall for pre-1940 and 1940–2009. Estimated stream base flows tended to be more variable during short periods, but simulated stream base flows were about correct as to magnitude and multiyear trend. For three of six major river basins, simulated base flow was less than 2 percent different from estimated base flow; and for a fourth, simulated base flow was about 12 percent different from estimated base flow. Small percent differences indicate that the models simulated about the right amount of regional groundwater discharge to streams (stream base flow), consistent with the purpose of a regional groundwater model. Graphical comparison of simulated and estimated stream base flows by streamgages suggested that the simulated stream base flows about matched the magnitude and multiyear trend of estimated stream base flows for most streams, with occasional exceptions when single-season estimates were much larger than estimates for neighboring seasons. These calibration results indicate that, on the whole, the model adequately simulated regional groundwater levels and stream base flow, consistent with the purpose of a regional groundwater model, and that the model can be used to examine the likely response of the regional aquifer system to potential future stresses.

Mean calibrated recharge for 1940–49 and 2000–2009 was smaller than that estimated with the soil-water-balance model. This indicated that though the general spatial patterns of recharge estimated with the soil-water-balance model were about correct at the regional scale of the Northern High Plains aquifer, the soil-water-balance model had overestimated recharge in some areas, and adjustments were needed to decrease it to improve the match of the groundwater model to calibration targets. In other areas, particularly for those with irrigated agriculture, calibrated recharge was increased above the initial estimate.

The largest components of the simulated groundwater budgets were recharge from precipitation, recharge from canal seepage, outflows to evapotranspiration, and outflows to stream base flow. Simulated outflows to irrigation wells increased from 7 percent of total outflows in 1940–49 to 38 percent of 1970–79 outflows and 49 percent of 2000–2009 total outflows. Outflows to irrigation wells for 2000–2009 were about equal to the sum of the next two largest outflows (stream base flow and evapotranspiration). Simulated groundwater outflows to evapotranspiration were only indirectly constrained by calibration targets but composed a similar percentage of the outflows budget as simulated by other groundwater-flow models covering parts of the study area, and were only about one-third of the maximum evapotranspiration from groundwater estimated in a previous study.

The model documented in this report is a regional groundwater-flow model meant to capture regional patterns of groundwater flow. Local hydrology may not be represented or may be lumped with regional features. Use of a regional model, such as the one documented in this report, to answer a local-scale question is most often inappropriate. Similarly, calibration targets and model temporal configuration were only refined to a seasonal level, so the model documented in this report would not be an appropriate tool for investigation of aquifer responses within seasons. In addition, the model documented in this report could be further enhanced. New data could provide additional refinement to model inputs or provide new calibration targets. More detailed simulation approaches could be applied that would further enhance the accuracy of the results through simulation of rejected recharge, groundwater discharge to land surface, and fully coupled simulation approaches to integrating landscape and groundwater hydrology, rather than the sequentially coupled approach used in this study.

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Appendix 1. Supplemental Information on Estimated and Simulated Stream Base Flow for 1940–2009

A summary of estimated and simulated 1940–2009 stream base flows for seven streamgages were presented in the “Calibration Targets, Estimated Stream Base Flows” section. Data from 91 streamgages in total were used for calibration of the 1940–2009 model, and a summary of the remaining streamgages is presented herein for readers interested in additional detail (table 1–1). The method for calculation of percent differences for table 1–1 was the same as described in the “Stream Base Flow Residuals” section and as used for streamgages reported in table 3. The summary statistics reported in table 1–1 include the average, minimum, and maximum estimated stream base flow; the number of estimated values within 1940–2009; and the average, minimum, and maximum simulated stream base flow. Comparative statistics, such as the mean and median residual, and the mean percent difference between estimated and simulated base flows are

included in table 1–1. Mean residuals indicate the bias of the residuals; an unbiased mean residual would be zero. Median residuals indicate the residual that happens most often. The mean percent difference regularizes size differences among streamgages and indicates, on average, how different simulated stream base flow was from estimated.

Graphs of simulated and estimated stream base flow for seven streamgages in the Northern High Plains aquifer were presented in the “Stream Base Flow Residuals” section (fig. 17). Graphs for the remaining 84 stations are presented herein for readers interested in additional detail (figs. 1–1 to 1–84). Lowess trend lines (Cleveland, 1979; Cleveland and Devlin, 1988) were used as a graphical technique to evaluate whether trends of simulated and estimated base flows were similar.

Table 1–1. Summary of estimated base flows and simulated base flows, at 84 streamgages used as calibration targets for the Northern High Plains aquifer 1940–2009 model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.

[Nebr., Nebraska; S. Dak., South Dakota; NC, not calculated, because simulated base flows were all zero, or because base flow targets were zero-weighted; Kans., Kansas; *, indicates that due to persistent low-flow conditions at sites, base flow targets considered suspect, and were zero-weighted and not used]

Streamgage	Streamgage number	Period of record	Estimated base flow (cubic foot per second)			Number of estimated base flow targets within 1940–2009	Simulated base flow (cubic foot per second)			Comparative statistics		
			Average	Minimum	Maximum		Average	Minimum	Maximum	Mean residual	Median residual	Mean percent difference
White River at Crawford, Nebr.	06444000	1931–2009	19	0	8	130	19	11	24	0	0	0
Little White River near VetaI, S. Dak.	06449100	1959–2009	50	17	144	100	42	27	53	9	-2	5
Little White River near Rosebud, S. Dak.	06449500	1943–2009	111	48	235	132	92	65	115	20	3	12
Niobrara River at Agate, Nebr.	06454100	1952–2009	12	5	26	79	7	4	11	5	5	55
Niobrara River above Box Butte Reservoir, Nebr.	06454500	1946–2009	25	3	60	115	27	8	51	-2	0	-13
Niobrara River near Gordon, Nebr.	06457500	1928–1994	100	29	173	97	162	115	215	-62	-60	-49
Snake River near Burge, Nebr.	06459500	1947–2009	178	8	320	85	235	193	289	-57	-37	-48
Minnechadua Creek at Valentine, Nebr.	06461000	1947–1994	28	8	68	94	29	21	44	-1	-4	-15
Niobrara River near Norden, Nebr.	06462000	1952–1986	865	541	1,248	61	842	723	1,038	23	0	1
Plum Creek at Meadville, Nebr.	06462500	1947–1994	108	75	250	91	94	83	118	14	5	11
Keya Paha River near Keyapaha, S. Dak.	06464100	1981–2009	38	10	118	56	38	26	50	0	-8	-20
Horse Creek near Lyman, Nebr.	06677500	1930–2009	42	2	186	137	12	1	24	30	24	102
Sheep Creek near Morrill, Nebr.	06678000	1931–2009	73	2	129	137	28	19	39	46	44	87
North Platte River at Mitchell, Nebr.	06679500	1909–2009	506	108	2,126	130	459	167	1,912	47	91	19
North Platte River near Minatare, Nebr.	06682000	1916–2009	838	280	2,601	130	592	272	2,005	246	267	38

Table 1–1. Summary of estimated base flows and simulated base flows, at 84 streamgages used as calibration targets for the Northern High Plains aquifer 1940–2009 model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.—Continued

[Nebr., Nebraska; S. Dak., South Dakota; NC, not calculated, because simulated base flows were all zero, or because base flow targets were zero-weighted; Kans., Kansas; *, indicates that due to persistent low-flow conditions at sites, base flow targets considered suspect, and were zero-weighted and not used]

Streamgage	Streamgage number	Period of record	Estimated base flow (cubic foot per second)			Number of estimated base flow targets within 1940–2009	Simulated base flow (cubic foot per second)			Comparative statistics		
			Average	Minimum	Maximum		Average	Minimum	Maximum	Mean residual	Median residual	Mean percent difference
North Platte River at Bridgeport, Nebr.	06684500	1930–2009	1,077	158	2,735	130	703	376	2,105	374	368	42
Pumpkin Creek near Bridgeport, Nebr.	06685000	1930–2009	19	0	50	129	5	1	9	13	11	95
North Platte River at Lisco, Nebr.	06686000	1931–2009	1,088	300	2,738	130	734	417	2,112	354	359	39
Blue Creek near Lewellen, Nebr.	06687000	1931–2009	69	1	103	138	66	48	82	3	12	-9
North Platte River near Lewellen, Nebr.	06687500	1931–2009	1,248	284	2,779	129	811	503	2,166	437	450	42
Birdwood Creek near Hershey, Nebr.	06692000	1931–2009	148	119	183	126	143	104	165	4	4	3
North Platte River at North Platte, Nebr.	06693000	1909–2009	958	238	4,607	138	935	178	4,681	22	119	6
Lodgepole Creek at Bushnell, Nebr.	06762500	1931–2009	7	0	16	119	5	2	9	2	3	24
Lodgepole Creek at Ralston, Nebr.	06763500	1951–2009	5	0	28	71	1	0	4	3	1	86
South Platte River at North Platte, Nebr.	06765500	1907–2009	192	39	1,530	136	155	0	1,068	37	100	12
Platte River at Brady, Nebr.	06766000	1939–2009	1,558	33	7,076	133	1,181	162	5,624	377	406	33
Platte River at Cozad, Nebr.	06766500	1940–2009	1,595	15	7,897	133	1,229	177	5,700	366	401	28
Platte River near Overton, Nebr.	06768000	1915–2007	966	0	8,010	135	1,239	163	5,750	-273	-235	-42
Platte River near Odessa, Nebr.	06770000	1936–2009	682	57	2,827	84	1,070	148	3,250	-388	-357	-56
Platte River near Grand Island, Nebr.	06770500	1933–2009	888	0	7,355	138	1,155	0	5,737	-267	-243	-45
Platte River near Duncan, Nebr.	06774000	1895–2009	1,017	0	8,330	138	1,101	0	5,732	-84	-80	-37

Table 1–1. Summary of estimated base flows and simulated base flows, at 84 streamgages used as calibration targets for the Northern High Plains aquifer 1940–2009 model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.—Continued

[Nebr., Nebraska; S. Dak., South Dakota; NC, not calculated, because simulated base flows were all zero, or because base flow targets were zero-weighted; Kans., Kansas; *, indicates that due to persistent low-flow conditions at sites, base flow targets considered suspect, and were zero-weighted and not used]

Streamgage	Streamgage number	Period of record	Estimated base flow (cubic foot per second)			Number of estimated base flow targets within 1940–2009	Simulated base flow (cubic foot per second)			Comparative statistics		
			Average	Minimum	Maximum		Average	Minimum	Maximum	Mean residual	Median residual	Mean percent difference
Middle Loup River at Dunning, Nebr.	06775500	1945–2009	423	313	556	128	334	284	424	89	83	23
Dismal River at Dunning, Nebr.	06776500	1932–1995	320	281	400	100	286	246	362	34	33	11
Middle Loup River at Arcadia, Nebr.	06779000	1937–1994	770	427	1,147	109	733	490	982	37	49	4
South Loup River at St. Michael, Nebr.	06784000	1943–2009	183	78	413	132	163	80	291	20	20	10
Middle Loup River at St. Paul, Nebr.	06785000	1934–2009	848	218	1,726	138	960	478	1,293	-112	-96	-16
North Loup River at Taylor, Nebr.	06786000	1936–2008	483	241	761	136	447	287	571	36	34	7
Calamus River near Burwell, Nebr.	06787500	1940–2009	290	55	588	117	228	155	287	62	62	21
North Loup River near St. Paul, Nebr.	06790500	1928–2009	862	385	1,548	138	854	522	1,058	8	11	0
Cedar River near Spalding, Nebr.	06791500	1944–2009	149	80	285	122	115	81	149	34	30	24
Cedar River near Fullerton, Nebr.	06792000	1930–2009	211	112	479	138	175	89	249	36	26	17
Loup River near Genoa, Nebr.	06793000	1928–2008	2,245	1,347	3,544	129	2,130	1,568	2,758	115	90	4
Beaver Creek at Genoa, Nebr.	06794000	1940–2009	99	36	290	138	103	30	174	-4	-10	-8
Loup River at Columbus, Nebr.	06794500	1909–1978	2,273	1,433	3,519	76	2,223	1,199	2,700	49	75	1
Platte River at North Bend, Nebr.	06796000	1949–2009	3,026	923	7,026	121	3,777	2,164	9,267	-751	-697	-28
Elkhorn River near Atkinson, Nebr.	06796973	1982–2009	68	1	334	53	36	5	70	33	2	18
Elkhorn River at Ewing, Nebr.	06797500	1947–2009	150	14	1,079	124	117	45	193	34	-27	-17

Table 1–1. Summary of estimated base flows and simulated base flows, at 84 streamgages used as calibration targets for the Northern High Plains aquifer 1940–2009 model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.—Continued

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Streamgage	Streamgage number	Period of record	Estimated base flow (cubic foot per second)			Number of estimated base flow targets within 1940–2009	Simulated base flow (cubic foot per second)			Comparative statistics		
			Average	Minimum	Maximum		Average	Minimum	Maximum	Mean residual	Median residual	Mean percent difference
South Fork Elkhorn River near Ewing, Nebr.	06798000	1947–2009	54	22	212	99	55	24	93	0	-6	-11
Elkhorn River at Neligh, Nebr.	06798500	1931–2009	276	62	1,484	133	288	142	446	-13	-83	-28
Elkhorn River at Norfolk, Nebr.	06799000	1945–2008	427	112	2,053	126	372	202	533	55	-43	-5
North Fork Elkhorn River near Pierce, Nebr.	06799100	1960–2009	76	12	435	98	19	6	41	56	28	112
Elkhorn River at West Point, Nebr.	06799350	1971–2009	755	157	3,034	74	408	183	645	348	170	42
Logan Creek Dredge at Pender, Nebr.	06799450	1964–2009	128	32	516	88	11	0	31	117	94	232
Logan Creek Dredge near Uehling, Nebr.	06799500	1940–1978	87	29	280	76	11	0	36	77	63	152
Maple Creek near Nicksen, Nebr.	06800000	1951–2009	34	0	245	116	15	0	64	19	9	31
Elkhorn River at Waterloo, Nebr.	06800500	1911–2009	1,008	139	3,481	137	478	211	771	531	305	57
Arikaree River at Haigler, Nebr.	06821500	1931–2009	8	0	32	138	8	8	8	0	-2	-64
North Fork Republican River at Colorado-Nebraska state line	06823000	1931–2009	33	5	73	138	33	23	42	0	0	-12
Republican River at Benkelman, Nebr.	06824500	1946–1994	66	7	152	95	62	47	79	3	-1	-8
South Fork Republican River near Benkelman, Nebr.	06827500	1903–2009	20	0	133	138	2	0	19	19	11	239
Republican River at Stratton, Nebr.	06828500	1950–2009	64	0	228	118	51	19	97	14	2	-30
Frenchman Creek near Imperial, Nebr.	06831500	1940–2009	37	3	72	137	22	0	56	15	16	92

Table 1–1. Summary of estimated base flows and simulated base flows, at 84 streamgages used as calibration targets for the Northern High Plains aquifer 1940–2009 model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.—Continued

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Streamgage	Streamgage number	Period of record	Estimated base flow (cubic foot per second)			Number of estimated base flow targets within 1940–2009	Simulated base flow (cubic foot per second)			Comparative statistics		
			Average	Minimum	Maximum		Average	Minimum	Maximum	Mean residual	Median residual	Mean percent difference
Frenchman Creek at Palisade, Nebr.	06834000	1950–2009	37	10	164	116	34	8	82	4	7	20
Stinking Water Creek near Palisade, Nebr.	06835000	1949–2009	29	7	50	120	23	5	41	6	6	26
Frenchman Creek at Culbertson, Nebr.	06835500	1932–2009	67	0	207	138	71	13	141	-4	0	-21
Driftwood Creek near McCook, Nebr.	06836500	1946–2009	3	0	8	127	2	0	5	1	1	20
Republican River at McCook, Nebr.	06837000	1954–2009	92	0	521	110	114	21	211	-22	-24	-57
Red Willow Creek near McCook, Nebr.	06837500	1940–1994	21	1	36	83	21	15	26	0	2	-21
Red Willow Creek near Red Willow, Nebr.	06838000	1939–2009	26	5	46	137	25	12	33	0	0	-9
Medicine Creek above Harry Strunk Lake, Nebr.	06841000	1962–1994	50	32	66	89	17	9	24	33	33	111
Medicine Creek below Harry Strunk Lake, Nebr.	06842500	1949–2009	51	4	125	118	19	7	30	32	36	75
Republican River at Cambridge, Nebr.	06843500	1945–2009	140	0	696	128	136	0	282	5	-7	-28
Republican River near Orleans, Nebr.	06844500	1947–2008	165	0	733	119	132	0	314	32	26	16
South Fork Sappa Creek near Achilles, Kans.	06844900	1959–2009	0	0	6	100	0	0	0	0	0	NC
Sappa Creek near Beaver City, Nebr.	06845200	1940–2002	6	0	50	66	0	0	0	6	2	NC
Beaver Creek at Cedar Bluffs, Kans.	06846500	1946–2009	3	0	47	126	0	0	0	3	0	NC

Table 1–1. Summary of estimated base flows and simulated base flows, at 84 streamgages used as calibration targets for the Northern High Plains aquifer 1940–2009 model in Colorado, Kansas, Nebraska, South Dakota, and Wyoming.—Continued

[Nebr., Nebraska; S. Dak., South Dakota; NC, not calculated, because simulated base flows were all zero, or because base flow targets were zero-weighted; Kans., Kansas; *, indicates that due to persistent low-flow conditions at sites, base flow targets considered suspect, and were zero-weighted and not used]

Streamgage	Streamgage number	Period of record	Estimated base flow (cubic foot per second)			Number of estimated base flow targets within 1940–2009	Simulated base flow (cubic foot per second)			Comparative statistics		
			Average	Minimum	Maximum		Average	Minimum	Maximum	Mean residual	Median residual	Mean percent difference
Beaver Creek near Beaver City, Nebr.	06847000	1937–2009	5	0	61	138	0	0	0	5	1	NC
Sappa Creek near Stamford, Nebr.	06847500	1944–2008	13	0	178	125	6	0	13	7	-2	-20
Prairie Dog Creek at Norton, Kans.	06848000	1944–2002	2	0	25	116	0	0	0	2	0	NC
Prairie Dog Creek near Woodruff, Kans.	06848500	1945–2009	4	0	48	128	0	0	0	4	1	NC
Thompson Creek at Riverton, Nebr.	06851500	1948–2009	20	13	29	95	13	5	24	8	8	54
Elm Creek at Amboy, Nebr.	06852000	1919–2009	14	9	23	74	0	0	0	14	15	NC
Republican River near Guide Rock, Nebr.	06853000	1950–2009	212	1	1,875	116	172	29	370	40	-25	-28
Republican River near Hardy, Nebr.	06853500	1940–2009	308	31	1,235	128	225	53	405	83	41	12
North Fork Smoky Hill River near McAllaster, Kans.	06858500	1947–1984	1	3	254	62*	4	3	7	NC	NC	NC
Saline River near Wakeeney, Kans.	06866900	1955–2009	5	31	326	79*	11	7	13	NC	NC	NC
Big Blue River at Seward, Nebr.	06880500	1954–2009	38	41	732	110	22	6	47	16	7	34
West Fork Big Blue River near Dorchester, Nebr.	06880800	1958–2008	79	27	213	100	96	31	160	-16	-25	-24
Big Blue River near Crete, Nebr.	06881000	1945–2008	165	12	81	89	149	53	264	16	-7	-2
Little Blue River near Deweese, Nebr.	06883000	1953–2008	68	0	0	107	73	37	119	-5	-6	-9
Big Sandy Creek at Alexandria, Nebr.	06883940	1979–2009	26	0	0	60	33	25	51	-7	-9	-31

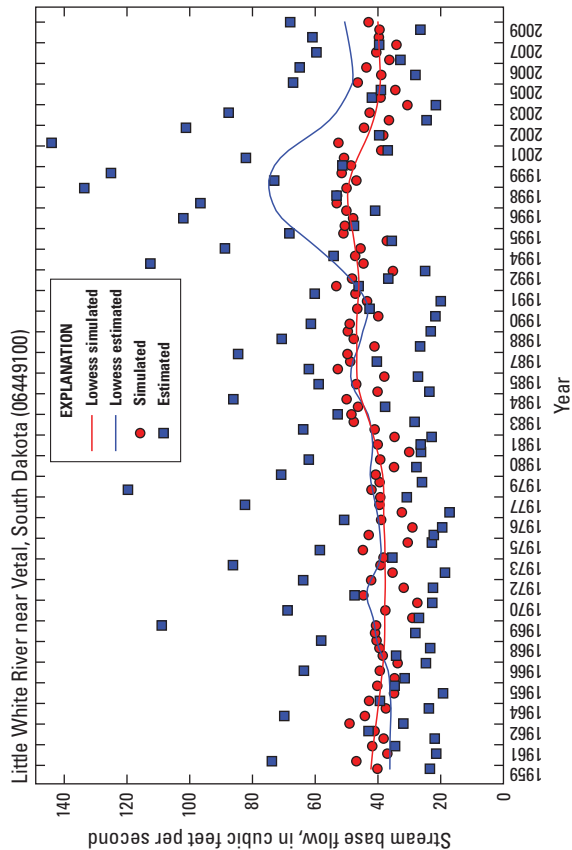


Figure 1–2. Simulated and estimated base flow, 1959–2009, for Little White River near Veta, S. Dak. (streamgage number 06449100).

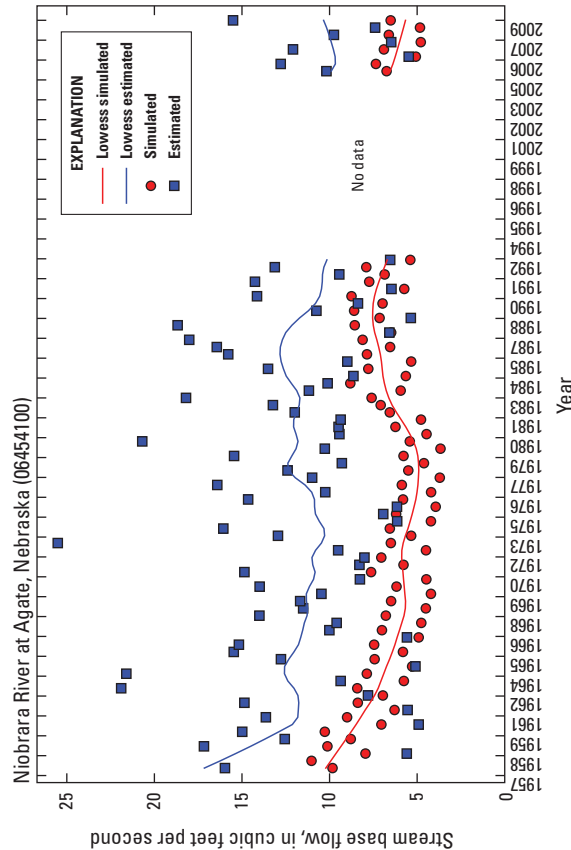


Figure 1–4. Simulated and estimated base flow, 1957–2009, for Niobrara River at Agate, Nebr. (streamgage number 06454100).

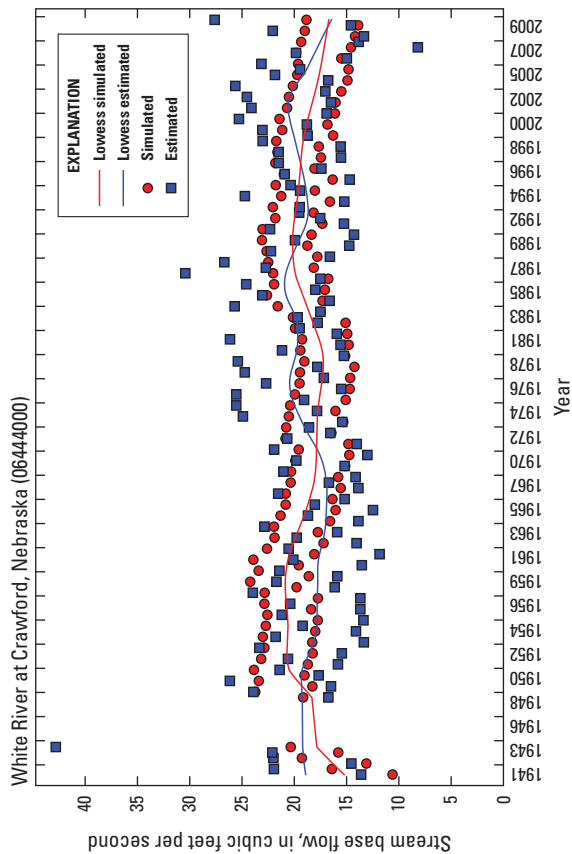


Figure 1–1. Simulated and estimated base flow, 1940–2009, for White River at Crawford, Nebr. (streamgage number 06444000).

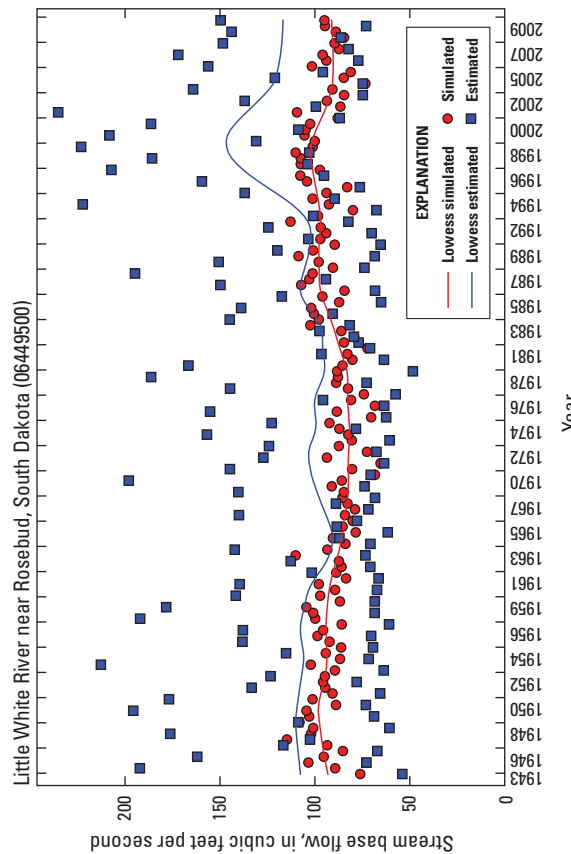


Figure 1–3. Simulated and estimated base flow, 1943–2009, for Little White River near Rosebud, S. Dak. (streamgage number 06449500).

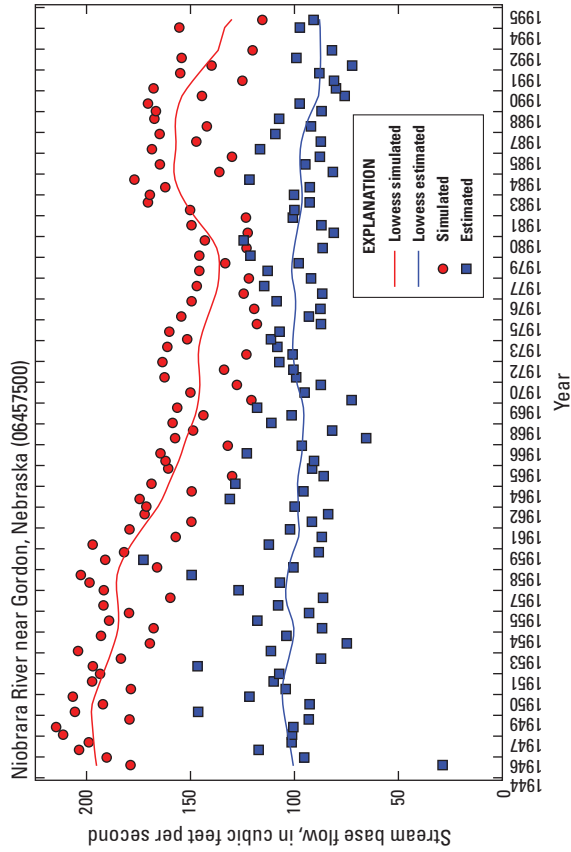


Figure 1-6. Simulated and estimated base flow, 1945-94, for Niobrara River near Gordon, Nebr. (streamgage number 06457500).

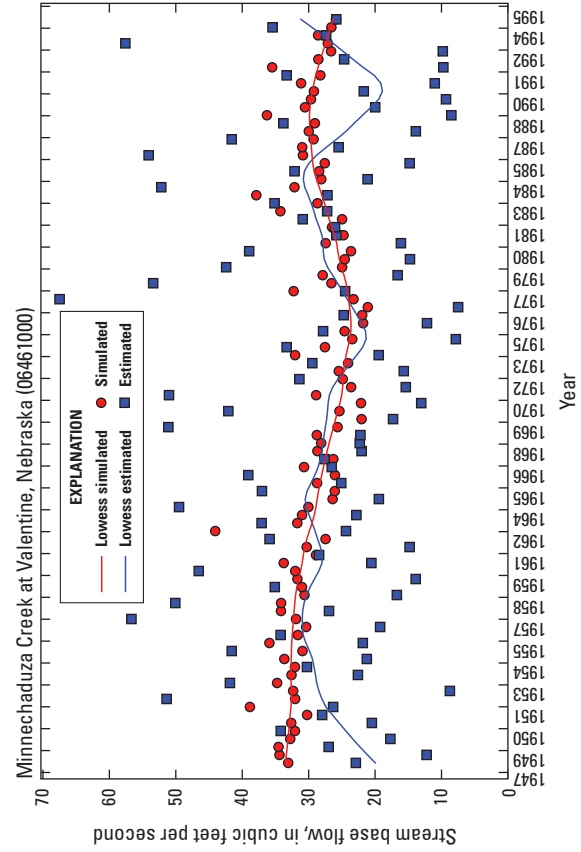


Figure 1-8. Simulated and estimated base flow, 1948-94, for Minnehadua Creek at Valentine, Nebr. (streamgage number 06461000).

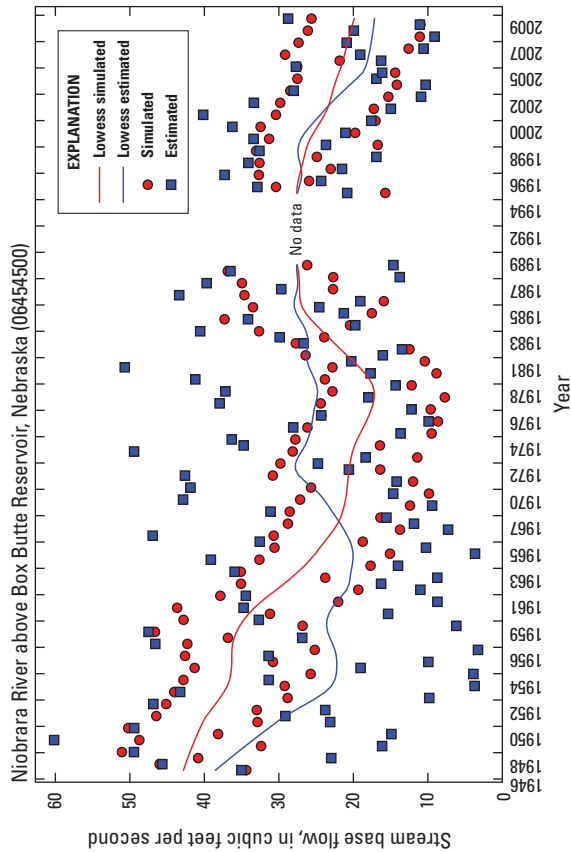


Figure 1-5. Simulated and estimated base flow, 1946-2009, for Niobrara River above Box Butte Reservoir, Nebr. (streamgage number 06454500).

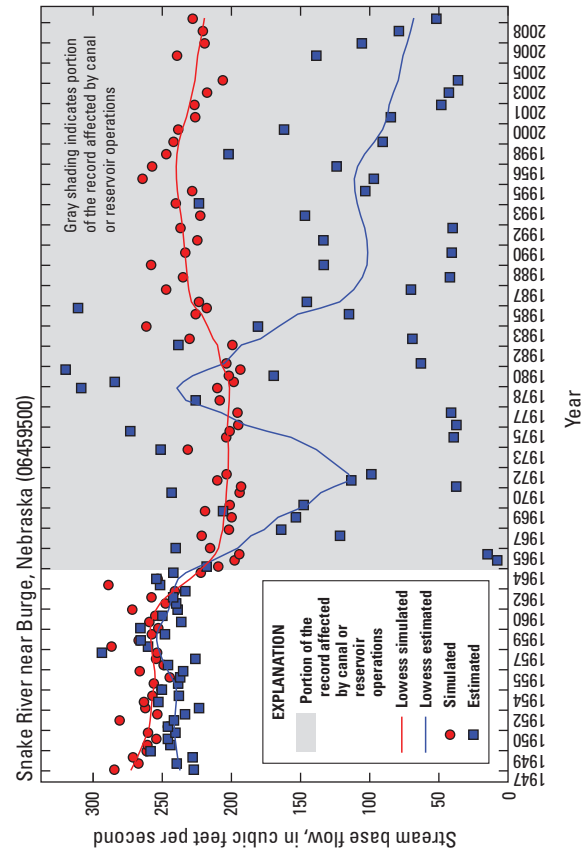


Figure 1-7. Simulated and estimated base flow, 1947-2009, for Snake River near Burge, Nebr. (streamgage number 06459500).

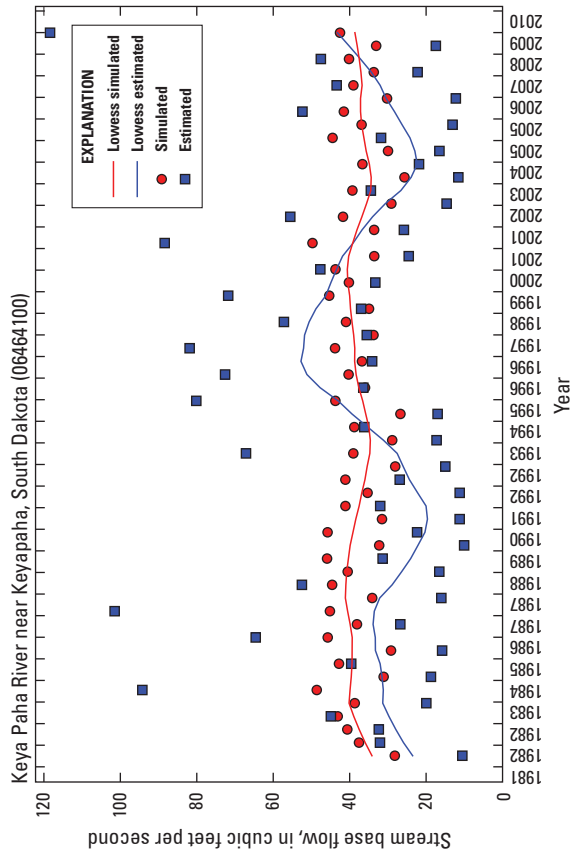


Figure 1–10. Simulated and estimated base flow, 1981–2009, for Keya Paha River near Keyapaha, S. Dak. (streamgage number 06464100).

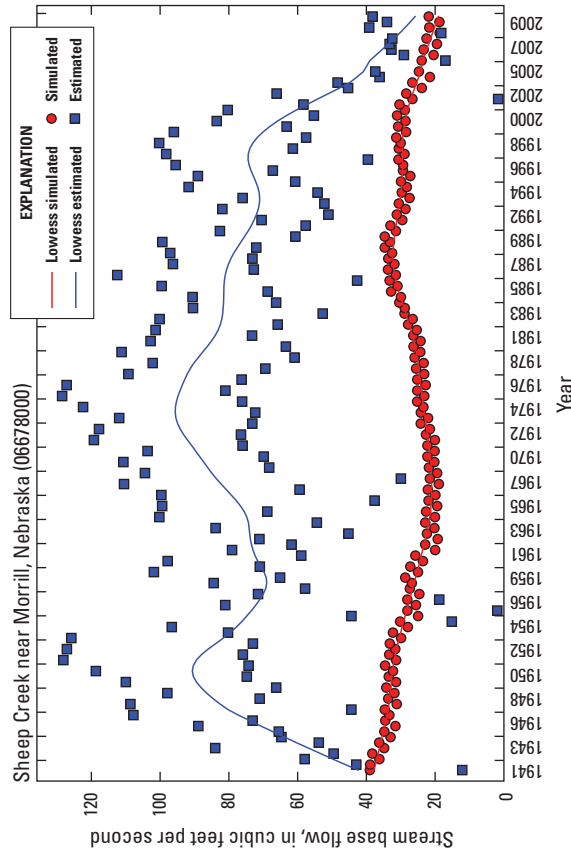


Figure 1–12. Simulated and estimated base flow, 1940–2009, for Sheep Creek near Morrill, Nebr. (streamgage number 06678000).

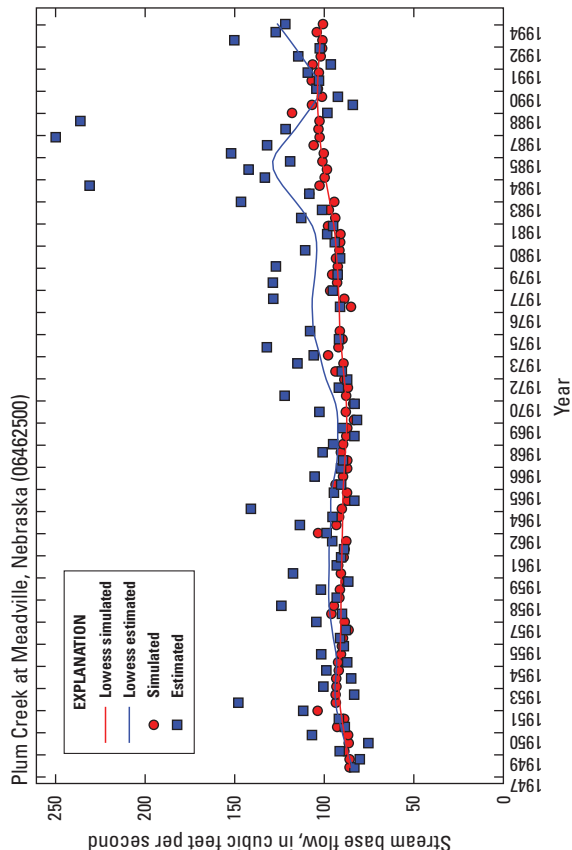


Figure 1–9. Simulated and estimated base flow, 1948–94, for Plum Creek at Meadville, Nebr. (streamgage number 06462500).

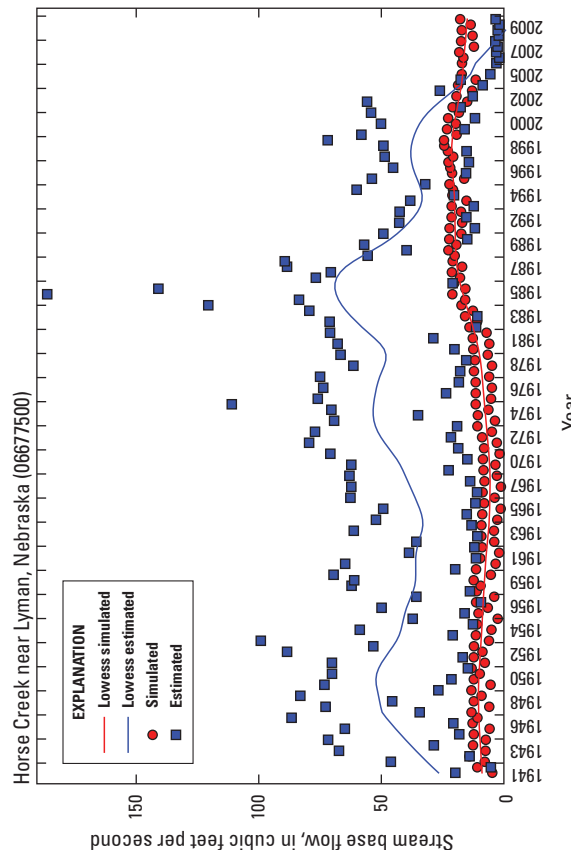


Figure 1–11. Simulated and estimated base flow, 1940–2009, for Horse Creek near Lyman, Nebr. (streamgage number 06677500).

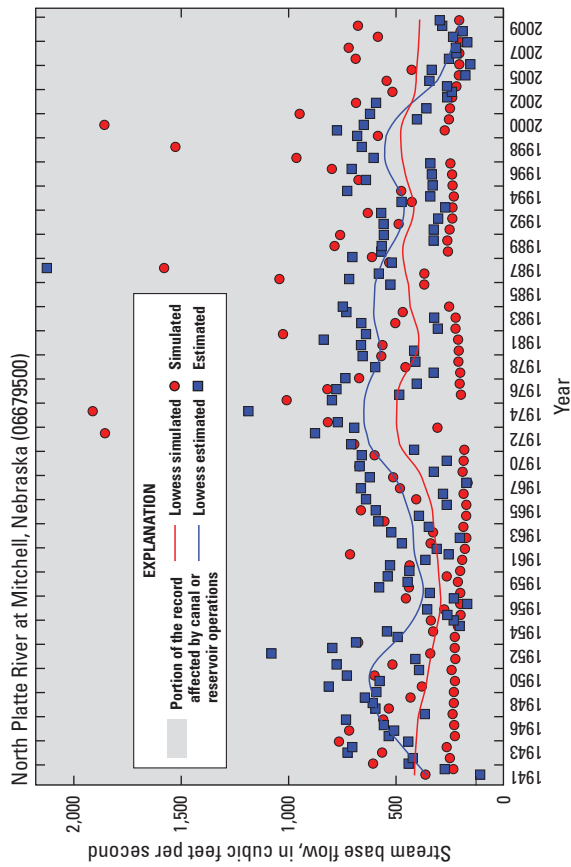


Figure 1-13. Simulated and estimated base flow, 1940–2009, for North Platte River at Mitchell, Nebr., (streamage number 06679500).

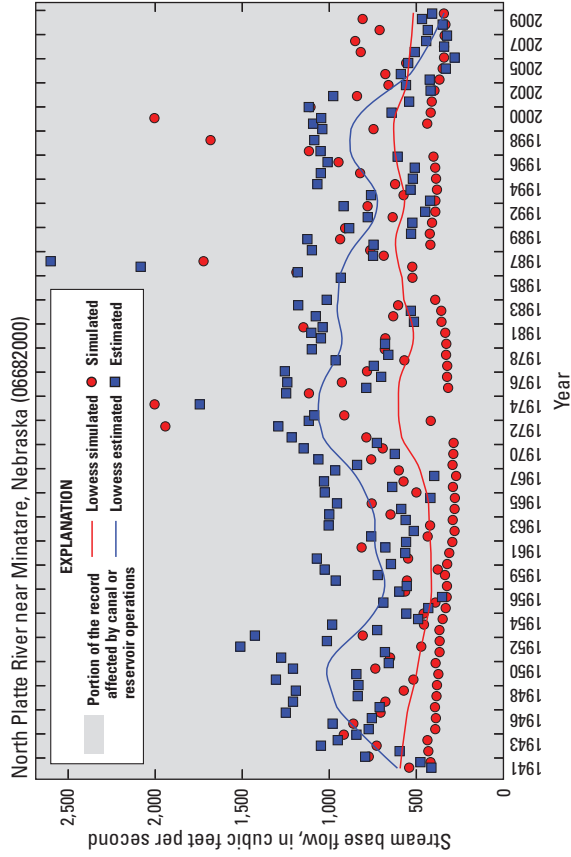


Figure 1-14. Simulated and estimated base flow, 1940–2009, for North Platte River near Minatare, Nebr. (streamage number 06682000).

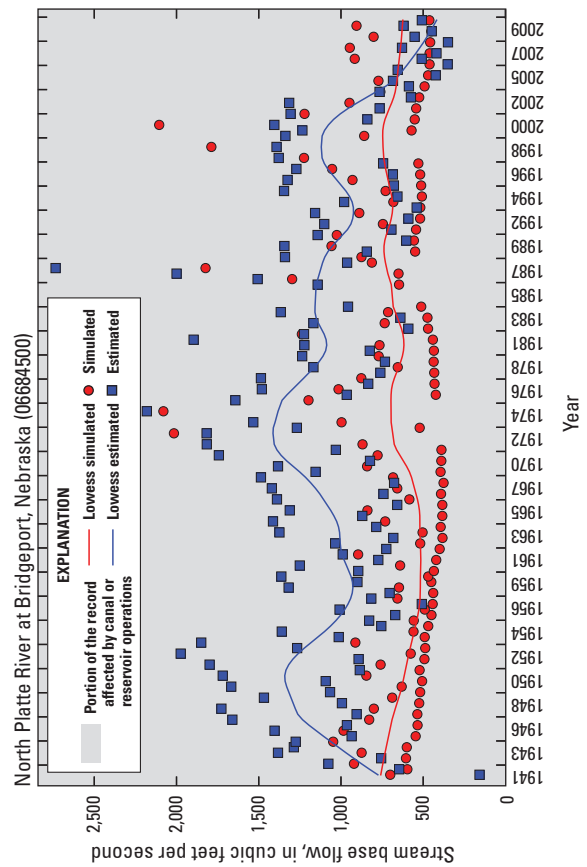


Figure 1-15. Simulated and estimated base flow, 1940–2009, for North Platte River at Bridgeport, Nebr. (streamage number 06684500).

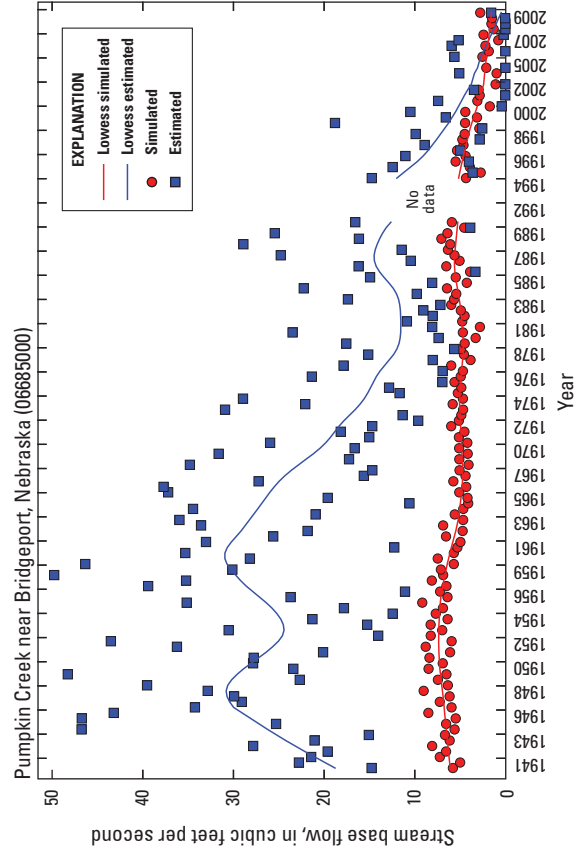


Figure 1-16. Simulated and estimated base flow, 1940–2009, for Pumpkin Creek near Bridgeport, Nebr. (streamage number 06685000).

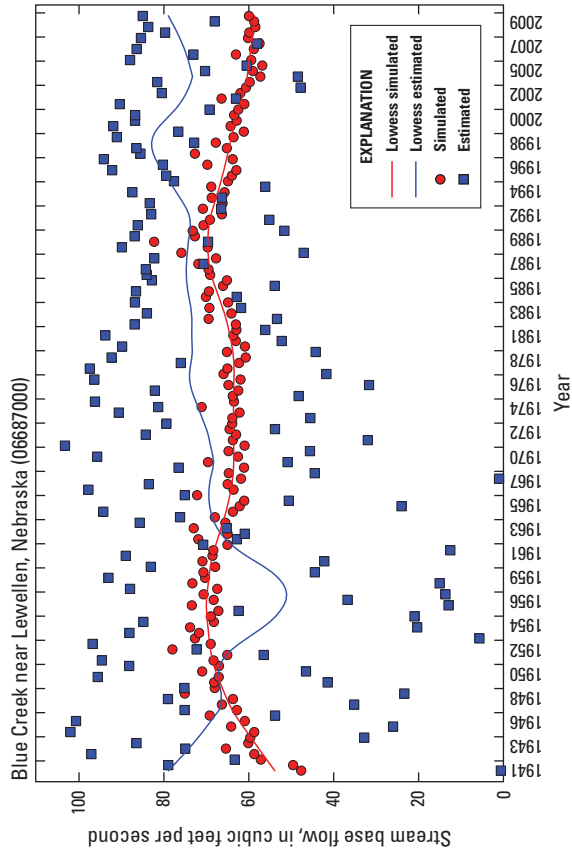


Figure 1-18. Simulated and estimated base flow, 1940–2009, for Blue Creek near Lewellen, Nebr., (streamgage number 06687000).

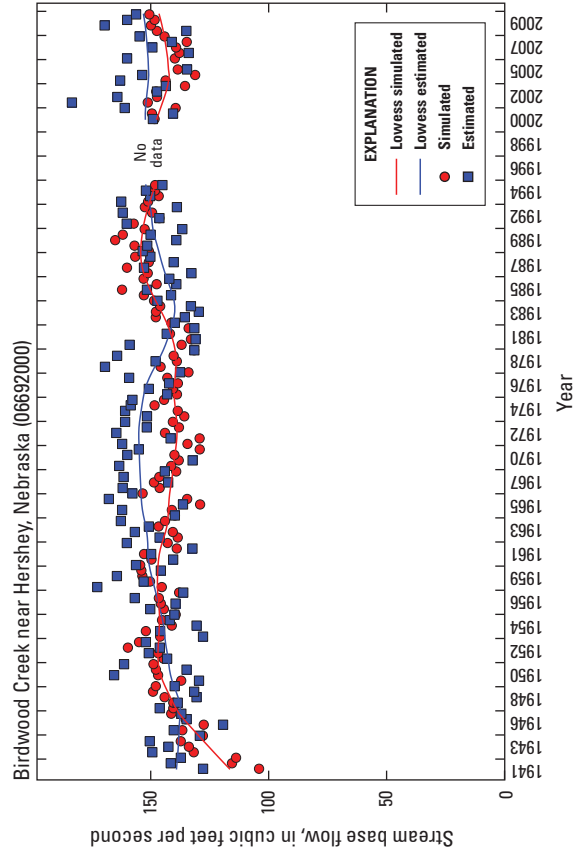


Figure 1-20. Simulated and estimated base flow, 1940–2009, for Birdwood Creek near Hershey, Nebr. (streamgage number 06692000).

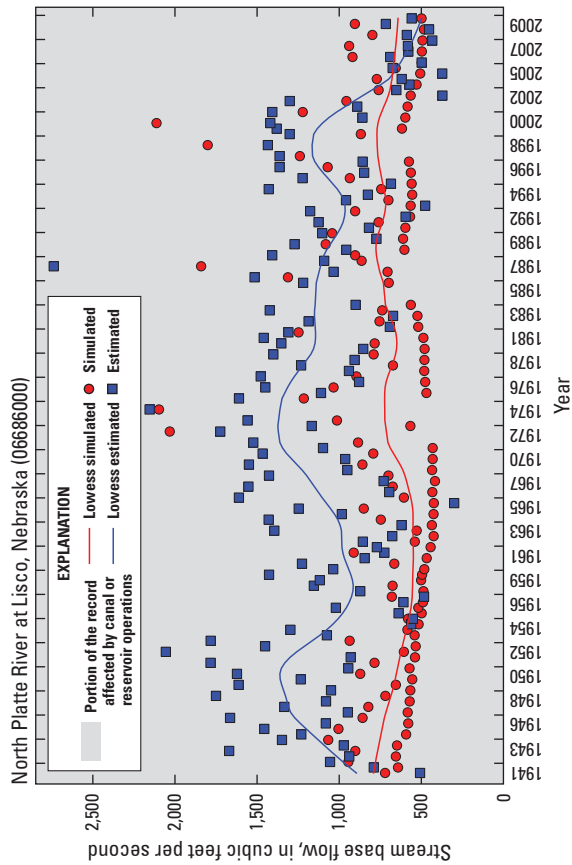


Figure 1-17. Simulated and estimated base flow, 1940–2009, for North Platte River at Lisco, Nebr. (streamgage number 06686000).

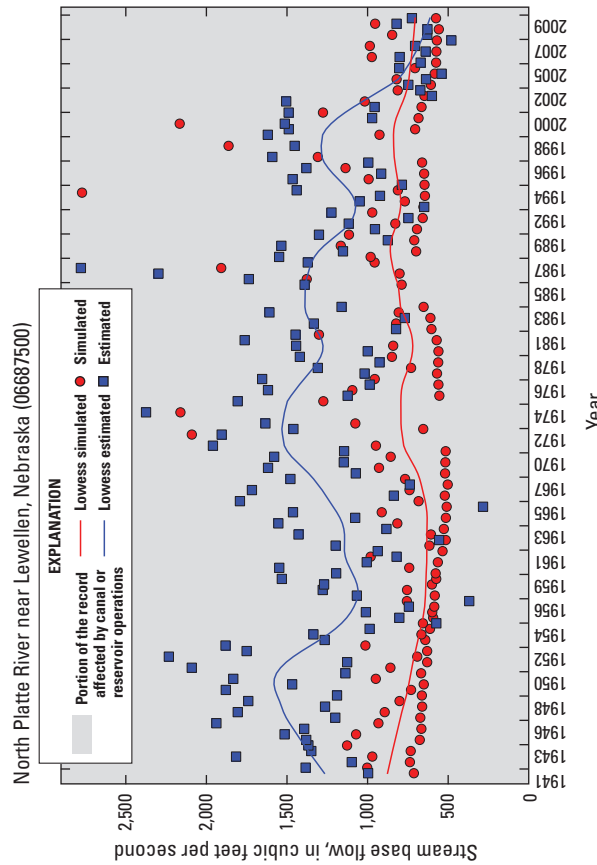


Figure 1-19. Simulated and estimated base flow, 1941–2009, for North Platte River near Lewellen, Nebr. (streamgage number 06687500).

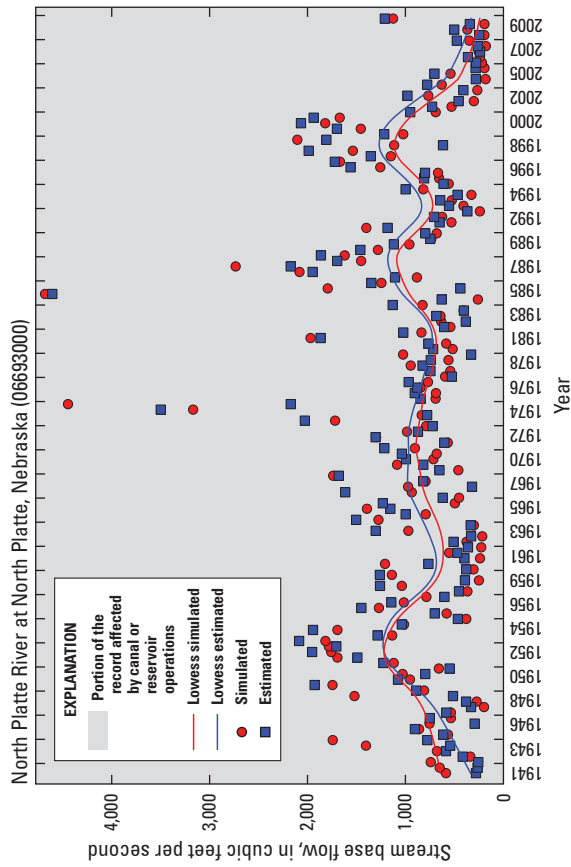


Figure 1–21. Simulated and estimated base flow, 1940–2009, for North Platte River at North Platte, Nebr. (streamage number 06693000).

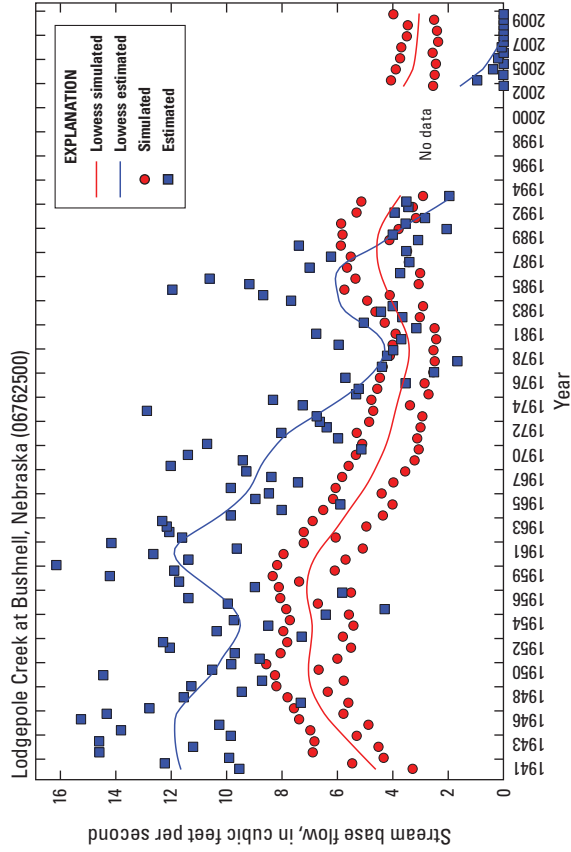


Figure 1–22. Simulated and estimated base flow, 1940–2009, for Lodgepole Creek at Bushnell, Nebr. (streamage number 06762500).

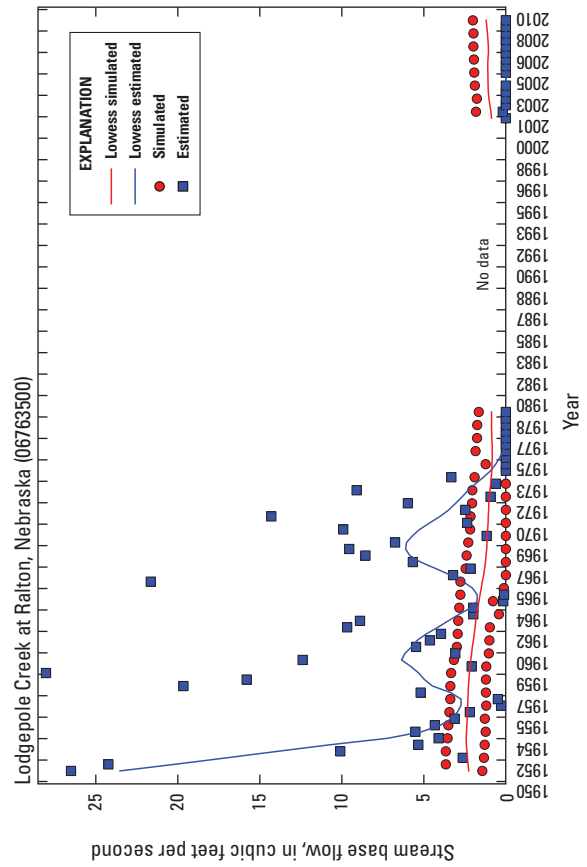


Figure 1–23. Simulated and estimated base flow, 1951–2009, for Lodgepole Creek at Ralton, Nebr. (streamage number 06763500).

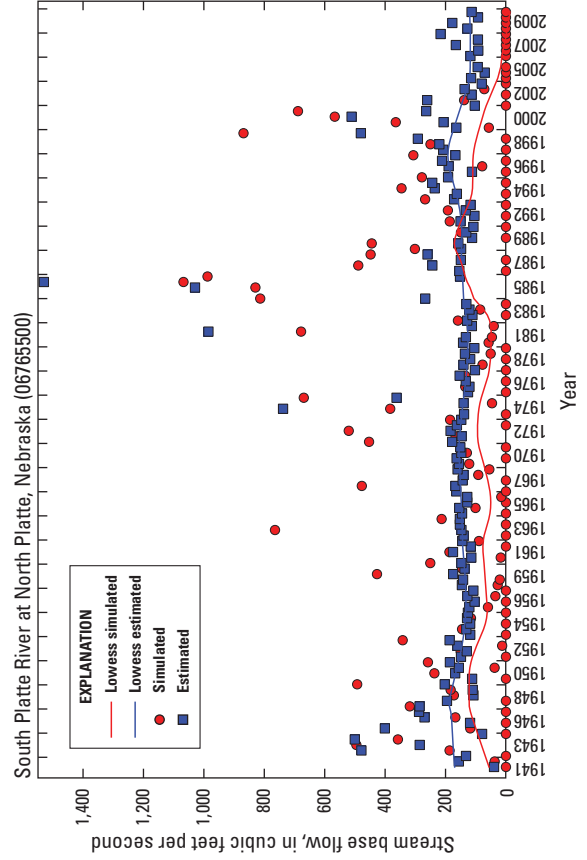


Figure 1–24. Simulated and estimated base flow, 1940–2009, for South Platte River at South Platte, Nebr., (streamage number 06765500).

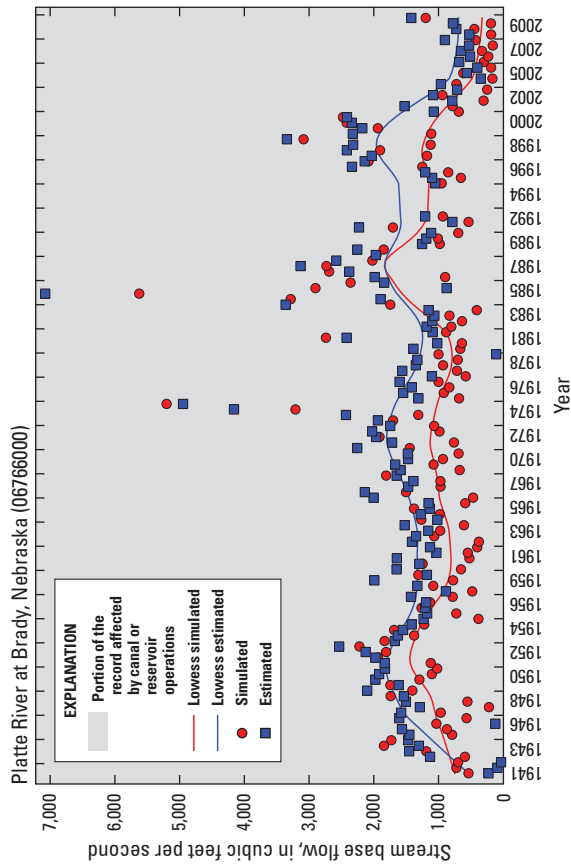


Figure 1-25. Simulated and estimated base flow, 1940–2009, for Platte River at Brady, Nebr. (streamgage number 06766000).

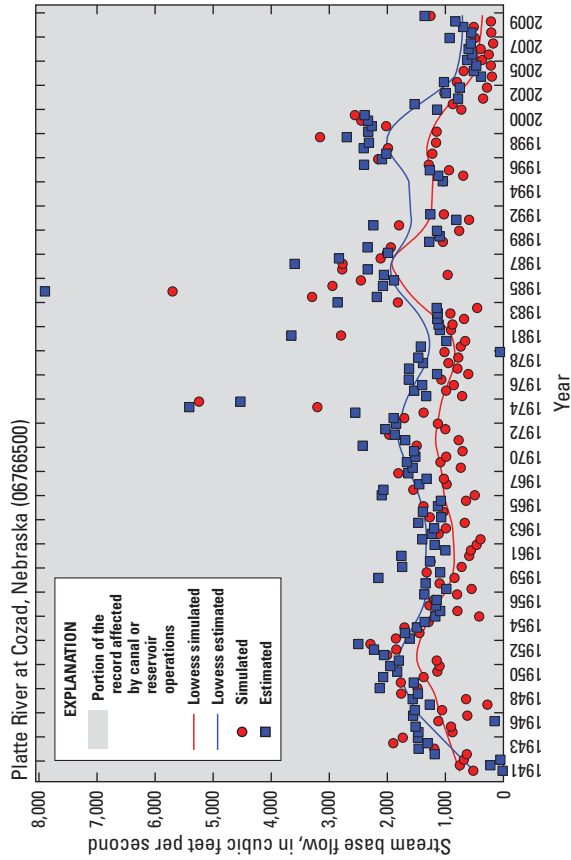


Figure 1-26. Simulated and estimated base flow, 1940–2009, for Platte River at Cozad, Nebr. (streamgage number 06766500).

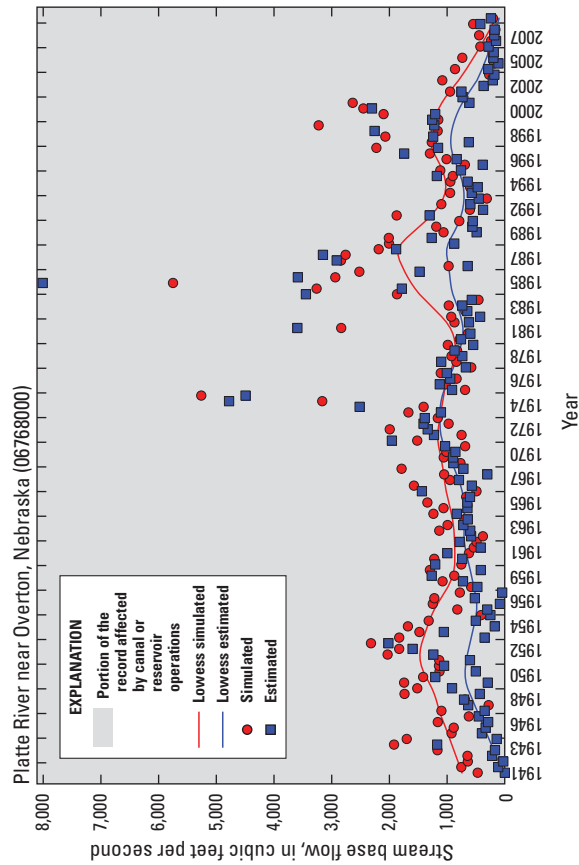


Figure 1-27. Simulated and estimated base flow, 1940–2009, for Platte River near Overton, Nebr. (streamgage number 06768000).

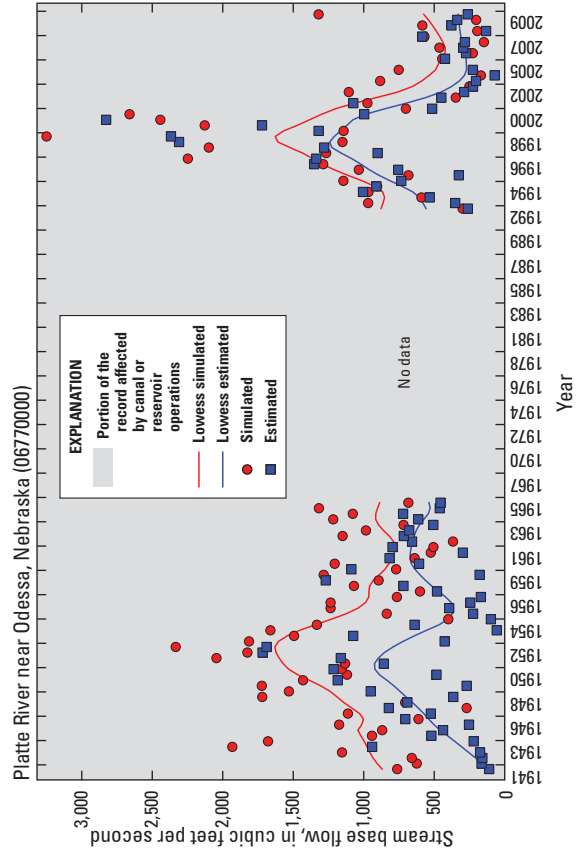


Figure 1-28. Simulated and estimated base flow, 1941–2009, for Platte River near Odessa, Nebr. (streamgage number 06770000).

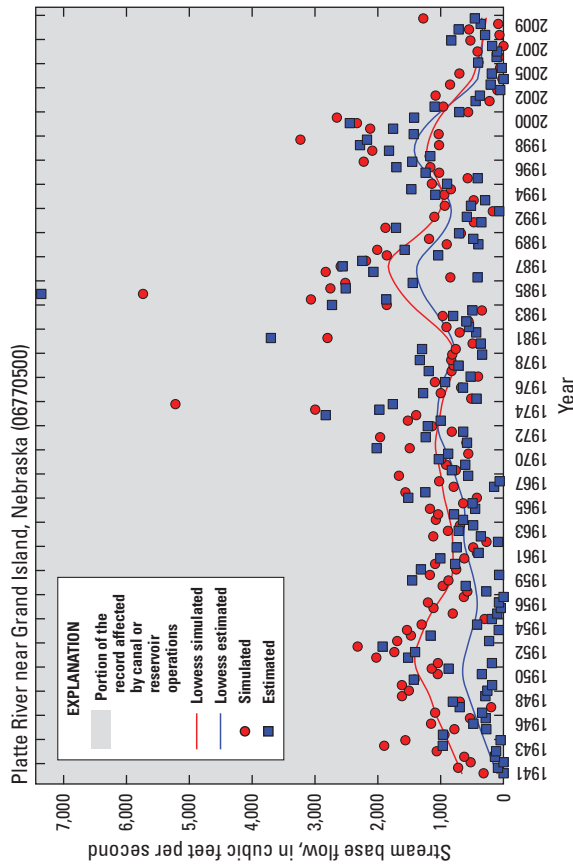


Figure 1–29. Simulated and estimated base flow, 1940–2009, for Platte River near Grand Island, Nebr. (streamgage number 06770500).

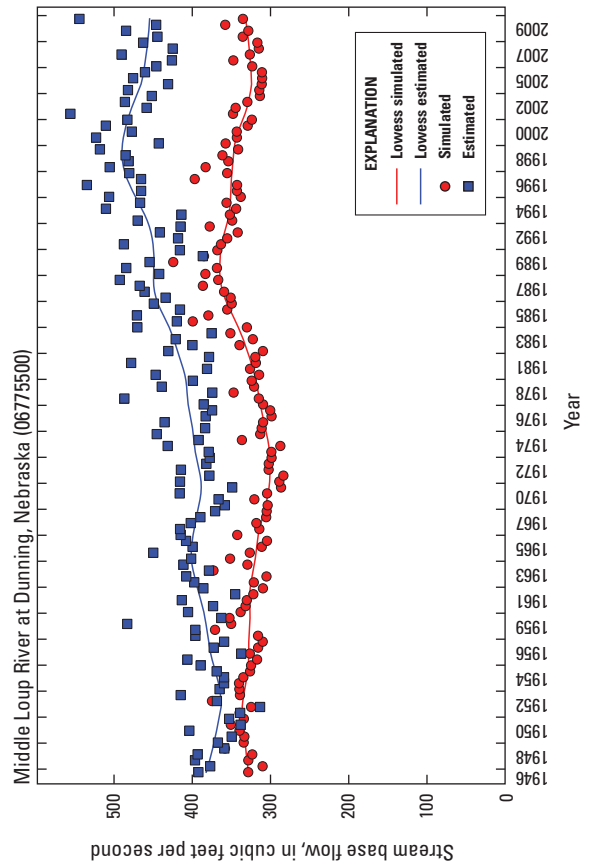


Figure 1–31. Simulated and estimated base flow, 1945–2009, for Middle Loup River at Dunning, Nebr. (streamgage number 06775500).

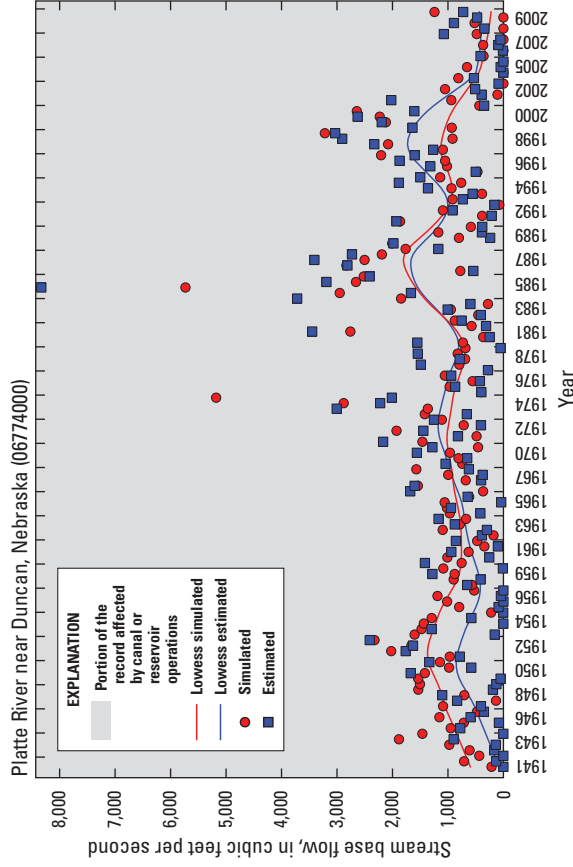


Figure 1–30. Simulated and estimated base flow, 1940–2009, for Platte River near Duncan, Nebr. (streamgage number 06774000).

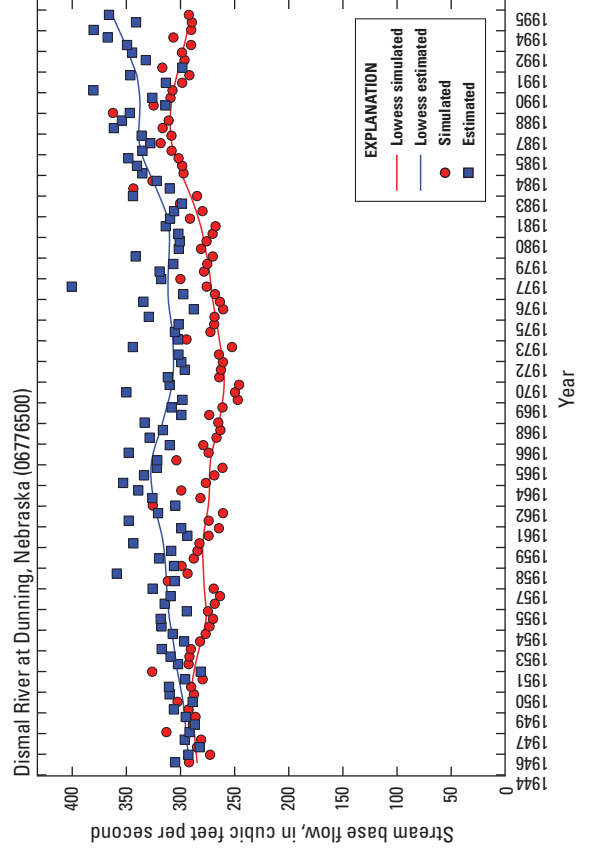


Figure 1–32. Simulated and estimated base flow, 1945–2009, for Dismal River at Dunning, Nebr. (streamgage number 06776500).

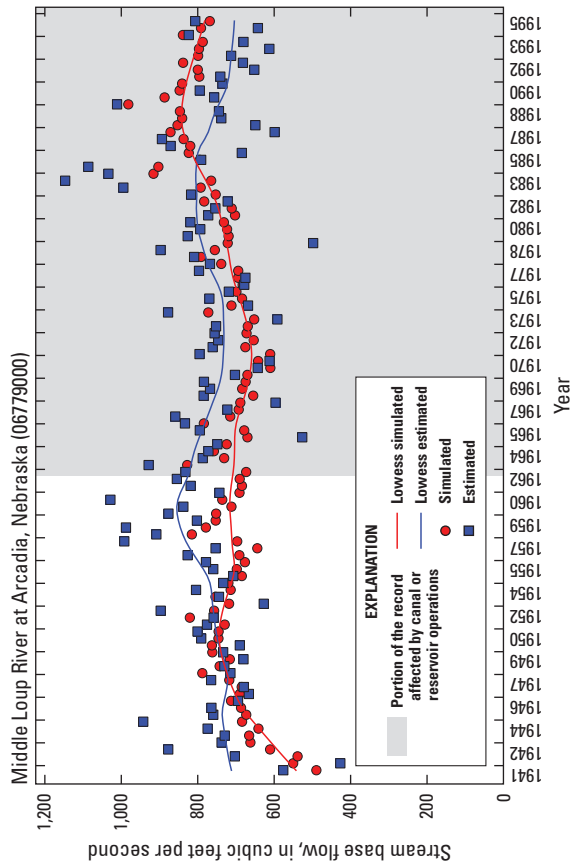


Figure 1-33. Simulated and estimated base flow, 1940–45, for Middle Loup River at Arcadia, Nebr. (streamage number 06779000).

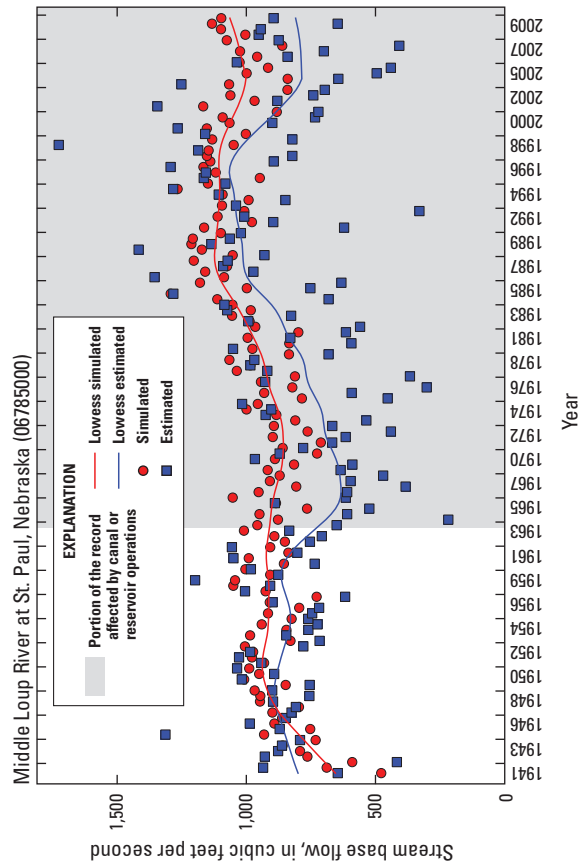


Figure 1-35. Simulated and estimated base flow, 1940–2009, for Middle Loup River at St. Paul, Nebr. (streamage number 06785000).

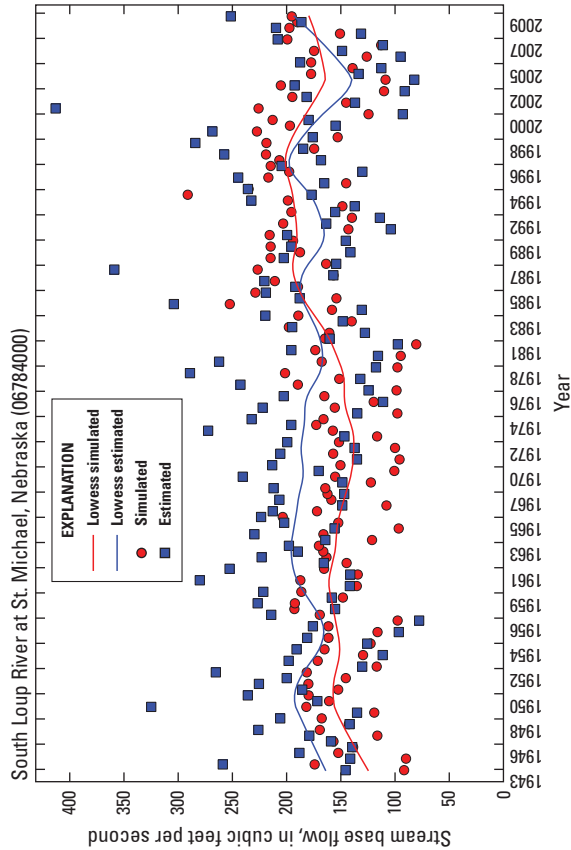


Figure 1-34. Simulated and estimated base flow, 1943–2009, for South Loup River at St. Michael, Nebr. (streamage number 06784000).

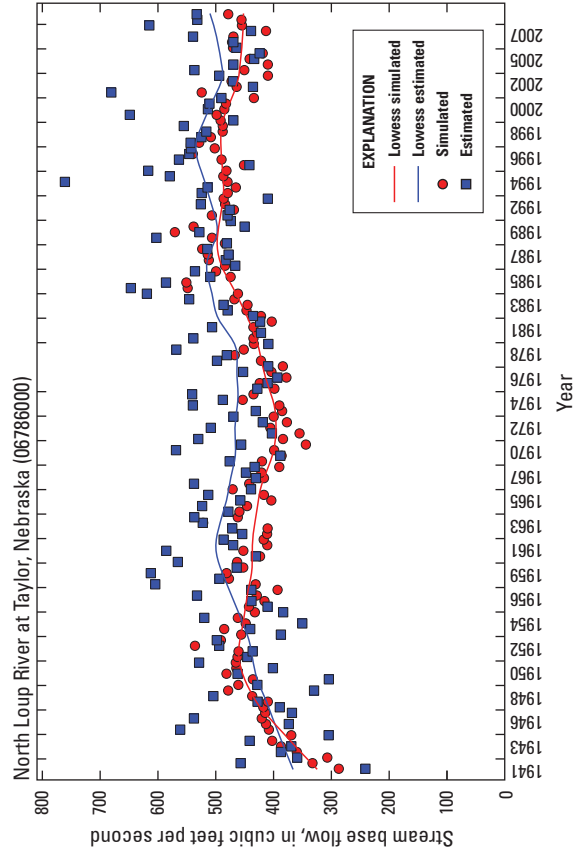


Figure 1-36. Simulated and estimated base flow, 1940–2008, for North Loup River at Taylor, Nebr. (streamage number 06786000).

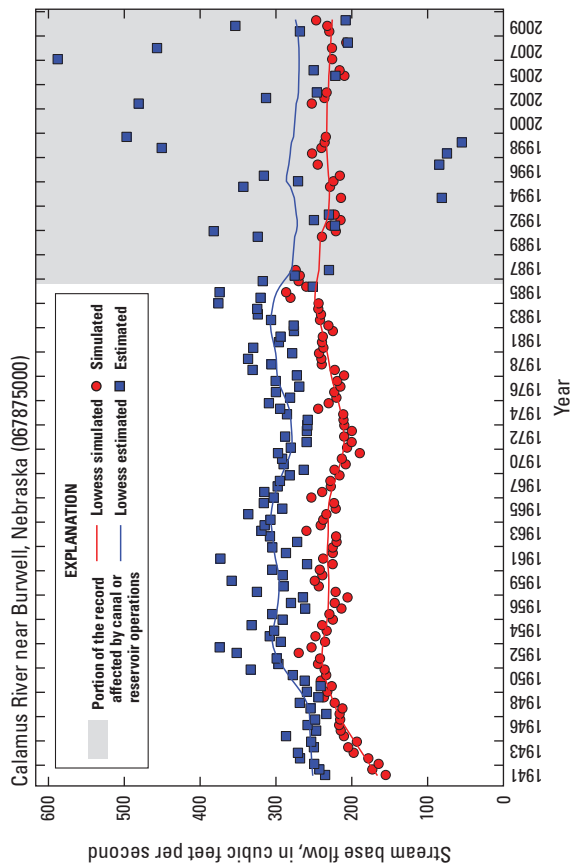


Figure 1-37. Simulated and estimated base flow, 1940–2009, for Calamus River near Burwell, Nebr. (streamgage number 06787500).

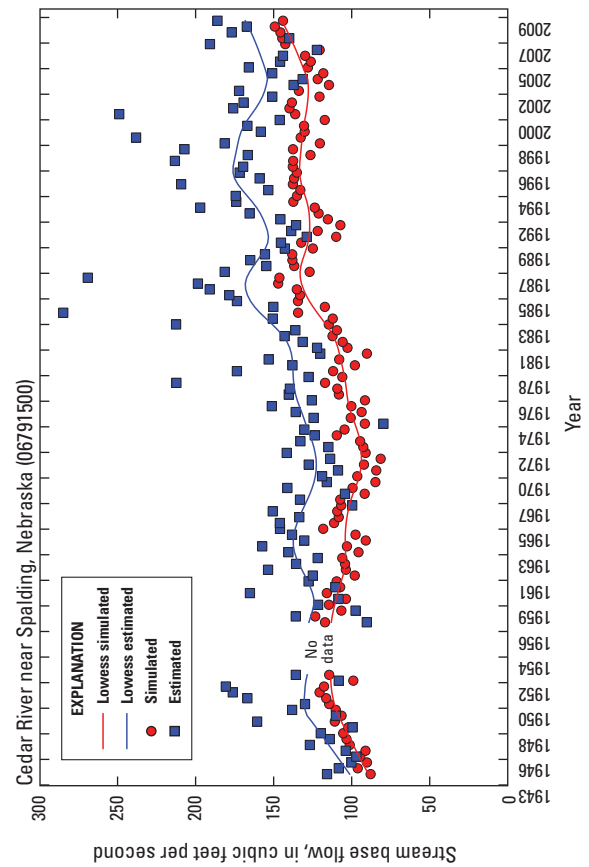


Figure 1-39. Simulated and estimated base flow, 1944–2009, for Cedar River near Spalding, Nebr. (streamgage number 06791500).

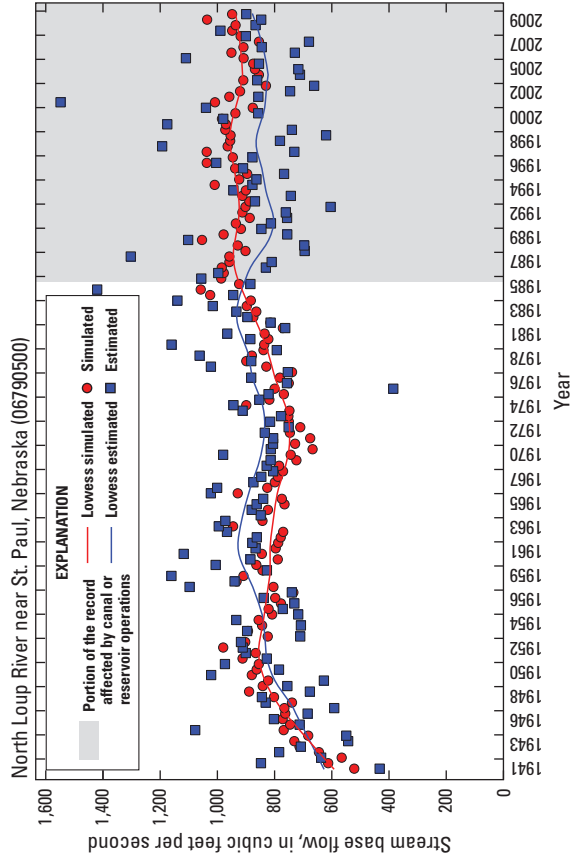


Figure 1-38. Simulated and estimated base flow, 1940–2009, for North Loup River near St. Paul, Nebr. (streamgage number 06790500).

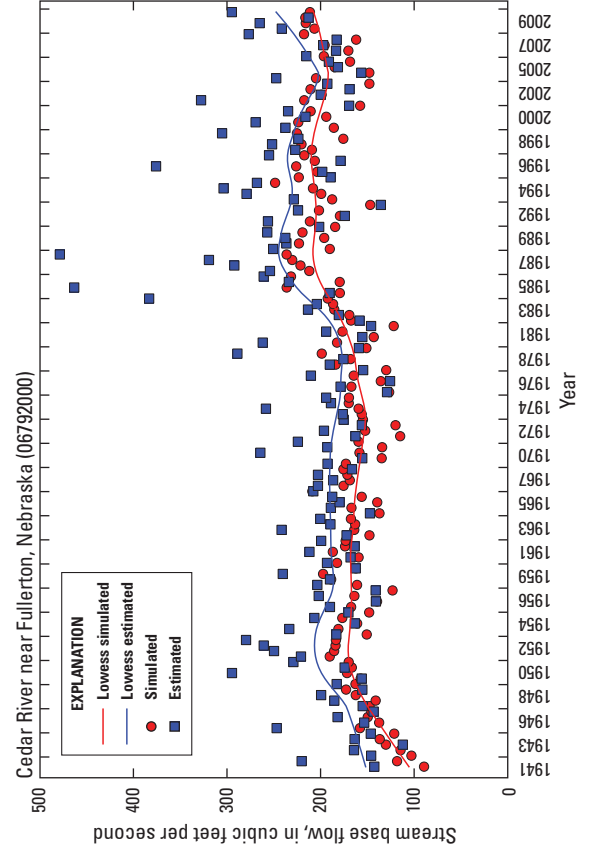


Figure 1-40. Simulated and estimated base flow, 1940–2009, for Cedar River near Fullerton, Nebr. (streamgage number 06792000).

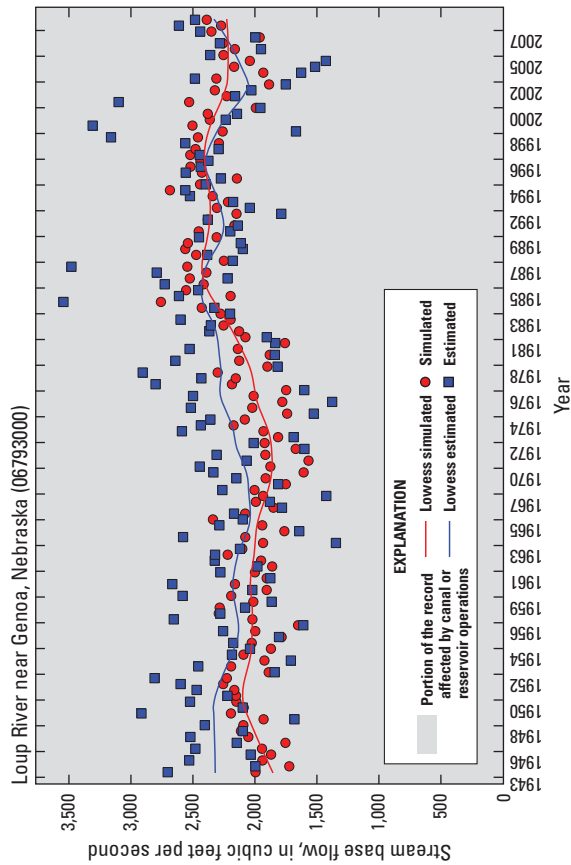


Figure 1-41. Simulated and estimated base flow, 1944–2008, for Loup River near Genoa, Nebr. (streamgage number 06793000).

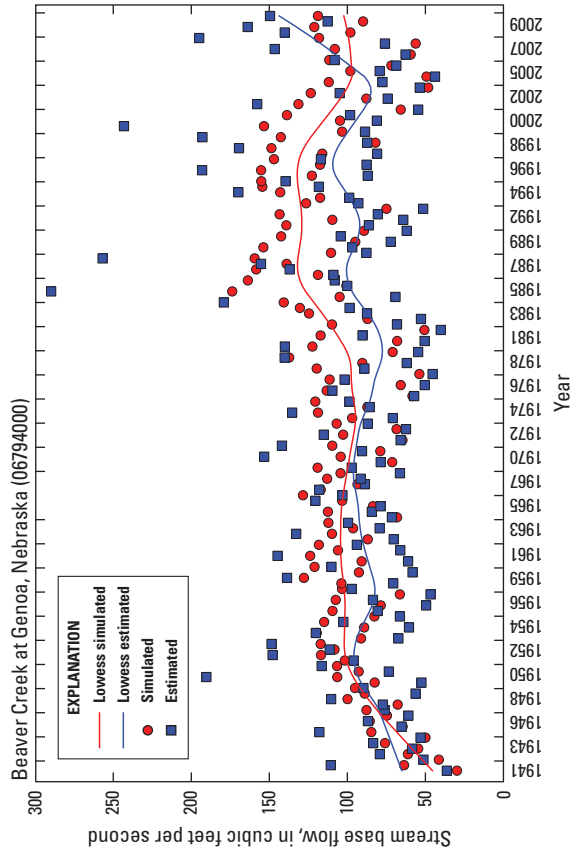


Figure 1-42. Simulated and estimated base flow, 1940–2009, for Beaver Creek at Genoa, Nebr. (streamgage number 06794000).

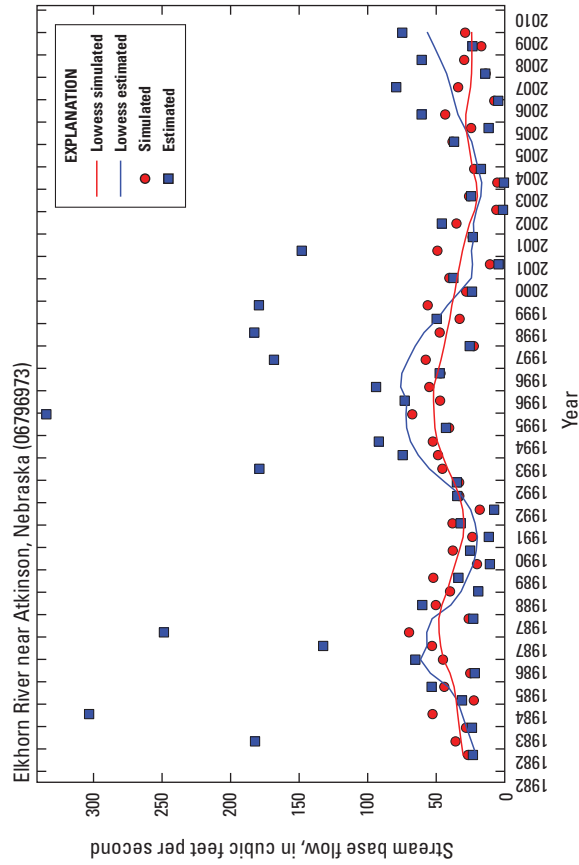


Figure 1-43. Simulated and estimated base flow, 1982–2009, for Elkhorn River near Atkinson, Nebr. (streamgage number 06796973).

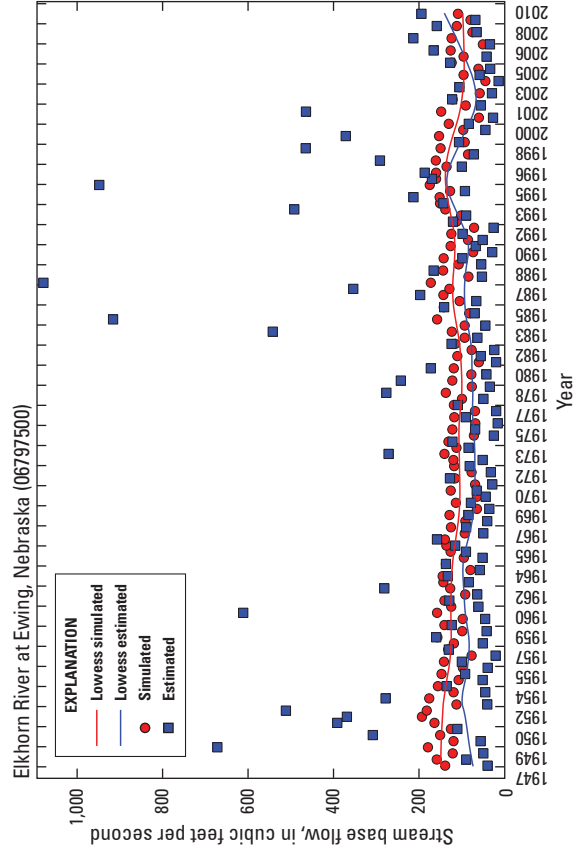


Figure 1-44. Simulated and estimated base flow, 1947–2009, for Elkhorn River at Ewing, Nebr. (streamgage number 06797500).

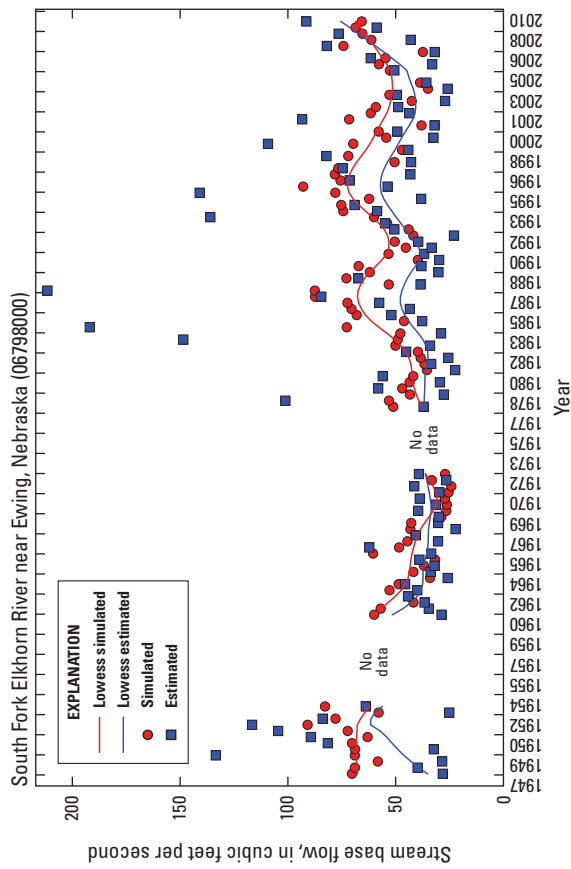


Figure 1-45. Simulated and estimated base flow, 1947–2009, for South Fork Elkhorn River near Ewing, Nebr. (streamage number 06798000).

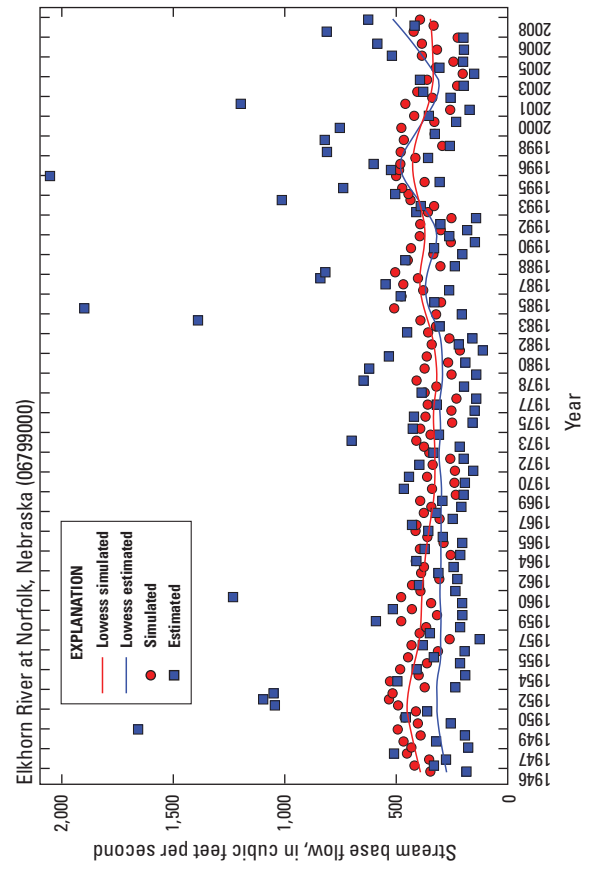


Figure 1-47. Simulated and estimated base flow, 1945–2008, for Elkhorn River at Norfolk, Nebr. (streamage number 06799000).

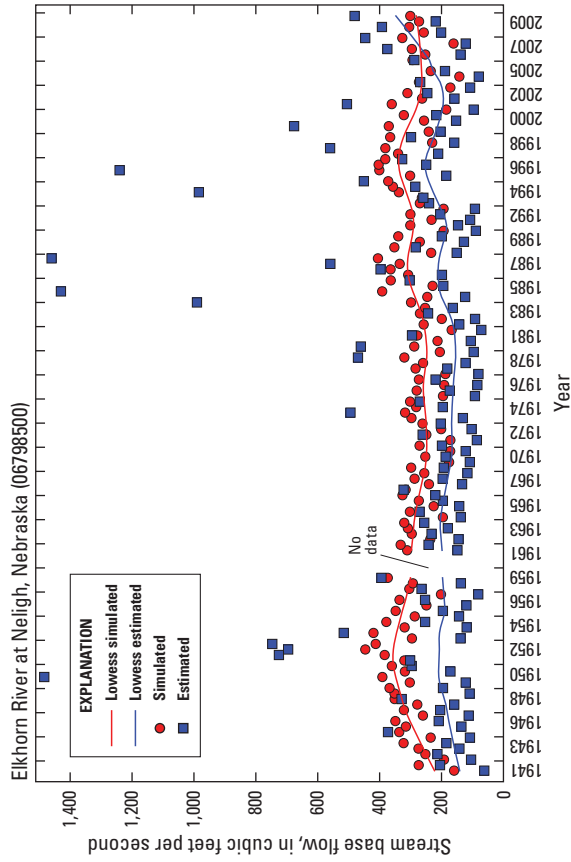


Figure 1-46. Simulated and estimated base flow, 1940–2009, for Elkhorn River at Neligh, Nebr. (streamage number 06798500).

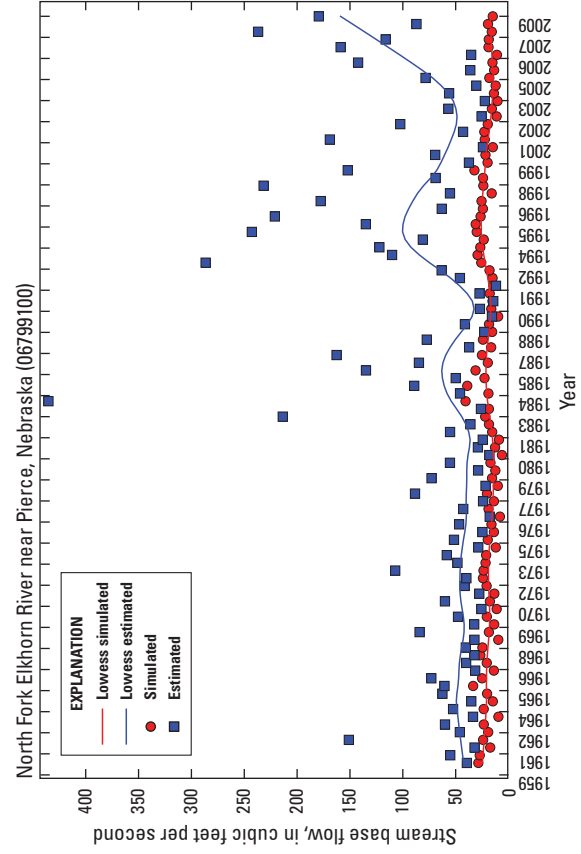


Figure 1-48. Simulated and estimated base flow, 1960–2009, for North Fork Elkhorn River near Pierce, Nebr. (streamage number 06799100).

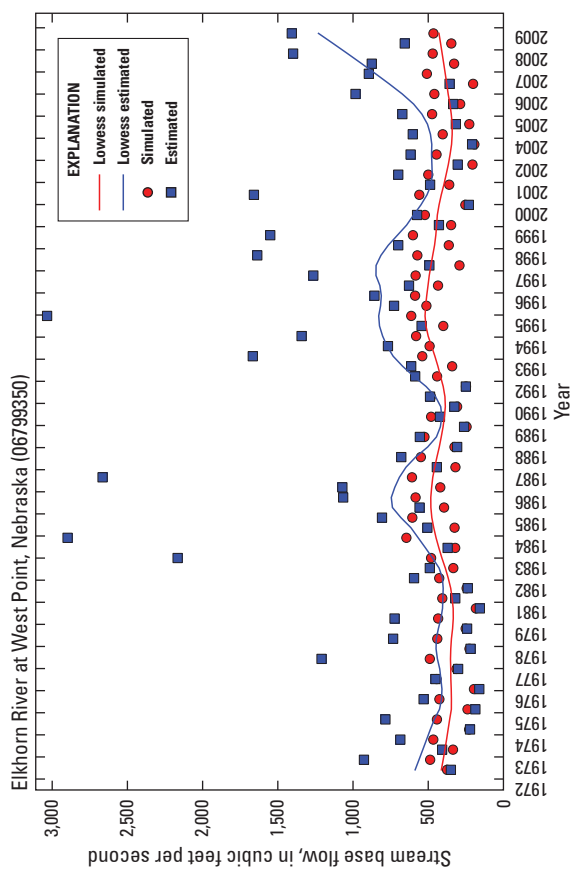


Figure 1-49. Simulated and estimated base flow, 1972–2009, for Elkhorn River at West Point, Nebr. (streamage number 06799350).

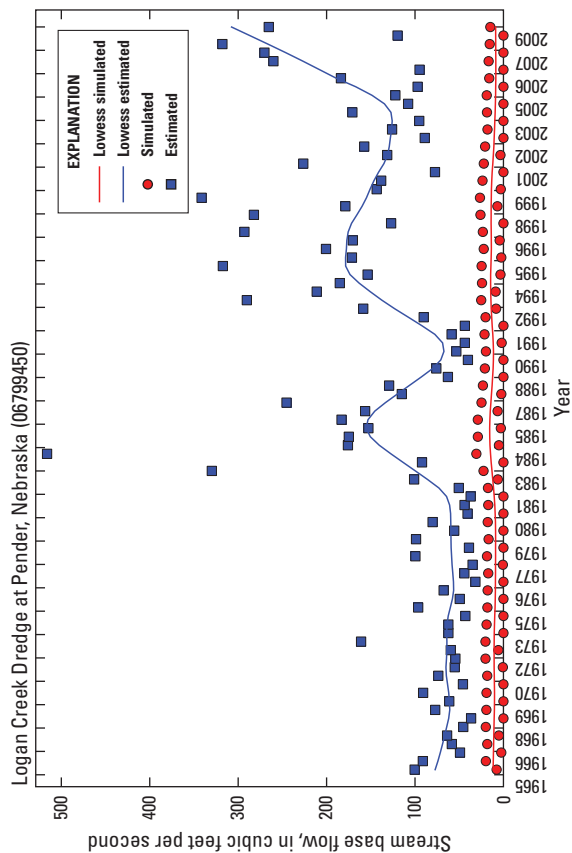


Figure 1-50. Simulated and estimated base flow, 1965–2009, for Logan Creek at Pender, Nebr., (streamage number 06799450).

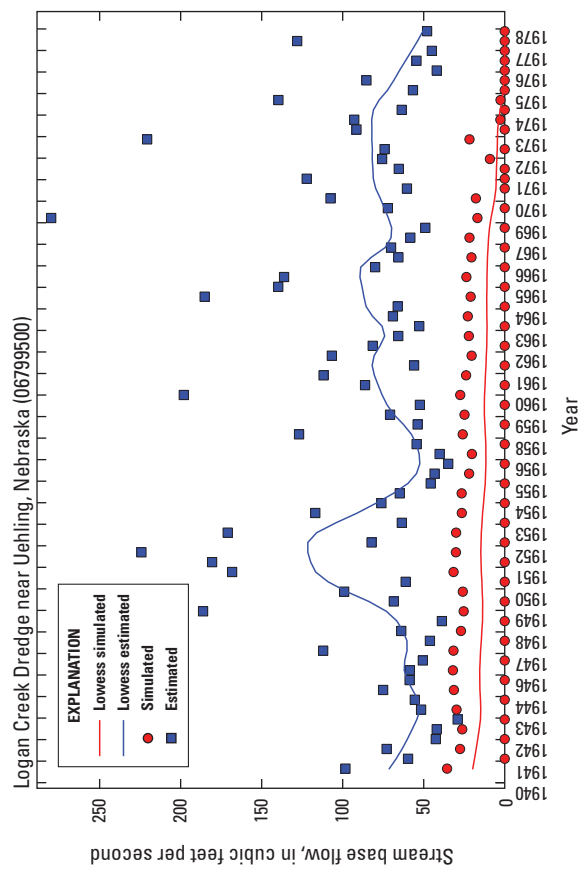


Figure 1-51. Simulated and estimated base flow, 1941–1978, for Logan Creek near Uehling, Nebr. (streamage number 06799500).

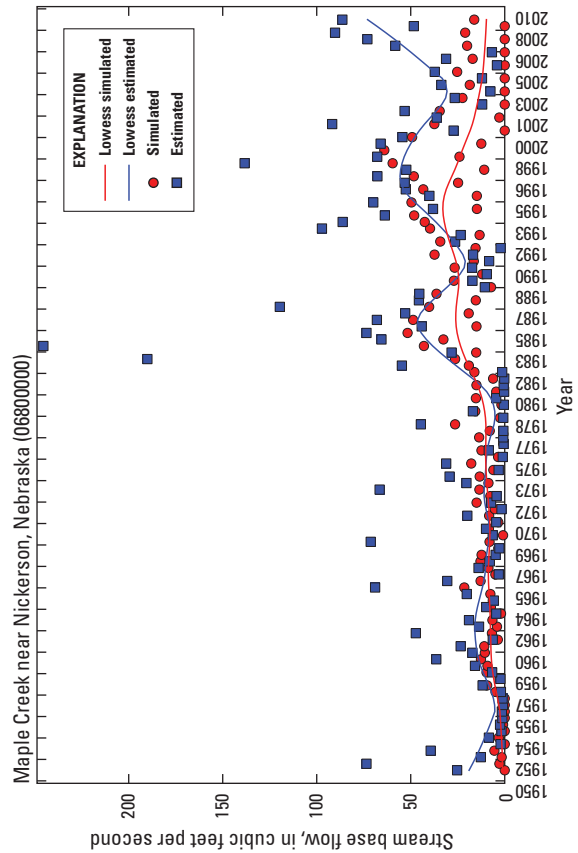


Figure 1-52. Simulated and estimated base flow, 1951–2009, for Maple Creek near Nickerson, Nebr. (streamage number 06800000).

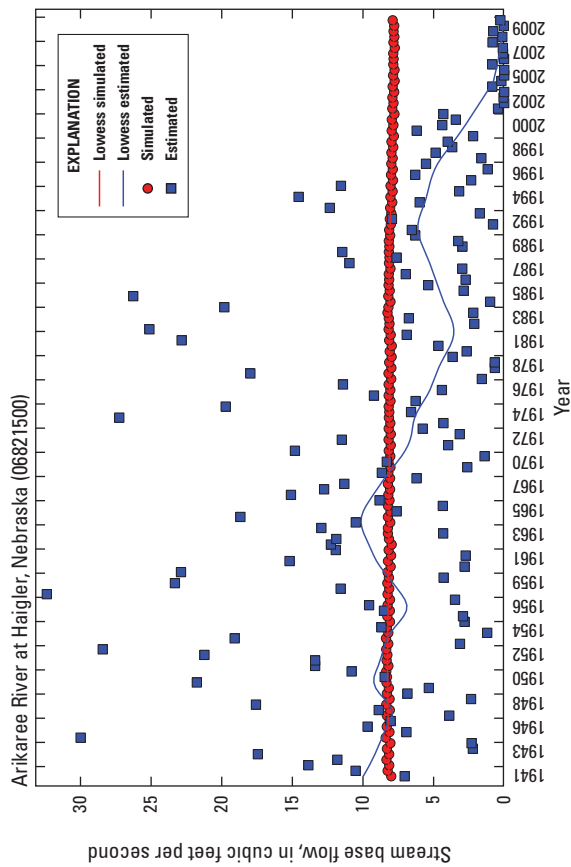


Figure 1-53. Simulated and estimated base flow, 1940-2009, for Arikaree River at Haigler, Nebr. (streamage number 06821500).

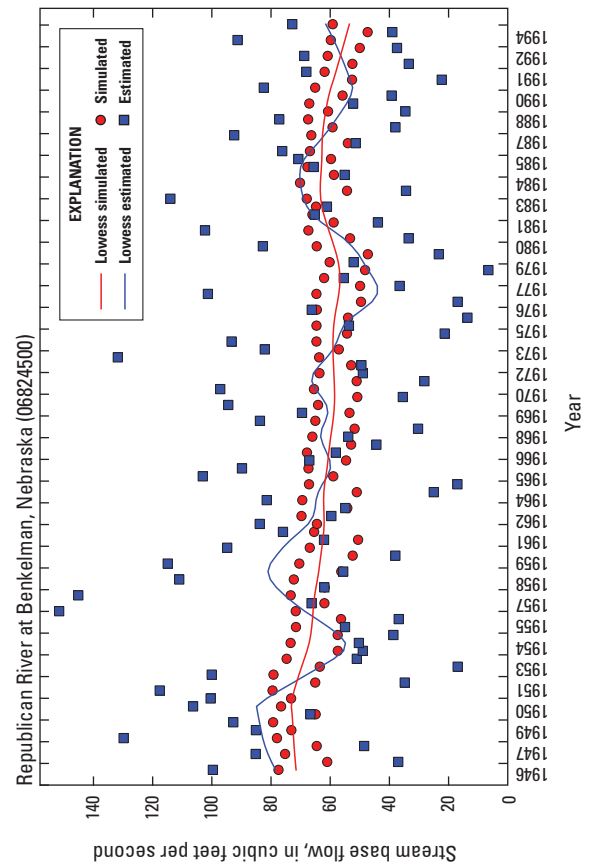


Figure 1-55. Simulated and estimated base flow, 1947-94, for Republican River at Benkelman, Nebr. (streamage number 06824500).

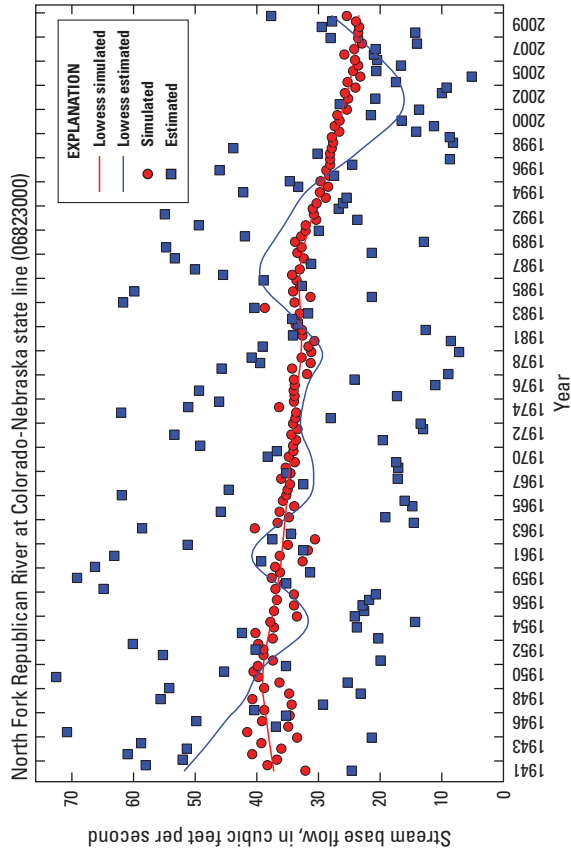


Figure 1-54. Simulated and estimated base flow, 1940-2009, for North Fork Republican River at the Colorado-Nebraska State Line (streamage number 06823000).

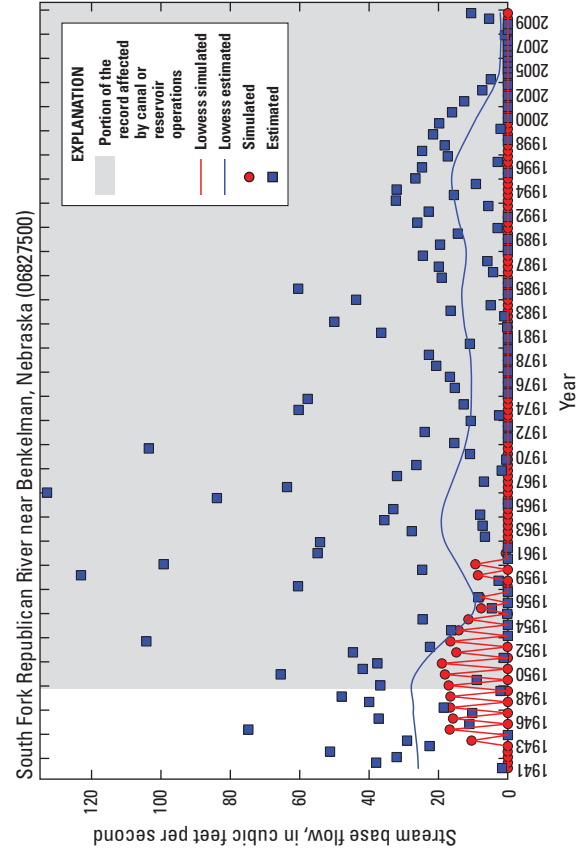


Figure 1-56. Simulated and estimated base flow, 1940-2009, for South Fork Republican River near Benkelman, Nebr. (streamage number ID 6827500).

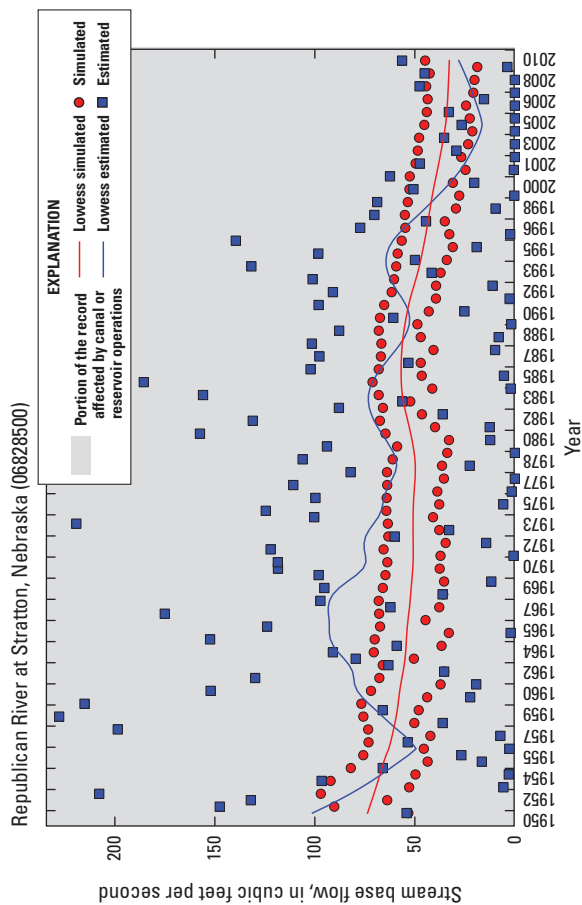


Figure 1-57. Simulated and estimated base flow, 1950–2009, for Republican River at Stratton, Nebr. (streamgage number 06828500).

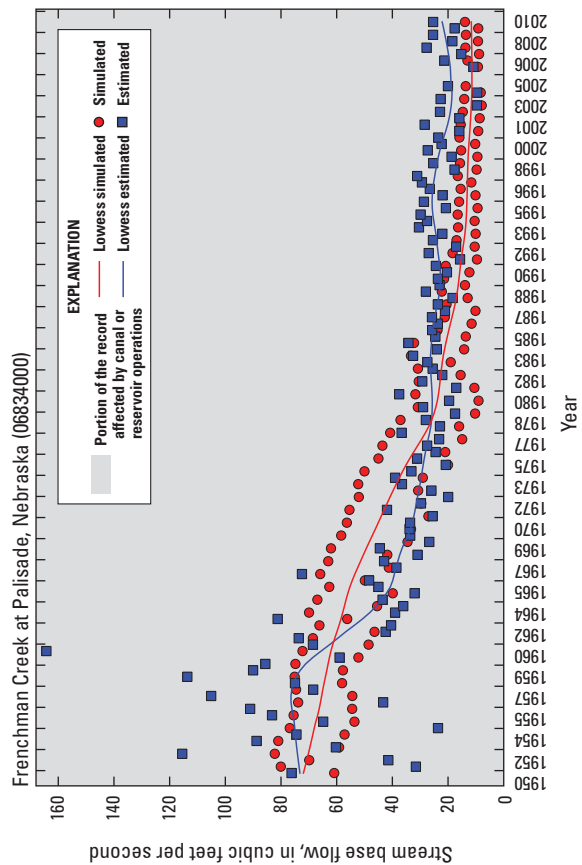


Figure 1-59. Simulated and estimated base flow, 1950–2009, for Frenchman Creek at Palisade, Nebr. (streamgage number 06834000).

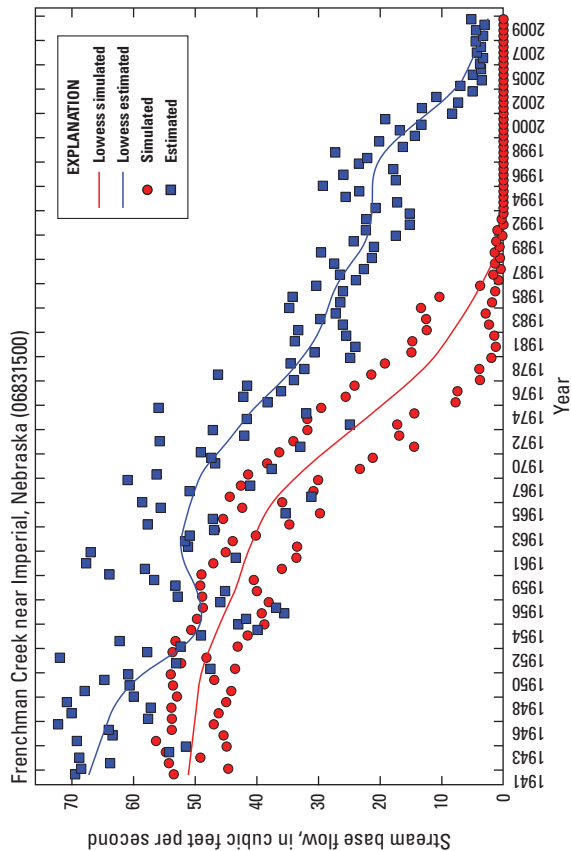


Figure 1-58. Simulated and estimated base flow, 1941–2009, for Frenchman Creek near Imperial, Nebr. (streamgage number 06831500).

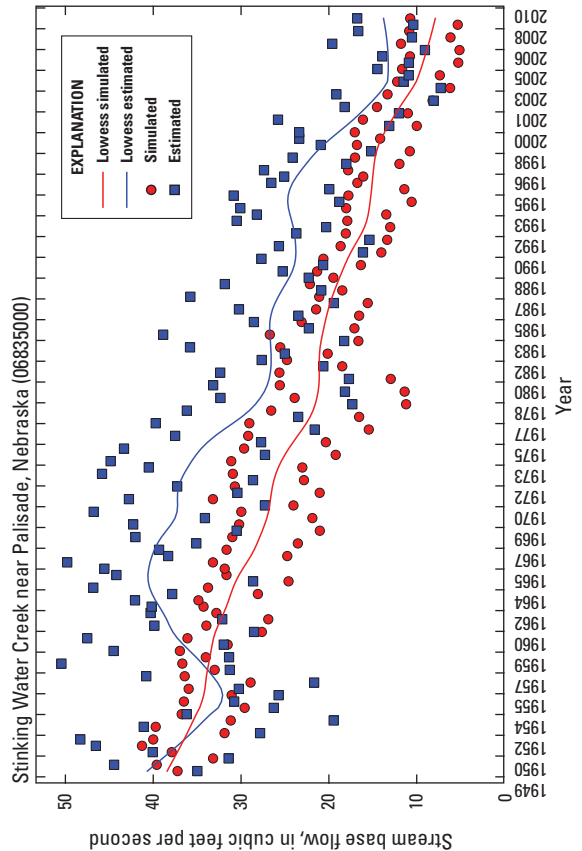


Figure 1-60. Simulated and estimated base flow, 1949–2009, for Stinking Water Creek near Palisade, Nebr. (streamgage number 06835000).

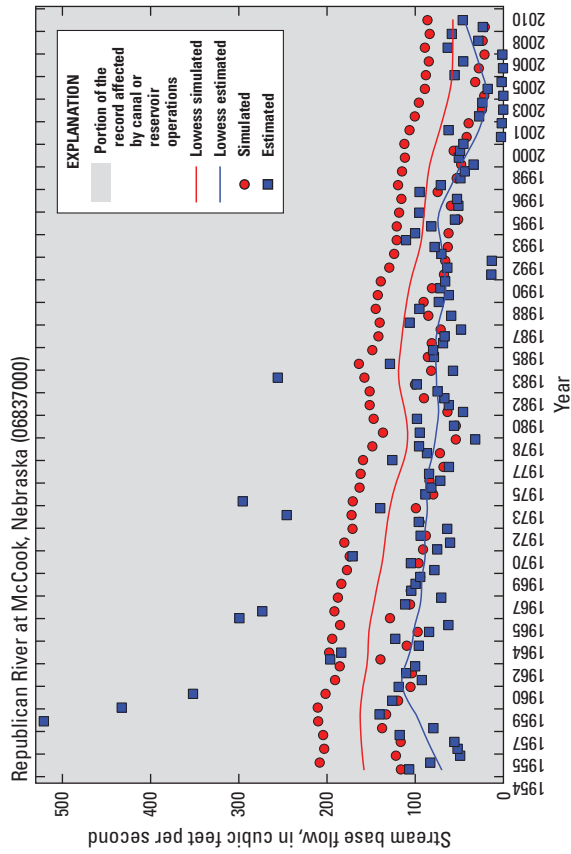


Figure 1-62. Simulated and estimated base flow, 1954–2009, for Republican River at McCook, Nebr. (streamgage number 06837000).

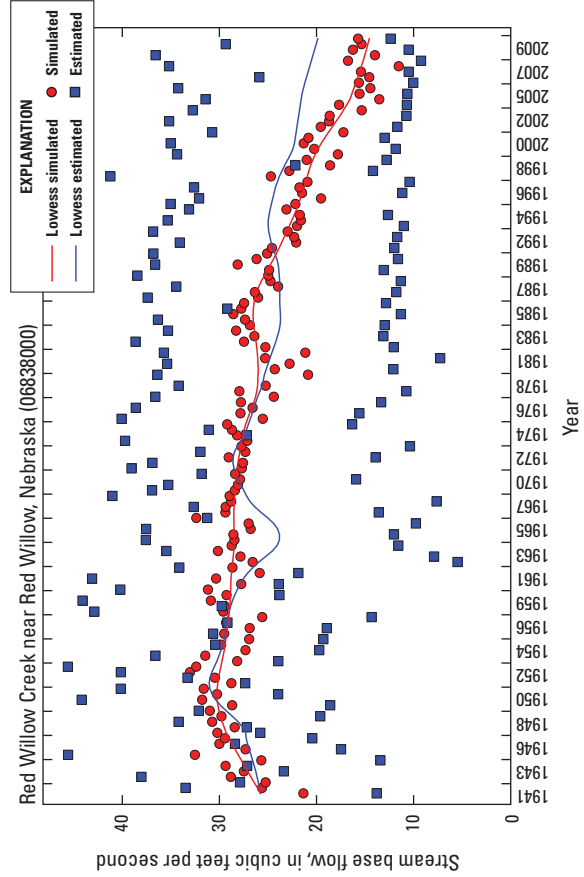


Figure 1-64. Simulated and estimated base flow, 1940–2009, for Red Willow Creek near Red Willow, Nebr. (streamgage number 06838000).

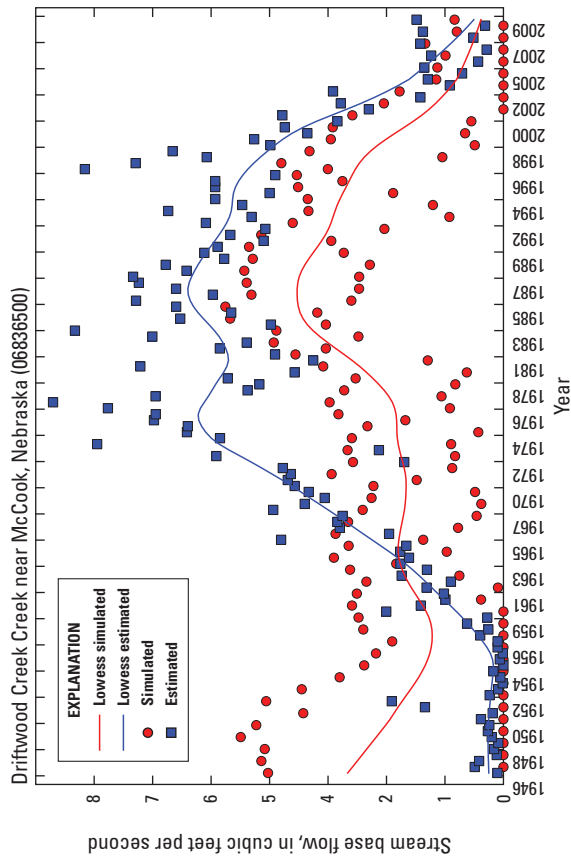


Figure 1-61. Simulated and estimated base flow, 1946–2009, for Driftwood Creek near McCook, Nebr. (streamgage number 06836500).

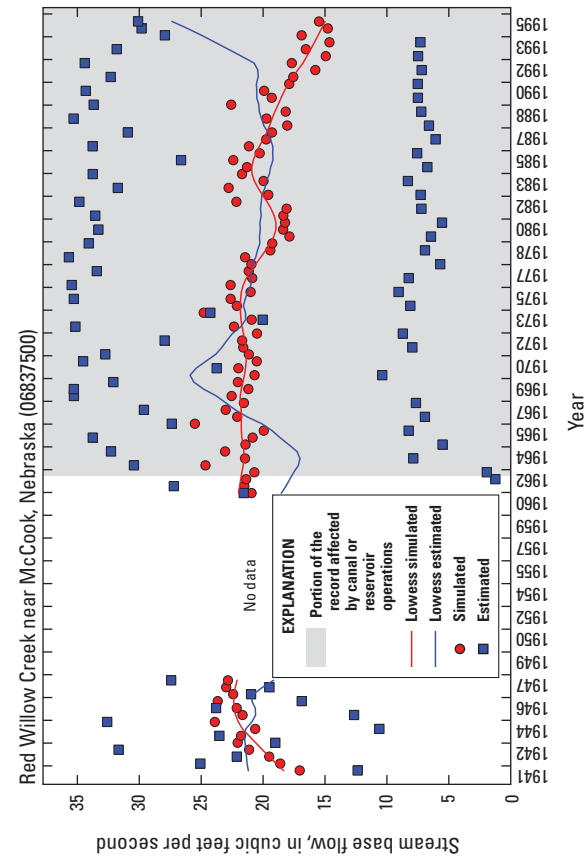


Figure 1-63. Simulated and estimated base flow, 1940–95, for Red Willow Creek near McCook, Nebr. (streamgage number 06837500).

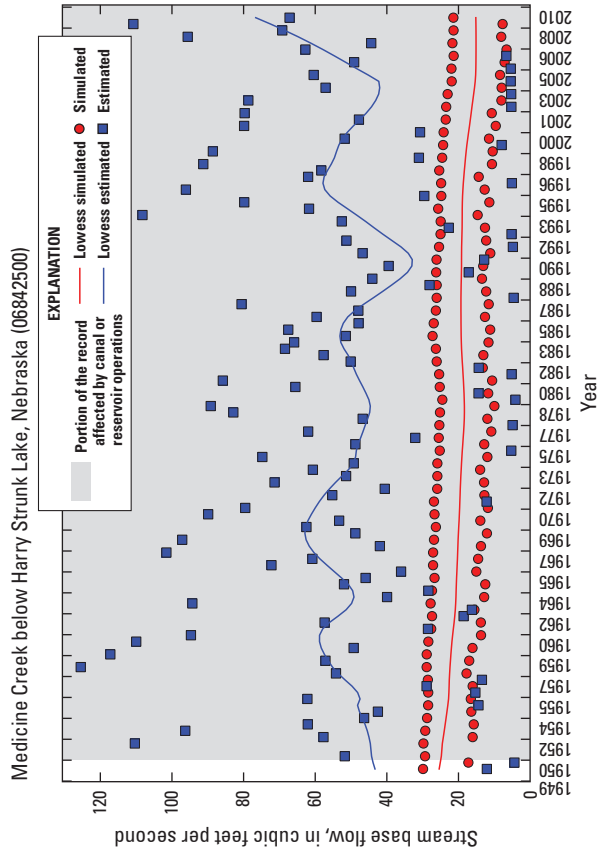


Figure 1-66. Simulated and estimated base flow, 1950–2009, for Medicine Creek below Harry Strunk Lake, Nebr. (streamgage number 06842500).

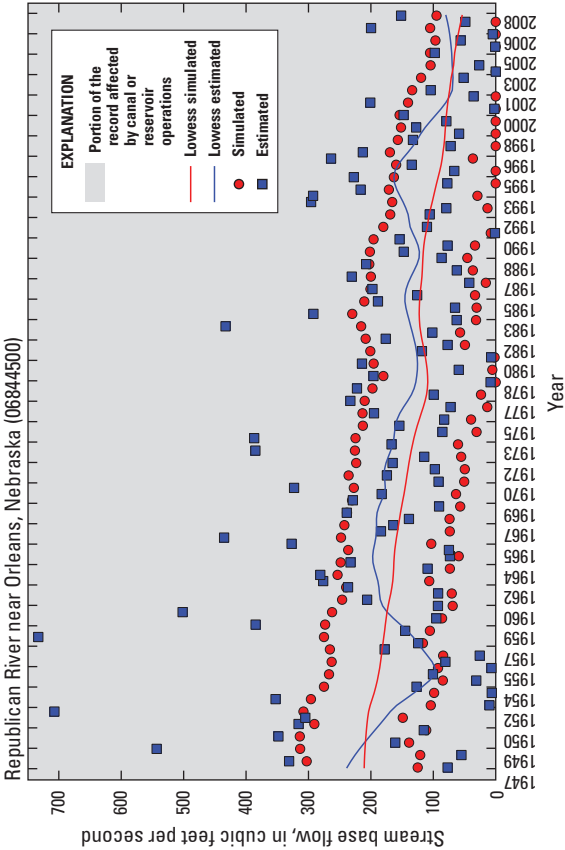


Figure 1-68. Simulated and estimated base flow, 1947–2008, for Republican River near Orleans, Nebr. (streamgage number 06844500).

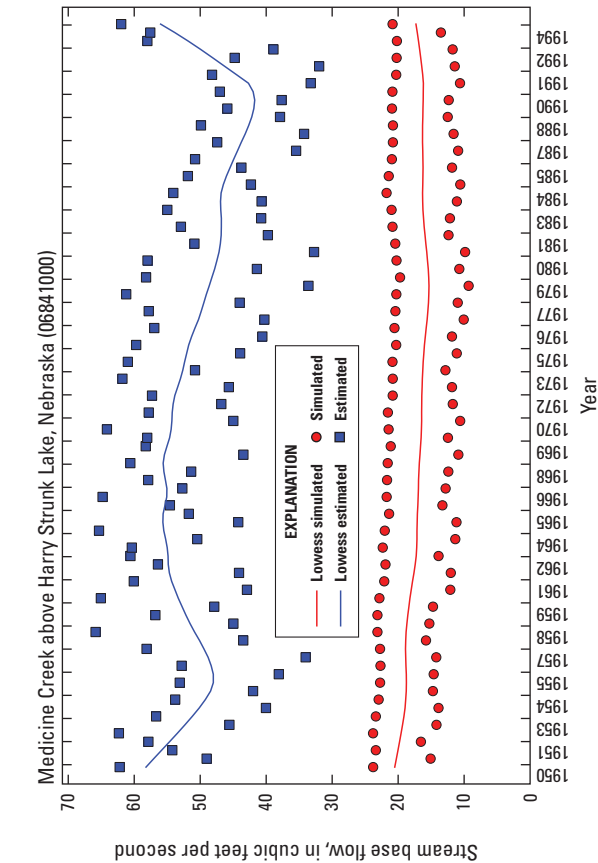


Figure 1-65. Simulated and estimated base flow, 1950–94, for Medicine Creek above Harry Strunk Lake, Nebr. (streamgage number 06841000).

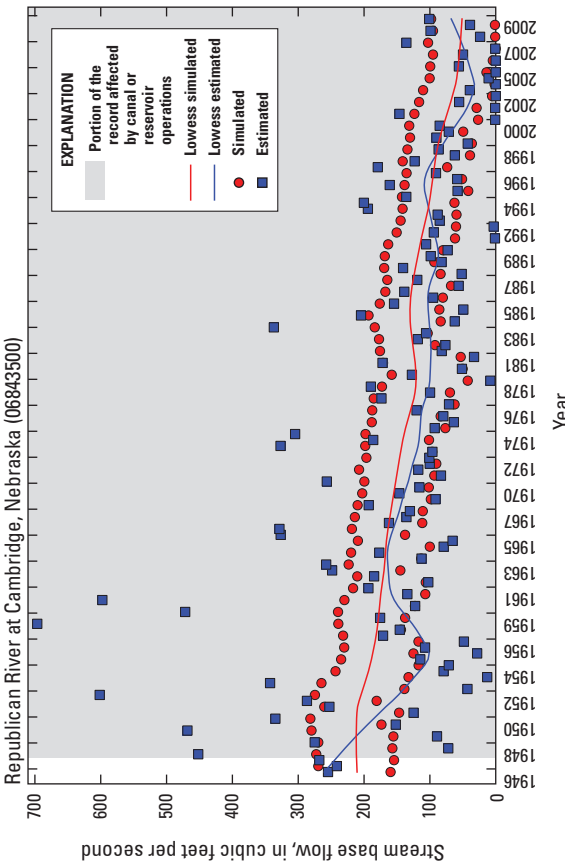


Figure 1-67. Simulated and estimated base flow, 1945–2009, for Republican River at Cambridge, Nebr. (streamgage number 06843500).

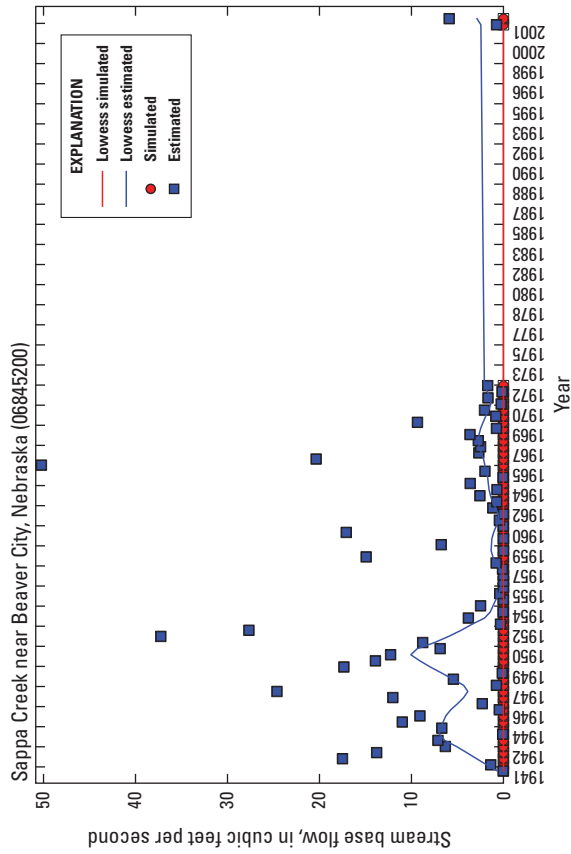


Figure 1-69. Simulated and estimated base flow, 1959–2009, for South Fork Sappa Creek near Achilles, Kans. (streamage number 06844900).

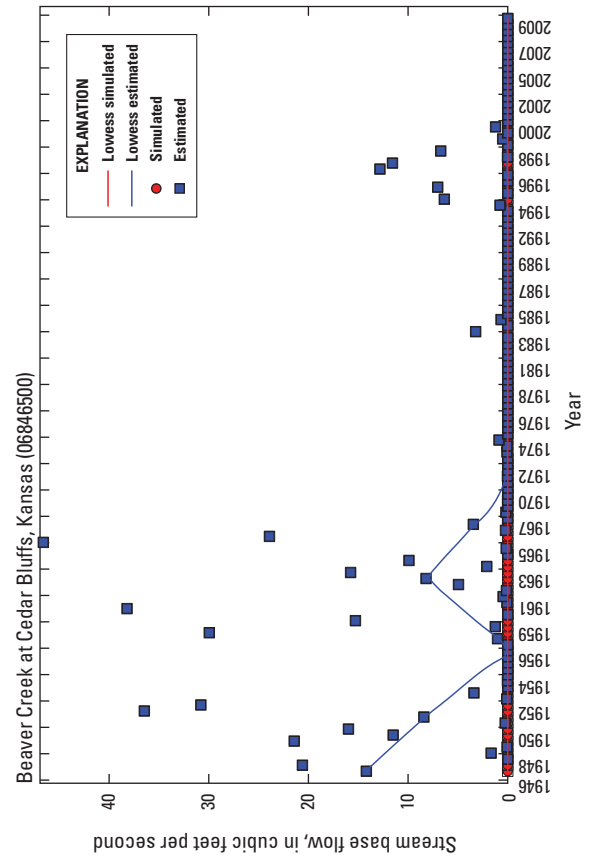


Figure 1-71. Simulated and estimated base flow, 1946–2009, for Beaver Creek at Cedar Bluffs, Kans. (streamage number 06846500).

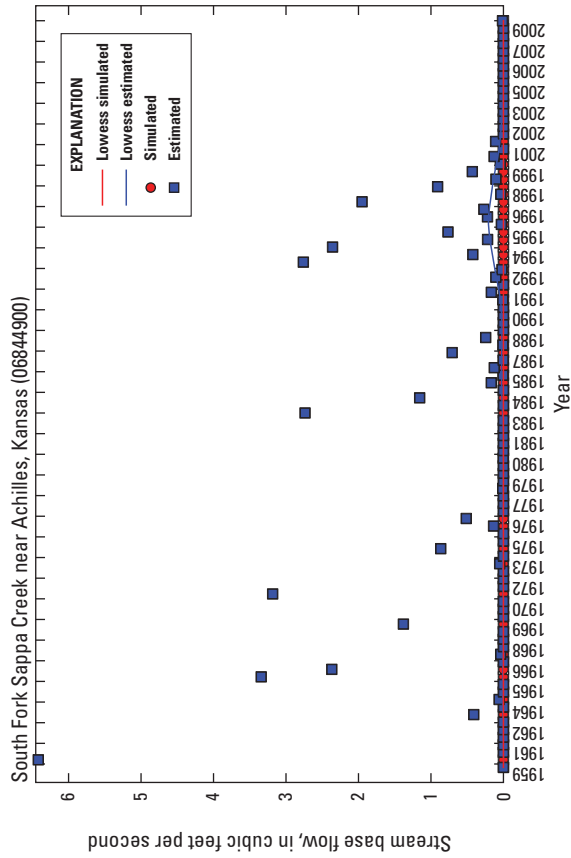


Figure 1-69. Simulated and estimated base flow, 1959–2009, for South Fork Sappa Creek near Achilles, Kans. (streamage number 06844900).

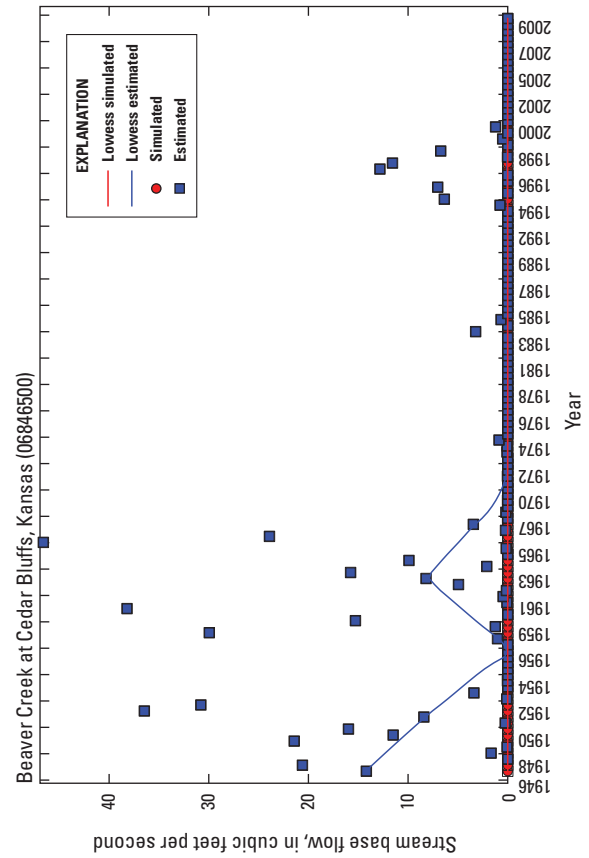


Figure 1-71. Simulated and estimated base flow, 1946–2009, for Beaver Creek at Cedar Bluffs, Kans. (streamage number 06846500).

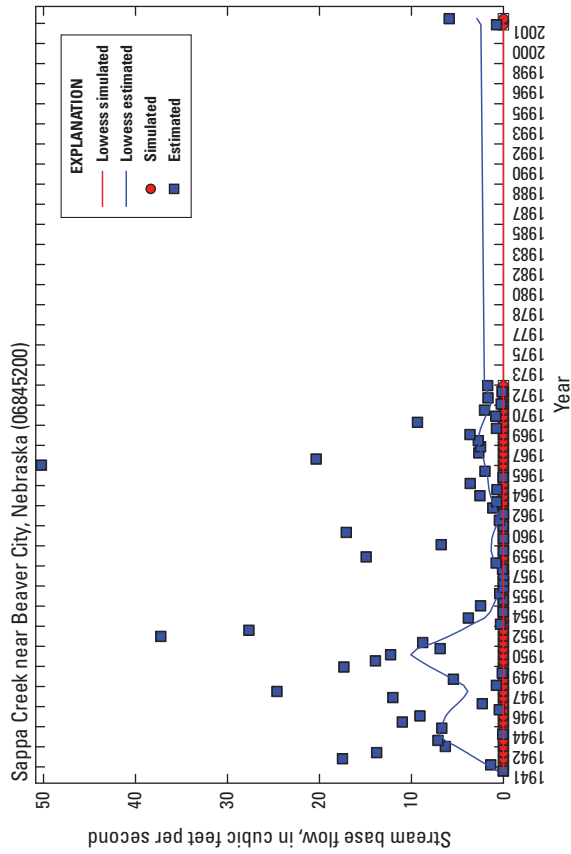


Figure 1-70. Simulated and estimated base flow, 1940–2002, for Sappa Creek near Beaver City, Nebr. (streamage number 06845200).

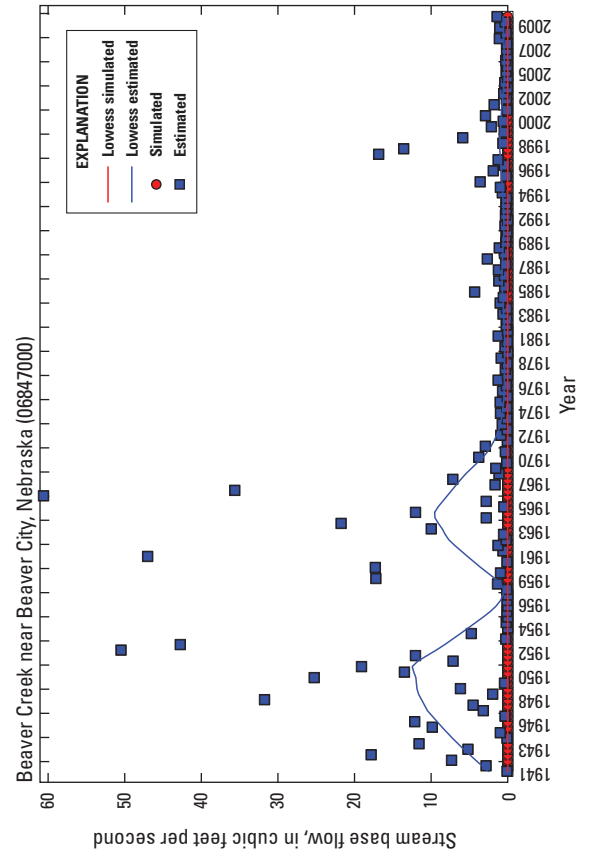


Figure 1-72. Simulated and estimated base flow, 1940–2009, for Beaver Creek near Beaver City, Nebr. (streamage number 06847000).

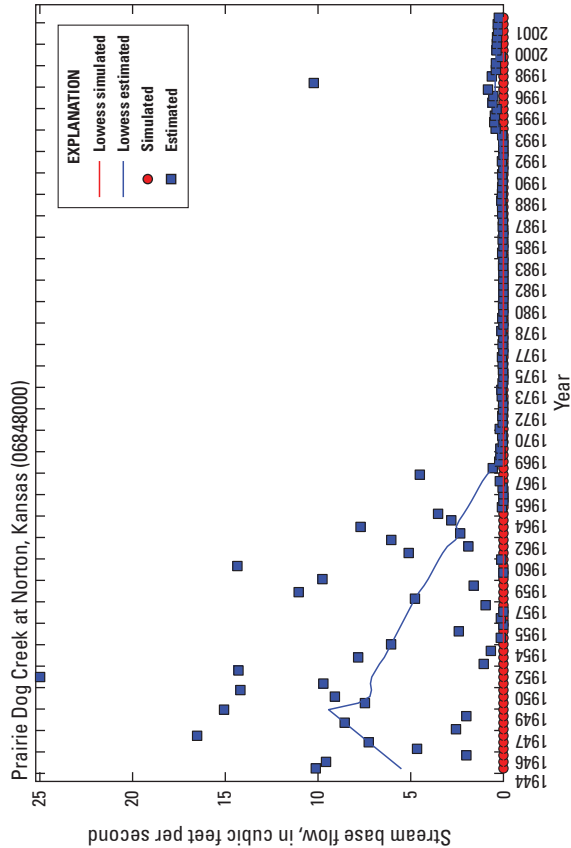


Figure 1-74. Simulated and estimated base flow, 1944–2002, for Prairie Dog Creek at Norton, Kans. (streamage number 06848000).

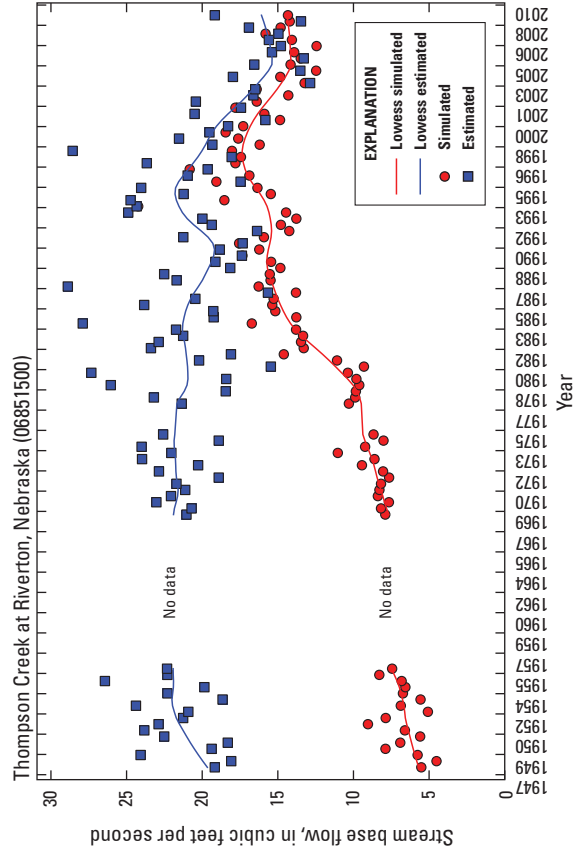


Figure 1-76. Simulated and estimated base flow, 1948–2009, for Thompson Creek at Riverton, Nebr. (streamage number 06851500).

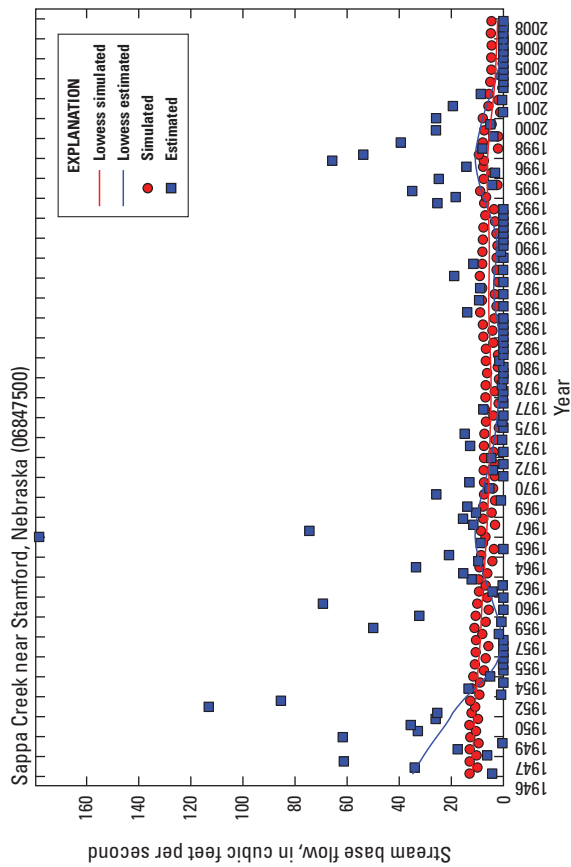


Figure 1-73. Simulated and estimated base flow, 1946–2008, for Sappa Creek near Stamford, Nebr. (streamage number 06847500).

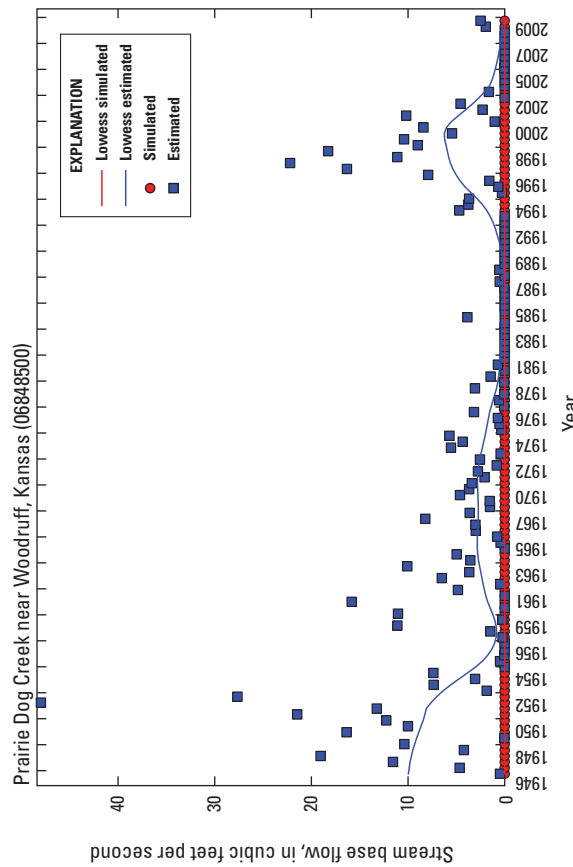


Figure 1-75. Simulated and estimated base flow, 1945–2009, for Prairie Dog Creek near Woodruff, Kans. (streamage number 06848500).

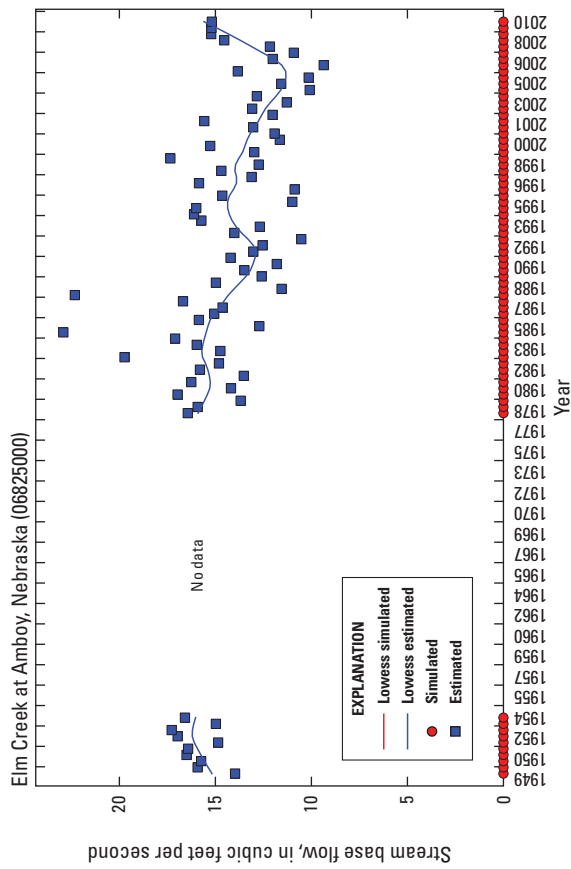


Figure 1-77. Simulated and estimated base flow, 1948–2009, for Elm Creek at Amboy, Nebr. (streamgage number 06825000).

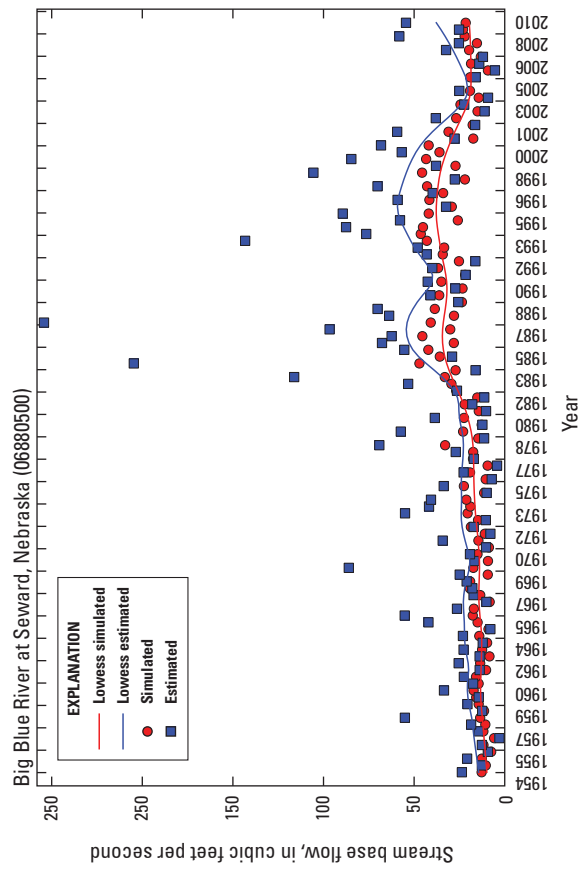


Figure 1-79. Simulated and estimated base flow, 1954–2009, for Big Blue River at Seward, Nebr. (streamgage number 06880500).

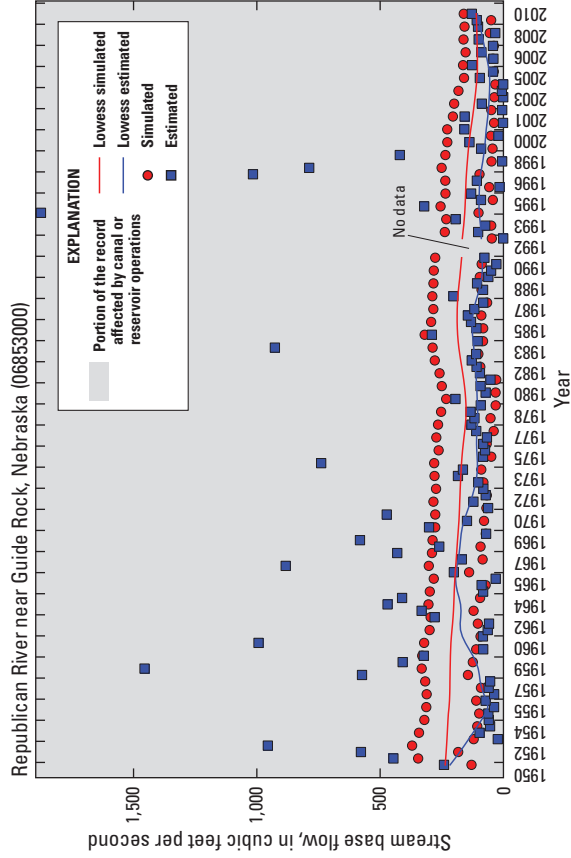


Figure 1-78. Simulated and estimated base flow, 1950–2009, for Republican River near Guide Rock, Nebr. (streamgage number 06853000).

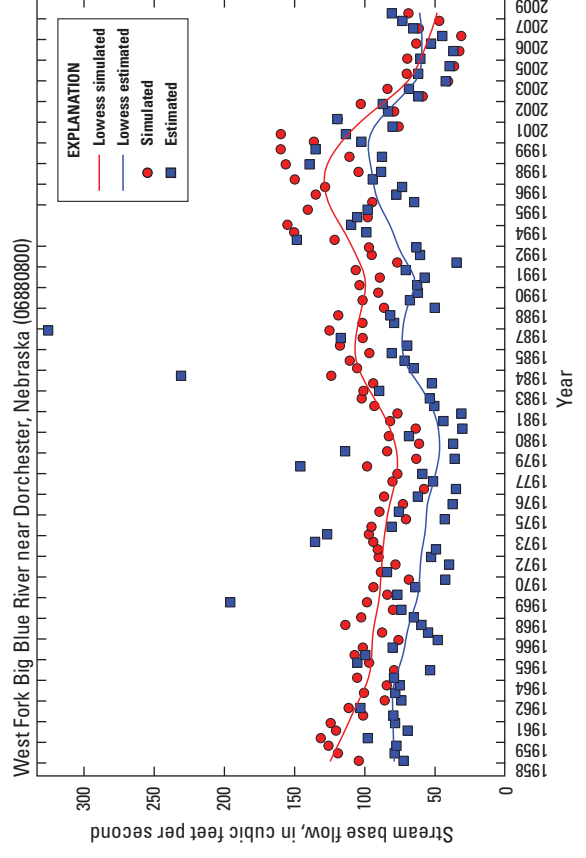


Figure 1-80. Simulated and estimated base flow, 1958–2008, for West Fork Big Blue River near Dorchester, Nebr. (streamgage number 06880800).

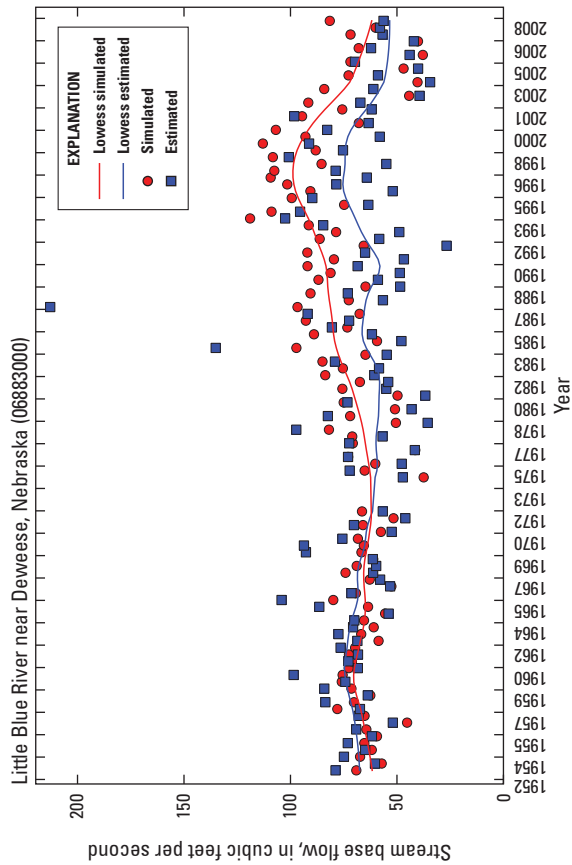


Figure 1-81. Simulated and estimated base flow, 1953–2008, for Little Blue River near Deweese, Nebr. (streamgage number 06883000).

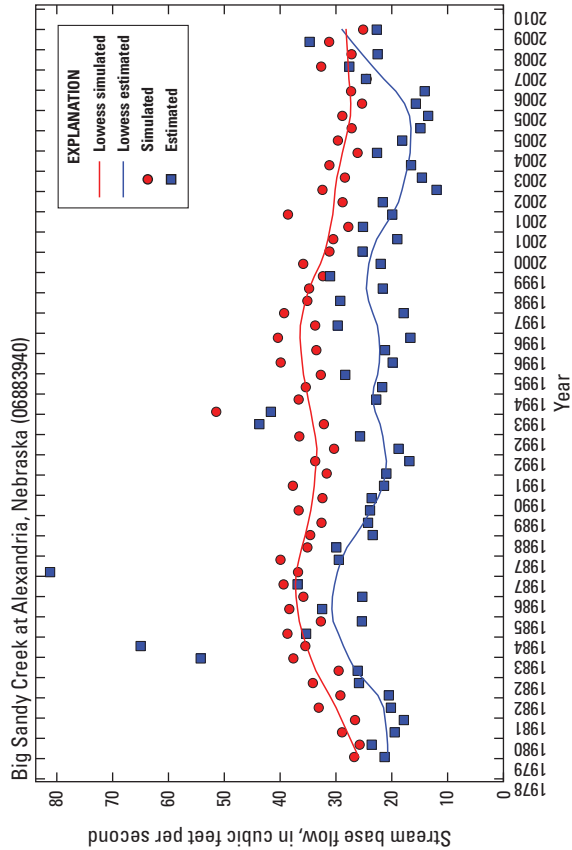


Figure 1-82. Simulated and estimated base flow, 1979–2009, for Big Sandy Creek at Alexandria, Nebr. (streamgage number 06883940).

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