

Prepared in cooperation with the city of Rapid City

Numerical Groundwater-Flow Model of the Minnelusa and Madison Hydrogeologic Units in the Rapid City Area, South Dakota



Scientific Investigations Report 2009–5205

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

Front cover. Swallow hole in Madison Limestone in Spring Creek Canyon (photograph by Ingrid E. Arlton).

Back cover. Doty Spring discharging from the Madison Limestone to Boxelder Creek (photograph by Ingrid E. Arlton).

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By Larry D. Putnam and Andrew J. Long

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

U.S. Department of the Interior

KEN SALAZAR, Secretary

U.S. Geological Survey

Suzette M. Kimball, Acting Director

U.S. Geological Survey, Reston, Virginia: 2009

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Conversion Factors and Datums

Multiply	By	To obtain
Length		
inch (in.)	2.54	centimeter (cm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
Area		
square mile (mi ²)	259.0	hectare (ha)
square mile (mi ²)	2.590	square kilometer (km ²)
Volume		
gallon (gal)	3.785	liter (L)
gallon (gal)	0.003785	cubic meter (m ³)
cubic foot (ft ³)	0.02832	cubic meter (m ³)
Flow rate		
cubic foot per second (ft ³ /s)	0.02832	cubic meter per second (m ³ /s)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
Hydraulic conductivity		
foot per day (ft/d)	0.3048	meter per day (m/d)
Conductance		
feet squared per day (ft ² /d)	0.0929	Meter squared per day (m ² /d)

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 the National Geodetic Vertical Datum of 1929 (NGVD 29).

Horizontal coordinate information is referenced to the North American Datum of 1927 (NAD 27).

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

Water year (WY) is the 12-month period, October 1 through September 30, and is designated by the calendar year in which it ends.

Abbreviations and Acronyms

SDDENR	South Dakota Department of Environment and Natural Resources
SDME	standard deviation of measurement error
USGS	U.S. Geological Survey
WY	water year

Numerical Groundwater-Flow Model of the Minnelusa and Madison Hydrogeologic Units in the Rapid City Area, South Dakota

By Larry D. Putnam and Andrew J. Long

Abstract

The city of Rapid City and other water users in the Rapid City area obtain water supplies from the Minnelusa and Madison aquifers, which are contained in the Minnelusa and Madison hydrogeologic units. A numerical groundwater-flow model of the Minnelusa and Madison hydrogeologic units in the Rapid City area was developed to synthesize estimates of water-budget components and hydraulic properties, and to provide a tool to analyze the effect of additional stress on water-level altitudes within the aquifers and on discharge to springs. This report, prepared in cooperation with the city of Rapid City, documents a numerical groundwater-flow model of the Minnelusa and Madison hydrogeologic units for the 1,000-square-mile study area that includes Rapid City and the surrounding area.

Water-table conditions generally exist in outcrop areas of the Minnelusa and Madison hydrogeologic units, which form generally concentric rings that surround the Precambrian core of the uplifted Black Hills. Confined conditions exist east of the water-table areas in the study area.

The Minnelusa hydrogeologic unit is 375 to 800 feet (ft) thick in the study area with the more permeable upper part containing predominantly sandstone and the less permeable lower part containing more shale and limestone than the upper part. Shale units in the lower part generally impede flow between the Minnelusa hydrogeologic unit and the underlying Madison hydrogeologic unit; however, fracturing and weathering may result in hydraulic connections in some areas. The Madison hydrogeologic unit is composed of limestone and dolomite that is about 250 to 610 ft thick in the study area, and the upper part contains substantial secondary permeability from solution openings and fractures. Recharge to the Minnelusa and Madison hydrogeologic units is from streamflow loss where streams cross the outcrop and from infiltration of precipitation on the outcrops (areal recharge).

MODFLOW-2000, a finite-difference groundwater-flow model, was used to simulate flow in the Minnelusa and Madison hydrogeologic units with five layers. Layer 1 represented the fractured sandstone layers in the upper 250 ft of the Minnelusa hydrogeologic unit, and layer 2 represented

the lower part of the Minnelusa hydrogeologic unit. Layer 3 represented the upper 150 ft of the Madison hydrogeologic unit, and layer 4 represented the less permeable lower part. Layer 5 represented an approximation of the underlying Deadwood aquifer to simulate upward flow to the Madison hydrogeologic unit. The finite-difference grid, oriented 23 degrees counterclockwise, included 221 rows and 169 columns with a square cell size of 492.1 ft in the detailed study area that surrounded Rapid City. The northern and southern boundaries for layers 1–4 were represented as no-flow boundaries, and the boundary on the east was represented with head-dependent flow cells. Streamflow recharge was represented with specified-flow cells, and areal recharge to layers 1–4 was represented with a specified-flux boundary. Calibration of the model was accomplished by two simulations: (1) steady-state simulation of average conditions for water years 1988–97 and (2) transient simulations of water years 1988–97 divided into twenty 6-month stress periods.

Flow-system components represented in the model include recharge, discharge, and hydraulic properties. The steady-state streamflow recharge rate was 42.2 cubic feet per second (ft^3/s), and transient streamflow recharge rates ranged from 14.1 to 102.2 ft^3/s . The steady-state areal recharge rate was 20.9 ft^3/s , and transient areal recharge rates ranged from 1.1 to 98.4 ft^3/s . The upward flow rate from the Deadwood aquifer to the Madison hydrogeologic unit was 6.3 ft^3/s . Discharge included springflow, water use, flow to overlying units, and regional outflow. The estimated steady-state springflow of 32.8 ft^3/s from seven springs was similar to the simulated springflow of 31.6 ft^3/s , which included 20.5 ft^3/s from Jackson-Cleghorn Springs. Simulated transient springflow ranged from 25.7 to 42.3 ft^3/s . Steady-state water-use rates for the Minnelusa and Madison hydrogeologic units were 3.4 and 6.7 ft^3/s , respectively. Total transient water-use rates ranged from 3.4 to 19.1 ft^3/s . Flow from the Minnelusa hydrogeologic unit to overlying units was 2.0 ft^3/s . Steady-state and transient regional outflows from the Minnelusa and Madison hydrogeologic units were 12.9 and 12.8 ft^3/s , respectively.

Linear regression of the 252 simulated and observed hydraulic head values for the steady-state simulation had a coefficient of determination (R^2 value) of 0.92 with an average

absolute difference of 37.6 ft. For the transient simulation, the average absolute difference between simulated and observed hydraulic head values for 19 observation wells ranged from 3.5 to 65.1 ft with a median value of 18.3 ft.

Calibrated horizontal hydraulic conductivity values for model layer 1 ranged from 1.0 to 5.2 feet per day (ft/d). Horizontal hydraulic conductivity values for model layer 3 ranged from 0.1 to 388.8 ft/d. Horizontal hydraulic conductivity for layers 2 and 4 were 10 percent of hydraulic conductivity values for layers 1 and layer 3, respectively, except near the outcrop where it was 50 percent of the values for layers 1 and 3, respectively. Vertical hydraulic conductivity for layers 1, 3, 4, and 5 was 10 percent of the respective horizontal hydraulic conductivity for those layers. Vertical hydraulic conductivity for layer 2 ranged from 0.000001 to 0.25 ft/d. Conductance for head-dependent cells representing springs ranged from 3,000 to 86,400 feet squared per day.

Simulation of increased hypothetical pumping of more than about 10 ft³/s may require modification of the boundaries to allow flow into the model. The model is limited by simplifying assumptions necessary to represent material having secondary porosity as a porous media. With additional data, further refinement of the model would be possible, which could improve the accuracy of model estimates of the effects of additional stresses on the system, such as increased withdrawals or drought. The model can yield simulations of future conditions, which can guide management decisions and planning. The model provides a useful tool for general characterization of the effects of stresses and management alternatives on a regional basis.

Introduction

The city of Rapid City, South Dakota (fig. 1), obtains more than one-half of its municipal water supplies from the Minnelusa and Madison aquifers through deep wells and springs, predominantly from the Madison aquifer. Numerous additional users in the Rapid City area obtain water from these aquifers for domestic, commercial, industrial, and irrigation usage. Groundwater flow within the Minnelusa and Madison aquifers is complex with extensive secondary porosity from fracturing, solution enhancement, and brecciation. The Minnelusa and Madison aquifers are contained within the Minnelusa and Madison hydrogeologic units, which include layers with less permeability.

The U.S. Geological Survey (USGS), as part of a long-term cooperative study with the city of Rapid City, has compiled numerous datasets designed to better understand groundwater flow in the Minnelusa and Madison hydrogeologic units. A numerical groundwater-flow model of the Minnelusa and Madison hydrogeologic units in the Rapid City area was developed to synthesize estimates of water-budget components and hydraulic properties, and to provide a tool to

analyze the effect of additional stress on water-level altitudes within the aquifers and on discharge to springs.

Purpose and Scope

The purposes of this report are to (1) document the development of a numerical groundwater-flow model of the Minnelusa and Madison hydrogeologic units, which contain the Minnelusa and Madison aquifers, in the Rapid City area in South Dakota, and (2) present simulated responses to stress and describe model limitations. The report describes the calibrated numerical groundwater-flow model including estimates of recharge, discharge, and hydraulic properties that characterize the Minnelusa and Madison hydrogeologic units.

Previous Investigations

This numerical modeling effort utilized datasets compiled in a report by Long and Putnam (2002) that documents a conceptual model of the Madison and Minnelusa aquifers. The concepts and datasets from that report were used extensively in constructing the numerical groundwater-flow model. Detailed description of the methods and interpretations that were used in compiling these datasets is available in Long and Putnam (2002). The conceptual-model report by Long and Putnam (2002) describes previous investigations pertinent to this report. Previous investigations include several publications from the Black Hills hydrology study (Hortness and Driscoll, 1998; Carter and Redden, 1999a, 1999b, 1999c; Strobel and others, 1999; Driscoll, Bradford, and Moran, 2000; Driscoll, Hamade, and Kenner, 2000; Carter and others, 2001) and studies specific to the Rapid City area (Greene, 1993, 1999; Anderson and others, 1999; Long, 2000; Long and Putnam, 2002).

Recent studies of the Minnelusa and Madison aquifers include linear modeling of three components of flow in karst aquifers by using oxygen isotopes (Long and Putnam, 2004). Hargrave (2005) described the vulnerability of the Minnelusa aquifer to contamination. Miller (2005) described the influences of geologic structures and stratigraphy on groundwater flow in the karstic Madison hydrogeologic unit in the study area. Putnam and Long (2007) characterized karst groundwater flow in the Rapid City area by using fluorescent dyes. Long and others (2008) described the use of age-determining tracers in conjunction with other tracers to characterize groundwater flow paths in the Madison aquifer.

Acknowledgments

The authors acknowledge the extensive support from the city of Rapid City for numerous studies from which datasets used in this modeling effort have been developed. West Dakota Water Development District and the South Dakota School of Mines and Technology participated in numerous

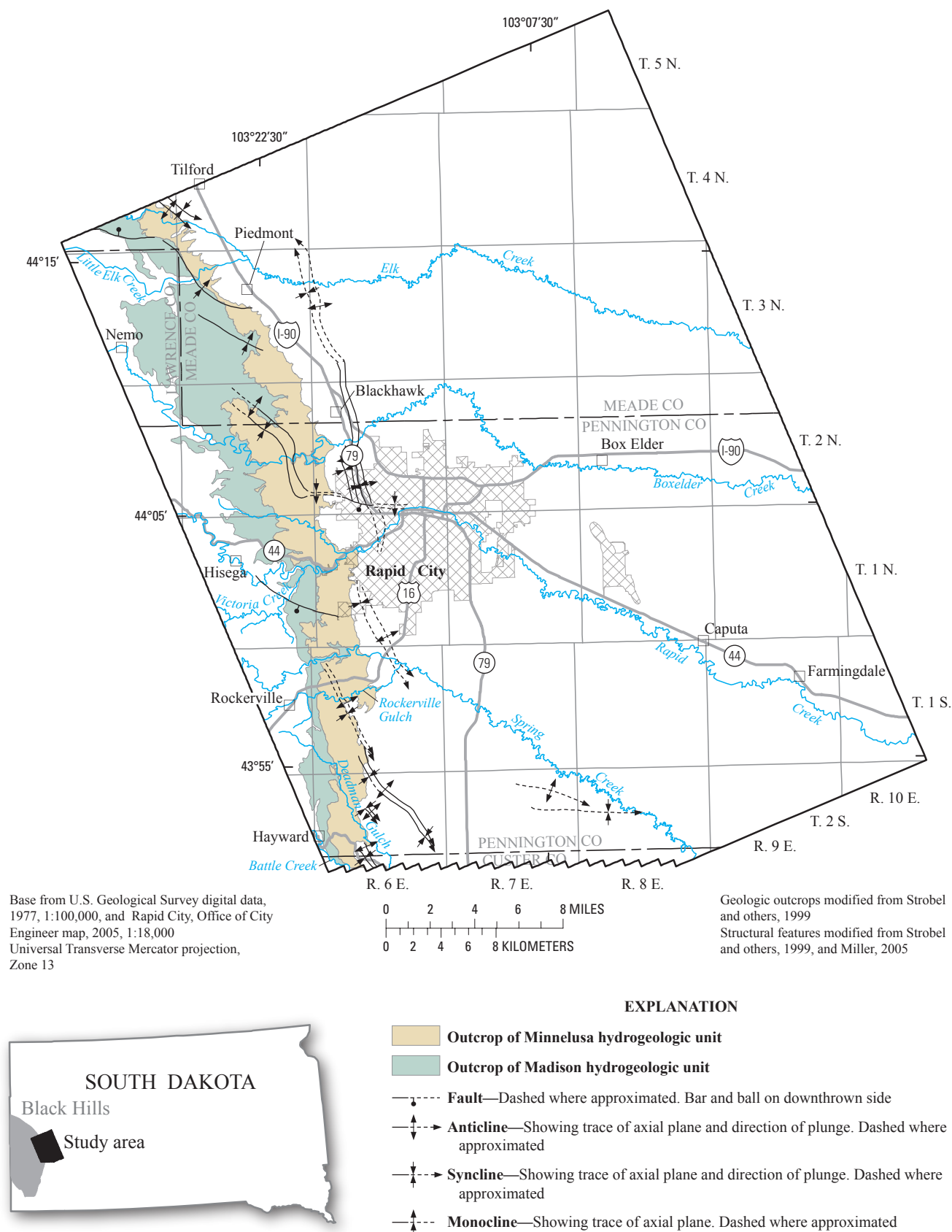


Figure 1. Location of study area.

studies to better understand the complex hydrogeology in the Rapid City area. The Water Rights Program of the South Dakota Department of Environment and Natural Resources (SDDENR) provided compilations of water-level data for numerous observation wells they operate in the study area. The SDDENR also provided water-use data.

Description of Study Area

The approximately 1,000-square-mile (mi²) study area (fig. 1) on the eastern flank of the Black Hills includes Rapid City and the surrounding area. The study area extends from Elk Creek on the north to Battle Creek on the south, and from the outcrop of the Madison hydrogeologic unit on the west to Farmingdale and Box Elder on the east (fig. 1). The population of Rapid City in the 2000 census was 59,607 (U.S. Census Bureau, 2000). The area around Rapid City includes numerous suburban subdivisions and smaller towns. The study area includes parts of Lawrence, Meade, Pennington, and Custer Counties. The Minnelusa and Madison aquifers are important sources of water for many of the smaller communities and subdivisions in these counties.

Physiography and Climate

Land-surface altitudes range from more than 5,000 feet (ft) above the National Geodetic Vertical Datum of 1929 on the western side of the study area to about 2,800 ft near Farmingdale. The western extent of the study area is characterized by high relief covered predominantly by pine and spruce forests. The eastern lowlands are characterized by rolling prairies with bottom lands along stream channels. The Minnelusa and Madison hydrogeologic units crop out in the western part of study area along the eastern flank of the Black Hills uplift. These outcrops are characterized by high-relief forested areas cut by deep canyons with entrenched meanders and steep cliffs formed by resistant limestone and sandstone.

Average (water years 1961–90) precipitation rates range from about 24 inches per year (in/yr) in the northwest to about 16 in/yr in the eastern lowlands with most precipitation occurring in March, April, May, and June (Driscoll, Hamade, and Kenner, 2000). The average (1960–90) temperature at Rapid City is about 22 degrees Fahrenheit in January and about 72 degrees Fahrenheit in July (National Oceanic and Atmospheric Administration, 1996).

Hydrogeologic Setting

Uplift at the end of the Cretaceous period followed by erosion has created the dome-like structure and geomorphology of the Black Hills. Metamorphic and igneous rocks of Precambrian age are exposed in the Black Hills' central core, whereas stratigraphic layers (figs. 2 and 3) of Paleozoic age

and younger are exposed on its flanks. The outcrops of Paleozoic units form generally concentric rings surrounding the Precambrian core and dip radially outward.

The Minnelusa and Madison hydrogeologic units dip away from the uplifted Black Hills at angles that can approach or exceed 15 to 20 degrees near the outcrops and decrease with distance from the uplift to less than 1 degree (Carter and Redden, 1999a, 1999b, 1999c). The Minnelusa and Madison aquifers are contained within the Minnelusa and Madison hydrogeologic units, and many studies refer to the aquifers as being the upper part of the hydrogeologic units. The Minnelusa hydrogeologic unit is composed of the Minnelusa Formation. The Madison hydrogeologic unit includes the Madison Limestone and the underlying Englewood Formation because the Englewood Formation is hydrogeologically similar to the lower part of the Madison Limestone. The outcrop areas for the Minnelusa and Madison hydrogeologic units in the study area are about 39 square miles (mi²) and 52 mi², respectively (fig. 1). Anticlines and synclines (fig. 1) result in local variations in the dip of beds near the outcrops.

A conceptual illustration of groundwater-flow components in an aquifer with hydrogeology similar to the Minnelusa and Madison aquifers is shown in figure 4. Infiltrating precipitation or streamflow losses may have an easterly flow component rather than a strictly vertical one because of the greater hydraulic conductivity parallel to easterly dipping bedding planes. The hydraulic head in the recharge area fluctuates with the changing recharge rate and causes a pressure wave to propagate through the confined part of the aquifer. This wave decreases in amplitude with distance traveled because of head losses in the aquifer. For this reason, hydraulic head fluctuations in downgradient locations east of the recharge area generally are less than in the recharge area unless other stresses, such as pumping, are introduced. Springs discharge through breccias pipes, fractures, and enlarged solution openings.

Water-table conditions generally exist in outcrop areas; however, the water table in the updip parts of the outcrops (approximately 1 to 2 miles along western edge) may be perched above the regional water table because of the higher altitudes. Recharge water may be stored under perched conditions before percolating downward to the regional water table. Water perched on discontinuous layers of low permeability material may exist in the Minnelusa Formation, and pools of perched water can be found in Madison Limestone caves. The unconfined area extends beyond the outcrop areas to the east a few hundred feet, where the dip is steepest, to more than 1 mile (mi), where the dip is less steep (Long and Putnam, 2002). Structural features near the outcrops (fig. 1) influence the shape and extent (described in the subsequent "Storage Properties" section) of the unconfined area. The estimated unconfined area is 36.3 mi² for the Minnelusa hydrogeologic unit and 52.9 mi² for the Madison hydrogeologic unit (Long and Putnam, 2002). East of the water-table areas, hydraulic head is above the tops of the hydrogeologic units owing to their easterly dip, and confined conditions exist. In some areas,

ERATHEM	SYSTEM	ABBREVIATION FOR STRATIGRAPHIC INTERVAL		STRATIGRAPHIC UNIT		THICKNESS IN FEET	DESCRIPTION
CENOZOIC	QUATERNARY & TERTIARY (?)	QTu		UNDIFFERENTIATED SANDS AND GRAVELS		0–50	Sand, gravel, and boulders
	TERTIARY	Tw		WHITE RIVER GROUP		0–600	Light colored clays with sandstone channel fillings and local limestone lenses.
		Tui		INTRUSIVE IGNEOUS ROCKS			Includes rhyolite, latite, trachyte, and phonite.
MESOZOIC	CRETACEOUS	Kps	PIERRE SHALE			1,200–2,000	Principal horizon of limestone lenses giving teepee buttes.
							Dark-gray shale containing scattered concretions.
							Widely scattered limestone masses, giving small teepee buttes.
			NIOBRARA FORMATION			100–225	Impure chalk and calcareous shale.
			CARLILE FORMATION Turner Sand Member Wall Creek Sands			400–750	Light-gray shale with numerous large concretions and sandy layers.
							Dark-gray shale.
			GREENHORN FORMATION			25–380	Impure slabby limestone. Weathers buff.
							Dark-gray calcareous shale, with thin Orman Lake limestone at base.
		GRANEROS GROUP	BELLE FOURCHE SHALE			300–550	Gray shale with scattered limestone concretions.
							Clay spur bentonite at base.
	MOWRY SHALE			150–250	Light-gray siliceous shale. Fish scales and thin layers of bentonite.		
	MUDDY SANDSTONE		NEWCASTLE SANDSTONE		20–60	Brown to light-yellow and white sandstone.	
	SKULL CREEK SHALE			170–270	Dark-gray to black siliceous shale.		
	Kik	INVAN KARA GROUP	FALL RIVER FORMATION			10–200	Massive to slabby sandstone.
			LAKOTA FORMATION			35–700	Coarse gray to buff crossbedded conglomeratic sandstone, interbedded with buff, red, and gray clay, especially toward top. Local fine-grained limestone.
			JURASSIC	Ju	MORRISON FORMATION		
UNKPAPA SS					0–225	Massive fine-grained sandstone.	
SUNDANCE FORMATION Redwater Member Lak Member Hulett Member Stockade Beaver Mem. Canyon Spr Member					250–450	Greenish-gray shale, thin limestone lenses.	
GYPSUM SPRING FORMATION					0–45	Red siltstone, gypsum, and limestone.	
TRIASSIC	TPs	SPEARFISH FORMATION Goose Egg Equivalent			250–700	Red sandy shale, soft red sandstone and siltstone with gypsum and thin limestone layers. Gypsum locally near the base.	
PALEOZOIC	PERMIAN	Pmk	MINNEKAHTA LIMESTONE			¹ 25–65	Thin to medium-bedded finely crystalline, purplish-gray laminated limestone.
		Po	OPECHE SHALE			50–135	Red shale and sandstone.
		PIPm	MINNELUSA FORMATION			^{1,2} 375–800	Yellow to red crossbedded sandstone, limestone, and anhydrite locally at top.
	Interbedded sandstone, limestone, dolomite, shale, and anhydrite.						
	Red shale with interbedded limestone and sandstone at base.						
	MISSISSIPPIAN	MDme	MADISON (PAHASAPA) LIMESTONE			² 250–550	Massive light-colored limestone. Dolomite in part. Cavernous in upper part.
							ENGLEWOOD FORMATION
	DEVONIAN	Ou	WHITEWOOD (RED RIVER) FORMATION			0–60	Buff dolomite and limestone.
			WINNIPEG FORMATION			0–100	Green shale with siltstone.
	CAMBRIAN	OCd	DEADWOOD FORMATION			³ 75–400	Massive to thin-bedded buff to purple sandstone. Greenish glauconitic shale, flaggy dolomite, and flat-pebble limestone conglomerate. Sandstone, with conglomerate locally at the base.
PRECAMBRIAN		pCu	UNDIFFERENTIATED METAMORPHIC AND IGNEOUS ROCKS			Unknown	Schist, slate, quartzite, and arkosic grit. Intruded by diorite, metamorphosed to amphibolite, and by granite and pegmatite.

¹ Modified on the basis of drill-hole data.

² Thickness based on structure contours of Minnelusa Formation, Madison Limestone, and Deadwood Formation tops (Carter and Redden, 1999a, 1999b, 1999c).

³ Based on Carter and others, 2001.

Modified from information furnished by the Department of Geology and Geological Engineering, South Dakota School of Mines and Technology (written commun., January 1994)

Figure 2. Stratigraphic section for the study area.

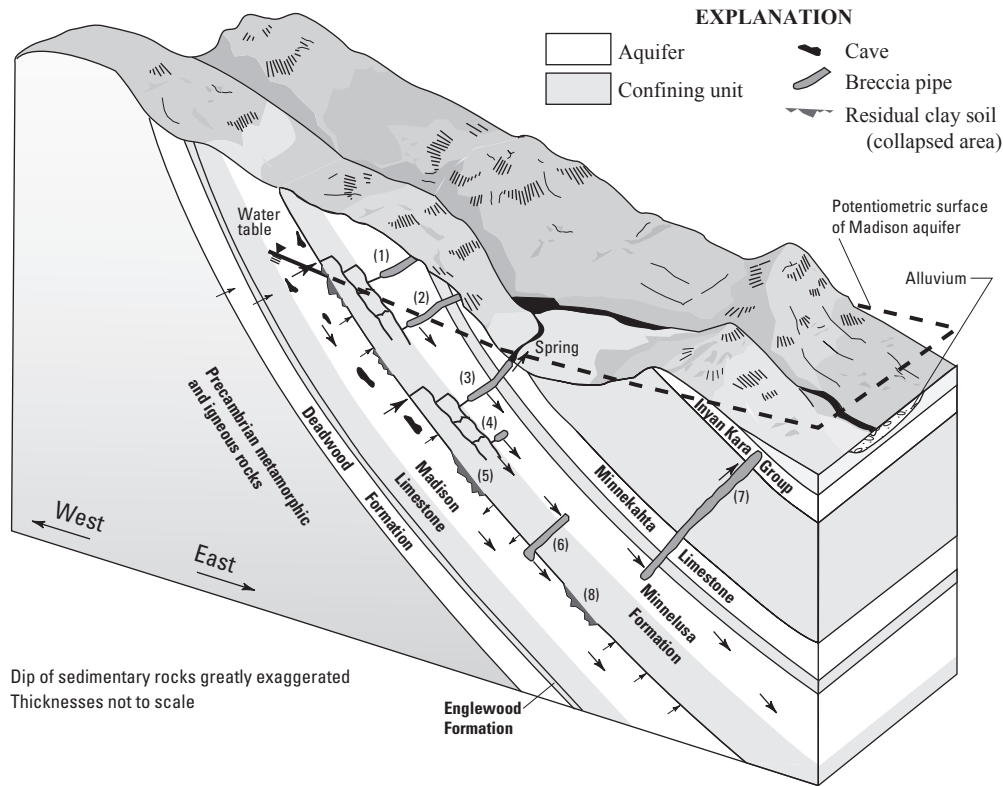


Figure 3. Conceptual hydrogeologic section of the study area (modified from Hayes, 1999). Each aquifer shown is separated from other aquifers by confining units. Hydraulic connection between aquifers is increased by vertical breccia pipes and fractures. The schematic shows (1) exposed breccia pipe above hydraulic head in Madison aquifer, (2) exposed breccia pipe with hydraulic head below land surface, (3) breccia pipe at active spring-discharge point, (4) developing breccia pipe, (5) fractures in confining unit, (6) breccia pipe originating in the Madison Limestone, (7) breccia pipe extending from Minnelusa Formation to the Inyan Kara Group, and (8) discontinuous residual clay soil. Arrows show general areal leakage, focused leakage at breccia pipes, or groundwater-flow directions.

the hydraulic head is above the land surface, and flowing wells may exist in these areas. Hydraulic connection to the overlying Minnekahta Limestone and Inyan Kara Group is limited by intervening shale layers.

Movement of groundwater between the Minnelusa and Madison aquifers is influenced by vertical hydraulic gradients, hydraulic properties of the intervening confining unit, and recharge rates (fig. 5; Long and Putnam, 2002). Although the confining units generally do not transmit water at a high rate, their capacity to store water could have substantial effects on

the hydraulics of the groundwater-flow system. Water that flows between the Minnelusa and Madison aquifers must pass through a confining unit that is several hundred feet thick with variable porosity and potentially a substantial amount of water held in storage. The confining units contain more secondary porosity in the western side of the study area where fracturing from the Black Hills uplift is more prevalent than in the eastern side of the study area. East of Rapid City, the hydraulic head is higher in the Minnelusa aquifer than in the Madison aquifer.

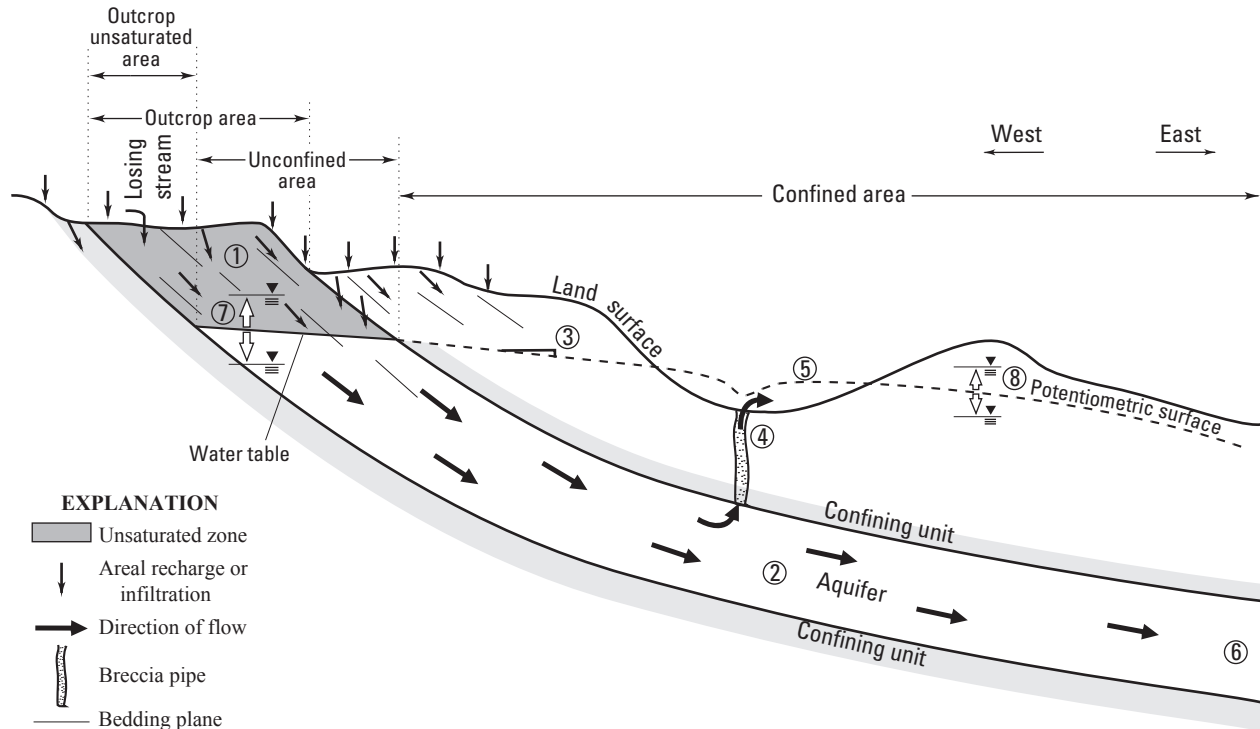


Figure 4. Generalized diagram with vertical exaggeration of a confined aquifer recharged at the updip end (from Long and Putnam, 2002). The diagram shows (1) recharge infiltrates and moves downward vertically or diagonally parallel to bedding planes, (2) near horizontal flow with head losses resulting from resistance from aquifer material, (3) sloping potentiometric surface results from head losses, (4) artesian spring discharges through high-conductivity breccia pipe or fracture because hydraulic head is above the land surface, (5) spring causes depression in the potentiometric surface, (6) outflow rate is controlled by hydraulic gradient and transmissivity, (7) hydraulic head fluctuation at recharge area is controlled by changes in recharge rate, and (8) smaller hydraulic head fluctuation downgradient is in response to larger fluctuation at recharge area.

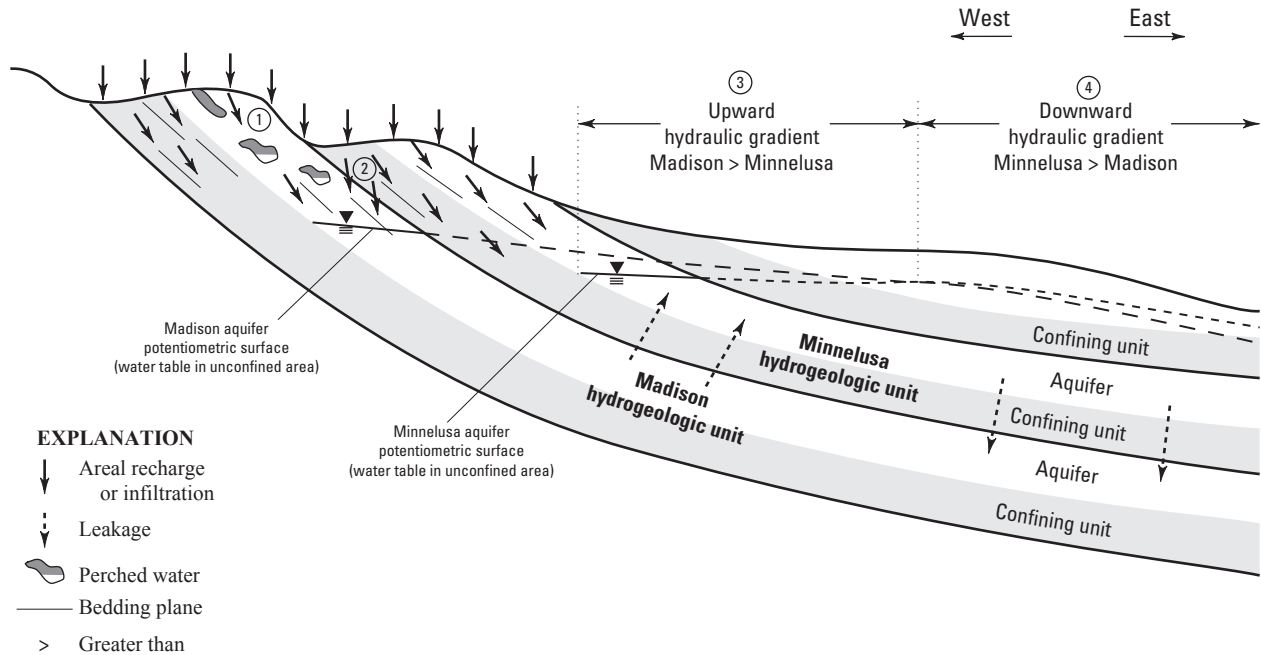


Figure 5. Recharge conditions and vertical gradients in the Minnelusa and Madison hydrogeologic units (from Long and Putnam, 2002). The diagram shows (1) perched water, (2) part of the recharge on the Minnelusa outcrop infiltrates to the Madison aquifer, (3) hydraulic head in Madison aquifer greater than in Minnelusa aquifer creating upward hydraulic gradient, and (4) hydraulic head greater in Minnelusa aquifer than in Madison aquifer creating downward hydraulic gradient.

Minnelusa Hydrogeologic Unit

The Pennsylvanian- and Permian-age Minnelusa Formation, which composes the Minnelusa hydrogeologic unit, is 375 to 800 ft thick (fig. 2) east of its outcrop. Bowles and Braddock (1963) described the upper part of the Minnelusa hydrogeologic unit as thick sandstone with thin limestone, dolomite, and mudstone, and the lower part as having less sandstone and more shale, limestone, and dolomite than the upper part. Siltstone, gypsum, and anhydrite also can be present. The upper 200 to 300 ft of the Minnelusa hydrogeologic unit is more permeable, because of the coarser sandstone, solution openings, breccias, and other karst features, than the lower part of the unit (Peter and others, 1988; Greene, 1993). At the base of the Minnelusa hydrogeologic unit is a red clay shale that ranges in thickness from 0 to 50 ft (Cattermole, 1969; Greene, 1993). This shale, which is discontinuous in the study area, is an ancient residual soil developed on the surface of the Madison Limestone (Gries, 1996). The lower part of the Minnelusa hydrogeologic unit is less permeable than the upper part and generally is an aquitard that impedes flow between the Minnelusa and Madison aquifers (Kyllonen and Peter, 1987; Peter and others, 1988; Greene, 1993). Near outcrop areas, however, the lower part can have greater permeability than in other areas owing to fracturing and weathering in outcrop areas. The hydraulic connection between the Minnelusa and Madison aquifers is highly varied with strong

connections possible in areas with breccia pipes (fig. 3) and structural features; however, over most of the area, the shale in the lower Minnelusa hydrogeologic unit limits the hydraulic connection.

Recharge to the Minnelusa hydrogeologic unit is from streamflow where streams cross the outcrop and from infiltration of precipitation on the outcrops (areal recharge). Most Black Hills streams lose their flow to various sedimentary rocks that are exposed around the periphery of the Black Hills (Hortness and Driscoll, 1998). When streamflow from the central core of the Black Hills is low, most streamflow is lost to the outcrop of the Madison hydrogeologic unit, which is upstream from the outcrop of the Minnelusa hydrogeologic unit.

Wells completed in the Minnelusa hydrogeologic unit in the study area range in depth from 80 ft near the outcrop to about 3,000 ft in the eastern part and are capable of producing from 5 to 700 gallons per minute (gal/min; Long and Putnam, 2002). Wells completed in the Minnelusa hydrogeologic unit are used extensively by small suburban developments and domestic households surrounding Rapid City.

Madison Hydrogeologic Unit

The Madison hydrogeologic unit includes the Mississippian-age Madison Limestone and the Devonian-age Englewood Formation and is 250 to 610 ft thick in the study

area. The Madison Limestone is composed of limestone and dolomite and is 250 to 550 ft thick east of the outcrop of the Madison hydrogeologic unit (fig. 2). The Englewood Formation, which underlies the Madison Limestone, is less than 60 ft thick, is composed of argillaceous, dolomitic limestone, and probably could logically be considered a member of the Madison Limestone because of its lithology (Gries and Martin, 1985).

The upper surface of the Madison Limestone is a weathered karst surface and is unconformably overlain by the Minnelusa Formation (Cattermole, 1969). The upper 150 ft of the Madison hydrogeologic unit contains substantial secondary permeability from solution openings and fractures (Greene, 1993), and solution enlargement of fractures has resulted in a predominance of conduit flow. Secondary permeability in the lower part of the Madison hydrogeologic unit generally is smaller than in the upper part (Greene, 1993); however, the lower unit can have greater permeability near outcrop areas than in the eastern part, especially along stream channels.

The Madison Limestone was divided into four cliff-forming geomorphic units by Miller (2005) through detailed mapping of the canyons in the Spring, Rapid, and Boxelder Creek Basins that are on the west-central edge of the study area (fig. 1). The thickness of the units from bottom to top are 130 to 165 ft for unit 1, 81 to 120 ft for unit 2, 140 to 150 ft for unit 3, and 0 to 85 ft for unit 4. Late Mississippian erosion removed the upper part of unit 4 in the Boxelder and Spring Creek canyons and all of unit 4 and part of unit 3 in the Rapid Creek canyon. Unit 1 is highly resistant to erosion, having thickly bedded sections and forming nearly vertical cliffs. Unit 2 is similar; however, the bedding is thinner and cliff faces tend to be more irregular. Unit 3 is characterized by massive collapse brecciation with large angular blocks and poorly preserved bedding. Caves are numerous in unit 3 and commonly are filled with cemented solution breccias and cave fill. Unit 4, where present, is characterized by collapse brecciation with angular blocks. Upper cliff surfaces in unit 4 are rounded off because of low resistance to erosion. Karst features are located throughout the Madison Limestone; however, they tend to be more common along the contacts between these geomorphic units (Miller, 2005).

Recharge to the Madison hydrogeologic unit is from streamflow loss where streams cross the outcrops and from infiltration of precipitation on the outcrops. The streamflow loss threshold represents the maximum amount of streamflow loss that can occur when water is available in the stream. Loss thresholds for the major streams in the study area (Hortness and Driscoll, 1998) for the Madison hydrogeologic unit range from about 10 to 25 cubic feet per second (ft^3/s). Because the Madison hydrogeologic unit outcrop is the most upstream unit with large loss thresholds, streamflow recharge is substantial and usually greater than areal recharge to the Madison hydrogeologic unit in the study area.

Wells completed in the Madison hydrogeologic unit in the study area are capable of producing from 5 to 2,500 gal/min. About 64 percent of the wells yield 5 to

50 gal/min, 11 percent yield 50 to 200 gal/min, and 25 percent yield 200 to 2,500 gal/min. The depth of wells ranges from 20 to 4,600 ft, with 78 percent of the wells less than 1,000 ft and 41 percent less than 500 ft (Long and Putnam, 2002). The deepest wells are on the eastern side of the study area, where the water level may be only a few hundred feet below the land surface owing to artesian pressure. The varied well yields are influenced by karst features, with the larger production wells connected to substantial solution openings.

Numerical Groundwater-Flow Model

MODFLOW-2000, a numerical, three-dimensional, finite-difference groundwater-flow model (Harbaugh and others, 2000), was used to simulate flow in the hydrogeologic units within the study area. Detailed descriptions of MODFLOW-2000 packages that were used in the model are presented in McDonald and Harbaugh (1988) and Harbaugh and others (2000). The MODFLOW-2000 parameter estimation process (Hill and others, 2000) was used to optimize estimates of hydraulic properties for calibration.

The design of the numerical groundwater-flow model was based on the conceptual model described in Long and Putnam (2002) and includes five model layers (fig. 6). The Minnelusa and Madison hydrogeologic units were each represented by two model layers. Although the entire thickness of these units may include secondary porosity, the more permeable parts of the hydrogeologic units generally occur in the upper part of both units (Long and Putnam, 2002). Layer 1 represents the fractured sandstone layers in the upper part of the Minnelusa hydrogeologic unit. The Minnelusa aquifer generally is contained in the upper 200 to 300 ft of the unit (Greene, 1993). Layer 2 represents the fractured limestone, minor sandstone layers, and shale units in the lower part of the Minnelusa hydrogeologic unit. Layer 3 represents the secondary porosity and karstic features of the Madison hydrogeologic unit that are more prevalent in the upper part than in the lower part. This layer approximates the upper two geomorphic units of the Madison Limestone described by Miller (2005) and described by Greene (1993) as being 100 to 200 ft thick in total. Layer 4 represents the less permeable lower part of the Madison hydrogeologic unit. Layer 5 represents the western part of the Deadwood aquifer to approximate upward flow to the Madison hydrogeologic unit from the underlying Deadwood aquifer. The Whitewood and Winnipeg Formations shown in figure 2 are absent throughout most of the study area (Long and Putnam, 2002).

Arrays representing the altitude of the tops of model layers 1, 3, and 5 were constructed from maps of the structural tops of the Minnelusa Formation, Madison Limestone, and Deadwood Formation (Carter and Redden, 1999a, 1999b, 1999c; Long and Putnam, 2002). A uniform thickness of 250 ft was assumed for layer 1 with the remainder of the Minnelusa hydrogeologic unit represented as layer 2. The upper 250 ft

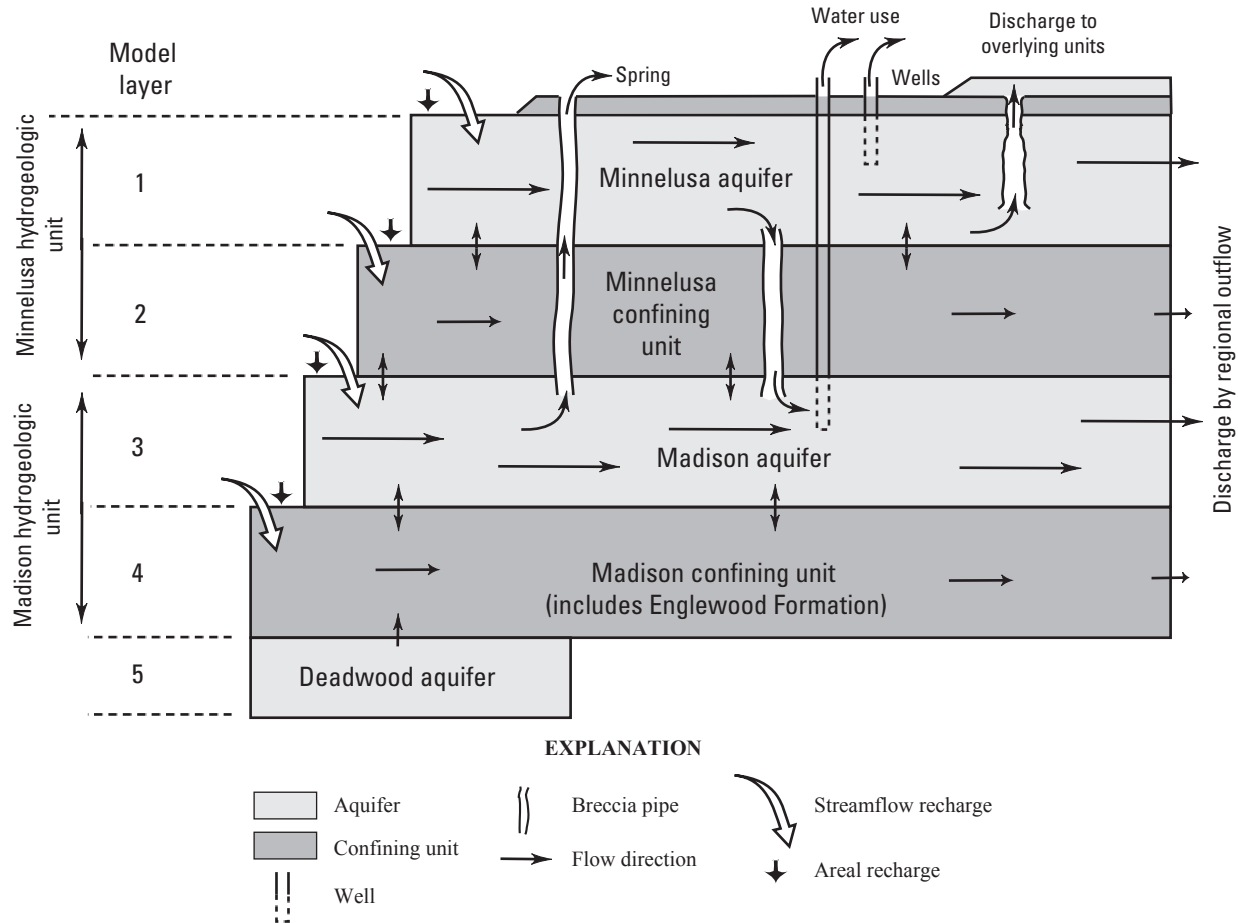


Figure 6. Conceptual schematic of model layers in relation to hydrogeology.

generally contains the sandstone layers. The thickness of layer 2 ranged from 106 to 590 ft with a mean of 293 ft. A uniform thickness of 150 ft was assumed for layer 3 with the remainder of the Madison hydrogeologic unit represented in layer 4. The thickness of layer 4 ranged from 139 to 599 ft with a mean of 341 ft.

Finite-Difference Grid and Boundary Conditions

The finite-difference grid defining the model area consisted of 221 rows and 169 columns (fig. 7). The cell size was 492.1 ft (150 meters) by 492.1 ft in the detailed study area that surrounded Rapid City. Cell sizes increased from the detailed area to 1,640 ft in width at the boundaries on the north and south and to 6,562 ft in width at the boundary on the east (fig. 7). The model was oriented 23 degrees counterclockwise so that the orientation of finite-difference cells approximated the orthogonal patterns in cave passageways and fracture patterns (Greene and Rahn, 1995). This orientation allowed for simulation of horizontal anisotropy if necessary. Miller (2005) mapped a major trend in fractures and joints in the study area

as north 65 degrees east with a less dominant trend of north 35 degrees west. The general pattern of structural features (fig. 1; Strobel and others, 1999) in the study area was about 23 degrees.

The boundary conditions used in the model are described in general terms in this section with quantitative descriptions presented in the subsequent "Representation of Flow System Components" section. The mathematical concepts for various types of cells used to represent boundaries are described in detail in McDonald and Harbaugh (1988), and the specific MODFLOW-2000 packages that were used are included with the model cell descriptions that follow.

Boundaries on the north and south for layers 1–4 were represented as no-flow boundaries (figs. 8 and 9) because the northern and southern boundaries generally occur along flow lines. The boundaries on the east for layers 1–4 were represented with head-dependent flow cells (Drain Package, McDonald and Harbaugh, 1988). With this type of boundary cell, flow across the boundary changes in relation to the hydraulic head in the aquifer. The calculated flow is the difference between a specified head and the head in the model cell multiplied by the area of the cell multiplied by the assigned

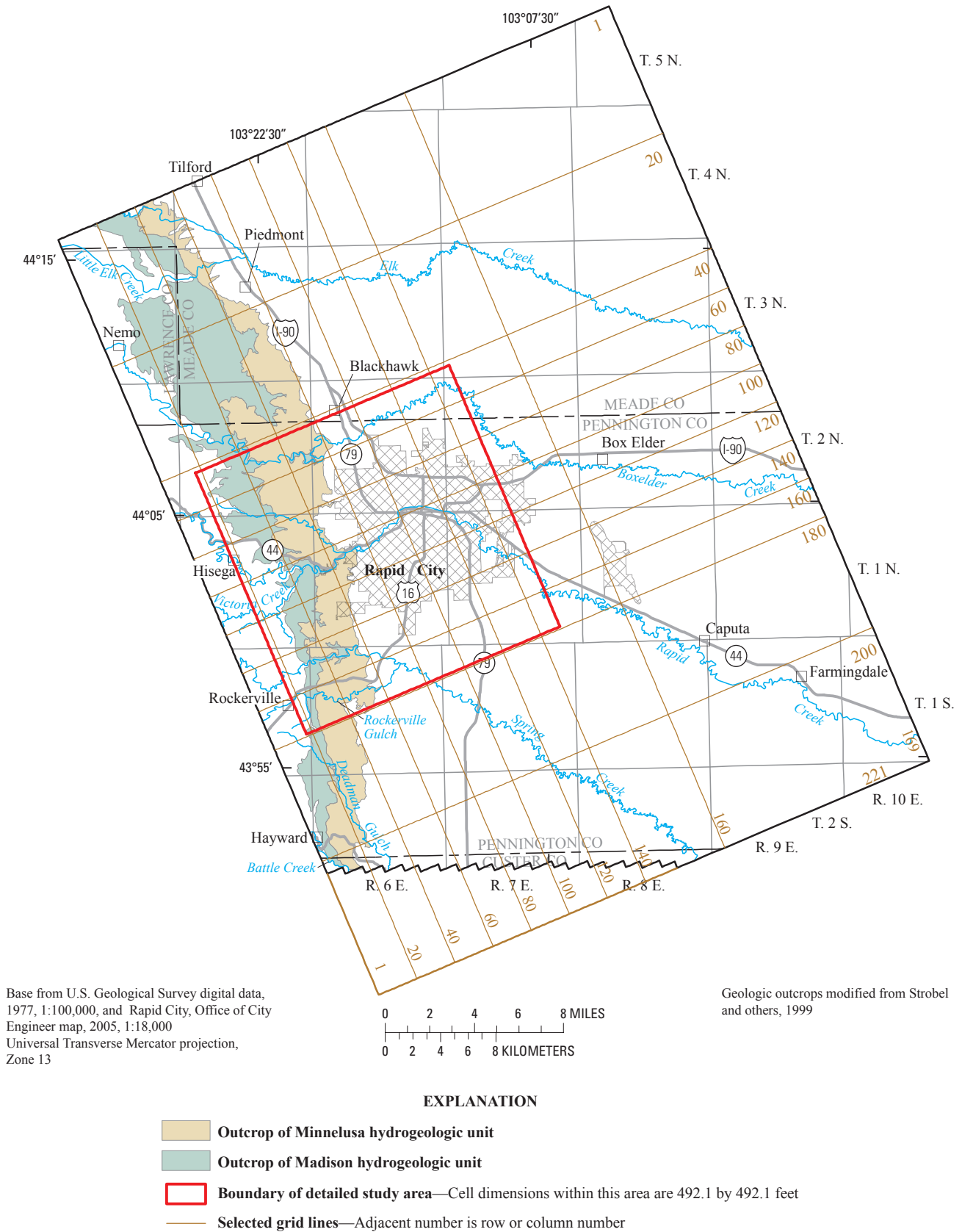


Figure 7. Location of finite-difference grid.

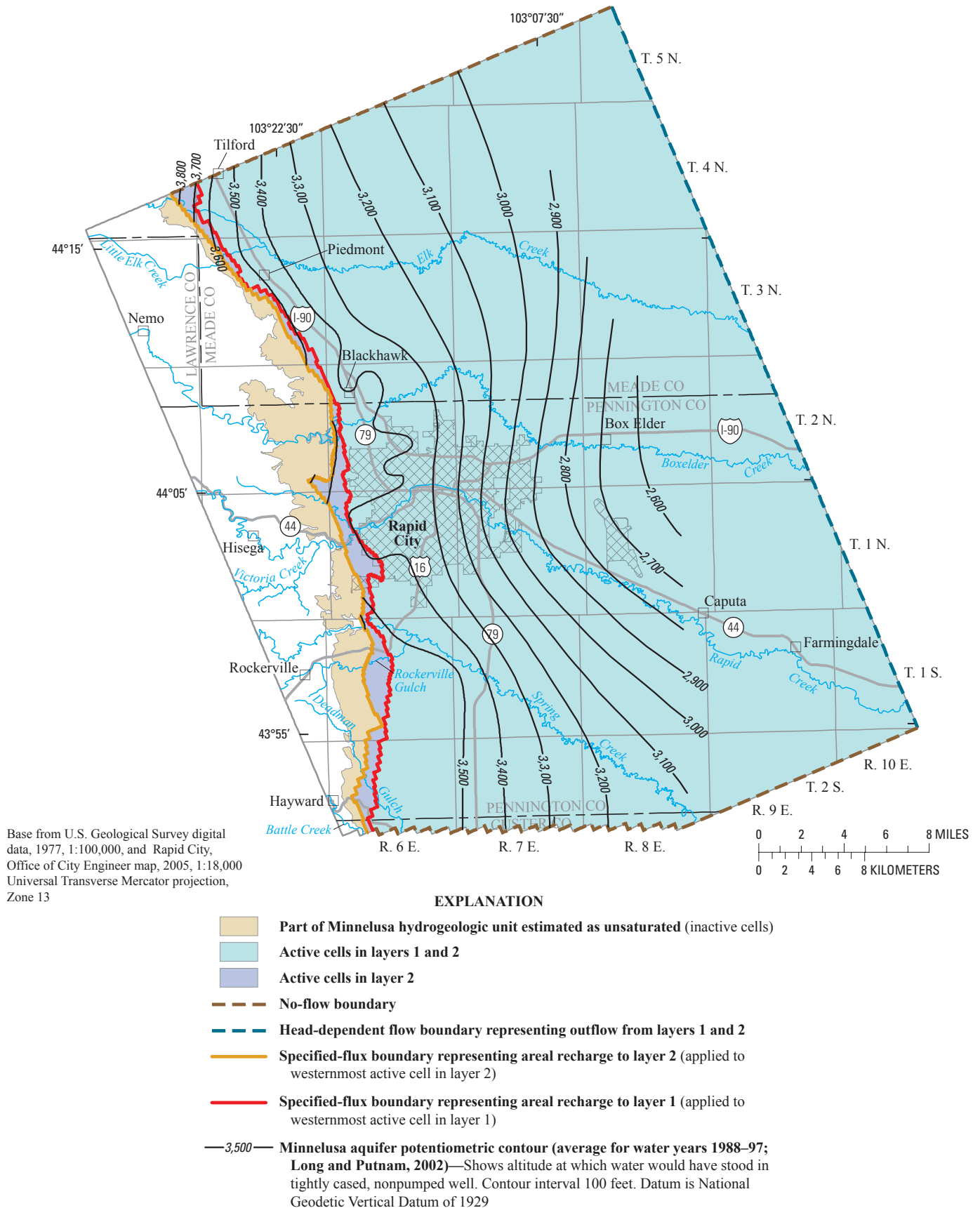


Figure 8. Active model cells and boundary conditions for layers 1 and 2 (Minnelusa hydrogeologic unit).

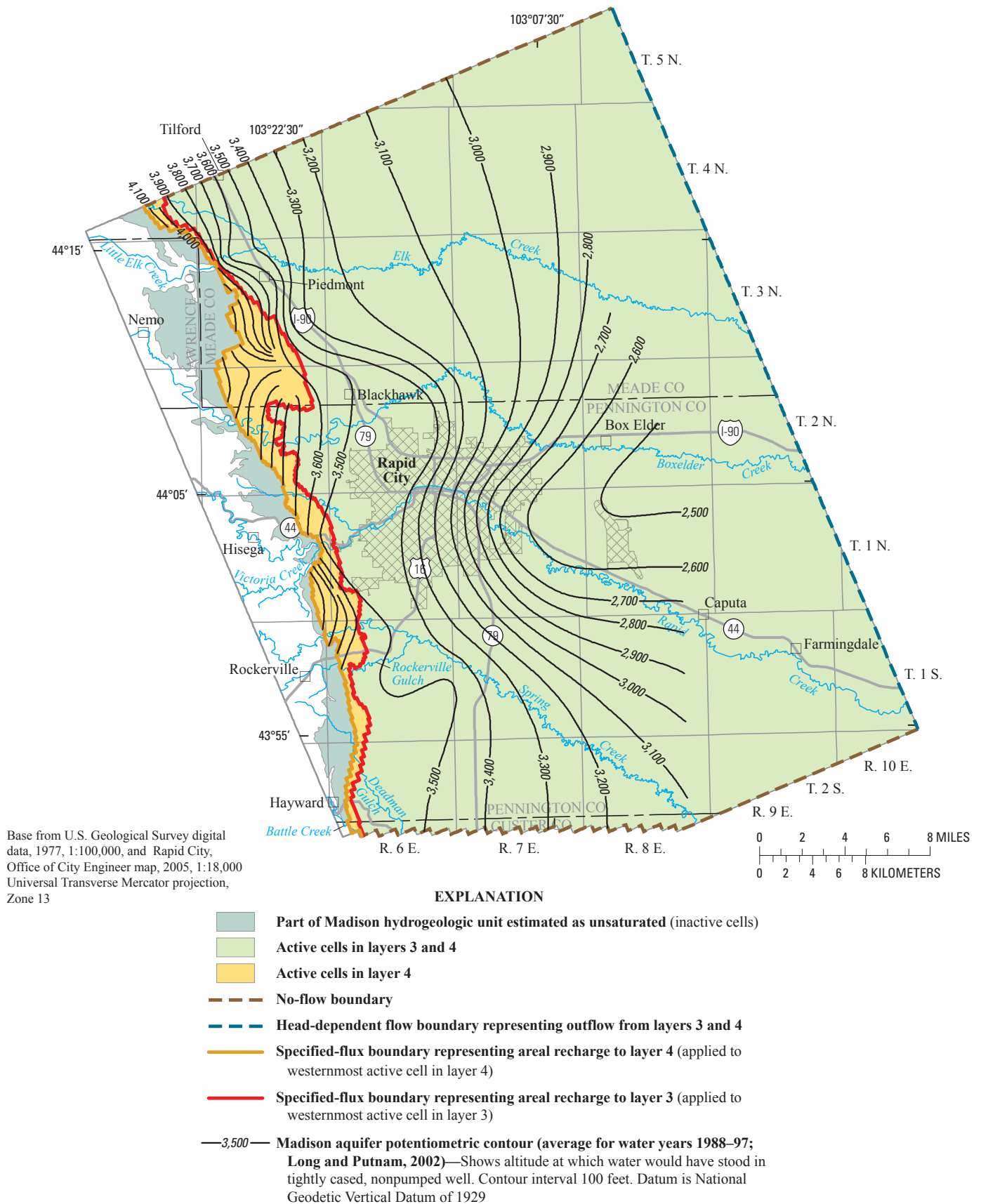


Figure 9. Active model cells and boundary conditions for layers 3 and 4 (Madison hydrogeologic unit).

boundary conductance. Conductance is a term that represents the hydraulic conductivity multiplied by a unit of length, and is expressed in units of feet squared per day (ft^2/d). The boundaries on the north, east, and south perimeters were located about 12 mi from the detailed study area to minimize their influence on analysis of groundwater flow in the detailed study area (fig. 7).

Areal recharge to layers 1–4 (figs. 8 and 9) was represented with a specified-flux boundary (Recharge Package, McDonald and Harbaugh, 1988) with the recharge flux assigned to the westernmost active cell. Groundwater flow in the western part of the outcrops of the Minnelusa and Madison hydrogeologic units was not simulated for the unsaturated areas (figs. 8 and 9). In these areas near the outcrops, groundwater flow includes unsaturated areas, perched water, and caves intermittently filled with water that spills and moves downgradient. This non-Darcian flow was not simulated; therefore, accumulated infiltration of precipitation on the outcrop was assigned to the first downgradient (westernmost) active model cells in layers 1–4. Streamflow recharge was represented with specified-flow cells assigned to selected cells in layers 1–4 to represent streamflow loss that occurred when streams crossed the outcrops of the Minnelusa and Madison hydrogeologic units (fig. 10). Streamflow loss for the larger streams was gaged and described in Hortness and Driscoll (1998). The streamflow recharge cells were located approximately where streamflow loss was observed for the larger streams. For some streams, the specified-flow cells were moved downstream to the westernmost active cell in the model layer.

A specified-flux boundary (Recharge Package, McDonald and Harbaugh, 1988) representing the source of water for upward flow from the Deadwood aquifer (layer 5) to layer 4 was assigned to the westernmost active cells in layer 5 (fig. 11). Active cells for layer 5 extend from one cell west of active cells in layer 4 to the east beyond the start of active cells in layer 3 (fig. 11) with no-flow boundaries on the north, east, and south. The purpose of layer 5 was to approximate the distribution of upward flow from underlying units to the Madison hydrogeologic unit and not to simulate groundwater flow in the Deadwood aquifer. Upward flow from the Minnelusa hydrogeologic unit to overlying units, which was assumed to be small, was represented with specified-flux cells (Recharge Package, McDonald and Harbaugh, 1988) in the western part of the model where more fracturing was likely (Long and Putnam, 2002; fig. 11).

Springs discharging from the Minnelusa and Madison hydrogeologic units (figs. 11 and 12) were represented with head-dependent flow cells (Drain package or River Package, McDonald and Harbaugh, 1998). The user-specified hydraulic head at the cell or cells was set at the altitude of the land surface at the spring. The springs generally appear as seepage from the alluvium along streams; therefore, spring-flow through the bedrock openings could emerge somewhat upgradient from observed flow in the streams. A more detailed

description of individual springs and estimates of spring discharge is included in the subsequent “Springflow” section.

Calibration and Simulated Stress Periods

Calibration of the model was accomplished by two simulations: (1) steady-state simulation of average conditions for water years (WY) 1988–97 and (2) transient simulations for WY 1988–97 divided into twenty 6-month stress periods. The model was calibrated with both the steady-state and transient simulations by comparison of hydraulic head values and flows. Average hydraulic heads for water years 1988–97 (Long and Putnam, 2002) were assumed to approximate hydraulic heads for the steady-state simulation because the 10-year period included a range of hydrologic conditions. Hydraulic heads from the steady-state simulation were used as starting heads for the transient simulation. The transient stress periods (table 1) include a dry period in the late 1980s and early 1990s (stress periods 1–11) and a wet period in the late 1990s (stress periods 12–20). Most recharge occurred during the summer stress periods from April through September, and water use from the Madison aquifer also increased substantially during the summer stress periods. Recharge and other model components are described with more detail in the subsequent “Representation of Flow-System Components” section.

Inverse modeling with the MODFLOW–2000 parameter estimation process (Hill and others, 2000) was used to calibrate the steady-state model, and trial-and-error methods were used to conjunctively calibrate the transient model. The term “parameter” in this report is used to describe any physical property that was estimated by model calibration. An “observation” is a direct measurement or an estimate based on measured data. A parameter can represent a particular property for a single cell or group of model cells. Simulated results were compared to measured or estimated hydraulic heads or flow rates and optimized.

In inverse modeling, simulated values of hydraulic head and flows are statistically compared to observed values. The program adjusts parameter values in an iterative process to produce the best possible match between the observed and simulated values. Parameterization is the process of identifying the aspects of the simulated system that can be optimized with these statistical algorithms. The number of parameters that can be estimated is limited by the amount and distribution of observed data (Hill, 1998). Some parameters may be insensitive to the observed data, or some parameters may be highly correlated with each other and cannot be optimized. The selection of parameters that could be estimated with statistical confidence by inverse modeling is described in the subsequent “Steady-State Calibration” section.

Transient simulations were calibrated by comparison of observed and simulated hydraulic head values and flows using a trial-and-error evaluation. Simulated steady-state hydraulic head values were used as the starting hydraulic head values for the transient simulation. Parameters that were not sensitive

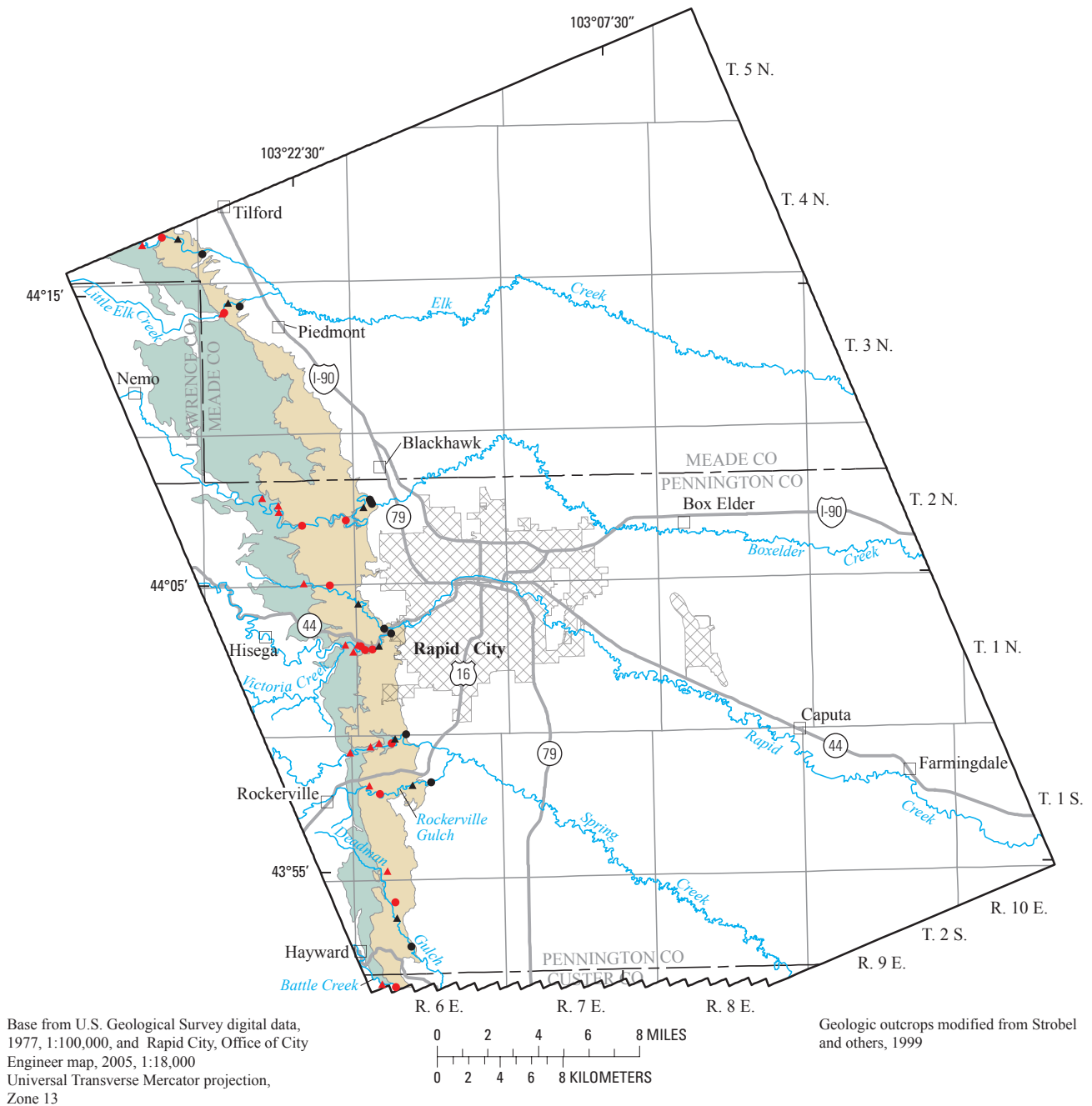
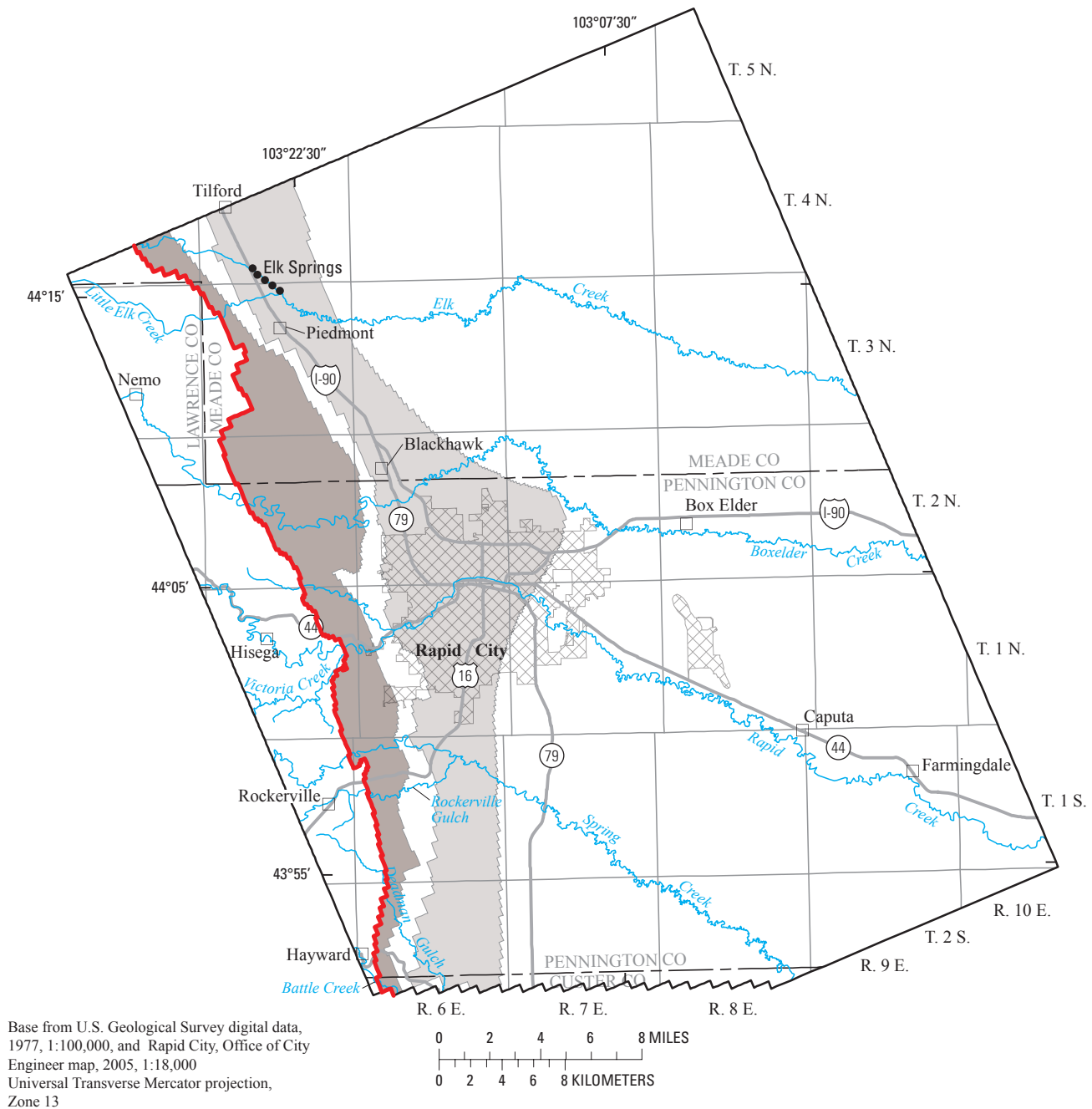


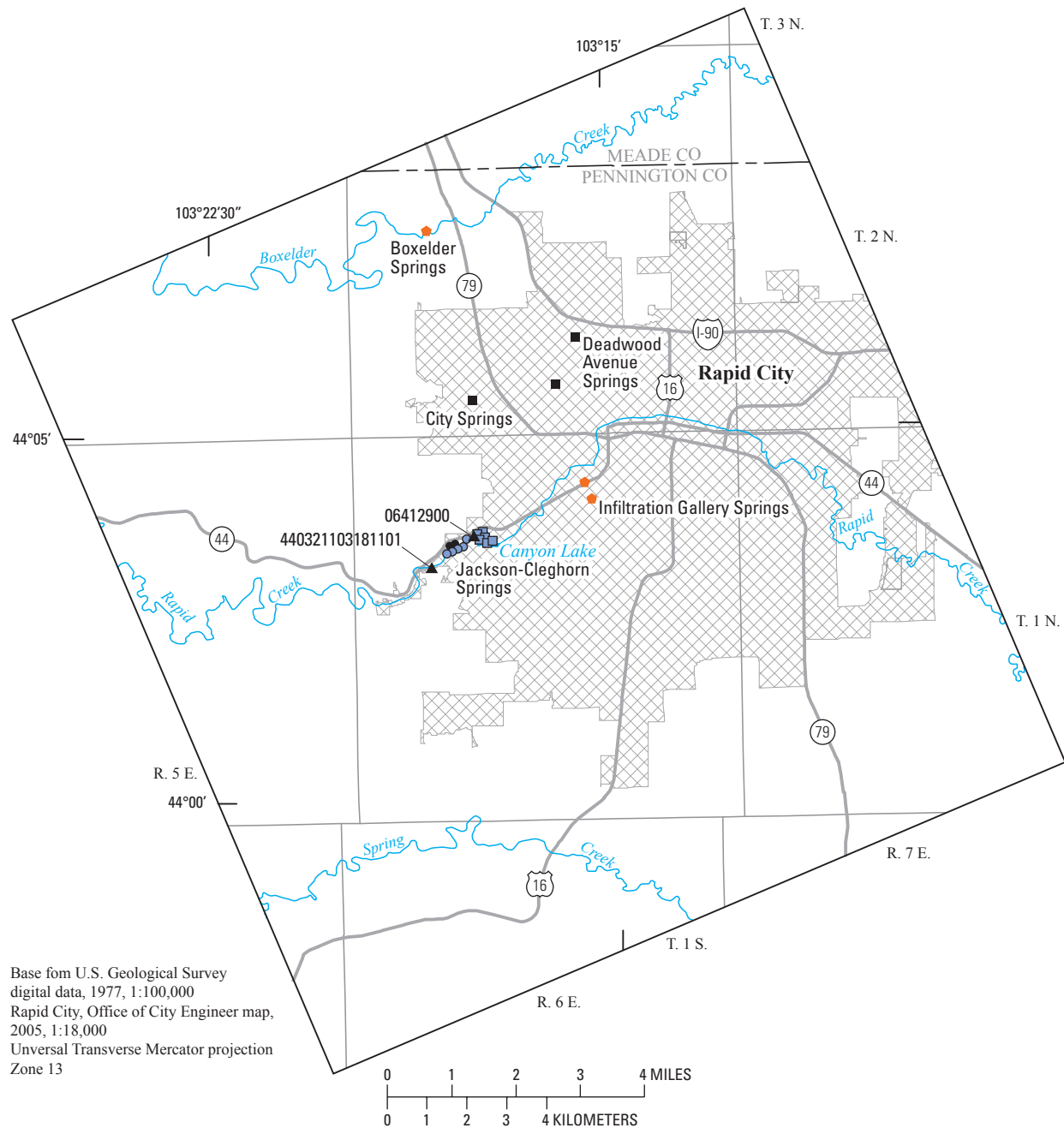
Figure 10. Specified-flow cells representing streamflow recharge.



EXPLANATION

- Specified-flux cells representing upward flow from the Minnelusa hydrogeologic unit to overlying units
- Active cells in layer 5 representing distribution of upward flow from the Deadwood aquifer
- Specified-flux boundary representing upward flow from the Deadwood aquifer (applied to westernmost active cell in layer)
- Head-dependent flow cell (Drain Package) representing discharge from layers 1 and 3 to Elk Springs

Figure 11. Model cells representing upward flow from the Deadwood aquifer to layer 4, flow from the Minnelusa hydrogeologic unit to overlying units, and discharge to Elk Springs.



EXPLANATION

- Head-dependent flow cell (Drain Package) representing discharge from model layer 1 to Boxelder and Infiltration Gallery Springs
- Head-dependent flow cell (River Package) representing hydraulic connection of model layer 1 to Canyon Lake
- Head-dependent flow cell (River Package) representing discharge from model layer 1 to Jackson-Cleghorn Springs
- Head-dependent flow cell (Drain Package) representing discharge from model layer 3 to Jackson-Cleghorn Springs
- Head-dependent flow cell (Drain Package) representing discharge from model layers 1 and 3 to City and Deadwood Avenue Springs
- ▲ Streamflow-gaging station—Number is station identification

Figure 12. Locations of head-dependent flow cells representing springs in the detailed study area.

Table 1. Stress periods for transient simulations.

[Stress period identifiers: “W” represents the 6-month stress period from October through March, and “S” represents the 6-month period from April through October. Total recharge and water-use rates from Long and Putnam (2002)]

Stress period number	Stress period identifier	Time period	Total recharge rate to Minnelusa and Madison hydrogeologic units (cubic feet per second)	Water-use rate for Minnelusa and Madison hydrogeologic units (cubic feet per second)
1	W-88	October 1, 1987–March 31, 1988	18.7	1.9
2	S-88	April 1, 1988–September 30, 1988	23.6	4.7
3	W-89	October 1, 1988–March 31, 1989	16.3	1.7
4	S-89	April 1, 1989–September 30, 1989	34.1	3.4
5	W-90	October 1, 1989–March 31, 1990	22.7	1.8
6	S-90	April 1, 1990–September 30, 1990	50.0	4.2
7	W-91	October 1, 1990–March 31, 1991	19.8	3.0
8	S-91	April 1, 1991–September 30, 1991	135.2	7.2
9	W-92	October 1, 1991–March 31, 1992	32.9	5.4
10	S-92	April 1, 1992–September 30, 1992	40.3	12.5
11	W-93	October 1, 1992–March 31, 1993	25.6	3.7
12	S-93	April 1, 1993–September 30, 1993	149.4	13.6
13	W-94	October 1, 1993–March 31, 1994	46.9	6.2
14	S-94	April 1, 1994–September 30, 1994	59.6	15.0
15	W-95	October 1, 1994–March 31, 1995	43.0	5.8
16	S-95	April 1, 1995–September 30, 1995	180.7	11.0
17	W-96	October 1, 1995–March 31, 1996	57.0	3.2
18	S-96	April 1, 1996–September 30, 1996	131.6	11.4
19	W-97	October 1, 1996–March 31, 1997	82.3	5.5
20	S-97	April 1, 1997–September 30, 1997	181.8	11.9

to the observed data in inverse modeling or by trial-and-error methods were assigned fixed values that were considered to be within the range of values expected for the parameters. The following section describes the properties that were represented with parameters and the remaining properties that were assigned values on the basis of previous estimates.

Representation of Flow-System Components

Flow-system components represented in the model include recharge, discharge, and hydraulic properties. Recharge includes streamflow recharge, areal recharge, and upward flow from the Deadwood aquifer. Upward flow from other underlying units was assumed to be small and included in upward flow represented in layer 5. Recharge to the hydrogeologic units occurs by two processes: streamflow recharge and areal recharge. Streamflow recharge originates from precipitation runoff in the central Black Hills upstream from the outcrops of the Minnelusa and Madison hydrogeologic units. These streams lose flow to swallow holes and fractures as they cross the outcrops. Areal recharge occurs

from infiltration of precipitation on the outcrops. Because of the fractures and openings on the outcrop, any precipitation that exceeds evapotranspiration is assumed to infiltrate. Discharge includes springflow, water use, flow to overlying units, and regional outflow. Hydraulic properties and model features include horizontal and vertical hydraulic conductivity, horizontal flow barriers, and storage properties.

A steady-state water budget (table 2) was estimated by modifying the water-balance analysis by Long and Putnam (2002). The water-balance analysis quantified budget components for twenty 6-month time steps for water years (WY) 1988–97 (table 1). The average water budget for that 10-year period included varied hydrologic conditions that collectively were assumed to approximate steady-state conditions. That water-balance analysis is described with more detail in the following subsections for each budget component.

Recharge and discharge were adjusted slightly to account for the change in storage that was included in the 10-year average water budget (1988–97) from Long and Putnam (2002). Inflow (recharge) was decreased by 4.4 ft³/s, and outflow was increased by 4.3 ft³/s to account for the change in storage of 8.7 ft³/s. Streamflow recharge was reduced by

3.1 ft³/s, and areal recharge was reduced by 1.3 ft³/s. Streamflow recharge for the Minnelusa hydrogeologic unit was distributed between model layers 1 and 2 with a larger portion assigned to model layer 1 because of its higher permeability. Streamflow recharge for the Madison hydrogeologic unit was distributed between model layers 3 and 4 with a larger portion assigned to model layer 3 because of its higher permeability. Areal recharge was distributed equally between model layers 1 and 2 for the Minnelusa hydrogeologic unit and between layers 3 and 4 for the Madison hydrogeologic unit. Although layers 1 and 3 represented more permeable parts of the units, the outcrop area associated with these layers was smaller than the outcrop areas for layers 2 and 4; therefore, areal recharge was distributed equally between the layers for each hydrogeologic unit. The estimated springflow was increased by 2.0 ft³/s, and regional outflow from the Minnelusa and Madison hydrogeologic units was increased by 1.3 ft³/s and 1.0 ft³/s, respectively.

Recharge

Recharge assigned in the model was based on rates calculated by Long and Putnam (2002) and was not adjusted during calibration (tables 18 and 19, in the “Supplemental Information” section). The calculated streamflow recharge rate was determined from streamflow-gaging records and loss thresholds for the major streams (Hortness and Driscoll, 1998).

Recharge rates for the small ungaged streams were estimated by making correlations with data from adjacent gaged streams (Long and Putnam, 2002).

Areal recharge from infiltration of precipitation on the outcrops of the Minnelusa and Madison hydrogeologic units was estimated by a stream-catchment-runoff-yield method applied in the Black Hills hydrology study (Carter and others, 2001) and applicable in mountainous areas with large variability in precipitation (Carter and Driscoll, 2005). Application of this method to determine areal recharge for the twenty 6-month time steps for WY 1988–97 is presented in Long and Putnam (2002). Because of fractures and openings in the exposed outcrops, any precipitation that exceeded evapotranspiration was assumed to infiltrate to the aquifers. The amount of precipitation that exceeded evapotranspiration was determined by analyzing runoff from adjacent drainage basins in the Precambrian rocks upgradient from the outcrops.

Streamflow Recharge

Drainage areas that contribute flow to the streams that cross the outcrops of the Minnelusa and Madison hydrogeologic units were delineated, and streamflow recharge rates were calculated for the 10 streams (table 3 and fig. 10) that cross these outcrops (Long and Putnam, 2002). Daily streamflow records were available for the larger streams that lose flow when crossing the outcrops, including Battle, Spring, Rapid, Boxelder, and Elk Creeks (fig. 10), and loss thresholds

Table 2. Estimated water-budget components for steady-state simulation.

Component		Water-budget rate (cubic feet per second)						Model total
		Minnelusa hydrogeologic unit			Madison hydrogeologic unit			
		Model layer		Total	Model layer		Total	
		1	2		3	4		
Inflow	Streamflow recharge	3.4	2.6	6.0	23.8	12.4	36.2	42.2
	Areal recharge	2.9	2.9	5.8	7.6	7.5	15.1	20.9
	Upward flow from Deadwood aquifer	(¹)	(¹)	(¹)	(¹)	6.3	6.3	6.3
Total inflow								69.4
Outflow	Springflow	(²)	(²)	(²)	(²)	(²)	(²)	32.8
	Water use	3.4	(³)	3.4	6.7	(³)	6.7	10.1
	Flow to units overlying the Minnelusa hydrogeologic unit	2.0	(⁴)	2.0	(⁴)	(⁴)	(⁴)	2.0
	Regional outflow	(⁵)	(⁵)	12.3	(⁵)	(⁵)	12.2	24.5
Total outflow								69.4

¹Upward flow from the Deadwood aquifer only occurred to layer 4.

²Distribution among layers was determined by model calibration; springflow measurements are described in the “Springflow” section.

³Because layers 1 and 3 represented the more permeable parts of the Minnelusa and Madison hydrogeologic units, water use from wells was simplified to a specified flow from cells in layer 1 or 3, respectively.

⁴Discharge from the model as upward flow only occurred from layer 1.

⁵Distribution of discharge among model layers was determined in model calibration.

for all but 2 of the 10 streams were determined by Hortness and Driscoll (1998). For the other two streams (Rockerville Gulch and Deadman Gulch), loss thresholds were estimated by Long and Putnam (2002). For the small ungaged streams, flow was estimated by correlation with flow at adjacent gaged streams as described in Long and Putnam (2002), and the distribution of recharge among the model layers was approximated. Measured or estimated daily streamflow less than or equal to the loss threshold recharged the Madison hydrogeologic unit, and once that threshold flow was exceeded, streamflow recharged the Minnelusa hydrogeologic unit at a rate equal to the loss threshold for this unit.

Streamflow recharge was represented with specified-flow cells (Well Package, McDonald and Harbaugh, 1998) in model layers 1–4 (fig. 10). The recharge was distributed between the four layers with larger amounts assigned to layers 1 and 3, which represented the more permeable parts of each hydrogeologic unit. Streamflow recharge rates assigned to the steady-state simulation (table 3) show the relatively large contributions from Boxelder, Rapid, and Spring Creeks. Because Battle Creek and Elk Creek were on the southern and northern boundaries of the model, respectively, it was assumed that one-half of the streamflow recharge for these streams entered the model area. When streamflow was adequate to flow through the loss zones, the more permeable layers 1 and 3 were assigned a larger portion of the recharge calculated for each hydrogeologic unit. When streamflow was less than the loss threshold, layer 4 had the first opportunity to receive recharge; therefore, for stress periods with relatively small streamflow, the recharge to layers 3 and 4 was about equal. As streamflow increased, layer 4 could not accept any more

recharge and layer 3 received the increase in streamflow. A similar scenario occurred for layers 1 and 2; however, available flow was reduced by the upstream loss to layers 3 and 4. Spring Creek and Boxelder Creek had the largest streamflow loss thresholds; thus, the largest streamflow recharge to layer 3 occurred along these two streams. Rapid Creek almost always exceeded the loss threshold because streamflow was regulated by releases from an upstream dam; therefore, the distribution of streamflow recharge for this stream was the same for all stress periods.

Transient streamflow recharge rates for the Minnelusa and Madison hydrogeologic units are presented in table 18 in the “Supplemental Information” section for each of the 10 streams. Total transient streamflow recharge rates for all streams by stress period for the Minnelusa and Madison hydrogeologic units ranged from 14.1 ft³/s (stress period 3, winter WY 1989) to 102.2 ft³/s (stress period 20, summer WY 1997). Distribution of transient streamflow recharge rates for 20 stress periods by model layer (fig. 13) shows the relatively large rates during wet periods (stress periods 12–20). Because the Minnelusa hydrogeologic unit receives recharge only when streamflow exceeds the loss threshold for the Madison hydrogeologic unit, streamflow recharge rates to model layers 1 and 2 were substantially less than streamflow recharge rates to model layers 3 and 4.

Areal Recharge

The methods used for calculating the areal recharge are described in detail in Long and Putnam (2002) and summarized here. The outcrop area of the Minnelusa and Madison

Table 3. Steady-state streamflow recharge rates distributed by model layer.

Stream	Streamflow recharge rate (cubic feet per second)						Total model
	Minnelusa hydrogeologic unit			Madison hydrogeologic unit			
	Model layer		Total	Model layer		Total	
	1	2		3	4		
Elk Creek	0.4	0.4	0.8	1.2	1.2	2.4	3.2
Little Elk Creek	.3	.3	.6	.3	.3	.6	1.2
Boxelder Creek	1.4	.6	2.0	9.1	2.8	11.9	13.9
Unnamed tributary on Rapid Creek	.0	.0	.0	.1	.1	.2	.2
Rapid Creek	.9	.9	1.8	4.5	3.0	7.5	9.3
Victoria Gulch	(¹)	(¹)	(¹)	(¹)	.8	.8	.8
Spring Creek	.4	.4	.8	6.9	2.3	9.2	10.0
Rockerville Gulch	.0	.0	.0	.4	.4	.8	.8
Deadman Gulch	.0	.0	.0	.0	.2	.2	.2
Battle Creek	(²)	(²)	(²)	1.3	1.3	2.6	2.6
Total	3.4	2.6	6.0	23.8	12.4	36.2	42.2

¹The stream did not cross the part of the hydrogeologic unit represented by layers 1, 2, or 3.

²Streamflow loss was offset by reemerging springs.

hydrogeologic units was subdivided into five zones for calculation of areal recharge from infiltration of precipitation on the outcrops (table 4 and fig. 14). The zones represented infiltration of precipitation on the outcrop areas between major streams. The fraction of precipitation that was available for recharge after evapotranspiration was represented by runoff in gaged stream basins in Precambrian rocks adjacent to the outcrop areas. Most of the groundwater infiltration in these Precambrian basins with high relief returns to the basins as springflow; therefore, the runoff per unit area was assumed to represent the fraction of precipitation that was available for infiltration on the adjacent outcrops.

The areal recharge was distributed equally between the two model layers representing each hydrogeologic unit. The distribution was assumed to be equal because the more permeable layers 1 and 3 represented a smaller percentage of the total outcrop area than the thicker layers 2 and 4. The recharge flux for each model layer was assigned to the first downgradient (westernmost) active model cell. Steady-state areal recharge by zone and model layer (table 5) shows larger recharge rates for the northern zones because of greater precipitation and larger outcrop areas for these zones than for the southern zones. Calculated areal recharge rates for the Minnelusa hydrogeologic unit were less than rates for the Madison

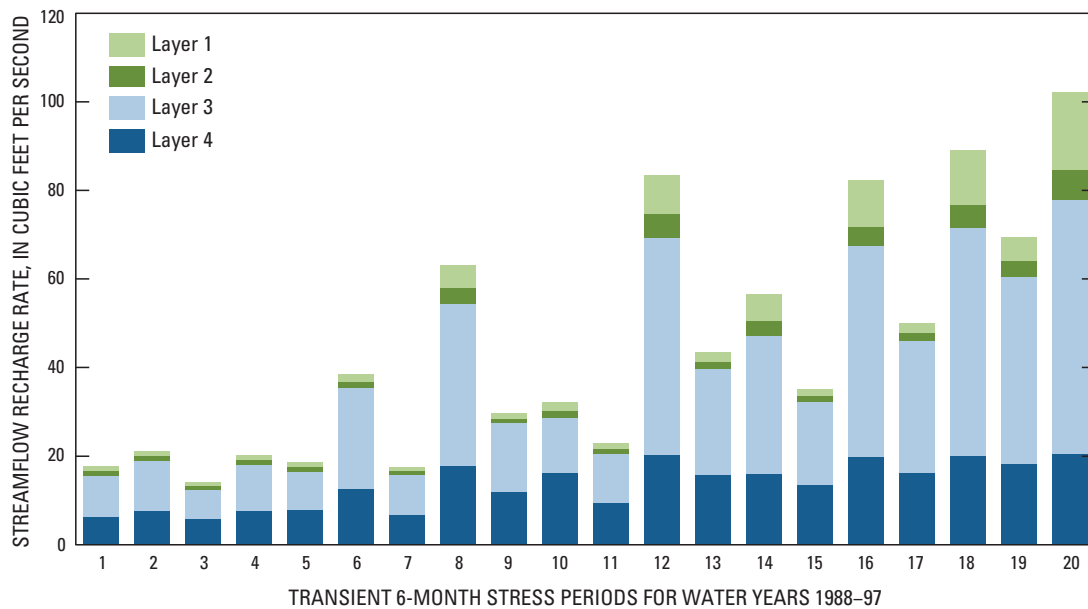


Figure 13. Total transient streamflow recharge rates for twenty 6-month stress periods by model layer.

Table 4. Areal recharge zones and average annual precipitation on outcrops of the Minnelusa and Madison hydrogeologic units.

[From Long and Putnam (2002). Outcrop zone represents extent of hydrogeologic unit outcrop bounded by streams crossing the outcrop. Average annual precipitation by hydrogeologic unit was calculated from precipitation data for water years 1961–98 by Driscoll, Hamade, and Kenner (2000). NA, not applicable]

Outcrop zone	Outcrop zone number shown on fig. 14	Outcrop area by hydrogeologic unit (square miles)		Average annual precipitation by hydrogeologic unit (inches)		Average annual precipitation rate by hydrogeologic unit (cubic feet per second)	
		Minnelusa	Madison	Minnelusa	Madison	Minnelusa	Madison
Elk Creek to Little Elk Creek	1	3.6	10.1	22.3	24.0	5.9	17.9
Little Elk Creek to Boxelder Creek	2	15.2	27.3	19.2	20.0	21.5	40.2
Boxelder Creek to Rapid Creek	3	12.5	9.8	18.9	19.6	17.4	14.2
Rapid Creek to Spring Creek	4	7.1	4.6	18.6	19.1	9.7	6.5
Spring Creek to Battle Creek	5	15.1	6.6	19.4	19.9	21.6	9.7
Total	NA	53.5	58.4	NA	NA	76.1	88.5

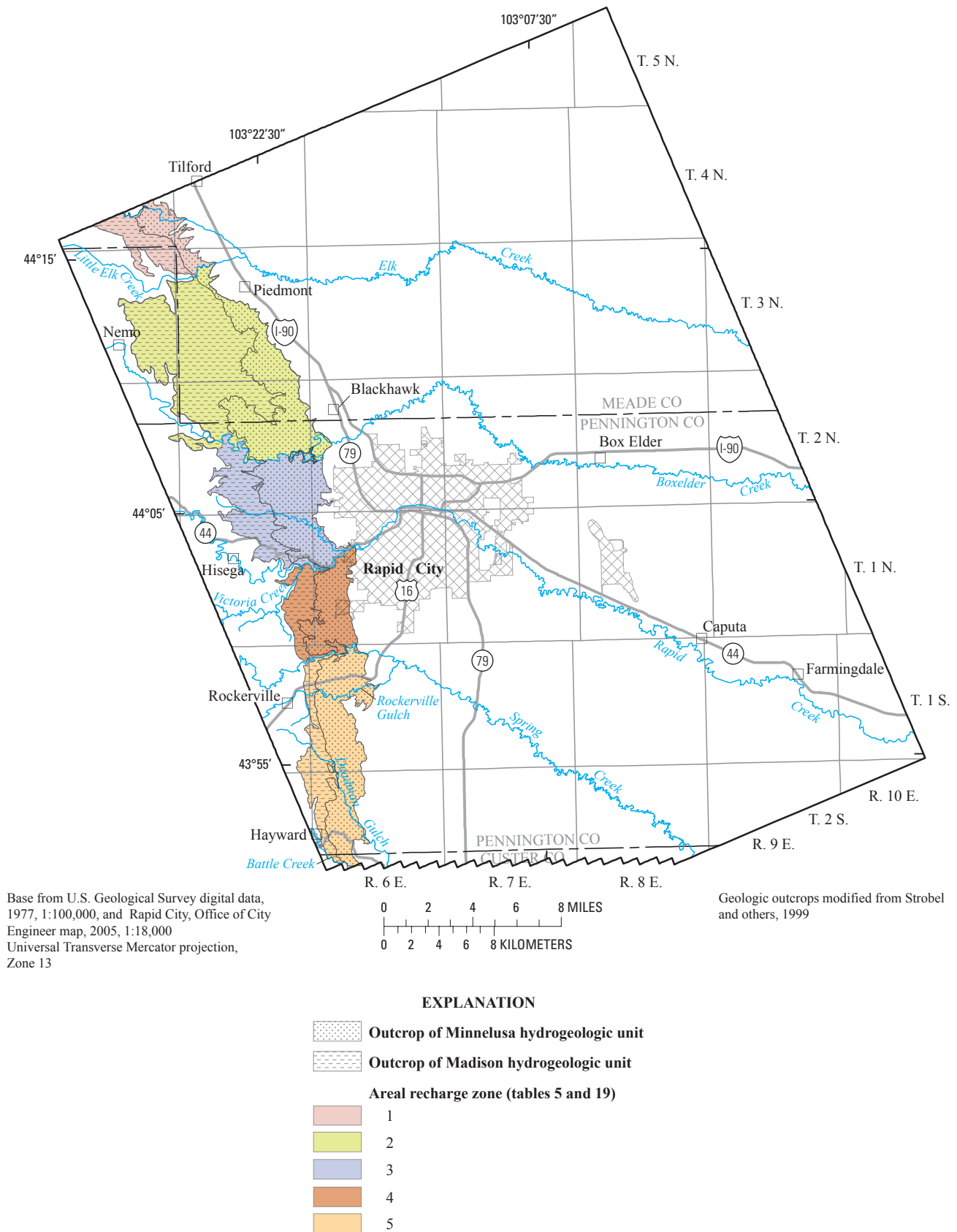


Figure 14. Locations of areal recharge zones.

Table 5. Areal recharge zones and steady-state areal recharge rates distributed by model layer.

[Outcrop zone represents extent of hydrogeologic unit outcrop bounded by streams crossing the outcrop. NA, not applicable]

Outcrop zone	Outcrop zone number shown on fig. 14	Areal recharge rate (cubic feet per second)						Total model
		Minnelusa hydrogeologic unit			Madison hydrogeologic unit			
		Model layer		Total	Model layer		Total	
		1	2		3	4		
Elk Creek to Little Elk Creek	1	0.3	0.3	0.6	1.6	1.6	3.2	3.8
Little Elk Creek to Boxelder Creek	2	1.1	1.1	2.2	3.8	3.7	7.5	9.7
Boxelder Creek to Rapid Creek	3	.6	.6	1.2	1.2	1.2	2.4	3.6
Rapid Creek to Spring Creek	4	.2	.2	.4	.4	.4	.8	1.2
Spring Creek to Battle Creek	5	.7	.7	1.4	.6	.6	1.2	2.6
Total	NA	2.9	2.9	5.8	7.6	7.5	15.1	20.9

hydrogeologic unit because 50 percent of the precipitation that fell on the thin western part of the Minnelusa hydrogeologic unit outcrop was assumed to have infiltrated downward to the Madison hydrogeologic unit.

Transient areal recharge rates for the Minnelusa and Madison hydrogeologic units (Long and Putnam, 2002) are presented in table 19 in the “Supplemental Information” section. Total transient areal recharge by stress period (fig. 15) shows the substantially larger areal recharge during wet stress periods (stress periods 12–20) compared to recharge during dry periods (stress periods 1–11). During the stress periods with large precipitation, a greater fraction of precipitation becomes recharge. During the summer stress periods 8, 12, 16, and 20, areal recharge was about the same as streamflow recharge (fig. 13); however, during the winter stress periods and drier summer stress periods, streamflow recharge was larger than areal recharge. Total transient areal recharge rates by stress period for the Minnelusa and Madison hydrogeologic units (table 19) ranged from 1.1 ft³/s (stress period 1, winter WY 1988) to 98.4 ft³/s (stress period 16, summer WY 1995). Most of the steady-state areal recharge occurred during the wet summer periods. The steady-state areal recharge rate (table 5) was largest for the Little Elk Creek to Boxelder Creek zone (zone 2) because of the relatively large outcrop area and greater precipitation in this zone than in the southern zones.

Upward Flow from Deadwood Aquifer

Layer 5 represented an approximation of the underlying Deadwood aquifer to simulate upward flow to the Madison hydrogeologic unit. The average upward flow from the Deadwood aquifer to the Madison hydrogeologic unit was estimated by Long and Putnam (2002) to be 6.3 ft³/s in the study area, representing about 9 percent of the total average inflow to the model. This upward flow was assumed to be constant during transient simulations. Although the hydraulic head in the Deadwood aquifer and the Madison hydrogeologic unit

changed over time, it was assumed that the vertical gradient between the two units did not change substantially. Input to model layer 5 was accomplished by a specified flux (fig. 11) equally distributed to the westernmost cells in layer 5.

Discharge

Springflow, water use, flow to overlying units, and regional outflow were estimated by Long and Putnam (2002). Water use was assigned to the model by using specified-flow cells. Flow to overlying units was assumed to be small and assigned a value that did not change in transient simulations. Simulated springflow and regional outflow rates were dependent on the conductance of head-dependent cells, which was estimated by model calibration.

Springflow

Discharge from springs (figs. 11 and 12) calculated by model simulation was the difference between the user-specified hydraulic head of the spring and the simulated hydraulic head multiplied by the conductance for the cell. Conductance is a term that groups the hydraulic conductivity and the length of flow path. The conductance term for springs was represented as a parameter that was adjusted in model calibration (described in the subsequent “Steady-State Calibration” section and “Transient Calibration” section) to produce the best fit with estimated flows and hydraulic head values. Generally, the Madison hydrogeologic unit has been interpreted as the major source of water for large springs in the model area on the basis of hydrochemistry and the presence of conduits (Anderson and others, 1999; Long and Putnam, 2002). Although the Madison hydrogeologic unit is the predominant source of water for the springs, substantial interaction between the upward flow and the Minnelusa hydrogeologic unit is possible. In representing discharge from these springs, both the Minnelusa and Madison hydrogeologic units

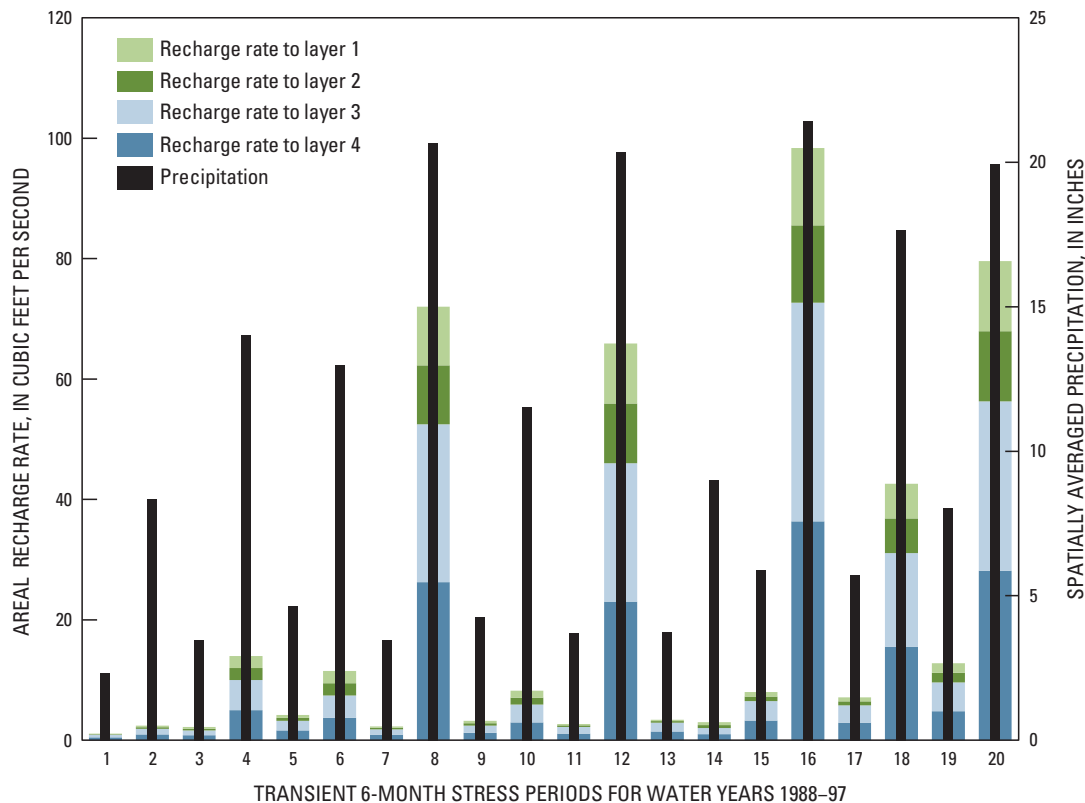


Figure 15. Total transient areal recharge rates for twenty 6-month stress periods by model layer and spatially averaged precipitation on outcrops of the Minnelusa and Madison hydrogeologic units.

were considered as potential sources. Steady-state springflow (table 6) was estimated by Anderson and others (1999) and Long and Putnam (2002).

The springflow of Jackson-Cleghorn Springs, the largest spring in the study area, was estimated as 21.6 ft³/s by a water-budget analysis of a Rapid Creek stream reach that extended through the Jackson-Cleghorn complex (Anderson and others, 1999). On the basis of stable isotope data, they concluded that the springs were a regional discharge point from the Madison aquifer. The spring's discharge from the Madison hydrogeologic unit in the current model was represented with two head-dependent flow cells (Drain Package, McDonald and Harbaugh, 1998) in layer 3 (fig. 12) with hydraulic head set equal to the hydraulic head in the alluvium (Anderson and others, 1999), and the conductance was represented with parameter CLEGmdsn that was estimated in model calibration. Although assumed to be small, a contribution to spring discharge from the Minnelusa hydrogeologic unit also is possible. The component of discharge from Jackson-Cleghorn Springs from the Minnelusa hydrogeologic unit was represented with five head-dependent cells in layer 1 located along the Rapid Creek reach near Jackson-Cleghorn Springs (fig. 12). This hydraulic connection was represented with special head-dependent flow cells (River Package, McDonald and Harbaugh, 1998) that allowed both outflow and inflow

to the model depending on the relation between the hydraulic head of the water body and the hydraulic head in the cell. Conductance was represented with parameter CLEGmnls that was estimated in calibration. This Rapid Creek stream reach and associated alluvium overlies the outcrop of the Minnelusa hydrogeologic unit and was assumed to be hydraulically connected to the unit.

Comparison of hydraulic head values in the Minnelusa and Madison hydrogeologic units and Rapid Creek at Jackson Springs, located in the southwest part of the Jackson-Cleghorn springs area provides some insight on the potential for interaction between the Minnelusa and Madison hydrogeologic units. On September 8, 1999, hydraulic head in the Minnelusa hydrogeologic unit in the observation well at site 231 (well JS-1B; table 20 in "Supplemental Information" section), which is located near Jackson-Cleghorn Springs, was about 17 ft above Rapid Creek. At site 52, a nested well completed in the Madison hydrogeologic unit (well JS-1A; table 21 in "Supplemental Information" section), hydraulic head was about 40 ft above Rapid Creek. Hydraulic head in the Minnelusa hydrogeologic unit was about 10 ft higher at site 231 than it was in the observation well at site 233 (well CHLN-1; table 20), which is located about 0.5 mi south of site 231. This higher hydraulic head indicated the possibility of a mound created in the Minnelusa hydrogeologic unit from

Table 6. Steady-state springflow.

[NA, not applicable]

Spring name	Conductance parameter names	Steady-state springflow (cubic feet per second)
Jackson-Cleghorn Springs	CLEGmdsn, CLEGmnls	¹ 21.6
City Springs	CITYSPR	² 1.7
Deadwood Avenue Springs	DWSPR	² 2.8
Boxelder Springs	BESPR	² 1.0
Elk Springs	ELKSPR	² 3.7
Infiltration Gallery Springs	GAL	¹ 1.5
Canyon Lake	CLAKE	³ .5
Total	NA	32.8

¹From Anderson and others, 1999.²From Long and Putnam, 2002.³No quantitative measurements were available; small discharge was assumed because of hydraulic connection.

water moving upward from the Madison hydrogeologic unit. The vertical hydraulic conductivity for cells in layer 2 in the Jackson-Cleghorn Springs area was assumed to be substantially higher than surrounding areas because of collapse features and solution enhanced openings. The multiplier for layer 2 vertical hydraulic conductivity was determined in model calibration.

Discharge at City Springs was represented by head-dependent cells (Drain Package, McDonald and Harbaugh, 1998) in layers 1 and 3 (fig. 12), and the conductance was assumed to be similar for the two cells and was represented with parameter CITYSPR. Because of the structural features in this area (fig. 1), vertical hydraulic conductivity for layer 2 in the City Springs area was assumed to be substantially higher than for other parts of layer 2. During dry periods, flow from this spring decreased to near zero (Long and Putnam, 2002).

Discharge at Deadwood Avenue Springs was represented with head-dependent cells (Drain Package, McDonald and Harbaugh, 1998) in both layers 1 and 3 (fig. 12). In the Deadwood Avenue Springs area, the average hydraulic head (Long and Putnam, 2002) was about 100 ft above the land surface in the Madison hydrogeologic unit and near the land surface in the Minnelusa hydrogeologic unit. Therefore, most spring discharge probably occurred from the Madison hydrogeologic unit. Conductance was assumed similar for both cells and was estimated with parameter DWSPR in model calibration.

Boxelder Springs was assumed to increase from no flow in dry periods to estimated flows of about 7 ft³/s in wet periods (Carter and others, 2001; Long and Putnam, 2002). Because the hydraulic head in the Madison hydrogeologic unit was always well above the land surface even during periods when there was no observed spring discharge, Boxelder Springs was assumed to discharge from the Minnelusa hydrogeologic

unit and was represented with a head-dependent cell (Drain Package, McDonald and Harbaugh, 1998) in layer 1 (fig. 12), and conductance was estimated with parameter BESPR in model calibration.

Elk Springs was represented by five head-dependent cells (Drain Package, McDonald and Harbaugh, 1998) in both layers 1 and 3 (fig. 11) upstream on Elk Creek from the point identified as the spring on the map in Long and Putnam (2002). During the wet periods (late 1990s), flow was observed to increase in the downstream direction in this reach. Flow was very low during the dry periods, but increased dramatically during the wet periods. On the basis of hydraulic head values, the source potentially could be both the Minnelusa and Madison hydrogeologic units. Conductance was assumed to be similar for each layer and was estimated with parameter ELKSPR in model calibration.

Anderson and others (1999) estimated a small discharge contribution of about 1.5 ft³/s from bedrock units to two infiltration galleries along Rapid Creek about 0.7 mi and 1.2 mi downstream, respectively, from Canyon Lake. This discharge was represented by two head-dependent cells (Drain Package, McDonald and Harbaugh, 1998) in layer 1 (fig. 12), and conductance was estimated with parameter GAL in model calibration.

When Canyon Lake was drained in the winter of 1995–96, hydraulic head in an observation well completed in the Minnelusa hydrogeologic unit near Canyon Lake (site 229, table 20) responded with a similar decline in hydraulic head (Driscoll, Bradford, and Moran, 2000). Canyon Lake was represented with six head-dependent cells (River Package, McDonald and Harbaugh, 1998) in layer 1 (fig. 12) that allowed inflow or outflow from the model depending on the hydraulic head relation, and conductance was estimated with parameter CLAKE in model calibration.

Water Use

Water-use estimates (table 7) from Long and Putnam (2002) were used in model simulations. All water use was represented by specified-flow cells (Well Package, McDonald and Harbaugh, 1998) in layer 1 for pumping in the Minnelusa hydrogeologic unit and in layer 3 for pumping in the Madison hydrogeologic unit (fig. 16). The average water-use rates for the Minnelusa and Madison hydrogeologic units for the 10-year period were 3.4 and 6.7 ft³/s, respectively. The average combined pumping rates were 6.3 ft³/s for the October–March (W) stress periods and 13.8 ft³/s for the April–September (S) stress periods. The highest average pumping rate of 19.1 ft³/s was for stress period 14 (S–94).

Flow to Overlying Units

Groundwater under confined pressure in the Minnelusa hydrogeologic unit probably flows into overlying units (figs. 2 and 3) including aquifers in the Minnekahta Limestone and Inyan Kara Group through fractures or breccia pipes. Because very little information was available to calculate the upward flow from the Minnelusa hydrogeologic unit, the estimate of 2.0 ft³/s from the water-budget calculation in Long and Putnam (2002) was uniformly assigned to the specified-flux cells (Recharge Package, McDonald and Harbaugh, 1998; fig. 11). Evidence supporting the potential upward flow is presented in Long and Putnam (2002).

Table 7. Water-use rates for steady-state and transient simulations.

[From Long and Putnam (2002). W, winter; S, summer (W–88 = winter, water year 1988)]

Stress period	Other stress period identifier	Water-use rate (cubic feet per second)								Total model
		Model layer 1 (Minnelusa hydrogeologic unit)				Model layer 3 (Madison hydrogeologic unit)				
		Rapid City	Other domestic and public water supplies	Industrial and irrigation use	Total layer 1	Rapid City	Other domestic and public water supplies	Industrial and irrigation use	Total layer 3	
1	W–88	1.5	1.2	0.2	2.9	0.0	1.3	0.6	1.9	4.8
2	S–88	1.4	3.1	1.2	5.7	.0	3.4	1.3	4.7	10.4
3	W–89	.5	1.0	.2	1.7	.0	1.1	.6	1.7	3.4
4	S–89	.7	1.8	1.2	3.7	.1	2.0	1.3	3.4	7.1
5	W–90	1.6	1.1	.2	2.9	.0	1.2	.6	1.8	4.7
6	S–90	1.9	1.8	1.2	4.9	.9	2.0	1.3	4.2	9.1
7	W–91	1.8	1.3	.2	3.3	1.0	1.4	.6	3.0	6.3
8	S–91	1.6	2.0	1.0	4.6	3.3	2.2	1.7	7.2	11.8
9	W–92	1.3	1.1	.2	2.6	3.6	1.2	.6	5.4	8.0
10	S–92	1.2	1.9	1.2	4.3	9.1	2.1	1.3	12.5	16.8
11	W–93	1.1	1.3	.2	2.6	1.7	1.4	.6	3.7	6.3
12	S–93	1.1	1.8	1.1	4.0	10.6	1.9	1.1	13.6	17.6
13	W–94	.6	1.3	.2	2.1	4.2	1.4	.6	6.2	8.3
14	S–94	.3	2.6	1.2	4.1	10.9	2.8	1.3	15.0	19.1
15	W–95	.8	1.3	.2	2.3	3.7	1.4	.7	5.8	8.1
16	S–95	.6	2.1	1.0	3.7	7.5	2.3	1.2	11.0	14.7
17	W–96	.5	1.4	.2	2.1	.9	1.6	.7	3.2	5.3
18	S–96	.6	2.4	1.0	4.0	7.6	2.6	1.2	11.4	15.4
19	W–97	.8	1.5	.2	2.5	3.2	1.6	.7	5.5	8.0
20	S–97	.9	2.2	1.0	4.1	8.4	2.3	1.2	11.9	16.0
Average (steady state)		1.0	1.7	.7	3.4	3.8	1.9	1.0	6.7	10.1

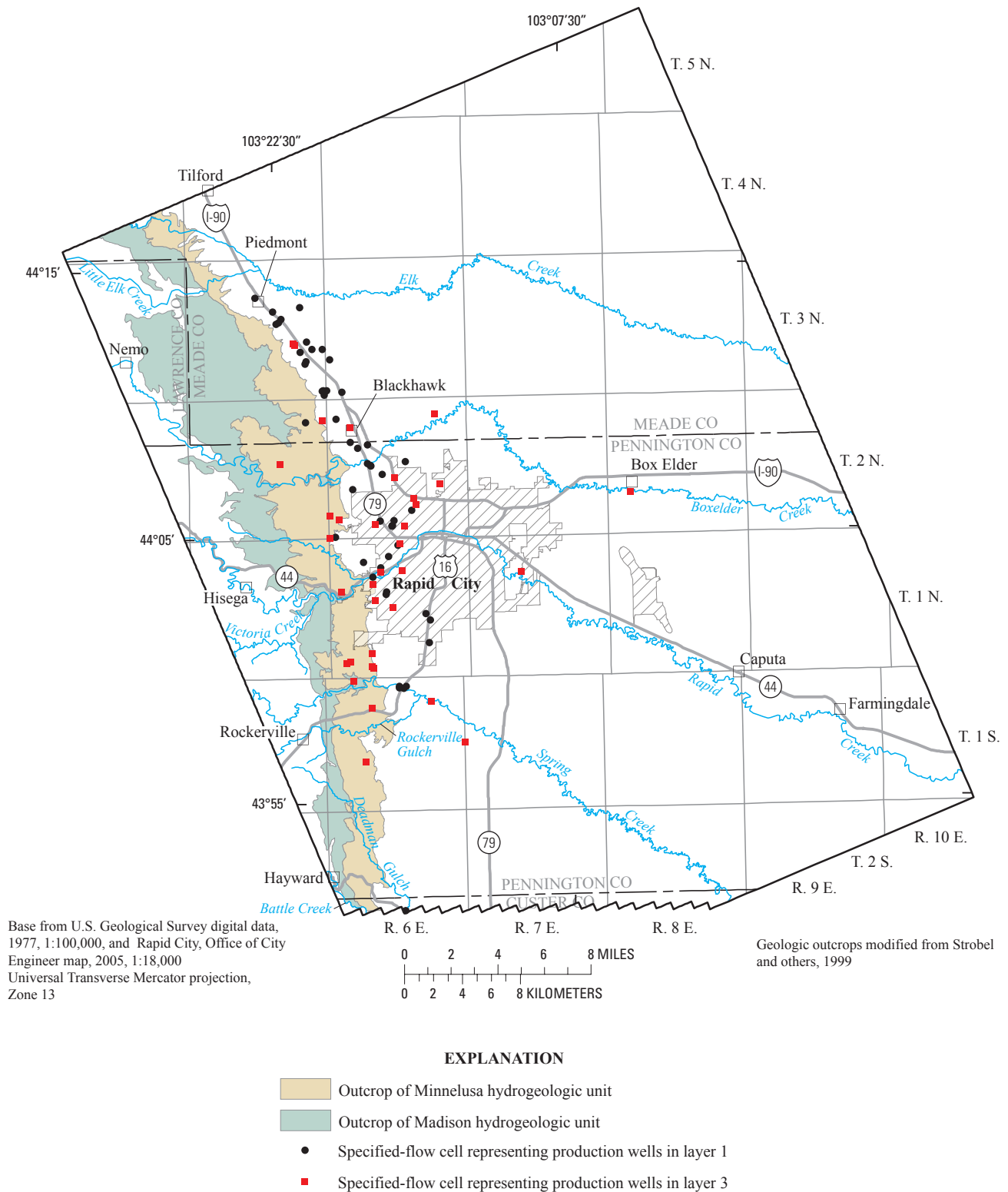


Figure 16. Locations of specified-flow cells representing water use in model layers 1 and 3.

Regional Outflow

Regional outflow to the east was estimated by Long and Putnam (2002) to be 22.2 ft³/s. The model boundary was extended about 6 mi farther east than the eastern boundary used to define the aquifer analysis area in Long and Putnam (2002) to minimize the effect of the boundary on the detailed study area. Head-dependent flow cells (Drain Package, McDonald and Harbaugh, 1998) represented the boundary (figs. 8 and 9) with the hydraulic head extrapolated from Long and Putnam (2002). Conductance for head-dependent cells in layers 1 and 2 at the eastern boundary was represented with parameter MnlEB, and conductance for head-dependent cells in layers 3 and 4 was represented with parameter MdsnEB. Steady-state discharge was estimated as 12.3 ft³/s from the Minnelusa hydrogeologic unit and 12.2 ft³/s from the Madison hydrogeologic unit.

Hydraulic Properties

Hydraulic properties represented include horizontal hydraulic conductivity, vertical hydraulic conductivity, conductance of horizontal flow barriers, and storage coefficients. Hydraulic properties were represented by parameters that included cells grouped into zones.

Horizontal Hydraulic Conductivity

Horizontal hydraulic conductivity values for active cells in layer 1 were estimated by using three parameters named HK1_1, HK1_2, and HK1_3 (fig. 17) and multiplier arrays. Parameter group HK1_1 represented zones where the relatively gradual hydraulic gradient is attributable to larger hydraulic conductivity. Parameter group HK1_3 represented a zone where the relatively steep hydraulic gradient indicated potentially lower hydraulic conductivity. Parameter group HK1_2 represented a zone where the hydraulic conductivity was assumed to be between the values in zones HK1_1 and HK1_3. Areas near streamflow loss zones included subzones with large hydraulic conductivity that were assigned values by using a multiplier of five times the hydraulic conductivity in each respective zone. These small areas are near streams with substantial streamflow recharge where solution and collapse features cause increased permeability.

Hydraulic conductivity for layer 2 was represented as the product of the hydraulic conductivity of layer 1 times a multiplier. The multiplier was set as 0.1 for most of the active cells in layer 2. The multiplier for active cells in layer 2 near the western boundary was set as 0.5 (fig. 18); this higher multiplier accounted for the enhanced permeability in layer 2 near the outcrop area of the Minnelusa hydrogeologic unit resulting from fracturing, weathering, and dissolution.

Hydraulic properties for active cells in layer 3 were represented with seven parameters named HK3_1, HK3_2, HK3_3, HK3_4, and HK3_5, HK3_6, and HK3_7 (fig. 19). Most of the discharge from Jackson-Cleghorn Springs represents a

regional discharge point from the Madison hydrogeologic unit (Anderson and others, 1999), and substantially enlarged solution openings probably exist in the vicinity of the spring. The zone surrounding Jackson-Cleghorn Springs (fig. 19) was grouped in parameter HK3_6. Parameter group HK3_7 represented a zone of large hydraulic conductivity in the general areas of conduit flow paths from Spring Creek (Putnam and Long, 2007; Long and others, 2008) and the area surrounding zone HK3_6. Parameter group HK3_4 represented the western area of Boxelder Creek that had a relatively gradual hydraulic gradient. Parameter group HK3_5 represented the areas near the outcrop that were assumed to have lower hydraulic conductivity than the areas near streams. Parameter group HK3_5 also represented other areas where the hydraulic gradient transitioned from gradual to steep. Parameter group HK3_3 represented an area where the hydraulic gradient was much steeper than for the surrounding areas and hydraulic conductivity was assumed to be substantially lower. Parameter group HK3_1 represented a high transmissivity zone described by Downey (1986). The horizontal hydraulic conductivity for the parameter group HK3_1 area included anisotropy represented by parameter HANI3_2. The hydraulic conductivity for parameter group HK3_1 was greater in the direction of model rows than in the direction of model columns. Parameter group HK3_2 represented the remaining areas that had relatively gradual hydraulic gradients.

Additional information was available to characterize hydraulic conductivity in the detailed study area (fig. 20). Wells completed in the Madison hydrogeologic unit that were used as the basis for the characterization are listed in table 8. Hydraulic conductivity parameter groups HK3_4, HK3_6, and HK3_7 in model layer 3 represented areas that probably include substantial solution enlargement. Parameter group HK3_4 included wells 32, 35, 36, and 40 (fig. 20) where dye was detected following injection of dye in Boxelder Creek (Greene, 1999). The relatively large recharge from Boxelder Creek and the dispersed detection of dye indicated the likelihood of a relatively large hydraulic conductivity. This zone also contains structural features that could have contributed to enlargement of solution openings in the Madison hydrogeologic unit. A small area south of this zone that includes well 44 was represented with parameter group HK3_5 because of a steep hydraulic gradient in the area. The difference in hydraulic head between wells 40 and 44, which are about 2,000 ft apart, was about 100 ft (table 21).

Parameter group HK3_7 represented an area that was assumed to have relatively high hydraulic conductivity because of preferential flow paths, structural features, and relatively gradual hydraulic gradient. The parameter zone includes a potential flow path from the Spring Creek loss zone towards Jackson-Cleghorn Springs. Dye injected in Spring Creek was detected at sites CRO, NON, 66 (HH), and NAY (Putnam and Long, 2007). Analysis of a time-series of oxygen isotope values for Spring Creek streamflow recharge and water samples from site 71 (HR-2), also indicated a potential flow path from Spring Creek toward site 71 (Long and Putnam,

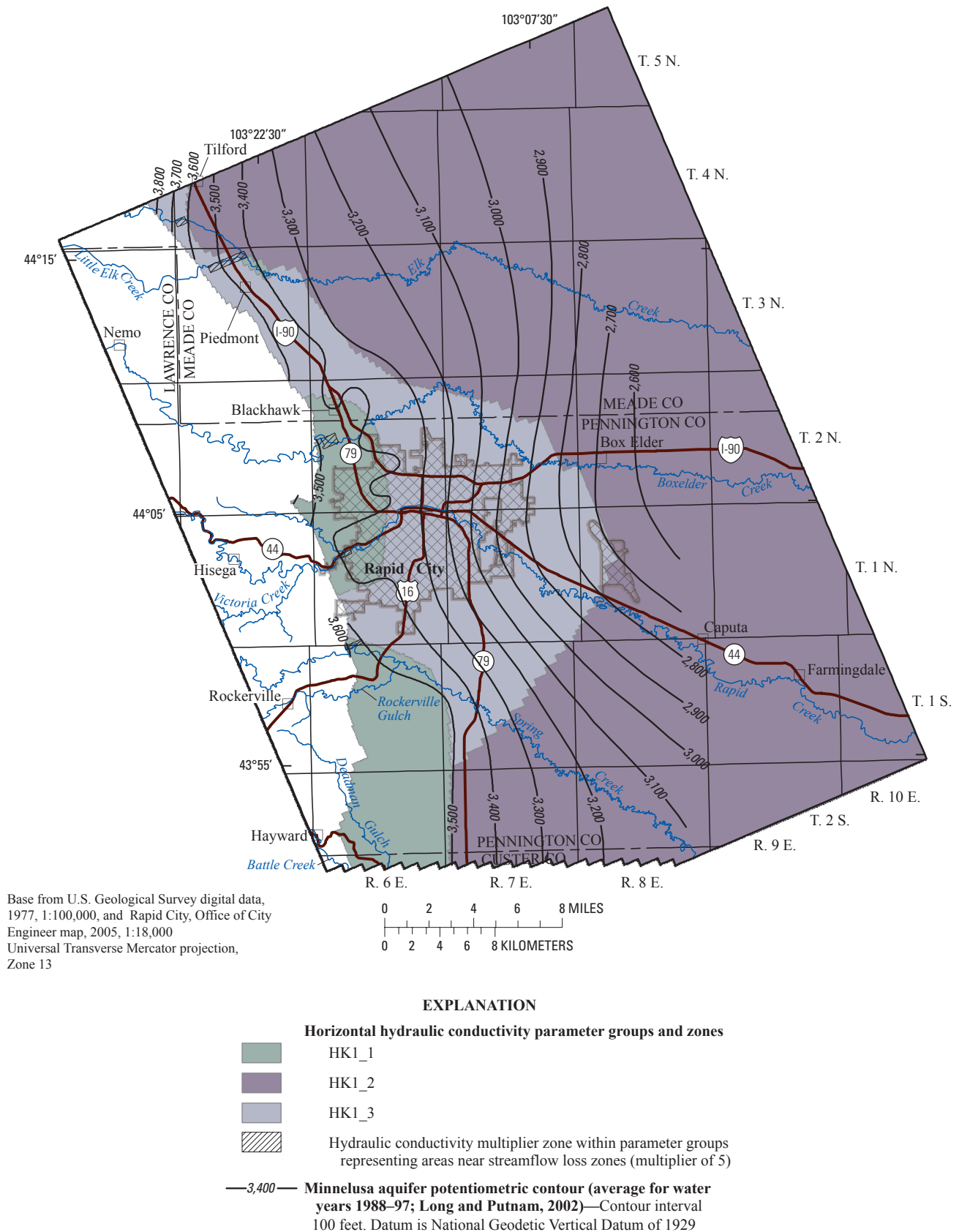


Figure 17. Horizontal hydraulic conductivity parameter groups and zones for model layer 1.

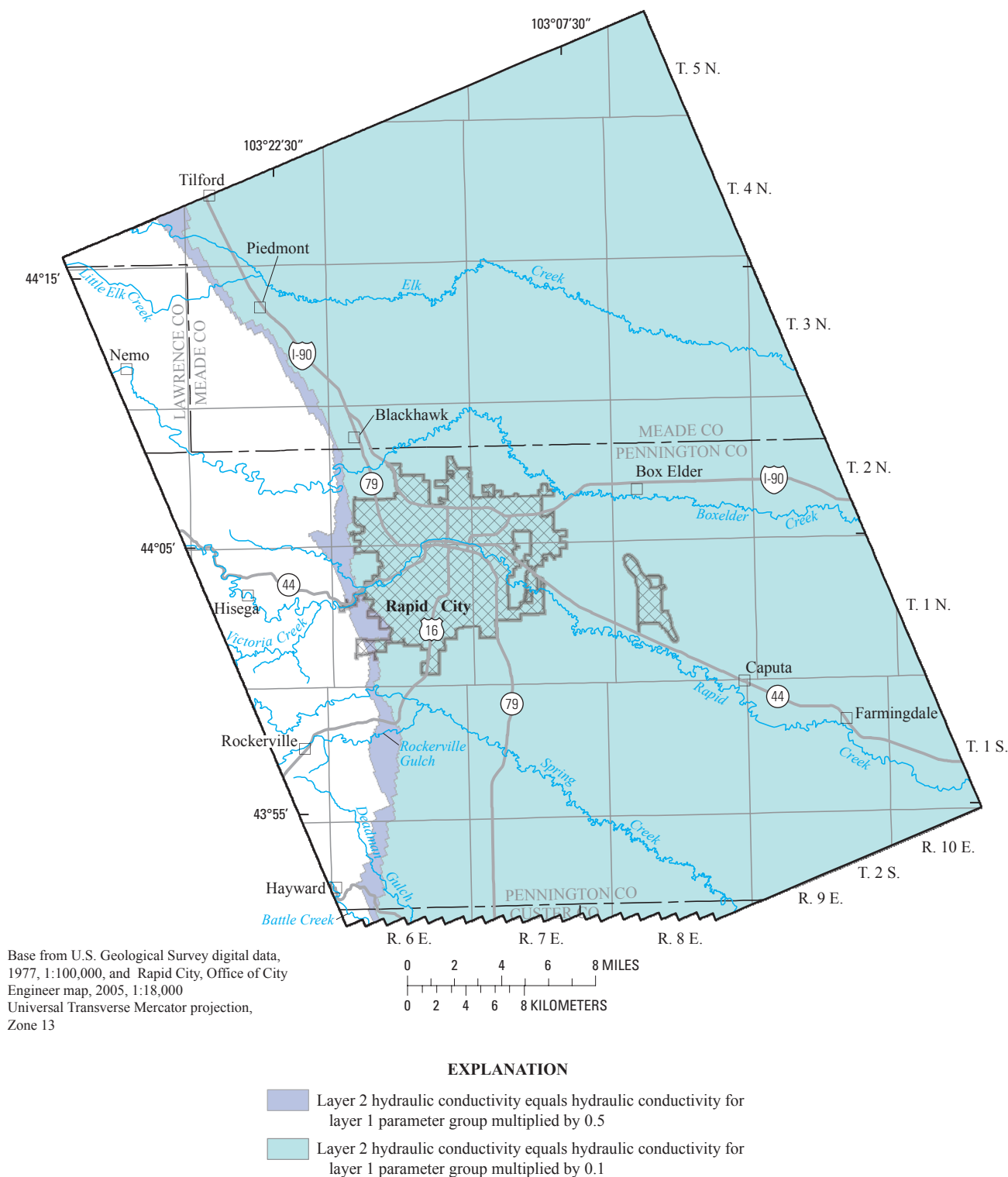


Figure 18. Horizontal hydraulic conductivity multiplier zones for model layer 2.

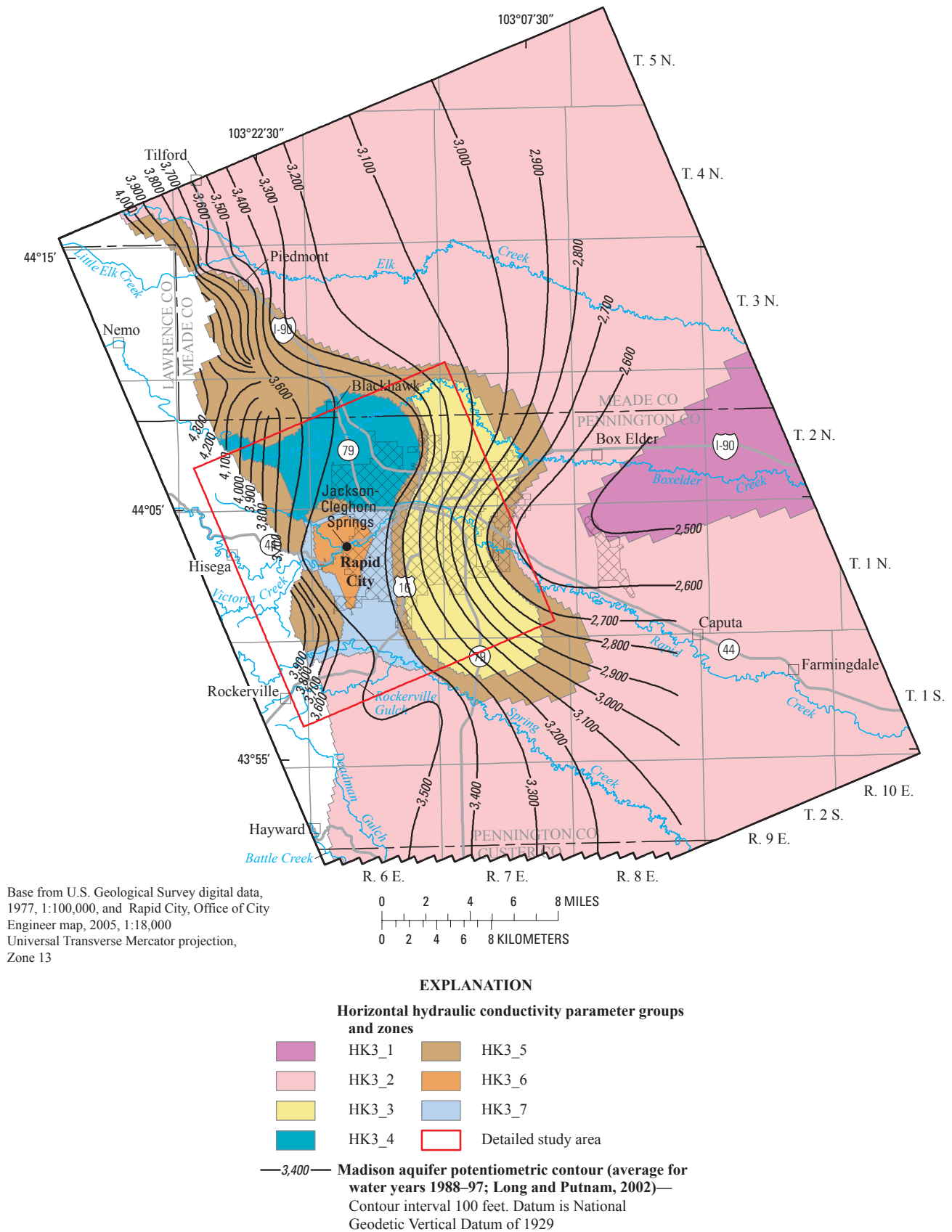
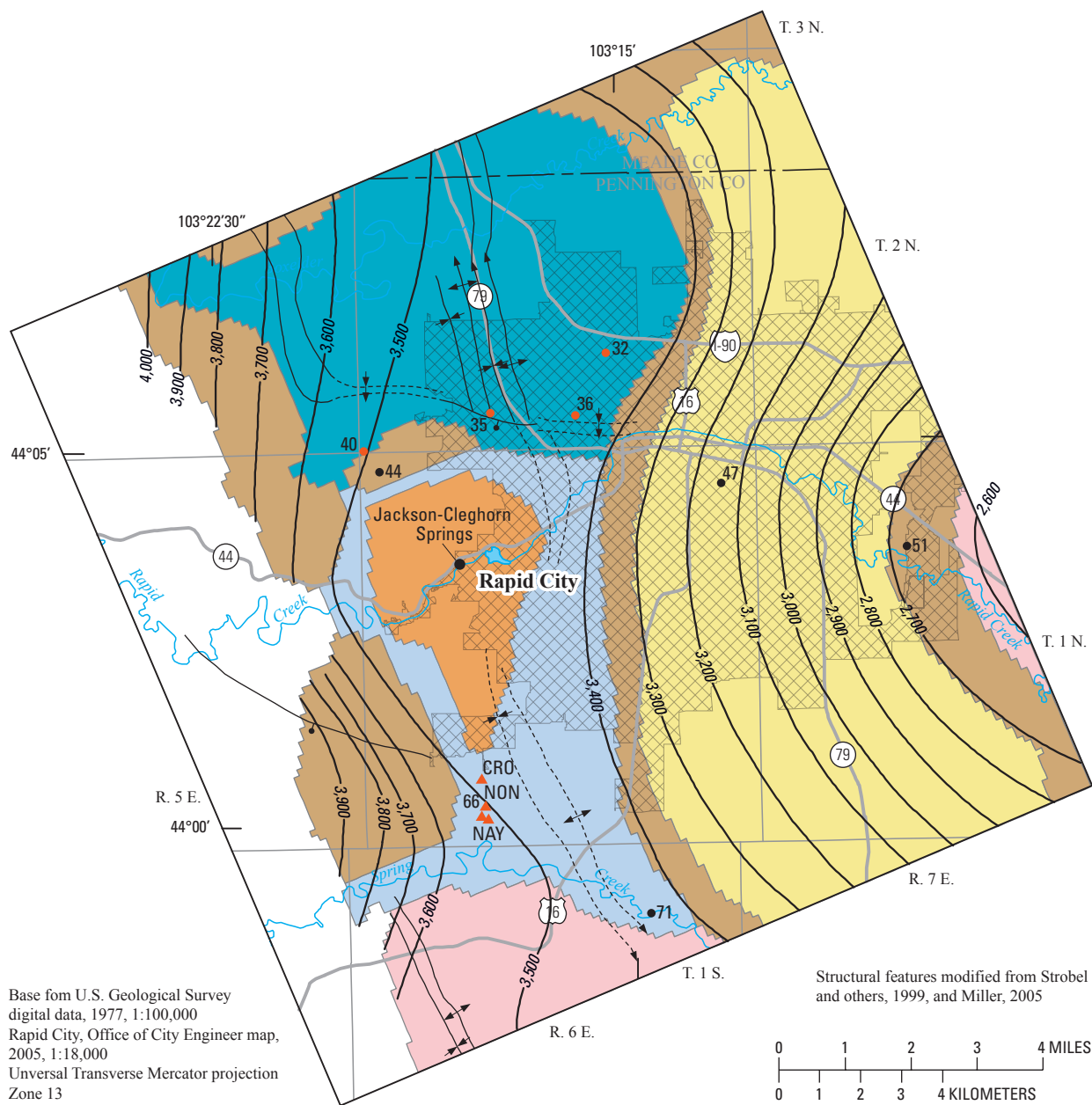


Figure 19. Horizontal hydraulic conductivity parameter groups and zones for model layer 3.



EXPLANATION

Horizontal hydraulic conductivity parameter group and zone



—3,600— Madison aquifer potentiometric contour (average for water years 1988–97; Long and Putnam, 2002)—Contour interval 100 feet. Datum is National Geodetic Vertical Datum of 1929

40 Well where dye was detected when dye was injected in Boxelder Creek (Greene, 1999) and number

NON Well where dye was detected when dye was injected in Spring Creek (Putnam and Long, 2007) and identifier

71 Selected observation well completed in Madison hydrogeologic unit and number

Fault—Dashed where approximated. Bar and ball on downthrown side

Anticline—Showing trace of axial plane and direction of plunge. Dashed where approximated

Syncline—Showing trace of axial plane and direction of plunge. Dashed where approximated

Monocline—Showing trace of axial plane. Dashed where approximated

Figure 20. Horizontal hydraulic conductivity parameter groups and zones for model layers 3 in the detailed study area.

Table 8. Sites used as basis for delineation of hydraulic conductivity parameter zones for model layer 3.

[Site numbers from table 20 in Long and Putnam (2002). Other identifier from Long and Putnam (2002) unless otherwise noted. NA, not applicable; gal/min, gallons per minute]

Site number (fig. 20)	Station identification number	Other identifier	Description
40	440500103193601	WT-2	Dye injected in Boxelder Creek was detected in well.
44	440441103193301	NA	Hydraulic head measurements indicate an area of steep hydraulic gradient.
32	440612103152001	RC-10	Dye injected in Boxelder Creek was detected in well; approximate well yield of 1,790 gal/min (Anderson and others, 1999).
35	440526103173001	RC-6	Dye injected in Boxelder Creek was detected in well; approximate well yield of 770 gal/min (Anderson and others, 1999).
36	440523103155701	BHPL	Dye injected in Boxelder Creek was detected in well.
47	440427103131701	RC-7	Approximate well yield of 150 gal/min (Anderson and others, 1999). Hydraulic head indicates steep hydraulic gradient.
51	440334103095601	RV-4	Hydraulic head measurements indicate an area of steep hydraulic gradient.
NA	440037103174401	CRO ¹	Dye injected in Spring Creek was detected in well.
NA	440017103174301	NON ¹	Dye injected in Spring Creek was detected in well.
66	440004103174001	HH	Dye injected in Spring Creek was detected in well.
NA	440007103175301	NAY ¹	Dye injected in Spring Creek was detected in well.
71	435851103143501	HR-2 ¹	Analysis of stable isotope tracers indicates conduit flow from Spring Creek (Long and Putnam, 2004).

¹From Putnam and Long, 2007.

2004; Long and others, 2008). Parameter zone HK3_7 was extended toward site 71 and included the anticline-syncline structural feature oriented towards Jackson-Cleghorn Springs. Anderson and others (1999) estimated that more than 50 percent of the discharge from Jackson-Cleghorn Springs originated from the area south of Jackson-Cleghorn Springs.

Potentiometric contours for the Madison aquifer (Long and Putnam, 2002) show the relatively steep hydraulic gradient in the area represented by parameter group HK3_3 (fig. 20). The decline in hydraulic head from site 36 to site 51 was about 800 ft (table 21) in approximately 5 miles.

Similarly to hydraulic conductivity for layer 2, hydraulic conductivity for layer 4 was set as 0.1 times that of layer 3. The multiplier for active cells in layer 4 near the western boundary was set as 0.5 times those in layer 3 (fig. 21); this higher multiplier accounted for enhanced permeability near the outcrop area of the Madison hydrogeologic unit in layer 4.

Horizontal hydraulic conductivity for layer 5 was represented by parameter group HK5_1. The active cells in model layer 5 (fig. 22) facilitated the distribution of flow to layer 4 from the underlying Deadwood aquifer.

Vertical Hydraulic Conductivity

The vertical hydraulic conductivity for model layer 2 was important in the model because it controls the hydraulic connection between the Minnelusa and Madison hydrogeologic units. Layer 2 represented the lower part of the Minnelusa hydrogeologic unit, which because of shale layers had a lower vertical hydraulic conductivity than other model layers. Vertical hydraulic conductivity determined from aquifer tests (Greene, 1993; Long and Putnam, 2002) conducted in the western part of the model area (fig. 23) ranged from 0.0053 to 2.7 feet per day (ft/d). Regional estimates of vertical conductivity for the lower part of the Minnelusa hydrogeologic unit are several orders of magnitude smaller (Long and Putnam, 2002). Collapse features related to dissolution of gypsum (hydrated calcium sulfate) in the Minnelusa hydrogeologic unit probably has resulted in greater vertical hydraulic conductivity on the western side of the model area than on the eastern side. A transition zone of increasing sulfate concentrations from west to east in the Minnelusa hydrogeologic unit (Williamson and Carter, 2001) provides an estimate of the

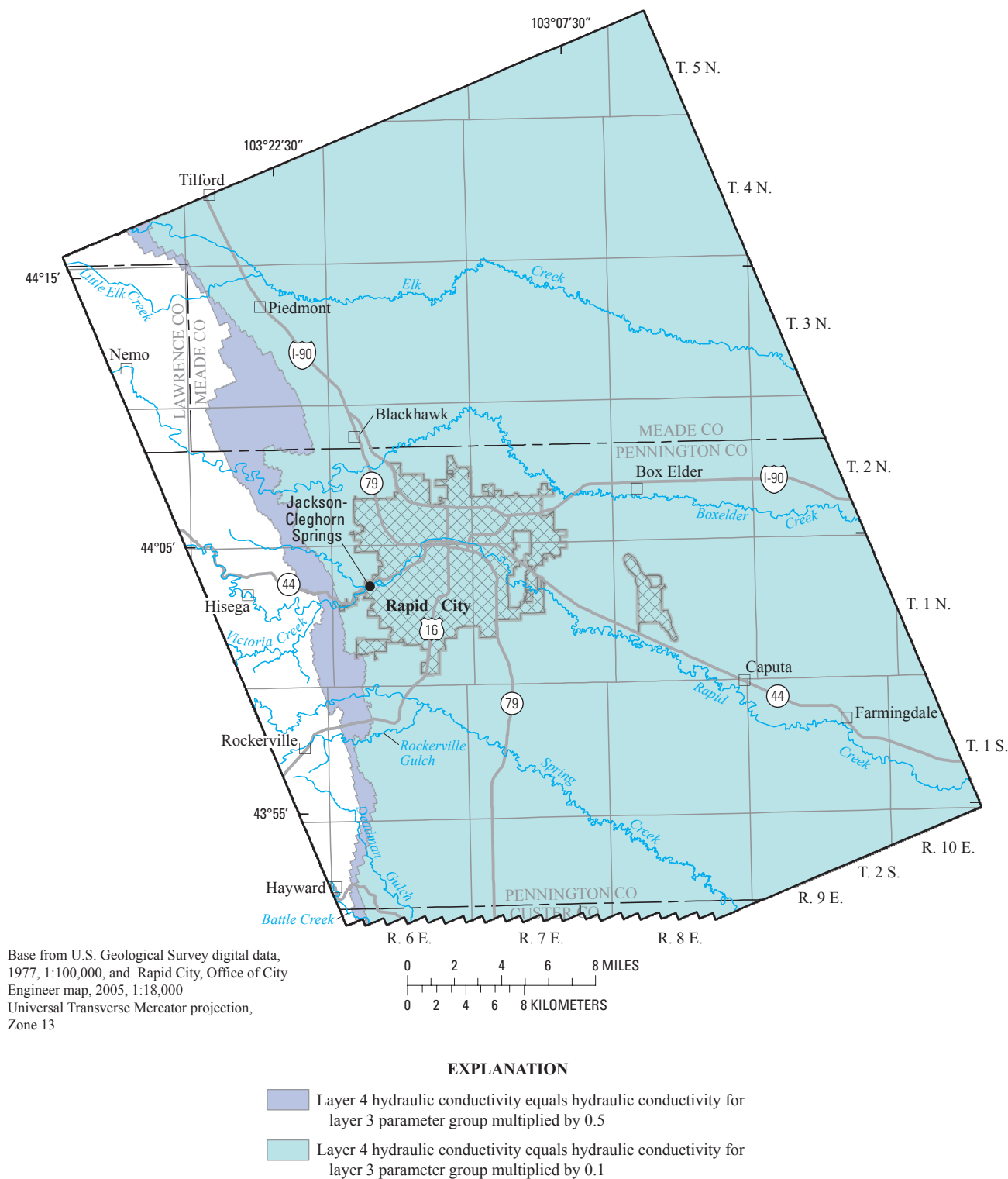


Figure 21. Horizontal hydraulic conductivity multiplier zones for model layer 4.

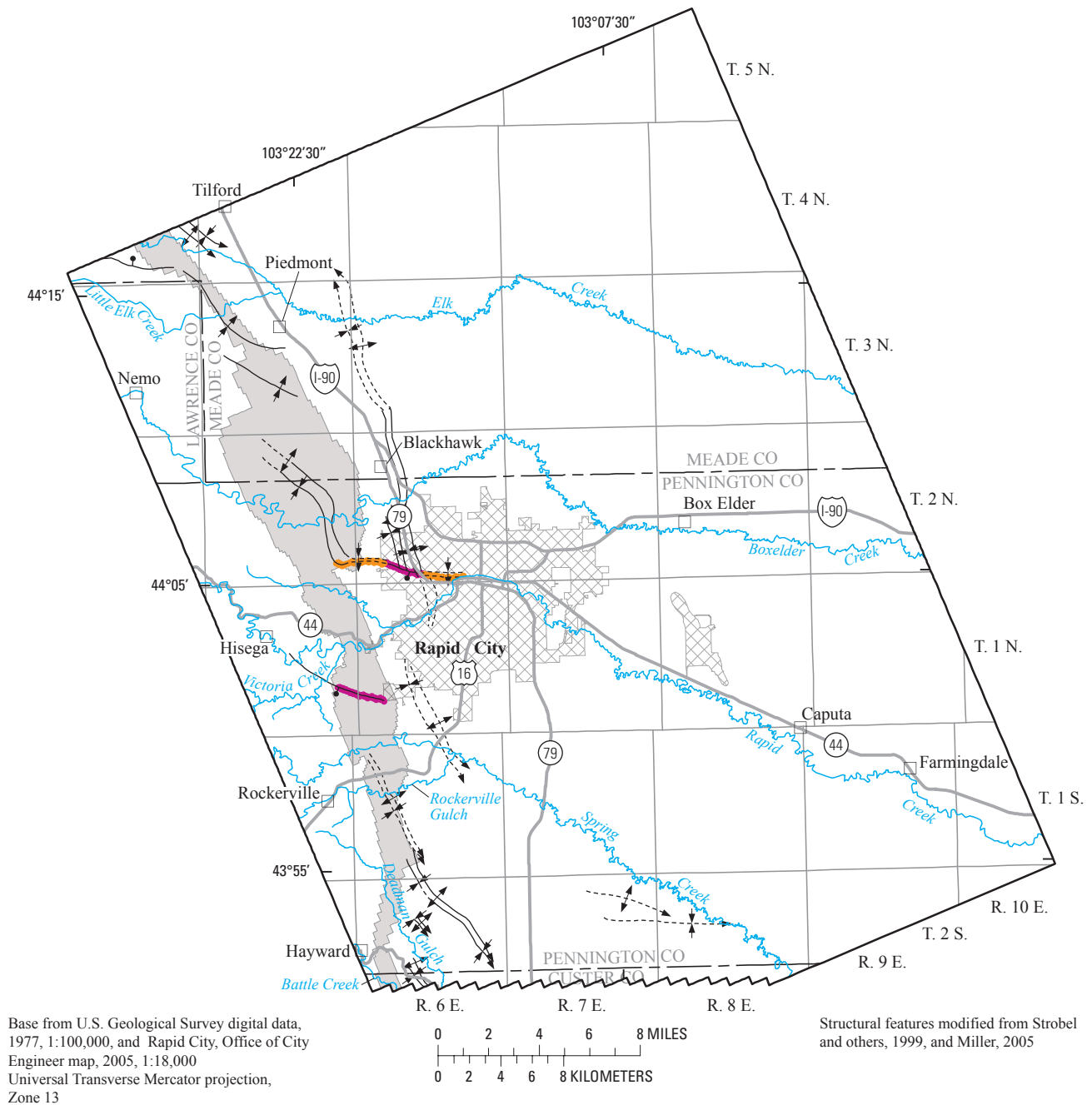


Figure 22. Horizontal hydraulic conductivity parameter group for model layer 5 and horizontal flow barriers for layers 1 to 5.

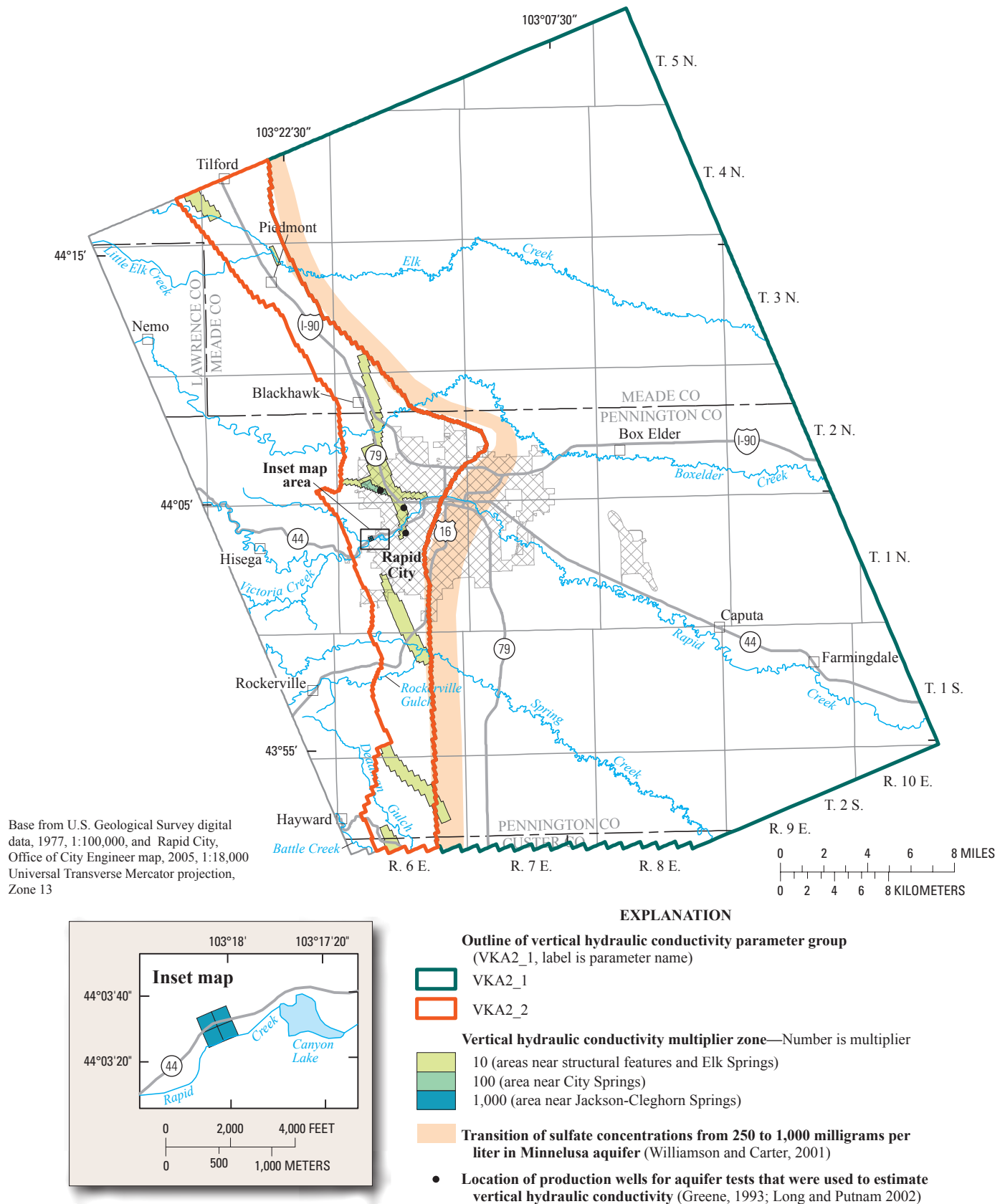


Figure 23. Vertical hydraulic conductivity parameter groups and zones for model layer 2, transition zone for sulfate concentrations in Minnelusa hydrogeologic unit, and location of aquifer tests in Madison aquifer.

extent of this area of greater vertical hydraulic conductivity (fig. 23); areas to the west of the transition zone indicate areas where gypsum has been dissolved in the Minnelusa hydrogeologic unit and collapse features exist, and areas within the zone indicate areas of active dissolution of gypsum. The transition zone of increasing sulfate concentrations was used as the basis for dividing the vertical hydraulic conductivity for layer 2 into parameter groups VKA2_1 and VKA2_2 (fig. 23), with group VKA2_2 having higher vertical hydraulic conductivity than group VKA2_1. Parameter group VKA2_2 includes multiplier zones of a factor of 10 near structural features and Elk Springs. The area near City Springs had a multiplier of 100 times the vertical hydraulic conductivity of the surrounding area (fig. 23) on the basis of upward flow from the Madison hydrogeologic unit through the Minnelusa hydrogeologic unit. The vertical hydraulic conductivity multiplier at Jackson-Cleghorn Springs was set as 1,000 because of the large volume of upward flow. Additional information about the potential variability in vertical hydraulic conductivity can be found in Long and Putnam (2002).

The vertical hydraulic conductivity of the other model layers was assumed to be larger than that of layer 2 and was represented by ratios of vertical to horizontal hydraulic conductivity. The ratios for layers 1, 3, 4, and 5 were represented by a single parameter for each layer (VANI1_1, VANI3_1, VANI4_1, and VANI5_1, respectively).

Horizontal Flow Barriers

Horizontal flow barriers restrict lateral flow between adjacent cells in a layer and are simulated by a parameter called the “barrier hydraulic characteristic,” which is the value of the hydraulic conductivity of the barrier divided by the width of the barrier. Horizontal flow barriers were used to simulate the hydraulic effect of the City Springs fault, the City Springs monocline, and the Victoria Creek fault (fig. 22). The offset along the faults placed impermeable areas against permeable areas with increased solution openings probable along the axis of the faults and monoclines. These barriers were assumed to extend vertically through all model layers having active cells in the area. Values used for the hydraulic characteristic assigned to the barriers by model layer (table 9) ranged from 0.0005 to 0.01 foot per day per foot. The Victoria Creek fault was present only in layers 3 and 4.

In layers 1 and 2, the City Springs monocline was interpreted to have little effect on horizontal flow in the sandstone layers of the Minnelusa hydrogeologic unit; therefore, the barrier hydraulic characteristic was set as 0.01 foot per day per foot. The barrier hydraulic characteristic for the City Springs and Victoria Creek faults for all layers was assumed to be less than for the City Springs monocline and was assigned a value of 0.0005. The maximum offset along the City Springs fault is about 500 ft, downthrown to the south, and the maximum offset for the Victoria Creek fault is about 350 ft, downthrown to the north (Miller, 2005). The hydraulic characteristic for the barrier representing the City Springs monocline in layers 3 and 4 was assigned a value of 0.001 to approximate the effects of preferential solution openings along the axis of the monocline. The hydraulic characteristic for the barriers in layer 5 were estimated as 0.01 where cells in layer 5 were active.

Storage Properties

A general definition of specific yield for the unconfined part of the Minnelusa and Madison hydrogeologic units (figs. 2 and 3) is the ratio of the volume of water that the rock, after being saturated, will yield by gravity to the volume of the rock (Lohman and others, 1972). A definition of storage coefficient for the confined part of the aquifer is the volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in hydraulic head (Lohman and others, 1972). Smaller changes in storage occur in a confined aquifer compared to an unconfined aquifer because storage changes in a confined aquifer result from the slight expansion or contraction of the aquifer material and water due to hydraulic head changes.

The unconfined areas (fig. 24) for the Minnelusa and Madison hydrogeologic units were delineated for average conditions for WY 1988–97 by Long and Putnam (2002). In MODFLOW–2000, a layer can be simulated as confined (constant saturated thickness), unconfined (variable saturated thickness), or convertible, in which cells convert from confined to unconfined depending on hydraulic head and layer thickness. To maintain numerical stability, convertible layers were not used in this simulation, and all layers were simulated as confined, except that storage coefficients for unconfined areas were set equal to estimated values of specific yield. Thus, the compromise between numerical stability,

Table 9. Horizontal barrier hydraulic characteristic for City Springs fault, City Springs monocline, and Victoria Creek fault.

[Hydraulic characteristic is hydraulic conductivity of the barrier divided by the width of the barrier. NA, not applicable]

Barrier name	Barrier hydraulic characteristic (feet per day per foot)				
	Model layer				
	1	2	3	4	5
City Springs fault	0.0005	0.0005	0.0005	0.0005	NA
City Springs monocline	.01	.01	.001	.001	0.01
Victoria Creek fault	NA	NA	.0005	.0005	NA

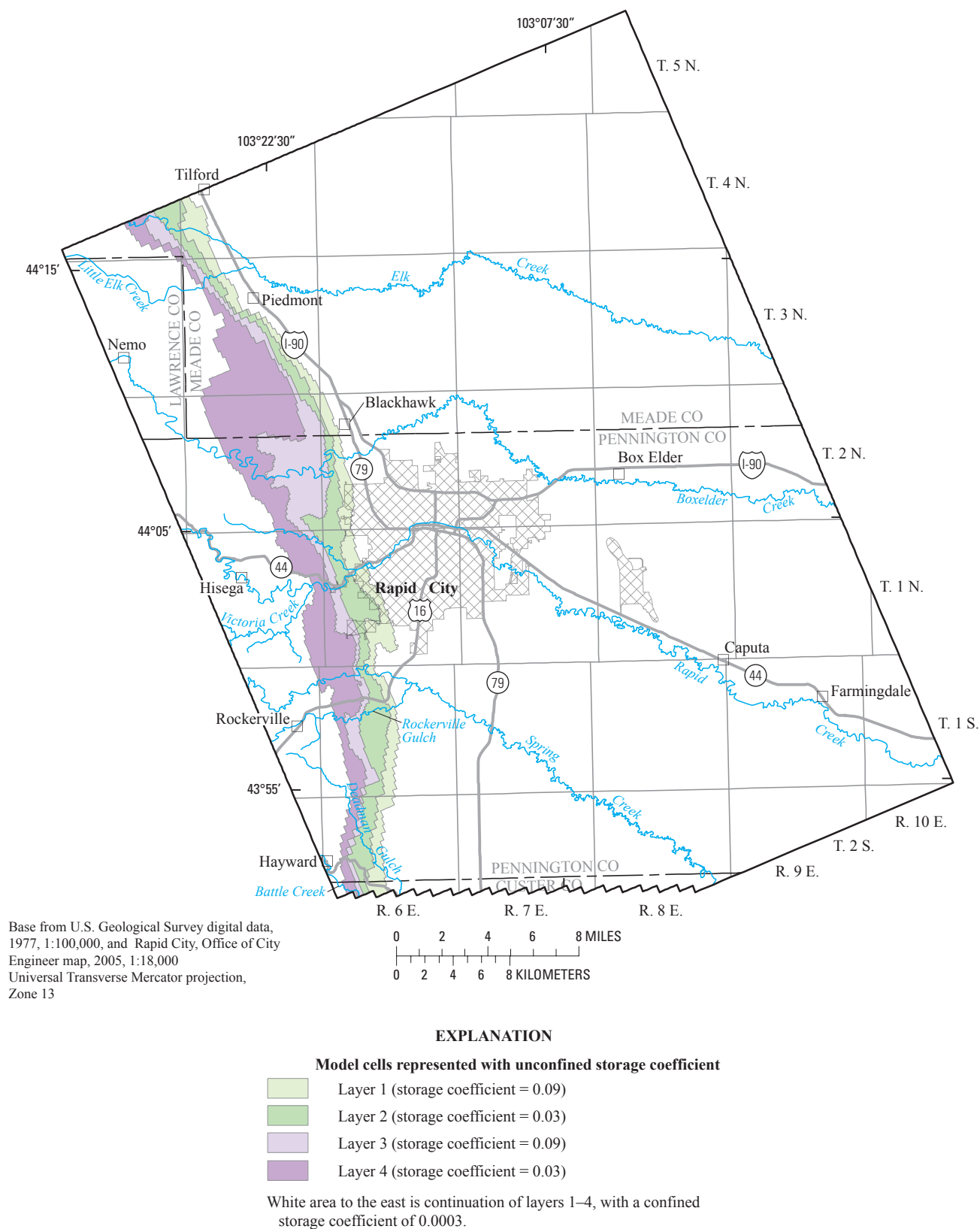


Figure 24. Model cells with storage represented by unconfined storage coefficient.

particularly for inverse modeling, and physical accuracy was to assume that hydraulic conductivity did not change with hydraulic head in unconfined areas. Also, the actual unconfined areas move up and down dip as water levels fluctuate, but their general shape and total areal coverage change negligibly. Therefore, unconfined areas were simulated as spatially fixed. Unconfined areas for layers 1–4 were represented in the simulation by assigning storage coefficient values that were much higher than for the confined areas to the east. As estimated by Long and Putnam (2002), the storage coefficient in unconfined areas (specific yield) was 0.09 for the upper parts of the hydrogeologic units (layers 1 and 3) and 0.03 for the lower parts of the hydrogeologic units (layers 2 and 4). The storage coefficient for all confined areas was 0.0003.

Specific storage, which is the storage coefficient divided by the aquifer thickness, is the required storage property in MODFLOW–2000. Therefore, the storage coefficients were divided by arrays of thickness for each model layer to create the required specific storage input.

Steady-State Calibration

The steady-state model was calibrated by using a hypothetical simulation of estimated average conditions with no changes in storage. A sensitivity analysis showed model parameters that were most sensitive to steady-state observations, and these parameters were optimized with inverse modeling as described in the following section. Comparisons of simulated steady-state hydraulic head values and flows to observed values and estimated flows were used to calibrate the model. The observed hydraulic head values used for steady-state calibration were based on 10-year average hydraulic head values for WY1988–97 for 176 wells completed in the Minnelusa hydrogeologic unit and 76 wells completed in the Madison hydrogeologic unit (Long and Putnam, 2002); selected information for these wells is presented in tables 20 and 21 in the “Supplemental Information” section. Forty interpolated hydraulic head values, 20 for model layer 1 and 20 for model layer 3 (table 22 in “Supplemental Information” section), were added to constrain the model in locations where information was sparse. These hydraulic head values were interpolated from the estimated average potentiometric contours for WY 1988–97 for the Minnelusa and Madison aquifers (Long and Putnam, 2002). The observed (measured) hydraulic head values (figs. 25 and 26) predominantly were available near the outcrops of the Minnelusa and Madison hydrogeologic units, whereas the interpolated hydraulic heads were used in areas east of the outcrops where measured water-level data were sparse. Estimated springflow and regional outflow (Anderson and others, 1999; Long and Putnam, 2002) also were used for steady-state flow calibration.

Standard deviation of measurement error (SDME) was used in the MODFLOW–2000 observation process for weighting observations and estimates for inverse modeling, in which the weight assigned to an observation or estimate was the inverse of the implied accuracy. The implied accuracy was 1 ft for observation wells with surveyed altitudes and continuous water-level records, 5 ft for wells with multiple measurements, and 10 ft for wells with water levels estimated from well driller’s completion reports or potentiometric surfaces. The implied accuracy was 0.1 ft³/s for discharge from Jackson-Cleghorn Springs, City Springs, and Deadwood Avenue Springs, and 0.2 ft³/s for discharge from other springs. Implied accuracy for discharge at the eastern boundary was 0.1 ft³/s.

Sensitivity Analysis

Composite-scaled sensitivity is the square root of the sum of squared scaled sensitivity values for all observations divided by the number of observations. Composite-scaled sensitivity indicates the information content of all of the observations for the estimation of the parameter (Hill and others, 2000). The composite-scaled sensitivities for hydraulic properties represented as parameters (table 10) indicated the parameters most sensitive to steady-state observations. Parameters with composite-scaled sensitivities greater than 1.2 were selected for optimization with inverse modeling. If some parameters have composite-scaled sensitivities that are less than about 0.01 times the largest composite-scaled sensitivity, the regression has trouble converging (Hill, 1998). Optimization for parameters that had composite-scaled sensitivity values less than 1.2 did not converge with reasonable values for the less-sensitive parameters.

The sensitivity of a parameter is related to the amount of information that is available to constrain the parameter. The ratio of the smallest to the largest composite-scaled sensitivities for the 12 parameters selected for optimization with parameter estimation (fig. 27) was 0.06 (ratio of sensitivities of DWSPR to HK3_2). The model was sensitive to most horizontal hydraulic conductivity parameters except parameter HK3_3, which represented the area in model layer 3 where the hydraulic gradient was very steep. Sensitivity to vertical hydraulic conductivity parameters was low except for VKA2_2 (fig. 23 and table 10), which represents an area in model layer 2 that extends east about 2 to 4 mi from the outcrop of the Minnelusa hydrogeologic unit. The model was substantially more sensitive to conductance for the cells in layer 3 that represented Jackson-Cleghorn Springs (CLEGmdsn) than conductance for other springs. The model was relatively insensitive to conductance parameters for head-dependent cells representing the eastern boundary (MnlEB and MdsnEB; table 10).

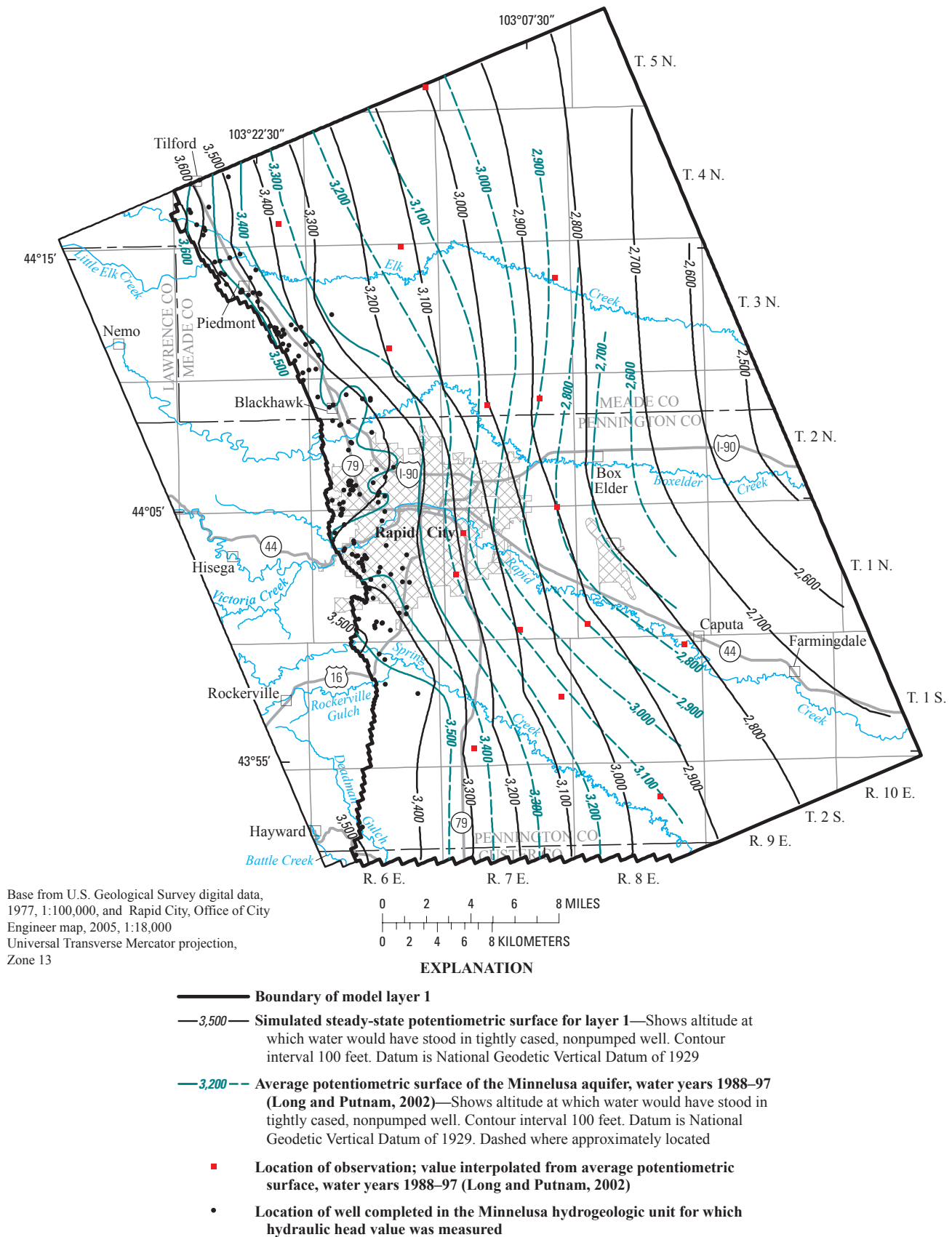
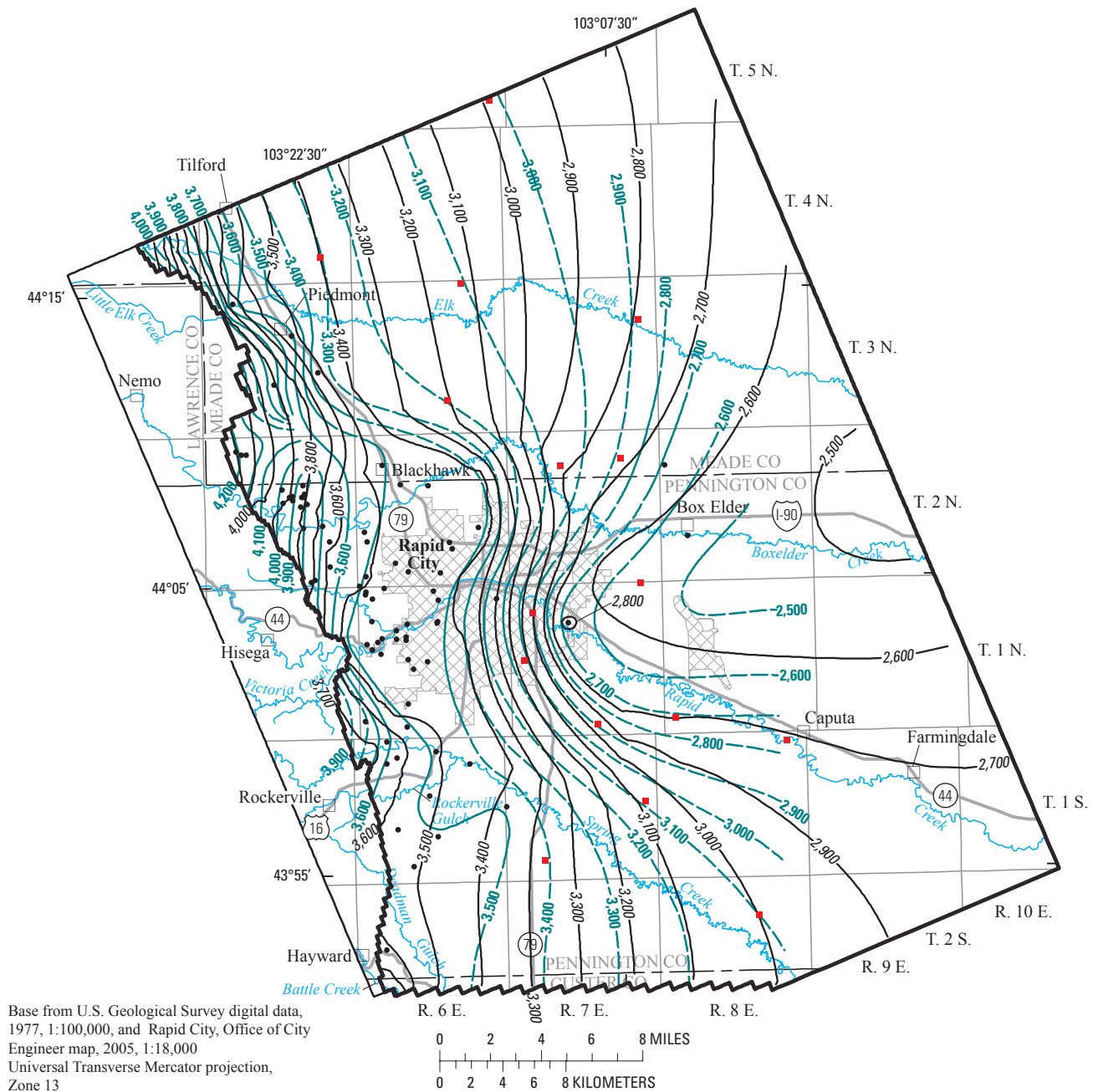


Figure 25. Comparison of simulated steady-state potentiometric surface for model layer 1 to the average potentiometric surface of the Minnelusa aquifer, water years 1988–97.



- EXPLANATION**
- Boundary of model layer 4
 - 3,100 — Simulated steady-state potentiometric surface for layer 3 and for layer 4 where layer 3 was inactive—Shows altitude at which water would have stood in tightly cased, nonpumped well. Contour interval 100 feet. Datum is National Geodetic Vertical Datum of 1929
 - 2,900 — Average potentiometric surface of the Madison aquifer, water years 1988–97 (Long and Putnam, 2002)—Shows altitude at which water would have stood in tightly cased, nonpumped well. Contour interval 100 feet. Datum is National Geodetic Vertical Datum of 1929. Dashed where approximately located
 - Location of observation; value interpolated from average potentiometric surface, water years 1988–97 (Long and Putnam, 2002)
 - Location of well completed in the Madison hydrogeologic unit for which hydraulic head value was measured

Figure 26. Comparison of simulated steady-state potentiometric surface for model layer 3 and western part of layer 4 to the average potentiometric surface of the Madison aquifer, water years 1988–97.

Table 10. Composite-scaled sensitivities for hydraulic properties represented as parameters.

[Shading indicates that parameter values were estimated by inverse modeling. Composite-scaled sensitivity calculated for final parameter values]

Parameter name	Location map	Parameter description	Composite-scaled sensitivity
CLEGmdsn	Jackson-Cleghorn Springs, fig. 12	Conductance parameter is the hydraulic conductivity of the interface between the head-dependent cells times the cell area divided by the thickness	4.63
CLEGmnls	Jackson-Cleghorn Springs, fig. 12		.39
CITYSPR	City Springs, fig. 12		.19
DWSPR	Deadwood Avenue Springs, fig. 12		1.25
BESPR	Boxelder Springs, fig. 12		.15
ELKSPR	Elk Springs, fig. 11		.05
GAL	Infiltration Gallery Springs, fig. 12		.03
CLAKE	Canyon Lake, fig. 12		.16
HK1_1	Fig. 17	Horizontal hydraulic conductivity parameters for model layer 1	2.38
HK1_2	Fig. 17		16.96
HK1_3	Fig. 17		5.98
HK3_1	Fig. 19	Horizontal hydraulic conductivity parameters for model layer 3	1.29
HK3_2	Fig. 19		21.42
HK3_3	Fig. 19		1.13
HK3_4	Fig. 19		15.51
HK3_5	Fig. 19		8.63
HK3_6	Fig. 19		2.24
HK3_7	Fig. 19		5.70
HANI3_2	Fig. 19	Anisotropy for area represented by parameter group HK3_1	.03
HK5_1	Fig. 22	Horizontal hydraulic conductivity for layer 5	.83
VANI1_1	Active cells in model layer 1, fig. 8	Ratio of vertical to horizontal hydraulic conductivity for model layer 1	.01
VKA2_1	Fig. 23	Vertical hydraulic conductivity for model layer 2	.20
VKA2_2	Fig. 23		4.13
VANI3_1	Active cells in model layer 3, fig. 9	Ratio of vertical to horizontal hydraulic conductivity for model layer 3	.01
VANI4_1	Active cells in model layer 4, fig. 9	Ratio of vertical to horizontal hydraulic conductivity for model layer 4	.04
VANI5_1	Fig. 22	Ratio of vertical to horizontal hydraulic conductivity for model layer 5	.01
MnlsEB	Fig. 8	Conductance parameter is the hydraulic conductivity of the interface between the head-dependent cell at the eastern boundary times the area divided by the thickness	.01
MdsnEB	Fig. 9		.01

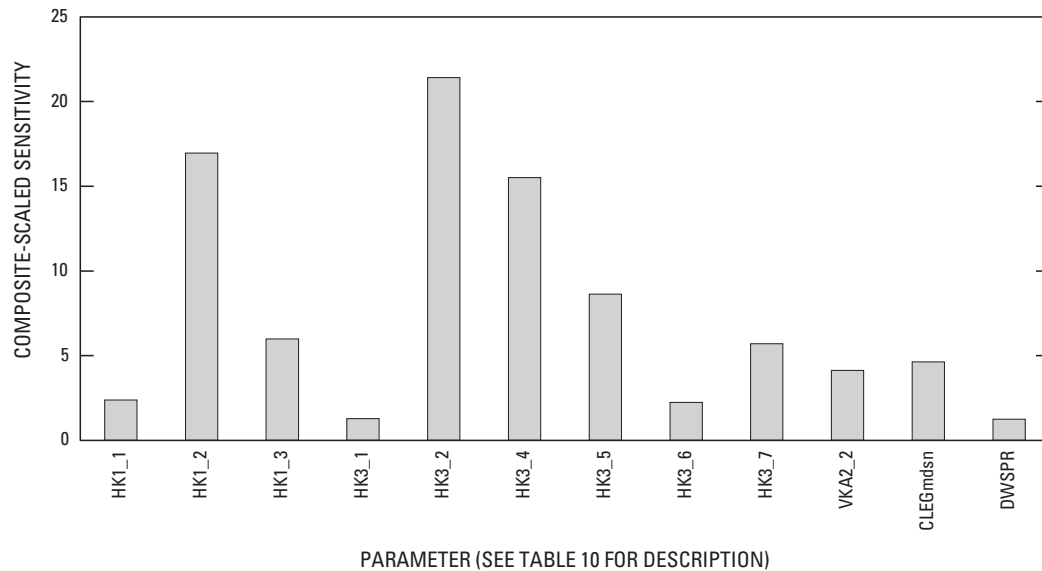


Figure 27. Composite-scaled sensitivities for parameter values that were determined with inverse modeling in steady-state simulation.

Comparison of Simulated and Observed Steady-State Hydraulic Head Values

Linear regression between the 252 simulated and observed hydraulic head values for the steady-state simulation had a coefficient of determination (R^2 value) of 0.92 with an average arithmetic difference of -7.6 ft and an average absolute difference of 37.6 ft. Most of the observed hydraulic heads used for comparison with simulated hydraulic heads were between 3,400 and 3,600 ft above NGVD 29 (fig. 28). A histogram of residuals between the 252 simulated and observed hydraulic head values (fig. 29) shows that about 65 percent of the residuals were less than ± 40 ft.

The simulated potentiometric surface for model layer 1 overlain on the average potentiometric surface for WY 1988–97 generally shows similar gradients in hydraulic head values, with the largest differences near the model boundaries (fig. 25). The differences between simulated and observed (average for WY 1988–97) hydraulic head values for the Minnelusa hydrogeologic unit in the detailed study area were within ± 30 ft for 73.7 percent of the observation wells (fig. 30). The difference between observed and simulated hydraulic head values that were not within ± 30 ft for some of the wells could be related to the effect of unknown breccia pipes or karst features that were not represented in the model.

The simulated potentiometric surface for model layer 3 matched the observed steep hydraulic gradients in eastern Rapid City (fig. 26). The differences between simulated and observed hydraulic head values for the Madison hydrogeologic unit in the detailed study area were within ± 30 ft for 76.9 percent of the observation wells (fig. 31). Wells with a

hydraulic head difference that was not within ± 30 ft generally were near the western model boundary.

Comparison of Simulated and Estimated Steady-State Flow

Total simulated steady-state springflow of 31.6 ft^3/s was similar to the sum of estimated springflow of 32.8 ft^3/s (table 11). Simulated discharge from Jackson-Cleghorn Springs was 20.5 ft^3/s compared to the estimated value of 21.6 ft^3/s . Simulated regional outflow from model layers 1 and 2 at the eastern boundary of 12.9 ft^3/s was slightly larger than the estimated value of 12.3 ft^3/s from Long and Putnam (2002). Simulated regional outflow from model layers 3 and 4 of 12.8 ft^3/s was slightly larger than the estimated value of 12.2 ft^3/s from Long and Putnam (2002).

Transient Calibration

The transient calibration was accomplished by simulating transient conditions for WY 1988–97 divided into twenty 6-month stress periods. Comparisons of simulated transient hydraulic head values and flows to observed values and estimated flows were used to calibrate the model. Simulated transient hydraulic head values were compared to observed (measured) hydraulic head values for 19 observation wells completed in the Minnelusa and Madison hydrogeologic units (fig. 32); water-level records were available for these 19 observation wells for all or parts of the transient simulation. Transient springflow was calibrated to springflow measurements

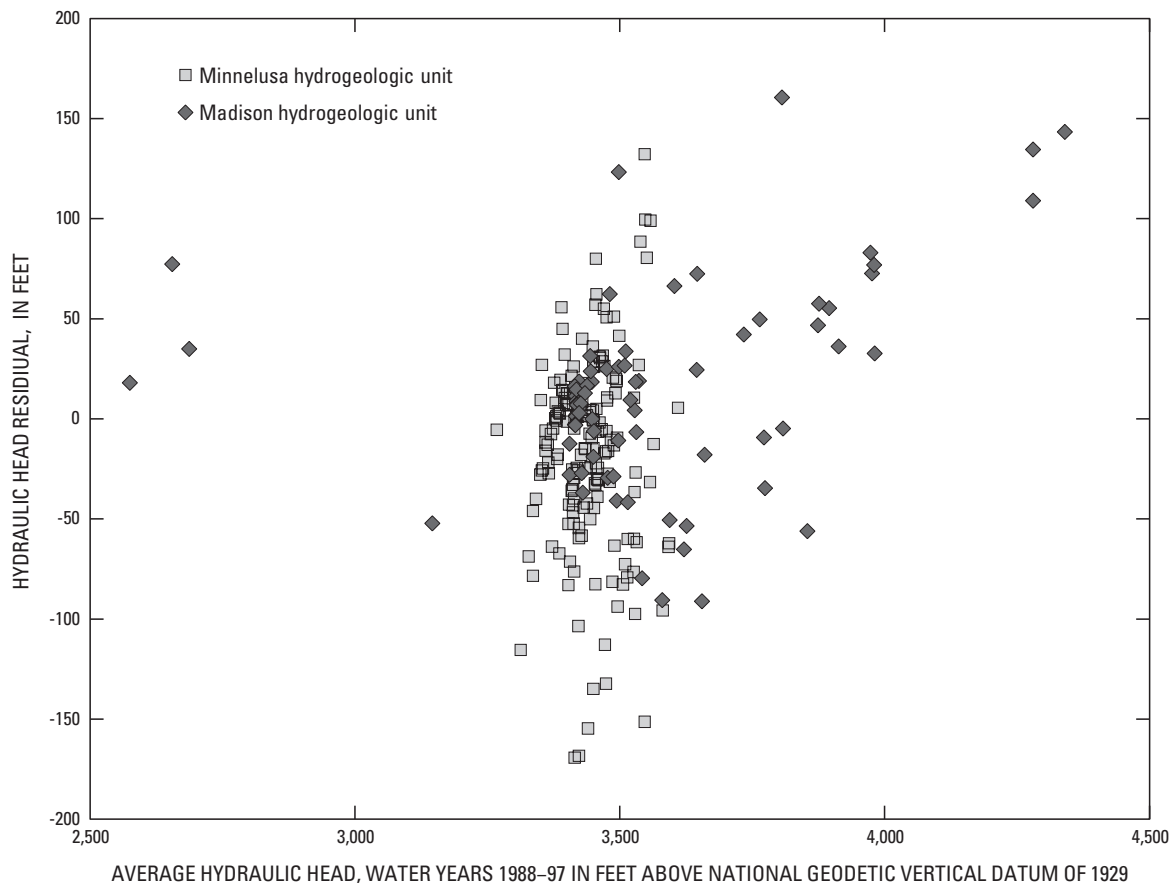


Figure 28. Residuals between simulated steady-state and observed hydraulic head values and their relation to average (water years 1988–97) hydraulic head values.

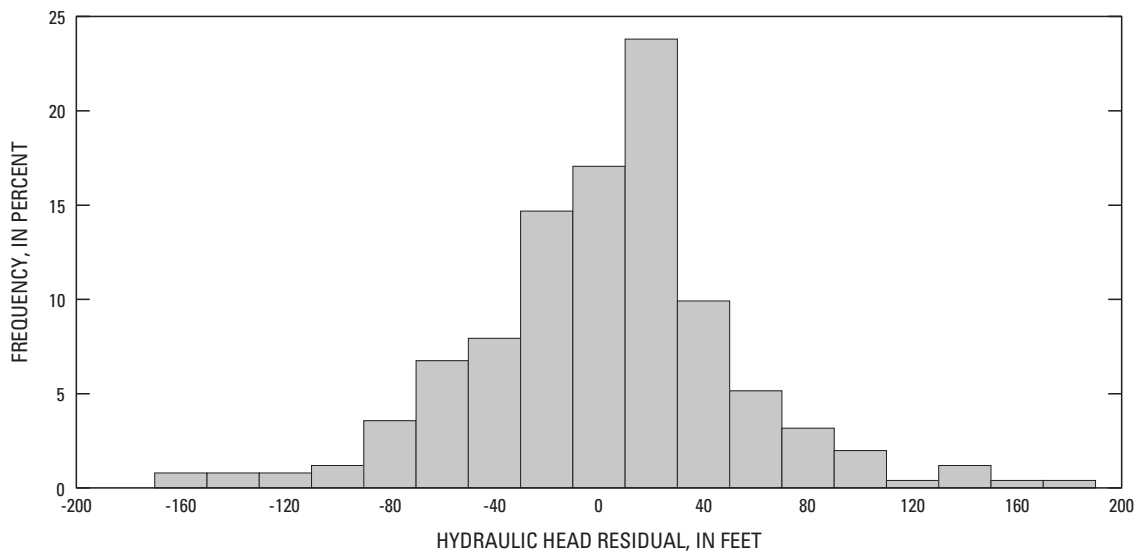
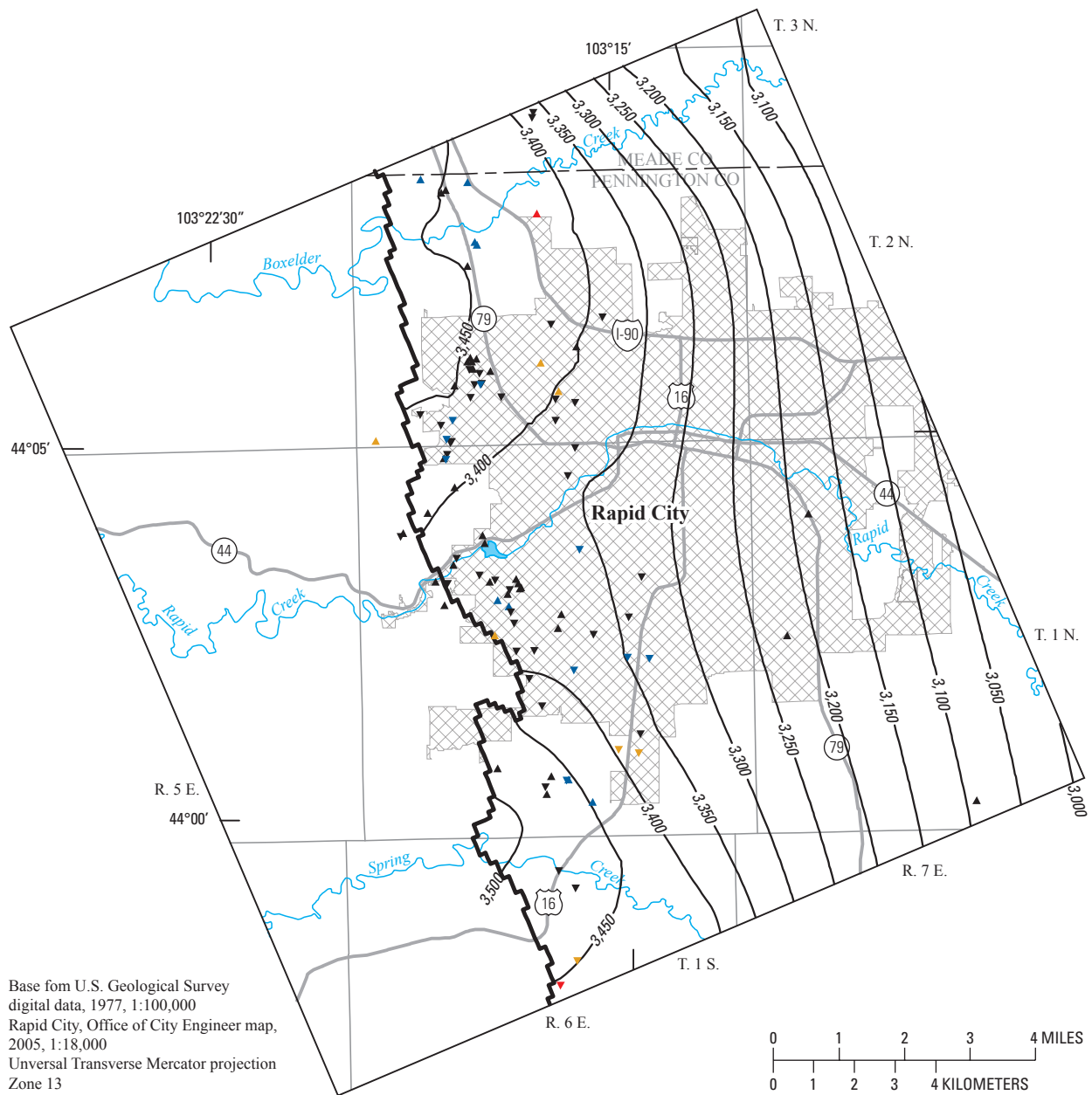


Figure 29. Histograms of residuals between simulated steady-state and observed (average for water years 1988–97) hydraulic head values.

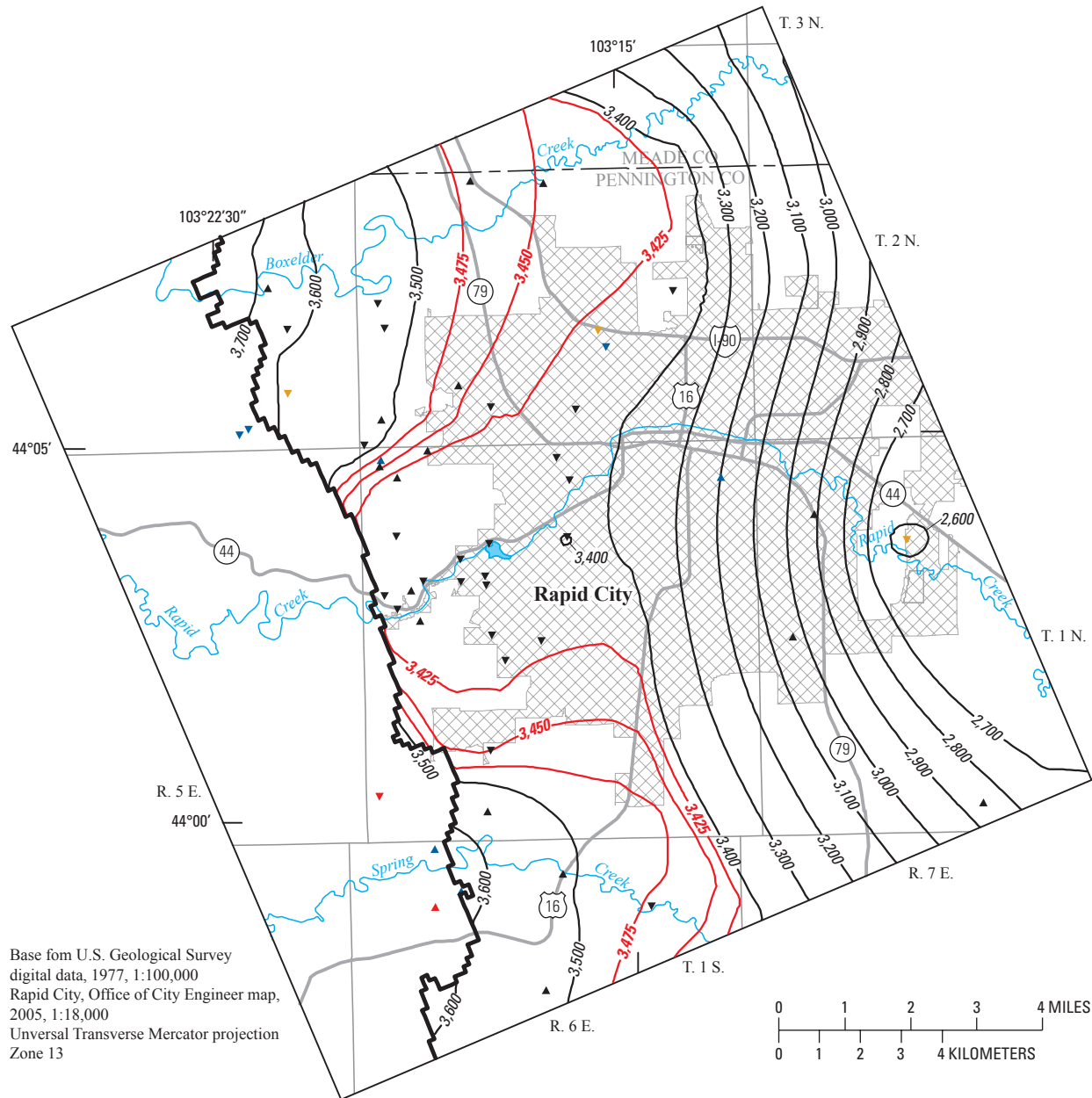


EXPLANATION

- Boundary of model layer 1
- 3,500 — Simulated steady-state potentiometric surface for model layer 1—Shows altitude at which water would have stood in tightly cased, nonpumped well. Contour interval 50 feet. Datum is National Geodetic Vertical Datum of 1929

Frequency of difference category	Difference, in feet
▲ 1.0	90 to less than 120
▲ 4.1	60 to less than 90
▲ 8.1	30 to less than 60
▲ 34.3	0 to less than 30
▼ 39.4	Less than 0 to greater than -30
▼ 9.1	-30 to greater than -60
▼ 3.0	-60 to greater than -90
▼ 1.0	-90 to greater than -120

Figure 30. Differences between simulated steady-state and observed (average for water years 1988–97) hydraulic head values for the Minnelusa hydrogeologic unit in the detailed study area.



EXPLANATION

- Boundary of model layer 3
 - 3,600 Simulated steady-state potentiometric surface for model layer 3—Shows altitude at which water would have stood in tightly cased, nonpumped well. Contour interval 100 feet. Datum is National Geodetic Vertical Datum of 1929
 - 3,475 Supplemental contours 3,425, 3,450, and 3,475
- | Frequency of difference category | Difference, in feet |
|----------------------------------|---------------------------------|
| ▲ 1.9 | 90 to less than 120 |
| ▲ .0 | 60 to less than 90 |
| ▲ 7.7 | 30 to less than 60 |
| ▲ 30.8 | 0 to less than 30 |
| ▼ 46.1 | Less than 0 to greater than -30 |
| ▼ 5.8 | -30 to greater than -60 |
| ▼ 5.8 | -60 to greater than -90 |
| ▼ 1.9 | -90 to greater than -120 |

Figure 31. Differences between simulated steady-state and observed (average for water years, 1988–97) hydraulic head values for the Madison hydrogeologic unit in the detailed study area.

Table 11. Steady-state springflow estimates and simulated springflow.

[Springflow rates in cubic feet per second]

Spring name	Steady-state springflow (from table 6)	Simulated springflow	Arithmetic difference
Jackson-Cleghorn Springs	¹ 21.6	20.5	-1.1
City Springs	² 1.7	1.4	-.3
Deadwood Avenue Springs	² 2.8	2.6	-.2
Boxelder Springs	² 1.0	1.8	.8
Elk Springs	² 3.7	4.2	.5
Infiltration Gallery Springs	¹ 1.5	.5	-1.0
Canyon Lake	³ .5	.6	.1
Total	32.8	31.6	-1.2

¹Anderson and others, 1999.²Long and Putnam, 2002.³No quantitative measurements were available; small discharge was assumed because of hydraulic connection.

or estimated springflow for seven springs (figs. 11 and 12). The estimated springflow for some of the springs represented an average discharge. Regional outflow at the eastern model boundary was calibrated to estimated average regional outflow used in the steady-state simulation. The regional outflow was assumed to be relatively constant because the boundary was about 20 mi east of most recharge, spring discharge, and pumping areas.

Comparison of Simulated and Observed Transient Hydraulic Head Values

The average absolute difference between simulated and observed hydraulic head values for the 19 observation wells ranged from 3.5 to 65.1 ft with a median value of 18.3 ft (table 12). Hydrographs of simulated and observed hydraulic head values for 18 of the wells (fig. 33) show the response of the heads to stresses. Hydrographs for the nested wells completed in both the Minnelusa and Madison hydrogeologic units at six sites are presented adjacent to each other. In general, the magnitude of changes in simulated hydraulic head values were less than the observed changes.

The changes in simulated hydraulic head values for wells 104 and 1 between stress periods are much less than observed changes. The simulated hydraulic head for well 104 does not increase as quickly as the observed hydraulic head during the wet period (late 1990s; stress periods 12–20). Wells 104 and 1 are located at the northern no-flow boundary (figs. 32, 8, and 9), which may have damped hydraulic head response. Simulated hydraulic head values for wells 124 and 3 were similar in trend; however, the simulated hydraulic head

values for well 3 were about 50 ft higher than observed values. Simulated hydraulic head values for wells 200 and 33 were similar to observed hydraulic head values; however, declines during the dry period and increases during the wet period were less for the simulated values.

Simulated hydraulic head values for wells 223 and 46 were similar to observed hydraulic head values; however, the influence of pumping resulted in some fluctuations in observed hydraulic head that were not matched in the simulated values, probably because of smoothing from using 6-month stress periods. Simulated hydraulic head values for wells 229 and 50 were similar to observed hydraulic head values. Simulated hydraulic head values for well 272 were similar in trend but consistently about 50 ft below observed values. Simulated and observed hydraulic head values for well 68 were very similar.

Simulated hydraulic head values for well 181 were similar in trend and consistently about 20 ft above observed values. Simulated hydraulic head values for well 209 were similar to observed hydraulic head values; however, simulated hydraulic head values were smoothed compared to observed values. Simulated hydraulic head values for wells 224 and 233 were similar to observed values. The simulated hydraulic head values for well 15 (table 12) were about 60 ft below observed values. Well 15 had only four observations (not shown in fig. 33) and is located near the western model boundary. Simulated hydraulic head values for well 47 were about 35 ft above observed values and were steady in comparison to observed values. Simulated hydraulic head values for well 43 were influenced by pumping in a nearby well, resulting in large fluctuation in observed hydraulic head values; however, the simulated values were similar to the recovered hydraulic head values.

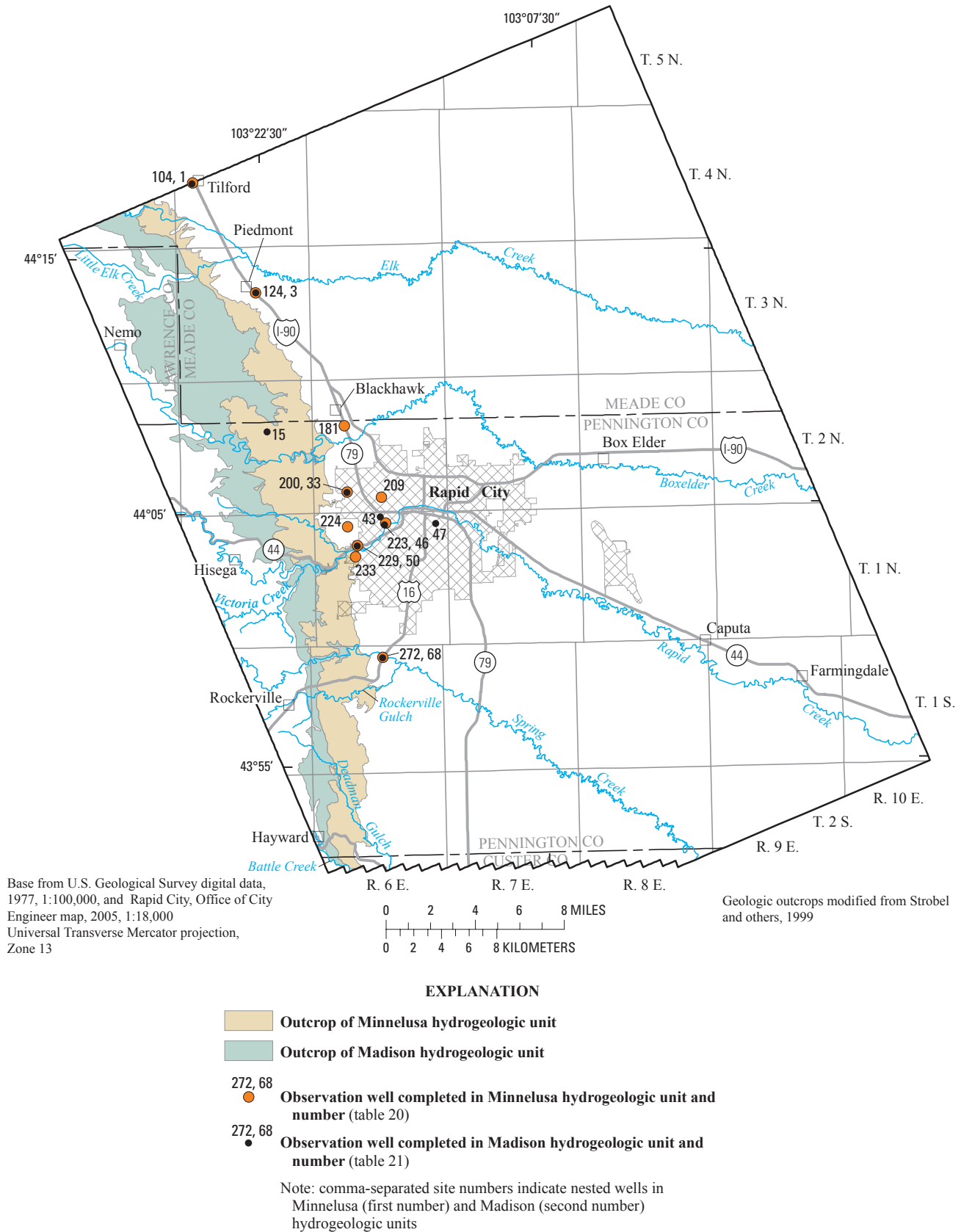


Figure 32. Locations of observation wells with transient hydraulic head values.

Table 12. Difference between simulated and observed hydraulic head values for transient simulation.

[Site numbers listed in tables 20 and 21; number corresponds to site number in tables 28 and 29 and plates 1 and 2 of Long and Putnam (2002). Smallest, largest, and average difference calculated by subtracting observed value from simulated value]

Station identification number	Model layer	Site number	Other identifier	Number of observations	Smallest difference (feet)	Largest difference (feet)	Average difference (feet)	Average absolute difference (feet)
441759103261201	1	104	TF-1	20	1.5	52.4	-8.3	19.0
441337103225001	1	124	PDMT-1	15	16.3	31.6	23.8	23.8
440818103180801	1	181	BLHK-1	20	5.0	30.5	19.5	19.5
440544103180001	1	200	CQ-1	16	1.6	25.4	-7.1	13.5
440528103161001	1	209	CP	16	2.9	37.1	-10.8	11.7
440430103160201	1	223	SP-1	15	8.9	26.9	-18.3	18.3
440423103180501	1	224	WCR-3	16	.8	6.8	.2	3.5
440338103173301	1	229	CL-1	16	3.6	6.7	5.4	5.4
440310103173801	1	233	CHLN-1	13	2.1	7.3	-4.4	4.4
435916103161802	1	272	RG-2	6	34.9	53.3	-47.3	47.3
441759103261202	3	1	TF-2	13	20.3	83.6	49.4	49.4
441337103225002	3	3	PDMT-2	6	48.5	66.4	-57.1	57.1
440544103180002	3	33	CQ-2	16	2.1	25.6	-4.1	14.2
440446103161701	3	43	LC	15	8.3	64.6	13.4	30.6
440430103160202	3	46	SP-2	16	2.6	27.6	-14.1	16.2
440427103131701	3	47	RC-7	14	28.6	52.4	39.0	39.0
440338103173302	3	50	CL-2	16	2.6	13.5	-9.1	9.1
435916103161801	3	68	RG	20	.2	21.2	-6.6	11.7
440811103222201	4	15	DOTY	4	58.4	76.1	-65.1	65.1

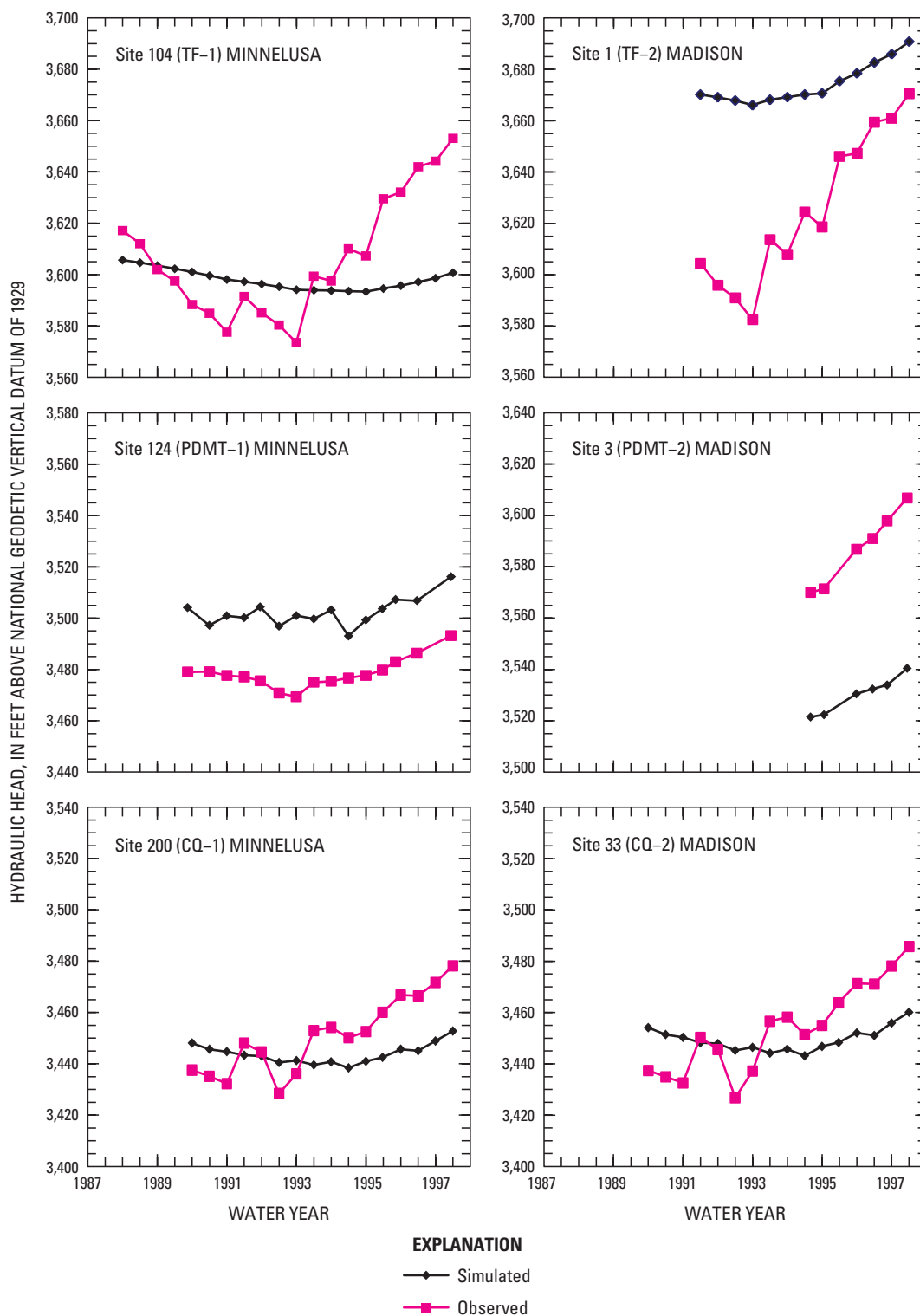


Figure 33. Simulated and observed hydraulic head values for model layers 1 and 3.

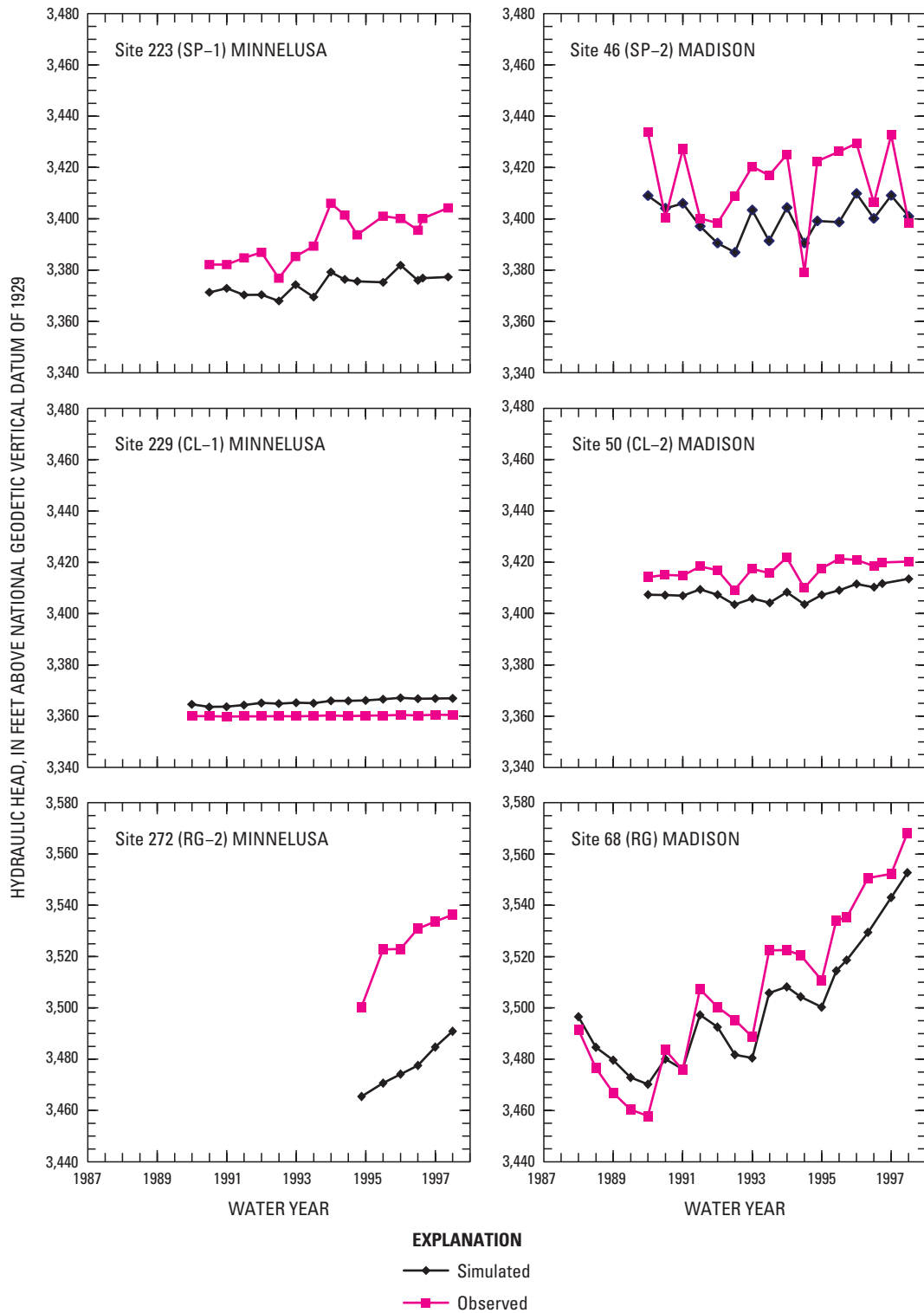


Figure 33. Simulated and observed hydraulic head values for model layers 1 and 3.—Continued

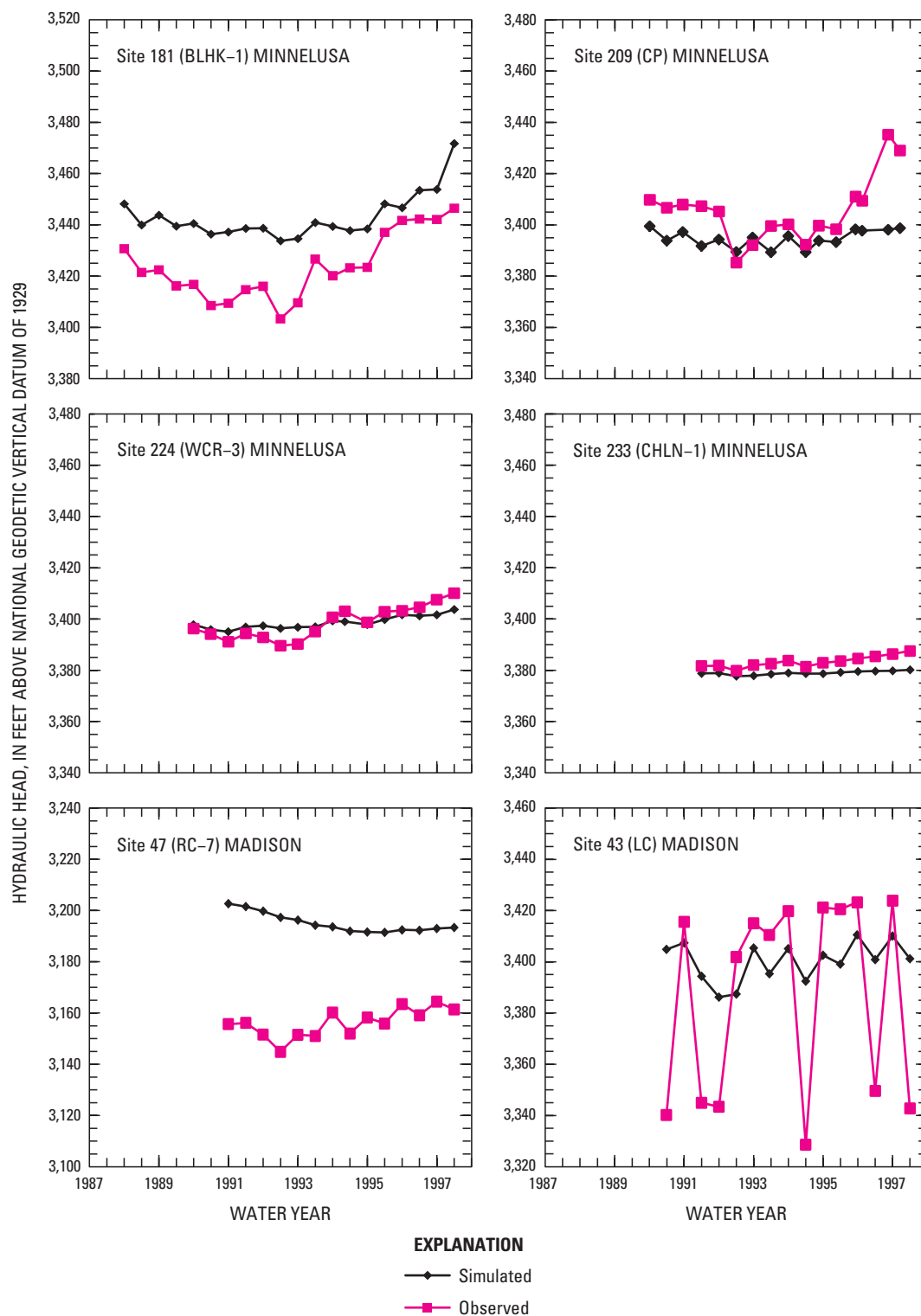


Figure 33. Simulated and observed hydraulic head values for model layers 1 and 3.—Continued

Comparison of Simulated and Observed Transient Springflow

Total simulated transient springflow ranged from 25.7 to 42.3 ft³/s (table 13). Simulated springflow for Jackson-Cleghorn Springs ranged from 17.0 to 23.6 ft³/s and averaged 20.0 ft³/s (table 13 and fig. 34) compared to an estimate of 21.6 ft³/s (Anderson and others, 1999). The Jackson-Cleghorn springflow from model layer 1 ranged from 1.5 to 2.7 ft³/s and averaged 2.2 ft³/s. Simulated springflow from model layer 3 ranged from 15.4 to 20.9 ft³/s and averaged 17.7 ft³/s. Measurements of Jackson-Cleghorn springflow during the transient simulation period were not available to compare to simulated transient springflow; however, measurements made during a dry period during WY 2005–07 (table 14) with pumping stresses indicated that the simulated range in transient springflow was plausible. Discharge from Jackson-Cleghorn Springs was most sensitive to the conductance of the head-dependent cells (CLEGmdsn; table 10; fig. 12) representing the spring followed by horizontal hydraulic conductivity for parameter group HK3_2 (table 10; fig. 19).

Observed and simulated springflow for six smaller springs also are listed in table 13. Simulated springflow for City Springs had similar trends to observed springflow (fig. 34); however, changes in simulated values generally were less than changes in observed values. Simulated springflow for Deadwood Avenue Springs was slightly less than the estimated average springflow (fig. 34). Simulated springflow for Boxelder Springs had similar trends to estimated springflow; however, the increase during the wet period (late 1990s) was less for simulated values than for estimated values (fig. 34).

The averages of estimated and simulated springflow for Elk Springs were about the same (table 13); however, simulated values had a much smaller range (fig. 34). Elk Springs is near the northern no-flow boundary, which influences simulated springflow. During very wet conditions, groundwater may flow into the study area because of recharge on the large outcrop of the Minnelusa and Madison hydrogeologic units northwest of the model area near the Vanocker Laccolith (Strobel and others, 1999). During extremely wet periods, the assumption of a no-flow boundary may not be the best approximation. Also, representation of conduits and heterogeneity associated with this spring may be difficult to approximate with the larger cell size near the northern boundary of the model.

Simulated springflow representing the combined bedrock contributions to the Meadowbrook and Girl Scout galleries (Infiltration Gallery Springs) was smaller than the springflow estimated by Anderson and others (1999). The potential error in the estimated springflow was relatively large (Anderson and others (1999)). The simulated springflow to Canyon Lake ranged from 0.3 to 0.9 ft³/s. No measured or estimated springflow information was available for Canyon Lake; however, a response in hydraulic head at a nearby observation well completed in the Minnelusa hydrogeologic unit (well 229) was

evident when Canyon Lake was drained from December 1995 through January 1996 (Driscoll, Bradford, and Moran, 2000).

Comparison of Simulated and Estimated Transient Regional Outflow

Simulated transient regional outflow from the Minnelusa and Madison hydrogeologic units at the eastern model boundary (table 15) was steady and slightly higher than the average outflow of 12.3 ft³/s for the Minnelusa hydrogeologic unit and 12.2 ft³/s for the Madison hydrogeologic unit estimated by Long and Putnam (2002). The eastern boundary was far enough east so that changes in recharge and pumping did not affect regional outflow at the boundary.

Calibrated Hydraulic Properties

Horizontal hydraulic conductivity for zones in layers 1 and 3, vertical conductivity for part of layer 2, and conductance for head-dependent cells representing selected springs were optimized with steady-state parameter estimation and trial-and-error transient simulations. Horizontal hydraulic conductivity for layers 2 and 4 was set to 0.1 times the hydraulic conductivity for layers 1 and 3, respectively, except near the outcrops, where it was set to 0.5 times the hydraulic conductivity for layers 1 and 3. The horizontal hydraulic conductivity for layers 2 and 4 was assumed to be smaller than for layers 1 and 3, respectively; however, the values selected are arbitrary. Vertical conductivity was assigned for layers 1, 3, 4, and 5 as 0.1 times the horizontal hydraulic conductivity for the respective layer. The model was not sensitive to these parameters for vertical conductivity; therefore, the assigned value was assumed.

Calibrated horizontal hydraulic conductivity values for model layer 1 parameter groups (fig. 17) ranged from 1.0 to 5.2 ft/d (table 16). The largest horizontal hydraulic conductivity (parameter group HK1_1) represented the zone near the outcrop of the Minnelusa hydrogeologic unit from Boxelder Creek south to the Jackson-Cleghorn Springs area. The smallest hydraulic conductivity (parameter group HK1_3) represented the north and central part of the study area where the hydraulic gradient generally was the steepest.

Horizontal hydraulic conductivity values for model layer 3 parameter groups (figs. 19 and 20) ranged from 0.1 to 388.8 ft/d (table 16). The largest value of hydraulic conductivity of 388.8 ft/d (zone HK3_6) represented the zone around Jackson-Cleghorn Springs where converging enlarged solution openings are likely. Zone HK3_7, which approximated conduit zones beginning at the Spring Creek loss zone and surrounding zone HK3_6, had a hydraulic conductivity of 22.9 ft/d. Hydraulic conductivity values determined from large-scale aquifer tests that overlap zones HK3_6 and HK3_7 were about 57 and 93 ft/d (Greene, 1993; Long and Putnam, 2002). Zone H3_4, which approximates the conduit zone

Table 13. Estimated and simulated transient springflow.

[Springflow in cubic feet per second. Springflow for Jackson-Cleghorn and Infiltration Gallery Springs from Anderson and others (1999); estimate was made for water years 1988–89 and values for remaining stress periods were assumed to be similar. Springflow for City Springs from Long and Putnam (2002); flow from City Springs and some unnamed springs about 0.3 mi to the east were estimated from streamflow record. Springflow for Deadwood Avenue, Boxelder, and Elk Springs from Long and Putnam (2002); an average value was estimated for Deadwood Avenue Springs on the basis of streamflow records for water years 1988–90. No quantitative measurements were available for Canyon Lake; small discharge was assumed]

Stress period	Jackson-Cleghorn Springs		City Springs		Deadwood Avenue Springs		Boxelder Springs		Elk Springs		Infiltration Gallery Springs		Canyon Lake		Total	
	Estimated	Simulated	Estimated	Simulated	Estimated	Simulated	Estimated	Simulated	Estimated	Simulated	Estimated	Simulated	Estimated	Simulated	Estimated	Simulated
1	21.6	20.9	1.9	1.8	2.8	3.2	0.0	1.3	0.3	6.2	1.5	0.5	0.5	0.6	28.6	34.5
2	21.6	20.0	1.0	1.6	2.8	2.8	.0	1.0	.3	5.9	1.5	.5	.5	.6	27.7	32.3
3	21.6	19.8	1.0	1.7	2.8	3.1	.0	.9	.0	5.7	1.5	.5	.5	.7	27.4	32.4
4	21.6	19.7	.7	1.5	2.8	2.8	.0	.8	.0	5.4	1.5	.5	.5	.7	27.1	31.4
5	21.6	19.1	.8	1.5	2.8	2.9	.0	.7	.0	5.3	1.5	.5	.5	.5	27.2	30.4
6	21.6	19.1	.8	1.2	2.8	2.5	.0	.6	.0	5.0	1.5	.4	.5	.3	27.2	29.2
7	21.6	18.9	.3	1.0	2.8	2.7	.0	.6	.1	4.8	1.5	.4	.5	.4	26.8	28.8
8	21.6	20.8	1.3	.7	2.8	2.1	.0	1.4	.1	5.1	1.5	.4	.5	.4	27.8	31.0
9	21.6	19.5	.6	.9	2.8	1.9	.0	.9	.0	5.1	1.5	.4	.5	.5	27.0	29.3
10	21.6	17.0	.4	.4	2.8	1.8	.0	.6	.0	5.0	1.5	.4	.5	.5	26.8	25.7
11	21.6	18.4	.3	.9	2.8	2.0	.2	.5	.1	4.8	1.5	.4	.5	.5	27.0	27.5
12	21.6	18.1	2.1	.3	2.8	1.5	.2	2.1	.1	5.1	1.5	.4	.5	.5	28.8	28.0
13	21.6	20.0	1.5	.5	2.8	1.7	.0	1.3	.4	5.2	1.5	.5	.5	.7	28.3	30.0
14	21.6	17.6	1.4	.1	2.8	1.3	.0	1.9	.4	5.1	1.5	.4	.5	.6	28.2	27.1
15	21.6	19.3	1.3	.8	2.8	1.9	.5	1.1	8.6	5.1	1.5	.5	.5	.7	36.8	29.4
16	21.6	20.8	3.3	.7	2.8	2.0	.5	3.1	8.6	5.8	1.5	.5	.5	.8	38.8	33.6
17	21.6	21.9	2.8	1.3	2.8	2.4	2.5	1.9	8.2	6.1	1.5	.5	.5	.9	39.9	34.9
18	21.6	21.5	4.0	.7	2.8	1.9	2.5	3.7	8.2	6.5	1.5	.5	.5	.8	41.1	35.5
19	21.6	23.0	3.5	1.5	2.8	2.5	6.8	2.7	18.9	6.6	1.5	.5	.5	.8	55.6	37.6
20	21.6	23.6	4.9	1.5	2.8	2.8	6.8	5.9	18.9	7.2	1.5	.5	.5	.8	57.0	42.3
Average	21.6	20.0	1.7	1.0	2.8	2.3	1.0	1.7	3.7	5.6	1.5	.5	.5	.6	32.8	31.5

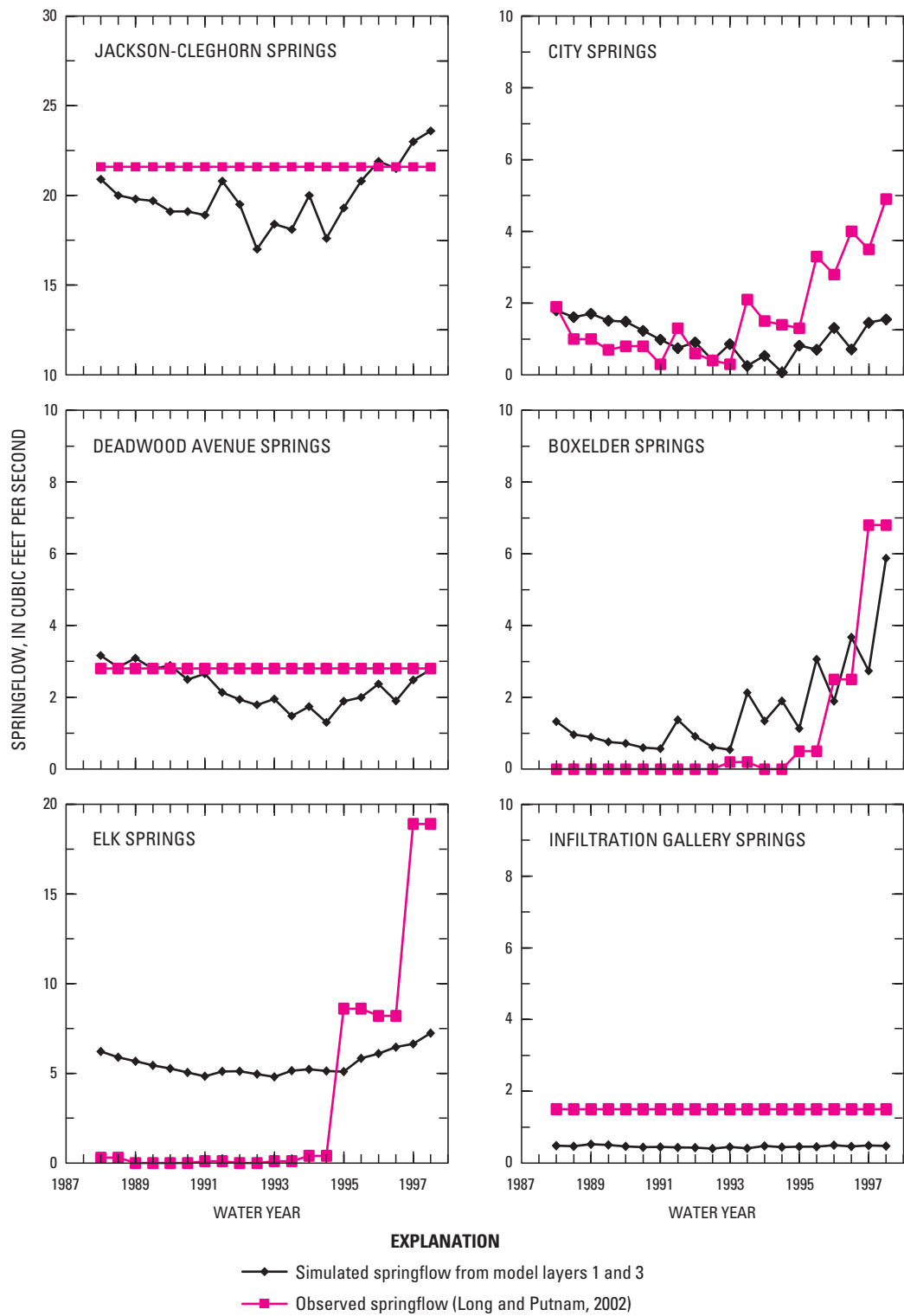


Figure 34. Simulated and observed or estimated springflow values.

Table 14. Jackson-Cleghorn springflow estimated from streamflow measurements on Rapid Creek.

[Upstream station is Rapid Creek at Braeburn Addition (fig. 12). Downstream station is Rapid Creek below Cleghorn Springs (fig. 12). Pumping withdrawal rate from Jackson Springs Gallery from Ron Barber, Rapid City Water Department, oral commun. (2004–05). Estimated Jackson-Cleghorn springflow calculated by subtracting streamflow at upstream station from downstream station and adding pumping withdrawal from Jackson Springs Gallery]

Date of paired streamflow measurements	Streamflow (cubic feet per second)		Pumping withdrawal rate from Jackson Springs Gallery (cubic feet per second)	Estimated Jackson-Cleghorn springflow (cubic feet per second)
	Upstream station (440321103181101)	Downstream station (06412900)		
10/19/2004	10.3	26.2	0.0	15.9
10/19/2004	10.7	25.2	.0	14.5
10/19/2004	11.0	28.6	.0	17.6
10/19/2004	10.2	26.5	.0	16.3
10/26/2004	9.9	22.8	4.0	16.9
10/26/2004	10.3	23.0	4.0	16.7
11/04/2004	10.1	24.8	4.0	18.7
11/04/2004	10.7	26.4	4.0	19.7
10/11/2005	8.3	16.6	10.0	18.3
10/11/2005	8.7	16.4	10.0	17.7
11/07/2005	6.4	19.6	4.0	17.2
11/07/2005	7.6	21.2	4.0	17.6
10/02/2006	6.3	21.2	.0	14.9
03/15/2007	8.5	25.2	.0	16.7
03/15/2007	7.9	23.5	.0	15.6
03/16/2007	7.2	23.2	.0	16.0
03/16/2007	7.0	22.0	.0	15.0

Table 15. Simulated transient regional outflow from Minnelusa and Madison hydrogeologic units at the eastern model boundary.

Stress period	Simulated regional outflow (cubic feet per second)		Stress period	Simulated regional outflow (cubic feet per second)	
	Minnelusa hydrogeologic unit	Madison hydrogeologic unit		Minnelusa hydrogeologic unit	Madison hydrogeologic unit
1	12.9	12.9	11	12.8	12.7
2	12.9	12.6	12	12.7	12.6
3	12.9	12.9	13	12.8	12.7
4	12.8	12.8	14	12.7	12.5
5	12.9	12.9	15	12.8	12.6
6	12.8	12.8	16	12.7	12.5
7	12.9	12.9	17	12.7	12.6
8	12.8	12.7	18	12.7	12.5
9	12.8	12.8	19	12.8	12.6
10	12.8	12.7	20	12.8	12.5

Table 16. Calibrated horizontal hydraulic conductivity for parameters representing zones of hydraulic conductivity in model layers 1, 3, and 5.

[Location of area represented by the parameter group name is shown in figures 17, 19, 20, and 22]

Parameter group name	Model layer	Area of parameter zone (square miles)	Effective horizontal hydraulic conductivity (feet per day)
HK1_1	1	67.9	5.2
HK1_2	1	566.0	3.3
HK1_3	1	153.2	1.0
¹ HK3_1	3	46.8	16.0
HK3_2	3	596.3	5.4
HK3_3	3	56.8	.1
HK3_4	3	28.9	39.3
HK3_5	3	91.5	1.4
HK3_6	3	6.9	388.8
HK3_7	3	16.7	22.9
HK5_1	5	59.4	.1

¹Hydraulic conductivity represented by this parameter included anisotropy of 0.1; hydraulic conductivity along rows (K_x) = 50.5 feet per day and columns (K_y) = 5.05 feet per day; effective hydraulic conductivity equals the square root of ($K_x * K_y$).

extending from Boxelder Creek that included an eastward bulge in the potentiometric surface, had a hydraulic conductivity of 39.3 ft/d. Substantial heterogeneity within this zone is likely on the basis of tracer tests and the prevalence of numerous structural features. Calibrated hydraulic conductivity for zone HK3_5 was 1.4 ft/d, which represented areas near the outcrop of the Madison hydrogeologic unit with relatively steep hydraulic gradients and areas that transition from gradual to steep hydraulic gradients. Calibrated hydraulic conductivity for zone HK3_2 was 5.4 ft/d, which is consistent with the assumption of higher hydraulic conductivity where the hydraulic gradient is relatively gradual.

Zone HK3_3, which represents the area where the hydraulic gradient is relatively steep, was assigned a hydraulic conductivity of 0.1 ft/d. In model calibration, hydraulic conductivity for this area converged towards a very small value. The parameter group HK3_3 became relatively insensitive at these small values; therefore, a small value was assigned. The effective hydraulic conductivity for zone HK3_1 was 16.0 ft/d with anisotropy of 0.1 (10 times greater along model rows than columns). This parameter group approximated the area with large hydraulic conductivity identified in regional modeling (Downey, 1986). The hydraulic properties represented by zones HK3_1 and HK3_3 are based on sparse information, and the areas probably include a combination of massive impermeable limestone and karst features that are poorly characterized.

Calibrated vertical hydraulic conductivity for layer 2 (fig. 23) was 0.0025 ft/d for parameter VKA2_2. Vertical hydraulic conductivity was 0.025 ft/d for the areas within

VKA2_2 near structural features and 0.25 ft/d for the area near Jackson-Cleghorn Springs. Parameter VKA2_2 was relatively sensitive to the concentration of observations in this parameter zone. The model was insensitive to parameter VKA2_1, and therefore a small value of 0.000001 ft/d was assigned.

The conductance term for head-dependent cells representing springs was calibrated using parameters CLEGmdsn and DWDSR. The sensitivity of the conductance parameters for the remaining springs to observed data was low and values were estimated. Conductance for head-dependent cells representing springs ranged from 3,000 to 86,400 feet squared per day (table 17).

Table 17. Conductance for head-dependent cells representing springs.

Parameter name	Spring name (location on fig. 11 or 12)	Conductance (feet squared per day)
CLEGmdsn	Jackson-Cleghorn Springs	32,000
CLEGmnl	Jackson-Cleghorn Springs	8,600
CITYSPR	City Springs	35,000
DWSPR	Deadwood Avenue Springs	3,000
BESPR	Boxelder Springs	35,000
ELKSPR	Elk Springs	86,400
GAL	Infiltration Gallery Springs	6,000
CLAKE	Canyon Lake	3,400

Response to Stress

In calibration of the transient model, the largest annual average pumping rate for large production wells in the Madison hydrogeologic unit was about 7 ft³/s. To test the relation between pumping stresses and boundary conditions, the 10-year transient simulation was made with production of 7 ft³/s from the large production wells in the Madison hydrogeologic unit. The hydraulic head values at the end of this simulation were saved, and the same simulation was made with pumping increased by 2-ft³/s increments for the same wells. Simulations were made for hypothetical pumping rates of 7, 9, 11, 13, 15 and 17 ft³/s. The increases in pumping were distributed to the production wells as a fraction of the productive capacity of each well relative to the total pumping from all the wells. The outline of the area where drawdown from the additional pumping exceeded 2 ft was calculated for each of these simulations (fig. 35).

When pumping was increased from 7 to 17 ft³/s, drawdown of 2 feet approached the northern and southern boundaries of the model (fig. 35). Simulation of increased hypothetical pumping rates of more than 10 ft³/s (total of about 17 ft³/s) may require modification to the no-flow boundaries to allow flow into the model. Outflow at the eastern boundary did not change with increases in pumping from 7 to 17 ft³/s. The drawdown at observation well 46 (SP-2; fig. 35) increased from 5.3 ft with an increase in pumping of 2 ft³/s to 27.1 ft with an increase in pumping of 10 ft³/s.

The influence of potential pumping stresses on springflow is an important consideration in evaluating management alternatives. Although some error may be associated with estimating the magnitude of decreases in springflow, transient simulation of pumping rates provides an assessment of how increased pumping could influence springflow. Compilation of the simulated springflows for Jackson-Cleghorn Springs (fig. 36) for each of the six previous hypothetical increases in pumping rates gives an indication of how springflow may decrease with time. The simulated effect of increased pumping from existing production wells on springflow from Jackson-Cleghorn Springs was a decrease in springflow equal to about 30 percent of the increase in pumping rate at 6 months increasing to about 50 percent of the increase in pumping rate after 10 years of pumping.

The drawdown in four observation wells during a 6-day aquifer test for production well 49 pumped at 2,550 gal/min (RC-9; Long and Putnam, 2002) was simulated to illustrate limitations of the model in representing the site-specific effect of pumping stresses. The locations of well 49 and the four observation wells in relation to layer 3 hydraulic conductivity zones are shown in figure 37. Comparison of the simulated and observed drawdown for the four wells (fig. 38) shows that the model approximates the average drawdown; however, the response at specific sites may not be matched. This probably results from generalization of heterogeneous hydraulic properties. The confined storage coefficient used in the model was

0.0003, whereas estimates from the aquifer test ranged from 0.000014 to 0.00027 (Long and Putnam, 2002). Additional model limitations are described in the following section.

Model Limitations

The model was developed to evaluate groundwater flow on a large scale and is limited by numerous simplifying assumptions. One of the most important simplifying assumptions is representation of a hydrogeologic system that contains substantial secondary porosity as a porous media. A bulk parameter that represents the average for a given model cell area provides only a generalized approximation of hydraulic properties, especially for the karstic Madison hydrogeologic unit. Representation of vertical heterogeneity also is very generalized, especially for the Minnelusa hydrogeologic unit, which contains many layers of different consolidated sediments. These simplifications probably result in a simulated hydraulic head response to stress that is more damped than observed responses. Site specific responses in some areas could be substantially in error.

An additional limitation is the influence of boundary conditions. Model boundaries were extended away from the detailed study area to limit the influence of errors in approximating boundary conditions; however, the potential for boundary conditions to influence simulation results increases as simulated stresses are increased. No-flow boundaries on the north and south may approximate average conditions; however, the boundaries may not represent extreme recharge conditions as well.

Estimates of streamflow recharge, which makes up a substantial portion of the total recharge, were based on measured values; however, the distribution of that recharge between model layers is more uncertain. Areal recharge was estimated by more indirect methods, and storage in the partially saturated zones near the outcrops is poorly understood. Regional outflow at the eastern boundary is based on a water balance, and adequate information was not available for the horizontal distribution among model layers. The distribution between hydrogeologic units was assumed to be about equal. Few springflow measurements were available, and many of the calibration constraints were based on estimates. Available aquifer tests represent only a small part of the model area, and interpretation of hydraulic properties from aquifer tests is difficult because of the complex hydrogeology.

With additional data, further refinement of the model would be possible, which could improve the accuracy of model estimates of the effects of additional stresses on the system, such as increased withdrawals or drought. The model can yield simulations of future conditions, which can guide management decisions and planning. The model provides a useful tool for general characterization of the effects of stresses and management alternatives on a regional basis.

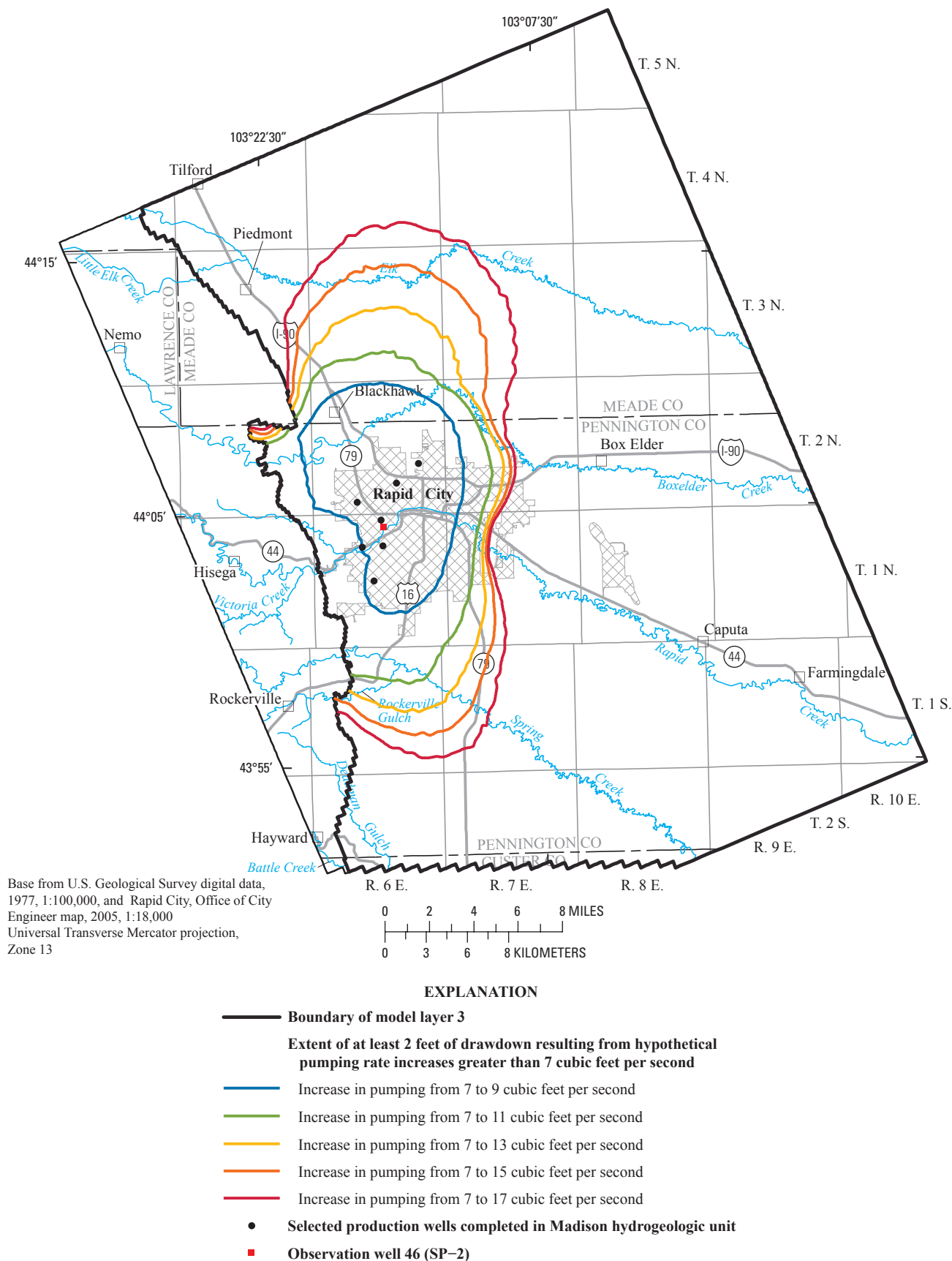


Figure 35. Simulated extent of drawdown resulting from hypothetical increases in pumping rates from the Madison hydrogeologic unit.

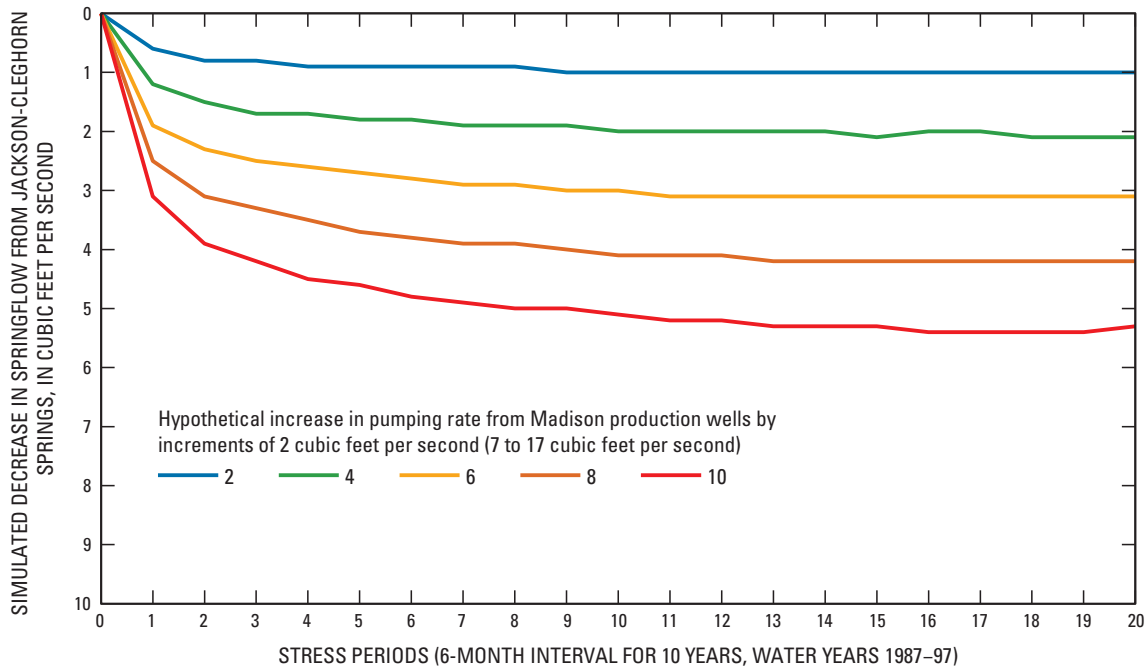


Figure 36. Simulated decrease in springflow from Jackson-Cleghorn Springs in response to hypothetical increases in pumping rates from the Madison hydrogeologic unit.

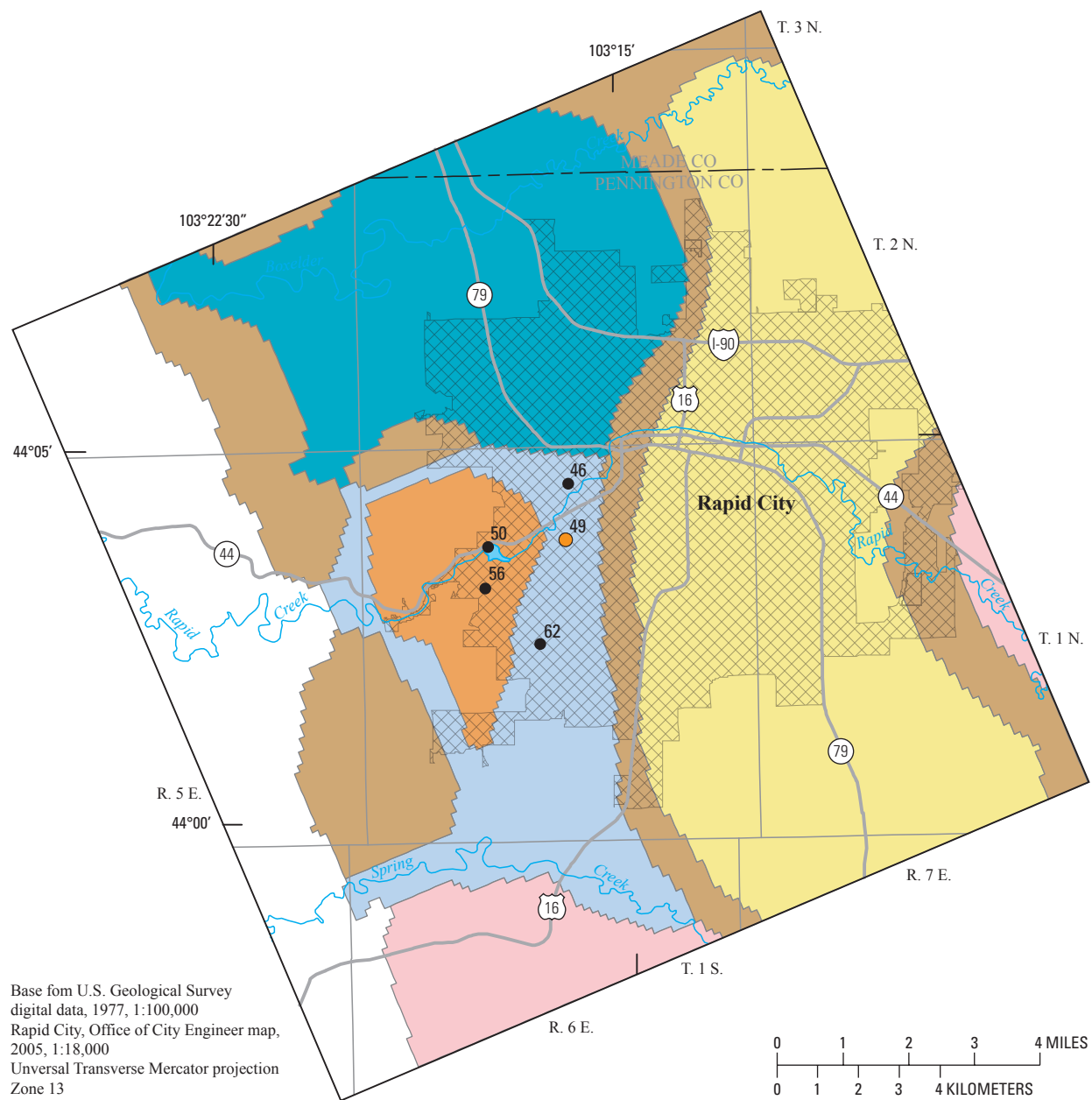
Summary

The city of Rapid City, South Dakota, obtains more than one-half of its municipal water supplies from the Minnelusa and Madison aquifers through deep wells and springs, predominantly from the Madison aquifer. Numerous additional users in the Rapid City area obtain water from the Minnelusa and Madison aquifers for domestic, commercial, industrial, and irrigation usage. The U.S. Geological Survey, as part of a long-term cooperative study with the city of Rapid City, has compiled numerous datasets designed to better understand groundwater flow in the Minnelusa and Madison hydrogeologic units. A numerical groundwater-flow model was developed of the Minnelusa and Madison hydrogeologic units in the Rapid City area to synthesize estimates of water-budget components and hydraulic properties, and to provide a tool to analyze the effect of additional stress on water-level altitudes within the aquifers and on discharge to springs. The purposes of this report are to (1) document the development of a numerical groundwater-flow model of the Minnelusa and Madison hydrogeologic units, which contain the Minnelusa and Madison aquifers, in the Rapid City area in South Dakota, and (2) present simulated responses to stress and describe model limitations.

The 1,000-square-mile study area on the eastern flank of the Black Hills includes Rapid City and the surrounding area. Land-surface altitudes range from more than 5,000 feet (ft)

on the western side of the study area to about 2,800 ft in the eastern lowlands. Average precipitation rates range from about 24 inches per year (in/yr) in the northwest to about 16 in/yr in the eastern lowlands. The outcrops of the Minnelusa and Madison hydrogeologic units, which are characterized by high-relief forested areas cut by deep canyons, form generally concentric rings surrounding the Precambrian core of the uplifted Black Hills. Water-table conditions generally exist in outcrop areas and extend to more than 1 mile east of the outcrop areas in parts of the Minnelusa and Madison hydrogeologic units in the study area. Confined conditions exist east of the water-table areas because of the easterly dip of the hydrogeologic units.

The Minnelusa hydrogeologic unit is 375 to 800 ft thick in the study area with the more permeable upper part containing predominantly sandstone, and the less permeable lower part containing more shale and limestone than the upper part. Siltstone, gypsum, and anhydrite also can be present. The shale units in the lower part generally impede flow between the Minnelusa hydrogeologic unit and underlying Madison hydrogeologic unit; however, fracturing and weathering may result in hydraulic connections in some areas. Recharge to the Minnelusa hydrogeologic unit is from streamflow loss that occurs where streams cross the outcrop and from infiltration of precipitation on the outcrops. Wells completed in the Minnelusa hydrogeologic unit are used extensively by small suburban developments and for domestic supplies surrounding Rapid City.



- EXPLANATION**
- Parameter group and hydraulic conductivity**
- HK3_2 (5.6 feet per day)
 - HK3_3 (0.1 feet per day)
 - HK3_4 (31.9 feet per day)
 - HK3_5 (1.0 feet per day)
 - HK3_6 (660 feet per day)
 - HK3_7 (100 feet per day)
- ⁴⁹ **Production well (site 49, table 20)**
- ⁶² **Well where drawdown was monitored during aquifer test (number is site number in table 20; Long and Putnam, 2002)**

Figure 37. Locations of production and observation wells for Madison aquifer test at well 49.

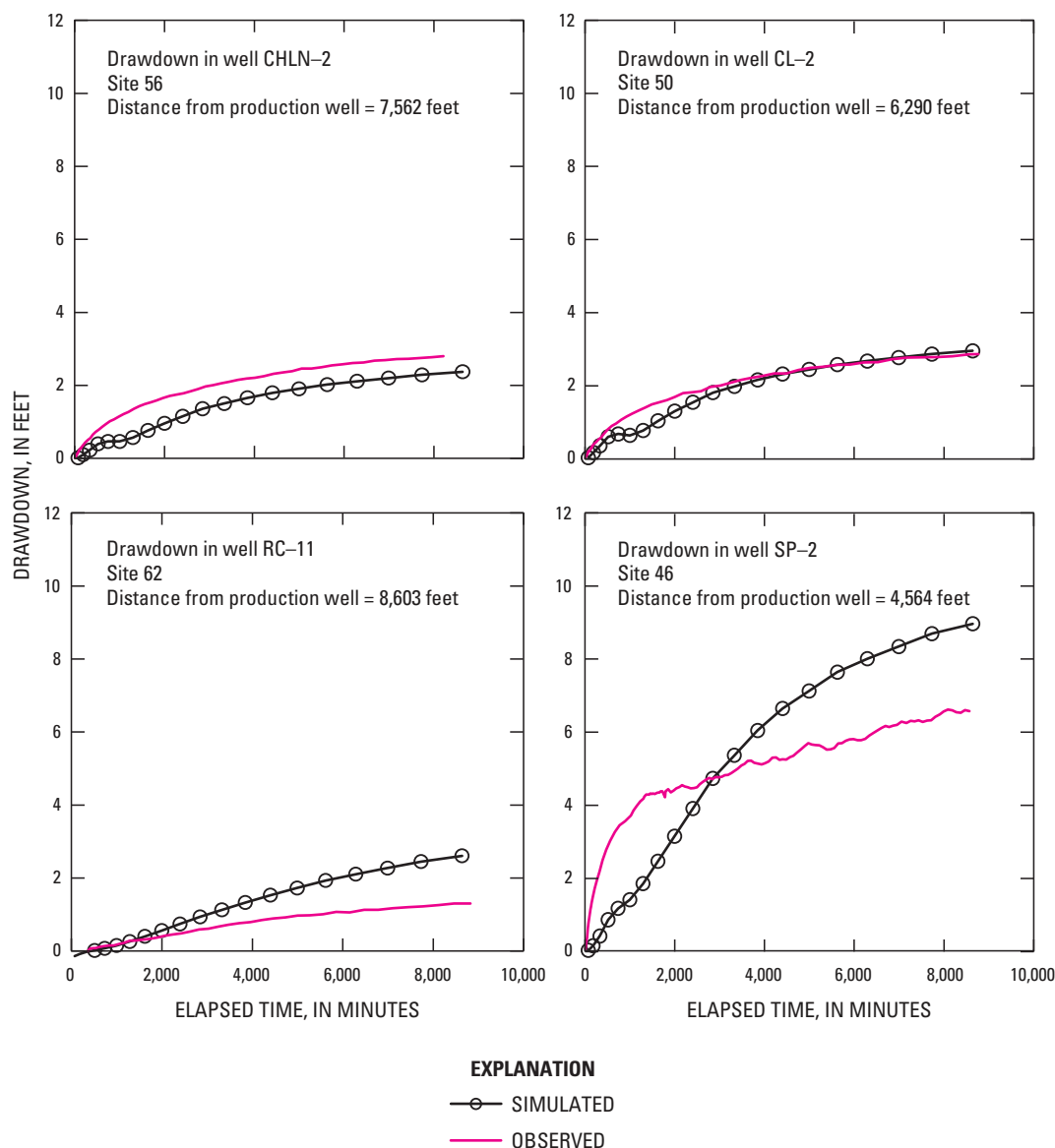


Figure 38. Simulated and observed drawdown for Madison aquifer test at well 49.

The Madison hydrogeologic unit, which includes the Englewood Formation, is composed of limestone and dolomite that is about 250 to 610 ft thick in the study area. The upper 150 ft of the Madison hydrogeologic unit contains substantial secondary permeability from solution openings and fractures, and permeability is smaller in the lower part of the unit. Karst features are found throughout the Madison Limestone; however, they tend to be more common along the contacts between mapped geomorphic units. Recharge to the Madison hydrogeologic unit is from streamflow loss where streams cross the outcrop and from infiltration of precipitation on the outcrops (areal recharge). The amount of streamflow loss to the Madison hydrogeologic unit outcrops can be as much as 25 cubic feet per second (ft^3/s) for some streams, and

streamflow recharge usually is greater than areal recharge in the study area. Wells completed in the Madison hydrogeologic unit are capable of producing from 5 to 2,500 gallons per minute.

MODFLOW-2000, a numerical, three-dimensional, finite-difference groundwater-flow model, was used to simulate flow in the Minnelusa and Madison hydrogeologic units, and the MODFLOW-2000 parameter estimation process was used to optimize estimates of hydraulic properties. The model includes five layers with layer 1 representing the fractured sandstone layers in the upper part of the Minnelusa hydrogeologic unit. Layer 2 represented the fractured limestone, minor sandstone layers, and shale units in the lower part of the Minnelusa hydrogeologic unit. Layer 3 represented the upper part

of the Madison hydrogeologic unit, and layer 4 represented the less permeable lower part of the Madison hydrogeologic unit. Layer 5 represented an approximation of the underlying Deadwood aquifer to simulate upward flow to the Madison hydrogeologic unit. Arrays representing the altitude of the tops of layers 1, 3, and 5 were constructed from maps of the structural tops of the Minnelusa Formation, the Madison Limestone, and the Deadwood Formation. A uniform thickness of 250 ft was assumed for layer 1 with the remainder of the Minnelusa hydrogeologic unit represented as layer 2. A uniform thickness of 150 ft was assumed for layer 3 with the remainder of the Madison hydrogeologic unit represented in layer 4.

The finite-difference grid included 221 rows and 169 columns with a square cell size of 492.1 ft in the detailed study area that surrounded Rapid City. Cell sizes increased outward from the detailed area to 1,640 ft at boundaries on the north and south and to 6,562 ft at the eastern boundary. The model was oriented 23 degrees counterclockwise to approximate orthogonal directions observed in cave passages and fractures. The northern and southern boundaries for layers 1–4 were represented as no-flow boundaries, and the eastern boundary was represented with head-dependent flow cells. Streamflow recharge was represented with specified-flow cells. Areal recharge to layers 1–4 was represented with a specified-flux boundary with the recharge flux assigned to the westernmost active cells. A specified-flux boundary assigned to the westernmost active cells in layer 5 represented the source of upward flow from the Deadwood aquifer to layer 4. Calibration of the model was accomplished by two simulations: (1) steady-state simulation of average conditions for water years 1988–97 and (2) transient simulations of water years 1988–97 divided into twenty 6-month stress periods.

Flow-system components represented in the model include recharge, discharge, and hydraulic properties. The steady-state streamflow recharge rate was 42.2 ft³/s for five major streams and five minor tributaries with 36.2 ft³/s recharging the Madison hydrogeologic unit. Total transient streamflow recharge rates by stress period for the Minnelusa and Madison hydrogeologic units ranged from 14.1 to 102.2 ft³/s. The steady-state areal recharge rate was 20.9 ft³/s and was distributed among five zones from north to south delineated by the major streams where they crossed the outcrops. Total transient areal recharge rates by stress period for the Minnelusa and Madison hydrogeologic units ranged from 1.1 to 98.4 ft³/s. The upward flow rate from the Deadwood aquifer to the Madison hydrogeologic unit was 6.3 ft³/s.

Discharge included springflow, water use, flow to overlying units, and regional outflow. The estimated steady-state springflow was 32.8 ft³/s from seven springs. Steady-state water-use rates for the Minnelusa and Madison hydrogeologic units for the 10-year period were 3.4 and 6.7 ft³/s, respectively. Total transient water-use rates ranged from 3.4 to 19.1 ft³/s. Steady-state flow from the Minnelusa hydrogeologic unit to overlying units was 2.0 ft³/s.

The hydraulic properties that the model was most sensitive to were horizontal hydraulic conductivity, vertical

hydraulic conductivity for layer 2 near the outcrop of the Minnelusa hydrogeologic unit, and conductance for Jackson-Cleghorn Springs. Linear regression of the 252 simulated and observed hydraulic heads for the steady-state simulation had a coefficient of determination (R^2 value) of 0.92 with an average arithmetic difference of -7.6 ft and an average absolute difference of 37.6 ft. The estimated steady-state springflow of 32.8 ft³/s from seven springs was similar to the simulated springflow of 31.6 ft³/s, which included springflow of 20.5 ft³/s from Jackson-Cleghorn Springs. Simulated steady-state regional outflow of 12.9 ft³/s from the Minnelusa hydrogeologic unit at the eastern boundary was slightly larger than the estimated outflow of 12.3 ft³/s. Simulated steady-state regional outflow from the Madison hydrogeologic unit of 12.8 ft³/s was slightly larger than the estimated outflow of 12.2 ft³/s.

For the transient simulation, the average absolute difference between simulated and observed hydraulic heads for the 19 observation wells ranged from 3.5 to 65.1 ft with a median value of 18.3 ft. Simulated transient springflow ranged from 25.7 to 42.3 ft³/s.

Horizontal hydraulic conductivity for layers 2 and 4 were 10 percent of horizontal hydraulic conductivity for layers 1 and 3, respectively, except near the outcrops where it was 50 percent of the conductivity for layers 1 and 3, respectively. Calibrated horizontal hydraulic conductivities for model layer 1 ranged from 1.0 to 5.2 feet per day (ft/d). Horizontal hydraulic conductivities for model layer 3 ranged from 0.1 to 388.8 ft/d. Vertical hydraulic conductivity for layers 1, 3, 4, and 5 was 10 percent of the respective horizontal hydraulic conductivity for those layers. Vertical hydraulic conductivity for layer 2 ranged from 0.000001 to 0.25 ft/d. Conductance for head-dependent cells representing springs ranged from 3,000 to 86,400 feet squared per day.

Simulation of increased hypothetical pumping of more than about 10 ft³/s may require modification to the no-flow boundaries to allow flow into the model. The simulated effect of increased pumping from existing production wells on springflow from Jackson-Cleghorn Springs was a decrease in springflow equal to about 30 percent of the increased pumping rate at 6 months increasing to about 50 percent of the increased pumping rate after 10 years of pumping. Simulation of a large-scale aquifer test indicated that the model is limited in describing pumping stresses with detail.

The model is limited by simplifying assumptions necessary to represent a hydrogeologic system that contains substantial secondary porosity as a porous media. A bulk parameter that represents the average for a given model cell area provides only a generalized approximation of hydraulic properties. Adequate information was not available for distribution of estimates of recharge and discharge among model layers. With additional data, further refinement of the model would be possible, which could improve the accuracy of model estimates of the effects of additional stresses on the system, such as increased withdrawals or drought. The model can yield simulations of future conditions, which can guide

management decisions and planning. The model provides a useful tool for general characterization of the effects of stresses and management alternatives on a regional basis.

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Supplemental Information

Table 18. Transient streamflow recharge rates for the Minnelusa and Madison hydrogeologic units.

[From Long and Putnam (2002). W, winter; S, summer (for example, W-88 = winter, water year 1988); Mnls, Minnelusa hydrogeologic unit; Mdsn, Madison hydrogeologic unit]

Stress period	Other stress period identifier	Streamflow recharge (cubic feet per second)									
		Elk Creek		Little Elk Creek		Boxelder Creek		Unnamed tributary on Rapid Creek		Rapid Creek	
		Mnls	Mdsn	Mnls	Mdsn	Mnls	Mdsn	Mnls	Mdsn	Mnls	Mdsn
1	W-88	0.0	1.2	0.0	0.4	0.0	4.2	0.0	0.0	2.0	8.0
2	S-88	.0	1.5	.2	.4	.0	5.7	.0	.1	2.0	8.0
3	W-89	.0	.8	.1	.3	.0	2.8	.0	.0	1.7	7.9
4	S-89	.0	1.5	.1	.5	.0	5.3	.0	.0	2.0	8.0
5	W-90	.0	1.2	.1	.4	.0	4.3	.0	.0	1.9	8.0
6	S-90	.0	2.1	.3	.6	.0	7.7	.0	.1	2.0	8.0
7	W-91	.0	.7	.0	.3	.0	2.4	.0	.0	1.8	8.0
8	S-91	1.0	3.8	.9	.9	2.7	17.1	.0	.2	2.0	8.0
9	W-92	.0	1.9	.2	.7	.0	7.8	.0	.1	2.0	8.0
10	S-92	.1	2.9	.1	.6	.0	6.9	.0	.1	2.0	8.0
11	W-93	.1	1.3	.1	.5	.1	5.0	.0	.0	1.9	8.0
12	S-93	2.3	4.8	1.6	1.1	5.5	25.3	.0	.4	2.0	8.0
13	W-94	.7	3.0	.5	.8	.5	11.2	.0	.1	2.0	8.0
14	S-94	1.6	4.1	1.1	.9	3.8	19.2	.0	.3	2.0	8.0
15	W-95	.2	3.5	.3	.8	.1	9.7	.0	.1	2.0	8.0
16	S-95	2.0	4.6	1.6	1.1	6.7	25.5	.1	.5	2.0	8.0
17	W-96	.5	4.2	.7	.9	.5	14.1	.0	.1	2.0	8.0
18	S-96	3.4	3.4	1.9	1.1	7.6	28.8	.0	.5	2.0	8.0
19	W-97	1.9	3.1	1.4	1.0	2.1	21.5	.0	.2	2.0	8.0
20	S-97	3.9	3.0	2.1	1.2	12.6	32.3	.1	.7	2.0	8.0
Average		.9	2.6	.7	.7	2.1	12.8	.0	.2	2.0	8.0

Table 18. Transient streamflow recharge rates for the Minnelusa and Madison hydrogeologic units.—Continued

[From Long and Putnam (2002). W, winter; S, summer (for example, W-88 = winter, water year 1988); Mnls, Minnelusa hydrogeologic unit; Mdsn, Madison hydrogeologic unit]

Stress period	Other stress period identifier	Streamflow recharge (cubic feet per second)												
		Victoria Creek		Spring Creek		Rockerville Gulch		Deadman Gulch		Battle Creek		Total of all streams		Total Mnls, Mdsn
		Mnls	Mdsn	Mnls	Mdsn	Mnls	Mdsn	Mnls	Mdsn	Mnls	Mdsn	Mnls	Mdsn	
1	W-88	0.0	0.1	0.0	1.1	0.0	0.2	0.0	0.1	0.0	0.3	2.0	15.6	17.6
2	S-88	.0	.2	.0	2.5	.0	.1	.0	.0	.0	.5	2.2	19.0	21.2
3	W-89	.0	.0	.0	.2	.0	.1	.0	.0	.0	.2	1.8	12.3	14.1
4	S-89	.0	.1	.0	1.8	.0	.1	.0	.0	.0	.7	2.1	18.0	20.1
5	W-90	.0	.1	.0	1.3	.0	.1	.0	.0	.0	1.1	2.0	16.5	18.5
6	S-90	.0	.9	.9	11.2	.0	.4	.0	.2	.0	4.1	3.2	35.3	38.5
7	W-91	.0	.2	.0	3.1	.0	.1	.0	.1	.0	.8	1.8	15.7	17.5
8	S-91	.0	1.4	1.9	16.8	.2	1.2	.1	.6	.0	4.4	8.8	54.4	63.2
9	W-92	.0	.4	.0	6.7	.0	.3	.0	.1	.0	1.5	2.2	27.5	29.7
10	S-92	.0	.5	.0	8.2	.0	.3	.0	.1	.0	2.3	2.2	29.9	32.1
11	W-93	.0	.3	.1	4.1	.0	.2	.0	.1	.0	1.1	2.3	20.6	22.9
12	S-93	.0	1.7	2.6	19.9	.1	1.7	.0	.8	.0	5.7	14.1	69.4	83.5
13	W-94	.0	.8	.2	11.1	.0	.9	.0	.4	.0	3.3	3.9	39.6	43.5
14	S-94	.0	.9	.9	11.0	.0	.5	.0	.2	.0	2.1	9.4	47.2	56.6
15	W-95	.0	.5	.1	7.1	.0	.4	.0	.2	.0	2.0	2.7	32.3	35.0
16	S-95	.0	1.5	2.1	18.0	.3	2.1	.1	1.1	.0	5.0	14.9	67.4	82.3
17	W-96	.0	.9	.1	13.1	.0	1.2	.0	.5	.0	3.1	3.8	46.1	49.9
18	S-96	.0	1.7	2.5	20.3	.1	1.7	.0	.8	.0	5.2	17.5	71.5	89.0
19	W-97	.0	1.4	1.5	18.5	.0	1.6	.0	.7	.0	4.6	8.9	60.6	69.5
20	S-97	.0	1.9	3.4	21.0	.2	2.5	.1	1.2	.0	6.0	24.4	77.8	102.2
Average		.0	.8	.8	9.9	.0	.8	.0	.4	.0	2.7	6.5	38.8	45.3

Table 19. Transient areal recharge rates for the Minnelusa and Madison hydrogeologic units by zones.

[From Long and Putnam (2002). Areal recharge zones shown in figure 14. W, winter; S, summer (for example, W-88 = winter, water year 1988); Mnl, Minnelusa hydrogeologic unit; Mdn, Madison hydrogeologic unit]

Stress period	Other stress period identifier	Areal recharge by zones (cubic feet per second)										Total areal recharge (cubic feet per second)		
		Elk Creek to Little Elk Creek (Zone 1)		Little Elk Creek to Boxelder Creek (Zone 2)		Boxelder Creek to Rapid Creek (Zone 3)		Rapid Creek to Spring Creek (Zone 4)		Spring Creek to Battle Creek (Zone 5)				
		Mnl	Mdn	Mnl	Mdn	Mnl	Mdn	Mnl	Mdn	Mnl	Mdn	Mnl	Mdn	Mnl, Mdn
1	W-88	0.1	0.5	0.1	0.3	0.0	0.3	0.0	0.0	0.0	0.0	0.2	0.9	1.1
2	S-88	.1	.9	.1	.6	.1	.6	.0	.1	.2	.2	.5	1.9	2.4
3	W-89	.1	.5	.2	.6	.2	.6	.0	.1	.1	.1	.6	1.6	2.2
4	S-89	.3	2.4	.9	3.1	1.0	3.1	.8	1.6	1.0	.9	4.0	10.0	14.0
5	W-90	.2	1.5	.3	1.1	.2	1.1	.1	.1	.2	.2	1.0	3.2	4.2
6	S-90	.1	.9	.7	2.2	.6	2.2	.5	.8	2.2	2.3	4.1	7.4	11.5
7	W-91	.1	.8	.2	.6	.1	.6	.0	.1	.1	.1	.5	1.8	2.3
8	S-91	1.3	7.5	6.4	24.0	4.9	24.0	2.6	5.5	4.3	4.5	19.5	52.5	72.0
9	W-92	.1	.9	.2	.8	.1	.8	.1	.1	.3	.3	.8	2.4	3.2
10	S-92	.2	1.0	.9	3.1	.4	3.1	.2	.4	.6	.5	2.3	5.9	8.2
11	W-93	.1	1.1	.2	.7	.1	.7	.0	.1	.1	.1	.5	2.2	2.7
12	S-93	1.7	10.0	5.3	19.6	3.5	19.6	2.3	4.0	7.1	6.2	19.9	46.0	65.9
13	W-94	.2	1.8	.2	.8	.1	.8	.0	.0	.0	.1	.5	2.9	3.4
14	S-94	.1	.4	.3	.8	.1	.8	.1	.1	.4	.4	1.0	2.0	3.0
15	W-95	.4	3.1	.6	2.4	.3	2.4	.0	.1	.2	.3	1.5	6.5	8.0
16	S-95	2.9	18.4	11.6	35.8	4.6	35.8	1.1	2.1	5.5	5.5	25.7	72.7	98.4
17	W-96	.3	2.8	.5	2.2	.2	2.2	.1	.1	.2	.2	1.3	5.8	7.1
18	S-96	1.1	4.5	4.4	17.8	2.1	17.8	.9	1.5	3.0	3.0	11.5	31.1	42.6
19	W-97	.6	2.6	1.2	4.8	.7	4.8	.2	.4	.5	.5	3.2	9.6	12.8
20	S-97	1.4	5.8	11.6	35.0	5.9	35.0	1.1	1.8	3.3	3.1	23.3	56.3	79.6
Average		.6	3.4	2.3	7.8	1.3	7.8	.5	1.0	1.5	1.4	6.1	16.1	22.2

Table 20. Observed and simulated hydraulic head values for the Minnelusa hydrogeologic unit in steady-state simulation.

[Observation name corresponds to site number in table 29 and plate 2 of Long and Putnam (2002). Observed hydraulic head from Long and Putnam (2002). Implied accuracy is standard deviation of measurement error (SDME), which is used in the parameter estimation process to weight observations (weight = $1/(\text{SDME})^2$) (Hill and others, 2000). Implied accuracy for sites with a single measurement was assigned a value of 10. Implied accuracy for sites with time-series measurements was assigned a value of 1. Implied accuracy for sites with multiple measurements was assigned a value of 5. Other identifier used in Long and Putnam (2002). NGVD 29, National Geodetic Vertical Datum of 1929; --, none]

Observation name and site number	Model cell location			Observed hydraulic head (feet above NGVD 29)	Simulated hydraulic head (feet above NGVD 29)	Difference between simulated and observed hydraulic head (feet)	Implied accuracy of hydraulic head (feet)	Station identification number	Other identifier
	Layer	Row	Column						
102	1	2	89	3,454	3,486	32	10	441812103230501	--
103	1	1	75	3,564	3,574	10	10	441826103263301	--
104	1	1	70	3,610	3,602	-8	1	441759103261201	TF-1
105	1	5	74	3,515	3,573	58	10	441701103251101	PDMT-East
106	1	4	65	3,592	3,653	61	10	441656103261601	--
107	1	4	65	3,593	3,652	59	10	441649103262401	--
108	1	6	65	3,529	3,624	95	10	441620103255301	--
109	1	7	68	3,532	3,591	59	10	441612103252801	--
110	1	8	68	3,557	3,586	29	10	441606103252501	--
111	1	8	66	3,514	3,591	77	10	441556103253201	--
112	1	13	79	3,428	3,486	58	10	441504103230301	--
113	1	13	74	3,422	3,524	102	10	441500103240601	--
114	1	13	69	3,486	3,563	77	10	441453103243801	--
115	1	12	66	3,527	3,582	55	10	442221103245501	--
116	1	16	83	3,412	3,458	46	10	441314103233301	--
117	1	14	67	3,506	3,583	77	10	441424103243601	--
118	1	15	68	3,510	3,576	66	10	441421103242302	--
119	1	17	79	3,478	3,491	13	10	441400103231501	--
120	1	16	67	3,526	3,594	68	10	441358103241801	--
121	1	18	73	3,490	3,547	57	10	441643103232801	--
122	1	18	73	3,530	3,550	20	10	441339103232701	--
123	1	18	72	3,528	3,557	29	10	441338103233201	--
124	1	19	78	3,481	3,508	27	1	441337103225001	PDMT-1
125	1	19	80	3,414	3,487	73	10	441335103223202	--
126	1	19	80	3,414	3,487	73	10	441335103223201	--
127	1	19	81	3,457	3,483	26	10	441333103222801	--
128	1	19	75	3,454	3,531	77	10	441331103230701	--
129	2	19	69	3,547	3,690	143	10	441322103234202	--
130	2	19	69	3,581	3,669	88	10	441322103234201	--
131	1	20	78	3,489	3,497	8	10	441315103224001	--
132	1	21	78	3,458	3,492	34	10	441257103222501	--
133	1	26	109	3,268	3,278	10	10	441241103183401	--
134	1	25	89	3,414	3,418	4	10	441219103204801	--
135	1	24	79	3,495	3,499	4	10	441242103215701	--
136	1	24	79	3,477	3,499	22	10	441216103215801	--

Table 20. Observed and simulated hydraulic head values for the Minnelusa hydrogeologic unit in steady-state simulation.—Continued

[Observation name corresponds to site number in table 29 and plate 2 of Long and Putnam (2002). Observed hydraulic head from Long and Putnam (2002). Implied accuracy is standard deviation of measurement error (SDME), which is used in the parameter estimation process to weight observations (weight = $1/(\text{SDME})^2$) (Hill and others, 2000). Implied accuracy for sites with a single measurement was assigned a value of 10. Implied accuracy for sites with time-series measurements was assigned a value of 1. Implied accuracy for sites with multiple measurements was assigned a value of 5. Other identifier used in Long and Putnam (2002). NGVD 29, National Geodetic Vertical Datum of 1929; --, none]

Observation name and site number	Model cell location			Observed hydraulic head (feet above NGVD 29)	Simulated hydraulic head (feet above NGVD 29)	Difference between simulated and observed hydraulic head (feet)	Implied accuracy of hydraulic head (feet)	Station identification number	Other identifier
	Layer	Row	Column						
137	1	25	81	3,483	3,488	5	10	441210103215101	--
138	1	26	93	3,366	3,393	27	10	441212103201501	--
139	1	25	81	3,458	3,483	25	10	441211103213701	--
140	1	25	84	3,476	3,462	-14	10	441202103211301	--
141	1	26	90	3,449	3,412	-37	10	441318103221301	--
142	1	25	81	3,444	3,489	45	10	442137103213801	--
143	1	25	81	3,474	3,485	11	10	441202103213301	--
144	1	26	81	3,471	3,483	12	10	441200103213001	--
145	1	26	84	3,445	3,465	20	10	441157103213701	--
146	1	26	80	3,451	3,490	39	10	441220103214001	--
147	1	26	82	3,459	3,479	20	10	441151103213002	--
148	1	29	95	3,413	3,388	-25	10	441147103270001	--
149	2	27	80	3,415	3,581	166	10	441136103221301	--
150	1	27	81	3,460	3,485	25	10	441134103211601	--
151	1	29	91	3,418	3,416	-2	10	441131103200101	--
152	1	28	84	3,466	3,467	1	10	441130103205601	--
153	1	28	82	3,453	3,480	27	10	441119103210301	--
154	1	31	95	3,454	3,398	-56	10	441105103192001	--
155	1	30	86	3,413	3,462	49	10	441125103203001	--
156	1	30	83	3,403	3,482	79	10	441038103203001	--
157	1	31	83	3,423	3,479	56	10	441049103203701	--
158	1	31	84	3,423	3,474	51	10	441047103203301	--
159	2	31	82	3,496	3,588	92	10	441323103232301	--
160	1	31	83	3,438	3,477	39	10	441040103203201	--
161	1	31	83	3,492	3,476	-16	10	441038103203002	--
162	2	31	83	3,472	3,583	111	10	441037103203502	--
163	2	31	83	3,450	3,583	133	10	441037103203501	--
164	1	33	91	3,463	3,433	-30	10	441033103193001	--
165	1	32	88	3,443	3,449	6	10	441031103195201	--
166	1	32	85	3,487	3,464	-23	10	441024103201301	--
167	2	32	81	3,474	3,604	130	10	441023103204001	--
168	2	32	82	3,440	3,593	153	10	441023103203101	--
169	2	32	82	3,423	3,590	167	10	441023103203001	--
170	1	33	84	3,432	3,474	42	10	441007103201101	--
171	1	39	100	3,489	3,439	-50	10	440930103175201	--

Table 20. Observed and simulated hydraulic head values for the Minnelusa hydrogeologic unit in steady-state simulation.—Continued

[Observation name corresponds to site number in table 29 and plate 2 of Long and Putnam (2002). Observed hydraulic head from Long and Putnam (2002). Implied accuracy is standard deviation of measurement error (SDME), which is used in the parameter estimation process to weight observations (weight = $1/(\text{SDME})^2$) (Hill and others, 2000). Implied accuracy for sites with a single measurement was assigned a value of 10. Implied accuracy for sites with time-series measurements was assigned a value of 1. Implied accuracy for sites with multiple measurements was assigned a value of 5. Other identifier used in Long and Putnam (2002). NGVD 29, National Geodetic Vertical Datum of 1929; --, none]

Observation name and site number	Model cell location			Observed hydraulic head (feet above NGVD 29)	Simulated hydraulic head (feet above NGVD 29)	Difference between simulated and observed hydraulic head (feet)	Implied accuracy of hydraulic head (feet)	Station identification number	Other identifier
	Layer	Row	Column						
172	1	37	87	3,447	3,460	13	10	440936103191801	Blackhawk Water Co. No. 3
173	1	42	106	3,416	3,402	-14	10	440919103170501	--
174	1	44	110	3,379	3,372	-7	10	440929103163501	--
175	1	45	110	3,392	3,379	-13	10	440916103163301	--
176	1	40	92	3,386	3,457	71	10	440907103183501	Blackhawk Water Co. No. 2
177	1	38	83	3,462	3,467	5	10	440908103193101	--
178	1	40	90	3,403	3,460	57	10	440901103184801	Blackhawk Water Co. No. 1
179	1	47	89	3,414	3,461	47	10	440829103183801	--
180	1	50	96	3,410	3,450	40	10	440826103174601	--
181	1	50	92	3,422	3,456	34	1	440818103180801	BLHK-1
182	1	50	91	3,449	3,457	8	10	440818103174701	--
183	1	59	104	3,313	3,430	117	10	440800103163001	--
184	1	60	93	3,409	3,450	41	10	440738103173901	--
185	1	60	93	3,412	3,450	38	10	440736103173701	--
186	1	63	90	3,434	3,453	19	10	440719103174801	--
187	1	79	107	3,400	3,385	-15	10	440636103152001	--
188	1	77	99	3,420	3,418	-2	10	440635103161801	--
189	1	82	102	3,383	3,399	16	10	440607103155901	--
190	1	77	85	3,419	3,445	26	10	440606103174701	--
191	1	77	86	3,419	3,442	23	10	440605103174701	--
192	1	77	84	3,421	3,445	24	10	440552103173401	--
193	1	77	85	3,427	3,443	16	10	440603103174501	--
194	1	82	95	3,336	3,412	76	10	440552103162301	--
195	1	79	85	3,414	3,441	27	10	440557103174401	--
196	1	79	84	3,449	3,443	-6	10	440556103174801	--
197	1	80	87	3,410	3,433	23	10	440542103172501	--
198	1	80	85	3,443	3,437	-6	10	440639103173701	--
199	1	81	85	3,438	3,434	-4	10	440545103173501	--
200	1	80	81	3,450	3,448	-2	1	440544103180001	CQ-1
201	1	81	84	3,444	3,437	-7	10	440544103174301	--

Table 20. Observed and simulated hydraulic head values for the Minnelusa hydrogeologic unit in steady-state simulation.—Continued

[Observation name corresponds to site number in table 29 and plate 2 of Long and Putnam (2002). Observed hydraulic head from Long and Putnam (2002). Implied accuracy is standard deviation of measurement error (SDME), which is used in the parameter estimation process to weight observations (weight = $1/(\text{SDME})^2$) (Hill and others, 2000). Implied accuracy for sites with a single measurement was assigned a value of 10. Implied accuracy for sites with time-series measurements was assigned a value of 1. Implied accuracy for sites with multiple measurements was assigned a value of 5. Other identifier used in Long and Putnam (2002). NGVD 29, National Geodetic Vertical Datum of 1929; --, none]

Observation name and site number	Model cell location			Observed hydraulic head (feet above NGVD 29)	Simulated hydraulic head (feet above NGVD 29)	Difference between simulated and observed hydraulic head (feet)	Implied accuracy of hydraulic head (feet)	Station identification number	Other identifier
	Layer	Row	Column						
202	1	82	85	3,468	3,434	-34	10	440544103173601	--
205	1	87	96	3,328	3,395	67	10	440538103161001	--
206	1	83	82	3,467	3,435	-32	10	440534103174801	--
207	1	85	87	3,450	3,421	-29	10	440534103171401	--
209	1	89	95	3,412	3,396	-16	1	440528103161001	CP
210	1	90	98	3,406	3,391	-15	10	440528103155201	--
211	1	82	74	3,471	3,442	-29	10	440521103184401	--
212	1	85	78	3,462	3,428	-34	10	440532103182001	--
213	1	92	94	3,403	3,393	-10	10	440514103161401	--
214	1	85	76	3,434	3,430	-4	10	450512103182101	--
215	2	83	65	3,406	3,478	72	10	440516103194001	WT-1
216	1	88	76	3,475	3,422	-53	10	440501103181501	--
217	1	88	77	3,450	3,419	-31	10	440459103181001	--
218	1	88	76	3,450	3,419	-31	10	440458103181101	--
219	1	97	95	3,396	3,383	-13	10	440452103155301	South Dakota ARNG
220	1	90	75	3,435	3,415	-20	10	440449103181501	--
221	1	90	75	3,415	3,414	-1	10	440446103181801	--
222	1	91	75	3,470	3,413	-57	10	440445103181601	--
223	1	101	92	3,392	3,375	-17	1	440430103160201	SP-1
224	1	95	74	3,400	3,399	-1	1	440423103180501	WCR-3
225	1	97	69	3,382	3,400	18	10	440403103183701	--
226	2	99	64	3,435	3,451	16	10	440347103190501	--
227	2	99	63	3,456	3,452	-4	10	440344103190801	--
228	1	104	75	3,360	3,373	13	10	440344103173701	--
229	1	106	75	3,360	3,369	9	1	440338103173301	CL-1
230	1	113	89	3,396	3,363	-33	10	440331103155101	--
231	1	106	70	3,395	3,384	-11	5	440326103180703	JS-1B
232	1	107	69	3,365	3,385	20	10	440321103181001	--
233	1	110	72	3,383	3,378	-5	1	440310103173801	CHLN-1
234	1	113	78	3,360	3,376	16	10	440309103170101	--
235	2	108	65	3,435	3,461	26	10	440308103183001	--
236	1	112	74	3,378	3,377	-1	10	440308103172501	--
237	1	121	97	3,353	3,325	-28	10	440308103144301	--
238	1	112	74	3,353	3,378	25	10	440336103165301	--

Table 20. Observed and simulated hydraulic head values for the Minnelusa hydrogeologic unit in steady-state simulation.—Continued

[Observation name corresponds to site number in table 29 and plate 2 of Long and Putnam (2002). Observed hydraulic head from Long and Putnam (2002). Implied accuracy is standard deviation of measurement error (SDME), which is used in the parameter estimation process to weight observations (weight = $1/(SDME)^2$) (Hill and others, 2000). Implied accuracy for sites with a single measurement was assigned a value of 10. Implied accuracy for sites with time-series measurements was assigned a value of 1. Implied accuracy for sites with multiple measurements was assigned a value of 5. Other identifier used in Long and Putnam (2002). NGVD 29, National Geodetic Vertical Datum of 1929; --, none]

Observation name and site number	Model cell location			Observed hydraulic head (feet above NGVD 29)	Simulated hydraulic head (feet above NGVD 29)	Difference between simulated and observed hydraulic head (feet)	Implied accuracy of hydraulic head (feet)	Station identification number	Other identifier
	Layer	Row	Column						
239	1	109	67	3,399	3,389	-10	10	440307103181501	--
240	1	114	78	3,364	3,377	13	10	440305103164501	--
241	1	115	78	3,370	3,377	7	10	440211103165601	--
242	1	115	78	3,350	3,378	28	10	440301103165701	--
243	1	114	76	3,380	3,378	-2	10	440300103170901	--
244	1	115	75	3,355	3,379	24	10	440257103171101	--
245	1	115	74	3,336	3,381	45	10	440252103172201	--
246	2	112	65	3,450	3,467	17	10	440249103182101	--
247	1	117	75	3,342	3,381	39	10	440247103165401	--
248	1	118	75	3,386	3,382	-4	10	440242103170801	--
249	1	121	82	3,374	3,378	4	10	440244103154102	--
250	1	126	92	3,351	3,341	-10	10	440236103145801	--
251	1	120	74	3,387	3,384	-3	10	440233103170501	--
252	1	123	81	3,381	3,381	0	5	440223103161701	Springbrook West Well No. 2
253	2	120	71	3,372	3,439	67	10	440224103172601	Carriage Hills No. 2
254	1	126	86	3,388	3,368	-20	5	440214103153501	Springbrook East Well
255	1	124	73	3,401	3,388	-13	10	440211103165201	--
256	1	125	76	3,409	3,387	-22	10	440211103164301	--
257	1	132	89	3,392	3,348	-44	10	440204103150001	--
258	1	134	93	3,390	3,335	-55	10	440203103143601	--
259	1	131	80	3,429	3,390	-39	5	440158103160401	Springbrook South No. 3
260	1	129	73	3,412	3,406	-6	1	440149103164901	Wildwood North
261	1	134	73	3,429	3,423	-6	5	440130103163401	Wildwood South
262	1	144	86	3,376	3,360	-16	10	440103103144801	--
263	1	145	82	3,456	3,396	-60	10	440059103154101	--
264	1	147	85	3,455	3,377	-78	10	440048103145001	--
265	1	141	63	3,475	3,485	10	10	440038103172601	--
266	1	145	70	3,459	3,469	10	10	440031103162701	--

Table 20. Observed and simulated hydraulic head values for the Minnelusa hydrogeologic unit in steady-state simulation.—Continued

[Observation name corresponds to site number in table 29 and plate 2 of Long and Putnam (2002). Observed hydraulic head from Long and Putnam (2002). Implied accuracy is standard deviation of measurement error (SDME), which is used in the parameter estimation process to weight observations (weight = $1/(SDME)^2$) (Hill and others, 2000). Implied accuracy for sites with a single measurement was assigned a value of 10. Implied accuracy for sites with time-series measurements was assigned a value of 1. Implied accuracy for sites with multiple measurements was assigned a value of 5. Other identifier used in Long and Putnam (2002). NGVD 29, National Geodetic Vertical Datum of 1929; --, none]

Observation name and site number	Model cell location			Observed hydraulic head (feet above NGVD 29)	Simulated hydraulic head (feet above NGVD 29)	Difference between simulated and observed hydraulic head (feet)	Implied accuracy of hydraulic head (feet)	Station identification number	Other identifier
	Layer	Row	Column						
267	1	147	73	3,412	3,458	46	10	440028103160801	--
268	1	147	72	3,499	3,460	-39	10	440027103161001	--
269	1	146	69	3,494	3,478	-16	10	440022103163401	--
270	1	148	69	3,455	3,480	25	10	440007103165301	--
271	1	152	75	3,404	3,448	44	10	440010103154201	--
272	1	160	65	3,493	3,473	-20	1	435916103161802	RG-2
273	1	164	67	3,476	3,466	-10	10	435858103155701	--
274	1	175	62	3,539	3,449	-90	10	435803103160301	--
275	1	177	58	3,548	3,447	-101	10	445809103162201	--
276	1	183	75	3,547	3,414	-133	10	435720103141601	Hart Ranch No. 3
277	1	193	36	3,558	3,456	-102	10	435352103170801	--
278	2	196	27	3,527	3,514	-13	10	435325103171701	--
279	1	199	25	3,551	3,467	-84	10	435225103172801	--
280	1	203	17	3,536	3,504	-32	10	435119103175001	--

Table 21. Observed and simulated hydraulic head values for the Madison hydrogeologic unit in steady-state simulation.

[Observation name corresponds to site number in table 29 and plate 2 of Long and Putnam (2002). Estimated hydraulic head from Long and Putnam (2002). Implied accuracy is standard deviation of measurement error (SDME), which is used in the parameter estimation process to weight observations (weight = $1/(SDME^2)$) (Hill and others, 2000). Implied accuracy for sites with a single measurement was assigned a value of 10. Implied accuracy for sites with time-series measurements was assigned a value of 1. Implied accuracy for sites with multiple measurements was assigned a value of 5. Other identifier used in Long and Putnam (2002). NGVD 29, National Geodetic Vertical Datum of 1929; --, none]

Observation name and site number	Model cell location			Estimated hydraulic head (feet above NGVD 29)	Simulated hydraulic head (feet above NGVD 29)	Difference between simulated and observed hydraulic head (feet)	Implied accuracy of hydraulic head (feet)	Station identification number	Other identifier
	Layer	Row	Column						
1	3	1	70	3,621	3,684	63	1	441759103261202	TF-2
2	3	12	60	3,655	3,740	85	10	441500103253501	--
3	3	19	78	3,603	3,535	-68	1	441337103225002	PDMT-2
4	3	24	82	3,494	3,534	40	10	441220103213601	--
5	4	24	63	3,854	3,902	48	10	441157103234301	--
6	4	30	37	4,340	4,185	-155	10	440958103253401	--
7	4	30	39	4,280	4,160	-120	10	440934103252001	--
8	4	30	40	4,280	4,135	-145	10	440933103250701	--
9	3	40	92	3,511	3,480	-31	10	441636103183801	Black Hawk Water Co. No. 4
10	3	86	163	2,687	2,656	-31	10	440851103044801	Ellsworth AFB
11	3	37	57	3,895	3,839	-56	10	440828103222601	--
12	3	50	96	3,450	3,471	21	10	440826103174701	--
13	3	55	106	3,448	3,448	0	10	440823103162701	--
14	4	37	49	3,981	3,944	-37	10	440820103232601	--
15	3	39	57	3,808	3,813	5	1	440811103222201	DOTY
16	3	38	51	3,973	3,887	-86	10	440807103225801	--
17	3	38	50	3,976	3,900	-76	10	440804103230501	--
18	3	40	55	3,874	3,827	-47	10	440804103223001	--
19	3	40	56	3,774	3,809	35	10	440804103222701	--
21	4	39	51	3,913	3,873	-40	10	440758103225601	--
22	4	39	47	3,980	3,899	-81	10	440753103232401	--
23	3	42	56	3,772	3,784	12	10	440749103221501	--
24	4	42	53	3,876	3,815	-61	10	440744103223401	--
25	3	53	58	3,660	3,698	38	10	440658103213001	--
26	3	79	119	3,423	3,414	-9	10	440655103140501	RC-8
27	3	63	74	3,528	3,529	1	10	440650103193201	--
28	3	61	59	3,645	3,634	-11	10	440631103211201	--
29	3	67	73	3,536	3,521	-15	10	440630103192501	--
30	3	117	163	2,575	2,564	-11	10	440629103040901	City of Box Elder
31	3	81	105	3,481	3,415	-66	10	440622103152701	Coke Plant
32	3	84	105	3,444	3,408	-36	10	440612103152001	RC-10
33	3	80	81	3,451	3,454	3	1	440544103180002	CQ-2

Table 21. Observed and simulated hydraulic head values for the Madison hydrogeologic unit in steady-state simulation.—Continued

[Observation name corresponds to site number in table 29 and plate 2 of Long and Putnam (2002). Estimated hydraulic head from Long and Putnam (2002). Implied accuracy is standard deviation of measurement error (SDME), which is used in the parameter estimation process to weight observations (weight = $1/(SDME^2)$) (Hill and others, 2000). Implied accuracy for sites with a single measurement was assigned a value of 10. Implied accuracy for sites with time-series measurements was assigned a value of 1. Implied accuracy for sites with multiple measurements was assigned a value of 5. Other identifier used in Long and Putnam (2002). NGVD 29, National Geodetic Vertical Datum of 1929; --, none]

Observation name and site number	Model cell location			Estimated hydraulic head (feet above NGVD 29)	Simulated hydraulic head (feet above NGVD 29)	Difference between simulated and observed hydraulic head (feet)	Implied accuracy of hydraulic head (feet)	Station identification number	Other identifier
	Layer	Row	Column						
34	3	70	55	3,646	3,581	-65	10	440540103211301	--
35	3	85	84	3,447	3,425	-22	10	440526103173001	RC-6
36	3	91	97	3,424	3,404	-20	10	440523103155701	BHPL
37	3	80	67	3,477	3,509	32	10	440518103193001	--
38	4	73	46	3,734	3,694	-40	10	440512103215701	--
39	4	73	45	3,764	3,716	-48	10	440508103220701	--
40	3	83	63	3,520	3,513	-7	10	440500103193601	WT-2
41	3	88	72	3,405	3,432	27	10	440453103184001	--
42	3	86	64	3,430	3,468	38	10	440446103193201	--
43	3	97	91	3,417	3,402	-15	10	440446103161701	LC
44	3	87	64	3,428	3,456	28	10	440441103193301	--
45	3	90	66	3,405	3,417	12	10	440432103191401	--
46	3	101	91	3,423	3,401	-22	1	440430103160202	SP-2
47	3	110	114	3,146	3,198	52	1	440427103131701	RC-7
48	3	99	62	3,418	3,416	-2	10	440337103191801	--
49	3	110	87	3,415	3,396	-19	10	440342103160701	RC-9
50	3	106	75	3,418	3,405	-13	1	440338103173302	CL-2
51	3	132	138	2,655	2,584	-71	10	440334103095601	Rapid Valley No. 4, RV-4
52	3	106	70	3,418	3,402	-16	1	440326103180702	JS-1A
53	3	110	73	3,418	3,406	-12	10	440310103173802	CHLN-3
54	3	107	63	3,417	3,415	-2	1	440308103184601	CLEGmdsn
55	3	110	69	3,419	3,405	-14	10	440308103180701	--

Table 21. Observed and simulated hydraulic head values for the Madison hydrogeologic unit in steady-state simulation.—Continued

[Observation name corresponds to site number in table 29 and plate 2 of Long and Putnam (2002). Estimated hydraulic head from Long and Putnam (2002). Implied accuracy is standard deviation of measurement error (SDME), which is used in the parameter estimation process to weight observations (weight = $1/(\text{SDME}^2)$) (Hill and others, 2000). Implied accuracy for sites with a single measurement was assigned a value of 10. Implied accuracy for sites with time-series measurements was assigned a value of 1. Implied accuracy for sites with multiple measurements was assigned a value of 5. Other identifier used in Long and Putnam (2002). NGVD 29, National Geodetic Vertical Datum of 1929; --, none]

Observation name and site number	Model cell location			Estimated hydraulic head (feet above NGVD 29)	Simulated hydraulic head (feet above NGVD 29)	Difference between simulated and observed hydraulic head (feet)	Implied accuracy of hydraulic head (feet)	Station identification number	Other identifier
	Layer	Row	Column						
56	3	112	72	3,419	3,408	-11	10	440300103173501	CHLN-2
57	3	108	61	3,415	3,419	4	10	440247103192401	--
58	3	107	56	3,438	3,427	-11	10	440302103194601	--
59	3	110	57	3,434	3,426	-8	10	440247103191701	--
60	3	113	60	3,416	3,422	6	10	440238103185201	--
61	3	120	70	3,426	3,420	-6	10	440223103173201	Carriage Hills No. 1, CH
62	3	124	77	3,423	3,422	-1	10	440220103164001	RC-11
63	3	124	70	3,445	3,426	-19	10	440205103172001	--
64	3	137	62	3,475	3,463	-12	10	440054103173801	--
65	4	137	43	3,806	3,626	-180	10	440523103194201	--
66	3	146	58	3,531	3,543	12	10	440004103174001	HH
67	4	148	48	3,626	3,639	13	10	435937103184101	--
68	3	160	65	3,497	3,511	14	1	435916103161801	RG
69	4	156	49	3,594	3,617	23	10	435903103181301	--
70	4	157	44	3,580	3,639	59	10	435851103184201	--
71	3	170	77	3,498	3,476	-22	10	435851103143501	HR-2
72	3	176	55	3,488	3,514	26	10	435746103160601	--
73	3	187	84	3,498	3,375	-123	10	435718103130301	HR-1
74	3	183	38	3,515	3,552	37	10	435635103181401	PG
75	3	187	52	3,509	3,481	-28	10	435619103161901	--
76	3	189	38	3,530	3,509	-21	10	435518103173001	--
77	3	198	13	3,542	3,620	78	10	435227103185301	HWRD

Table 22. Supplemental interpolated hydraulic head values for the Minnelusa and Madison hydrogeologic units in steady-state simulation.

[Interpolated hydraulic head from estimated potentiometric contours in Long and Putnam (2002, plates 1 and 2). Implied accuracy is standard deviation of measurement error (SDME), which is used in the parameter estimation process to weight observations ($\text{weight} = 1/(\text{SDME})^2$) (Hill and others, 2000). The SDME or implied accuracy for these supplemental sites was set at 10 feet. NGVD 29, National Geodetic Vertical Datum of 1929]

Observation name	Model cell location			Interpolated hydraulic head (feet above NGVD 29)	Simulated hydraulic head (feet above NGVD 29)	Difference between simulated and interpolated hydraulic head (feet)	Implied accuracy of hydraulic head (feet)
	Layer	Row	Column				
Minnelusa hydrogeologic unit							
801	1	17	162	3,020	3,001	-19	10
802	1	11	102	3,315	3,382	67	10
803	1	21	152	3,160	3,131	-29	10
804	1	34	128	3,266	3,197	-69	10
805	1	34	164	2,880	2,849	-31	10
806	1	69	156	3,070	3,010	-60	10
807	1	176	133	3,100	3,094	-6	10
808	1	193	163	2,780	2,816	36	10
809	1	193	139	3,130	3,043	-87	10
810	1	195	90	3,430	3,269	-161	10
811	1	122	126	3,150	3,184	34	10
812	1	76	161	2,900	2,890	-10	10
813	1	184	157	2,900	2,927	27	10
814	1	128	159	2,800	2,903	103	10
815	1	139	115	3,200	3,225	25	10
816	1	212	157	3,105	2,953	-152	10
817	1	104	124	3,200	3,227	27	10
818	1	127	154	2,900	3,006	106	10
819	1	122	142	3,000	3,093	93	10
820	1	162	141	3,000	3,063	63	10

Table 22. Supplemental interpolated hydraulic head values for the Minnelusa and Madison hydrogeologic units in steady-state simulation. —Continued

[Interpolated hydraulic head from estimated potentiometric contours in Long and Putnam (2002, plates 1 and 2). Implied accuracy is standard deviation of measurement error (SDME), which is used in the parameter estimation process to weight observations (weight = $1/(\text{SDME})^2$ (Hill and others, 2000). The SDME or implied accuracy for these supplemental sites was set at 10 feet. NGVD 29, National Geodetic Vertical Datum of 1929]

Observation name	Model cell location			Interpolated hydraulic head (feet above NGVD 29)	Simulated hydraulic head (feet above NGVD 29)	Difference between simulated and interpolated hydraulic head (feet)	Implied accuracy of hydraulic head (feet)
	Layer	Row	Column				
Madison hydrogeologic unit							
901	3	17	162	3,010	2,989	-21	10
902	3	11	102	3,300	3,399	99	10
903	3	21	152	3,100	3,126	26	10
904	3	34	128	3,200	3,208	8	10
905	3	34	164	2,880	2,792	-88	10
906	3	69	156	3,050	2,822	-228	10
907	3	176	133	2,900	2,854	-46	10
908	3	193	163	2,775	2,708	-67	10
909	3	193	139	3,090	2,973	-117	10
910	3	195	90	3,400	3,266	-134	10
911	3	122	126	2,910	2,914	4	10
912	3	76	161	2,880	2,724	-156	10
913	3	184	157	2,710	2,700	-10	10
914	3	128	159	2,570	2,588	18	10
915	3	139	115	3,020	3,038	18	10
916	3	212	157	3,100	2,897	-203	10
917	3	104	124	3,100	3,127	27	10
918	3	127	154	2,600	2,604	4	10
919	3	122	142	2,700	2,623	-77	10
920	3	162	141	2,700	2,648	-52	10

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For more information concerning this publication, contact:
Director, USGS South Dakota Water Science Center
1608 Mt. View Rd.
Rapid City, SD 57702
(605) 394-3200

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