science for a changing world

Prepared in cooperation with the
CITY OF RAPID CITY

Flow-System Analysis of the Madison and Minnelusa Aquifers in the Rapid City Area, South Dakota—Conceptual Model

Water-Resources Investigations Report 02-4185

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

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

Flow-System Analysis of the Madison and Minnelusa Aquifers in the Rapid City Area, South Dakota—Conceptual Model

By Andrew J. Long and Larry D. Putnam

Water-Resources Investigations Report 02-4185

Prepared in cooperation with the
CITY OF RAPID CITY

U.S. Department of the Interior

GALE A. NORTON, Secretary

U.S. Geological Survey

Charles G. Groat, Director

Rapid City, South Dakota: 2002

For additional information write to:

District Chief
U.S. Geological Survey
1608 Mt. View Road
Rapid City, SD 57702

Copies of this report can be purchased from:

U.S. Geological Survey
Information Services
Building 810
Box 25286, Federal Center
Denver, CO 80225-0286

CONTENTS

CONTENTS—Continued

PLATES

[Plates are in pocket]

FIGURES

FIGURES—Continued

TABLES

CONVERSION FACTORS AND VERTICAL DATUM

Multiply	By	To obtain
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot per mile (ft/mi)	0.1894	meter per kilometer
foot squared per day (ft^2/d)	0.09290	meter squared per day
gallon per day (gal/d)	0.003785	cubic meter per day
gallon per minute (gal/min)	0.06309	liter per second
million gallons per day (Mgal/d)	0.04381	cubic meter per second
inch	2.54	centimeter
inch	25.4	millimeter
inch per year (in/yr)	25.4	millimeter per year
mile (mi)	1.609	kilometer
square mile (mi^2)	259.0	hectare
square mile (mi^2)	2.590	square kilometer

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

$$°C = (°F - 32) / 1.8$$

Sea level: In this report, “sea level” refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)—a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called Sea Level Datum of 1929.

Water year (WY): Water year is the 12-month period, October 1 through September 30, and is designated by the calendar year in which it ends. Thus, the water year ending September 30, 1999, is called “WY99.”

SYMBOL DEFINITIONS AND DIMENSIONS

Symbol	Dimensions	Description
K	L/T	Horizontal hydraulic conductivity
K_v	L/T	Vertical hydraulic conductivity
S	dimensionless	Storage coefficient
S_y	dimensionless	Specific yield (unconfined aquifer)
T	L^2/T	Transmissivity
Q	V/T	Discharge rate
A	A	Area
ΔH	L	Hydraulic head potential
ΔL	L	Length over which ΔH applies
ET	L	Evapotranspiration

L = length

T = time

A = area

V = volume

OTHER DEFINITIONS

Aquifer analysis area - Includes the outcrops and aquifer areas of the Madison and Minnelusa hydrogeologic units (fig. 1).

Areal recharge - Infiltration from precipitation falling directly on the outcrop recharge areas.

Basin yield - The rate of streamflow leaving a basin divided by the area of the basin, usually expressed in inches per year.

Unsaturated area - The area of a formation that does not contain a water table because the bottom of the dipping formation is at a higher altitude than the elevation of the water table.

High-flow area - The near-outcrop area that includes Rapid City and extends 2 to 3 mi north of Boxelder Creek where extensive tectonic activity, carbonate and sulfate dissolution, brecciation, ground-water recharge, and ground-water circulation has taken or is taking place.

Paired wells - Two wells in the same location or very close proximity that are each open in a different aquifer.

Streamflow recharge - Recharge from streamflow that loses water to the outcrop recharge areas.

Summer period - April through September (S-88 = summer 1988 = April 1988 to September 1988).

Winter period - October through March (W-88 = winter 1988 = October 1987 to March 1988).

Flow-System Analysis of the Madison and Minnelusa Aquifers in the Rapid City Area, South Dakota—Conceptual Model

By Andrew J. Long *and* Larry D. Putnam

ABSTRACT

The conceptual model of the Madison and Minnelusa aquifers in the Rapid City area synthesizes the physical geography, hydraulic properties, and ground-water flow components of these important aquifers. The Madison hydrogeologic unit includes the karstic Madison aquifer, which is defined as the upper, more permeable 100 to 200 ft of the Madison Limestone, and the Madison confining unit, which consists of the lower, less permeable part of the Madison Limestone and the Englewood Formation. Overlying the Madison hydrogeologic unit is the Minnelusa hydrogeologic unit, which includes the Minnelusa aquifer in the upper, more permeable 200 to 300 ft and the Minnelusa confining unit in the lower, less permeable part. The Madison and Minnelusa hydrogeologic units outcrop in the study area on the eastern flank of the Black Hills where recharge occurs from streamflow losses and areal recharge. The conceptual model describes streamflow recharge, areal recharge, ground-water flow, storage in aquifers and confining units, unsaturated areas, leakage between aquifers, discharge from artesian springs, and regional outflow.

Effective transmissivities estimated for the Madison aquifer range from 500 to 20,000 ft^2/d and for the Minnelusa aquifer from 500 to 10,000 ft^2/d. Localized anisotropic transmissivity in the Madison aquifer has tensor ratios as high as 45:1. Vertical hydraulic conductivities for the Minnelusa confining unit determined from aquifer tests range from $1.3x10^{-3}$ to $3.0x10^{-1}$ ft/d. The confined storage coefficient of the Madison and Minnelusa hydrogeologic units was estimated as $3x10^{-4}$ ft/d. Specific yield was estimated as 0.09 for the Madison and Minnelusa aquifers and 0.03 for the Madison and Minnelusa confining units. Potentiometric surfaces for the Madison and Minnelusa aquifers have a general easterly gradient of about 70 ft/mi with local variations. Temporal change in hydraulic head in the Madison and Minnelusa aquifers ranged from about 5 to 95 ft in water years 1988-97. The unconfined areas were estimated at about 53 and 36 mi^2 for the Madison and Minnelusa hydrogeologic units, respectively, in contrast to an aquifer analysis area of 629 mi^2.

Dye-tracer tests, stable isotopes, and hydrogeologic features were analyzed conjunctively to estimate generalized ground-water flowpaths in the Madison aquifer and their influences on the Minnelusa aquifer. The western Rapid City area between Boxelder Creek and Spring Creek was characterized as having undergone extensive tectonic activity, greater brecciation in the Minnelusa Formation, large transmissivities, generally upward hydraulic gradients from the Madison aquifer to the Minnelusa aquifer, many karst springs, and converging flowpaths.

Water-budget analysis included: (1) a dry-period budget for declining water levels; October 1, 1987, to March 31, 1993; (2) a wet-period budget for rising water levels, April 1,

1993, to September 30, 1997; and (3) a full 10-year period budget for water years 1988-97. By simultaneously balancing these water budgets, initial estimates of recharge, discharge, change in storage, and hydraulic properties were refined. Inflow rates for the 10-year budget included streamflow recharge of about 45 ft^3/s or 61 percent of the total budget and areal recharge of 22 ft^3/s or 30 percent. Streamflow recharge to the Madison hydrogeologic unit was about 86 percent of the total streamflow recharge. Outflow for the 10-year budget included springflow of 31 ft^3/s or 42 percent of the total budget, water use of about 10 ft^3/s or 14 percent, and regional outflow of 22 ft^3/s or 30 percent. Ground-water storage increased 9 ft^3/s during the 10-year period, and net ground-water movement from the Madison to Minnelusa hydrogeologic unit was about 8 ft^3/s.

INTRODUCTION

The Madison and Minnelusa aquifers are the main source of ground water in the Black Hills area (Driscoll and Carter, 2001). The city of Rapid City obtains more than one-half of its municipal water supply from these two bedrock aquifers via deep wells and springs. Numerous additional users in the Rapid City area obtain water from the Madison or Minnelusa aquifers for domestic, industrial, and irrigation usage. Ground-water flow within the Madison and Minnelusa aquifers is complex. Extensive fracturing, solution enhancement, and brecciation contribute to heterogeneity, anisotropic transmissivity, and spatially variable ground-water seepage between the two aquifers. A long-term cooperative study between the U.S. Geological Survey and the city of Rapid City has provided hydrogeologic data and interpretation for planning and management of the Madison and Minnelusa aquifers.

Purpose and Scope

The primary purpose of this report is to present a conceptual model of ground-water flow in the Madison and Minnelusa aquifers in the Rapid City area leading to a better understanding of the unique concepts involved. The conceptual model consists of discussions of hydraulic properties, hydraulic head, and ground-water flow. A detailed water-budget analysis encompassing WY88-97 characterizes and quantifies recharge and discharge for the study area and is used to refine estimates of hydraulic properties. An additional purpose of the conceptual model was to compile data for numerical modeling and other research efforts in the future.

Description of Study Area

The study area, which includes Rapid City and the surrounding area, extends north of Elk Creek, to the south near Battle Creek, east of the city of Boxelder, and west into the central Black Hills (fig. 1). Streams flowing from the central Black Hills contribute to streamflow losses that recharge the Madison and Minnelusa aquifers. The western extent of the study area comprises these stream basins, which include Elk, Boxelder, Rapid, Spring, and Battle Creeks. This western area is included in the study area for analysis of streamflow recharge and evapotranspiration. The eastern part of the study area, where the Madison and Minnelusa hydrogeologic units exist, is referred to as the aquifer analysis area (figs. 1 and 2). These hydrogeologic units are described in subsequent sections of the report.

Land-surface altitudes range from more than 7,000 ft in the western highlands of the study area to about 3,000 ft in the eastern lowlands. The western extent of the study area is characterized by high relief with predominantly pine and spruce forests. The eastern lowlands comprise approximately the eastern one-half of the aquifer analysis area and are characterized by rolling prairies with bottom lands along stream channels. The outcrops of the Madison and Minnelusa hydrogeologic units are located in the western part of the aquifer analysis area (fig. 2) 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 precipitation rates range from about 27 in/yr in the northwest to 17 in/yr in the eastern lowlands with most precipitation occurring in March, April, May, and June (Driscoll, Hamade, and Kenner, 2000). Summer temperatures in the western highlands generally are cooler and have less variation during the winter (table 1). As of 1999, about 88,000 people lived in the Rapid City metropolitan area (U.S. Census Bureau, 2000), whereas population is much sparser in other parts of the study area.

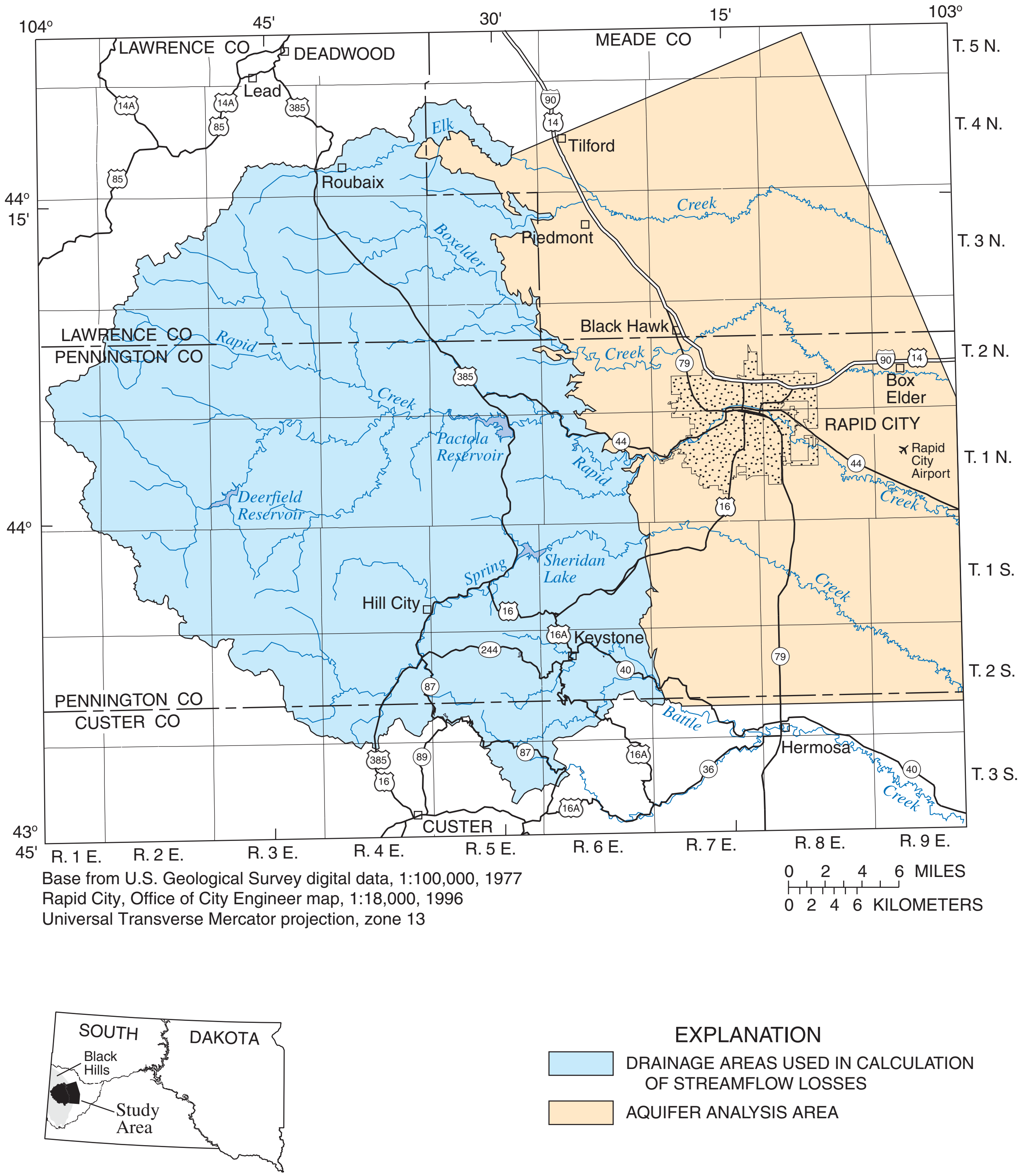


Figure 1. Location of study area.

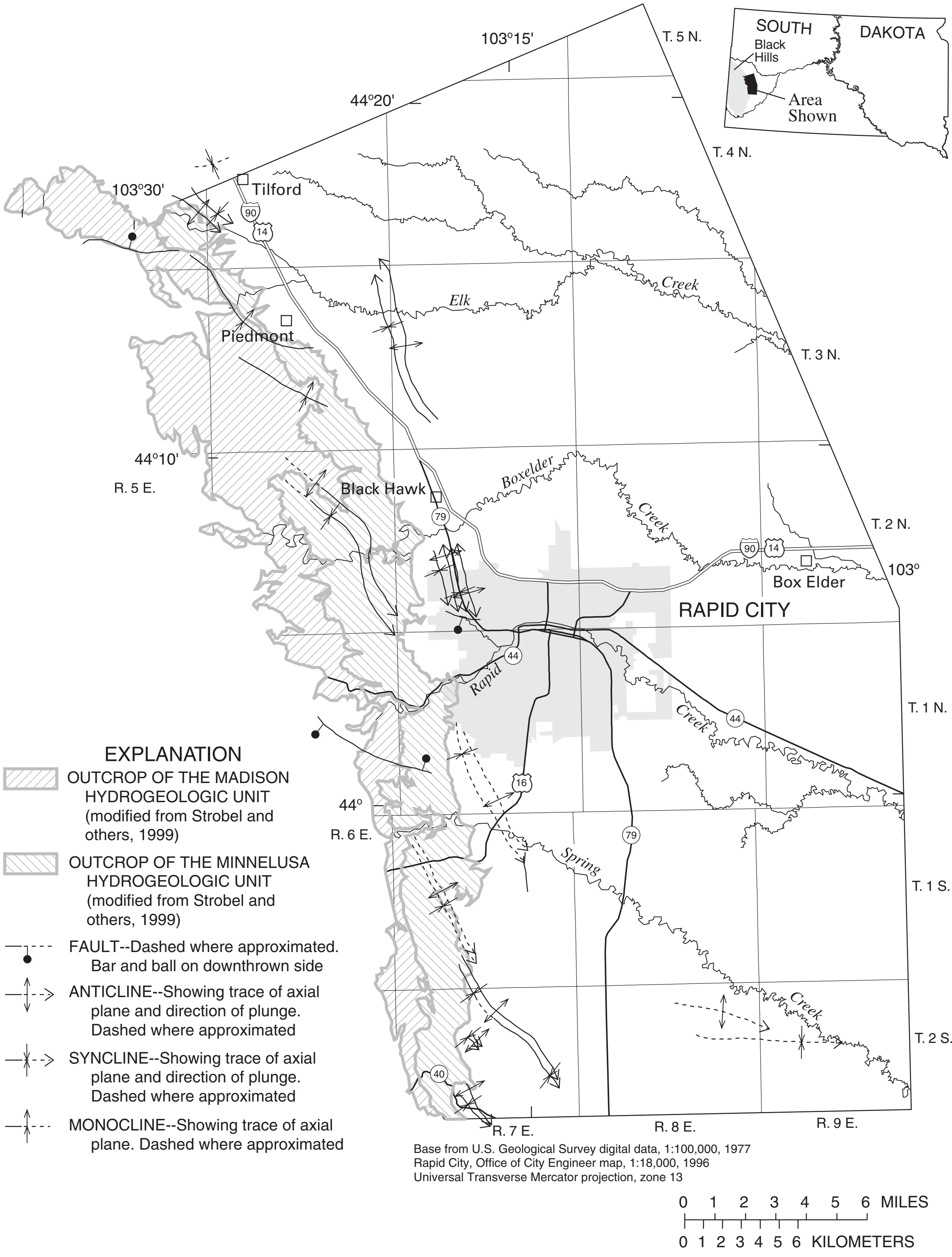


Figure 2. Detail of aquifer analysis area shown in figure 1.

Table 1. Long-term average temperatures in the study area for months of January and July

[Data averaged from 1961 to 1990 (National Oceanic and Atmospheric Administration, 1996)]

	January	July
	(degrees Fahrenheit)	
Pactola Dam[1] (western highlands)	21.2	64.4
Rapid City Airport[1] (eastern lowlands)	22.1	72.2
Difference	.9	7.8

[1]Shown in figure 1.

Previous Investigations

Greene (1993) analyzed aquifer tests and geophysical well logs to determine aquifer properties for the Madison and Minnelusa aquifers in the Rapid City area. Greene and Rahn (1995) presented evidence based on cave-passageway orientations, dye tests, aquifer tests, and well-bore geophysics that the directional orientation of anisotropic transmissivity is localized. Long and Derickson (1999) analyzed hydraulic response to recharge in the Madison aquifer using a linear-systems approach. Long (2000) modeled flow in the Madison aquifer in the Rapid City area by incorporating localized anisotropy. Carter, Driscoll, Hamade, and Jarrell (2001) presented a water budget for the Madison and Minnelusa aquifers for the Black Hills area.

The Madison aquifer has been characterized as both a dual-porosity aquifer (Greene and others, 1998) and a leaky aquifer (Greene, 1993). In the northern Black Hills, Greene and others (1998) adapted a method similar to the Theis (1935) method to determine aquifer properties for dual porosity from an aquifer test. In the eastern Black Hills, Greene (1993) used the method of Neuman and Witherspoon (1969a, 1969b) to determine aquifer properties from an aquifer test for a leaky, two-aquifer system. Long (2000) used numerical modeling to show that dual porosity and leakage might be simultaneously affecting the hydraulics. Rahn (1992) characterized the permeability of the Madison aquifer and summarized published data related to the Madison aquifer in the Black Hills area.

Studies of Madison and Minnelusa springs include Rahn and Gries (1973), Klemp (1995), Wenker (1997), and Anderson and others (1999). Dye-tracer tests in the Madison aquifer include Rahn (1971), Rahn and Gries (1973), and Greene (1999). Studies of Madison Limestone cave development in the Black Hills include Howard (1964) and Ford (1989). Geochemical studies on the Madison and Minnelusa aquifers include Busby and others (1991, 1995) and Naus and others (2001). Regional studies include Downey (1984, 1986), Cooley and others (1986), Downey and Dinwiddie (1988), Kyllonen and Peter (1987), and Plummer and others (1990).

Acknowledgments

The authors thank the city of Rapid City for providing extensive assistance with data collection associated with this study. The Water Rights Program of the South Dakota Department of the Environment and Natural Resources (DENR) provided monitoring wells and assisted in the collection and compilation of water-level data. The South Dakota DENR also provided water-use data. Other assistance with data collection was provided by Chapel Lane Water Company, Carriage Hills Water User's Association, Hart Ranch, and Black Hills Power and Light Company.

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

The Madison and Minnelusa aquifers are contained within the Madison and Minnelusa hydrogeologic units, which are exposed in the western part of the aquifer analysis area (fig. 2). These units coincide with stratigraphic units shown in figures 3 and 4 by the symbols MDme and IPPm. In the aquifer analysis area, stratigraphic units dip in an easterly direction away from the Precambrian core (fig. 5). Water-table conditions generally exist in outcrop areas; however, the updip parts of the outcrops (western extent) may not have a water table because of their higher altitudes. East of the water-table areas, hydraulic head is above the tops of the units due to their easterly dip causing artesian (confined) conditions to exist (fig. 5).

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
		Tw	WHITE RIVER GROUP	0-600	Light colored clays with sandstone channel fillings and local limestone lenses.
	TERTIARY	Tui	INTRUSIVE IGNEOUS ROCKS	- -	Includes rhyolite, latite, trachyte, and phondite.
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. Black fissile shale with concretions.
			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.
			GRANEROS GROUP — MOWRY SHALE	150-250	Light-gray siliceous shale. Fish scales and thin layers of bentonite.
			GRANEROS GROUP — MUDDY SANDSTONE / NEWCASTLE SANDSTONE	20-60	Brown to light-yellow and white sandstone.
			GRANEROS GROUP — SKULL CREEK SHALE	170-270	Dark-gray to black siliceous shale.
		Kik	INYAN KARA GROUP — FALL RIVER FORMATION	10-200	Massive to slabby sandstone.
			INYAN KARA GROUP — LAKOTA FORMATION	35-700	Coarse gray to buff cross-bedded conglomeratic sandstone, interbedded with buff, red, and gray clay, especially toward top. Local fine-grained limestone.
	JURASSIC	Ju	MORRISON FORMATION	0-220	Green to maroon shale. Thin sandstone.
			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. Glauconitic sandstone; red sandstone near middle.
			GYPSUM SPRING FORMATION	0-45	Red siltstone, gypsum, and limestone.
	TRIASSIC	TRPs	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-crystaline, purplish-gray laminated limestone.
		Po	OPECHE SHALE	50-135	Red shale and sandstone.
	PENNSYLVANIAN	PIPm	MINNELUSA FORMATION	[1,2]375-800	Yellow to red cross-bedded 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.
	DEVONIAN		ENGLEWOOD FORMATION	30-60	Pink to buff limestone. Shale locally at base.
	ORDOVICIAN	Ou	WHITEWOOD (RED RIVER) FORMATION	0-60	Buff dolomite and limestone.
			WINNIPEG FORMATION	0-100	Green shale with siltstone.
	CAMBRIAN	OЄd	DEADWOOD FORMATION	[3]75-400	Massive to thin-bedded buff to purple sandstone. Greenish glauconitic shale, flaggy dolomite, and flatpebble limestone conglomerate. Sandstone, with conglomerate locally at the base.
PRECAMBRIAN		pЄu	UNDIFFERENTIATED METAMORPHIC AND IGNEOUS ROCKS		Schist, slate, quartzite, and arkosic grit. Intruded by diorite, metamorphosed to amphibolite, and by granite and pegmatite.

[1] Modified based on 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 3. Stratigraphic section for the study area.

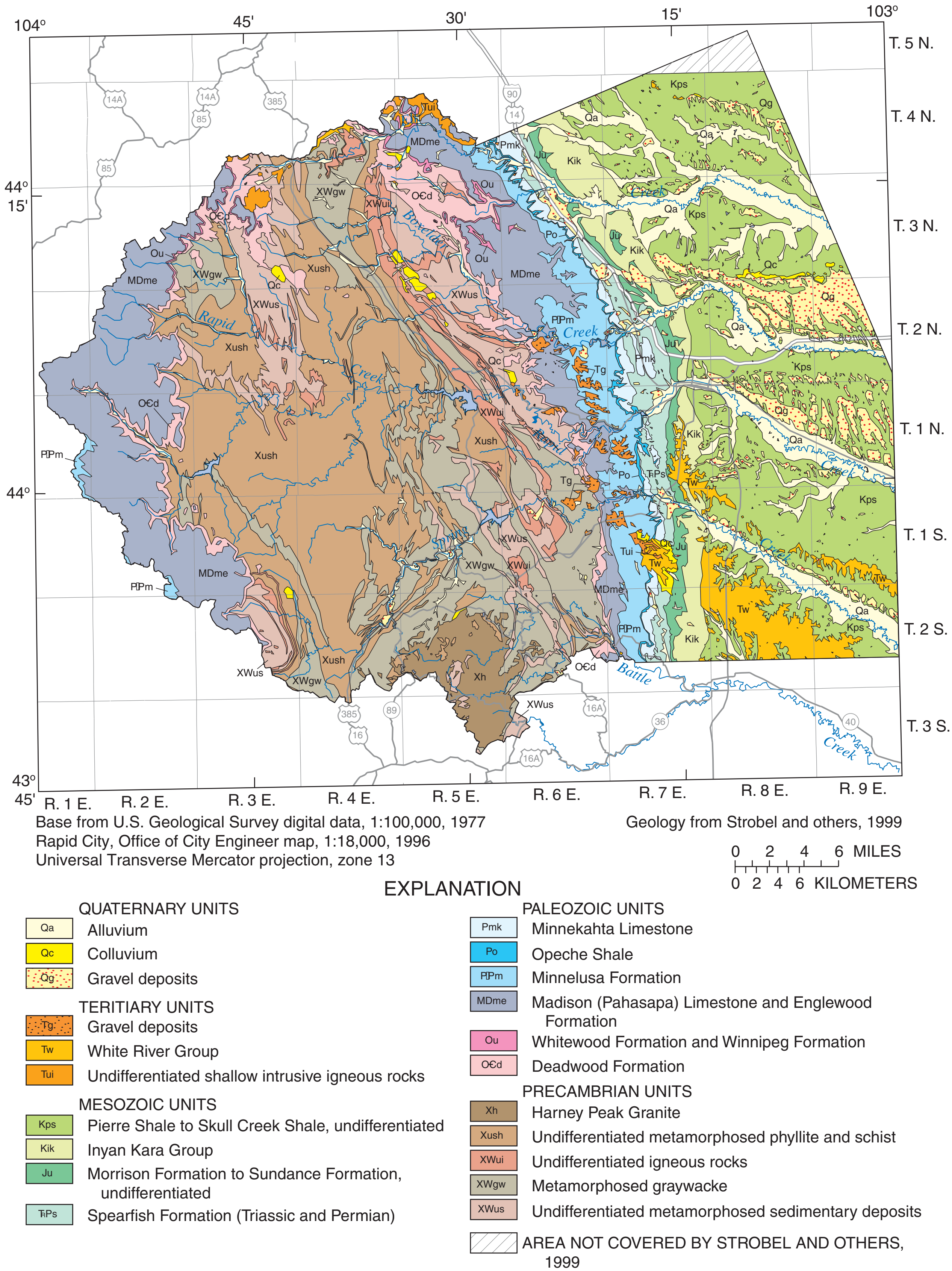


Figure 4. Geologic map of study area.

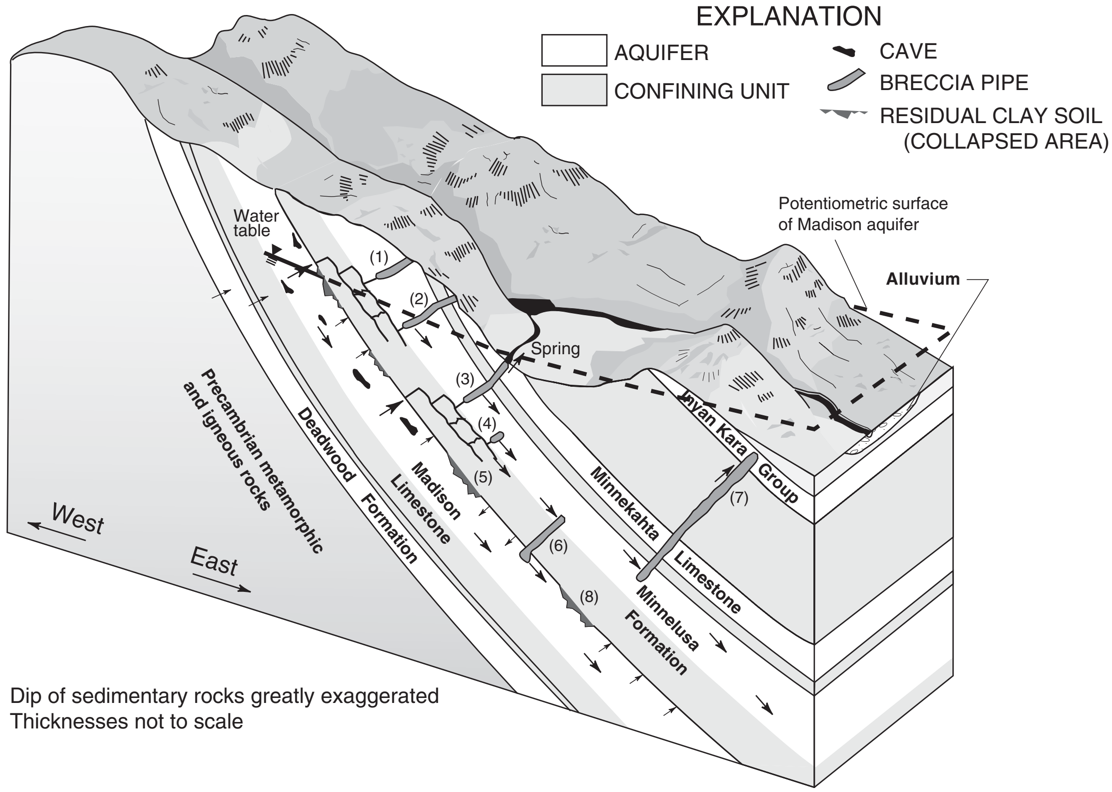


Figure 5. Schematic showing conceptual hydrogeologic section of the study area (modified from Hayes, 1999). Each aquifer shown is separated from other aquifers by confining units. Hydraulic connection between aquifers is increased by vertical breccia pipes and fractures. The schematic shows: (1) exposed 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 ground-water flow directions.

Figure 5 also illustrates potential ground-water flowpaths and features that influence flow in the Madison and Minnelusa aquifers. Breccia pipes, which have the potential to enhance hydraulic connection between aquifers, are discussed in greater detail in the “Vertical Hydraulic Conductivity” section. Hydrogeologic units that may hydraulically affect the Madison and Minnelusa aquifers are composed of rocks of Precambrian age through Early Cretaceous age (Inyan Kara Group) and also Quaternary surficial deposits, which include alluvial aquifers (saturated sand and gravel along streams).

Madison Hydrogeologic Unit

The Mississippian-age Madison Limestone is composed of limestone and dolomite. In the study area to the east of its outcrop, the formation is 250 to 550 ft thick (fig. 3). The Madison hydrogeologic unit is defined in this report as the Madison aquifer and underlying confining unit, which includes the lower Madison Limestone and Englewood Formation. The Madison aquifer is defined as the upper 100 to 200 ft of the Madison hydrogeologic unit where secondary permeability generally is high because of solution openings and fractures (Greene, 1993). The upper surface of the formation is a weathered karst surface, unconformable with the overlying Minnelusa Formation (Cattermole, 1969). The aquifer is considered karstic because of the extensive solution enlargement of fractures that have resulted in a predominance of conduit flow.

Secondary permeability in the lower part of the Madison Limestone (Madison confining unit) generally is much smaller than in the upper part (Greene, 1993); however, the confining unit can have greater permeability near outcrop areas, especially along stream channels. The Englewood Formation, which underlies the Madison Limestone and is less than 60 ft thick, is considered part of the confining unit. The Englewood Formation 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). Strobel and others (1999) combined the Madison Limestone and Englewood Formation as a single hydrogeologic unit. For simplicity, the outcrop of the Madison hydrogeologic unit is referred to as the Madison outcrop in this report.

Wells completed in the Madison aquifer in the study area are capable of producing 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.

Minnelusa Hydrogeologic Unit

In the study area to the east of its outcrop, the Pennsylvanian- and Permian-age Minnelusa Formation is 375 to 800 ft thick (fig. 3). Bowles and Braddock (1963) describe the upper part as thick sandstone with thin limestone, dolomite, and mudstone, and the lower part as having less sandstone and more shale, limestone, and dolomite. Siltstone, gypsum, and anhydrite also can be present. At the base of the Minnelusa Formation is a red clay shale that varies between 0 and 50 ft thick (Cattermole, 1969; Greene, 1993). This shale, which is discontinuous in the aquifer analysis area, is an ancient residual soil developed on the surface of the Madison Limestone (Gries, 1996).

The Minnelusa hydrogeologic unit is defined in this report as the Minnelusa aquifer and underlying confining unit. The Minnelusa aquifer is defined as the upper, more permeable 200 to 300 ft of the Minnelusa Formation because of the coarser sandstone, solution openings, breccias, and other collapse features (Peter and others, 1988; Greene, 1993). The aquifer is confined by the overlying Opeche Shale. The lower part of the formation, which is less permeable and generally impedes flow between the Minnelusa and Madison aquifers (Kyllonen and Peter, 1987; Peter and others, 1988; Greene, 1993) is defined as the Minnelusa confining unit. Near outcrop areas, however, the lower part can have greater permeability due to weathering. For simplicity, the outcrop of the Minnelusa hydrogeologic unit is referred to as the Minnelusa outcrop in this report.

Wells completed in the Minnelusa aquifer in the study area are capable of producing 5 to 700 gal/min. About 66 percent yield from 5 to 50 gal/min, 28 percent yield 50 to 200 gal/min, and 6 percent yield 200 to 700 gal/min. The depth of wells ranges from 80 to 3,000 ft with 90 percent of the wells less than 1,000 ft and 60 percent less than 500 ft.

CONCEPTS OF THE GROUND-WATER-FLOW SYSTEM

Pertinent concepts of the ground-water-flow system include general concepts such as basic hydraulics, recharge, spring discharge, and aquifer interaction. Hydraulic properties described in this section, such as transmissivity, anisotropy, vertical hydraulic conductivity, and storage properties, influence hydraulic head and ground-water flow. Estimated potentiometric surfaces give valuable insight into the spatial distribution of these hydraulic properties. Long-term observation wells provide essential data for the analysis of hydraulic response to stress. The areal extent and location of unconfined areas is important because of the large changes in ground-water storage occurring there. All of the items mentioned above influence ground-water flowpaths, which can be analyzed using natural and artificial tracers.

General Concepts

Figure 6 conceptually illustrates an artesian aquifer with hydrogeology similar to that of the Madison and Minnelusa aquifers. 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 bedding planes dipping easterly. The hydraulic head at 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 are less than at the recharge area unless other stresses such as pumping are introduced.

In a setting such as that shown in figure 6, the unconfined area occurs on the downdip side of the outcrop area. The western updip part of the outcrop not containing a water table is defined as the "unsaturated area" in this report. The unsaturated area may contain infiltrating or perched water, but should not be confused with the space directly above the water table often called the "vadose zone" or "unsaturated zone."

Movement of ground water between the Madison and Minnelusa aquifers is influenced by vertical hydraulic gradients, hydraulic properties of the intervening confining unit, and recharge rates (fig. 7). Recharge water may be stored under perched conditions before percolating downward to the regional water table. Pools of perched water can be found in Madison Limestone caves, and water perched on discontinuous layers of low permeability material may exist in the Minnelusa Formation.

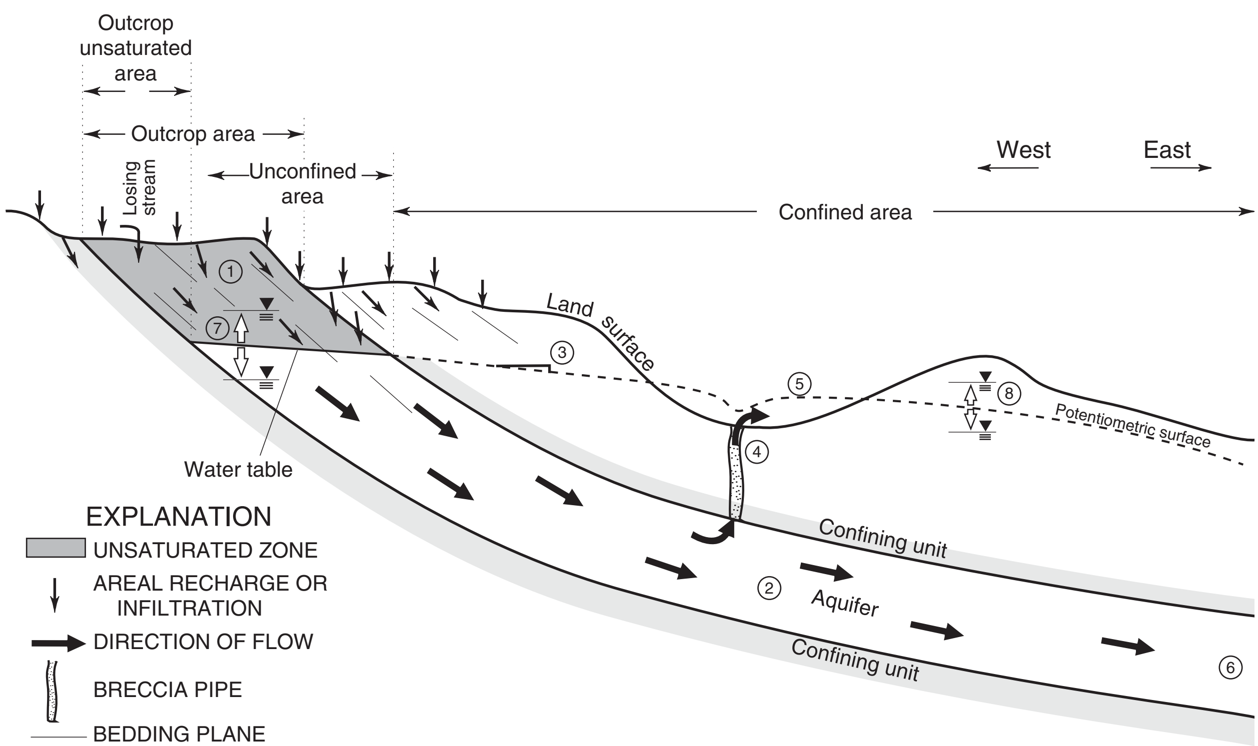


Figure 6. Generalized diagram with vertical exaggeration of an artesian aquifer recharged at the updip end. The diagram shows: (1) recharge infiltrates and moves downward vertically or diagonally parallel to bedding planes; (2) near horizontal flow with head losses resulting from resistance from aquifer 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.

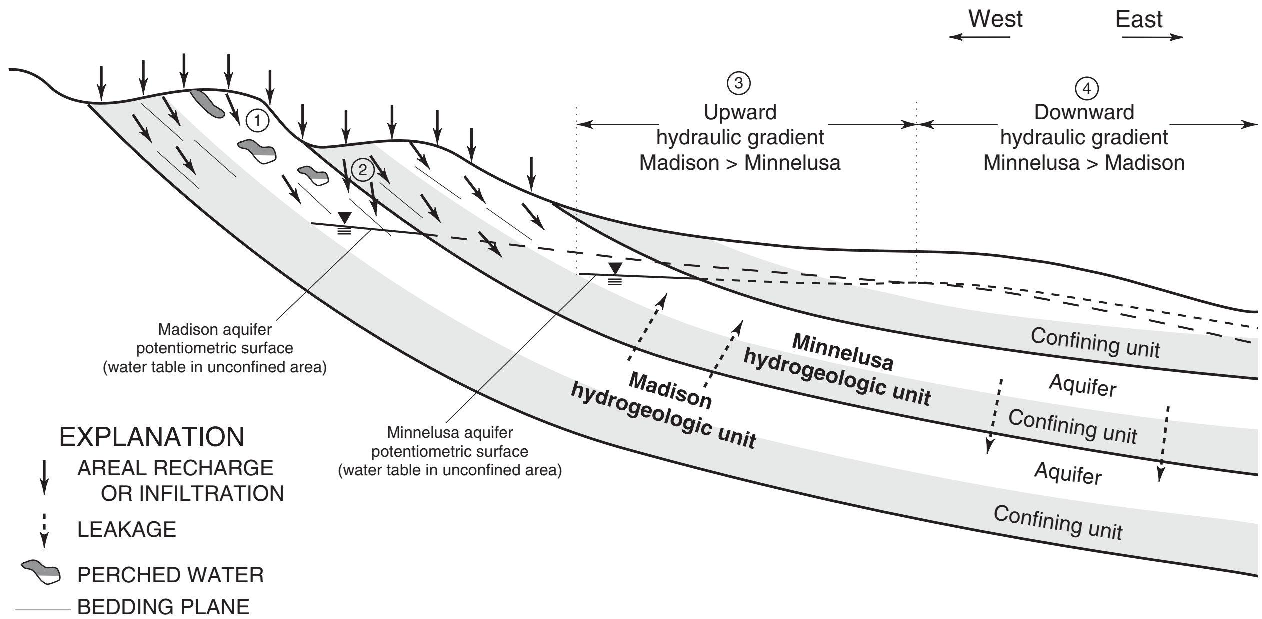


Figure 7. Recharge conditions and vertical gradients in the Madison and Minnelusa hydrogeologic units. The diagram shows: (1) perched water; (2) part of the recharge on the Minnelusa outcrop infiltrates to the Madison aquifer; (3) hydraulic head in Madison aquifer greater than in Minnelusa aquifer creating upward hydraulic gradient; and (4) hydraulic head greater in Minnelusa aquifer than in Madison aquifer creating downward hydraulic gradient.

Although the confining units generally do not transmit water at a high rate, their capacity to store water could have significant effects on the hydraulics of the ground-water-flow system. Water that leaks between the Madison and Minnelusa aquifers must pass through a confining unit as much as 500 ft thick that is composed of material of variable porosity where a substantial amount of water can be held in storage. The confining units also can affect solute transport and hydraulic response to recharge because ground water can move into and out of these layers in response to changes in hydraulic head.

Discharge from artesian springs is an important consideration in the analysis of the Madison and Minnelusa aquifers. There can be difficulty determining whether an artesian spring originates in the Madison aquifer, the Minnelusa aquifer, or both. An example of the hydraulics of ground-water flow from a hypothetical artesian spring includes a vertical conduit of high permeability, such as a breccia pipe, beginning in the Madison aquifer and ending in an alluvial aquifer (fig. 8). In this example, hydraulic head in the Madison aquifer is higher than in the Minnelusa aquifer, and the permeability of the Madison aquifer, breccia pipe, and alluvium is greater than the Minnelusa aquifer. Based on these assumptions, ground water would flow from the Madison aquifer up through the breccia pipe, discharging partly into the Minnelusa aquifer and partly into the alluvium where it may emerge as seepage to the stream. This discharge would create a depression in the Madison aquifer potentiometric surface. Minnelusa aquifer water also may flow into the alluvium, mixing with Madison aquifer water, thus ultimately discharging as a mixture of the two waters.

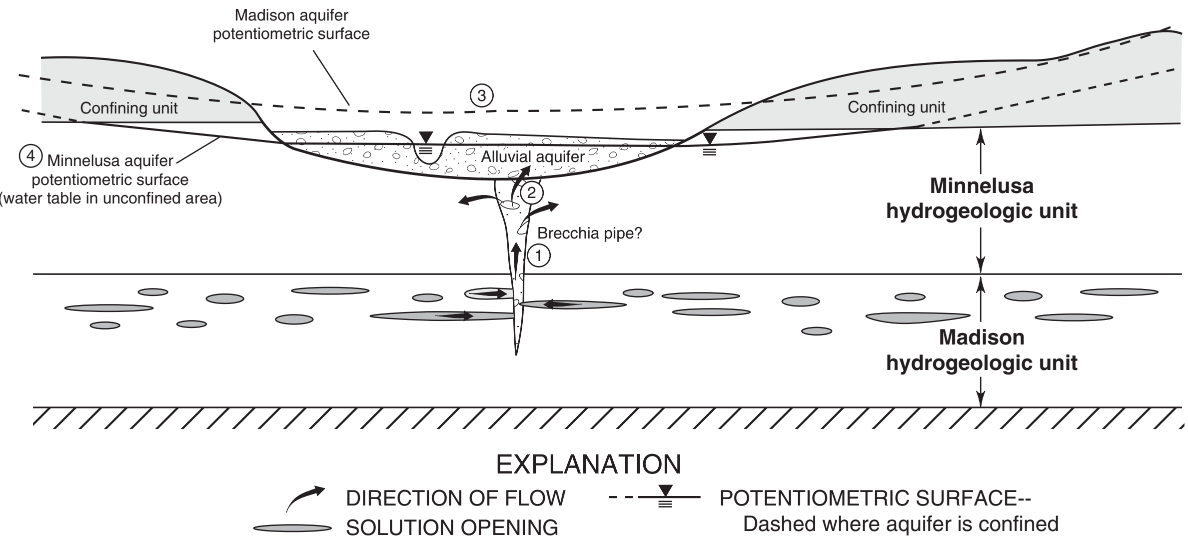


Figure 8. Generalized diagram of an artesian spring. The diagram shows: (1) the Madison aquifer hydraulic head is higher than that of the Minnelusa aquifer causing an upward hydraulic gradient inducing flow through a natural pipe; (2) the spring discharges into the alluvial aquifer; (3) spring discharge causes a depression in the Madison aquifer potentiometric surface surrounding the spring; and (4) Minnelusa aquifer hydraulic head is influenced by the stream and water table in the alluvial aquifer.

Hydraulic Properties

Hydraulic properties, including transmissivity, anisotropic transmissivity, vertical hydraulic conductivity, storage coefficient, and specific yield, were estimated based on (1) previously published work, (2) a water-budget analysis presented later in this report, and (3) an aquifer test at well RC-9 (appendix A). In areas where little or no data were available, aquifer properties were estimated based on a combination of factors. Well locations and average potentiometric surfaces included in this discussion are shown on plates 1 and 2 and are discussed in more detail in the "Hydraulic Head and Ground-Water Flow" section. Preliminary investigations (Long, 2000) also were helpful in estimating aquifer properties, and water-budget analysis served to refine estimates because these properties needed to be adjusted to achieve a balance between inflow and outflow.

Transmissivity

An aquifer test in the Madison aquifer at well RC-9 is described in appendix A, and results are summarized in table 2 along with the results of two previously published aquifer tests in the aquifer analysis area. Locations of pumped wells and observation wells are shown in figures 9 and 10, which also show the estimated transmissivity (*T*) distributions for the Madison and Minnelusa aquifers. These distributions are generally in agreement with table 2; however, anisotropy and heterogeneity could account for variations. For example, an aquifer test measures the directional *T* between a pumped well and observation well. A general or "effective" *T* can be estimated based on multiple observation wells in an aquifer test (see "Anisotropic Transmissivity" section). Estimates of this effective *T* (figs. 9 and 10) are based on aquifer tests, well yields, potentiometric surfaces, water budgets, and other hydrogeologic information. Table 3 lists aquifer properties reported for the Madison and Minnelusa aquifers in the general region but not necessarily within the study area. Additional data compiled for the Black Hills area by Rahn (1992) show a similar range in Madison aquifer transmissivity.

Well production rates in comparison to drawdown and potentiometric surfaces are also indicative of relative *T* values. The production rates and drawdowns shown in table 4 are in general agreement with the effective *T* distributions shown in figures 9 and 10. The low production rate and large drawdown in well RC-7 (fig. 9) indicate that *T* decreases to the east of Rapid City. In addition, a water sample from well RC-7 in 1991 (Feb. 7, 1991) had a specific conductance of 3,490 microsiemens per centimeter at 25 degrees Celsius indicating a high concentration of dissolved solids in contrast to water from many other Madison aquifer wells. The high specific conductance is interpreted to be a result of limited ground-water flow in that area. Inferences in some local areas can be made from the potentiometric surfaces (pls. 1 and 2) by assuming that *T* is generally larger in areas where the hydraulic gradient is small. Potentiometric surfaces for both aquifers have low gradients in the west-central part of the aquifer analysis area. Although this is only a general indication of *T* because recharge rates affect hydraulic gradients, aquifer tests also indicate high *T* values in that area.

A zone of relatively large *T* for the Madison aquifer in the eastern part of the area (fig. 9) is based in part on a trough in the potentiometric surface at that location. This type of potentiometric feature would most likely be related to highly transmissive rocks. The production rate of 350 gal/min and relatively low dissolved solids concentration (Vince Finkhouse, City of Boxelder Public Works, oral commun., 2001) from a Madison aquifer production well in the city of Boxelder (pl. 2, site 30) is consistent with large *T* within this zone. In addition, a regional analysis of the Madison aquifer by Downey (1986, p. E54) shows a similar spatial *T* distribution. The lower *T* values in the northeast and southeast parts of the area for the Madison aquifer were estimated by balancing the water budget (see “Water Budget” section) and are consistent with those estimated by Downey (1986, p. E54).

Table 2. Hydraulic properties reported for the Madison and Minnelusa hydrogeologic units determined from aquifer tests in the aquifer analysis area

[*T*, transmissivity; K_v, vertical hydraulic conductivity; *S*, storage coefficient; ft, feet; ft^2/d, feet squared per day; ft/d, feet per day; NA, not applicable; --, no data available]

Pumped well and date of test	Layer	Observation well and site number (pls. 1 and 2)	Distance from pumped well (ft)	*T* (ft^2/d)	K_v of the Minnelusa confining unit (ft/d)	*S* (dimensionless)	Source
RC-5[1] Spring 1990 (site 79)	Madison aquifer	LC (43)	685	1,600	6.8×10^{-3}	1.0×10^{-4}	Greene (1993)
		SP-2 (46)	1,700	2,600	1.6×10^{-2}	1.0×10^{-4}	
		BHPL (36)	3,950	5,200	1.1×10^{-2}	1.0×10^{-4}	
		CL-2 (50)	8,900	40,000	9.1×10^{-3}	3.0×10^{-4}	
		CHLN-2 (56)	11,700	40,000	5.3×10^{-3}	3.0×10^{-4}	
RC-6 Spring 1990 (site 35)	Minnelusa aquifer	CQ-1 (200)	2,930	12,000	NA	3.0×10^{-3}	Greene (1993)
	Minnelusa confining unit	CQ-1 (200) CQ-2 (33)	2,930 2,919	--	3.0×10^{-1}	2×10^{-7}	
	Madison aquifer	CQ-2 (33)	2,919	17,000	NA	2×10^{-3}	
RC-9 Fall 1995 (site 49)	Madison aquifer	CL-2 (50)	6,290	14,700	1.7	2.1×10^{-5}	Appendix A
		RC-11 (62)	8,603	11,500	2.7	2.7×10^{-4}	
		CHLN-2 (56)	7,562	14,100	0.9	1.4×10^{-5}	
		SP-2 (46)	4,564	13,900	0.3	6.8×10^{-5}	

[1]Anisotropic transmissivity of 56,000 ft^2/d was determined with the major axis at an angle of 42 degrees east of north. The minor axis of transmissivity was 1,300 ft^2/d at an angle of 48 degrees west of north.

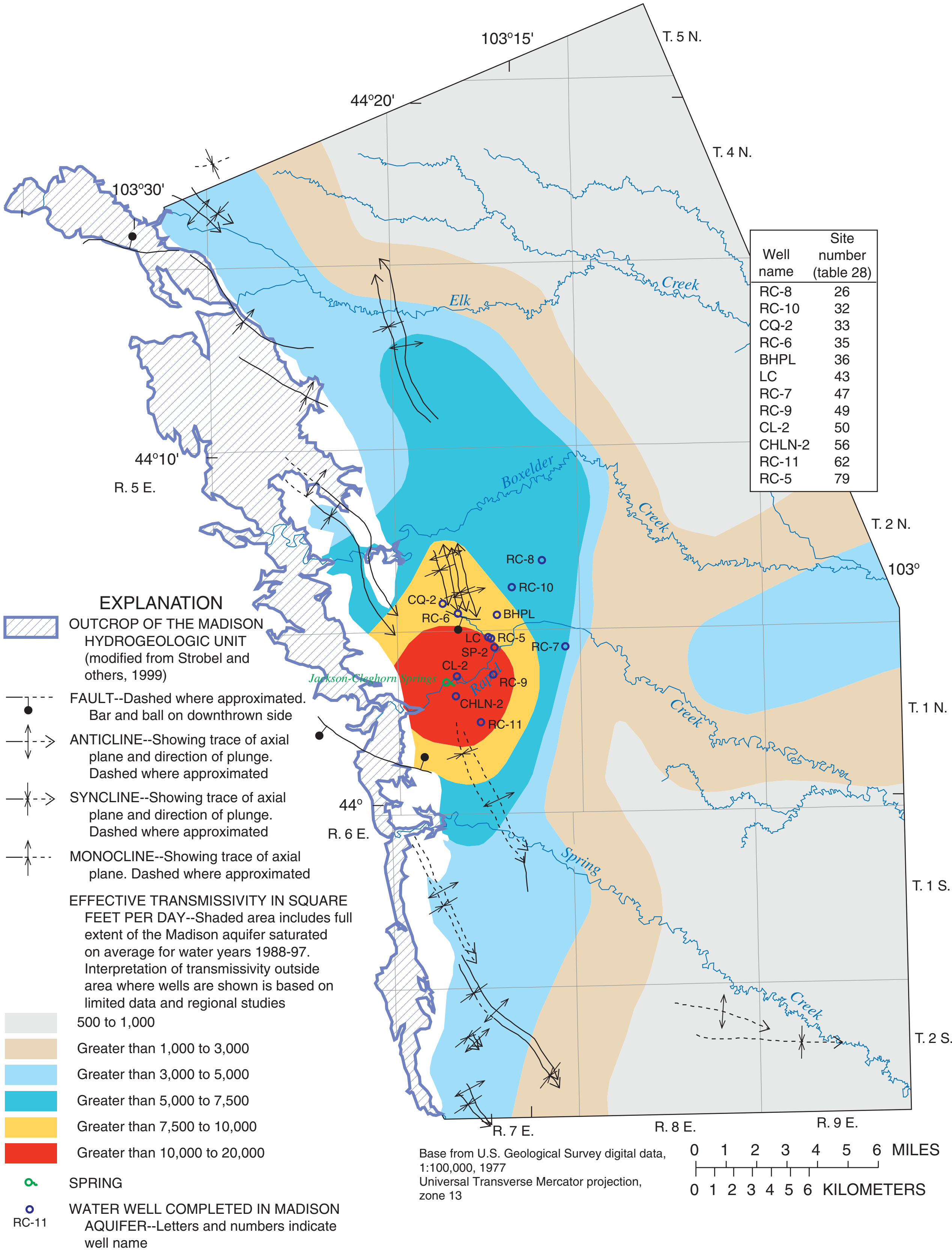


Figure 9. Effective transmissivity distribution estimates for the Madison aquifer. Transmissivity estimates are provided for areas where the Madison aquifer exists as shown on plate 1 and do not extend into areas where only the Madison confining unit exists.

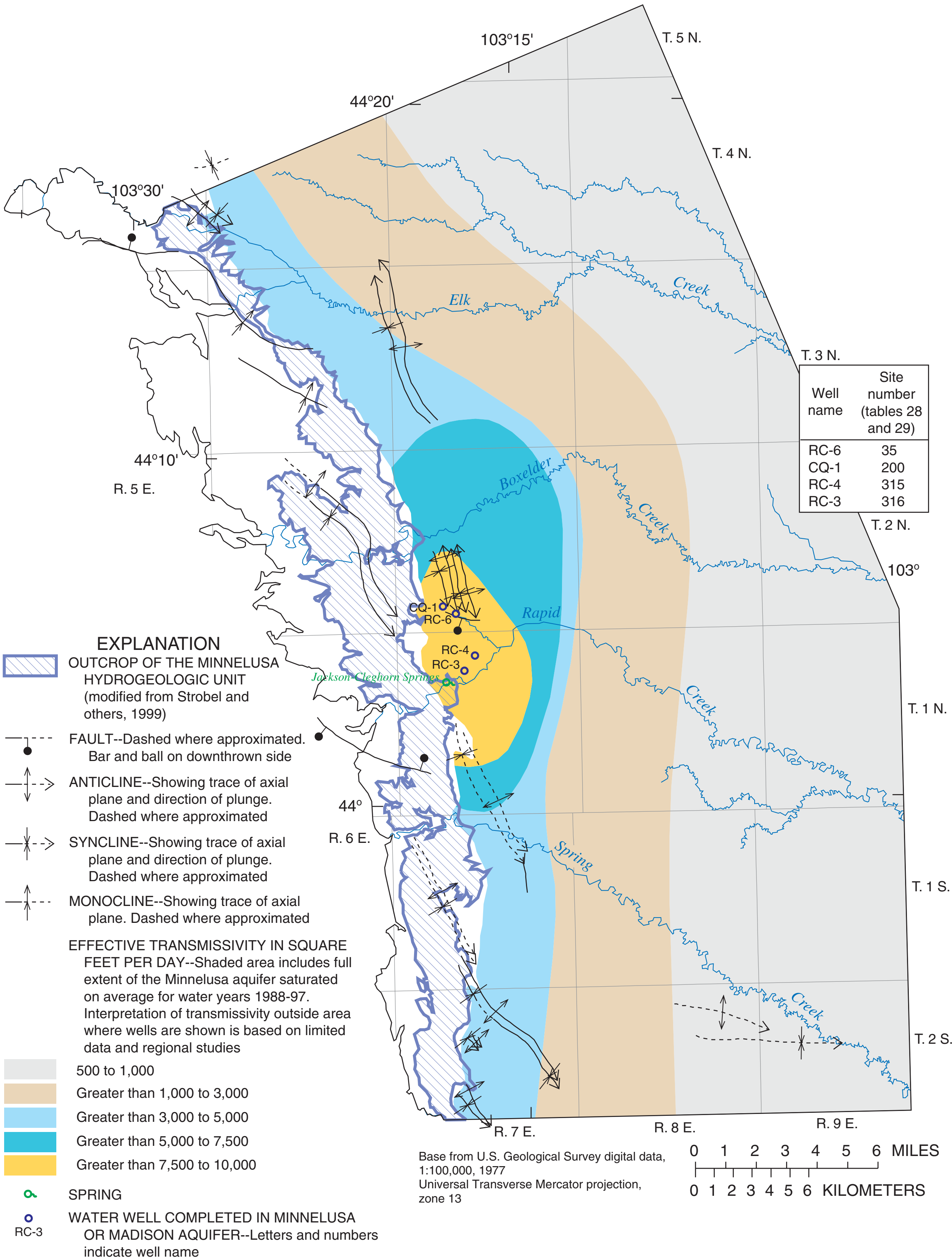


Figure 10. Effective transmissivity distribution estimates for the Minnelusa aquifer. Transmissivity estimates are provided for areas where the Minnelusa aquifer exists as shown on plate 2 and do not extend into areas where only the Minnelusa confining unit exists.

Table 3. Estimates of regional values of hydraulic conductivity, transmissivity, vertical hydraulic conductivity, and storage coefficient

[Modified from Kyllonen and Peter, 1987, p. 20-21. ft/d, feet per day; ft^2/d, feet squared per day; --, no data available]

Original source	Hydraulic conductivity (ft/d)	Transmis-sivity (ft^2/d)	Vertical hydraulic conductivity (ft/d)	Storage coefficient (dimensionless)	Data source or method
Madison Aquifer					
Konikow (1976, p. 41)	--	860 - 2,200	--	--	Flow net analysis and model, includes correction for temperature variation.
Miller (1976, p. 25)	--	0.01 - 5,400	--	--	Drill-stem tests in southeastern Montana.
Blankennagel and others (1977, p. 52-53)	2.4×10^{-5} - 1.9	--	--	--	Permeability test core.
Woodward-Clyde Consultants (1980, p. 4-13)	--	3,000	--	2×10^{-4} - 3×10^{-4}	Aquifer test, long-term response of aquifer to pumping in western Black Hills region, and model.
Blankennagel and others (1981, p. 50)	--	5,090	--	2×10^{-5}	Step-drawdown tests.
Downey (1984, p. 45)	--	250 - 1,500	--	--	The range given is for Black Hills part of Downey's model.
Kyllonen and Peter (1987, p. 21)	--	4.3 - 8,600	--	--	Model calibrated values.
Downey (1986, p. E54)	--	less than 250 to 3,000	--	--	Model calibrated values.
Carter, Driscoll, Hamade, and Jarrel (2001)	--	100 - 7,400	--	--	Water-budget analysis.
Greene and others (1998)	--	41,700		3×10^{-4}	Interference test.
Minnelusa Aquifer					
Blankennagel and others (1977, p. 50)	less than 2.4×10^{-5} to 1.4	--	--	--	Permeability test of core.
Pakkong (1979, p. 41)	--	880	--	--	Aquifer test.
Woodward-Clyde Consultants (1980, p. 4-12)	--	30 - 300	--	6.6×10^{-5} - 2.0×10^{-4}	Aquifer test, flow and specific capacity data, permeability data, and lithologic considerations.
J.S. Downey, U.S. Geological Survey, Denver, Colo., written commun., 1982)	--	700 - 1,000	--	--	Model calibrated values.
Kyllonen and Peter (1987, p. 21)	--	0.86 - 8,600	--	--	Model calibrated values.
Downey (1986, p. E55)	--	less than 250 to 1,000	--	--	Model calibrated values.
Greene and others (1998)		9,600		7×10^{-5}	Interference test.
Minnelusa Confining Unit					
Blankennagel and others (1977, p. 50-51)	--	--	less than 2.4×10^{-5} to 0.01	--	Permeability test of core.
Downey (1982, p. 74)	--	--	5.0×10^{-7} - 7.0×10^{-7}	--	Model calibrated values.
Kyllonen and Peter (1987, p. 21)	--	--	3.4×10^{-6} - 3.4×10^{-4}	--	Model calibrated values.

Table 4. Selected information for Rapid City production wells

[From Anderson and others, 1999. --, no data available]

Well	Site number (pls. 1 and 2)	Year drilled	Major aquifer	Depth of hole (feet below land surface)	Approximate static water level (feet below or above (-) land surface)	Approximate pumping water level (feet below land surface)	Drawdown (pumping level minus static level)	Approximate well yield (gallons per minute)
RC-1	317	1935	Minnelusa[1]	1,460	-32	--	--	640
RC-3	316	1936	Minnelusa[2]	957	30	--	--	670
RC-4	315	1939	Minnelusa	1,070	-5	--	--	700
RC-5	79	1989	Madison	1,292	-102	210	312	1,700
RC-6	35	1990	Madison[3]	1,300	8	426	418	770
RC-7	47	1991	Madison	3,280	250	773	523	150
RC-8	26	1991	Madison	2,680	125	440	315	545
RC-9	49	1991	Madison	1,050	-85	--	--	2,580
RC-10	32	1991	Madison	1,790	-73	277	350	1,790
RC-11	62	1991	Madison	1,280	64	374	310	820

[1]Also may produce from Madison and Deadwood aquifers.
[2]Also may produce from Madison aquifer.
[3]Also may produce from Minnelusa aquifer.

Transmissivity in the Madison aquifer is primarily influenced by fractures and solution openings. Geophysical well logs indicate that in the Rapid City area, the relative volume of solution openings is generally largest near the outcrop areas (Greene, 1993), which is a result of fracturing caused by the Black Hills uplift. The potential for enlargement of fracture openings by dissolution is greatest near recharge areas where carbon dioxide is readily available and the dissolved solids concentration is low. Carbon dioxide in recharge water, which increases in concentration within the soil zone, combines with water to form carbonic acid causing the dissolution of calcite. As carbonate rock is dissolved and water becomes more saturated along the flowpath, dissolution potential decreases and secondary porosity is less developed.

Fracturing was assumed to coincide with areas where there is curvature in strata indicated by a change of dip. Areas where changes in dip for the Madison and Minnelusa hydrogeologic units are greater than 2 percent where determined from contour maps of altitude of tops (Carter and Redden, 1999a,1999b) and are concentrated near the outcrops (figs. 11 and 12). Curvature also could be the result of an erosional surface of the top of the Madison Limestone; however, formation thickness in the area showing curvature is relatively uniform. Areas where change in dip is greater than 2 percent generally extend about 3 mi to the east of the outcrop and about 5 mi in the area east of the Boxelder and Rapid Creek streamflow-loss zones. This wider area could indicate more extensive tectonic activity and help explain the higher T values. Because of this tectonic activity and other reasons to be discussed later, this area is referred to as the high-flow area. Near the center of the high-flow area is a syncline-anticline set (a, figs. 11 and 12) with a fault at the southern end. Jackson-Cleghorn Springs are also located in the central part of the high-flow area, and another syncline-anticline set (b, figs. 11 and 12) crosses the Boxelder Creek loss zone and extends to the southeast toward Jackson-Cleghorn Springs. A third syncline-anticline set (c, figs. 11 and 12) is located south of Jackson-Cleghorn Springs extending toward Spring Creek.

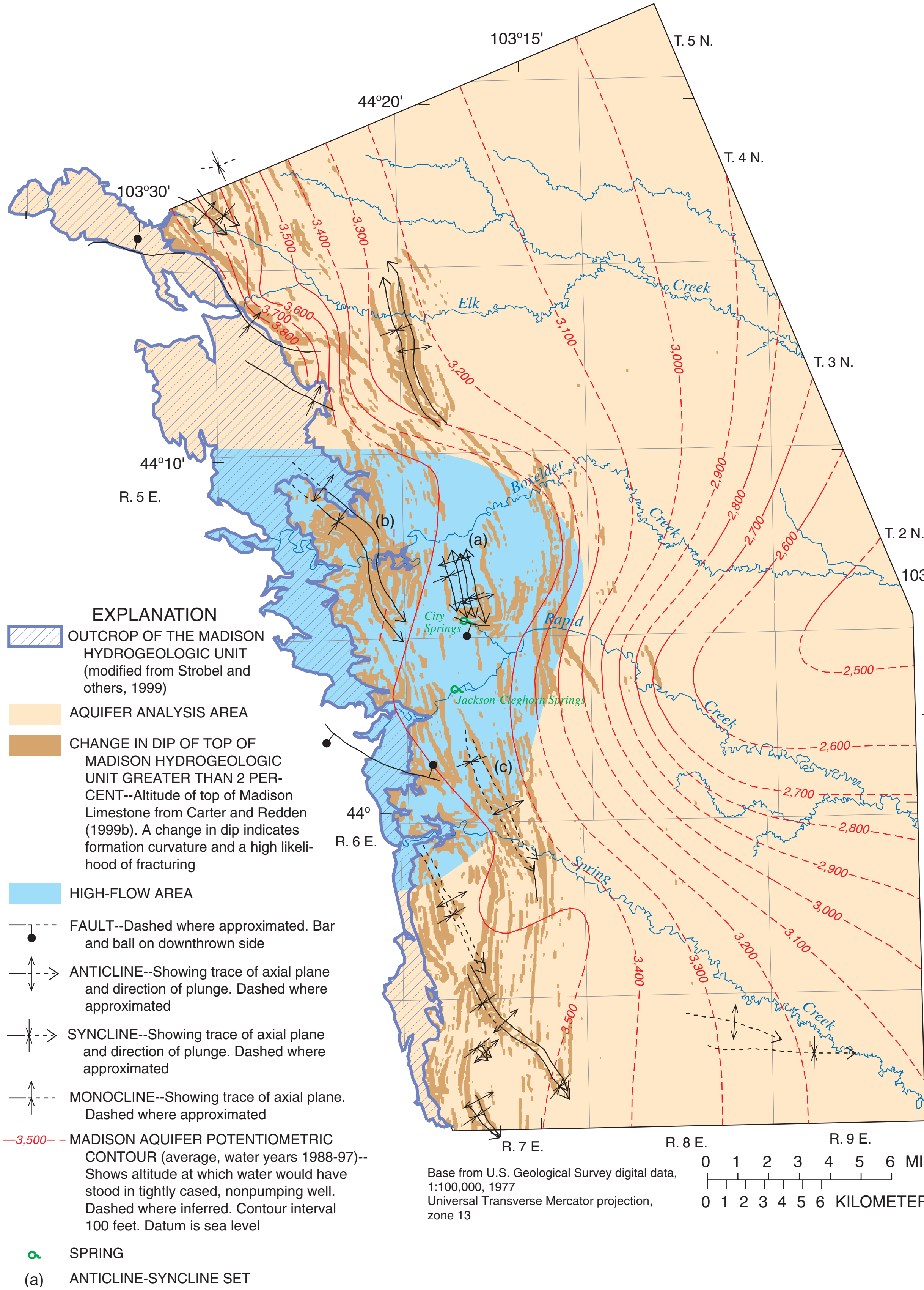


Figure 11. Hydrogeologic features indicating transmissivity distribution in the Madison aquifer. A change in dip indicates formation curvature and a likelihood of fracturing.

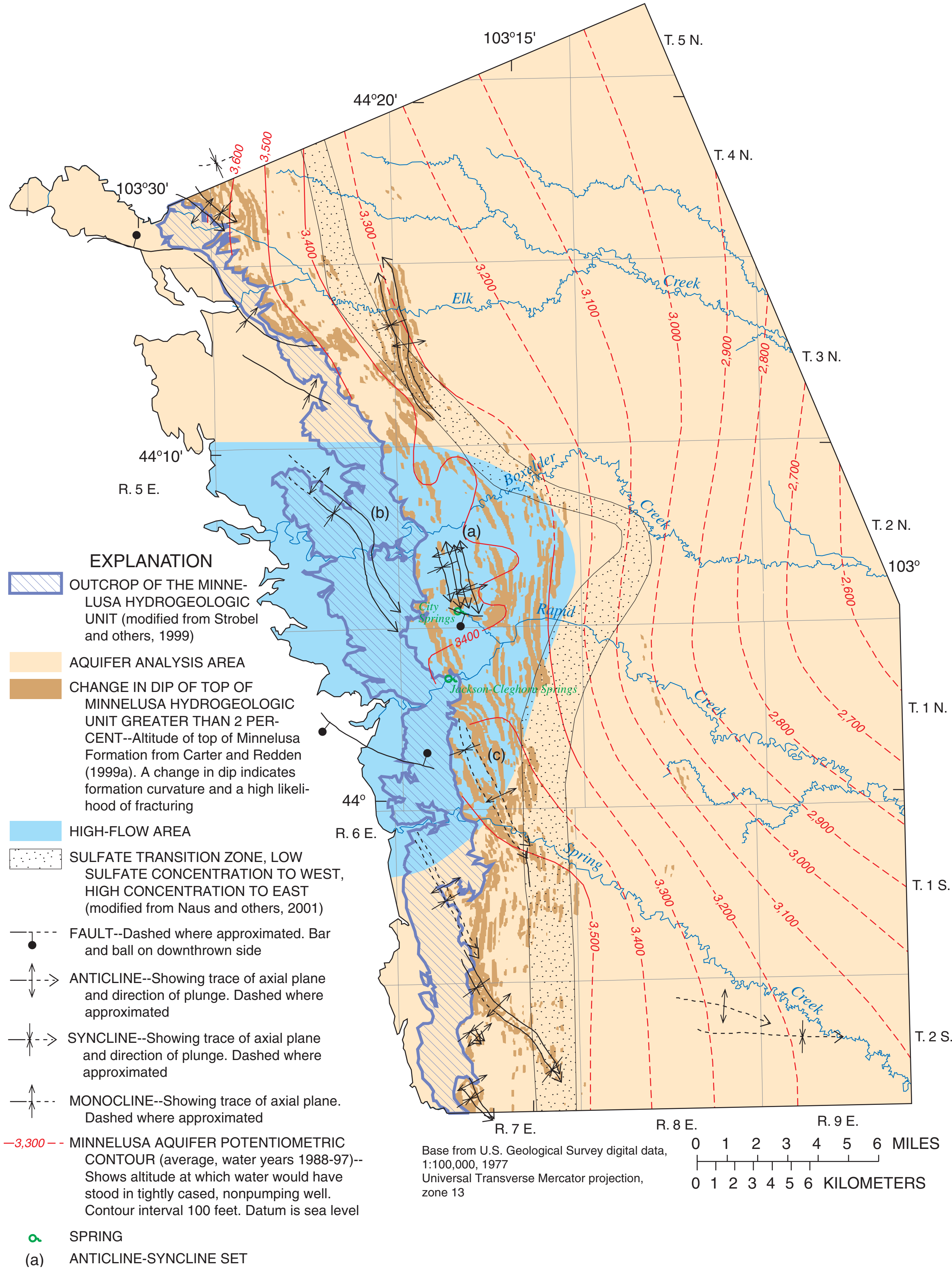


Figure 12. Hydrogeologic features indicating transmissivity distribution in the Minnelusa aquifer. A change in dip indicates formation curvature and a likelihood of fracturing.

Estimated T values greater than 7,500 ft^2/d in the high-flow area (figs. 11 and 12) are consistent with aquifer test results in the Madison aquifer (table 2). Also, aggressive dissolution of carbonate rocks in the Jackson-Cleghorn Springs area is likely to have occurred because of the convergence of large volumes of recharge water moving toward the springs, which probably flow primarily from the Madison aquifer (Rahn and Gries, 1973; Back and others, 1983; Anderson and others, 1999). Karst solution enlargement of fractures often is initiated at a spring and proceeds radially outward from the spring (Clemens and others, 1997), which can create large T values in areas surrounding a spring. Jackson-Cleghorn Springs are likely to be similar to the generalized diagram of an artesian spring shown in figure 8.

A large part of the transmissivity of the Minnelusa aquifer (fig. 10) is due to the primary porosity of the sandstone layers; secondary porosity results from brecciation and from fracturing due to faulting, folding, and separation of bedding planes. Dissolution of interbedded carbonate rock layers and carbonate cements also can increase secondary porosity. Although there are very little T data for the Minnelusa aquifer in the study area, a similarity to the spatial distribution of T in the Madison aquifer is likely because the same tectonic forces have lead to increased secondary porosity in both aquifers.

A process that probably has increased T in the Minnelusa aquifer in the near-outcrop area is the collapse resulting from the dissolution of gypsum and anhydrite, which are calcium sulfate minerals. Greater dissolution is more likely to occur closer to the outcrops because of smaller concentrations of dissolved solids in recharge waters. Brobst and Epstein (1963) discuss the removal of up to 200 to 300 ft of gypsum and anhydrite layers in the southwestern Black Hills due to this process, which could result in significant collapse and fracturing. The downdip advancement of evaporite removal beginning near the outcrop was documented by Naus and others (2001) by an analysis of sulfate concentrations in Minnelusa aquifer water. The zone of transition for sulfate concentrations (fig. 12) separates concentrations less than 250 mg/L (milligrams per liter) to the west, which indicates removal of gypsum and anhydrite, from concentrations greater than 1,000 mg/L to the east. The transition zone generally ranges from 2 to 3 mi downdip from the Minnelusa outcrop and about 7 mi downdip from the outcrop congruent with the high-flow area (fig. 12). The wider area probably results from increased ground-water circulation due to a combination of factors within the high-flow area.

Anisotropic Transmissivity

In well-developed karst aquifers, solution-enhanced fractures and bedding planes provide secondary porosity that dominates ground-water flow and results in anisotropic transmissivity. Anisotropic transmissivity can be represented by an ellipse with perpendicular axes that define the magnitude and direction of the maximum and minimum transmissivity tensors (T_{max} and T_{min}) (Freeze and Cherry, 1979). Effective transmissivity (T_e), as defined by Hantush (1966a), is $\sqrt{(T_{max} \times T_{min})}$, or the geometric mean of T_{max} and T_{min}. Greene (1993) reported a ratio of T_{max} to T_{min} of 45:1 from an aquifer test at RC-5 (site 79, table 3) in the Madison aquifer.

The distribution of T shown in figures 9 and 10 refer to effective transmissivity. For the aquifer test at RC-5 (table 2), the large discrepancy in T calculated for each of the five observation wells was interpreted as resulting from anisotropy with the major axis of transmissivity (T_{max}) trending northeast-southwest (Greene, 1993). To determine this anisotropy, Greene (1993) applied the method of Hantush (1966a, 1966b), which assumes the aquifer is homogeneous and anisotropic. However, the results of the aquifer test at RC-5 also could be interpreted as resulting, at least in part, from heterogeneity. In addition to this, Greene and Rahn (1995) and Long (2000) indicated that anisotropy in the Madison could be localized in its horizontal orientation.

Anisotropic transmissivity not only can affect flow direction but also hydraulic head and hydraulic gradients as well. Long (2000) used a numerical model to show how the potentiometric surface changes when comparing isotropic to anisotropic conditions. A ratio of 40:1 can cause flow direction to deflect as much as 70 degrees from perpendicular to equipotential lines (Long, 2000). There is much evidence in support of the presence of anisotropy in the Madison aquifer but very little data to approximate its spatial distribution.

Some formations have been found to contain two mutually perpendicular fracture sets. McQuillan (1973) described the fractured-limestone Asmari Formation of southwestern Iran, which has such fracture sets. Stearns and Friedman (1972) described a major class of fracture systems called “regional orthogonal fractures.”

Also, according to Price (1959), uplift can create fractures in one orientation, and further uplift can create a new set of fractures orthogonal to the first.

Greene and Rahn (1995) determined that anisotropic transmissivity in the Madison aquifer near the Black Hills had predominantly one of two mutually perpendicular orientations, depending on the local area, and called this “localized anisotropic transmissivity.” The orientation was found to be mainly either northeast or northwest. Those authors presented evidence based on analysis of cave orientations, fracture traces, aquifer tests, tracer tests, and geophysical methods. Within the study area for this report, seven out of the eight diagrams of the orientations of cave passageways from Greene and Rahn (1995) generally display a predominant orientation that ranges from about 40 to 70 degrees east of north. Interpretation of an aquifer test at RC-5 resulted in an anisotropic orientation of 42 degrees east of north (Greene, 1993). Interpretation of bedding plane solution openings from an acoustic televiewer log at RC-6 showed a predominant orientation approximately perpendicular to anisotropy determined from the RC-5 aquifer test (Greene and Rahn, 1995). Synclines and anticlines in the aquifer analysis area (fig. 2) are generally oriented about 20 to 25 degrees west of north. Downey (1984) inferred a lineament pattern from satellite imagery in the northern Great Plains, where in the Black Hills area, the predominant lineament pattern appears to be either 45 degrees west of north or 45 degrees east of north. Using localized anisotropy in these two general orientations was effective in calibrating a numerical model of the Madison aquifer for the Rapid City area (Long, 2000).

This localized anisotropy may have occurred when one set of orthogonal fractures was selectively enhanced by dissolution in comparison to the other fracture set. Many researchers have studied the conditions that influence conduit network patterns (Howard and Groves, 1995; Clemens and others, 1997; Kaufman and Braun, 1999, 2000; Gabrovsek and Dreybrodt, 2000). Tectonic movements can contribute to the hydraulic conditions that allow fracture permeability to be enhanced in a particular direction in a particular area. Local hydraulic gradients can cause the enhancement of one set of fractures over another because the fracture set that is aligned most closely with the ground-water flow direction will capture the largest flows and become preferentially enlarged. As the increasingly dominant fracture set captures progressively more of the flow, the gradient and the direction of the resultant flow vector also would change. Another factor that may have influenced the present conduit network is development of cave and solution openings prior to the Black Hills uplift. Because the Madison Limestone was being eroded at the land surface after the Mississippian sea retreated (Gries, 1996), the surface topography and drainage would have been a factor in the development of these paleo-cave networks.

Vertical Hydraulic Conductivity

Vertical hydraulic conductivities (K_v) of the Minnelusa confining unit estimated from aquifer tests in the aquifer analysis area range from 5.3×10^{-3} to 2.7 ft/d (table 2), which indicates potential for large spatial variability in leakage between the Madison and Minnelusa aquifers. These K_v values are notably larger than regional values reported by various investigators (table 3). Because the aquifer tests reported in table 2 represent only a small sample of the aquifer analysis area, K_v values calculated do not necessarily represent the entire area. Indeed, these K_v values could be some of the largest in the study area because the aquifer tests were conducted in the high-flow area (figs. 11 and 12) where extensive structural deformation has occurred as previously discussed in the “Transmissivity” section. However, because of the variable distribution of structural features, this high-flow area probably has some of the largest variability of K_v values in the study area as well.

Part of the variability of K_v in the Minnelusa confining unit determined from aquifer tests (table 2) could be due to the selection of the analytical model and associated assumptions. Three different models were used for the RC-5, RC-6, and RC-9 aquifer tests, which included Hantush and Jacob (1955), Neuman and Witherspoon (1969a), and Hantush (1960), respectively. An important difference in the assumptions made by these models relates to the storage properties of the confining units. The models selected for RC-6 and RC-9 take into account storage in the confining units, whereas the model used for RC-5 assumes no storage in the confining units. The lower K_v values determined for RC-5 as compared to RC-6 and RC-9 (table 2) could be partly due to the different assumptions related to storage in confining units.

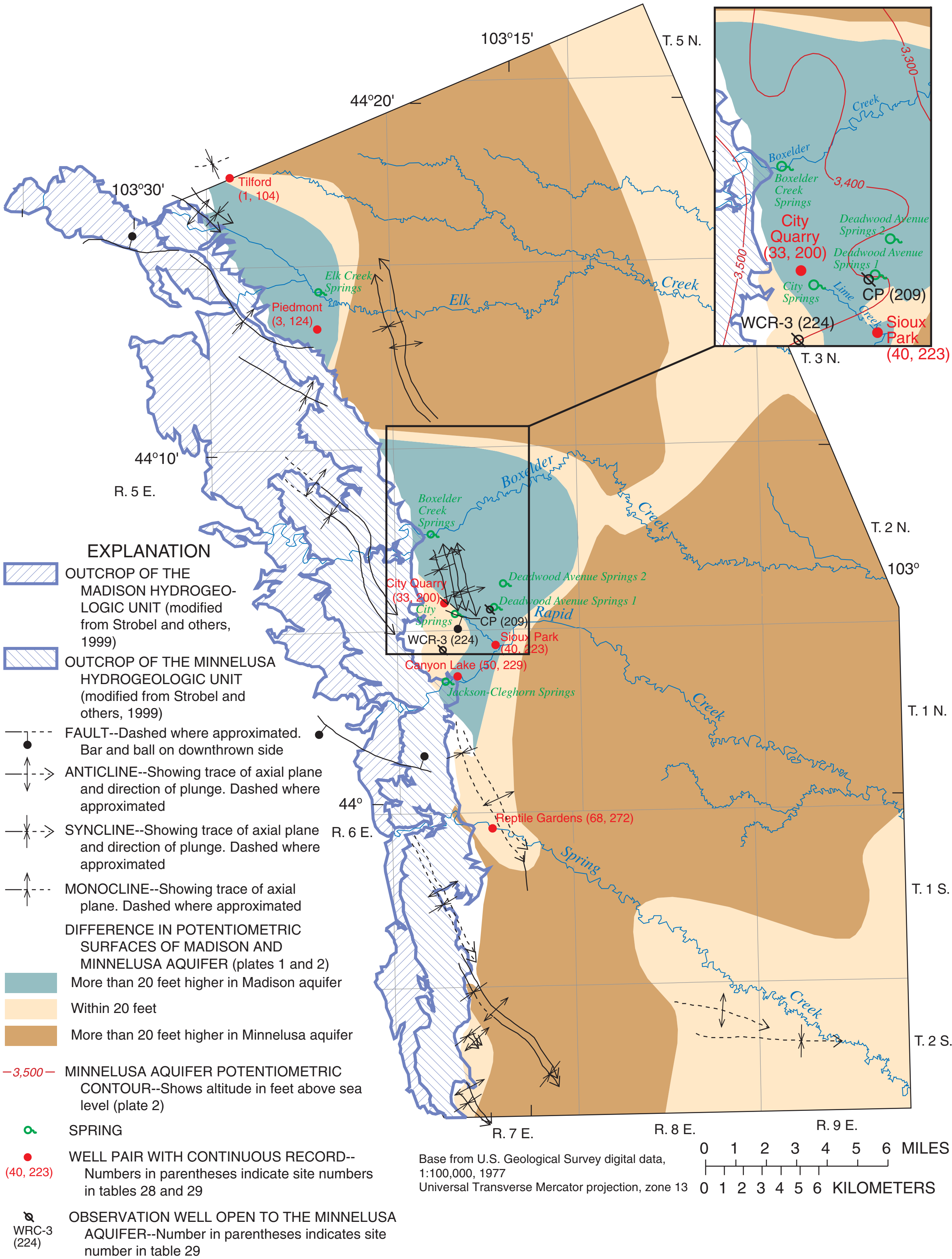


Figure 13. Comparison of hydraulic head in Madison and Minnelusa aquifers.

The uneven distribution of a residual clay soil on top of the Madison Limestone after weathering and reworking during the Mississippian and Pennsylvanian Periods (Gries, 1996) could cause heterogeneities in hydraulic connection. Greater leakage rates might occur where the residual clay is not present. Cattermole (1969) and Greene (1993) described this residual deposit as 0 to 50 ft of red, clayey shale at the base of the Minnelusa aquifer.

Other possible features influencing variability of hydraulic connection are faults, fractures, or breccia pipes. Hayes (1999) concluded that collapse brecciation within the Minnelusa Formation was the cause of episodic sediment discharge at Cascade Springs, which is located in the southern Black Hills. Hayes (1999) also used geochemical modeling to conclude that dissolution of anhydrite within the Minnelusa Formation by upward leakage from the Madison aquifer was the mechanism for development of the breccia pipes, which are the spring throats at Cascade Springs. Hayes (1999) further concluded that collapse brecciation resulting from upward leakage of water from the Madison aquifer is a probable mechanism that has contributed to development of numerous other artesian springs around the Black Hills area. Dye-tracer tests and geochemical analysis of artesian springs in the Black Hills area that flow from outcrops of geologic units overlying the Madison Limestone often indicate a source from the Madison aquifer (Klemp, 1995; Anderson and others, 1999, p. 37; Greene, 1999). Many breccia pipes that are visible in outcrop sections of the Minnelusa Formation probably are throats of previous artesian springs that have been abandoned over geologic time (Hayes, 1999). Breccia pipes that connect the Madison and Minnelusa aquifers but do not presently extend to the land surface could allow highly localized leakage between aquifers.

Breccia pipes can form when ground water dissolves gypsum and anhydrite in the Minnelusa Formation, thus, creating voids that initiate collapse brecciation, which may propagate upward to form vertical breccia pipes (fig. 14). Many breccia pipes probably formed along fractures, especially the intersection of fractures (fig. 14), where increased vertical ground-water flow could occur (Brobst and Epstein, 1963). Breccia pipes also can be initiated by collapse of the lower Minnelusa Formation into Madison Limestone caves (figs. 14 and 15) and propagate upward through the Minnelusa aquifer (Bowles and Braddock, 1963; Brobst and Epstein, 1963; Gott and others, 1974). These localized breccia pipes are in addition to the areally extensive brecciation in the Minnelusa Formation documented by Brobst and Epstein (1963).

Comparisons of hydrographs for paired observation wells indicate possible hydraulic connection between the Madison and Minnelusa aquifers at some locations but not at others. There are six paired wells within the aquifer analysis area (fig. 16). Three of these pairs (Tilford, City Quarry, Reptile Gardens) have similar hydrographs indicating a possible hydraulic connection between aquifers. The City Quarry wells, which have nearly identical hydrographs (fig. 16c), are located about one-half mile from City Springs and RC-6, where an aquifer test (Greene, 1993) and dye-tracer test (Greene, 1999) have indicated nearly certain hydraulic connection.

If recharge rates for the Madison and Minnelusa aquifers were similar, this could explain similarities in hydrographs; however, streamflow recharge rates, which are generally larger than areal recharge within the study area, are very different for the Madison and Minnelusa aquifers as discussed later. This is especially true in the southern part of the aquifer analysis area near the Reptile Gardens wells where the outcrop area is very small and streamflow losses are much larger than areal recharge, yet the paired wells have similar hydrographs (fig. 16f). This indicates that at least part of the hydrograph similarity could be the result of hydraulic connection. For example, the rise in hydraulic head in both aquifers in the summer of 1995 was about 20 ft. However, the estimated streamflow recharge rate from Spring Creek into the Madison aquifer was about 18 ft^3/s, whereas the estimated recharge rate to the Minnelusa aquifer was about 2 ft^3/s (see “Recharge from Streamflow Loss” section).

Conversely, dissimilarity of hydrographs does not necessarily indicate the absence of hydraulic connection. The Canyon Lake paired wells (fig. 16e) are located about 0.5 mi from Jackson-Cleghorn Springs where Madison aquifer water passes through the Minnelusa aquifer before discharging at the surface. Hydraulic head in the Minnelusa aquifer is generally more than 50 ft lower than in the Madison aquifer and shows no resemblance to the Madison hydrograph at Canyon Lake. Hydraulic head in the Minnelusa aquifer is apparently influenced by the level of Canyon Lake, as evidenced by a sharp water-level decline when the lake was drained near the end of 1995.

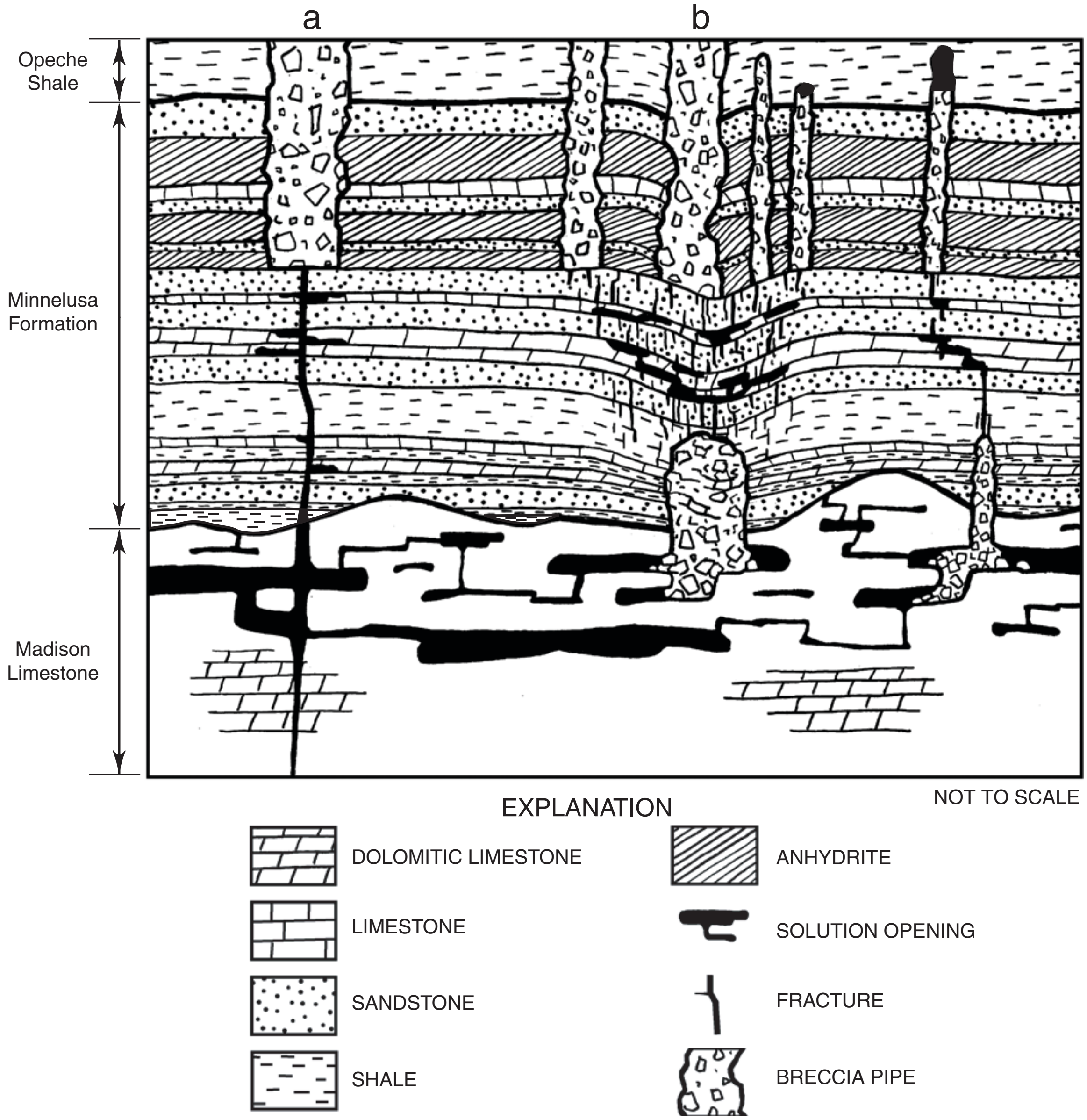


Figure 14. Geologic features that could enhance vertical hydraulic conductivity including (a) breccia pipe that formed along a fracture and (b) breccia pipe initiated by collapse of Minnelusa Formation into Madison Limestone cave.

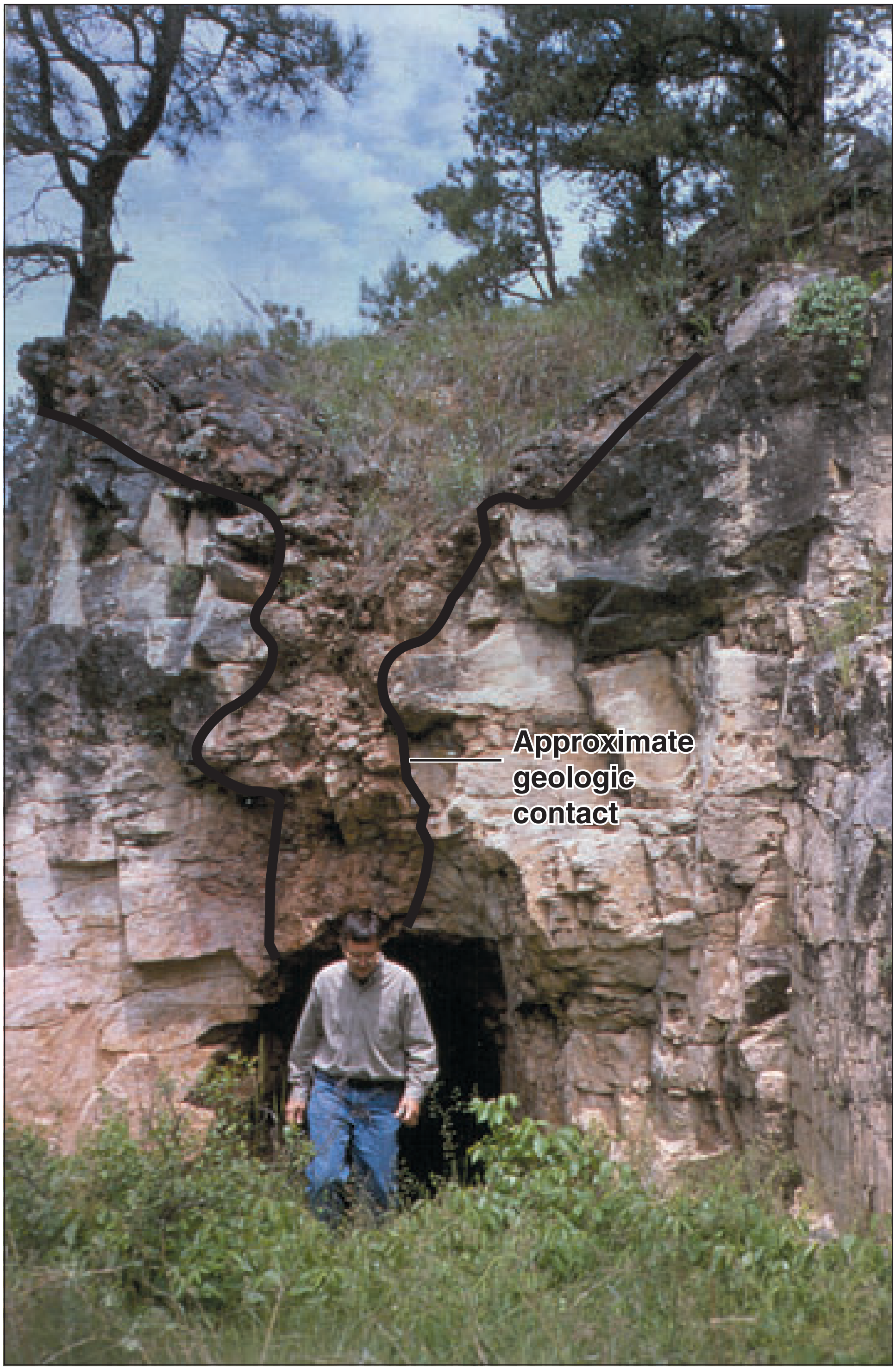


Figure 15. Entrance to Onyx Cave in the Madison Limestone in Wildcat Canyon of the southern Black Hills. Just above the cave opening are brecciated rocks of the Minnelusa Formation that have collapsed.

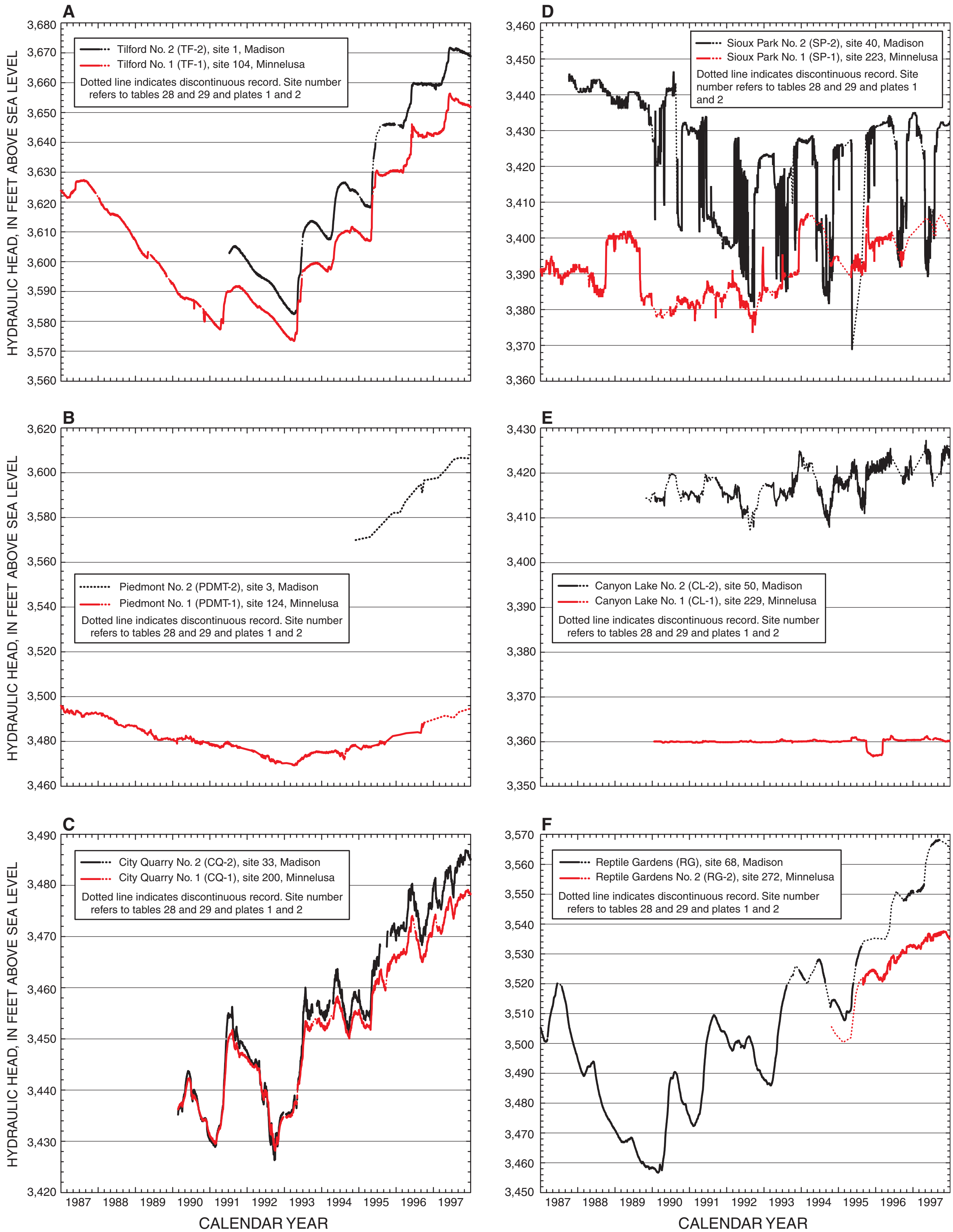


Figure 16. Hydrographs for paired wells.

Additional evidence of hydraulic connection between the aquifers is observed in the hydrograph of the Minnelusa observation well WCR-3 (figs. 13 and 17), which is located about 2 mi south of the City Quarry paired observation wells. Well WCR-3 shows declines in hydraulic head that are coincident with pumping from the Madison aquifer during September and October 1995. In October, an aquifer test (appendix A) was conducted in the Madison aquifer using production well RC-9 (fig. 9). Except for a short period of pumping from well RC-10, municipal withdrawals from the Madison and Minnelusa aquifers were terminated in that area from September 25 through October 16. On September 25, the WCR-3 hydrograph shows an upward deflection in hydraulic head (fig. 17) presumably due to the termination of pumping. When RC-10 began pumping on September 29, hydraulic head in WCR-3 declined until pumping ceased less than 2 days later. When RC-9 began pumping for the aquifer test beginning October 4, there was a sharp decline in hydraulic head. Hydraulic head began to increase on October 6 before pumping from RC-9 was terminated, which may have resulted because the aquifer was in a general recovery period and drawdown due to pumping RC-9 was beginning to flatten. The Black Hills Power and Light well (BHPL) was either not pumping or pumping intermittently because of plant maintenance until October 12 when the pumping rate increased, which is coincident with a decline at WCR-3. When Chapel Lane Water Company began using its production well (CHLN-2) on October 16, a similar decline was noted. These observations also indicate that hydraulic connection between the two aquifers is spatially variable, because noticeable responses did not occur in the Sioux Park Minnelusa well (fig. 13), which is located closer to the pumping wells than is WCR-3. Also, the Sioux Park paired wells have dissimilar hydrographs (fig. 16d).

Irregularities in the potentiometric surface of the Minnelusa aquifer near the 3,400-ft contour (fig. 13 inset) could be the result of localized upward leakage from the Madison aquifer. The central area of

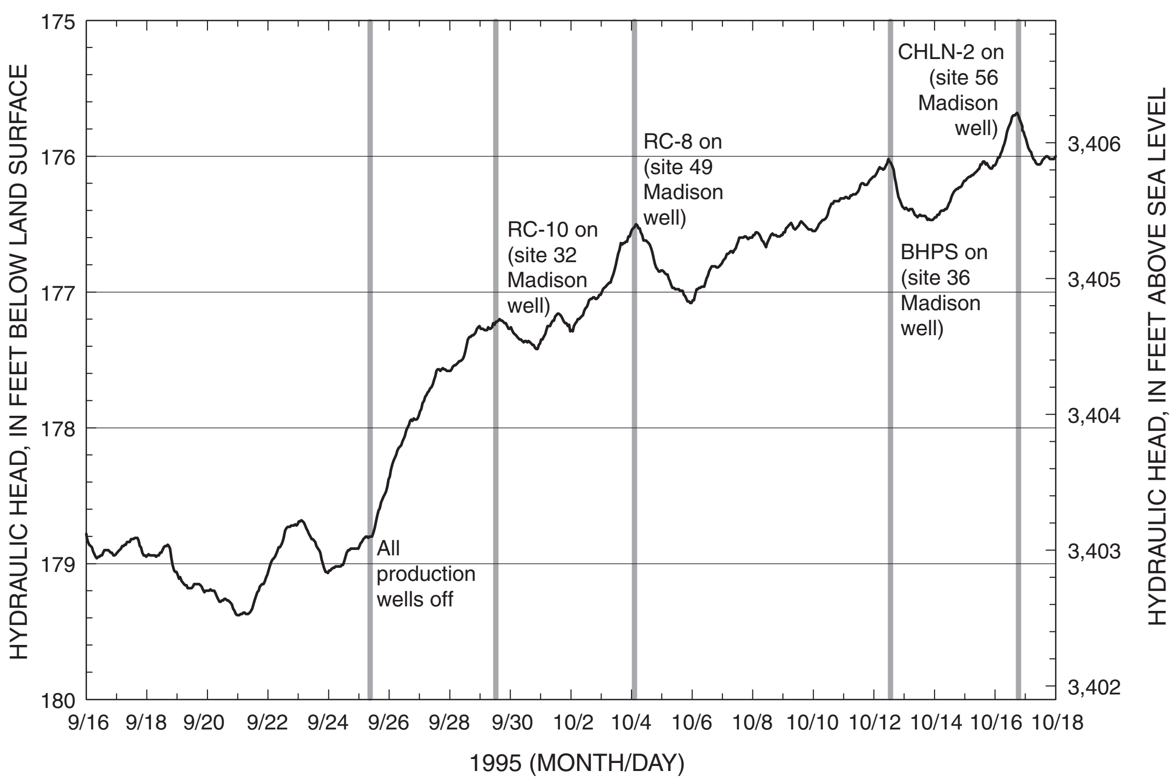


Figure 17. Possible response of hydraulic head at West Camp Rapid Minnelusa well (WCR-3) to pumping from Madison aquifer.

upward hydraulic gradient coincides with the high-flow area (figs. 11 and 12) and contains several artesian springs that are assumed or known to originate from the Madison aquifer. These upward hydraulic gradients could result in movement of Madison aquifer water into the Minnelusa aquifer through vertical breccia pipes to produce an irregular potentiometric surface in the Minnelusa aquifer.

Specific Yield

Specific yield (S_y) is the storage term used to describe and make calculations for storage in unconfined areas described in this report. Lohman and others (1972) defined S_y and related properties, which include total porosity, effective porosity, and specific retention. S_y of a rock or soil is defined as the ratio of (1) the volume of water that the rock or soil, after being saturated, will yield by gravity to (2) the volume of the rock or soil. S_y is equal to total porosity minus specific retention. The specific retention of a rock or soil is the ratio of (1) the volume of water that the rock or soil, after being saturated, will retain against the pull of gravity to (2) the volume of the rock or soil. Specific retention increases with decreasing grain size or pore size. Total porosity is the ratio of the total volume of pore space in a material to the total volume of the material. Effective porosity, which is an upper limit for S_y, is the ratio of interconnected pore space available for the transmission of water to the total volume of the material.

Due to heterogeneity, a wide range of porosity values for the Madison hydrogeologic unit have been reported (table 5). Although direct measurements of S_y were not available for the study area, estimates were made based on porosities and other hydrogeologic information. The average S_y was estimated as 0.09 for the Madison and Minnelusa aquifers and 0.03 for the Madison and Minnelusa confining units.

The definitions of S_y and specific retention imply sufficient time for gravity drainage to complete, which often is not the case in natural environments (Lohman and others, 1972). Therefore, the effective value of S_y is generally lower than that based strictly on the definition depending on factors such as particle or pore size, rate of water-table change, and time. Perching of water also can decrease the effective value of S_y because water is retained in these perched spaces as the water table declines. Other than in solution openings and fractures, the permeability of the Madison hydrogeologic unit is generally low, and water collects in poorly drained depressions on the floors of solution openings above the water table. These depressions often form on the floors of enlarged bedding planes, which have irregular surfaces and are nearly horizontal. The Minnelusa hydrogeologic unit also has a high potential to retain perched water because of low-permeability shale layers.

Estimates of S_y for the Madison and Minnelusa confining units are lower than for the aquifer units because of differences in lithology. S_y of the Madison confining unit is less than that of the aquifer because of smaller effective porosity and larger specific retention, which results from porosity dominated by small fractures rather than solution openings. The lower Minnelusa Formation generally contains less sandstone and more limestone, dolomite, and shale than the upper part (Bowles and Braddock, 1963). Therefore, smaller effective porosity and greater specific retention in the Minnelusa confining unit results in smaller S_y than in the aquifer.

Storage Coefficient

Storage coefficient, S, 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). In a confined aquifer, storage coefficient can be orders of magnitude smaller that in an unconfined aquifer. This is because in an unconfined aquifer, large changes in storage can occur due to the rise and fall of the water table; in a confined aquifer, smaller changes in storage occur as a result of the slight expansion or contraction of the aquifer material and water due to hydraulic-head changes. For an unconfined aquifer, storage coefficient is virtually equal to the specific yield (Lohman and others, 1972). Aquifer tests in the aquifer analysis area indicate that S varies between 1×10^{-4} and 2×10^{-3} for confined conditions of the Madison aquifer (table 2). The only value for the Minnelusa aquifer from these aquifer tests is 3×10^{-3}. However, table 3 shows that S as low as 7×10^{-5} has been estimated for the Minnelusa aquifer outside of the study area. Because there doesn't appear to be a definitive pattern, S for all confined areas of the Madison and Minnelusa aquifers is taken at a middle value of 3×10^{-4} for this report.

Table 5. Estimates of porosity and specific yield (S_y) for the Madison and Minnelusa hydrogeologic units from previous investigations and this study

[--, no data available]

Total porosity (as a fraction of 1)	Effective porosity (as a fraction of 1)	Specific yield (dimensionless)	Data source or method	Source
		Madison Limestone		
0.11	0.05	--	Oil tests	Rahn (1985)
		Madison Aquifer		
--	0.35 (average)	--	Well RC-6 resistivity log	Greene (1993)
--	0.35 (average)	--	Well LC resistivity log	Greene (1993)
--	--	0.09	Water budget[1]	This report
		Madison Confining Unit		
--	--	0.03	Water budget[1]	This report
		Minnelusa Formation		
0.10	0.05	--	Oil tests	Rahn (1985)
		Minnelusa Aquifer		
--	0.10 - 0.15	--	Well RC-6 neutron porosity log	Greene (1993)
--	0.05 - 0.10	--	Well RC-5 neutron porosity log	Greene (1993)
--	--	0.09	Water budget[1]	This report
		Minnelusa Confining Unit		
--	0.05 (average)	--	Well RC-6 neutron porosity log	Greene (1993)
--	0.05 (average)	--	Well RC-5 neutron porosity log	Greene (1993)
--	--	0.03	Water budget[1]	This report

[1]See “Water-Budget Analysis” section.

Hydraulic Head and Ground-Water Flow

Potentiometric surfaces, hydraulic response to stress, the extent of unconfined areas in relation to confined areas, and flowpaths are important considerations in understanding the dynamics of the ground-water-flow system. Analysis of hydraulic response to stress can be used to characterize the aquifers and estimate properties such as storage and transmissivity. Hydraulic gradients can be useful for estimating transmissivities, recharge areas, discharge areas, and flow directions. The areal extent of unconfined areas is important for analysis of storage and ground-water flow near recharge areas. Flowpaths are important when considering sources of springs and wells.

Potentiometric Surfaces

Potentiometric maps of the Madison and Minnelusa aquifers (pls. 1 and 2; figs. 11 and 12) were interpreted based on available hydraulic heads in observation wells, public-supply wells, and private wells. The potentiometric surfaces for both aquifers show a general easterly gradient of about 1,000 ft in 15 mi. Detailed information for the wells used in interpreting potentiometric maps is listed in tables 28 and 29 (appendix B). Where data are sparse in the eastern part of the aquifer analysis area, contours were based on Downey (1986).

Because the hydraulic heads were measured at various times, a potentiometric surface could not be interpreted for a particular date; therefore, the potentiometric maps represent an average potentiometric surface for WY88-97. To achieve this, each hydraulic-head measurement was either adjusted up if the measurement was made during a period of low water levels or adjusted down if made during a period of high water levels. Adjustment to each hydraulic head measurement was based on the WY88-97 hydrograph of the nearest continuous-record well (figs. 18 and 19). Hydrographs of longer periods of record are available for some wells (Driscoll, Bradford, and Moran, 2000). WY88-97 included a range of climatic conditions from dry during the late 1980's through wet conditions in the middle to late 1990's. Periods of continuous records that included the influence of pumping were eliminated from the analysis. Water levels for many wells have been obtained from drillers' reports, which generally are fairly accurate; however, locations can be inaccurate. In areas of large topographic relief, accurate locations are essential to determine land-surface altitudes. However, because it was infeasible to visit all of the wells to check or verify locations, hydraulic head measurements not considered reliable were not used (tables 28 and 29).

In a few cases, unusually high hydraulic heads were measured in wells located on or near the outcrop areas and are likely to represent perched water, which is known to exist near the Madison outcrop area. Wind Cave in the southern Black Hills contains a 300-ft-long body of water (Phantom Lake) that is perched considerably above the water table (Marc Ohms, Wind Cave National Park, oral commun., 2001). Hydraulic head measurements thought to be from perched water were omitted from the potentiometric-surface analysis.

The trough in the potentiometric surface in the eastern part of the Madison aquifer, where data are sparse, is based on the regional potentiometric surface of the Madison aquifer shown in Downey (1986, p. E39). The few hydraulic heads measurements available for the eastern part of the Madison aquifer (pl. 1) are in agreement with Downey's (1986) interpretation. Downey and Dinwiddie (1988, p. A47) show a narrow path of high easterly ground-water velocity in the Madison aquifer along the axis of this area, and Downey (1986, p. E54) shows a zone of anomalously high transmissivity in this area. If there is a high-transmissivity pathway in this area of the Madison aquifer that draws down the hydraulic head, it is likely to draw down hydraulic head in the Minnelusa aquifer as well, especially if there is enhanced vertical hydraulic conductivity in the Minnelusa confining unit. Therefore, the Minnelusa aquifer was interpreted as having a lowered potentiometric surface in this area (pl. 2).

Hydraulic Response to Recharge

Hydraulic-head changes in the Madison and Minnelusa aquifers in the study area are highly variable (figs. 18 and 19). The difference of the maximum and minimum hydraulic head in continuous-record observation wells during WY88-97 ranged from about 5 to 120 ft (fig. 20). With the effects of pumping from RC-5 removed from the LC and SP-2 hydrographs, a pattern emerges whereby hydraulic-head change is small in the Jackson-Cleghorn Springs area and increases with distance from the springs (fig. 20). The transmissivity distribution probably is part of the reason for this pattern because higher transmissivity tends to damp hydraulic head fluctuations in response to stresses such as recharge or pumping. Based on aquifer tests and the hydraulic gradients shown on plates 1 and 2, transmissivity near the Jackson-Cleghorn Springs area is high and generally decreases with distance from the springs.

Hydraulic head fluctuates in response to changing recharge rates at the outcrop areas. In the Spring Creek area where areal recharge is small compared to streamflow recharge, a direct correlation of streamflow loss to hydraulic head in the Madison aquifer was described by Long and Derickson (1999) by invoking a time-invariant transfer function (fig. 21). Hydraulic head can be predicted based on this transfer function applied to the streamflow-loss rate. Based on data from the Reptile Gardens Madison well (RG) about 3 mi from the recharge area, the transfer function showed that (1) the peak response is less than one month, (2) the system has memory of about 4 years, and (3) ground-water recession follows a logarithmic curve. The very long memory (elapsed time before all effects of a stress have diminished) in contrast to the short response time of the system probably results from dual porosity. A long-term system memory can result from delayed storage within the aquifer's matrix of tiny fractures and small pore spaces in contrast to the very fast flowpaths of large solution openings. This also can result from leakage to and from overlying or underlying confining units, which can store water in a similar way to that of a dual-porosity system. Numerical modeling has shown that hydraulic head fluctuation in the Madison aquifer can be damped by leakage into and out of vertically adjacent layers and also by dual-porosity effects (Long, 2000). Storage properties and/or the hydraulic response of dual-porosity reservoirs are described in Barenblatt and others (1960), Warren and Root (1963), Streltsova (1988), and Bai and others (1993). Streltsova (1988) also describes these properties for leaky aquifers.

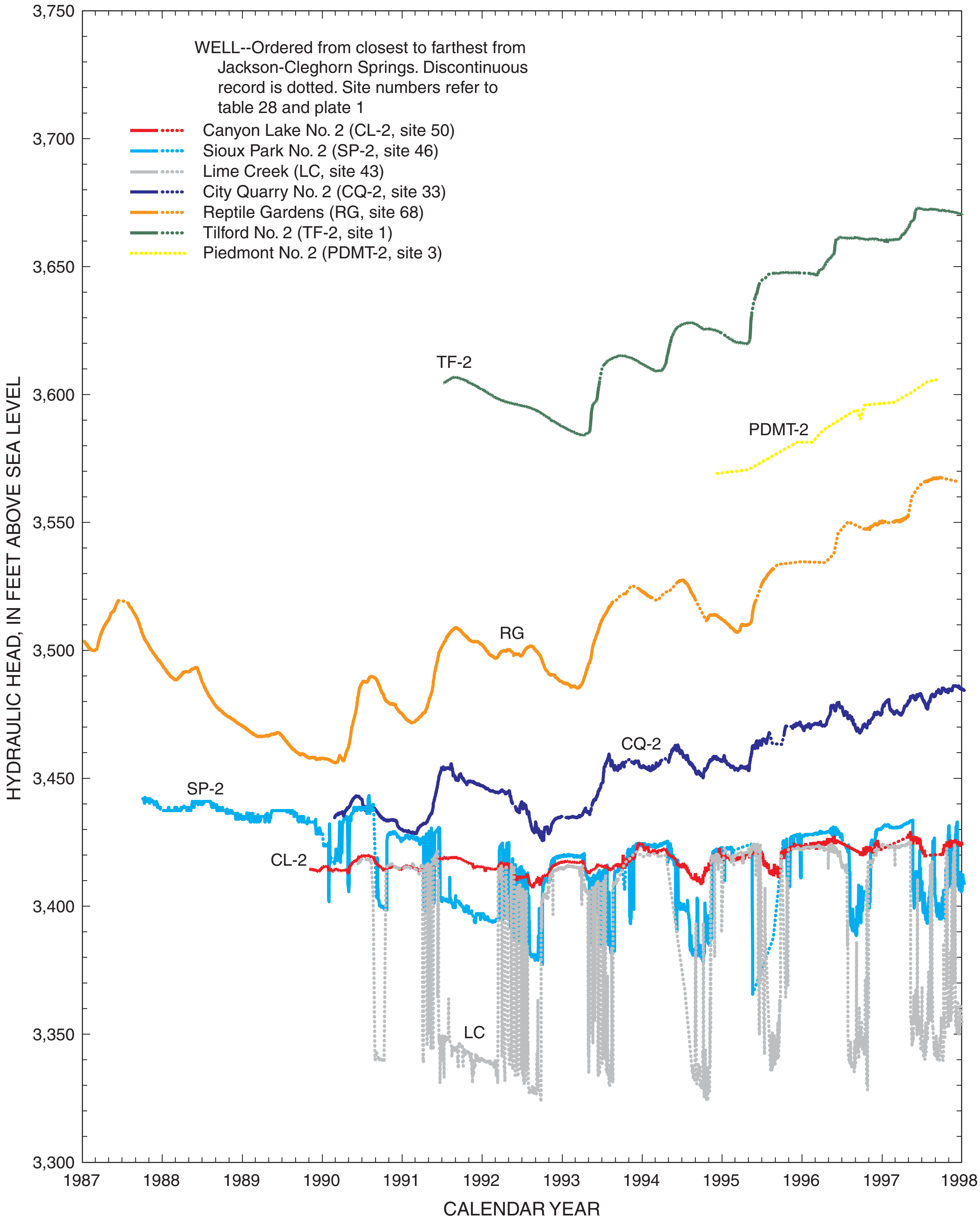


Figure 18. Hydrographs of continuous-record observation wells in the Madison aquifer.

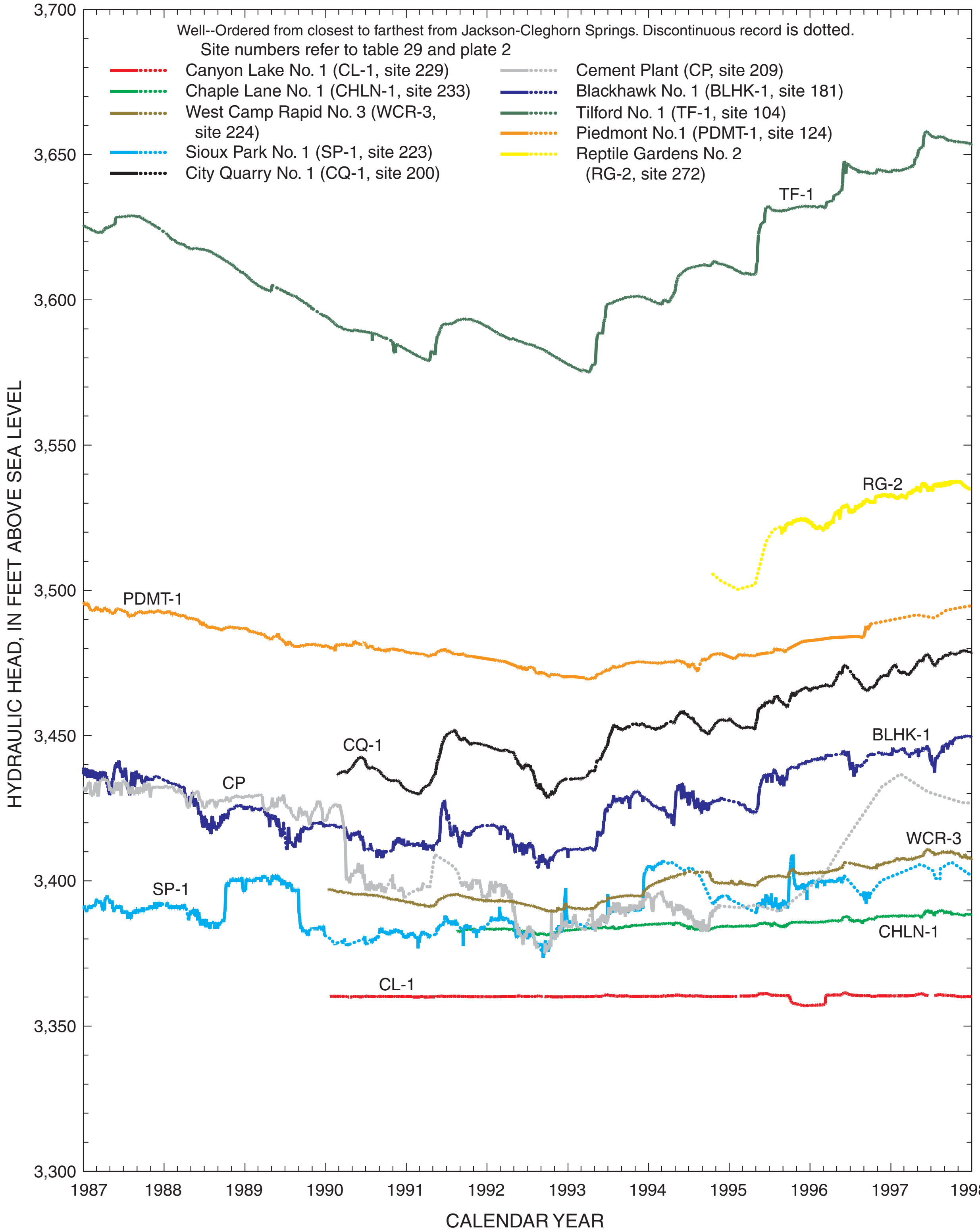
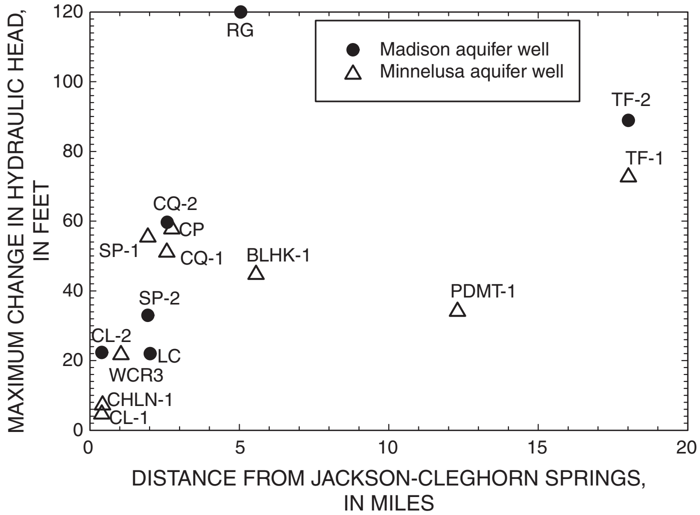


Figure 19. Hydrographs of continuous-record observation wells in the Minnelusa aquifer.



Well name	Site number
TF-2	1
CQ-2	33
LC	43
SP-2	46
CL-2	50
RG	68
TF-1	104
PDMT-1	124
BLHK-1	181
CQ-1	200
CP	209
SP-1	223
WCR-3	224
CL-1	229
CHLN-1	233

Site number refers to tables 28 and 29 and plates 1 and 2

Figure 20. Maximum hydraulic head change in selected continuous-record observation wells during WY88-97 versus distance from Jackson-Cleghorn Springs. In general, maximum hydraulic head change increases with distance from Jackson-Cleghorn Springs. Maximum hydraulic head change in wells with shorter periods of record was estimated. Hydrographs are plotted in figures 18 and 19.

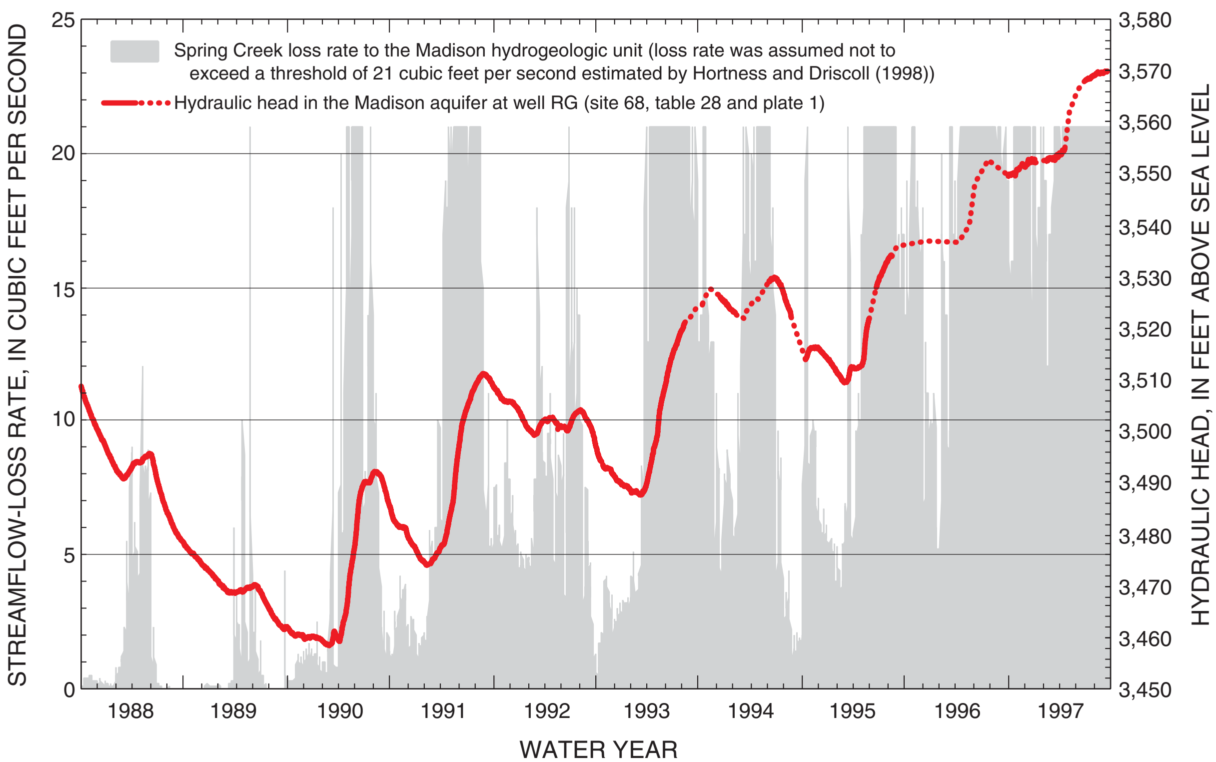


Figure 21. Relation between Spring Creek streamflow loss and hydraulic head in Madison aquifer.

Unconfined Areas

The boundaries of unconfined areas (pls. 1 and 2) are determined by the locations where the average potentiometric surfaces (WY88-97) contact the tops and bottoms of the Madison and Minnelusa aquifers and associated confining units. Altitudes for top and bottom of the Madison and Minnelusa hydrogeologic units were taken from maps of the structural tops of the Minnelusa Formation, Madison Limestone, and Deadwood Formation from Carter and Redden (1999a, 1999b, 1999c). The top of the Minnelusa Formation was taken as the top of the Minnelusa hydrogeologic unit. The top of the Madison Limestone was taken as the top of the Madison hydrogeologic unit and the bottom of the Minnelusa hydrogeologic unit. The top of the Deadwood Formation was taken as the bottom of the Madison hydrogeologic unit. A uniform thickness of 150 ft (assumed thickness of the Madison aquifer) was subtracted from the top of the Madison hydrogeologic unit to estimate the altitude of the surface between the aquifer and confining unit. The same was done for the Minnelusa hydrogeologic unit, with a uniform thickness of 250 ft assumed for the aquifer thickness. These aquifer thicknesses are the midpoints of ranges given by Greene (1993).

Because potentiometric surfaces are generally some distance below the land surface in the outcrop areas, and the aquifers and confining units dip to the east (fig. 6), unconfined areas are present eastward of outcrop areas. The western part of outcrop areas generally are not saturated but may contain infiltrating or perched water.

Boundaries of the unconfined areas of the Madison and Minnelusa aquifers shift to the east or west as hydraulic head fluctuates. During WY88-97, the maximum fluctuation in measured hydraulic head was about 100 ft, which would cause these boundaries to shift a distance of 600 to 2,000 ft or about 1 to 2 percent of the total east-west dimension of the aquifer analysis area. Although unconfined areas shift horizontally, the total areal coverage changes negligibly. Because of these comparatively small changes in position and area, the unconfined zones shown on plates 1 and 2 were assumed to be spatially constant in time for water-budget purposes. Saturated thickness and transmissivity of the unconfined zones were assumed constant in time, which is justifiable if the dip of the beds are relatively constant within areas where the unconfined areas shift from east to west. It follows that even if unconfined areas change position vertically and horizontally as hydraulic heads rise and fall, saturated thicknesses also would be assumed constant.

The location and extent of the unconfined areas are significant in analyzing aquifer storage because most of the change in storage occurs in these areas. Estimates of specific yield (unconfined storage term) for the Madison and Minnelusa aquifers are three or more orders of magnitude larger than estimates of storage coefficient for confined conditions (tables 2, 3, and 5). Estimates of unconfined area coverages for the Madison and Minnelusa hydrogeologic units are 52.9 mi^2 and 36.3 mi^2, respectively. Although the unconfined areas represent a small percentage of the aquifer analysis area (629.4 mi^2), the amount of water these areas are capable of releasing from storage with a given change in head is orders of magnitude larger than that released in the confined part of the aquifer.

Eight hydrogeologic sections (fig. 22), A-A′ through H-H′ , (pls. 1 and 2), show differences in unconfined areas. Section A-A′ shows how the steep dip of a monocline results in a small unconfined area in the Madison aquifer relative to that of the Minnelusa aquifer. Section B-B′ illustrates an area where the Madison outcrop is large, but most of the outcrop area is unsaturated with a small unconfined area near the eastern edge of the outcrop. Although most of the outcrop is not saturated at this location, the areal extent of the outcrop makes it a large source of areal recharge. The “unsaturated area” in section B-B′ consists mainly of the lower part of the Madison hydrogeologic unit, which is generally considered a confining unit; however, weathering and dissolution of fractures has probably increased its permeability in outcrop areas allowing recharge water to quickly infiltrate and flow laterally toward water-table areas. This flow might consist of water cascading through a series of subsurface pools or non-Darcian channel flow, and could have very high velocities in some areas. For example, dye testing (Rahn and Gries, 1973; Greene, 1999) indicated ground-water flow velocities on the order of miles per hour along a flowpath starting in a swallow hole in Boxelder Creek, located in the Madison confining unit outcrop, and emerging from Gravel and Doty Springs (see cover photo).

Sections C-C′ and D-D′ illustrate how an anticline-syncline structural feature with a southeasterly plunge influences ground-water flow and the areal extent of the unconfined area near Boxelder Creek (pl. 1). Section C-C′ shows that saturated areas of the Madison hydrogeologic unit are separated by an anticline. Section D-D′ shows how an anticline blocks much easterly flow in the Madison aquifer, thus, diverting flow to the southeast along the syncline axis (pl. 1). Therefore, a large part of the streamflow recharge from Boxelder Creek probably is directed toward Rapid City. Some easterly flow within the Madison confining unit also is plausible but probably is minor compared to southeasterly flow in the aquifer.

Section E-E′ shows the uniform dip of the hydrogeologic strata in the Rapid Creek and Jackson-Cleghorn Springs area. A fault separates section E-E′ from section F-F′ on the upthrown side (pl. 1) where strata have been lifted higher than on the downthrown side in relation to the potentiometric surfaces. In addition, gently dipping beds and structural features have resulted in large unconfined areas or unsaturated areas along section F-F′ .

Section G-G′ shows how variations in dip create a wide Madison outcrop area in comparison to that of the Minnelusa outcrop. Section H-H′ illustrates the reason that the unconfined area of the Madison hydrogeologic unit is farther to the east of the outcrop than at section G-G′ .

Flowpaths

Dye-tracer tests, stable isotopes, and hydrogeologic features were analyzed conjunctively to estimate generalized ground-water flowpaths in the Madison aquifer and analyze the influences of flowpaths in the Minnelusa aquifer. A dye-tracer test (Greene, 1999) showed that ground water moved rapidly from the Boxelder Creek loss zone to five ground-water sampling sites in the Madison aquifer. Natural tracers in the form of stable isotopes provided more generalized evidence but were available for a larger part of the study area. The spatial configuration of saturated areas in relation to unsaturated areas indicated probable flowpaths at some locations.

As discussed in the previous section, the anticline-syncline set that crosses Boxelder Creek (fig. 23) has the potential to impede easterly ground water recharged near Boxelder Creek. Section D-D′ (fig. 22,) shows that easterly ground-water flow across the anticline, if any, would be within the Madison confining unit. The more likely flowpath, however, is to the southeast along the axis of the syncline where the Madison aquifer is fully saturated.

The dye-tracer test (Greene, 1993) indicates a focused flowpath, which probably follows the syncline southeasterly before turning to the east (fig. 23). The dye-tracer test consisted of injection of Rhodamine WT dye at the Boxelder Creek loss zone, which was detected at City Springs, RC-6, RC-10, BHPL, WT-2, and CQ-2 but was not detected at Jackson-Cleghorn Springs, RC-5, RC-8, or RC-9 (fig. 23). Except for observation well CQ-2, all these wells are production wells that were being pumped. The sites where dye was detected are grouped in a semi-linear pattern oriented downgradient, bounded by RC-8 on the north and RC-5 on the south. Greene (1993) measured breakthrough curves for all sites where dye was detected except for CQ-2. Dye detection at City Springs began 30 days after injection and continued until the 261st day with a peak concentration at 48 days. Dye detection at RC-10 and BHPL began less than 50 days after injection and continued until the 198th and 159th day respectively with peak concentrations less than 10 days after that of City Springs. If the flowpath was dispersive rather than focused, dye may have been detected at RC-5 and RC-8, especially considering the duration of breakthrough curves at the other sites. In addition, the mass recovery of 36 percent of the injected dye reported by Greene (1993) probably would not be possible without a focused flowpath.

Stable isotopes of oxygen and hydrogen (^{18}O and ^{2}H) in recharge water can be used as natural tracers to determine source areas and flowpaths for ground water. Isotope values are reported as a ratio of $^{18}O/^{16}O$ or $^{2}H/^{1}H$ of a sample water compared to a standard. The equation for oxygen given in parts per thousand (per mil) is as follows:

$$\delta^{18}O_{sample} = \frac{(^{18}O/^{16}O)_{sample} - (^{18}O/^{16}O)_{standard}}{(^{18}O/^{16}O)_{standard}} \times 1,000. \tag{1}$$

A smaller delta (δ) value is referred to as being isotopically lighter, while a larger value is considered heavier. For further discussion on stable isotope analysis in natural waters and distribution in the Black Hills area, see Naus and others (2001).

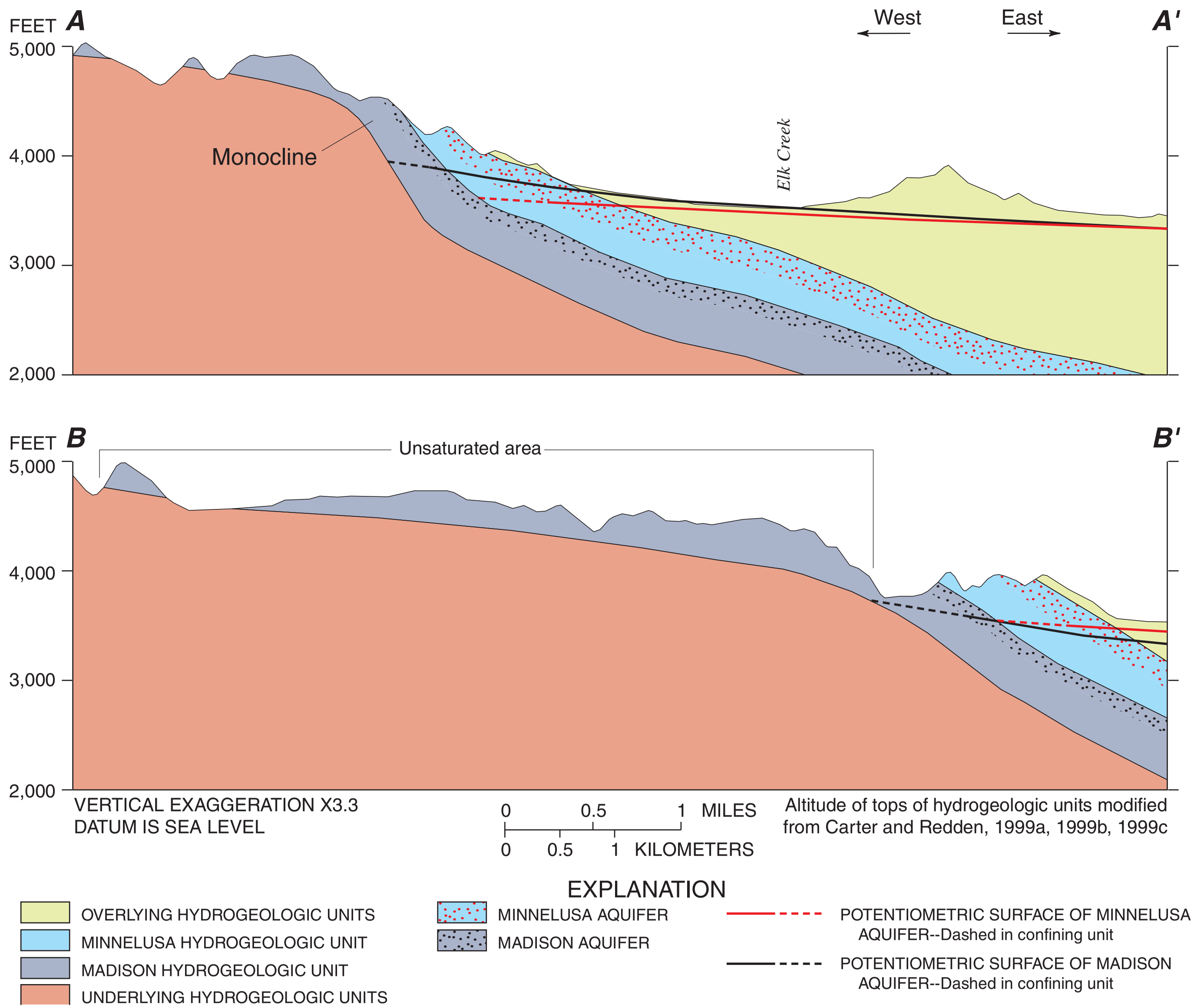


Figure 22. Hydrogeologic section showing average potentiometric surfaces of the Madison and Minnelusa aquifers (plates 1 and 2) in relation to hydrogeologic strata (location of sections shown on plates 1 and 2).

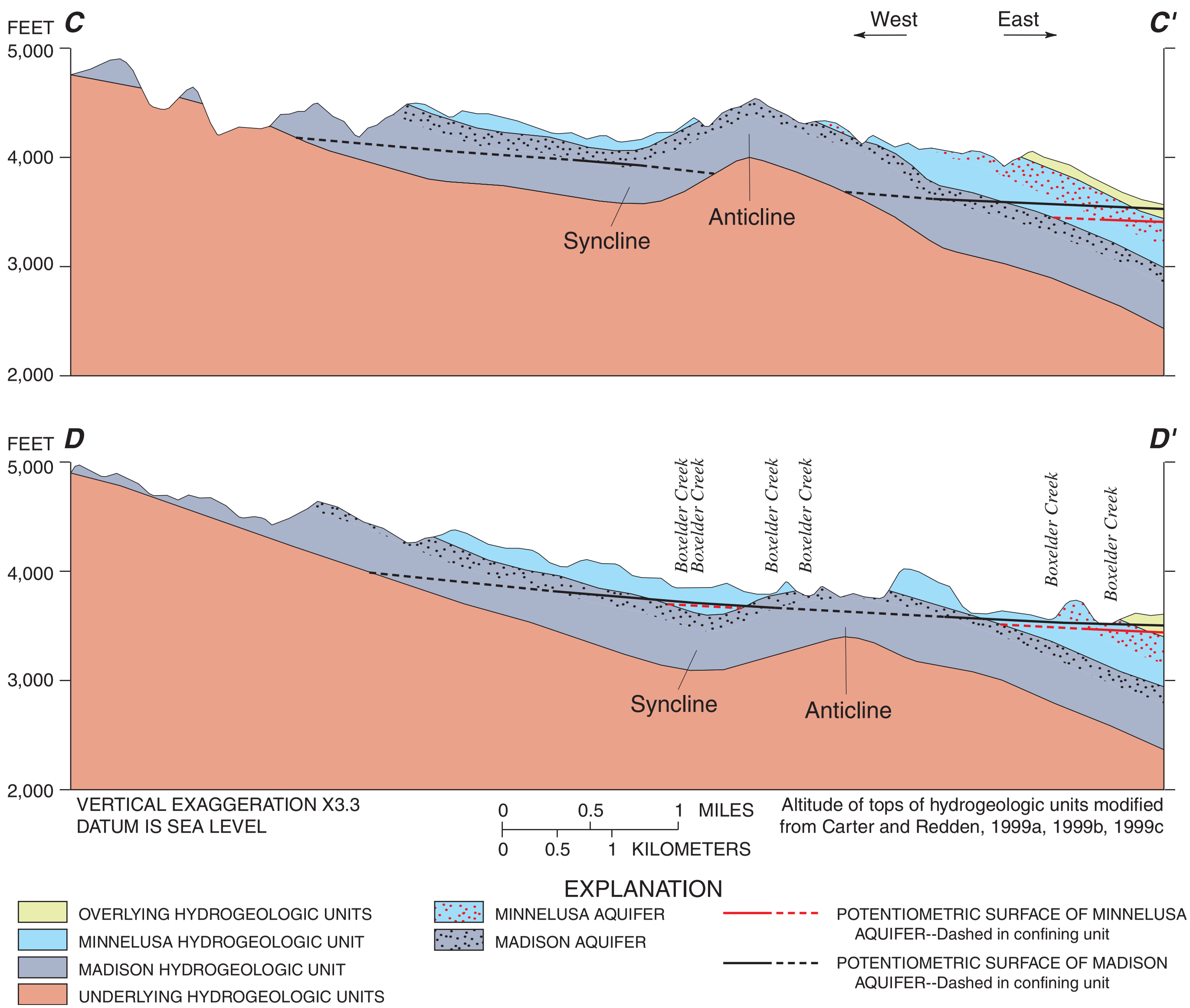


Figure 22. Hydrogeologic section showing average potentiometric surfaces of the Madison and Minnelusa aquifers (plates 1 and 2) in relation to hydrogeologic strata (location of sections shown on plates 1 and 2).—Continued

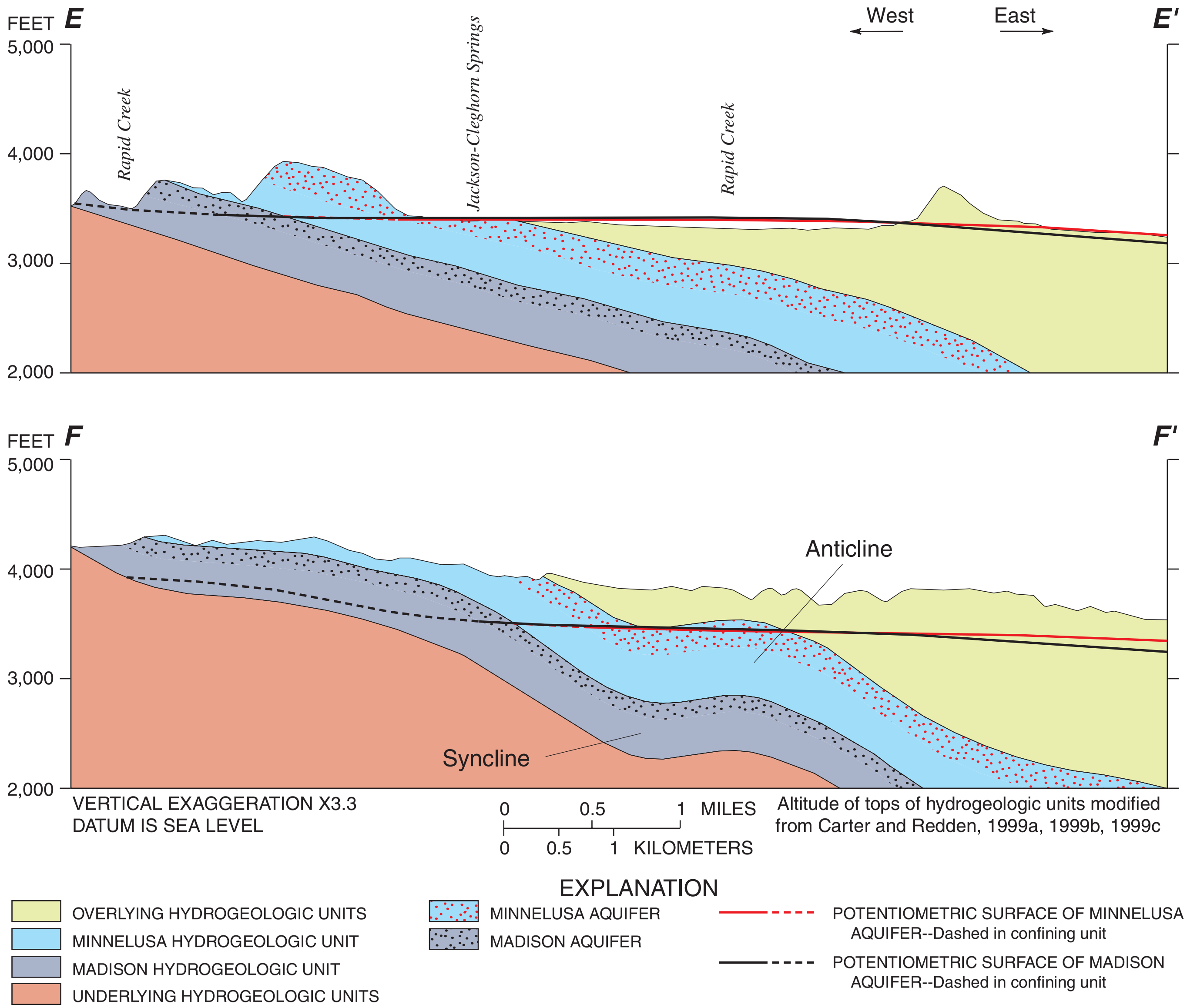


Figure 22. Hydrogeologic section showing average potentiometric surfaces of the Madison and Minnelusa aquifers (plates 1 and 2) in relation to hydrogeologic strata (location of sections shown on plates 1 and 2).—Continued

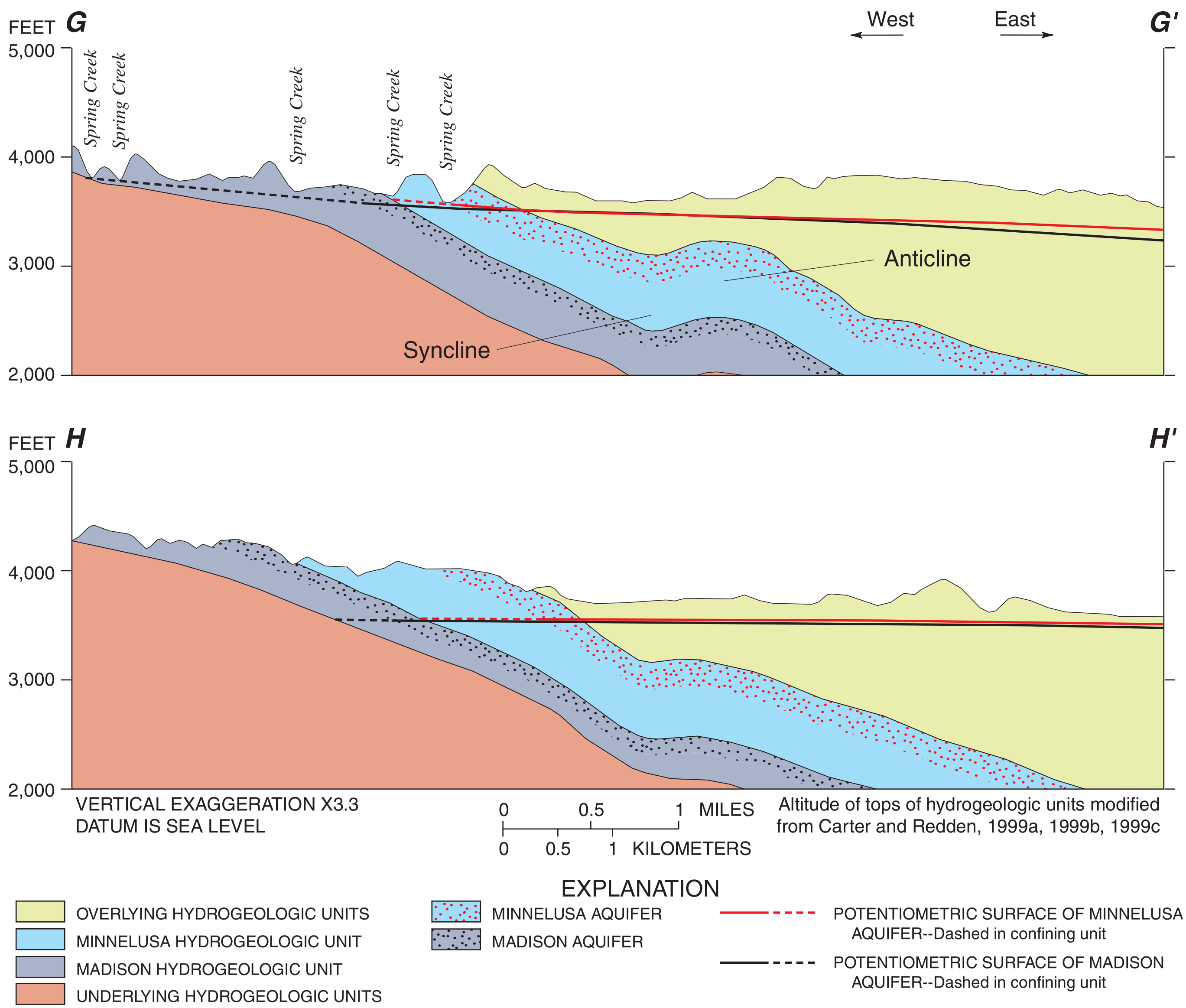


Figure 22. Hydrogeologic section showing average potentiometric surfaces of the Madison and Minnelusa aquifers (plates 1 and 2) in relation to hydrogeologic strata (location of sections shown on plates 1 and 2).—Continued

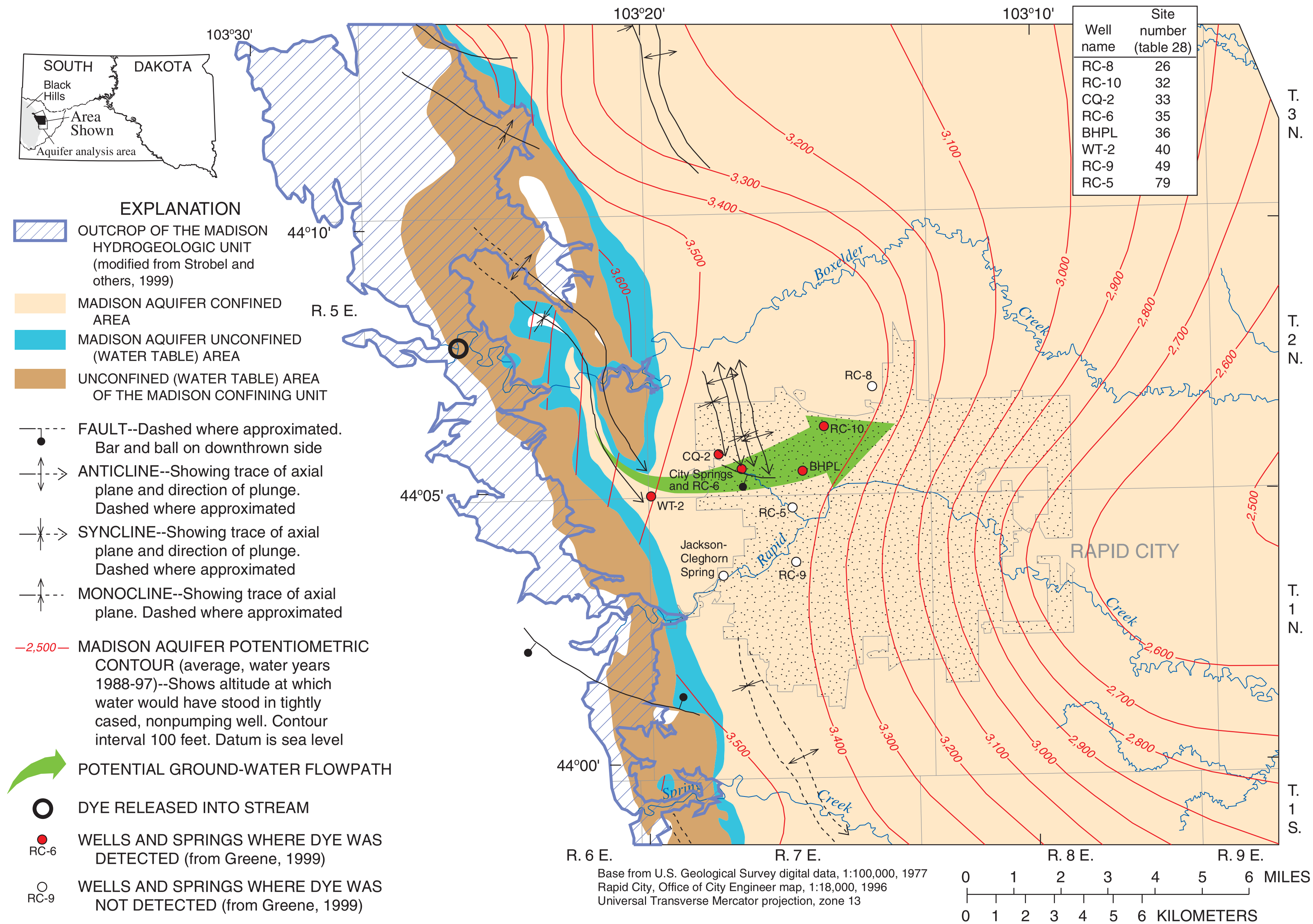


Figure 23. Generalized ground-water flowpath in the Madison aquifer determined from Boxelder Creek dye test.

Boxelder Creek, Rapid Creek, and Spring Creek have unique isotopic signatures in recharge water, which makes estimation of flowpaths possible in some locations. With respect to $\delta^{18}O$, Spring Creek is the isotopically heaviest of the three streams, Boxelder Creek is lightest, and Rapid Creek is between the two (table 6, fig. 24). Precipitation is isotopically lightest in the northwestern part of the study area and becomes heavier to the southeast. Figure 25 presents a generalized distribution of $\delta^{18}O$ in surface water and ground water near recharge areas (Naus and others, 2001).

Figure 26 shows flowpaths interpreted from isotope analysis and other information previously described. The flowpaths originate from streamflow-loss zones in the central and southern areas and from areal outcrop recharge in the northern area where the Madison outcrop is large. $\delta^{18}O$ values in the Madison aquifer in the Rapid City area (Naus and others, 2001) were grouped into three ranges: a heavy range (greater than -13 per mil) indicating Spring Creek origin, a medium range (-13 to -14 per mil) indicating Rapid Creek origin, and a light range (less than -14 per mil) indicating Boxelder Creek origin. The -13 to -14 per mil range also is representative of areal recharge to the north of Boxelder Creek (fig. 25) where the influence of areal recharge is relatively large because of the large outcrop area. Ranges indicating recharge from Rapid Creek and Boxelder Creek are heavier than the values shown in table 6 because of mixing with areal recharge from the outcrop, which is isotopically heavier. Figure 25 shows that stream basins collect water from isotopically lighter areas and recharge the aquifers in areas where areal recharge is isotopically heavier. The exception to this is Spring Creek, where the basin centroid is isotopically similar to the outcrop area near its loss zone.

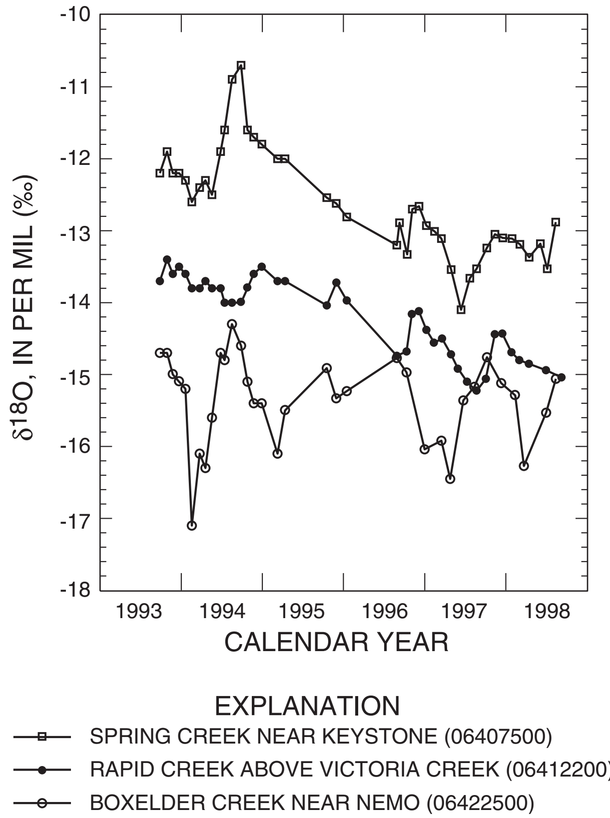
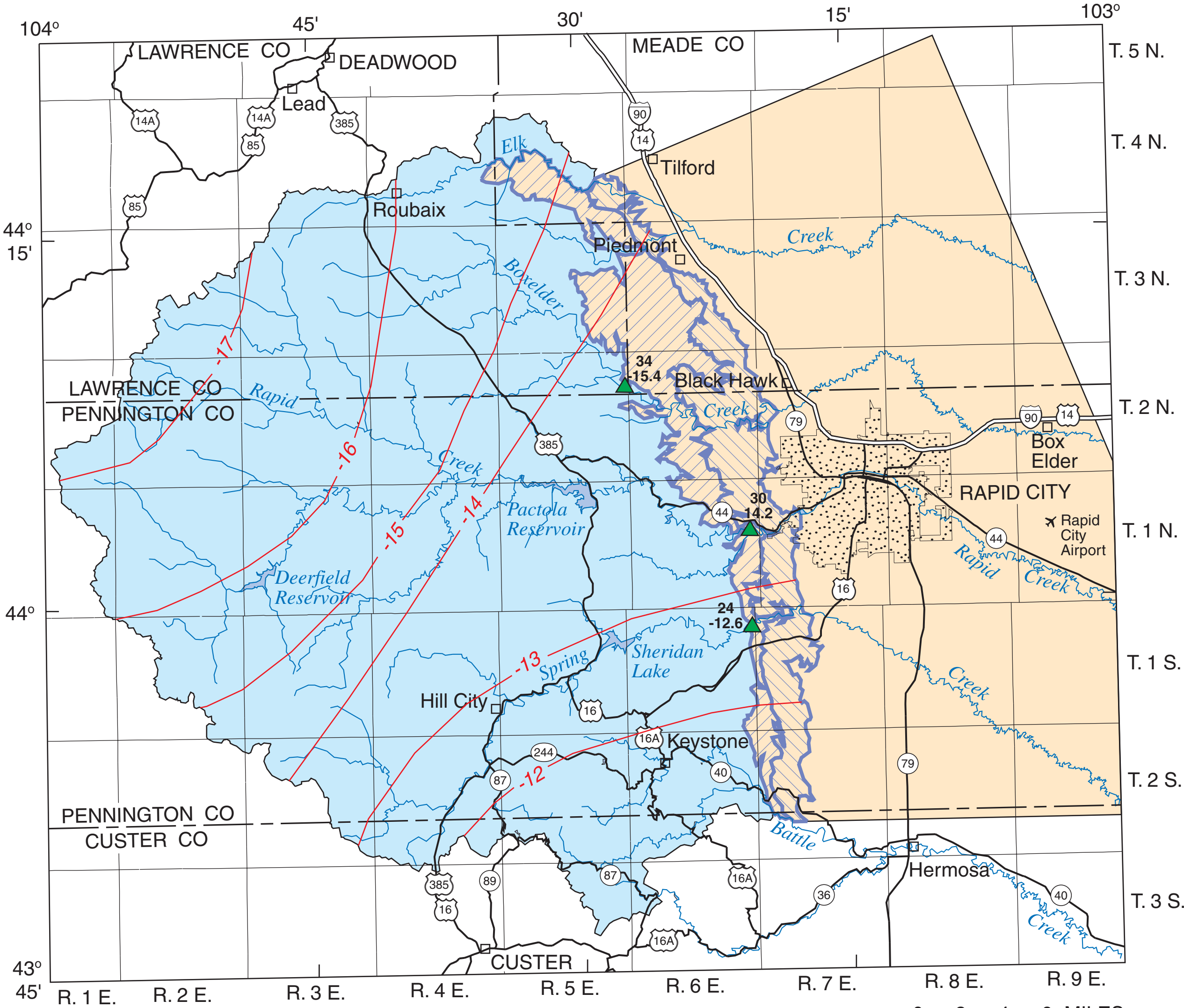


Figure 24. Temporal variation of $\delta^{18}O$ in Spring Creek, Rapid Creek, and Boxelder Creek (modified from Naus and others, 2001).

Table 6. Summary of $\delta^{18}O$ for Spring Creek, Rapid Creek, and Boxelder Creek

[Values reported by Naus and others (2001) for sampling locations upstream from loss zones]

Site number (fig. 25)	Gaging station name and number	Number of samples	$\delta^{18}O$ value (per mil)		
			Average	Maximum	Minimum
24	Spring Creek near Keystone (06407500)	41	-12.6	-10.7	-14.1
30	Rapid Creek above Victoria Creek (06412200)	39	-14.2	-13.4	-15.2
34	Boxelder Creek near Nemo (06422500)	33	-15.4	-14.3	-17.1



Base from U.S. Geological Survey digital data, 1:100,000, 1977
Rapid City, Office of City Engineer map, 1:18,000, 1996
Universal Transverse Mercator projection, zone 13

EXPLANATION

OUTCROP OF THE MADISON HYDROGEOLOGIC UNIT (modified from Strobel and others, 1999)

OUTCROP OF THE MINNELUSA HYDROGEOLOGIC UNIT (modified from Strobel and others, 1999)

DRAINAGE AREAS CONTRIBUTING TO STREAMFLOW LOSS

AQUIFER ANALYSIS AREA

—-15— LINE OF EQUAL AVERAGE $\delta^{18}O$ VALUE--Contour interval 1 per mil (modified from Naus and others, 2001)

34 -15.4 STREAMFLOW-GAGING STATION--Numbers indicate gage-site table and average $\delta^{18}O$ value in per mil

Figure 25. Generalized distribution of $\delta^{18}O$ in surface water and ground water near recharge areas. Contours indicate $\delta^{18}O$ values of areal-recharge water, whereas $\delta^{18}O$ values of streamflow-gaging stations indicate that of streamflow recharge.

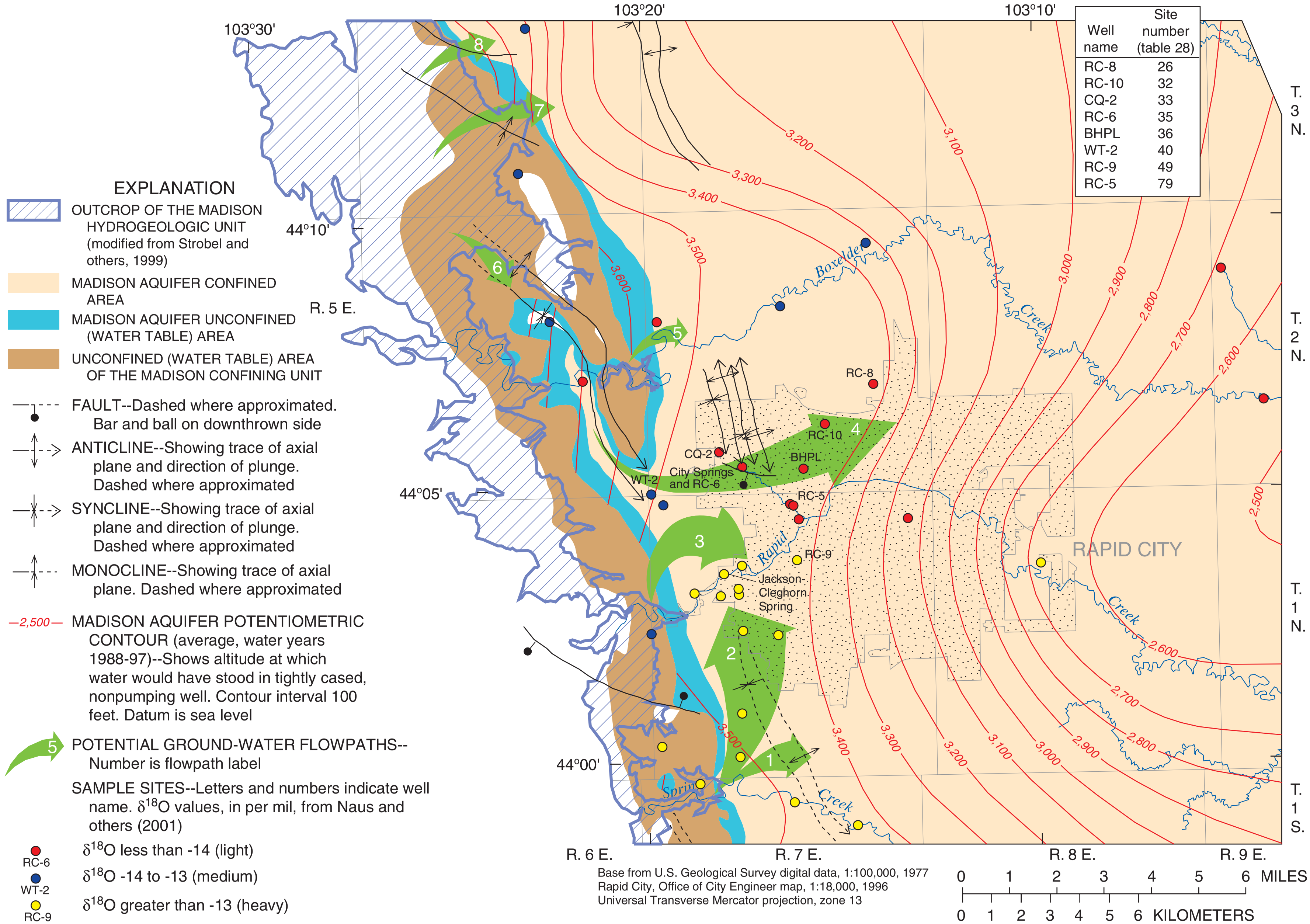


Figure 26. Generalized ground-water flowpaths in the Madison aquifer based on $\delta^{18}O$.

Ground-water flowpaths in the Madison aquifer probably are influenced by the large and stable discharge (about 22 ft^3/s) of Jackson-Cleghorn Springs (Anderson and others, 1999). Naus and others (2001) reported a $\delta^{18}O$ value of -12.92 per mil (average from 1986 to 1998) for Jackson-Cleghorn Springs, which indicates a large, and probably dominant, contribution from the south (table 6). Numerous other sites in the heavy range (greater than -13 per mil) indicate a northward flowpath (fig. 26, flowpath 2) from Spring Creek as well. This interpretation is consistent with conclusions of Anderson and others (1999) and Greene (1997). Streamflow recharge from Spring Creek combined with outcrop recharge between Spring Creek and Rapid Creek is estimated at about 12 ft^3/s on average (see "Water Budget" section). The steady streamflow-loss rate from Rapid Creek of about 10 ft^3/s to the Madison and Minnelusa outcrops (Hortness and Driscoll, 1998) is the maximum that Rapid Creek could supply to Jackson-Cleghorn Springs. Based on these recharge rates and discharge at the spring complex, all of the streamflow recharge from Spring Creek probably moves northward during periods of lower streamflows; some of this recharge probably moves easterly during higher streamflow periods (flowpath 1). Because of these temporal variations in flow direction, ground water surrounding the Spring Creek loss zone could be a mixture of Spring Creek water and water from farther south.

Because of the sites east of the Rapid Creek loss zone indicating water from the south, Rapid Creek streamflow recharge probably moves initially northward, then eastward in the Madison aquifer (fig. 26, flowpath 3). Anderson and others (1999) drew similar conclusions based on isotope analysis. Probably, some of the water recharged from Rapid Creek discharges from Jackson-Cleghorn Springs and some continues to the east.

The light-range isotope value near flowpath 5 (fig. 26) is similar to values for Boxelder Creek and probably results from streamflow recharge that occurs within the isolated Madison outcrop on the east side of the anticline. $\delta^{18}O$ values for two sites that are farther east near Boxelder Creek are slightly heavier and probably indicate influence from areal recharge north of Boxelder Creek (flowpaths 6, 7, and 8).

The Boxelder Creek dye-tracer flowpath (fig. 23) coincides with $\delta^{18}O$ values in the light range as well. The $\delta^{18}O$ values along flowpath 4 that are influenced by water originating from Boxelder Creek are isotopically heavier than the stream water due to mixing with areal recharge water. The two $\delta^{18}O$ values to the northeast of Rapid City in the light range indicate that flowpath 4 probably extends farther eastward in that direction.

Although light $\delta^{18}O$ at RC-5 and RC-8 (fig. 26) indicate the presence of water recharged from Boxelder Creek, dye was not detected in these wells. Non detection of dye does not necessarily indicate the absence of water recharged from Boxelder Creek, but does indicate that preferential flowpaths were not being sampled. Furthermore, flowpaths could be somewhat transient in nature, influenced by factors such as changing hydraulic head, pumping, or dual porosity.

Sufficient information is not available for estimation of flowpaths in the Minnelusa aquifer; however, increased ground-water circulation in the Minnelusa aquifer could, in part, be the result of collapse of solution features in the Madison aquifer where there is a convergence of preferential flowpaths in the high-flow area (fig. 11). The easterly bulge in the transition zone between low and high sulfate concentrations in the Minnelusa aquifer (fig. 12) is congruent with flowpath 4 (fig. 26). Some of this converging water in the Madison aquifer discharges to springs; some probably leaks upward through fractures, faults, and breccia pipes into the Minnelusa aquifer, then flows outward; and some flows eastward in the Madison aquifer. Much of the water that flows eastward in the Madison aquifer probably flows toward the zone of high transmissivity in the east-central part of the aquifer analysis area (fig. 9).

WATER-BUDGET ANALYSIS

Water-budget analysis of the Madison and Minnelusa hydrogeologic units included a 10-year period, WY88-97. Adequate data were available for this period, which includes periods of low and high recharge rates that represent long-term variability (Carter, Driscoll, and Hamade, 2001) fairly well.

Generally, ground-water levels declined during the first 5 or 6 years, then steadily rose during the remaining time (fig. 16). The 10-year period was divided into 20 seasonal stress periods for analysis. Each water year was divided into a winter period (October 1 through March 31), which generally has relatively low precipitation, and a summer period (April 1 through September 30), which generally has greater precipitation.

The water-budget conceptual model (fig. 27) includes the inflows and outflows as well as the general flow interactions within the model. The northern and southern boundaries of the aquifer analysis area are approximately perpendicular to potentiometric contours (pls. 1 and 2) and, therefore, were assumed to be boundaries where ground-water flow does not cross (pl. 3). Isotropic conditions near boundaries also were assumed. In order to balance a water budget, it was necessary to include many other simplifying assumptions and approximations because of incomplete data in many cases.

General Concepts

General concepts and water-budget summaries are described in this section, followed by detailed discussions of methods and results for individual budget components. Three budgets were developed: (1) a dry-period budget for declining water levels, (2) a wet-period budget for rising water levels, and (3) a full-period budget. All inflows and outflows were estimated separately for the Madison and Minnelusa aquifers. In general, hydraulic head declined during the dry-period budget (October 1, 1987, to March 31, 1993, 5.5 years) resulting in a decrease in storage. Hydraulic head rose during the wet-period budget (April 1, 1993, to September 30, 1997, 4.5 years) resulting in an increase in storage. By simultaneously balancing three water budgets, initial estimates of recharge, discharge, change in storage, and hydraulic properties were refined.

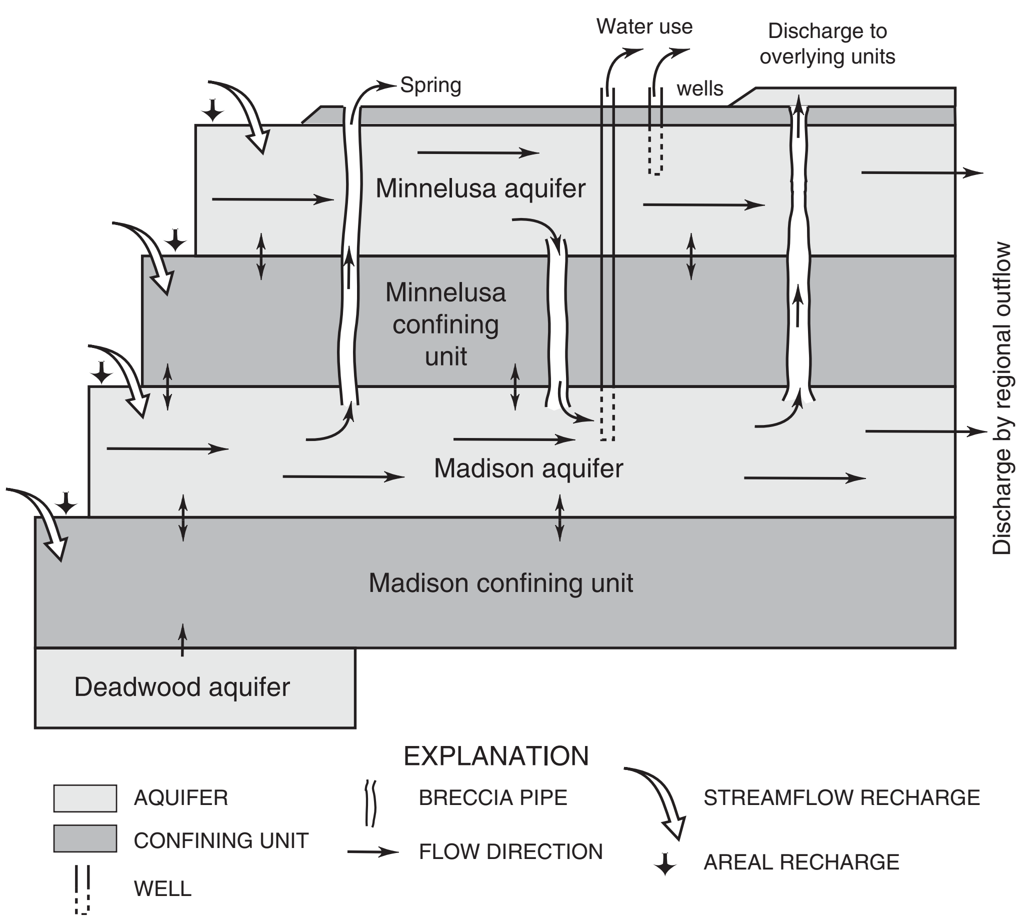


Figure 27. Conceptual model of water budget.

Water-budget results are summarized in tables 7 through 9. The equation for balancing the water budgets is the sum of inflows minus the sum of outflows, which is equal to the change in storage or:

$$\Delta S = (SD + SR + AR) - (SF + WU + OO + OE) \quad (2)$$

where

ΔS = change in ground-water storage;
SD = seepage from Deadwood aquifer;
SR = streamflow recharge;
AR = areal recharge;
SF = springflow;
WU = water use;
OO = outflow to overlying units; and
OE = outflow across eastern boundary.

Independent estimates of changes in storage for each of the three budgets could be made because there were two distinct periods of storage change during the 10-year period. Change in ground-water storage was computed by estimating the rise or fall in the potentiometric surface and multiplying this difference by area and specific yield for unconfined areas or by storage coefficient for confined areas. Hydraulic-head records for continuous-record observation wells (figs. 18 and 19, pls. 1 and 2) were used to estimate change in hydraulic head over the aquifer analysis area for the two periods. Some of the wells did not exist in October 1987, so water levels for part of the 10-year period were estimated based on comparisons to other wells with continuous records. Hydraulic head was interpolated between wells and extrapolated out to the boundaries of the aquifer analysis area. Most of the wells are located near the unconfined areas, and therefore gave a fairly accurate estimate of hydraulic head in the unconfined area. Potential error in the estimate of hydraulic-head change was estimated to be 15 percent.

Specific yield in the unconfined area is about three orders of magnitude larger than storage coefficient in the confined areas. Thus, change in storage in the confined areas is negligible by comparison, and the accuracy of hydraulic-head change in the confined areas is of little concern. The estimated hydraulic-head change over the aquifer analysis area was discretized into square cells 100 ft on a side to accurately quantify the change in storage in all areas.

Changes in hydraulic gradient across the eastern boundary were assumed to be small; therefore, regional outflow across the eastern boundary was assumed constant in all three budgets. Inflow from and outflow to adjacent hydrogeologic units also were assumed to be constant. Seepage from the Deadwood aquifer was estimated based on the approximate hydraulic head difference between the Madison and Deadwood aquifers and an estimated vertical hydraulic conductivity. A smaller amount was assumed to seep upward from the Minnelusa aquifer into overlying units.

Table 7. Average water budget for WY88-97 (full 10-year budget)

[Flow rate in cubic feet per second (ft^3/s); Mdsn, Madison hydrogeologic unit; Mnls, Minnelusa hydrogeologic unit]

Component		Mdsn		Mnls		Total		Relative confidence in estimates
		Flow rate (ft^3/s)	Percent of Mdsn budget	Flow rate (ft^3/s)	Percent of Mnls budget	Flow rate (ft^3/s)	Percent of total budget	
Inflow	Streamflow recharge	38.8	63.4	6.5	32.2	45.3	61.4	High
	Areal recharge	16.1	26.3	6.1	30.2	22.2	30.1	Medium
	Seepage from Deadwood aquifer	6.3	10.3	0	0	6.3	8.5	Low
Outflow	Springflow	-30.8	-50.3	0.0	0.0	-30.8	41.7	High
	Water use	-6.7	10.9	-3.4	16.8	-10.1	13.7	High
	Leakage to units overlying Mnls	0	0	-2.0	9.9	-2.0	2.7	Low
	Regional outflow	-11.0	18.0	-11.2	55.4	-22.2	30.1	Medium
Leakage between Madison and Minnelusa hydrogeologic units		-7.6	12.4	7.6	37.6	0	0	Medium
Change in storage (as a flow rate)		5.1	8.3	3.6	17.8	8.7	11.8	Medium

Table 8. Average water budget for October 1987 through March 1993 (dry period)

[Flow rate in cubic feet per second (ft^3/s); Mdsn, Madison hydrogeologic unit; Mnls, Minnelusa hydrogeologic unit]

Component		Mdsn		Mnls		Total		Relative confidence in estimates
		Flow rate (ft^3/s)	Percent of Mdsn budget	Flow rate (ft^3/s)	Percent of Mnls budget	Flow rate (ft^3/s)	Percent of total budget	
Inflow	Stream-loss recharge	24.1	51.9	2.8	16.7	26.9	46.7	High
	Areal recharge	8.2	17.7	3.1	18.5	11.3	19.6	Medium
	From Deadwood aquifer	6.3	13.6	0	0	6.3	10.9	Low
Outflow	Springflow	-25.3	54.5	0	0	-25.3	43.9	High
	Water use	-4.5	9.7	-3.6	21.4	-8.1	14.1	High
	Leakage to units overlying Mnls	0	0	-2.0	11.9	-2.0	3.5	Low
	Regional outflow	-11.0	23.7	-11.2	66.7	-22.2	38.5	Medium
Leakage between Madison and Minnelusa hydrogeologic units		-5.6	12.1	5.6	33.3	0	0	Medium
Change in storage (as a flow rate)		-7.8	16.8	-5.3	31.5	-13.1	22.7	Medium

Table 9. Average water budget for April 1993 to September 1997 (wet period)

[Flow rate in cubic feet per second (ft^3/s); Mdsn, Madison hydrogeologic unit; Mnls, Minnelusa hydrogeologic unit]

Component		Mdsn		Mnls		Total		Relative confidence in estimates
		Flow rate (ft^3/s)	Percent of Mdsn budget	Flow rate (ft^3/s)	Percent of Mnls budget	Flow rate (ft^3/s)	Percent of total budget	
Inflow	Stream-loss recharge	56.9	63.9	11.1	35.6	68.0	61.8	High
	Areal recharge	25.9	29.1	9.8	31.4	35.7	32.5	Medium
	From Deadwood aquifer	6.3	7.1	0	0	6.3	5.7	Low
Outflow	Springflow	-37.4	42.0	0	0	-37.4	34.0	High
	Water use	-9.3	10.4	-3.2	10.3	-12.5	11.4	High
	Leakage to units overlying Mnls	0	0	-2.0	6.4	-2.0	1.8	Low
	Regional outflow	-11.0	12.3	-11.2	35.9	-22.2	20.2	Medium
Leakage between Madison and Minnelusa hydrogeologic units		-10.3	11.6	10.3	33.0	0	0	Medium
Change in storage (as a flow rate)		21.1	23.7	14.8	47.4	35.9	32.6	Medium

Streamflow recharge, springflow, and water use were known or estimated with a relatively high level of confidence because these estimates generally were based on measured values. Estimates of areal recharge, outflow across the eastern boundary, leakage between aquifer units, inflow and outflow to adjacent hydrogeologic units, and change in storage were less certain. Simultaneously balancing the three water budgets provided an additional constraint to test estimates of the less certain water-budget components. Preliminary estimates of these less certain components were modified within plausible ranges. Transmissivities were adjusted within reasonable ranges to modify calculations of outflow across the eastern boundary. Estimates of changes in storage were modified by adjusting specific yield and estimated hydraulic-head change. Leakage between the Madison and Minnelusa hydrogeologic units was estimated by balancing the three water budgets.

Although the solution to the budget equation was not unique, the requirement that all three budgets simultaneously balance constrained the solution and improved estimates of specific yield and aquifer transmissivity near the eastern boundary. The two short-period budgets were especially sensitive to changes in specific yield, and thus, storage volume. Because the net change in storage for the 10-year period was small, the 10-year budget was more sensitive to transmissivity near the eastern boundary, which affected regional outflow.

Flow rates for selected water-budget components are listed in table 10 for each 6-month stress period. Average streamflow recharge to the Madison hydrogeologic unit was about 2.4 times that of areal recharge, while streamflow recharge to the Minnelusa hydrogeologic unit was about equal to areal recharge (table 7). This contrast is primarily due to the larger loss thresholds on the Madison outcrop and because streams lose to the Madison outcrop first, which often leaves little or no flow to cross the Minnelusa outcrop.

The Minnelusa hydrogeologic unit received about 7.6 ft^3/s or 38 percent of its total inflow (table 7) in the form of net leakage from the Madison hydrogeologic unit. The mechanisms that produce leakage between the Madison and Minnelusa aquifers are probably similar to those responsible for artesian springflow from the Madison aquifer. Therefore, the estimated net leakage from the Madison to the Minnelusa hydrogeologic unit of about 7.6 ft^3/s is plausible in comparison to the average flow from all Madison aquifer springs passing through the Minnelusa hydrogeologic unit, which was about 31 ft^3/s (tables 7 and 10). In general, hydraulic head was higher at the end of the 10 years than at the beginning. The largest discharge component from the Minnelusa hydrogeologic unit was regional outflow, whereas the largest discharge from the Madison hydrogeologic unit was springflow.

In addition to the 10-year water budget, two additional water budgets representing the first and second parts of the 10-year period were useful for refining the estimates of less certain water budget components (tables 8 and 9). The three largest total-budget components for the 10-year period (table 7) were streamflow recharge, springflow, and areal recharge. The largest total-budget components for the dry period (table 8) were streamflow recharge, springflow, and regional outflow, whereas the largest components for the wet period (table 9) were streamflow recharge, springflow, and change in storage. The total recharge rate during the wet period was about 110 ft^3/s or about 2.4 times the recharge rate during the dry period, which was 45 ft^3/s.

Seepage from Deadwood Aquifer

Underlying the Madison hydrogeologic unit are older rocks that potentially provide an additional source of recharge to the Madison hydrogeologic unit. These rocks primarily consist of the Cambrian-age Deadwood Formation, which contains a sandstone aquifer, and underlying Precambrian-age igneous and metamorphic rocks. Although the lower part of the Madison hydrogeologic unit generally is a confining unit, fractured and weathered areas near the outcrop provide paths for movement of water from these older rocks into the Madison hydrogeologic unit.

The seepage rate from the Deadwood aquifer into the Madison aquifer was estimated as 6.3 ft^3/s within the aquifer analysis area using Darcy's Law. Darcy's Law is given by the equation:

$$Q = K_v A \frac{\Delta H}{\Delta L} \tag{3}$$

where

Q = flow rate [V/T];
K_v = vertical hydraulic conductivity for the Madison confining unit [L/T];
A = area where upward leakage takes place [L^2];
ΔH = hydraulic-head difference between the Deadwood and Madison aquifers [L]; and
ΔL = vertical distance that corresponds to the hydraulic head difference [L].

Table 10. Selected water-budget components for 6-month stress periods

[Flow rates in cubic feet per second (ft^3/s)]

Stress period	Dates	Madison hydrogeologic unit				Minnelusa hydrogeologic unit		
		Streamflow recharge	Areal recharge	Springflow	Water use	Streamflow recharge	Areal recharge	Water use
Dry period								
Winter 1988 (W-88)	October 1, 1987 - March 31, 1988	15.6	0.9	26.6	1.9	2.0	0.2	2.9
Summer 1988 (S-88)	**April 1, 1988 - September 30, 1988**	**19.0**	**1.9**	**25.7**	**4.7**	**2.2**	**.5**	**5.7**
Winter 1989 (W-89)	October 1, 1988 - March 31, 1989	12.3	1.6	25.4	1.7	1.8	.6	1.7
Summer 1989 (S-89)	**April 1, 1989 - September 30, 1989**	**18.0**	**10.0**	**25.1**	**3.4**	**2.1**	**4.0**	**3.7**
Winter 1990 (W-90)	October 1, 1989 - March 31, 1990	16.5	3.2	25.2	1.8	2.0	1.0	2.9
Summer 1990 (S-90)	**April 1, 1990 - September 30, 1990**	**35.3**	**7.4**	**25.2**	**4.2**	**3.2**	**4.1**	**4.9**
Winter 1991 (W-91)	October 1, 1990 - March 31, 1991	15.7	1.8	24.8	2.9	1.8	.5	3.3
Summer 1991 (S-91)	**April 1, 1991 - September 30, 1991**	**54.4**	**52.5**	**25.8**	**7.2**	**8.8**	**19.5**	**4.6**
Winter 1992 (W-92)	October 1, 1991 - March 31, 1992	27.5	2.4	25.0	5.4	2.2	.8	2.6
Summer 1992 (S-92)	**April 1, 1992 - September 30, 1992**	**29.9**	**5.9**	**24.8**	**12.5**	**2.2**	**2.3**	**4.3**
Winter 1993 (W-93)	October 1, 1992 - March 31, 1993	20.6	2.2	25.0	3.7	2.3	.5	2.6
	Average for dry period	24.1	8.2	25.3	4.5	2.8	3.1	3.6
Wet period								
Summer 1993 (S-93)	**April 1, 1993 - September 30, 1993**	**69.4**	**46.0**	**26.8**	**13.7**	**14.1**	**19.9**	**4.0**
Winter 1994 (W-94)	October 1, 1993 - March 31, 1994	39.6	2.9	26.3	6.2	3.9	.5	2.1
Summer 1994 (S-94)	**April 1, 1994 - September 30, 1994**	**47.2**	**2.0**	**26.2**	**15.1**	**9.4**	**1.0**	**4.1**
Winter 1995 (W-95)	October 1, 1994- March 31, 1995	32.3	6.5	34.8	5.8	2.7	1.5	2.3
Summer 1995 (S-95)	**April 1, 1995 - September 30, 1995**	**67.4**	**72.7**	**36.8**	**11.0**	**14.9**	**25.7**	**3.7**
Winter 1996 (W-96)	October 1, 1995 - March 31, 1996	46.1	5.8	37.9	3.1	3.8	1.3	2.1
Summer 1996 (S-96)	**April 1, 1996 - September 30, 1996**	**71.5**	**31.1**	**39.1**	**11.3**	**17.5**	**11.5**	**4.0**
Winter 1997 (W-97)	October 1, 1996 - March 31, 1997	60.6	9.6	53.6	5.5	8.9	3.2	2.5
Summer 1997 (S-97)	**April 1, 1997 - September 30, 1997**	**77.8**	**56.3**	**55.0**	**12.0**	**24.4**	**23.3**	**4.1**
	Average for wet period	56.9	25.9	37.4	9.3	11.1	9.8	3.2
	Overall average	38.8	16.1	30.8	6.7	6.5	6.1	3.4

The area where upward seepage is considered to take place is between the western extent of the Madison outcrop and the Inyan Kara Group outcrop (171 mi^2; fig. 4). A general range of K_v values for limestone and dolomite is about 10^{-4} to 10^{-2} ft/d (Domenico and Schwartz, 1990, p. 67). K_v is likely to be higher toward the outcrop because of uplift and fracturing and lower toward the east. East of the Inyan Kara Group outcrop, upward seepage is assumed to be minimal and is neglected. As an average for the area west of the Inyan Kara outcrop, K_v was taken as $3x10^{-4}$ ft/d. Hydraulic head in the Deadwood aquifer was estimated to be about 100 ft on average above the Madison aquifer in the seepage area. The estimated length of ΔL includes the thickness of the Madison confining unit (250 ft) plus about one-half of the Deadwood Formation thickness (100 ft) for a total of 350 ft. The Whitewood and Winnipeg Formations are absent throughout most of the aquifer analysis area.

Streamflow Recharge

Streams lose much or all of their flow into swallow holes and fractures on the Madison and Minnelusa outcrops. Swallow holes are solutional features that extend upward to the land surface and intercept stream water. The Madison outcrop receives preferential recharge because of its upstream location. Large streamflow recharge rates of more than 25 ft^3/s to the Madison aquifer (Hortness and Driscoll, 1998) may have led to extensive karst development near streamflow-loss areas. Streamflow recharge to the Minnelusa aquifer also occurs but generally in smaller quantities. A description of methods used, a summary of streamflow-recharge estimates, and details on individual loss zones follow.

Methods

Drainage areas that contribute flow to the streamflow recharge are delineated for the western part of the study area (pl. 3). Although streamflow-loss zones are shown extending across the entire Madison and Minnelusa outcrop areas, losses are generally concentrated in certain areas of the outcrops. Because the Madison aquifer is more permeable, much of the streamflow lost to the confining unit located on the western part of the outcrop probably reaches the aquifer via underground conduits. An example of this can be seen along Boxelder Creek where much of the streamflow lost to the western part of the Madison outcrop reemerges as springflow and then disappears again in the eastern part of the loss zone. Rahn and Gries (1973) determined from dye testing that Gravel, Doty, and Dome Springs (pls. 1 and 2) on the Madison outcrop along Boxelder Creek are directly connected to upstream losses. At outcrop areas of the Minnelusa hydrogeologic unit, ground water probably moves from confining unit to aquifer in a similar way because of weathering and increased permeability of the confining unit.

Streamflow recharge is calculated for 10 streams that cross the Madison and Minnelusa outcrops in the study area. Daily streamflow records are available for the larger streams that lose flow to the outcrops, including Battle, Spring, Rapid, Boxelder, and Elk Creeks (pl. 3 and table 11). Loss thresholds were determined by Hortness and Driscoll (1998) and were used if available, but were not determined for some of the smaller streams and were not always separated by formation (table 11). According to Hortness and Driscoll (1998), all of the flow in these streams up to a threshold is lost to outcrops of Paleozoic rocks. Measured or estimated daily streamflow up to an estimated loss threshold is assumed to recharge the Madison hydrogeologic unit, and once that threshold flow is exceeded, recharge to the Minnelusa hydrogeologic unit (not to exceed its threshold) can occur. A diagram representing streamflow losses from Elk Creek (fig. 28) illustrates that during low-flow periods, such as WY88-90, the Minnelusa hydrogeologic unit may receive little or no streamflow recharge.

In some cases, gages were not located at the western contact of the Madison hydrogeologic unit, but were farther upstream. For these streams, the gaged flow was adjusted proportional to drainage area (table 11). Some streams were continuously gaged for only part of the 10-year period, and a few smaller basins were not gaged at all. Missing data for continuously gaged streams were synthesized using linear regression of the measured streamflow against streamflow in a nearby basin. The equation of the regression line was used to estimate missing parts of the hydrograph. Streamflow records for ungaged streams were estimated by proportioning flows of nearby gaged streams relative to drainage area size.

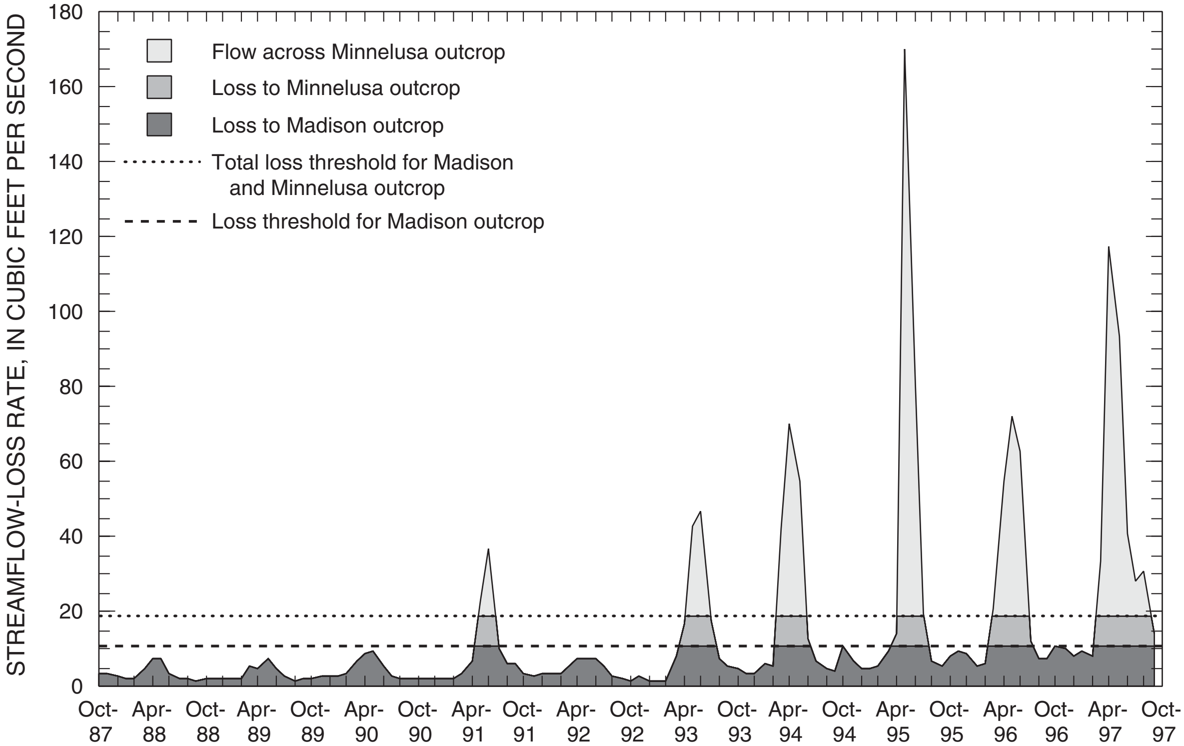


Figure 28. Hydrograph of monthly mean streamflow losses from Elk Creek to the Madison and Minnelusa outcrops. Loss thresholds estimated by Hortness and Driscoll (1998).

In the study area, the Minnelusa hydrogeologic unit contains "unsaturated areas," as previously defined (see "Concepts of the Ground-Water Flow System" section), across about 73 percent of the outcrop. Plate 2 shows the extent of saturated and unsaturated areas in relation to the outcrops. About 36 percent of the total length of the losing streams crossing the Minnelusa outcrop is on these unsaturated areas. Much of the streamflow loss to the unsaturated areas of the Minnelusa outcrop probably reaches the Madison aquifer because of fractures and breccias that extend through the outcrop. According to Gott and others (1974), solution breccias in the Minnelusa Formation permit rapid infiltration of ground water that probably recharges the Madison aquifer. However, because of easterly dipping beds, the movement of infiltrating water probably has a horizontal component and some will reach the Minnelusa hydrogeologic unit's water table. Because of the lack of data to quantify the proportion, 50 percent of this water infiltrating the unsaturated area on the Minnelusa outcrop was assumed to recharge the Madison aquifer and 50 percent to recharge the Minnelusa aquifer. Streamflow losses on the Minnelusa outcrop were assumed to be evenly distributed across the length of the stream reach. Therefore, the fraction of redistributed recharge from the Minnelusa outcrop is equal to the ratio of the unsaturated-area reach length to the total Minnelusa outcrop reach length multiplied by one-half of the loss across the Minnelusa outcrop:

$$\text{Redistributed recharge} = \left(\frac{\text{Dry reach length}}{\text{Total outcrop reach length}}\right) \times \left(\frac{\text{Minnelusa outcrop loss}}{2}\right). \qquad (4)$$

Table 11. Selected data for drainage areas, streamflow, and gaging stations

[WY, water year; Mdsn, Madison hydrogeologic unit; Mnls, Minnelusa hydrogeologic unit; mi^2, square miles; ft^3/s, cubic feet per second; mi, miles; --, undetermined; NA, not applicable]

Drainage areas contributing to streamflow-loss zones						Streamflow-loss-zone data: Loss-zone stream reach length		Streamflow-loss-zone data: Approximate loss threshold		Streamflow-gaging stations used in calculation of streamflow loss or springflow[1]			
Drainage area identifier (pl. 3)	Name	Area (mi^2)	Average annual precipitation[2] 1961-98 (inches)	Average annual streamflow, WY88-97 (ft^3/s)	Average annual yield, WY88-97 (inches)	Mdsn (mi)	Mnls (mi)	Mdsn (ft^3/s)	Mnls (ft^3/s)	Gage-site number[3] (pl. 3)	Station identification number	Station name	Drainage area[4] (mi^2)
A	Battle Creek	66.0	21.1	13.1	2.70	1.2	--	[5]14	[6]0	14	06404000	Battle Creek near Keystone	58.4
B	Deadman Gulch	1.49	19.9	[7].2	[7]1.82	1.5	4.8	[7]2	[7]4				
C	Rockerville Gulch	3.68	19.5	[7].7	[7]2.58	.8	4.7	[7]2	[7]4				
D	Spring Creek	163	20.6	25.3	2.11	2.3	2.4	[8]21	[5]3.5	24	06407500	Spring Creek near Keystone	163
E	Victoria Creek	11.1	19.8	[7]1.6	[7]1.96	.6	NA	[8]2.1	NA				
F	Rapid Creek	355	22.4	59.0	2.26	2.2	1.8	8	2	30	06412200	Rapid Creek above Victoria Creek	355
G	Unnamed tributary	.92	20.1	[7].1	[7]1.48	1.5	5.4	[7]2	[7]4				
H	Boxelder Creek	103	23.1	25.1	3.31	7.9	4.6	[7]30	[7]16	34	06422500	Boxelder Creek near Nemo	96
I	Little Elk Creek	11.6	23.4	[7]1.9	[7]2.22	.1	1.5	[8].7	[8]2.6				
J	Elk Creek	35.1	26.1	14.0	5.42	8.5	2.4	[8]11	[8]8	39	06424000	Elk Creek near Roubaix	21.5
	NA	--	--	--	--	NA	NA	NA	NA	43	44182310324100	Elk Creek below trib from north, near Tilford	--
	NA	--	--	--	--	NA	NA	NA	NA	44	441701103282700	Elk Creek below Madison outcrop	--
	NA	--	--	--	--	NA	NA	NA	NA	45	441614103253300	Elk Creek at Minnekahta outcrop	--

Table 11. Selected data for drainage areas, streamflow, and gaging stations–Continued

[WY, water year; Mdsn, Madison hydrogeologic unit; Mnls, Minnelusa hydrogeologic unit; mi^2, square miles; ft^3/s, cubic feet per second; mi, miles; --, undetermined; NA, not applicable]

Drainage areas contributing to streamflow-loss zones						Streamflow-loss-zone data: Loss-zone stream reach length		Streamflow-loss-zone data: Approximate loss threshold		Streamflow-gaging stations used in calculation of streamflow loss or springflow[1]			
Drainage area identifier (pl. 3)	Name	Area (mi^2)	Average annual precipitation[2] 1961-98 (inches)	Average annual streamflow, WY88-97 (ft^3/s)	Average annual yield, WY88-97 (inches)	Mdsn (mi)	Mnls (mi)	Mdsn (ft^3/s)	Mnls (ft^3/s)	Gage-site number[3] (pl. 3)	Station identification number	Station name	Drainage area[4] (mi^2)
	NA	--	--	--	--	NA	NA	NA	NA	90	[9]06413650	Lime Creek at mouth, at Rapid City[9]	--
	NA	--	--	--	--	NA	NA	NA	NA	91	[9]06413800	Deadwood Avenue drain at mouth, at Rapid City[9]	--

[1]Flow rates from U.S. Geological Survey (1989-98).

[2]Calculated from precipitation distribution for 1961-98 (Driscoll, Hamade, and Kenner, 2000).

[3]Site number corresponds to site number in Hortness and Driscoll, 1998.

[4]Drainage area from U.S. Geological Survey (1967-97) except for site number 14, which is a corrected value.

[5]Carter, Driscoll, and Hamade, 2001.

[6]Generally, there is a net gain in streamflow in this reach. During very dry periods, some streamflow loss to the Minnelusa hydrogeologic unit could be possible, however, the impact on the water budget was assumed to be negligible.

[7]Estimated.

[8]Hortness and Driscoll, 1998.

[9]Gaging station used to estimate springflow.