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Hydrology of the Black Hills Area, South Dakota

By Daniel G. Driscoll, Janet M. Carter, Joyce E. Williamson, and Larry D. Putnam

Water-Resources Investigations Report 02-4094

Prepared in cooperation with the
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CONVERSION FACTORS

Multiply	By	To obtain
acre	4,047	square meter
acre	0.4047	hectare
acre-foot (acre-ft)	1,233	cubic meter
acre-foot (acre-ft)	0.001233	cubic hectometer
acre-foot per year (acre-ft/yr)	1,233	cubic meter per year
acre-foot per year (acre-ft/yr)	0.001233	cubic hectometer per year
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 squared per day (ft^2/d)	0.09290	meter squared per day
inch	2.54	centimeter
inch	25.4	millimeter
inch per year (in/yr)	25.4	millimeter per year
mile (mi)	1.609	kilometer
square foot (ft^2)	929.0	square centimeter
square mile (mi^2)	259.0	hectare
square mile (mi^2)	2.590	square kilometer

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

$$°F = (1.8 \times °C) + 32$$

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

$$°C = (°F - 32) / 1.8$$

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, 1998, is called the “1998 water year.”

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ABSTRACT

The Black Hills Hydrology Study was initiated in 1990 to assess the quantity, quality, and distribution of surface water and ground water in the Black Hills area of South Dakota. This report summarizes the hydrology of the Black Hills area and the results of this long-term study.

The Black Hills area of South Dakota and Wyoming is an important recharge area for several regional, bedrock aquifer systems and various local aquifers; thus, the study focused on describing the hydrologic significance of selected bedrock aquifers. The major aquifers in the Black Hills area are the Deadwood, Madison, Minnelusa, Minnekahta, and Inyan Kara aquifers. The highest priority was placed on the Madison and Minnelusa aquifers, which are used extensively and heavily influence the surface-water resources of the area.

Within this report, the hydrogeologic framework of the area, including climate, geology, ground water, and surface water, is discussed. Hydrologic processes and characteristics for ground water and surface water are presented. For ground water, water-level trends and comparisons and water-quality characteristics are presented. For surface water, streamflow characteristics, responses to precipitation, annual yields and yield efficiencies, and water-quality characteristics are presented. Hydrologic budgets are presented for ground water, surface water, and the combined ground-water/surface-water system. A summary of study findings regarding the complex flow systems within the Madison and Minnelusa aquifers also is presented.

INTRODUCTION

The Black Hills area is an important resource center that provides an economic base for western South Dakota through tourism, agriculture, the timber industry, and mineral resources. In addition, water originating from the area is used for municipal, industrial, agricultural, and recreational purposes throughout much of western South Dakota. The Black Hills area also is an important recharge area for aquifers in the northern Great Plains.

Population growth, resource development, and periodic droughts have the potential to affect the quantity, quality, and availability of water within the Black Hills area. Because of this concern, the Black Hills Hydrology Study was initiated in 1990 to assess the quantity, quality, and distribution of surface water and ground water in the Black Hills area of South Dakota (Driscoll, 1992). This long-term study has been a cooperative effort between the U.S. Geological Survey (USGS), the South Dakota Department of Environment and Natural Resources, and the West Dakota Water Development District, which represents various local and county cooperators.

The specific objectives of the Black Hills Hydrology Study included:

1. Inventorying and describing precipitation amounts, streamflow rates, ground-water levels of selected aquifer units, and selected water-quality characteristics for the Black Hills area.

2. Developing hydrologic budgets to define relations among precipitation, streamflow, and aquifer response for selected Black Hills watersheds.
3. Describing the significance of the bedrock aquifers in the Black Hills area hydrologic system, with an emphasis on the Madison and Minnelusa aquifers, through determination of:
   a. aquifer properties (depth, thickness, structure, storage coefficient, hydraulic conductivity, etc.);
   b. the hydraulic connection between the aquifers;
   c. the source aquifer(s) of springs;
   d. recharge and discharge rates, and gross volumetric budgets; and
   e. regional flow paths.
4. Developing conceptual models of the hydrogeologic system for the Black Hills area.

Purpose and Scope

The purpose of this report is to summarize the hydrology of the Black Hills area and present major findings pertinent to the objectives of the Black Hills Hydrology Study. The information summarized in this report has been presented in more detail in previous reports prepared as part of the study. Because the Black Hills area of South Dakota and Wyoming is an important recharge area for several regional, bedrock aquifers and various local aquifers, the study concentrated on describing the hydrogeology and hydrologic significance of selected bedrock aquifers. The highest priority was placed on the Madison and Minnelusa aquifers because: (1) these aquifers are heavily used and could be developed further; (2) these aquifers are connected to surface-water resources through streamflow loss zones and large springs; and (3) hydraulic connection between these aquifers is extremely variable. The Deadwood and Minnekahta aquifers had a lower priority because they are used less and have less influence on the hydrologic system. The fractured Precambrian rocks, Inyan Kara Group, and various local aquifers, including minor bedrock aquifers and unconsolidated aquifers, had the lowest priorities because: (1) the Precambrian and local aquifers are not regional aquifers with regional flowpaths; and (2) the Inyan Kara Group is not used as extensively in the Black Hills area as the other priority units.

Hydrologic analyses within this report generally are by water year, which represents the period from October 1 through September 30. Discussions of timeframes refer to water years, rather than calendar years, unless specifically noted otherwise.

Description of Study Area

The study area for the Black Hills Hydrology Study consists of the topographically defined Black Hills and adjacent areas located in western South Dakota (fig. 1). Outcrops of the Madison Limestone and Minnelusa Formation, as well as the generalized outer extent of the Inyan Kara Group, which approximates the outer extent of the Black Hills area, also are shown in figure 1. The Black Hills are situated between the Cheyenne and Belle Fourche Rivers. The Belle Fourche River is the largest tributary to the Cheyenne River. The study area includes most of the larger communities in western South Dakota and contains about one-fifth of the State's population.

The Black Hills uplift formed as an elongated dome about 60 to 65 million years ago during the Laramide orogeny (Darton and Paige, 1925). The dome trends north-northwest and is about 120 mi long and 60 mi wide. Land-surface altitudes range from 7,242 ft above sea level at Harney Peak to about 3,000 ft in the adjacent plains. Most of the higher altitudes are heavily forested with ponderosa pine, which is the primary product of an active timber industry. White spruce, quaking aspen, paper birch, and other native trees and shrubs are found in cooler, wetter areas (Orr, 1959). The lower altitude areas surrounding the Black Hills primarily are urban, suburban, and agricultural. Numerous deciduous species such as cottonwood, ash, elm, oak, and willow are common along streams in the lower altitudes. Rangeland, hayland, and winter wheat farming are the principal agricultural uses for dryland areas. Alfalfa, corn, and vegetables are produced in bottom lands and in irrigated areas. Various other crops, primarily for cattle fodder, are produced in both dryland areas and in bottom lands.

Beginning in the 1870's, the Black Hills have been explored and mined for many commodities including gold, silver, tin, tungsten, mica, feldspar, bentonite, beryl, lead, zinc, uranium, lithium, sand, gravel, and oil (U.S. Department of Interior, 1967). Mines within the study area have used various techniques including placer mining, underground mining, and open-pit mining.

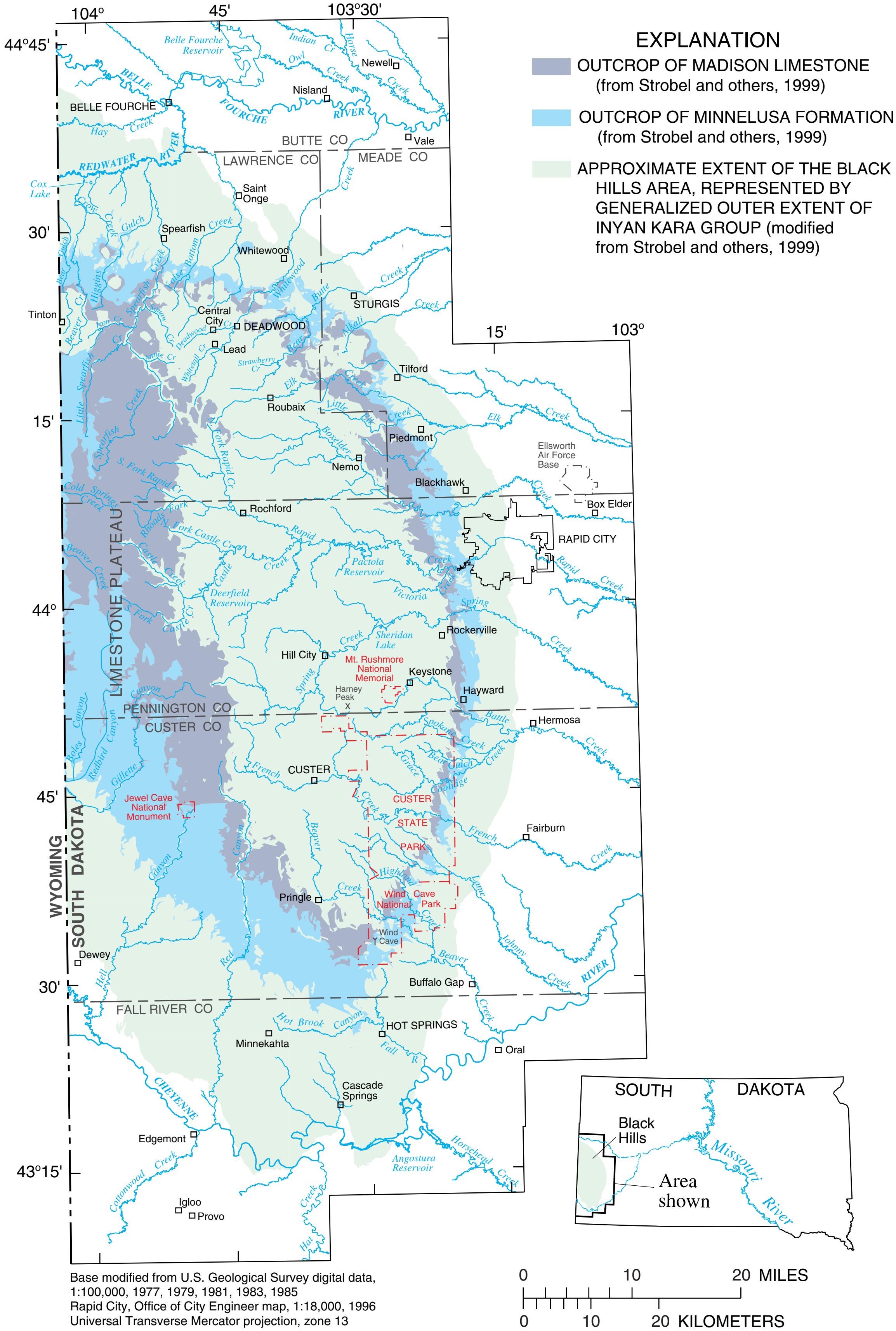


Figure 1. Area of investigation for the Black Hills Hydrology Study.

Acknowledgments

The authors acknowledge the efforts of the West Dakota Water Development District for helping to develop and support the Black Hills Hydrology Study. West Dakota's coordination of various local and county cooperators has been a key element in making this study possible. The authors also recognize the numerous local and county cooperators represented by West Dakota, as well as the numerous private citizens who have helped provide guidance and support for the Black Hills Hydrology Study. The South Dakota Department of Environment and Natural Resources has provided support and extensive technical assistance to the study. In addition, the authors acknowledge the input and technical assistance from many faculty and students at the South Dakota School of Mines and Technology.

HYDROGEOLOGIC FRAMEWORK

The Black Hills are located within the Great Plains physiographic province in western South Dakota and eastern Wyoming (fig. 2). The Black Hills strongly influence the hydrology of western South Dakota and northeastern Wyoming. Many streams in western South Dakota originate in the Black Hills, and major bedrock aquifers are recharged along outcrop areas in the Black Hills. Ground and surface water interact extensively in the Black Hills, and both streamflow and aquifer recharge are influenced by climatic conditions. Overviews of the climate, geology, ground water, and surface water are provided in the following sections.

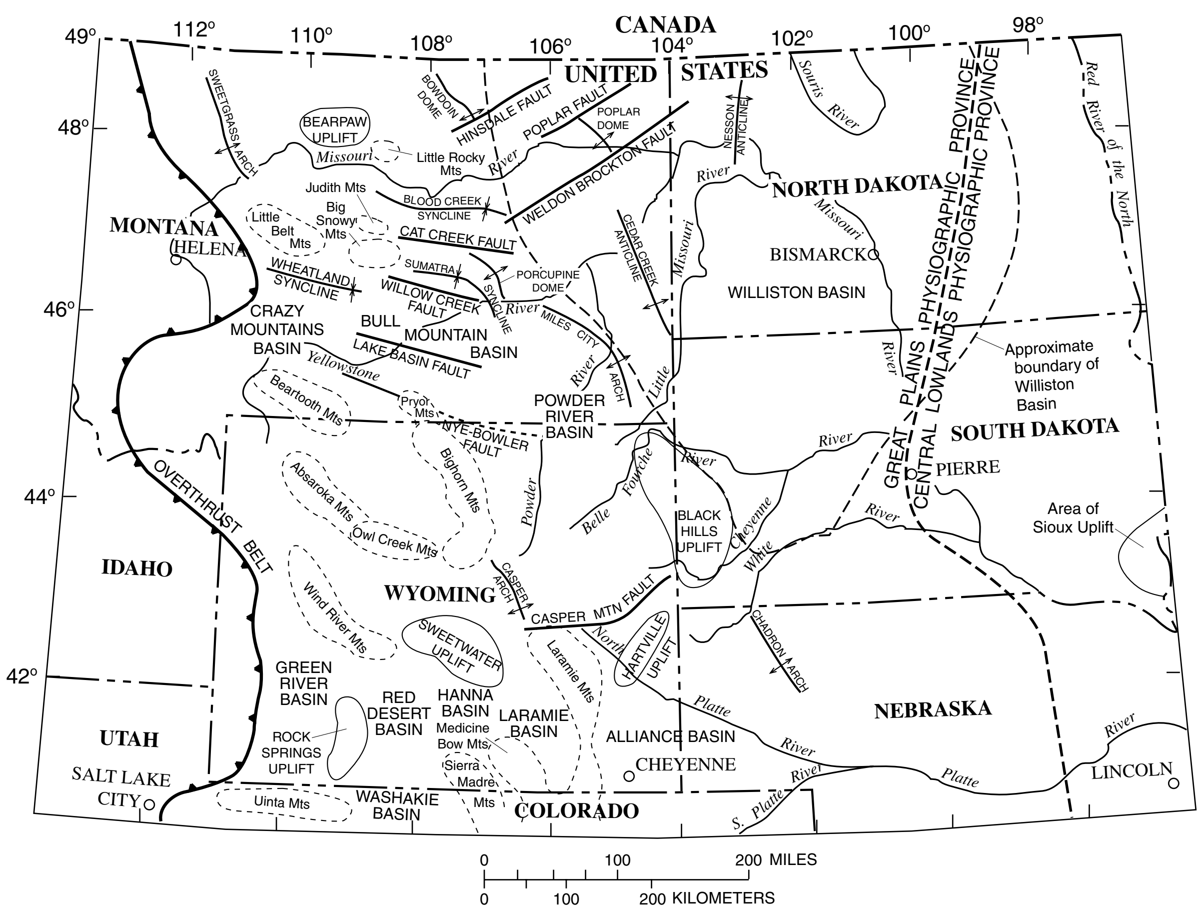


Figure 2. Present-day structural and physiographic features in the northern Great Plains area (modified from Peterson, 1981, and Busby and others, 1995).

Climatic Framework

The overall climate of the Black Hills area is continental, with generally low precipitation amounts, hot summers, cold winters, and extreme variations in both precipitation and temperatures (Johnson, 1933). Local climatic conditions are affected by topography, with generally lower temperatures and higher precipitation at the higher altitudes. The average annual temperature is 43.9°F (U.S. Department of Commerce, 1999) and ranges from 48.7°F at Hot Springs to approximately 37°F near Deerfield Reservoir.

Precipitation data sets used for this study generally were taken from Driscoll, Hamade, and Kenner (2000), who summarized available precipitation data (1931-98) for the Black Hills area. These investigators compiled monthly precipitation records for 52 long-term precipitation gages operated by National Oceanic and Atmospheric Administration (1998) and 42 short-term precipitation gages operated by the USGS. These data sets are available on the World Wide Web at http://sd.water.usgs.gov/projects/bhhs/precip/home.htm. A geographic information system (GIS) was used by Driscoll, Hamade, and Kenner (2000) to generate spatial distributions of monthly precipitation data for 1,000-by-1,000-meter grid cells for the study area; an example is shown in figure 3. Monthly distributions were composited to produce annual distributions for counties within the study area and for drainage areas of selected streamflow-gaging stations; these data sets were presented by Driscoll and Carter (2001). The precipitation distributions were used extensively for various applications including evaluating responses of ground-water levels and streamflow to precipitation, estimating precipitation recharge for bedrock aquifers, and developing long-term hydrologic budgets.

Spatial precipitation patterns in the Black Hills area are highly influenced by orography, as shown by an isohyetal map (fig. 4) for 1950-98, which is the period commonly used for hydrologic budgets presented in this report. Areas of relatively low precipitation occur in the low altitudes around the periphery of the Black Hills. Most areas with altitudes exceeding 6,000 ft above sea level have average annual precipitation in excess of 19 inches, with the largest amounts occurring in the northern Black Hills near Lead, where the average annual precipitation (1950-98) exceeds 28 inches. Orographic effects also are apparent in the high-altitude areas near Harney Peak.

Local conditions also are affected by regional climatic patterns, with the northern Black Hills influenced primarily by moist air currents from the northwest, and the southern Black Hills influenced primarily by drier air currents from the south-southeast. As a result, annual precipitation averages about 16 to 17 inches for most of Fall River County (fig. 4) and is much less than parts of Lawrence and Meade Counties that have comparable altitudes. Boxplots showing the distribution of annual precipitation for the study area and for counties within the study area during 1931-98 are presented in figure 5. For the study area, the long-term average of 18.61 inches is slightly larger than the median (50th percentile) of 17.96 inches. The 90th percentile indicates that annual precipitation over the study area is less than about 23.70 inches 90 percent of the time. Annual precipitation for both Butte and Fall River Counties is less than the long-term average for the study area about 75 percent of the time.

The largest precipitation amounts typically occur during May and June, and the smallest amounts typically occur during November through February (fig. 6). The most variable month is May, during which precipitation has ranged from a minimum of about 0.4 inch to a maximum of 8.5 inches. The seasonal distribution of precipitation is fairly uniform throughout the study area; however, Lawrence County receives slightly larger proportions of its annual precipitation during winter months than the other counties (fig. 7).

Long-term (1931-98) trends in precipitation (fig. 8) are an important consideration for hydrologic analysis for the Black Hills area. Figure 8A shows that annual precipitation for the study area averages 18.61 inches and has ranged from 10.22 inches in 1936 to 27.39 inches in 1995. Figure 8B shows that the associated departures (from the average) have ranged from a deficit (-) of 8.39 inches to a surplus (+) of 8.78 inches, respectively. The cumulative trends are readily apparent from figure 8C, with the most pronounced trends identified by the longest and steepest line segments. Sustained periods of generally deficit precipitation occurred during 1931-40 and 1948-61. Sustained periods of generally surplus precipitation occurred during 1941-47, 1962-68, and 1993-98. The middle to late 1990's stand out as the wettest period since 1931.

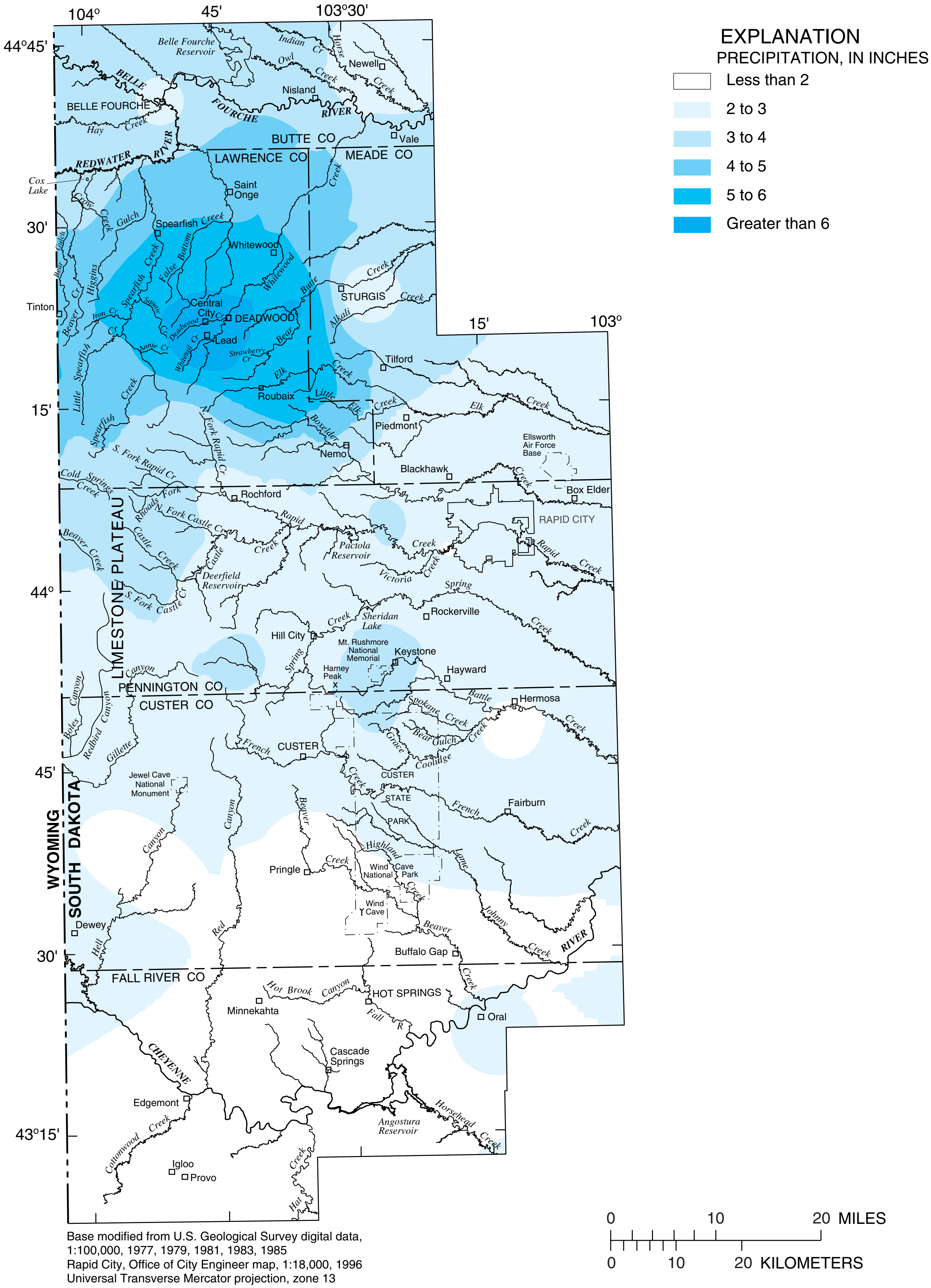


Figure 3. Monthly precipitation distribution for October 1995 (from Driscoll, Hamade, and Kenner, 2000).

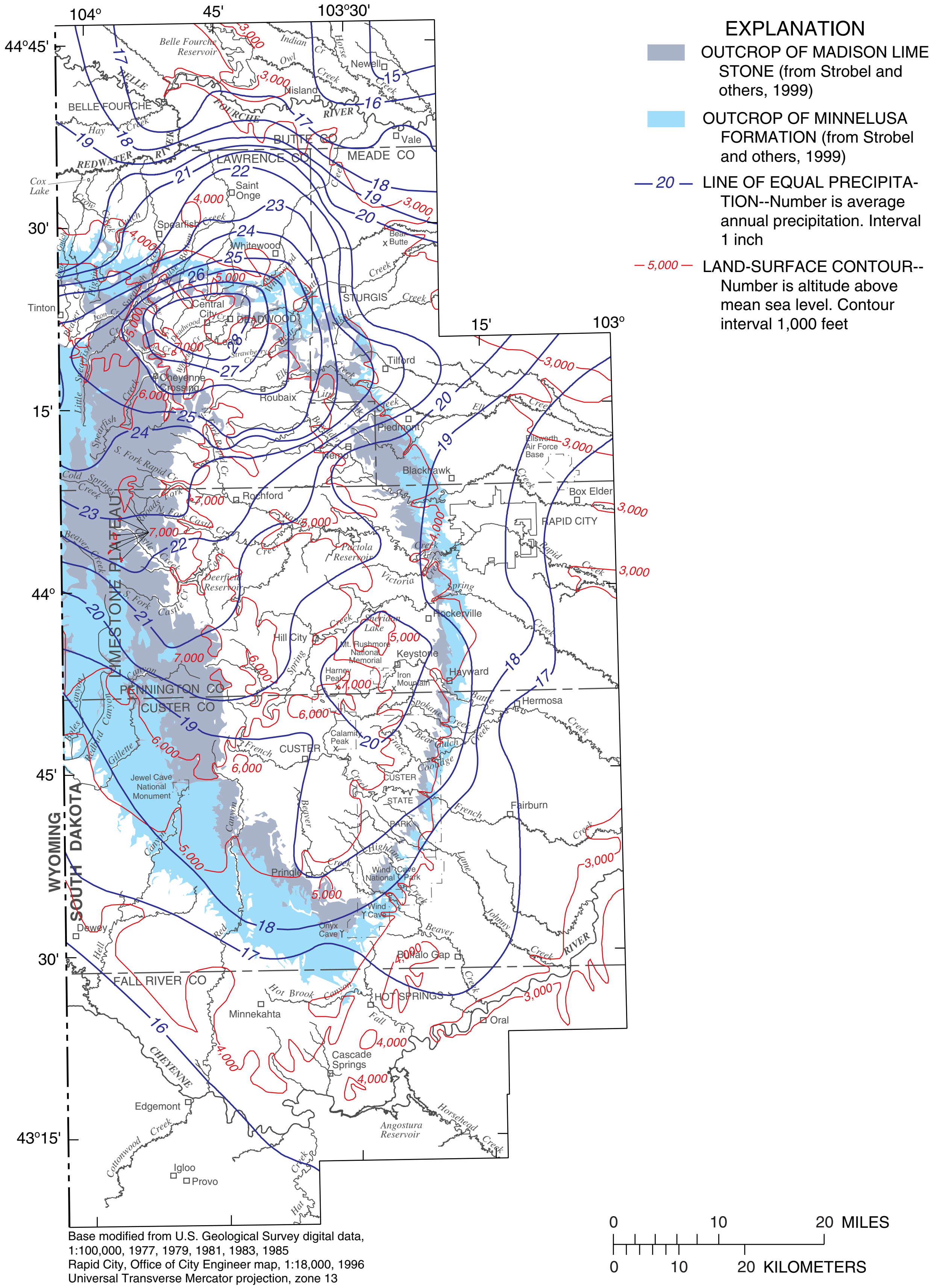


Figure 4. Isohyetal map showing distribution of average annual precipitation for Black Hills area, water years 1950-98 (from Carter, Driscoll, and Hamade, 2001).

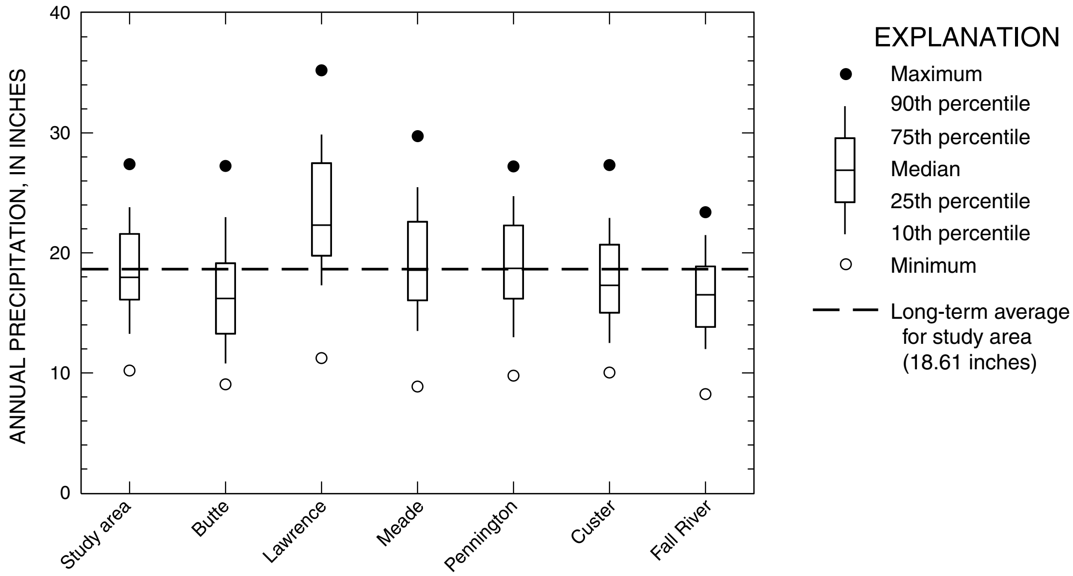


Figure 5. Distribution of annual precipitation for the study area and counties within the study area, water years 1931-98 (modified from Driscoll and Carter, 2001).

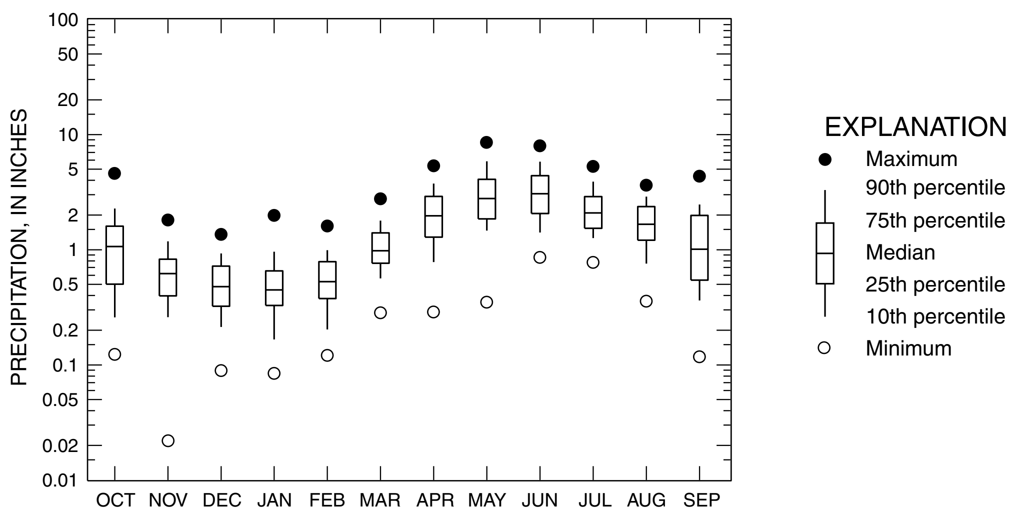


Figure 6. Distribution of monthly precipitation for the study area, water years 1931-98 (from Driscoll, Hamade, and Kenner, 2000).

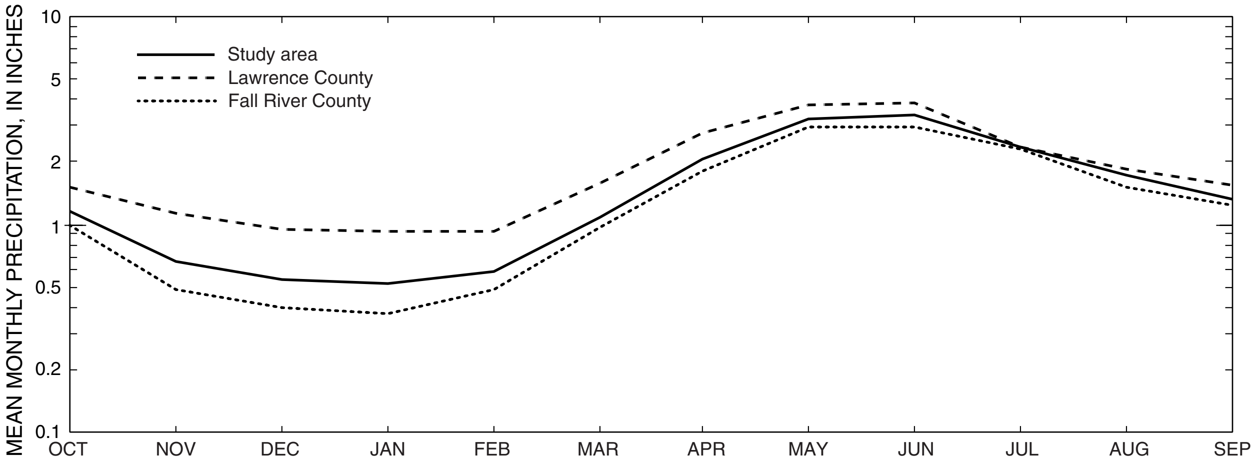


Figure 7. Mean monthly precipitation for study area and selected counties, water years 1931-98 (from Driscoll and Carter, 2001).

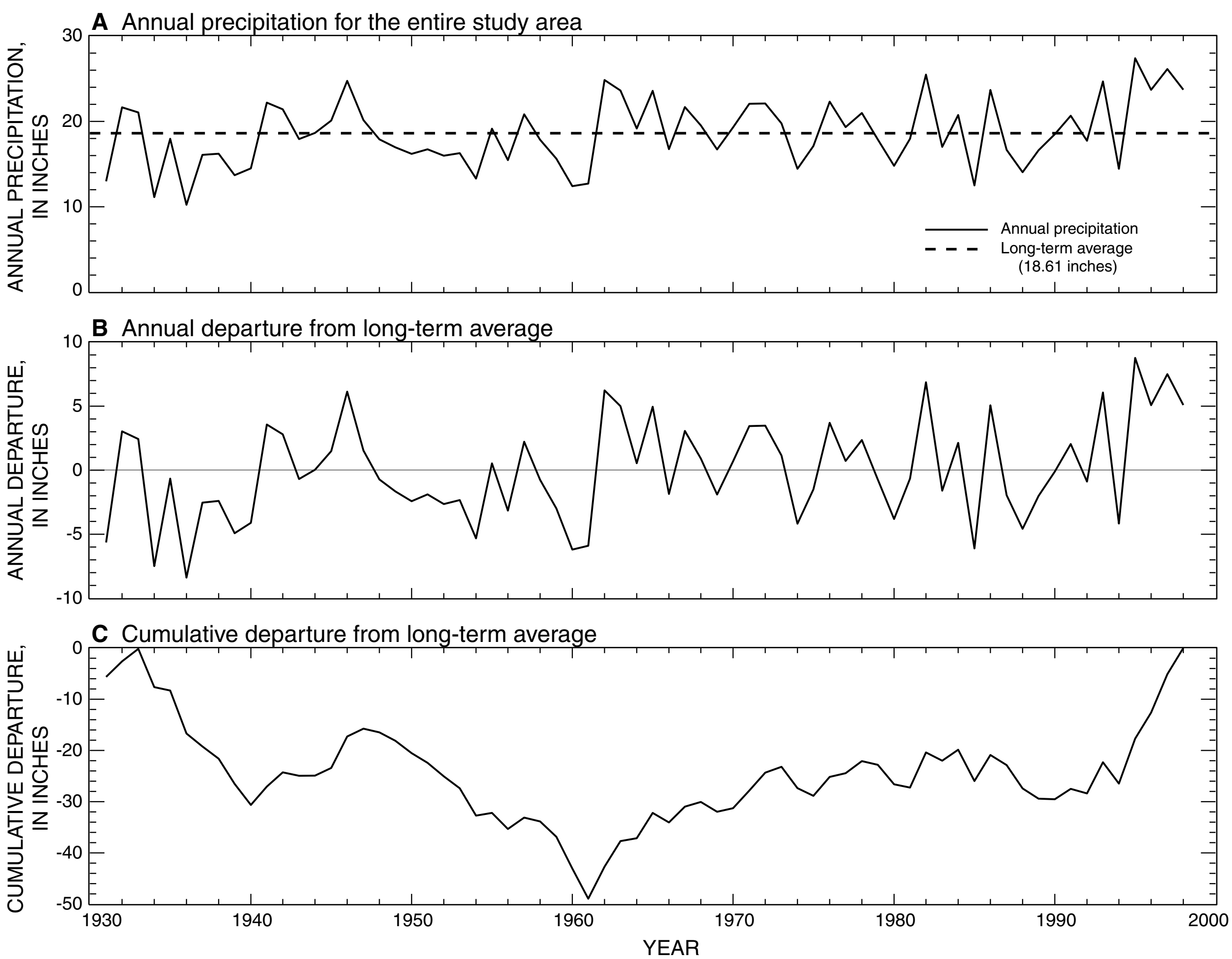


Figure 8. Long-term trends in precipitation for the Black Hills area, water years 1931-98 (from Driscoll, Hamade, and Kenner, 2000).

The long-term precipitation trends are especially important because of potential for bias in analysis and interpretation of available hydrologic data sets, which are much more abundant for the recent wet years. Water-level records are available for 71 observation wells in the Black Hills area for 1998, compared with five wells for 1965 (Driscoll, Bradford, and Moran, 2000). Miller and Driscoll (1998) reported streamflow records for 65 gages for 1993, compared with 30 gages for 1960. Thus, the potential for bias is an important consideration in analysis of hydrologic data sets for the Black Hills area.

Average annual potential evaporation generally exceeds average annual precipitation throughout the study area. Thus, evapotranspiration generally is limited by precipitation amounts and availability of soil moisture. Average pan evaporation for April through October is about 30 inches at Pactola Reservoir and about 50 inches at Oral (U.S. Department of Commerce, 1999).

Geologic Framework

The stratigraphic and structural features in the Black Hills area are complex. Many of the geologic formations, such as the Deadwood Formation, Madison Limestone, Minnelusa Formation, Minnekahta Limestone, and Inyan Kara Group, in the Black Hills (fig. 9) are regionally extensive. Several formations thin or pinch out in southern and eastern South Dakota. To better understand the stratigraphic and structural settings in the Black Hills, an overview of the regional geologic setting is provided first and is followed by an overview of the local geologic setting.

Regional Geologic Setting

Parts of Montana, North Dakota, South Dakota, and Wyoming are included in the Northern Great Plains area. The present-day structural features (fig. 2) of the Northern Great Plains are directly related to the geologic history of the Cordilleran platform, which is a part of the stable interior of the North American Continent (Downey, 1986). The present-day structure probably was controlled by the pre-existing structural grain in the Precambrian basement and modified during the Laramide orogeny (Downey, 1984).

During Paleozoic time, the area generally was broad, flat, and covered by shallow, warm seas (Downey, 1984). Numerous disconformities during Paleozoic time indicate intermittent transgressions and regressions when seas advanced from west to east in response to tectonic activity of the Antler orogeny to the west (Sandberg and Poole, 1977). Deposits generally were beach, shallow marine, carbonate, sabkha, and evaporite units (Redden and Lisenbee, 1996).

During Cretaceous time, the area was covered by a north-south trending sea, which extended from the Gulf of Mexico to the Arctic Ocean (Downey, 1986). During Late Cretaceous time, the sea was at its widest extent, but marine deposition was interrupted by frequent east-west regressions (Anna, 1986).

Paleostructure

The Northern Great Plains area was part of the Cordilleran platform throughout most of Paleozoic time. The Williston Basin, which covers parts of North Dakota, South Dakota, southern Saskatchewan, southwestern Manitoba, and eastern Montana (fig. 10), began to take shape during Ordovician time (Carlson and Anderson, 1965). Other major Jurassic and Cretaceous (pre-Laramide) paleostructural elements (fig. 10) include the Powder River Basin, the Central Montana trough and uplift, the Cedar Creek anticline, and the Alberta shelf (Anna, 1986).

The Laramide orogeny, which affected the eastern Rocky Mountains of the United States, began during late Cretaceous time and continued in the Eocene period (Redden and Lisenbee, 1996). The Laramide orogeny was characterized by large-scale warping, deep erosion of uplifts, and deposition of orogenic sediments into basins (Tweto, 1975). Most, if not all, pre-Laramide structural features (fig. 10) were reactivated and became more prominent during the Laramide orogeny (Anna, 1986). During the Laramide orogeny, the Bighorn and Laramie Mountains, the Black Hills, and the Central Montana uplift formed, and the Williston and Powder River Basins (fig. 2) were downwarped into essentially their present configuration (Anna, 1986).

ERATHEM	SYSTEM	ABBREVIATION FOR STRATIGRAPHIC INTERVAL	STRATIGRAPHIC UNIT	THICKNESS IN FEET	DESCRIPTION
CENOZOIC	QUATERNARY & TERTIARY (?)	QTac	UNDIFFERENTIATED ALLUVIUM AND COLLUVIUM	0-50	Sand, gravel, boulder, and clay.
		Tw	WHITE RIVER GROUP	0-300	Light colored clays with sandstone channel fillings and local limestone lenses.
	TERTIARY	Tui	INTRUSIVE IGNEOUS ROCKS	--	Includes rhyolite, latite, trachyte, and phonolite.
MESOZOIC	CRETACEOUS	Kps	PIERRE SHALE	1,200-2,700	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	[1]80-300	Impure chalk and calcareous shale.
			CARLILE SHALE — Turner Sandy Member; Wall Creek Member	[1]350-750	Light-gray shale with numerous large concretions and sandy layers. Dark-gray shale
			GREENHORN FORMATION	225-380	Impure slabby limestone. Weathers buff. Dark-gray calcareous shale, with thin Orman Lake limestone at base.
			GRANEROS GROUP: BELLE FOURCHE SHALE	150-850	Gray shale with scattered limestone concretions. Clay spur bentonite at base.
			GRANEROS GROUP: MOWRY SHALE	125-230	Light-gray siliceous shale. Fish scales and thin layers of bentonite.
			GRANEROS GROUP: MUDDY SANDSTONE / NEWCASTLE SANDSTONE	0-150	Brown to light-yellow and white sandstone.
			GRANEROS GROUP: SKULL CREEK SHALE	150-270	Dark-gray to black siliceous shale.
		Kik	INYAN KARA GROUP: FALL RIVER FORMATION	10-200	Massive to thin-bedded, brown to reddish-brown sandstone.
			INYAN KARA GROUP: LAKOTA FM — Fuson Shale; Minnewaste Limestone; Chilson Member	10-190 0-25 25-485	Yellow, brown, and reddish brown massive to thinly bedded sandstone, pebble conglomerate, siltstone, and claystone. Local fine-grained limestone and coal.
	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	375-800	Red silty 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, fine-grained, purplish gray laminated limestone.
		Po	OPECHE SHALE	[1]25-150	Red shale and sandstone.
	PENNSYLVANIAN	PIPm	MINNELUSA FORMATION	[1]375-1,175	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	[1]<200-1,000	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	[1]0-235	Buff dolomite and limestone.
			WINNIPEG FORMATION	[1]0-150	Green shale with siltstone.
	CAMBRIAN	OЄd	DEADWOOD FORMATION	[1]0-500	Massive to thin-bedded buff to purple sandstone. Greenish glauconitic shale flaggy dolomite and flat-pebble limestone conglomerate. Sandstone, with conglomerate locally at the base.
PRECAMBRIAN		pЄu	UNDIFFERENTIATED IGNEOUS AND METAMORPHIC 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

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

Figure 9. Stratigraphic section for the Black Hills.

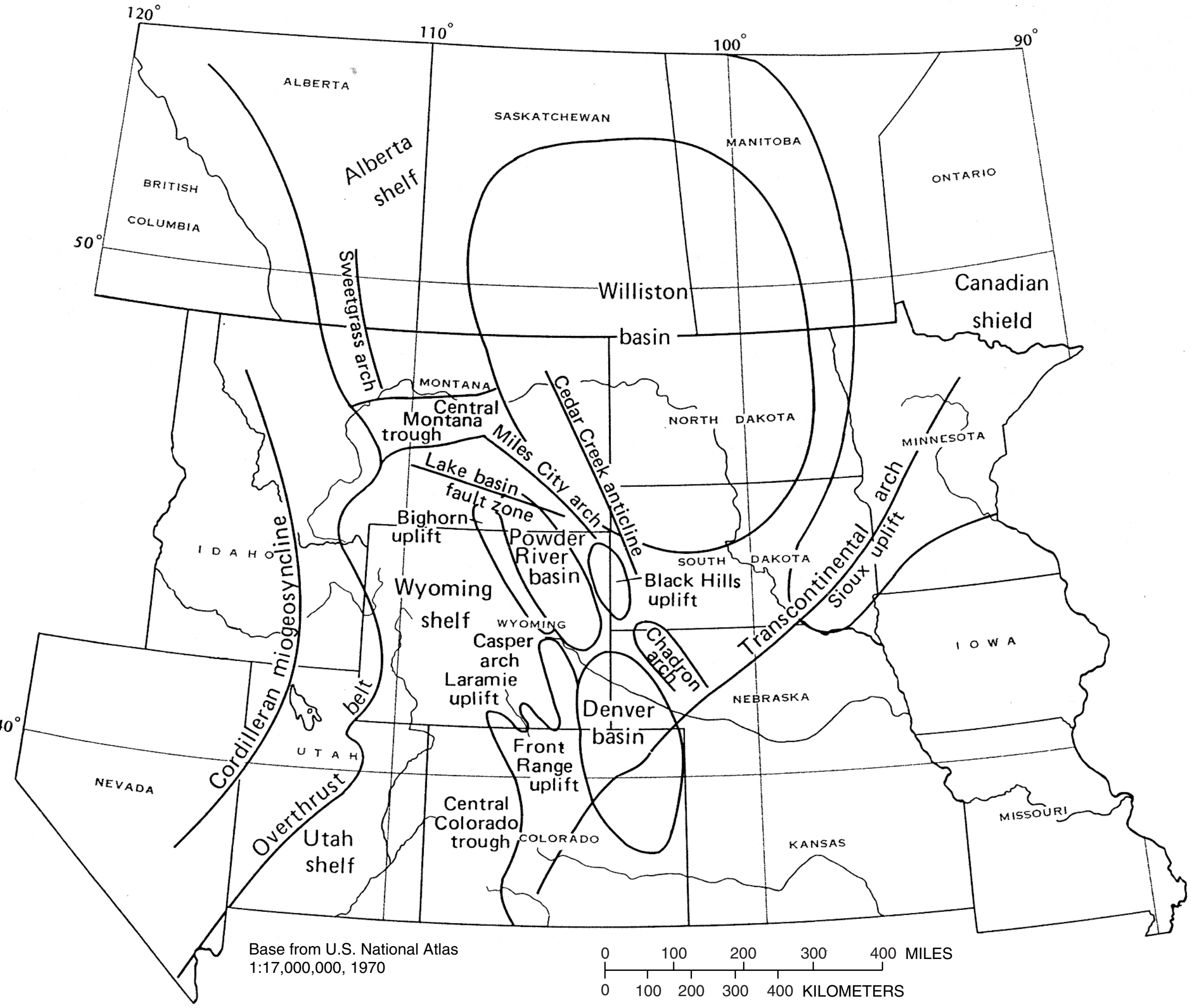


Figure 10. Regional paleostructure during Jurassic and Cretaceous time in the western interior of the United States (modified from Anna, 1986).

Stratigraphy

Precambrian rocks form the basement in the northern Great Plains area. Precambrian rocks are exposed in the central core of many of the mountain ranges, but lie greater than 15,000 ft below land surface at the center of the Williston Basin (Downey and Dinwiddie, 1988).

Rocks of Cambrian and Ordovician age consist of sandstone, shale, limestone, and dolomite and represent the shoreward facies of a transgressive sea (Peterson, 1981). The extent of the Cambrian and Ordovician rocks in the northern Great Plains area is shown in figure 11. The principal geologic units of Cambrian and Ordovician age are the Deadwood Formation, Emerson Formation, Winnipeg Formation, Red River Formation (Whitewood Formation), and Stony Mountain Formation (fig. 12). Rocks of Cambrian and Ordovician age extend into Canada where they are exposed along the Precambrian shield (Downey, 1986). Erosion during Devonian time truncated the Ordovician geologic units in South Dakota and Wyoming to the south of a line extending between the central Black Hills and southern Bighorn Mountains (Peterson, 1981). Rocks of Silurian age are not present in the Black Hills area.

The extent of Mississippian rocks in the northern Great Plains area is shown in figure 11. These rocks overlying the Bakken Formation (where present) are termed the Madison Limestone, or Madison Group where divided (fig. 12). The Madison Limestone consists of a sequence of marine carbonates and evaporites deposited mainly in a warm, shallow-water environment (Downey, 1986). Development of karst (solution) features in the Madison Limestone was common because the carbonate rocks are relatively soluble in water (Downey, 1986). Complex and interconnected solution features developed in the Madison Limestone during tropical conditions when it was exposed at or near land surface (Busby and others, 1995). Large and extensive cave systems have formed in the outcrop areas of the Madison Limestone in the Bighorn Mountains and in the Black Hills.

Rocks of Pennsylvanian age consist primarily of marine sandstone, shale, siltstone, and carbonate. The Pennsylvanian rocks are divided into many different geologic units (fig. 12). Rocks of Pennsylvanian-age have been truncated by pre-Jurassic erosion progressively northward across central Montana; these rocks thin to zero thickness near the axis of the central Montana trough (Downey, 1986; figs. 10 and 11).

A sequence of red shale, siltstone, and evaporite deposits belonging to the upper part of the Goose Egg and Spearfish Formations of Triassic age overlie the Minnelusa Formation (Downey and Dinwiddie, 1988). Jurassic rocks, which include the Nesson, Piper, Rierdon, and Sundance Formations and their equivalents (fig. 12) are predominantly carbonate, shale, and calcareous shale (Anna, 1986).

Deposits during Cretaceous time primarily were sandstones, shales, and minor carbonates (Redden and Lisenbee, 1996). A number of formation names have been applied to the various Cretaceous units in the northern Great Plains area; however, in several instances, these formation names are used only in one State or subregion (fig. 13). Lower Cretaceous rocks (fig. 13) range in thickness from zero in eastern North Dakota and South Dakota to more than 1,400 ft in west-central Wyoming (Anna, 1986). The extent of the Lower Cretaceous sandstones, which include the Inyan Kara Group, Muddy Sandstone, and Newcastle or Dakota Sandstone, is shown in figure 11. The sedimentary pattern of Upper Cretaceous rocks (fig. 13) is associated with four main transgressions and regressions of shallow seas.

Tertiary units (fig. 13) generally were deposited in a continental environment (Downey, 1986). Deposits of Quaternary age in the northern Great Plains area consist of alluvium, glacial materials, and other surficial deposits. Alluvial deposits fill major drainages in the area. Glacial deposits are located only in the eastern parts of North Dakota and South Dakota and in the northernmost part of Montana (Downey, 1986).

Local Geologic Setting

The Black Hills uplift is a northwest-trending, asymmetric, elongate dome, or doubly plunging anticline. Uplift began about 62 million years ago during the Laramide orogeny and probably continued in the Eocene period (Redden and Lisenbee, 1996). Large anticlines occur on the northern and southern flanks of the Black Hills and plunge away from the uplift into the surrounding plains. Numerous smaller folds, faults, domes, and monoclines also occur in the Black Hills (fig. 14). Igneous intrusions were emplaced on the northern flanks of the uplift during the Tertiary Period.

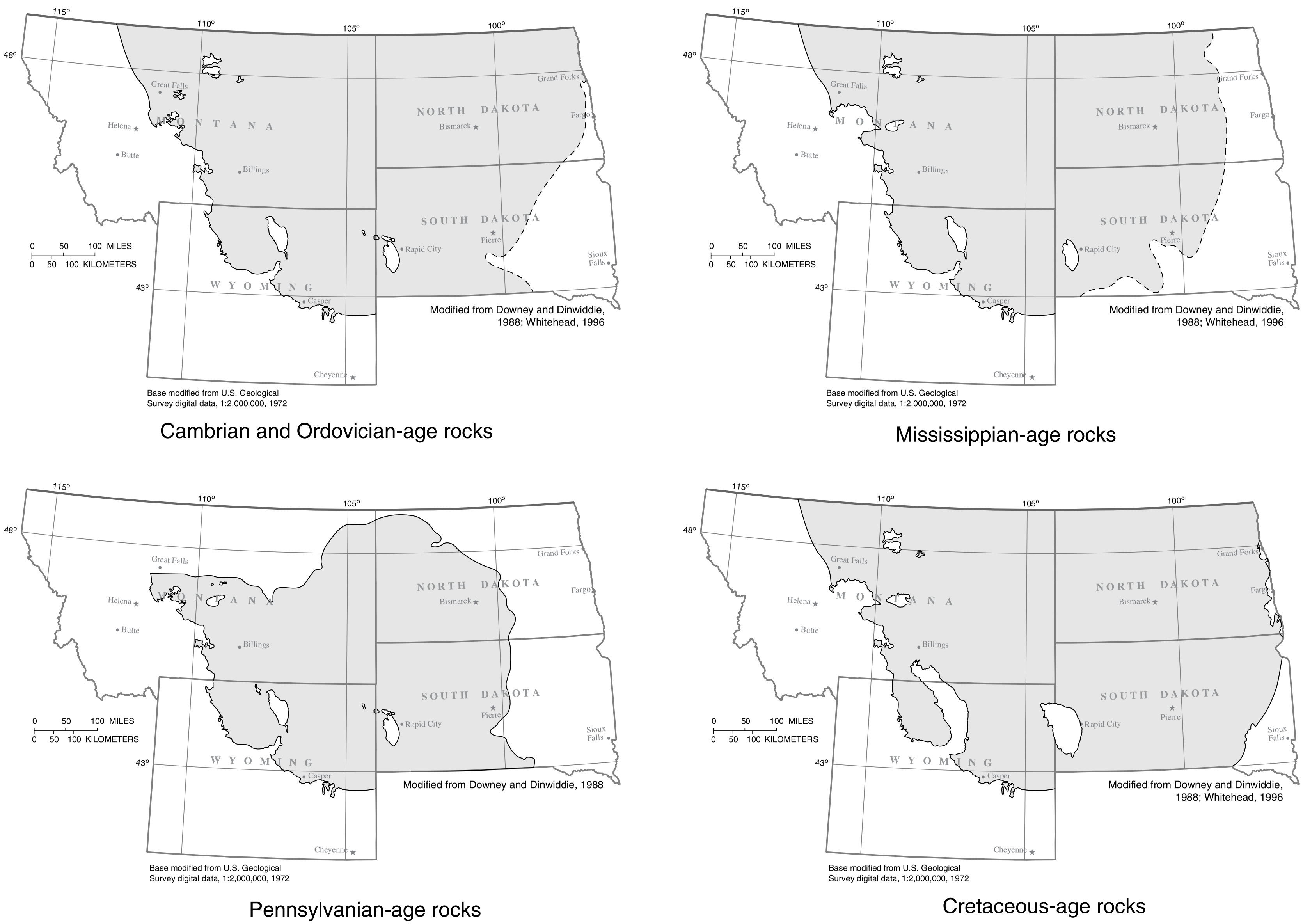


Figure 11. Approximate extent of rocks in the northern Great Plains area for selected geologic periods.

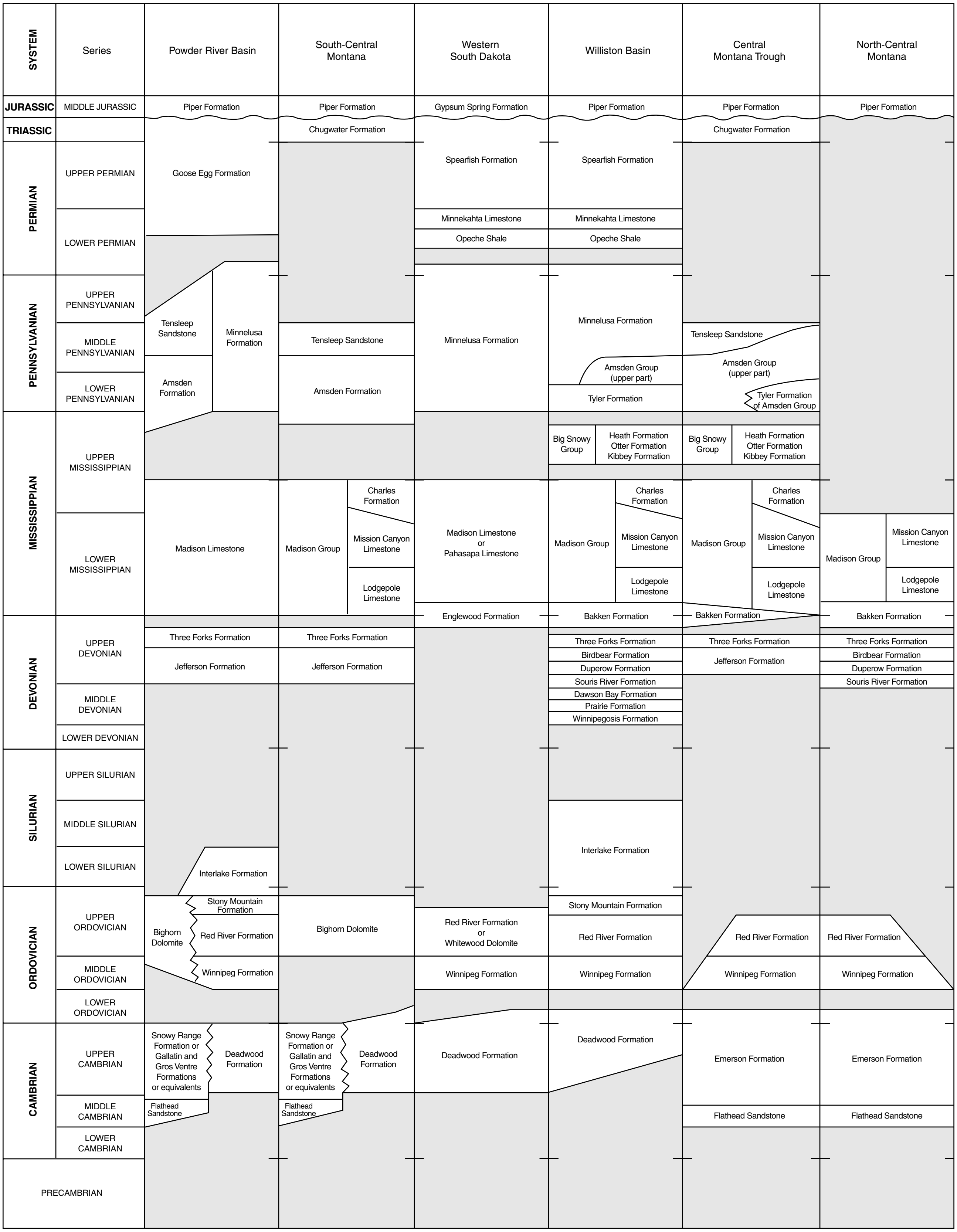
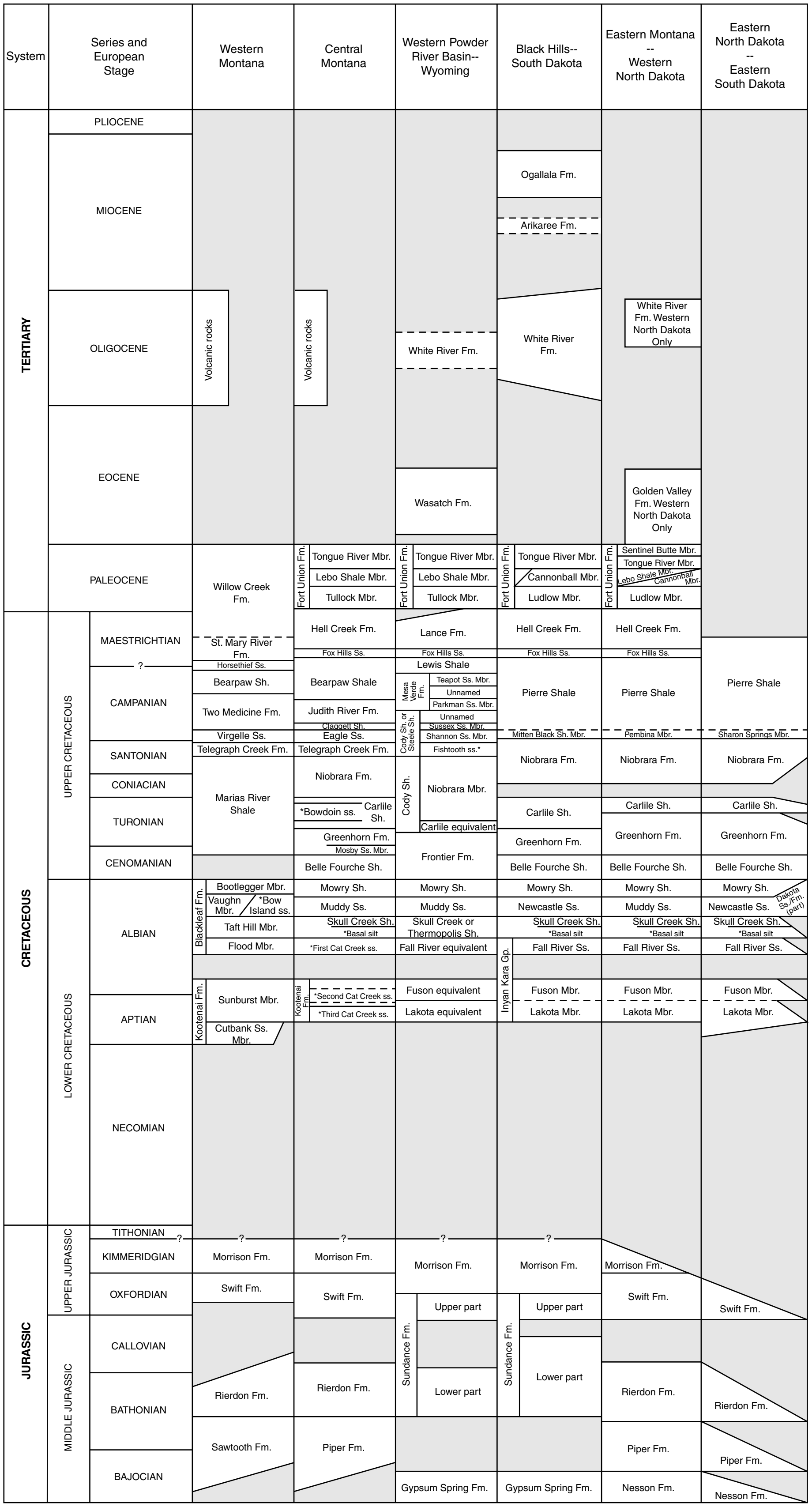


Figure 12. Generalized correlation chart for Paleozoic-age rocks in Montana, North Dakota, South Dakota, and Wyoming (modified from Downey, 1986).



* Of informal or subsurface usage

Figure 13. Generalized correlation chart for Mesozoic- and Cenozoic-age rocks in Montana, North Dakota, South Dakota, and Wyoming (modified from Downey, 1986).

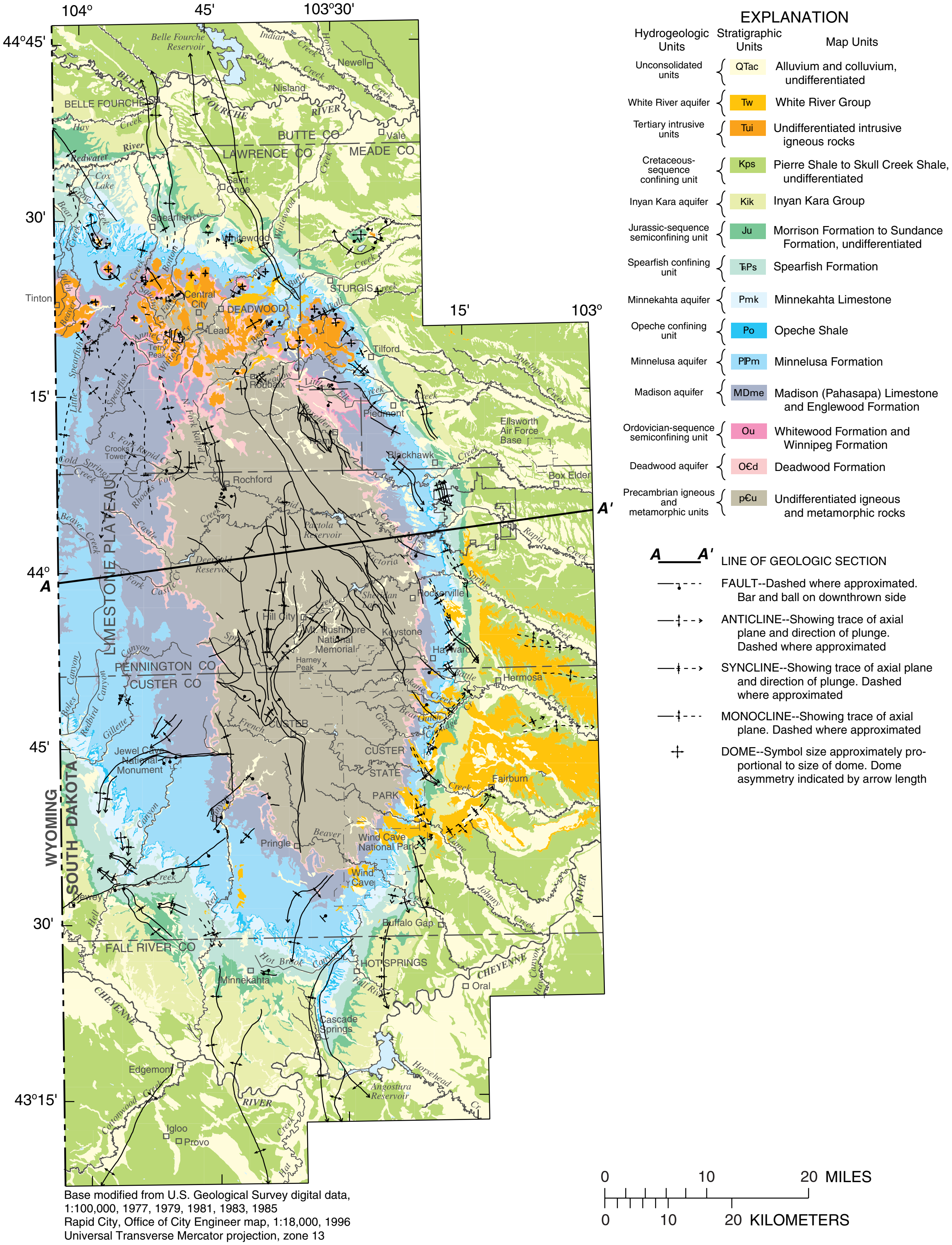


Figure 14. Distribution of hydrogeologic units in the Black Hills area (modified from Strobel and others, 1999).

The oldest stratigraphic units in the study area are the Precambrian igneous and metamorphic rocks (fig. 9), which underlie the Paleozoic, Mesozoic, and Cenozoic rocks and sediments. These Precambrian rocks range in age from 1.7 to about 2.5 billion years and were eroded to a gentle undulating plain at the beginning of the Paleozoic Era (Gries, 1996). The Precambrian rocks are highly variable in composition and are composed mostly of metasediments, such as schists and graywackes. The Paleozoic and Mesozoic rocks were deposited on the Precambrian rocks as nearly horizontal beds. Subsequent uplift during the Laramide orogeny and related erosion exposed the Precambrian rocks in the central core of the Black Hills, with many of the Paleozoic and Mesozoic sedimentary rocks exposed in roughly concentric rings around the core. The exposed Precambrian rocks commonly are referred to as the crystalline core.

The layered series of sedimentary rocks surrounding the crystalline core includes outcrops of the Madison Limestone (also locally known as the Pahasapa Limestone) and the Minnelusa Formation. The bedrock sedimentary formations typically dip away from the uplifted Black Hills (fig. 15) at angles that can approach or exceed 15 to 20 degrees near the outcrops, and decrease with distance from the uplift to less than 1 degree (Carter and Redden, 1999a, 1999b, 1999c, 1999d, 1999e). Following are descriptions for the bedrock formations that contain major aquifers in the Black Hills area.

The oldest sedimentary unit in the study area is the Cambrian- and Ordovician-age Deadwood Formation, which is composed primarily of brown to light-gray glauconitic sandstone, shale, limestone, and local basal conglomerate (Strobel and others, 1999). These sediments were deposited on top of a generally horizontal plain of Precambrian rocks in a coastal- to near-shore environment (Gries, 1975). The thickness of the Deadwood Formation increases from south to north in the study area and ranges from 0 to 500 ft (Carter and Redden, 1999e). In the northern and central Black Hills, the Deadwood Formation is disconformably overlain by Ordovician rocks, which include the Whitewood and Winnipeg Formations. The Winnipeg Formation is absent in the southern Black Hills, and the Whitewood Formation has eroded to the south and is not present south of the approximate latitude of Nemo (DeWitt and others, 1986). In the southern Black Hills, the Deadwood Formation is unconformably overlain by the Devonian- and Mississippian-age Englewood Formation because of the absence of the Ordovician sequence. The Englewood Formation is overlain by the Madison Limestone.

The Mississippian-age Madison Limestone is a massive, gray to buff limestone that is locally dolomitic (Strobel and others, 1999). The Madison Limestone, which was deposited as a marine carbonate, was exposed at land surface for approximately 50 million years. During this period, significant erosion, soil development, and karstification occurred (Gries, 1996). There are numerous caves and fractures within the upper part of the formation (Peter, 1985). The thickness of the Madison Limestone increases from south to north in the study area and ranges from almost zero in the southeast corner of the study area (Rahn, 1985) to 1,000 ft east of Belle Fourche (Carter and Redden, 1999d). Because the Madison Limestone was exposed to erosion and karstification for millions of years, its contact with the overlying Minnelusa Formation is unconformable.

The Pennsylvanian- and Permian-age Minnelusa Formation consists mostly of yellow to red cross-stratified sandstone, limestone, dolomite, and shale (Strobel and others, 1999). In addition to sandstone and dolomite, the middle part of the formation consists of shale and anhydrite (DeWitt and others, 1986). The upper part of the Minnelusa Formation also may contain anhydrite, which generally has been removed by dissolution in or near the outcrop areas, occasionally forming collapse features filled with breccia (Braddock, 1963). The Minnelusa Formation was deposited in a coastal environment, and dune structures at the top of the formation may represent beach sediments (Gries, 1996). The thickness of the Minnelusa Formation increases from north to south and ranges from 375 ft near Belle Fourche to 1,175 ft near Edgemont in the study area (Carter and Redden, 1999c). In the northeastern part of the central Black Hills, little anhydrite occurs in the subsurface due to a change in the depositional environment. On the south and southwest side of the study area, the thickness of clastic units increases and a thick section of anhydrite occurs. In the southern Black Hills, the upper part of the Minnelusa Formation thins due to leaching of anhydrite. The Minnelusa Formation is disconformably overlain by the Permian-age Opeche Shale, which is overlain by the Minnekahta Limestone.

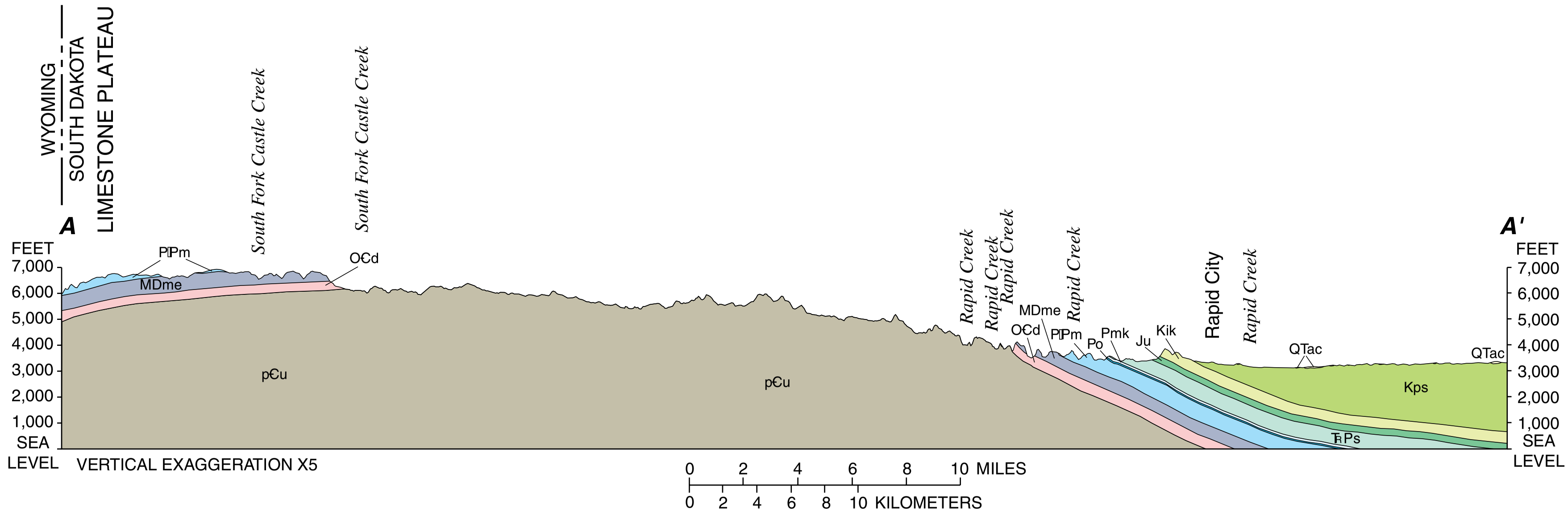


Figure 15. Geologic cross section A-A′ (modified from Strobel and others, 1999). Location of section is shown in figure 14. Abbreviations for stratigraphic intervals are explained in figure 9.

The Permian-age Minnekahta Limestone is a fine-grained, purple to gray laminated limestone (Strobel and others, 1999), which ranges in thickness from 25 to 65 ft in the study area. The Minnekahta Limestone is overlain by the Triassic- and Permian-age Spearfish Formation.

The Cretaceous-age Inyan Kara Group consists of the Lakota Formation and overlying Fall River Formation. The Lakota Formation consists of the Chilson, Minnewaste Limestone, and Fuson Shale members. The Lakota Formation consists of yellow, brown, and reddish-brown massive to thinly bedded sandstone, pebble conglomerate, siltstone, and claystone of fluvial origin (Gott and others, 1974); locally there are lenses of limestone and coal. The Fall River Formation is a brown to reddish-brown, fine-grained sandstone, thin bedded at the top and massive at the bottom (Strobel and others, 1999). The thickness of the Inyan Kara Group ranges from 135 to 900 ft in the study area (Carter and Redden, 1999a).

Ground-Water Framework

The hydrogeologic setting of the Black Hills area is schematically illustrated in figure 16, and the areal distribution of the hydrogeologic units is shown in figure 14. Four of the major aquifers in the Black Hills area (Deadwood, Madison, Minnelusa, and Inyan Kara aquifers) are regionally extensive and are discussed in the following sections in the context of regional and local hydrologic settings. A fifth major aquifer (Minnekahta aquifer) generally is used only locally, as are aquifers in the igneous and metamorphic rocks within the crystalline core area and in alluvium. In some local areas, wells are completed in strata that generally are considered to be semiconfining and confining units.

Regional Aquifers

The major aquifers underlie parts of Montana, North Dakota, South Dakota, Wyoming, and Canada. The parts of the regional aquifers in Canada are not described or shown in this report.

The Paleozoic aquifers include the Cambrian-Ordovician aquifer (Deadwood aquifer in the Black Hills), Mississippian aquifer (Madison aquifer in the Black Hills), and the Pennsylvanian aquifer (Minnelusa aquifer in the Black Hills). Recharge to the Paleozoic aquifers occurs in high-altitude outcrop areas around the major uplifts such as the Black Hills uplift (fig. 17).

The Cambrian-Ordovician (or Deadwood) aquifer is contained within the sandstones of Cambrian age (Deadwood Formation and equivalents) and limestones of Ordovician age (Red River Formation and equivalents) (fig. 12). Generally, flow in the Cambrian-Ordovician aquifer is from the high-altitude recharge areas to the northeast. Discharge (fig. 17) from the Cambrian-Ordovician aquifer is to adjacent aquifers, lakes and springs in eastern North Dakota, and springs and seeps where the aquifer crops out in Canada (Downey, 1984). Within the Great Plains region, the Cambrian-Ordovician aquifer contains fresh water (dissolved solids concentrations less than 1,000 mg/L (milligrams per liter)) only in an area surrounding the Black Hills and in a small area in north-central Wyoming (Whitehead, 1996). The aquifer is a brine (dissolved solids concentration greater than 35,000 mg/L) in eastern Montana and western and central North Dakota (Whitehead, 1996).

The Mississippian (or Madison) aquifer is contained within the limestones, siltstones, sandstones, and dolomite of the Madison Limestone or Group. Generally, water in the Mississippian aquifer is confined except in outcrop areas. Flow in the Mississippian aquifer generally is from the recharge areas to the northeast. Discharge (fig. 17) from the Mississippian aquifer occurs by upward leakage to the lower Cretaceous aquifer in central South Dakota and eastern flow to the Cambrian-Ordovician aquifer in eastern North Dakota (Downey, 1984). Water in the Mississippian aquifer is fresh only in small areas near recharge areas and becomes saline to slightly saline as it moves downgradient. The water is a brine with dissolved solids concentrations greater than 300,000 mg/L in the deep parts of the Williston Basin (Whitehead, 1996).

The Pennsylvanian (or Minnelusa) aquifer is contained within the sandstones and limestones of the Minnelusa Formation, Tensleep Sandstone, Amsden Formation, and equivalents of Pennsylvanian age (fig. 12). Water in the Pennsylvanian aquifer moves from recharge areas to the northeast to discharge areas in eastern South Dakota (Downey and Dinwiddie, 1988). Some water discharges by upward leakage to the lower Cretaceous aquifer (Swenson, 1968, Gott and others, 1974).

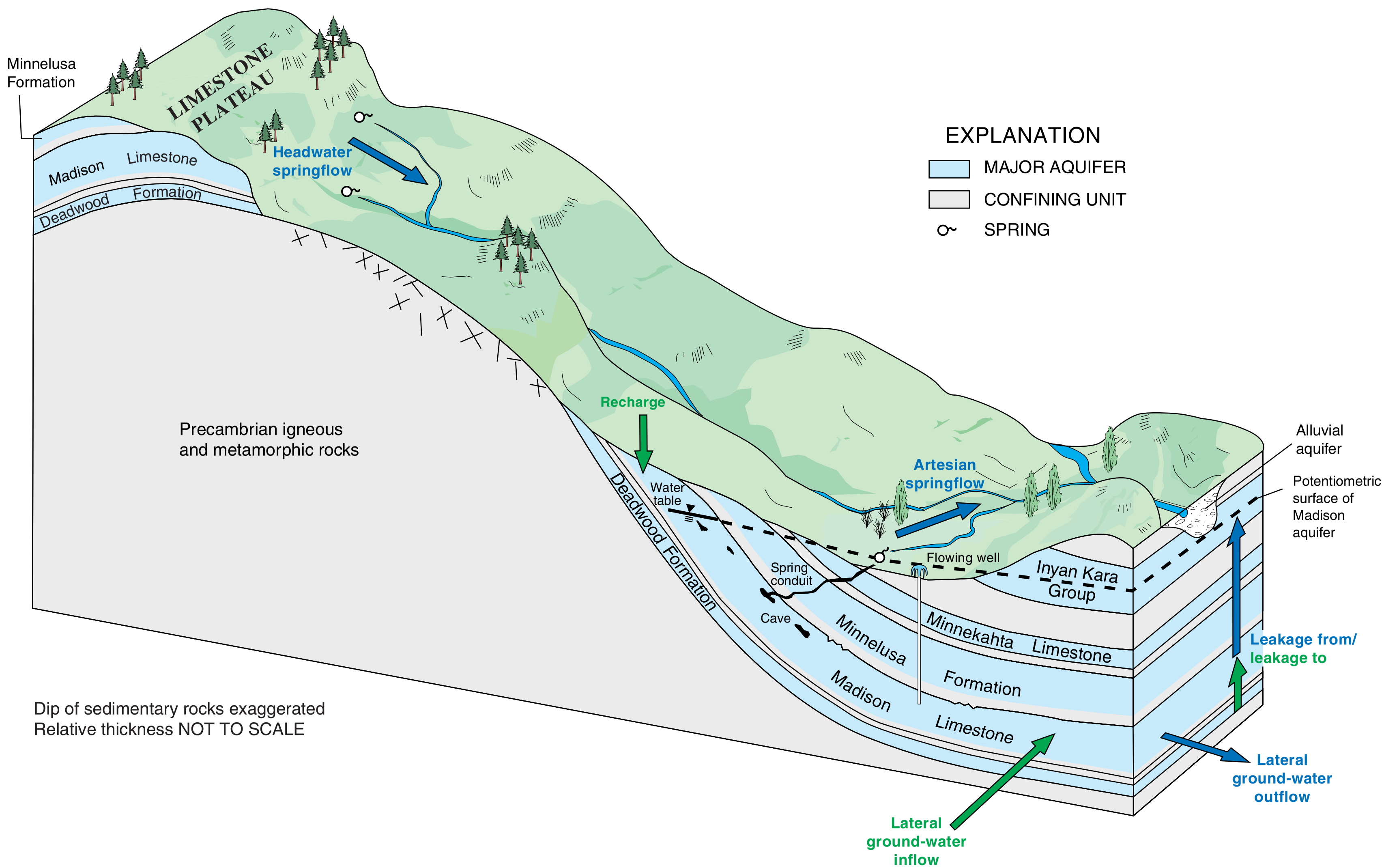


Figure 16. Schematic showing simplified hydrogeologic setting of the Black Hills area. Schematic generally corresponds with geologic cross section shown in figure 15. Components considered for hydrologic budget of the Madison aquifer also are shown with inflow components shown in green and outflow components shown in blue.

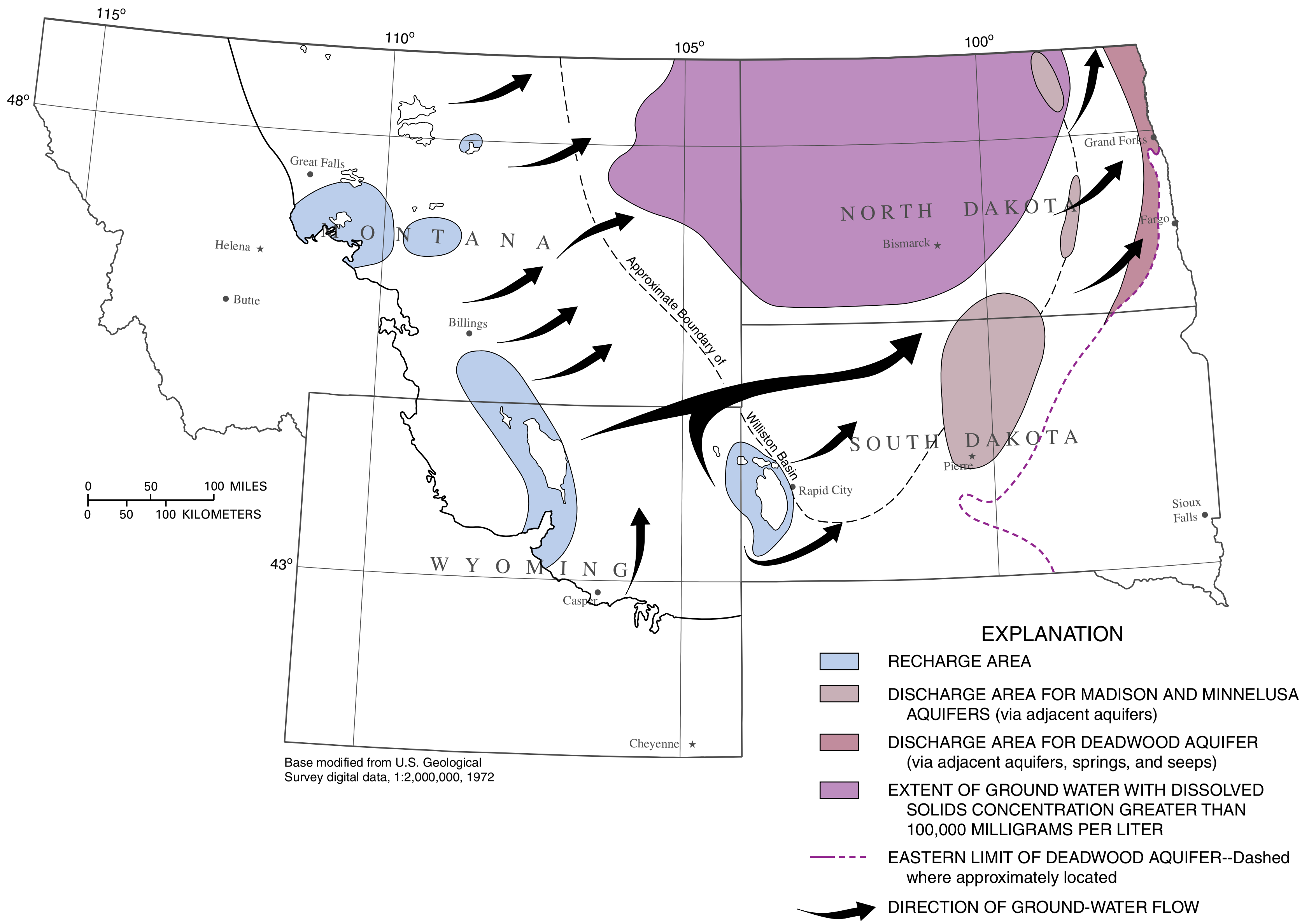


Figure 17. General direction of ground-water flow in regional aquifer system within Paleozoic aquifer units (modified from Downey and Dinwiddie, 1988; Whitehead, 1996).

Several sandstone units (fig. 13) compose the lower Cretaceous aquifer, which is known as the Inyan Kara aquifer in South Dakota. Generally, water in the lower Cretaceous aquifer is confined by several thick shale layers except in aquifer outcrop areas around structural uplifts, such as the Black Hills. Water in the lower Cretaceous aquifer generally moves northeasterly from high-altitude recharge areas to discharge areas in eastern North Dakota and South Dakota (Whitehead, 1996). Although the aquifer is widespread, it contains little fresh water. Water is fresh only in small areas in central and south-central Montana and north and east of the Black Hills uplift (Whitehead, 1996). More than one-half of the water in the lower Cretaceous aquifer is moderately saline, and the water is very saline or a brine in the deep parts of the Williston and Powder River Basins (Whitehead, 1996). Much of the saline water is believed to be from upward leakage of mineralized water from the Paleozoic aquifers.

Local Aquifers

Many of the sedimentary units contain aquifers, both within and beyond the study area. Within the Paleozoic rock interval, aquifers in the Deadwood Formation, Madison Limestone, Minnelusa Formation, and Minnekahta Limestone are used extensively. These aquifers are collectively confined by the underlying Precambrian rocks and the overlying Spearfish Formation. Individually, these aquifers are separated by minor confining layers or by low-permeability layers within the individual units. In general, ground-water flow in these aquifers is radially outward from the central core of the Black Hills. Although the lateral component of flow generally predominates, the vertical component of flow, and thus leakage between these aquifers, can be extremely variable (Peter, 1985; Greene, 1993).

Although the Precambrian basement rocks generally have low permeability and form the lower confining unit for the series of sedimentary aquifers (fig. 16), localized aquifers occur in many locations in the crystalline core of the Black Hills, where enhanced secondary permeability has resulted from weathering and fracturing. Where the Precambrian rocks are saturated, unconfined (water-table) conditions generally occur and topography can strongly control ground-water flow directions.

The Deadwood Formation contains the Deadwood aquifer, which overlies the Precambrian rocks. The Deadwood aquifer, which is used mainly by domestic and municipal users near its outcrop area, receives recharge primarily from precipitation on the outcrop. There may be some hydraulic connection between the Deadwood aquifer and the underlying weathered Precambrian rocks, but regionally the Precambrian rocks act as a lower confining unit to the Deadwood aquifer. The Whitewood and Winnipeg Formations, where present, act as overlying semiconfining units to the Deadwood aquifer (Strobel and others, 1999). The Whitewood and Winnipeg Formations locally may transmit water and exchange water with the Deadwood aquifer, but regionally are not considered aquifers. Where the Whitewood and Winnipeg Formations are absent, the Deadwood aquifer is in contact with the overlying Englewood Formation, which was included as part of the Madison aquifer for this study.

The Madison aquifer generally occurs within the karstic upper part of the Madison Limestone, where numerous fractures and solution openings have created extensive secondary porosity and permeability. Strobel and others (1999) included the entire Madison Limestone and the Englewood Formation in their delineation of the Madison aquifer. Thus, in this report, outcrops of the Madison Limestone and Englewood Formation (fig. 14) are referred to as the outcrop of the Madison Limestone for simplicity. The Madison aquifer receives significant recharge from streamflow losses and precipitation on the outcrop. Low-permeability layers in the lower part of the Minnelusa Formation generally act as an upper confining unit to the Minnelusa aquifer. However, karst features in the upper part of the Madison Limestone may have reduced the effectiveness of the overlying confining unit in some locations.

The Minnelusa aquifer occurs within layers of sandstone, dolomite, and anhydrite in the lower portion of the Minnelusa Formation and sandstone and anhydrite in the upper portion. The Minnelusa aquifer has primary porosity in the sandstone units and secondary porosity from collapse breccia associated with dissolution of interbedded evaporites and fracturing. The Minnelusa aquifer receives significant recharge from streamflow losses and precipitation on the outcrop. Streamflow recharge to the Minnelusa aquifer generally is less than to the Madison aquifer (Carter, Driscoll, and Hamade, 2001), which is preferentially recharged because of its upslope location. The Minnelusa aquifer is confined by the overlying Opeche Shale.

Both the Madison and Minnelusa aquifers are potential sources for numerous large artesian springs in the Black Hills area, and hydraulic connections between the two aquifers are possible in other locations (Naus and others, 2001). Ground-water flowpaths and velocities in both aquifers are influenced by anisotropic and heterogeneous hydraulic properties caused by secondary porosity.

The Minnekahta aquifer, which overlies the Opeche Shale, typically is very permeable, but well yields can be limited by the small aquifer thickness. The Minnekahta aquifer receives significant recharge from precipitation on the outcrop and some additional recharge from streamflow losses. The overlying Spearfish Formation acts as a confining unit to the aquifer and to other underlying Paleozoic aquifers. Hence, most of the artesian springs occur near the outcrop of the Spearfish Formation.

Within the Mesozoic rock interval, the Inyan Kara Group comprises an aquifer that is used extensively. Aquifers in various other units of the Mesozoic rock interval are used locally to lesser degrees. The Inyan Kara aquifer receives recharge primarily from precipitation on the outcrop. The Inyan Kara aquifer also may receive recharge from leakage from the underlying Paleozoic aquifers (Swenson, 1968; Gott and others, 1974). As much as 4,000 ft of Cretaceous shales act as the upper confining unit to aquifers in the Mesozoic rock interval.

Confined (artesian) conditions generally exist within the sedimentary aquifers where an upper confining layer is present. Under confined conditions, water in a well will rise above the top of the aquifer in which it is completed. Flowing wells will result when drilled in areas where the water level, or potentiometric surface, is above the land surface. Flowing wells and artesian springs that originate from confined aquifers are common around the periphery of the Black Hills.

Numerous headwater springs originating from the Paleozoic units at high altitudes on the western side of the study area provide base flow for many streams. These streams flow across the crystalline core of the Black Hills, and most streams generally lose all or part of their flow as they cross the outcrops of the Madison Limestone (Rahn and Gries, 1973; Hortness and Driscoll, 1998). Karst features of the Madison Limestone, including sinkholes, collapse features, solution cavities, and caves, are responsible for the Madison aquifer's capacity to accept recharge from streamflow. Large streamflow losses also occur in many locations within the outcrop of the Minnelusa Formation (Hortness and Driscoll, 1998). Large artesian springs, originating primarily from the Madison and Minnelusa aquifers, occur in many locations downgradient from these loss zones, most commonly within or near the outcrop of the Spearfish Formation. These springs provide an important source of base flow in many streams beyond the periphery of the Black Hills (Rahn and Gries, 1973; Miller and Driscoll, 1998).

Characteristics and Properties of Major Aquifers

Aquifer characteristics and properties for the major aquifers in the study area (Deadwood, Madison, Minnelusa, Minnekahta, and Inyan Kara aquifers) are presented in this section. Aquifer characteristics, including areal extent, thickness, and storage volume, are presented in table 1. Aquifer characteristics for the Precambrian aquifer also are presented because numerous wells are completed in this aquifer in the crystalline core of the Black Hills. The areal extent of the aquifers was determined using a geographic information system (GIS) coverage by Williamson and others (2000) of the hydrogeologic unit map by Strobel and others (1999) for the study area.

Localized aquifers occur in the Precambrian igneous and metamorphic rocks that make up the crystalline core of the Black Hills and are referred to collectively as the Precambrian aquifer. The Precambrian aquifer is not continuous, and ground-water flow is mainly controlled by secondary permeability caused by fracturing and weathering. The Precambrian aquifer is considered to be contained in the area where the Precambrian rocks are exposed in the central core, which has an area of approximately 825 mi^2 in the study area. The thickness of the aquifer has been estimated by Rahn (1985) to be generally less than 500 ft, which was considered the average thickness (table 1). Wells in the Custer area have been completed at depths greater than 1,000 ft, indicating that the Precambrian aquifer is thicker in some locations. The Precambrian aquifer is mostly unconfined, but may have locally confined conditions. The area of the sedimentary aquifers is smaller than the area of the Precambrian rocks because erosion has removed the sedimentary rocks in the central core of the Black Hills.

Table 1. Summary of the characteristics of major and Precambrian aquifers in the study area

[mi^2, square miles; ft, feet; acre-ft, acre-feet]

Aquifer	Area extent (mi^2)	Maximum formation thickness (ft)	Average saturated thickness (ft)	Effective porosity[1]	Estimated amount of recoverable water in storage[2] (million acre-ft)
Precambrian	[3]5,041	--	[1]500	0.01	2.6
Deadwood	4,216	500	226	.05	30.5
Madison	4,113	1,000	[4]521	.05	[5]62.7
Minnelusa	3,623	1,175	[6]736	.05	[5]70.9
Minnekahta	3,082	65	50	.05	4.9
Inyan Kara	2,512	900	310	.17	84.7
Combined storage for major and Precambrian aquifers					256.3

[1]From Rahn (1985).
[2]Storage estimated by multiplying area times average thicknesses times effective porosity.
[3]The area used in storage calculation was the area of the exposed Precambrian rocks, which is 825 mi^2.
[4]Average saturated thickness of the confined area of the Madison aquifer. The unconfined area had an average saturated thickness of 300 ft.
[5]Storage values are the summation of storage in the confined and unconfined areas.
[6]Average saturated thickness of the confined area of the Minnelusa aquifer. The unconfined area had an average saturated thickness of 142 ft.

Large amounts of water are stored within the major aquifers, but not all of it is recoverable because some of the water is contained in unconnected pore spaces. Thus, effective porosity, which is the porosity of a rock that consists of interconnected voids, was used in estimating the amount of recoverable water in storage (table 1). Where aquifer units are not fully saturated (generally in and near outcrop areas), the saturated thickness is less than the formation thickness and the aquifer is unconfined. For the Madison and Minnelusa aquifers, it was possible to delineate the saturated thickness of the unconfined portions of these aquifers, as discussed later in this section. Average saturated thicknesses of the unconfined and confined portions of the Madison and Minnelusa aquifers were used in storage estimates for these aquifers. For the other major aquifers, full saturation was assumed because more detailed information was not available.

The total volume of recoverable water stored in the major aquifers (including the Precambrian aquifer) within the study area is estimated as 256 million acre-ft. The largest storage volume is for the Inyan Kara aquifer because of the large effective porosity (0.17). Storage in the Minnelusa aquifer is larger than in the Madison aquifer, primarily because of larger average saturated thickness.

Well yields (fig. 18) for the major aquifers were obtained from the USGS Ground Water Site Inventory (GWSI) database. The mean well yields for the aquifers generally are much higher than the median well yields because some well yields are very high. Well yields generally are lower for wells completed in the Precambrian rocks than the major aquifers because the Precambrian aquifer is not continuous and most of the available water is stored in fractures. The Madison aquifer has the potential for high well yields, and the mean and median well yields are higher in the Madison aquifer than the other major aquifers. The Minnelusa aquifer also has the potential for high well yields. Low well yields are possible in some locations for all the major aquifers.

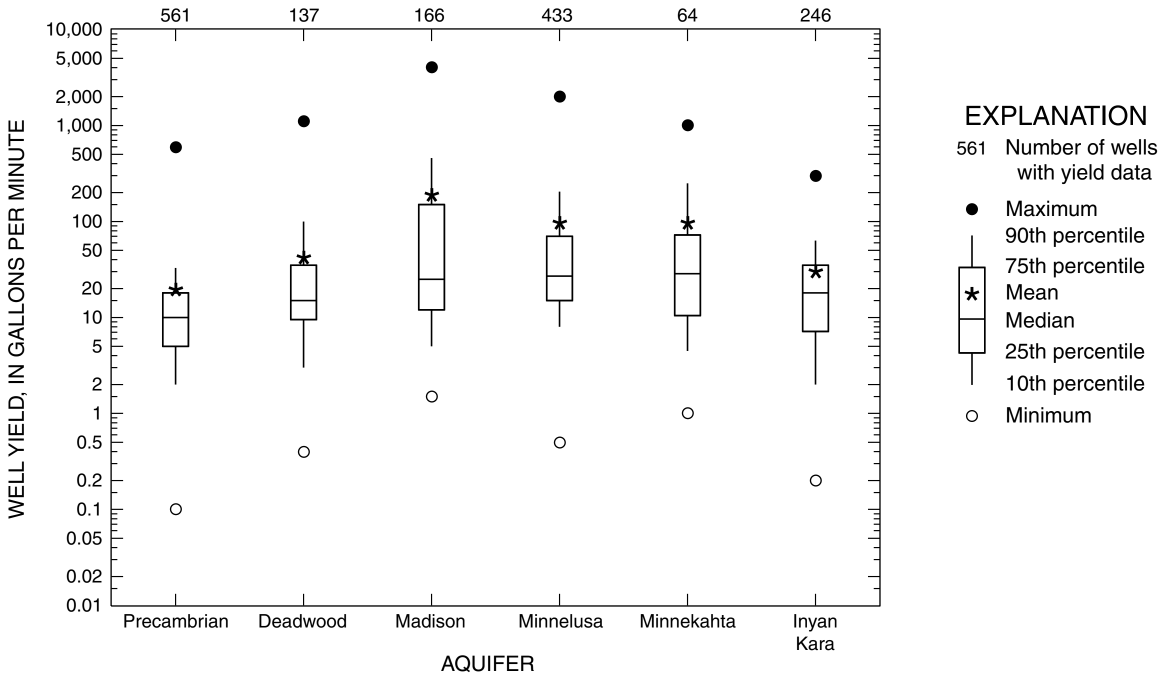


Figure 18. Boxplots showing distribution of well yields from selected aquifers (data obtained from U.S. Geological Survey Ground Water Site Inventory database).

Aquifer properties, including hydraulic conductivity, transmissivity, storage coefficient, and porosity, are presented in table 2 for the major aquifers and the Precambrian aquifer. The estimates presented for the various aquifer properties are based on previous studies. In general, the Madison aquifer has the highest hydraulic conductivity and transmissivity estimates of the major aquifers. Transmissivity and hydraulic conductivity also may be high in the Minnelusa aquifer. The Inyan Kara aquifer generally has the highest effective porosity of the major aquifers.

The potentiometric surfaces of the Madison and Minnelusa aquifers are shown in figures 19 and 20, respectively. In many locations, ground-water flow in these aquifers follows the bedding dip, which generally is radially away from the central part of the uplift. Structural features, such as folds and faults, may have local influence on ground-water flowpaths. Ground-water flowpaths and velocities also are heavily influenced by anisotropic and heterogeneous hydraulic properties of the Madison aquifer. Flowpaths are not necessarily orthogonal to equipotential lines because of highly variable directional transmissivities and may be further influenced by vertical flow components between the Madison and Minnelusa aquifers. Long (2000) described anisotropy in the Madison aquifer in the Rapid City area that causes ground-water flow to be nearly parallel to mapped equipotential lines in some cases. Regional ground-water flow from the west may influence the potentiometric surface in both aquifers in the northern and southwestern parts of the study area. Locations of artesian springs that probably originate from ground-water discharge from the Madison or Minnelusa aquifers and have potential to influence potentiometric surfaces also are shown in figures 19 and 20.

Table 2. Estimates of hydraulic conductivity, transmissivity, storage coefficient, and porosity from previous investigations

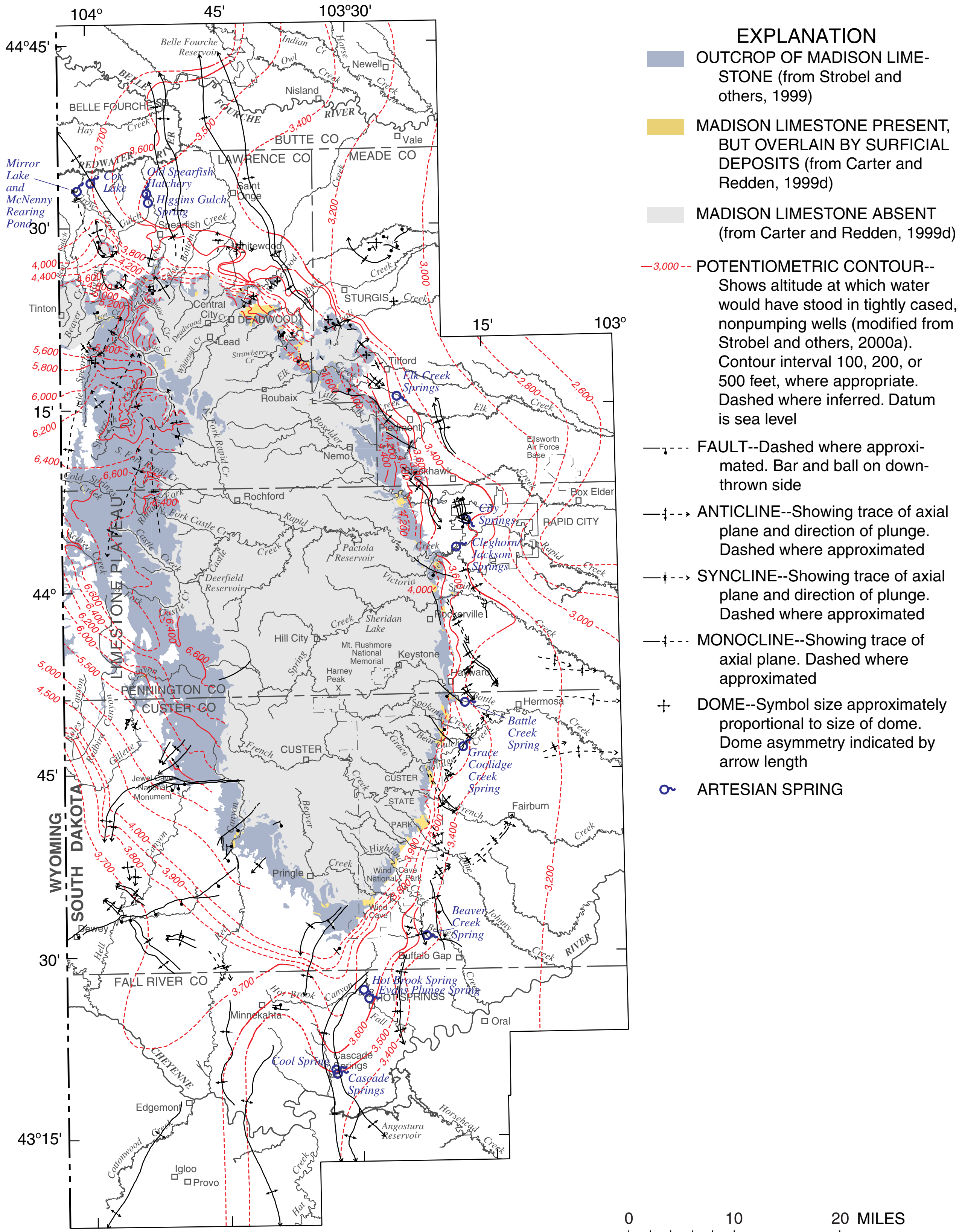
[ft/d, feet per day; ft^2/d, feet squared per day; --, no data; <, less than]

Source	Hydraulic conductivity (ft/d)	Transmissivity (ft^2/d)	Storage coefficient	Total porosity/ effective porosity	Area represented
Precambrian aquifer					
Rahn, 1985	--	--	--	0.03/0.01	Western South Dakota
Galloway and Strobel, 2000		450 - 1,435		0.10/--	Black Hills area
Deadwood aquifer					
Downey, 1984	--	250 - 1,000	--	--	Montana, North Dakota, South Dakota, Wyoming
Rahn, 1985	--	--	--	0.10/0.05	Western South Dakota
Madison aquifer					
Konikow, 1976	--	860 - 2,200	--	--	Montana, North Dakota, South Dakota, Wyoming
Miller, 1976	--	0.01 - 5,400	--	--	Southeastern Montana
Blankennagel and others, 1977	$2.4\text{x}10^{-5}$ - 1.9	--	--	--	Crook County, Wyoming
Woodward-Clyde Consultants, 1980	--	3,000	$2\text{x}10^{-4}$ - $3\text{x}10^{-4}$	--	Eastern Wyoming, western South Dakota
Blankennagel and others, 1981	--	5,090	$2\text{x}10^{-5}$	--	Yellowstone County, Montana
Downey, 1984	--	250 - 3,500	--	--	Montana, North Dakota, South Dakota, Wyoming
Plummer and others, 1990	--	--	$1.12\text{x}10^{-6}$ - $3\text{x}10^{-5}$	--	Montana, South Dakota, Wyoming
Rahn, 1985	--	--	--	0.10/0.05	Western South Dakota
Cooley and others, 1986	1.04	--	--	--	Montana, North Dakota, South Dakota, Wyoming, Nebr.
Kyllonen and Peter, 1987	--	4.3 - 8,600	--	--	Northern Black Hills
Imam, 1991	$9.0\text{x}10^{-6}$	--	--	--	Black Hills area
Greene, 1993	--	1,300 - 56,000	0.002	0.35/--	Rapid City area
Tan, 1994	5 - 1,300	--	--	0.05	Rapid City area
Greene and others, 1999	--	2,900 - 41,700	$3\text{x}10^{-4}$ - $1\text{x}10^{-3}$	--	Spearfish area
Carter, Driscoll, Hamade, and Jarrell, 2001	--	100 - 7,400	--	--	Black Hills area
Minnelusa aquifer					
Blankennagel and others, 1977	$<2.4\text{x}10^{-5}$ - 1.4	--	--	--	Crook County, Wyoming
Pakkong, 1979	--	880	--	--	Boulder Park area, South Dakota
Woodward-Clyde Consultants, 1980	--	30 - 300	$6.6\text{x}10^{-5}$ - $2.0\text{x}10^{-4}$	--	Eastern Wyoming, western South Dakota

Table 2. Estimates of hydraulic conductivity, transmissivity, storage coefficient, and porosity from previous investigations–Continued

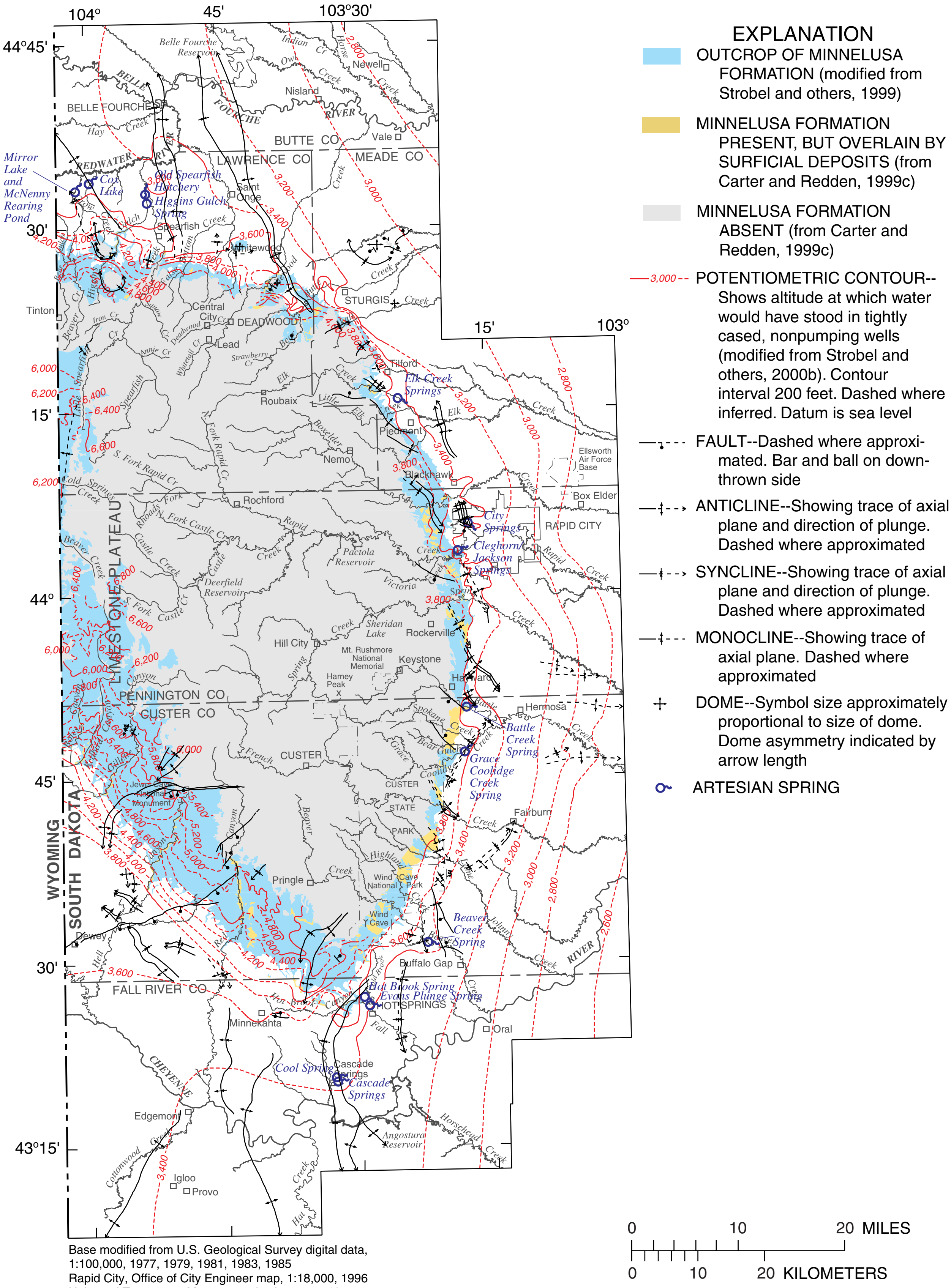
[ft/d, feet per day; ft^2/d, feet squared per day; --, no data; <, less than]

Source	Hydraulic conductivity (ft/d)	Transmissivity (ft^2/d)	Storage coefficient	Total porosity/ effective porosity	Area represented
Minnelusa aquifer—Continued					
Rahn, 1985	--	--	--	0.10/0.05	Western South Dakota
Kyllonen and Peter, 1987	--	0.86 - 8,600	--	--	Northern Black Hills
Greene, 1993	--	12,000	0.003	0.1/--	Rapid City area
Tan, 1994	32	--	--	--	Rapid City area
Greene and others, 1999	--	267 - 9,600	$5.0x10^{-9}$ - $7.4x10^{-5}$	--	Spearfish area
Carter, Driscoll, Hamade, and Jarrell, 2001	--	100 - 7,400	--	--	Black Hills area
Minnekahta aquifer					
Rahn, 1985	--	--	--	0.08/0.05	Western South Dakota
Inyan Kara aquifer					
Niven, 1967	0 - 100	--	--	--	Eastern Wyoming, western South Dakota
Miller and Rahn, 1974	0.944	178	--	--	Black Hills area
Gries and others, 1976	1.26	250 - 580	$2.1x10^{-5}$ - $2.5x10^{-5}$	--	Wall area, South Dakota
Boggs and Jenkins, 1980	--	50 - 190	$1.4x10^{-5}$ - $1.0x10^{-4}$	--	Northwestern Fall River County
Bredehoeft and others, 1983	8.3	--	$1.0x10^{-5}$	--	South Dakota
Rahn, 1985	--	--	--	0.26/0.17	Western South Dakota
Kyllonen and Peter, 1987	--	0.86 - 6,000	--	--	Northern Black Hills



Base modified from U.S. Geological Survey digital data, 1:100,000, 1977, 1979, 1981, 1983, 1985
Rapid City, Office of City Engineer map, 1:18,000, 1996
Universal Transverse Mercator projection, zone 13

Figure 19. Potentiometric surface of the Madison aquifer and locations of major artesian springs.



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Rapid City, Office of City Engineer map, 1:18,000, 1996
Universal Transverse Mercator projection, zone 13

Figure 20. Potentiometric surface of the Minnelusa aquifer and locations of major artesian springs.

Maps showing the saturated thickness of the unconfined areas of the Madison and Minnelusa aquifers are shown in figures 21 and 22, respectively. Both the Madison and Minnelusa aquifers are unconfined near their outcrops and confined (fully saturated) at some distance downdip from their outcrops. In general, the saturated thickness is less than 200 ft for most of the outcrop areas. These areas are especially susceptible to drought conditions, and the formations may even be dry in these areas regardless of precipitation conditions. In most areas, the aquifers are fully saturated within a short distance downdip of the outcrops. However, in the southwest part of the study area, neither aquifer is fully saturated for a distance of about 6 mi downdip of the respective outcrops.

Although the Limestone Plateau area is a large recharge area for the Madison and Minnelusa aquifers, saturated thicknesses generally are small within these aquifers in this area. Very few wells have been successfully completed in this area, especially within the Madison Limestone, where saturated conditions generally occur only near the bottom of the formation. Saturated thicknesses are limited by the discharge of springs along the eastern edge of the Plateau and by ground-water flow to the west. Fluctuations in ground-water levels in this area generally are smaller than other areas.

Overview of Other Aquifers

In addition to the major aquifers, many other aquifers are used throughout the study area. The Newcastle Sandstone, White River Group, and the unconsolidated units are considered aquifers where saturated (Strobel and others, 1999). In addition, many of the semiconfining and confining units shown in figure 14 may contain local aquifers. This section provides a brief overview from Strobel and others (1999) of other aquifers in the study area that are contained in various units from oldest to youngest.

The Whitewood Formation, where present, may contain a local aquifer, but seldom is used because of more reliable sources in the adjacent Madison or Deadwood aquifers. Local aquifers may exist in the Spearfish confining unit where gypsum and anhydrite have been dissolved, increasing porosity and permeability; these aquifers are referred to as the Spearfish aquifer in this report. The Jurassic-sequence semiconfining unit consists of shales and sandstones. Overall, this unit is semiconfining because of the low permeability of the interbedded shales; however, local aquifers exist in some formations such as the Sundance and Morrison Formations. These aquifers are referred to as the Sundance and Morrison aquifers in this report.

The Cretaceous-sequence confining unit mainly includes shales of low permeability, such as the Pierre Shale; local aquifers in the Pierre Shale are referred to as the Pierre aquifer in this report. Within the Graneros Group, the Newcastle Sandstone contains an important minor aquifer referred to as the Newcastle aquifer. Because water-quality characteristics (discussed in a subsequent section of this report) are very different between the Newcastle aquifer and the other units in the Graneros Group, data are presented for the Newcastle aquifer separately from the other units in the Graneros Group, known as the Graneros aquifer in this report.

Tertiary intrusive units are present only in the northern Black Hills, and generally are relatively impermeable, although "perched" ground water often is associated with intrusive sills. The White River aquifer consists of various discontinuous units of sandstone and channel sands along the eastern flank of the Black Hills and is considered a minor aquifer where saturated. Unconsolidated deposits of Tertiary or Quaternary age, including alluvium, colluvium, and wind-blown deposits, all have the potential to be local aquifers where they are saturated.

Surface-Water Framework

Streamflow within the study area is highly influenced by climatic and geologic conditions. The base flow of most streams in the Black Hills originates in the higher altitudes, where relatively large precipitation and small evapotranspiration result in more water being available for springflow and streamflow. Many streams have headwater springs originating from the Paleozoic carbonate rocks on the western side of the study area. Most of these streams flow in a generally eastward direction across the Precambrian rocks of the crystalline core and subsequently lose all or part of their flow where Paleozoic outcrops are crossed farther downstream (Rahn and Gries, 1973). Large artesian springs occur in many locations downgradient from loss zones, most commonly within or near the outcrop of the Spearfish Formation. These springs provide an important source of base flow in many streams beyond the periphery of the Black Hills (Rahn and Gries, 1973; Miller and Driscoll, 1998).

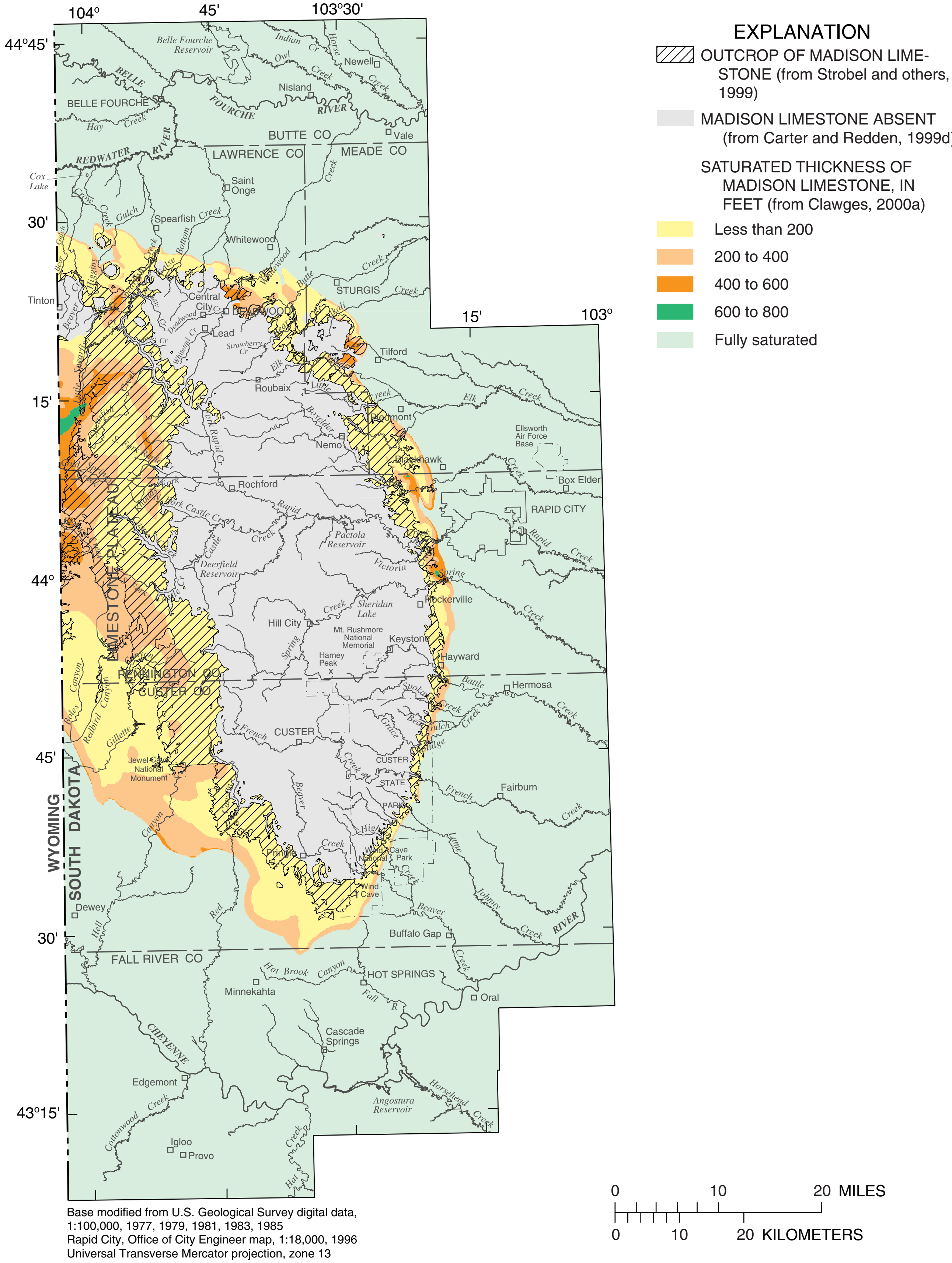

Figure 21. Saturated thickness of the Madison aquifer.

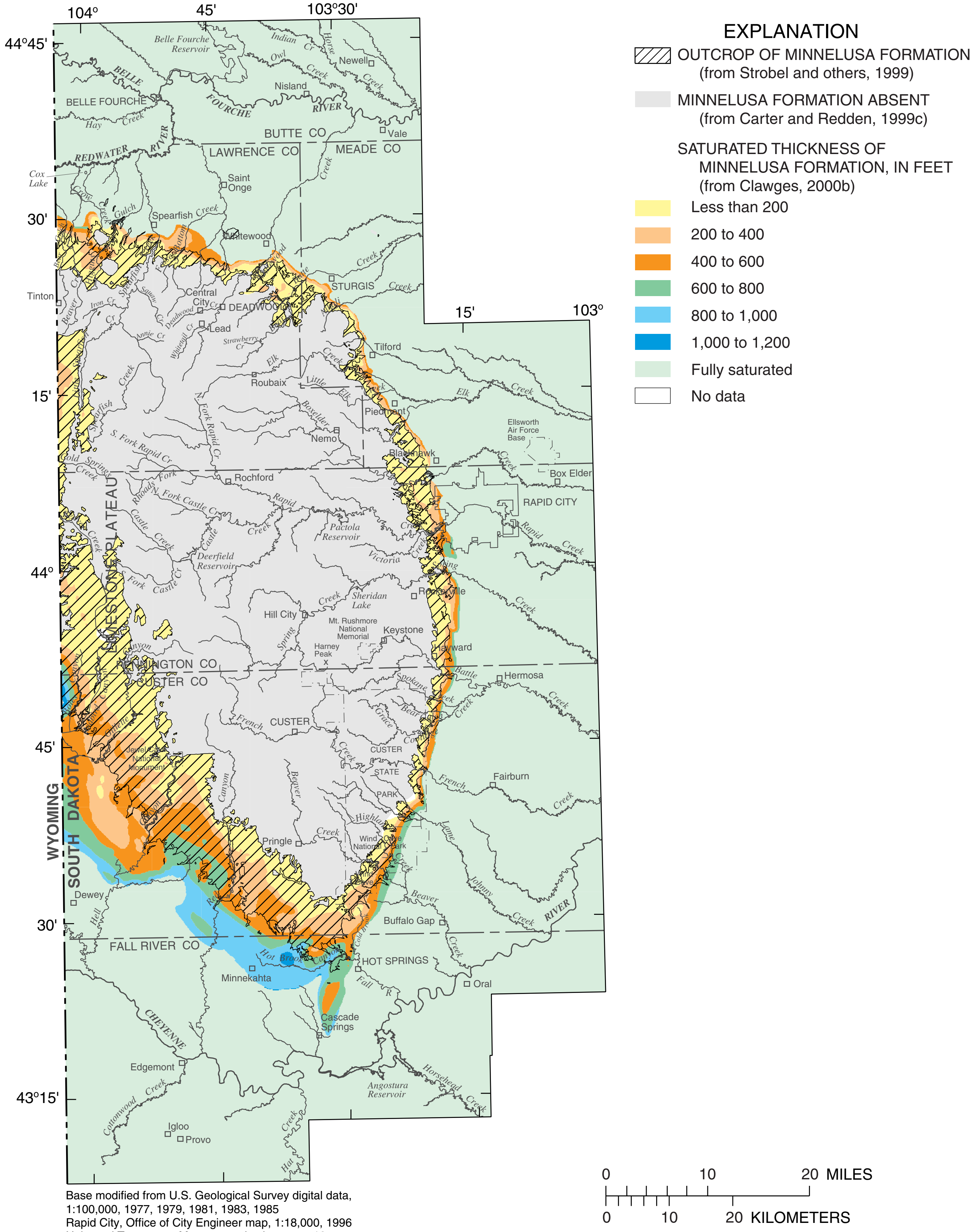

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Figure 22. Saturated thickness of the Minnelusa aquifer.

Five hydrogeologic settings have been identified for the Black Hills area that influence both streamflow (Driscoll and Carter, 2001) and water-quality (Williamson and Carter, 2001) characteristics. These settings are described in the following section, which is followed by sections describing streamflow losses and streamflow regulation, both of which have large influence on many area streams.

Hydrogeologic Settings

The five hydrogeologic settings identified for the Black Hills area include the limestone headwater, crystalline core, loss zone, artesian spring, and exterior settings, which are represented by four areas (fig. 23). The loss zone and artesian spring settings have distinctly different streamflow characteristics but share a common area because many artesian springs are located along stream channels that also are influenced by streamflow losses. Locations of representative streamflow-gaging stations for the five hydrogeologic settings, which are used in subsequent descriptions of streamflow and water-quality characteristics, also are shown in figure 23.

The limestone headwater setting is considered to be within or near the Limestone Plateau area (fig. 23), where large outcrops of the Madison Limestone and Minnelusa Formation occur in a high-altitude area of generally low relief, along the South Dakota-Wyoming border. Most of the limestone headwater springs occur near the eastern edge of the Limestone Plateau in areas where the contact between the Madison Limestone and underlying geologic units (fig. 9) is exposed (fig. 14). Various low-permeability layers in the underlying units can act as confining layers, which result in lateral movement of ground water prior to discharge as springflow. Most recharge for these headwater springs is from infiltration of precipitation on outcrops of the Madison Limestone and Minnelusa Formation. Ground-water discharge from the Deadwood aquifer also can contribute to springflow. Sustained streamflow within the Madison and Minnelusa outcrops is very uncommon (Driscoll and Carter, 2001) and generally occurs only in limited areas where low permeability “perching” layers occur. Small perched springs are common especially within outcrops of the Minnelusa Formation along the Limestone Plateau.

The crystalline core setting consists primarily of igneous and metamorphic Precambrian rocks within the central part of the Black Hills, but also includes numerous Tertiary intrusives in the northern Black Hills (fig. 14). Unconsolidated Quaternary and Tertiary deposits also occur in various locations. Within this setting, ground-water discharge contributes to base flow of many streams; however, base flow can diminish rather quickly during periods of minimal precipitation.

The loss-zone setting consists of areas that are heavily influenced by streamflow losses that occur as streams cross outcrops of the Madison Limestone and Minnelusa Formation. The outer extent of this area is represented by the outer extent of the outcrop of the Inyan Kara Group (fig. 23). This same area is used to represent the artesian spring setting because many artesian springs are located along stream channels that also are influenced by streamflow losses. Most artesian springs are located downgradient from the outcrop of the Minnelusa Formation (fig. 23). Complex interactions between bedrock aquifers, alluvial aquifers, and surface water can occur within this setting.

No artesian springs are known to be located beyond the outcrop of the Inyan Kara Group; the area beyond this outcrop is referred to as the exterior setting. Within this setting, the influence of ground water on streamflow generally is relatively minor or negligible, with the exception of upstream influences from streamflow losses or artesian springs. Many streams also are influenced by irrigation withdrawals or other forms of regulation, as described in a subsequent section.

Streamflow Losses

Streamflow losses influence the flow of most streams that cross Paleozoic outcrops, especially the Madison Limestone and Minnelusa Formation. Most streams lose all of their flow up to some loss threshold. When streamflow exceeds this threshold, flow is maintained through the loss zones. Loss thresholds for most large streams were quantified by Hortness and Driscoll (1998) and are summarized in table 3. Individual loss thresholds range from negligible (no loss) to as much as 50 ft^3/s. Streamflow losses can occur along Iron Creek and Higgins Gulch; however, loss thresholds are noted as zero because these streams receive net ground-water discharge from the Madison and Minnelusa aquifers. Newton and Jenny (1880) observed losses in Whitewood Creek; however, loss zones apparently were sealed by fine-grained mine tailings that have been discharged to the stream (Hortness and Driscoll, 1998).

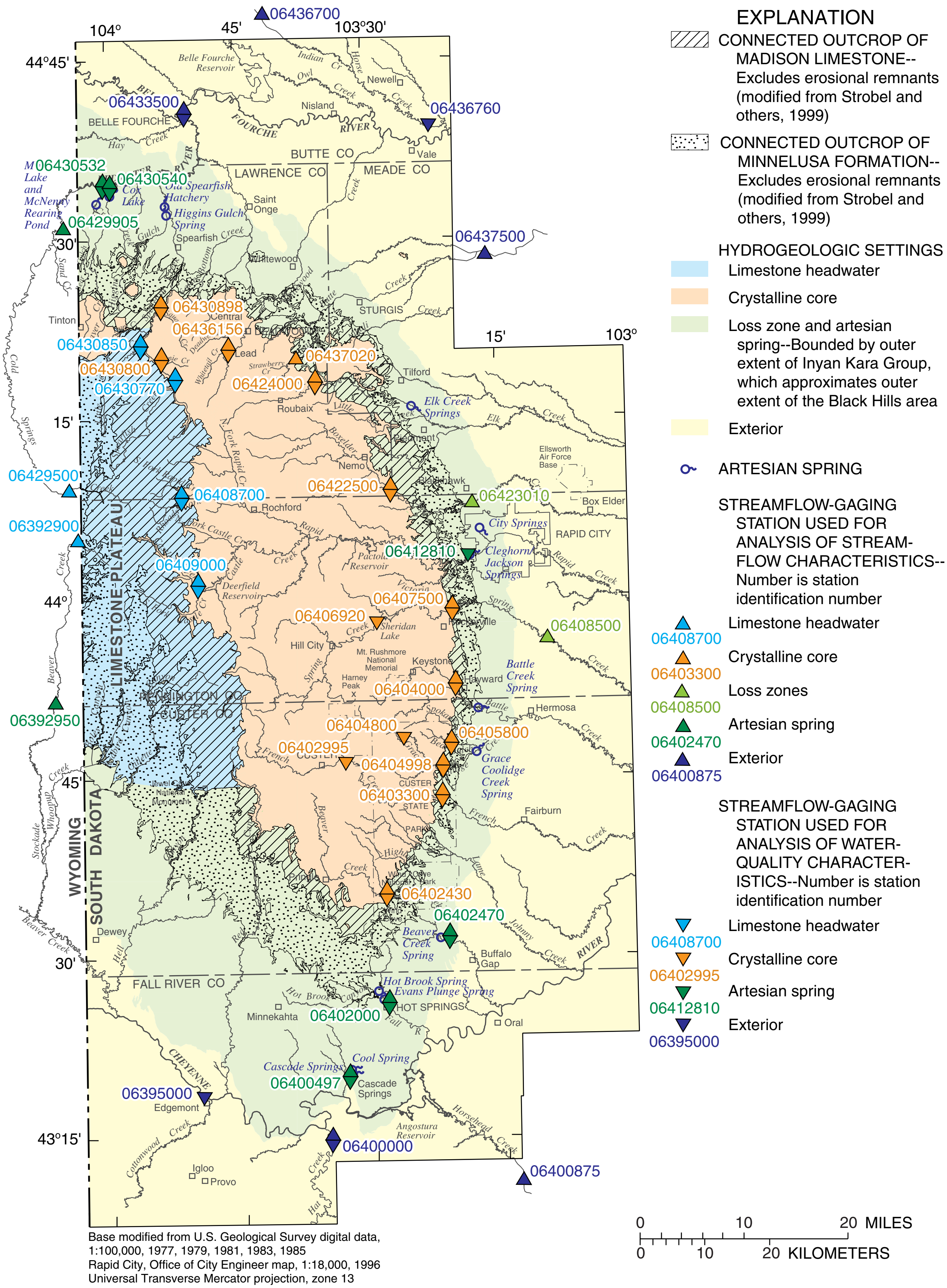


Figure 23. Hydrogeologic settings for the Black Hills area. Locations of streamflow-gaging stations representative of the settings also are shown (from Driscoll and Carter, 2001).

Table 3. Summary of loss thresholds from Black Hills streams to bedrock aquifers

[From Hortness and Driscoll (1998). ft^3/s, cubic feet per second]

Stream name	Approximate loss threshold (ft^3/s)
Beaver Creek[1]	5
Highland Creek	10
South Fork Lame Johnny Creek	1.4
North Fork Lame Johnny Creek	2.3
French Creek	15
Battle Creek	12
Grace Coolidge Creek	21
Bear Gulch[1]	.4
Spokane Creek	2.2
Spring Creek	28
Rapid Creek	10
Victoria Creek	1.0
Boxelder Creek	50
Elk Creek	19
Little Elk Creek	3.3
Bear Gulch[2]	4
Beaver Creek[2]	9
Iron Creek	0
Spearfish Creek	23
Higgins Gulch	0
False Bottom Creek	15
Whitewood Creek	0
Bear Butte Creek	12

[1]Located in southern Black Hills.
[2]Located in northern Black Hills.

Although most losses occur within outcrops of the Madison Limestone and Minnelusa Formation, small losses may occur to other bedrock units. Losses to the Deadwood Formation are minimal. Losses to the Minnekahta Limestone generally are small, relative to losses to the Madison and Minnelusa Formations; however, they are difficult to quantify because of potential losses to extensive alluvial deposits that commonly are located near Minnekahta Limestone outcrops.

Loss thresholds generally are relatively constant, without measurable effects from flow rate or duration of flow through loss zones. Changes in loss thresholds resulting from changes in channel conditions have been documented for Whitewood Creek (previously discussed), Grace Coolidge Creek, and Spring Creek (Hortness and Driscoll, 1998). The loss threshold for Grace Coolidge Creek probably was reduced by deposition of large quantities of fine-grained sediment mobilized after the Galena Fire, which occurred during July 1988. Streamflow losses along Spring Creek apparently were temporarily reduced as a result of efforts to seal the channel with bentonite and rocks during 1937-40 (Brown, 1944).

Streamflow Regulation

Many streams in the study area are affected by withdrawals, diversions, or reservoir regulation. The largest consumptive use of surface water within the study area is withdrawals for irrigation supplies (Amundson, 1998). The largest withdrawals are associated with irrigation projects along Rapid Creek and the Cheyenne and Belle Fourche Rivers, where Bureau of Reclamation storage reservoirs provide reliable water supplies. Angostura Reservoir (fig. 1) supplies the Angostura Unit; Deerfield and Pactola Reservoirs supply the Rapid Valley Project; and Keyhole (located in northeastern Wyoming) and Belle Fourche Reservoirs supply the Belle Fourche Project (Bureau of Reclamation, 1999). Details about these reservoirs, along with storage records through 1993, were reported by Miller and Driscoll (1998).

Large irrigation withdrawals also are made from Beaver Creek near Buffalo Gap and from Spearfish Creek and the Redwater River in the northern Black Hills, where streamflow is sufficiently reliable to provide consistent supplies. Smaller irrigation withdrawals are made from many other area streams.

Streamflow in many area streams is influenced by a variety of other generally non-consumptive diversions and regulation mechanisms (such as smaller reservoirs). Diversions from Rapid, Elk, and Spearfish Creeks have historically provided water for mining operations (Homestake Mining Company) and municipal supplies (Central City, Deadwood, and Lead) in the Whitewood Creek Basin (Miller and Driscoll, 1998). Homestake Mining Company also diverts water from Spearfish Creek for two hydroelectric power plants; these flows are returned to Spearfish Creek below the loss zone. Substantial withdrawals for municipal supplies also are made from Rapid Creek.

HYDROLOGIC PROCESSES AND CHARACTERISTICS

This section describes the characteristics of the ground-water and surface-water resources in the study area, including the response of ground water and streamflow to variations in hydrologic conditions and water-quality characteristics. A brief discussion of hydrologic processes relevant to the Black Hills area is first presented.

Hydrologic Processes

A schematic diagram illustrating hydrologic processes is presented in figure 24. Precipitation falling on the earth's surface generally infiltrates into the soil horizon, unless the soil is saturated or the infiltration capacity is exceeded, in which case overland flow or direct runoff will occur. Some water may be returned from the soil horizon to the land surface through interflow, contributing to relatively short-term increases in streamflow. In the Black Hills area, where potential evaporation generally exceeds precipitation, most water is eventually returned to the atmosphere through evapotranspiration (ET). Water infiltrating beyond the root zone may eventually recharge ground-water systems; however, ground-water discharge (in the form of springflow or seepage) also may contribute to streamflow.

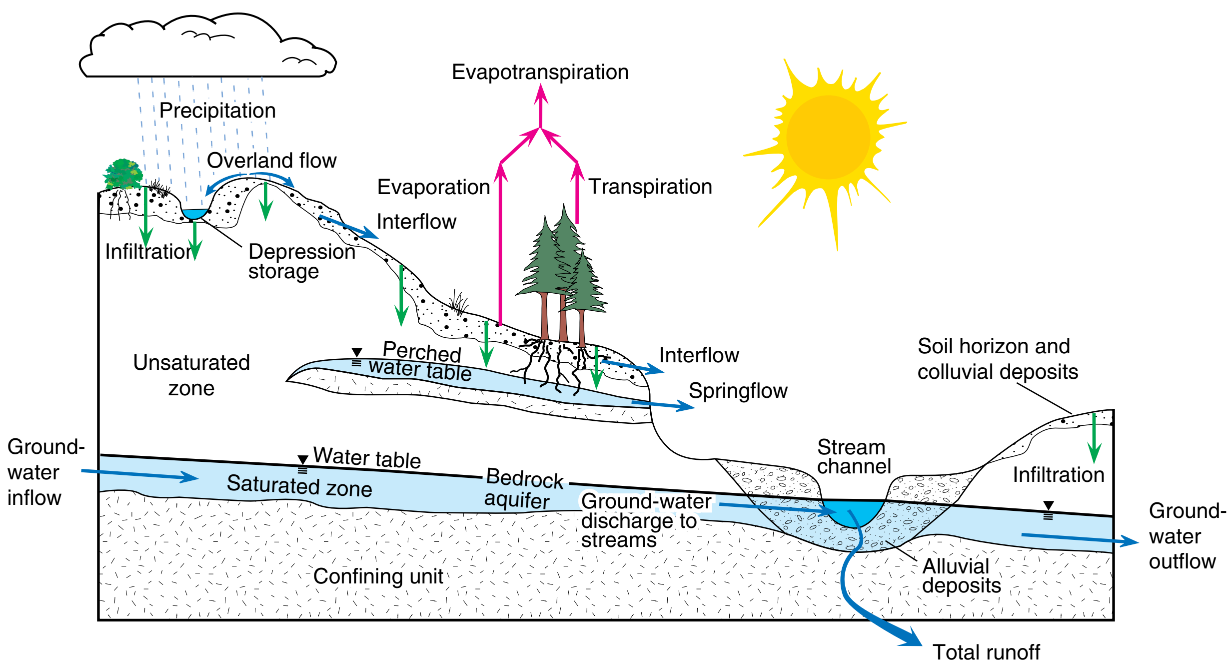


Figure 24. Schematic diagram illustrating hydrologic processes (modified from Driscoll and Carter, 2001).

In this report, the term runoff is used to include all means by which precipitation eventually contributes to streamflow. Direct runoff includes overland flow and that portion of interflow that arrives in stream channels relatively quickly. Base flow generally includes all ground water discharging to streams and also includes some interflow. Springflow is generally considered to be ground-water discharge that occurs in somewhat discrete and identifiable locations, as opposed to more general ground-water seepage. Streamflow is inclusive of runoff and also may include water from other sources such as diversions or well discharges.

Within this report, streamflow is most commonly expressed in units of cubic feet per second, but frequently is expressed in acre-feet per year (1.0 ft^3/s = 724.46 acre-ft for a year consisting of 365.25 days). Units of acre-feet (1.0 ft over an acre, which is equivalent to 43,560 ft^2) are especially convenient for calculating annual yield (annual runoff per unit of area), which generally is expressed in inches.

Ground-Water Characteristics

Water-level trends and comparisons for the major aquifers in the study area are described in this section. In addition, water-quality characteristics for the major aquifers are described, and a brief summary for other aquifers is provided.

Water-Level Trends and Comparisons

Well hydrographs provide information regarding temporal water-level trends, comparisons between aquifers, and water-level response to climatic conditions. Hydrographs (by calendar year) for 49 wells are presented in this section; the location of these wells is shown in figure 25. On these hydrographs, solid lines indicate continuous records and dashed lines indicate periods with discontinuous records, which may be based only on periodic manual measurements in some cases. Hydrographs for 22 additional wells were presented by Driscoll, Bradford, and Moran (2000).

Temporal Trends

Temporal trends in water levels are examined for eight wells with relatively long-term records (fig. 26). Most of these wells are in locations that may be affected by withdrawals from production wells. The Hermosa South Inyan Kara well (fig. 26G), with a steady decline in water level of about 4 ft from 1983 to 1998, is the only observation well in the Black Hills area that shows a steadily declining trend throughout its period of record. The Hermosa West Inyan Kara well (fig. 26F), which is located several miles farther north (fig. 25), does not show a similar decline.

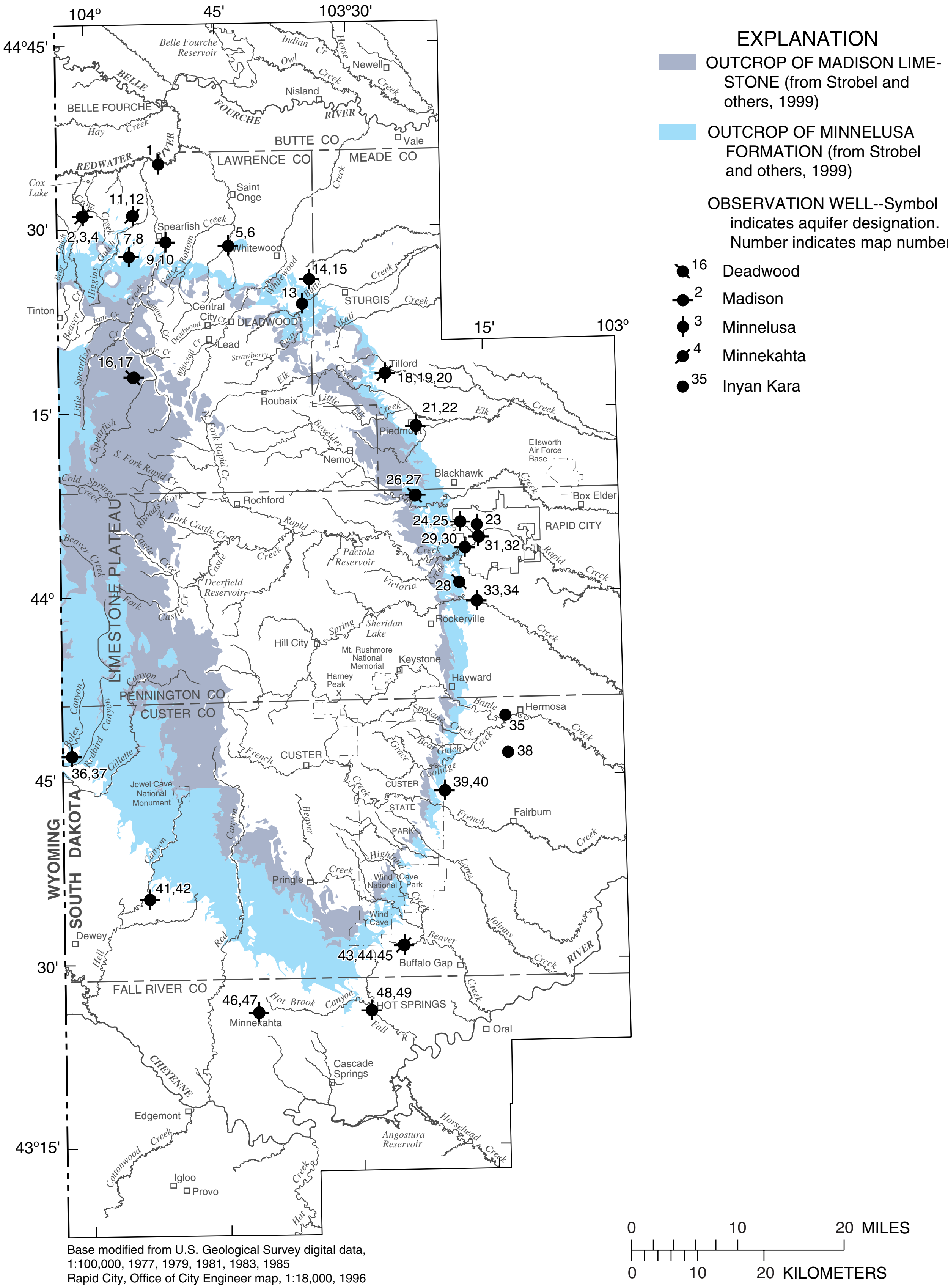
The water level at the Redwater Minnelusa well (fig. 26A) shows a seasonal response to withdrawals for irrigation, but generally recovers each year. The water level at the Boulder Canyon Minnelusa well (fig. 26B) declined steadily during the 1980’s and early 1990’s, but recovered during subsequent years.

The Sioux Park Madison well (fig. 26E) shows response to increased production by the city of Rapid City from the Madison aquifer beginning in the late 1980’s. Recovery occurs during winter months when production is reduced. An adjacent Minnelusa well shows no influence from production from the Madison aquifer; however, a decline during the 1990’s in the Cement Plant Minnelusa well (fig. 26C) may be related to the increased production from the Madison aquifer.

The Countryside Deadwood well (fig. 26D) is located southwest of Rapid City in an area where substantial production from the Deadwood aquifer occurs. Increasing demand in this area occasionally has caused water-supply shortages during recent periods of peak demand; however, long-term water-level declines are not apparent.

Comparisons Between Madison and Minnelusa Aquifers

In many locations, two or more observation wells are colocated. The most common colocated wells are paired Madison and Minnelusa wells, which can provide information regarding interactions between these aquifers. A variety of factors have potential to contribute to reduced competency of confining layers between the Madison and Minnelusa aquifers, which can result in hydraulic connection. Interactions between the Madison and Minnelusa aquifers were investigated by Naus and others (2001) and are discussed in more detail in a subsequent section of this report.



Base modified from U.S. Geological Survey digital data, 1:100,000, 1977, 1979, 1981, 1983, 1985
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Figure 25. Location of observation wells for which hydrographs are presented.

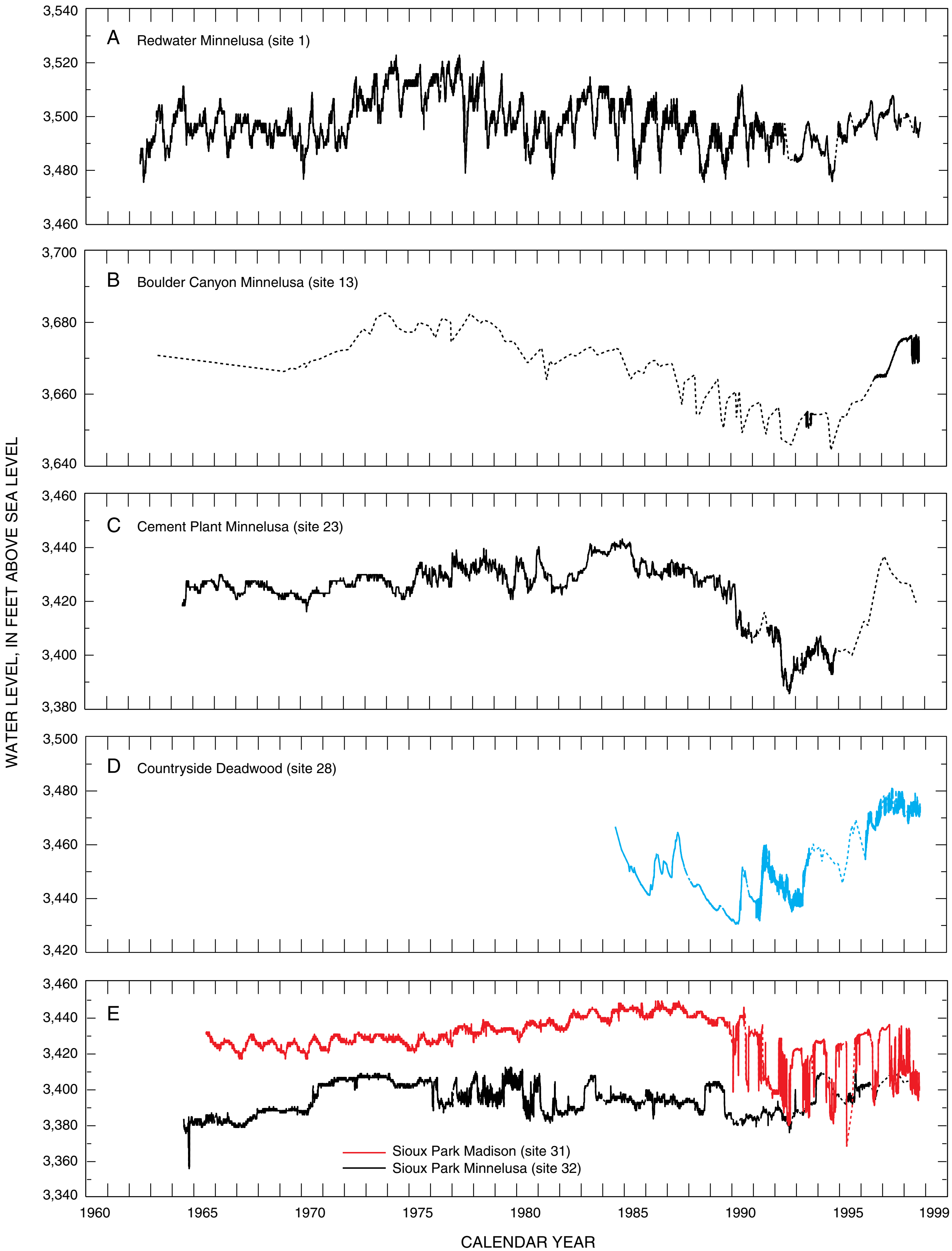


Figure 26. Hydrographs illustrating temporal trends in ground-water levels.

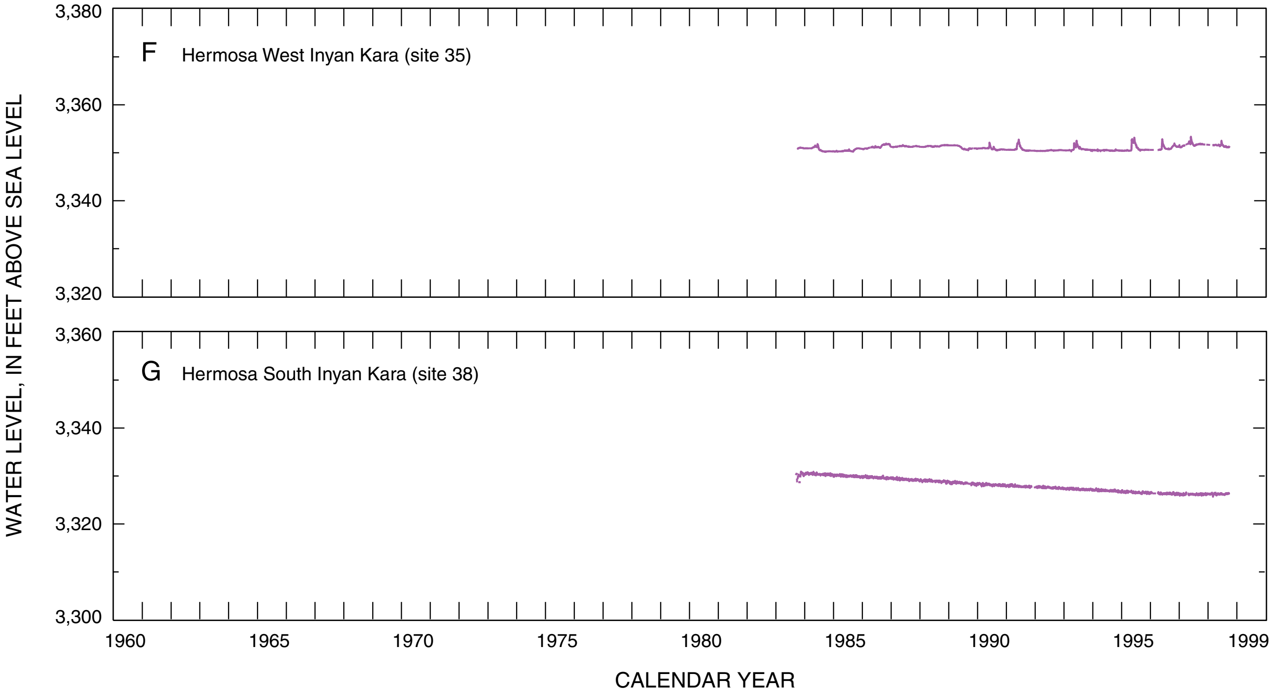


Figure 26. Hydrographs illustrating temporal trends in ground-water levels.—Continued

Hydrographs illustrating general similarities in water levels for some colocated Madison and Minnelusa wells are presented in figure 27. All of the wells are located where confined conditions exist in both aquifers. Hydraulic connection between the aquifers has been confirmed through aquifer testing (Greene, 1993) for the City Quarry wells (fig. 27D), which have hydrographs that are nearly identical. Hydraulic connection in this area also has been confirmed by dye testing (Greene, 1997), which identified a Madison aquifer source for nearby City Springs (fig. 23) that discharges through the Minnelusa Formation.

Similarities in hydrographs do not necessarily indicate hydraulic connection between the aquifers. Hydrographs for the Spearfish Golf Course wells (fig. 27A) are very similar during 1995-98, but have little similarity prior to that period. Aquifer testing (Greene and others, 1999) provided no indication of hydraulic connection in the vicinity of these wells. Hydrographs for the Custer State Park (CSP) wells (fig. 27F) are nearly identical, and other pairs shown have general similarities. At most pairs of wells, hydraulic connection cannot be confirmed or refuted because aquifer testing has not been performed.

Distinct hydraulic separation between the Madison and Minnelusa aquifers is apparent for two well pairs (fig. 28) where water-level altitudes differ by about 500 to 600 ft in locations where unconfined (water-table) conditions occur. In both locations, the water table in the Minnelusa aquifer is much higher than that in the Madison aquifer, which also is not fully saturated. Both well pairs are located within or near the Minnelusa Formation outcrops (fig. 25) and measured water-level altitudes in the Minnelusa aquifer are higher than for most other observation wells in unconfined areas.

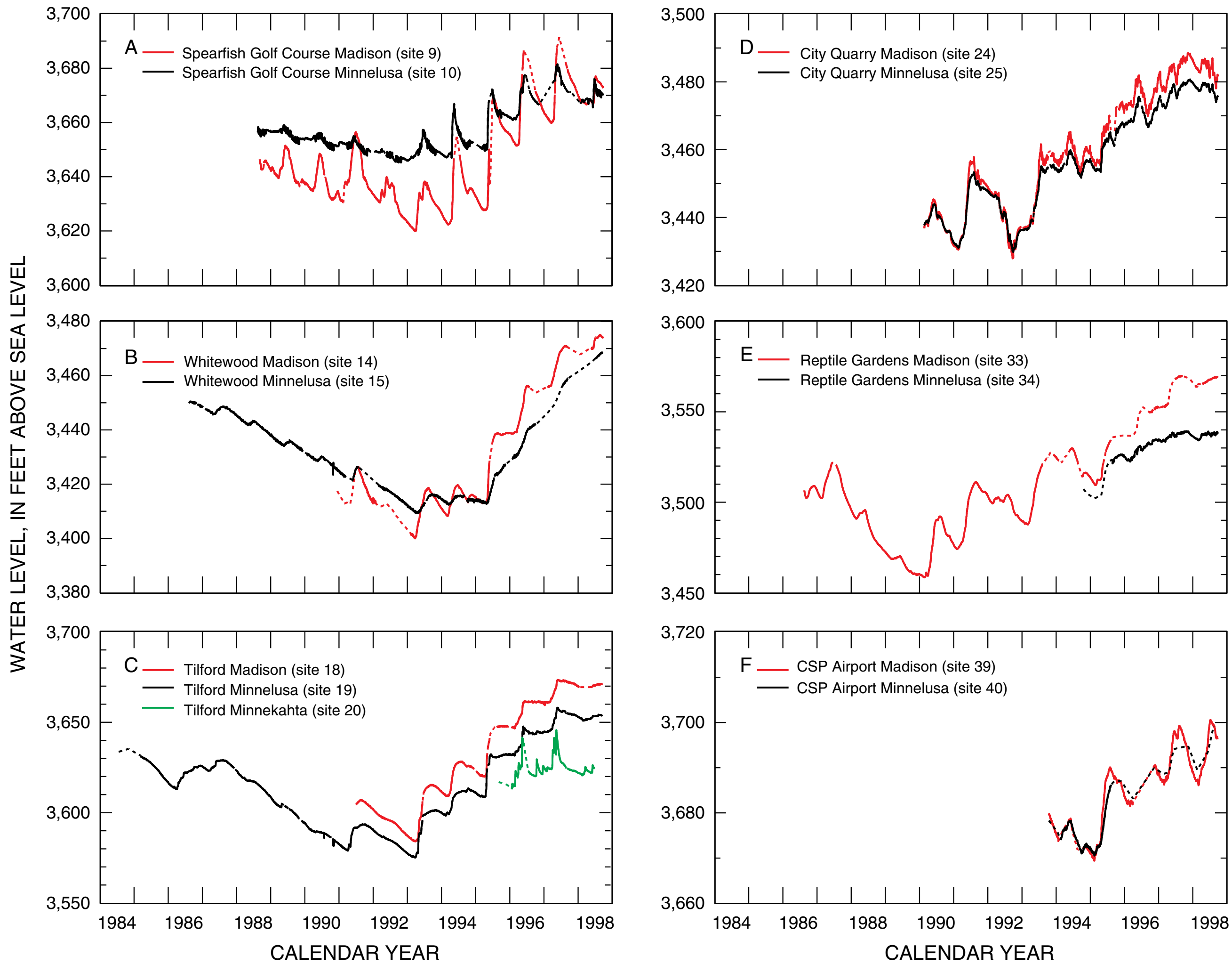


Figure 27. Hydrographs illustrating general similarities in water levels for some colocated Madison/Minnelusa wells with confined conditions.

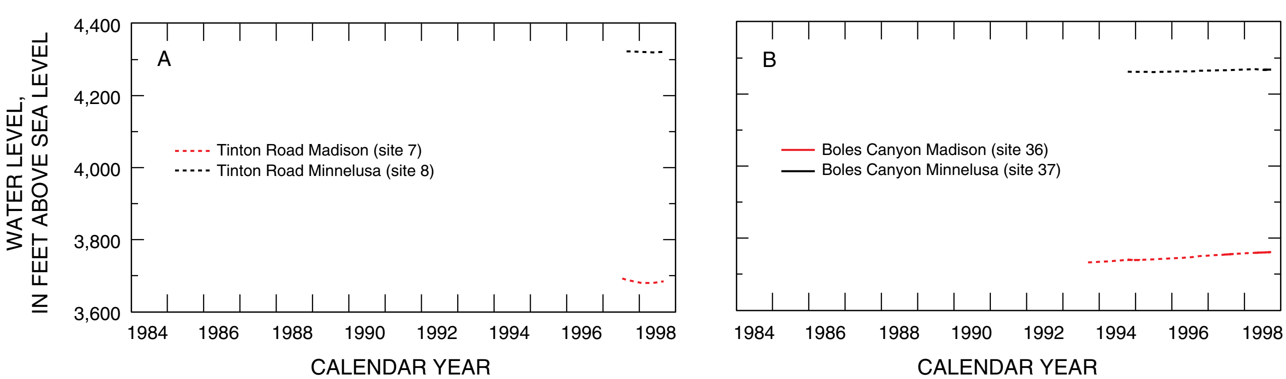


Figure 28. Hydrographs illustrating distinct hydraulic separation for two Madison/Minnelusa well pairs with unconfined conditions.

Figure 29 shows hydrographs for other colocated Madison and Minnelusa wells, most of which are in locations with confined conditions (figs. 21 and 22). Most of these well pairs show distinct hydraulic separation between the aquifers, with hydraulic heads separated by as much as 100 to 150 ft. Hydraulic separation is consistently less than about 30 ft for three well pairs, however—the Frawley Ranch, Hell Canyon, and Minnekahta Junction wells (figs. 29B, 29E, and 29G, respectively). Periods of record for these wells may be insufficient to indicate similarity or dissimilarity of hydrograph shapes.

Hydraulic connection between aquifers does not necessarily mean hydrographs will be similar. The Madison and Minnelusa aquifers probably are connected hydraulically at Cleghorn and Jackson Springs (fig. 23), which are located within the outcrop of the Minnelusa Formation, but probably originate primarily from the Madison aquifer (Naus and others, 2001). Hydrographs for the Canyon Lake wells, which are located about one-quarter mile from the spring complex, show no indication of hydraulic connection, however (fig. 29D). Hydraulic head in the Minnelusa aquifer is about 50 to 60 ft lower than in the Madison aquifer in this area, indicating probable recharge from the Madison aquifer (Driscoll and Carter, 2001). The Minnelusa aquifer apparently is connected hydraulically to Rapid Creek at this location, as evidenced by a sharp water-level decline during a period when Canyon Lake was drained near the end of 1995.

Another observation that can be made from comparisons of hydrographs for paired wells is that the hydraulic head in the Madison aquifer equals or exceeds the hydraulic head in the Minnelusa aquifer in most locations where confined conditions occur. The Madison aquifer has the potential for higher hydraulic head than the Minnelusa aquifer because of generally higher altitude of recharge area for the Madison aquifer. An exception to this generality occurs along the northeastern flank of the Black Hills. The hydraulic head in the Minnelusa aquifer generally equals or exceeds that in the Madison aquifer for the Spearfish Golf Course wells (fig. 27A), the Whitewood wells (fig. 27B), and the Frawley Ranch wells (fig. 29B).

Comparisons for Other Aquifers

Hydrograph comparisons for colocated wells completed in other aquifers are presented in figure 30. Hydrographs for the Spearfish West Minnelusa and Minnekahta wells are shown in figure 30A. Hydrographs for the Minnekahta aquifer also are available for the Tilford wells (fig. 27C), State Line wells (fig. 29A), and 7-11 Ranch wells (fig. 29F). Many artesian springs emerge through the Minnekahta Limestone; thus, hydraulic connections with the underlying Madison and Minnelusa aquifers are possible. Hydrographs for the Minnekahta and Madison wells are very similar for the State Line wells, which are located about 3 mi south of a group of large artesian springs (fig. 23). Hydraulic heads in the Minnelusa and Minnekahta wells are quite similar in the 7-11 Ranch wells, which are located 3 mi west of Beaver Creek Springs.

Hydrographs for colocated Deadwood and Madison wells (fig. 30) are available for two locations. For the Cheyenne Crossing wells (fig. 30B), the water table in the Madison aquifer is about 250 ft higher than the water table in the Deadwood aquifer. For the Doty wells (fig. 30C), the Deadwood aquifer is confined, and hydraulic head is about 200 ft higher than in the Madison aquifer.

Responses to Climatic Conditions

Ground-water levels are directly affected by recharge rates that are influenced by annual precipitation amounts; however, numerous other factors can affect ground-water response. The timing and intensity of precipitation, along with evaporative factors such as temperature, humidity, wind speed, and solar radiation, can have a large effect on annual recharge. Streamflow losses (especially for the Madison and Minnelusa aquifers) also contribute to recharge. Ground-water levels also can be affected by well withdrawals, spring discharges, and various hydraulic properties of aquifers. Hydrographs for many wells in figures 26-30 show a distinct response to annual precipitation patterns (fig. 8); thus, other influences probably are relatively minor for many wells. Many of these wells with sufficient periods of record show short-term declines during the late 1980's, with generally increasing water levels during the wetter conditions of the middle to late 1990's.

Water-level records are not available for the Black Hills area for the prolonged drought conditions that occurred during the 1930's and late 1950's. Cumulative precipitation deficits during these periods were more severe than for the short-term drought conditions that occurred during the late 1980's (fig. 8). Recharge estimates for 1931-98 for the Madison and Minnelusa aquifers (Carter, Driscoll, and Hamade, 2001) indicate that recharge also was minimal during the 1930's and late 1950's; thus, water-level declines exceeding those of the late 1980's probably occurred.

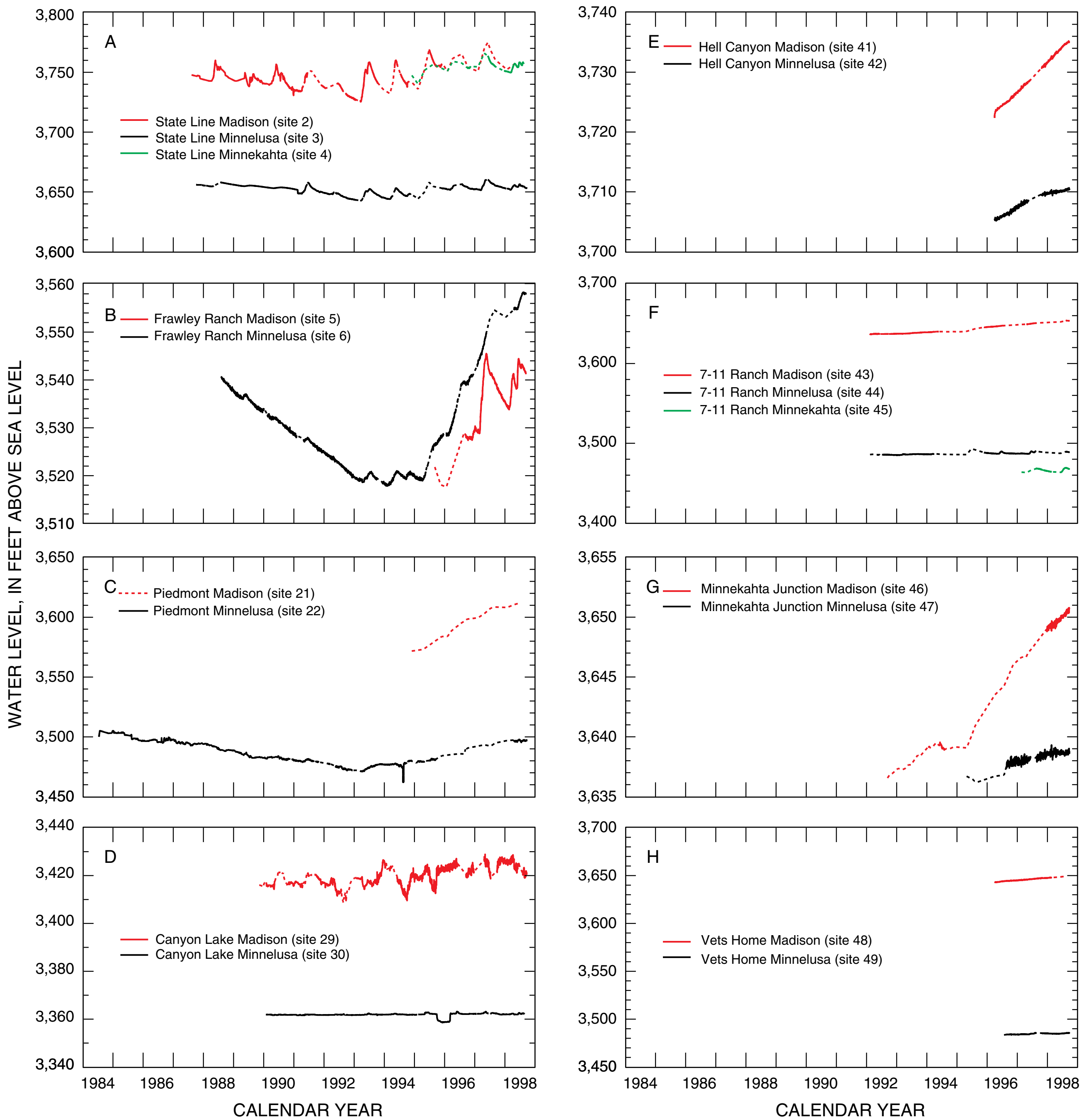


Figure 29. Hydrographs illustrating generally separated water levels for some colocated Madison/Minnelusa wells.

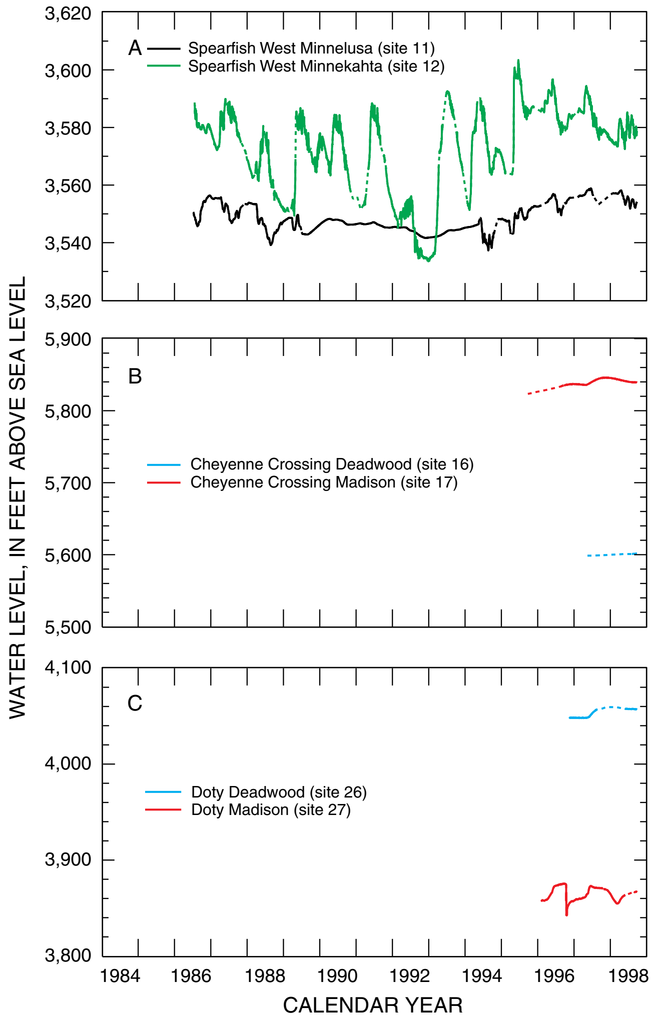


Figure 30. Hydrographs for colocated Minnelusa/Minnekahta and Deadwood/Madison wells.

Hydrographs for the Inyan Kara wells (figs. 26F and G) show minimal response to climatic conditions. Hydrographs for other Inyan Kara wells that are not shown also show minimal response to climatic conditions (Driscoll, Bradford, and Moran, 2000).

All of the other aquifers show a wide range of water-level responses to climatic conditions, ranging from minimal response to several tens of feet. The largest overall water-level change is for the Reptile Gardens Madison well (fig. 27E), which increased by about 110 ft during 1990-98. Increases of about 80 ft have been recorded for the Tilford Madison and Minnelusa wells (fig. 27C).

Driscoll and Carter (2001) noted that for the Madison and Minnelusa aquifers, the smallest water-level fluctuations occur in the extreme southern Black Hills. Smaller recharge than in other areas probably is a contributing factor. Another factor may be large storage capacity in unconfined parts of the aquifers, which are especially extensive in the southern Black Hills (figs. 21 and 22). Caves, which are especially prevalent in the Madison aquifer, probably are more common in the southern Black Hills than in other areas and can provide large storage capacity .

Water Quality

This section includes a summary of water-quality characteristics for the major aquifers and selected minor aquifers in the Black Hills area. More detailed descriptions of ground-water quality are presented by Williamson and Carter (2001), who considered water-quality data collected for the Black Hills Hydrology Study and other studies from October 1, 1930, to September 30, 1998. A brief discussion of the susceptibility of aquifers to contamination also is presented, as well as a summary of water quality relative to water use. Table 4 describes the significance of selected properties and constituents and any related U.S. Environmental Protection Agency (USEPA) water-quality standards for drinking water.

Maximum Contaminant Levels (MCL's) are established for contaminants that, if present in drinking water, may cause adverse human health effects; MCL's are enforceable health-based standards (U.S. Environmental Protection Agency, 1994a). Secondary Maximum Contaminant Levels (SMCL's) are established for contaminants that can adversely affect the taste, odor, or appearance of water and may result in discontinuation of use of the water; SMCL's are nonenforceable, generally non-health-based standards that are related to the aesthetics of water use (U.S. Environmental Protection Agency, 1994a). Action levels, which are concentrations that determine whether treatment requirements may be necessary (U.S. Environmental Protection Agency, 1997), have been established for copper and lead.

Concentrations of constituents were compared by Williamson and Carter (2001) to drinking-water standards set by the USEPA. Although USEPA standards apply only to public-water supplies, individuals using water from private wells should be aware of the potential health risks associated with drinking water that exceeds drinking-water standards.

Table 4. Water-quality criteria, standards, or recommended limits for selected properties and constituents

[All standards are from U.S. Environmental Protection Agency (1994a) unless noted. MCL, Maximum Contaminant Level; SMCL, Secondary Maximum Contaminant Level; USEPA, U.S. Environmental Protection Agency; mg/L, milligrams per liter; µS/cm, microsiemens per centimeter at 25 degrees Celsius; µg/L, micrograms per liter; pCi/L, picocuries per liter; --, no limit established]

Constituent or property	Standard	Significance
Specific conductance	--	A measure of the ability of water to conduct an electrical current; varies with temperature. Magnitude depends on concentration, kind, and degree of ionization of dissolved constituents; can be used to determine the approximate concentration of dissolved solids. Values are reported in microsiemens per centimeter at 25°Celsius.
pH	6.5-8.5 units SMCL	A measure of the hydrogen ion concentration; pH of 7.0 indicates a neutral solution, pH values smaller than 7.0 indicate acidity, pH values larger than 7.0 indicate alkalinity. Water generally becomes more corrosive with decreasing pH; however, excessively alkaline water also may be corrosive.
Temperature	--	Affects the usefulness of water for many purposes. Generally, users prefer water of uniformly low temperature. Temperature of ground water tends to increase with increasing depth to the aquifer.
Dissolved oxygen	--	Required by higher forms of aquatic life for survival. Measurements of dissolved oxygen are used widely in evaluations of the biochemistry of streams and lakes. Oxygen is supplied to ground water through recharge and by movement of air through unsaturated material above the water table (Hem, 1985).
Carbon dioxide	--	Important in reactions that control the pH of natural waters.
Hardness and noncarbonate hardness (as mg/L $CaCO_3$)	--	Related to the soap-consuming characteristics of water; results in formation of scum when soap is added. May cause deposition of scale in boilers, water heaters, and pipes. Hardness contributed by calcium and magnesium, bicarbonate and carbonate mineral species in water is called carbonate hardness; hardness in excess of this concentration is called noncarbonate hardness. Water that has a hardness less than 61 mg/L is considered soft; 61-120 mg/L, moderately hard; 121-180 mg/L, hard; and more than 180 mg/L, very hard (Heath, 1983).
Alkalinity	--	A measure of the capacity of unfiltered water to neutralize acid. In almost all natural waters alkalinity is produced by the dissolved carbon dioxide species, bicarbonate and carbonate. Typically expressed as mg/L $CaCO_3$.
Dissolved solids	500 mg/L SMCL	The total of all dissolved mineral constituents, usually expressed in milligrams per liter. The concentration of dissolved solids may affect the taste of water. Water that contains more than 1,000 mg/L is unsuitable for many industrial uses. Some dissolved mineral matter is desirable, otherwise the water would have no taste. The dissolved solids concentration commonly is called the water's salinity and is classified as follows: fresh, 0-1,000 mg/L; slightly saline, 1,000-3,000 mg/L; moderately saline, 3,000-10,000 mg/L; very saline, 10,000-35,000 mg/L; and briny, more than 35,000 mg/L (Heath, 1983).
Calcium plus magnesium	--	Cause most of the hardness and scale-forming properties of water (see hardness).
Sodium plus potassium	--	Large concentrations may limit use of water for irrigation and industrial use and, in combination with chloride, give water a salty taste. Abnormally large concentrations may indicate natural brines, industrial brines, or sewage.
Sodium-adsorption ratio (SAR)	--	A ratio used to express the relative activity of sodium ions in exchange reactions with soil. Important in irrigation water; the greater the SAR, the less suitable the water for irrigation.
Bicarbonate	--	In combination with calcium and magnesium forms carbonate hardness.

Table 4. Water-quality criteria, standards, or recommended limits for selected properties and constituents–Continued

[All standards are from U.S. Environmental Protection Agency (1994a) unless noted. MCL, Maximum Contaminant Level; SMCL, Secondary Maximum Contaminant Level; USEPA, U.S. Environmental Protection Agency; mg/L, milligrams per liter; µS/cm, microsiemens per centimeter at 25 degrees Celsius; µg/L, micrograms per liter; pCi/L, picocuries per liter; --, no limit established]

Constituent or property	Standard	Significance
Sulfate	250 mg/L SMCL	Sulfates of calcium and magnesium form hard scale. Large concentrations of sulfate have a laxative effect on some people and, in combination with other ions, give water a bitter taste.
Chloride	250 mg/L SMCL	Large concentrations increase the corrosiveness of water and, in combination with sodium, give water a salty taste.
Fluoride	4.0 mg/L MCL 2.0 mg/L SMCL	Reduces incidence of tooth decay when optimum fluoride concentrations present in water consumed by children during the period of tooth calcification. Potential health effects of long-term exposure to elevated fluoride concentrations include dental and skeletal fluorosis (U.S. Environmental Protection Agency, 1994b).
Nitrite (mg/L as N)	1.0 mg/L MCL	Commonly formed as an intermediate product in bacterially mediated nitrification and denitrification of ammonia and other organic nitrogen compounds. An acute health concern at certain levels of exposure. Nitrite typically occurs in water from fertilizers and is found in sewage and wastes from humans and farm animals. Concentrations greater than 1.0 mg/L, as nitrogen, may be injurious to pregnant women, children, and the elderly.
Nitrite plus nitrate (mg/L as N)	10 mg/L MCL	Concentrations greater than local background levels may indicate pollution by feedlot runoff, sewage, or fertilizers. Concentrations greater than 10 mg/L, as nitrogen, may be injurious to pregnant women, children, and the elderly.
Ammonia	--	Plant nutrient that can cause unwanted algal blooms and excessive plant growth when present at elevated levels in water bodies. Sources include decomposition of animal and plant proteins, agricultural and urban runoff, and effluent from waste-water treatment plants.
Phosphorus, orthophosphate	--	Dense algal blooms or rapid plant growth can occur in waters rich in phosphorus. A limiting nutrient for eutrophication since it is typically in shortest supply. Sources are human and animal wastes and fertilizers.
Arsenic	[1]10 µg/L MCL	No known necessary role in human or animal diet, but is toxic. A cumulative poison that is slowly excreted. Can cause nasal ulcers; damage to the kidneys, liver, and intestinal walls; and death. Recently suspected to be a carcinogen (Garold Carlson, U.S. Environmental Protection Agency, written commun., 1998).
Barium	2,000 µg/L MCL	Toxic; used in rat poison. In moderate to large concentrations can cause death; smaller concentrations can cause damage to the heart, blood vessels, and nerves.
Boron	--	Essential to plant growth, but may be toxic to crops when present in excessive concentrations in irrigation water. Sensitive plants show damage when irrigation water contains more than 670 µg/L and even tolerant plants may be damaged when boron exceeds 2,000 µg/L. The recommended limit is 750 µg/L for long-term irrigation on sensitive crops (U.S. Environmental Protection Agency, 1986).
Cadmium	5 µg/L MCL	A cumulative poison; very toxic. Not known to be either biologically essential or beneficial. Believed to promote renal arterial hypertension. Elevated concentrations may cause liver and kidney damage, or even anemia, retarded growth, and death.
Copper	1,300 µg/L (action level)	Essential to metabolism; copper deficiency in infants and young animals results in nutritional anemia. Large concentrations of copper are toxic and may cause liver damage. Moderate levels of copper (near the action level) can cause gastro-intestinal distress. If more than 10 percent of samples at the tap of a public water system exceed 1,300 µg/L, the USEPA requires treatment to control corrosion of plumbing materials in the system.

Table 4. Water-quality criteria, standards, or recommended limits for selected properties and constituents–Continued

[All standards are from U.S. Environmental Protection Agency (1994a) unless noted. MCL, Maximum Contaminant Level; SMCL, Secondary Maximum Contaminant Level; USEPA, U.S. Environmental Protection Agency; mg/L, milligrams per liter; µS/cm, microsiemens per centimeter at 25 degrees Celsius; µg/L, micrograms per liter; pCi/L, picocuries per liter; --, no limit established]

Constituent or property	Standard	Significance
Iron	300 µg/L SMCL	Forms rust-colored sediment; stains laundry, utensils, and fixtures reddish brown. Objectionable for food and beverage processing. Can promote growth of certain kinds of bacteria that clog pipes and well openings.
Lead	15 µg/L (action level)	A cumulative poison; toxic in small concentrations. Can cause lethargy, loss of appetite, constipation, anemia, abdominal pain, gradual paralysis in the muscles, and death. If 1 in 10 samples of a public supply exceed 15 µg/L, the USEPA recommends treatment to remove lead and monitoring of the water supply for lead content (U.S. Environmental Protection Agency, 1991).
Lithium	--	Reported as probably beneficial in small concentrations (250-1,250 µg/L). Reportedly may help strengthen the cell wall and improve resistance to genetic damage and to disease. Lithium salts are used to treat certain types of psychosis.
Manganese	50 µg/L SMCL	Causes gray or black stains on porcelain, enamel, and fabrics. Can promote growth of certain kinds of bacteria that clog pipes and wells.
Mercury (inorganic)	2 µg/L MCL	No known essential or beneficial role in human or animal nutrition. Liquid metallic mercury and elemental mercury dissolved in water are comparatively nontoxic, but some mercury compounds, such as mercuric chloride and alkyl mercury, are very toxic. Elemental mercury is readily alkylated, particularly to methyl mercury, and concentrated by biological activity. Potential health effects of exposure to some mercury compounds in water include severe kidney and nervous system disorders (U.S. Environmental Protection Agency, 1994b).
Nickel	--	Very toxic to some plants and animals. Toxicity for humans is believed to be very minimal.
Selenium	50 µg/L MCL	Essential to human and animal nutrition in minute concentrations, but even a moderate excess may be harmful or potentially toxic if ingested for a long time (Callahan and others, 1979). Potential human health effects of exposure to elevated selenium concentrations include liver damage (U.S. Environmental Protection Agency, 1994b).
Silver	100 µg/L SMCL	Causes permanent bluish darkening of the eyes and skin (argyria). Where found in water is almost always from pollution or by intentional addition. Silver salts are used in some countries to sterilize water supplies. Toxic in large concentrations.
Strontium	--	Importance in human and animal nutrition is not known, but believed to be essential. Toxicity believed very minimal—no more than that of calcium.
Zinc	5,000 µg/L SMCL	Essential and beneficial in metabolism; its deficiency in young children or animals will retard growth and may decrease general body resistance to disease. Seems to have no ill effects even in fairly large concentrations (20,000-40,000 mg/L), but can impart a metallic taste or milky appearance to water. Zinc in drinking water commonly is derived from galvanized coatings of piping.
Gross alpha-particle activity	15 pCi/L MCL	The measure of alpha-particle radiation present in a sample. A limit is placed on gross alpha-particle activity because it is impractical at the present time to identify all alpha-particle-emitting radionuclides due to analytical costs. Gross alpha-particle activity is a radiological hazard. The 15 pCi/L standard also includes radium-226, a known carcinogen, but excludes any uranium or radon that may be present in the sample. Thorium-230 radiation contributes to gross alpha-particle activity.
Beta-particle and photon activity (formerly manmade radionuclides)	4 millirem/yr MCL (under review)	The measure of beta-particle radiation present in a sample. Gross beta-particle activity is a radiological hazard. See strontium-90 and tritium.

Table 4. Water-quality criteria, standards, or recommended limits for selected properties and constituents–Continued

[All standards are from U.S. Environmental Protection Agency (1994a) unless noted. MCL, Maximum Contaminant Level; SMCL, Secondary Maximum Contaminant Level; USEPA, U.S. Environmental Protection Agency; mg/L, milligrams per liter; µS/cm, microsiemens per centimeter at 25 degrees Celsius; µg/L, micrograms per liter; pCi/L, picocuries per liter; --, no limit established]

Constituent or property	Standard	Significance
Radium-226 & 228 combined	5 pCi/L MCL	Radium locates primarily in bone; however, inhalation or ingestion may result in lung cancer. Radium-226 is a highly radioactive alkaline-earth metal that emits alpha-particle radiation. It is the longest lived of the four naturally occurring isotopes of radium and is a disintegration product of uranium-238. Concentrations of radium in most natural waters are usually less than 1.0 pCi/L (Hem, 1985).
Radon[2]	300 or 4,000 pCi/L proposed MCL	Inhaled radon is known to cause lung cancer (MCL for radon in indoor air is 4 pCi/L). Ingested radon also is believed to cause cancer. A radon concentration of 1,000 pCi/L in water is approximately equal to 1 pCi/L in air. The ultimate source of radon is the radioactive decay of uranium. Radon-222 has a half-life of 3.8 days and is the only radon isotope of importance in the environment (Hem, 1985).
Strontium-90 (contributes to beta-particle and photon activity)	Gross beta-particle activity (4 millirem/yr) MCL	Strontium-90 is one of 12 unstable isotopes of strontium known to exist. It is a product of nuclear fallout and is known to cause adverse human health affects. Strontium-90 is a bone seeker and a relatively long-lived beta emitter with a half-life of 28 years. The USEPA has calculated that an average annual concentration of 8 pCi/L will produce a total body or organ dose of 4 millirem/yr (U.S. Environmental Protection Agency, 1997).
Thorium-230 (contributes to gross alpha-particle activity)	15 pCi/L MCL	Thorium-230 is a product of natural radioactive decay when uranium-234 emits alpha-particle radiation. Thorium-230 also is a radiological hazard because it is part of the uranium-238 decay series and emits alpha-particle radiation through its own natural decay to become radium-226. The half-life of thorium-230 is about 80,000 years.
Tritium (^3H) (contributes to beta-particle and photon activity)	Gross beta-particle activity (4 millirem/yr) MCL	Tritium occurs naturally in small amounts in the atmosphere, but largely is the product of nuclear weapons testing. Tritium can be incorporated into water molecules that reach the Earth's surface as precipitation. Tritium emits low energy beta particles and is relatively short-lived with a half-life of about 12.4 years. The USEPA has calculated that a concentration of 20,000 pCi/L will produce a total body or organ dose of 4 millirem/yr (CFR 40 Subpart B 141.16, revised July 1997, p. 296).
Uranium	30 µg/L MCL (under review)	Uranium is a chemical and radiological hazard and carcinogen. It emits alpha-particle radiation through natural decay. It is a hard, heavy, malleable metal that can be present in several oxidation states. Generally, the more oxidized states are more soluble. Uranium-238 and uranium-235, which occur naturally, account for most of the radioactivity in water. Uranium concentrations range between 0.1 and 10 µg/L in most natural waters.

[1]USEPA currently is implementing a revised MCL for arsenic from 50 to 10 µg/L; public-water systems must meet the revised MCL by January 2006 (U.S. Environmental Protection Agency, 2001).

[2]USEPA currently is working to set an MCL for radon in water. The proposed standards are 4,000 pCi/L for States that have an active indoor air program and 300 pCi/L for States that do not have an active indoor air program (Garold Carlson, U.S. Environmental Protection Agency, oral commun., 1999). At this time, it is not known whether South Dakota will participate in an active indoor air program (Darron Busch, South Dakota Department of Environment and Natural Resources, oral commun., 1999).

General Characteristics for Major Aquifers

A summary of water-quality characteristics from Williamson and Carter (2001) for the major aquifers in the study area (Deadwood, Madison, Minnelusa, Minnekahta, and Inyan Kara aquifers) is presented in this section. Characteristics for the Precambrian aquifer also are included in this section because numerous wells are completed in this aquifer in the crystalline core of the Black Hills.

Most pH values for the major aquifers are within the specified range for the SMCL (6.5 to 8.5 standard units). About 13 percent of the samples from wells completed in Precambrian rocks had pH values less than the lower limit specified for the SMCL, which indicates acidity. In general, pH values are lower in wells completed in Precambrian rocks than in the other major aquifers, which is indicative of a unit containing little carbonate material.

Water temperatures generally increase with well depth. The deepest wells in the study area are completed in the Madison aquifer; thus, measured temperatures in the Madison aquifer generally are the warmest of the major aquifers. The Madison aquifer is the primary source of water to warm artesian springs in the southern Black Hills, where water temperatures may be influenced by factors other than aquifer depth (Whalen, 1994).

Williamson and Carter (2001) quantified relations between dissolved solids and specific conductance concentrations for the major aquifers. The r^2 (coefficient of determination) values are high for all of the major aquifers (fig. 31); thus, the equations provided could be used confidently for estimating dissolved solids concentrations from specific conductance measurements.

Specific conductance generally is low in water from the Precambrian, Deadwood, and Minnekahta aquifers. Dissolved constituents tend to increase with residence time as indicated by the general increase in specific conductance in the Madison aquifer with distance from the outcrop (fig. 32). Generally, water from the Inyan Kara aquifer is high in specific conductance even in some outcrop areas and is higher in specific conductance than the other major aquifers due to greater amounts of shale within the Inyan Kara Group. Water obtained from shales may contain rather high concentrations of dissolved solids (Hem, 1985) and, hence, high specific conductance.

Geologic units that contain little carbonate material, such as the Precambrian rocks, generally contain water with lower carbonate hardness and alkalinity than geologic units that are composed primarily of carbonate rocks. Water in the Madison, Minnelusa, and Minnekahta aquifers generally is hard to very hard because these units consist primarily of carbonate rocks. Water in the Deadwood aquifer also is hard to very hard because this unit consists primarily of sandstone with a calcium carbonate cement. The Inyan Kara aquifer may yield soft water, with hardness generally decreasing with increasing distance from the outcrop (fig. 33). Although concentrations of dissolved solids in the Inyan Kara aquifer actually increase with increasing distance from the outcrop, hardness decreases because calcium and bicarbonate are replaced by sodium and sulfate as water moves downgradient.

In the Black Hills area, water from the major aquifers generally is low in dissolved solids in and near outcrop areas. The Madison, Minnelusa, and Inyan Kara aquifers may yield slightly saline water (dissolved solids concentrations between 1,000 and 3,000 mg/L) at distance from the outcrops, especially in the southern Black Hills. The water in these aquifers generally is highly mineralized outside of the Black Hills area, as previously described and shown in figure 17 for aquifers in the Paleozoic units.

Many of the major aquifers yield a calcium bicarbonate type water in and near outcrop areas, with concentrations of sodium, chloride, and sulfate increasing with distance from outcrops. High concentrations of sodium, chloride, and sulfate occur in the Madison aquifer (fig. 34) in the southwestern part of the study area relative to the rest of the study area. These high concentrations could be due to long residence times, long flowpaths associated with regional flow from the west (Wyoming), or greater amounts of evaporite minerals, such as anhydrite and gypsum, available for dissolution (Naus and others, 2001). In the southern part of the study area, the common-ion chemistry of the water in the Minnelusa aquifer also is characterized by higher concentrations of sodium and chloride (fig. 35). The high chloride concentrations in this area could reflect hydraulic connection between the Madison and Minnelusa aquifers (Naus and others, 2001). The dissolution of evaporite minerals and long residence time also are possible explanations for the occurrence of this water type in the Minnelusa aquifer (Naus and others, 2001).

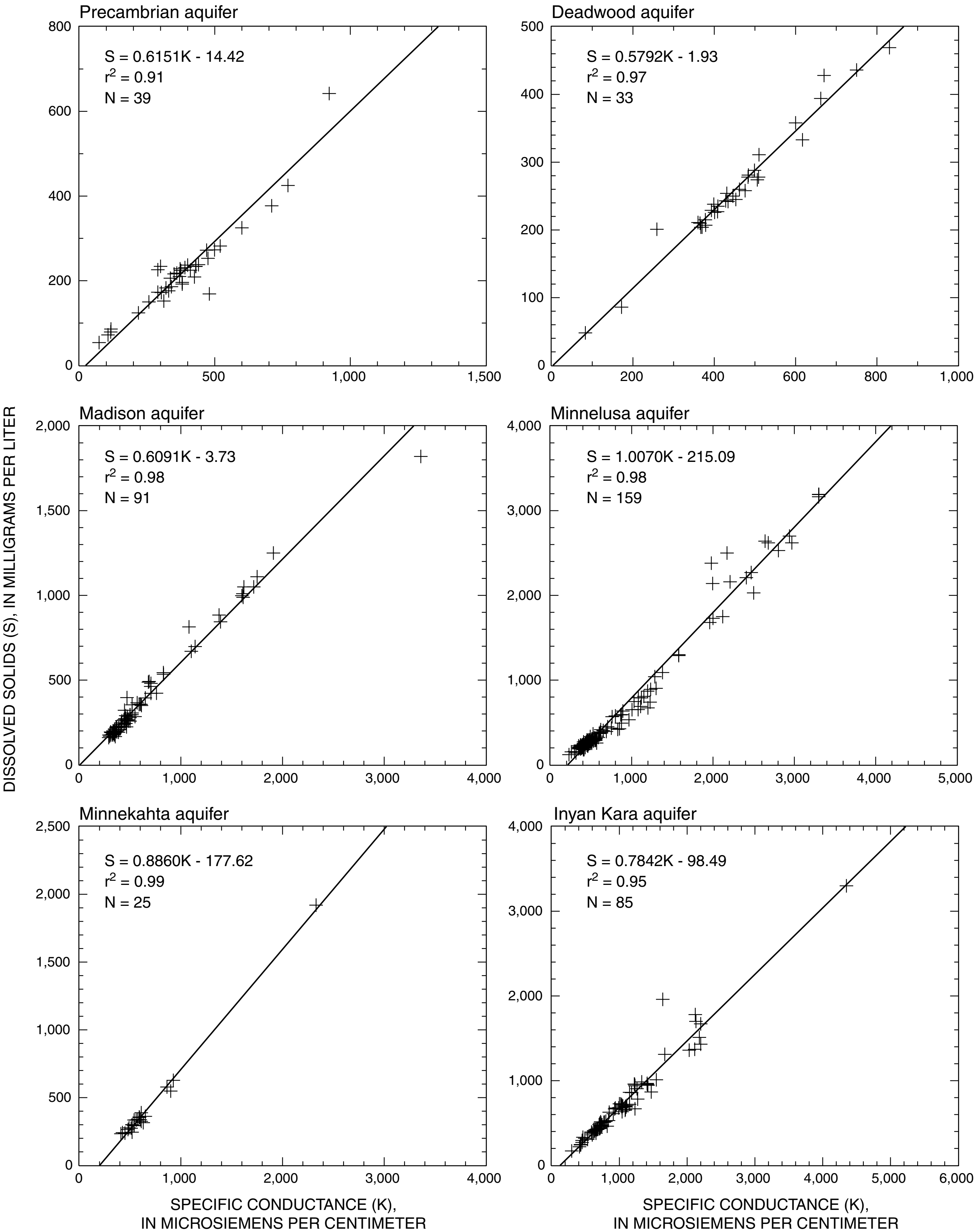


Figure 31. Relations between dissolved solids and specific conductance for the major aquifers.

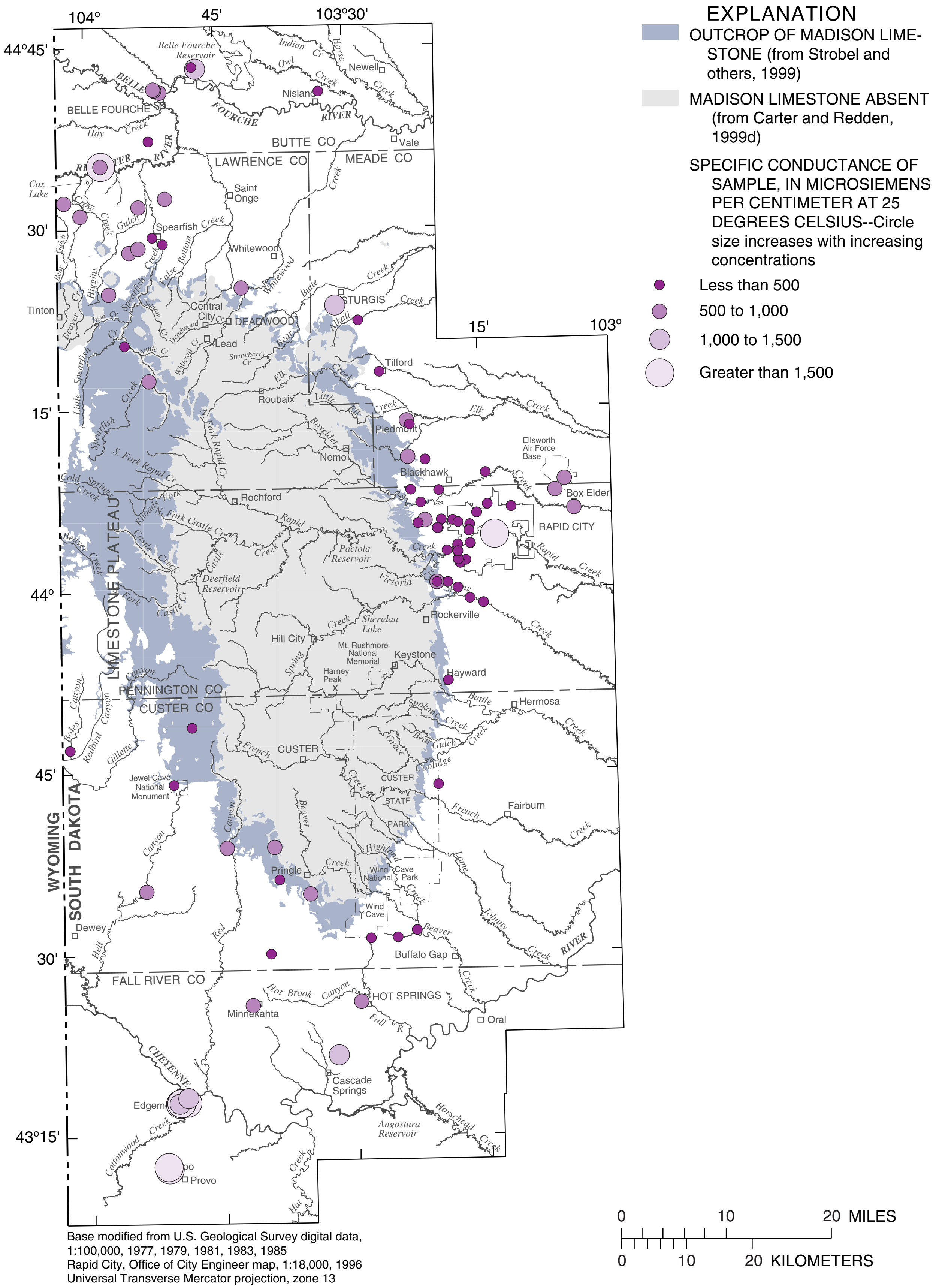


Figure 32. Distribution of specific conductance in the Madison aquifer (modified from Williamson and Carter, 2001).

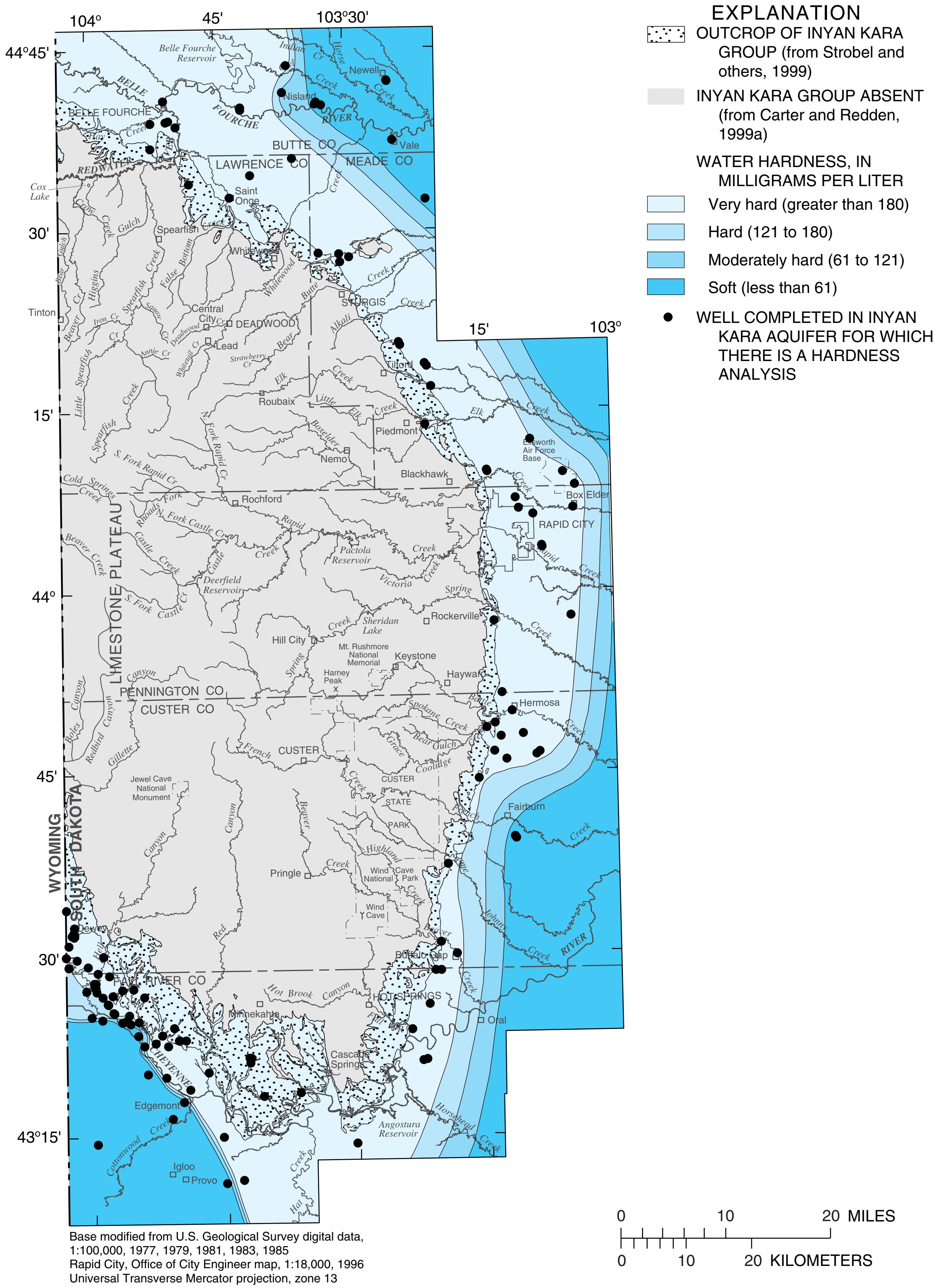


Figure 33. Distribution of hardness in the Inyan Kara aquifer (modified from Williamson and Carter, 2001).

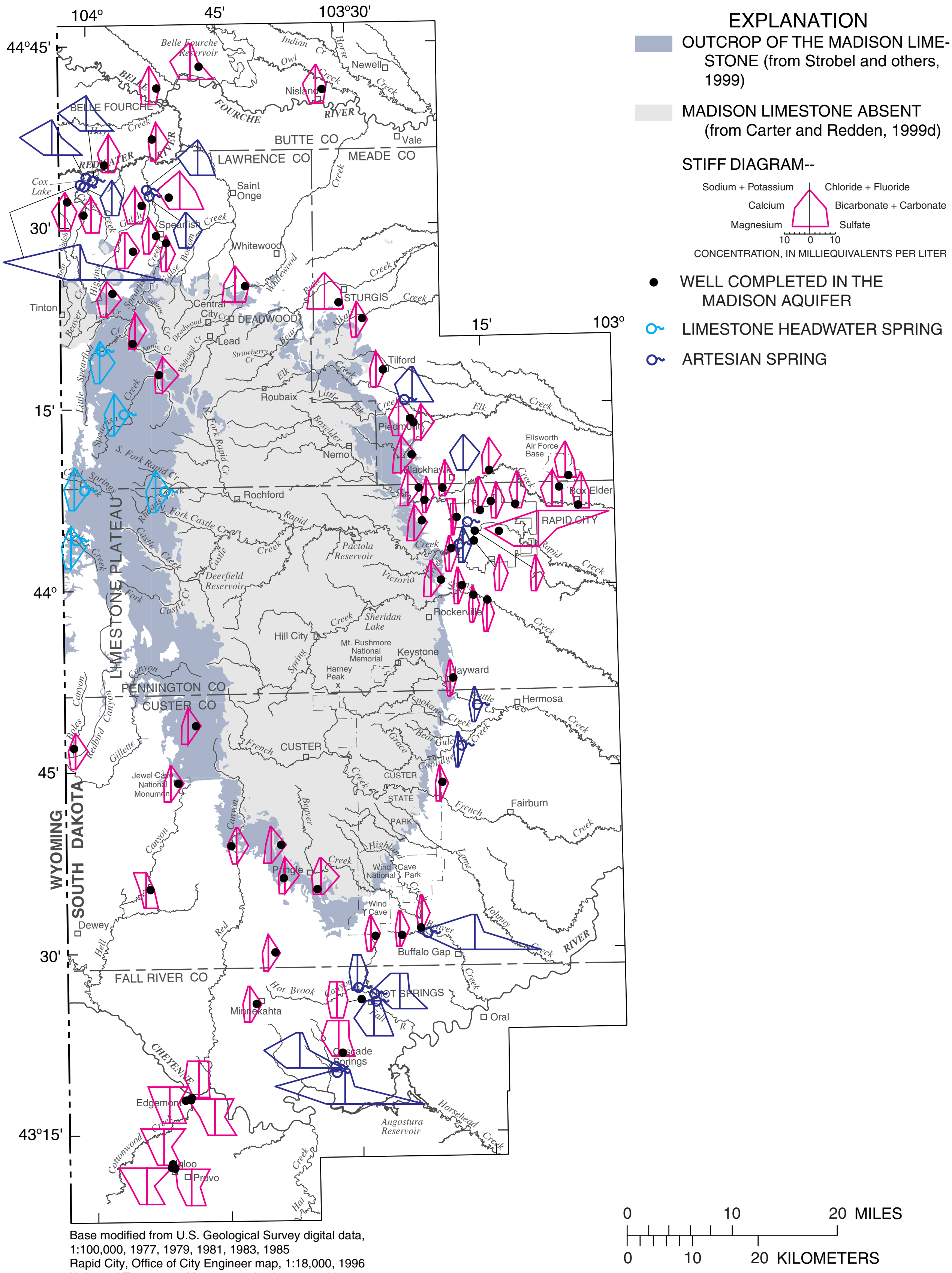


Figure 34. Stiff diagrams (Stiff, 1951) showing the distribution of common-ion chemistry in the Madison aquifer (from Naus and others, 2001).

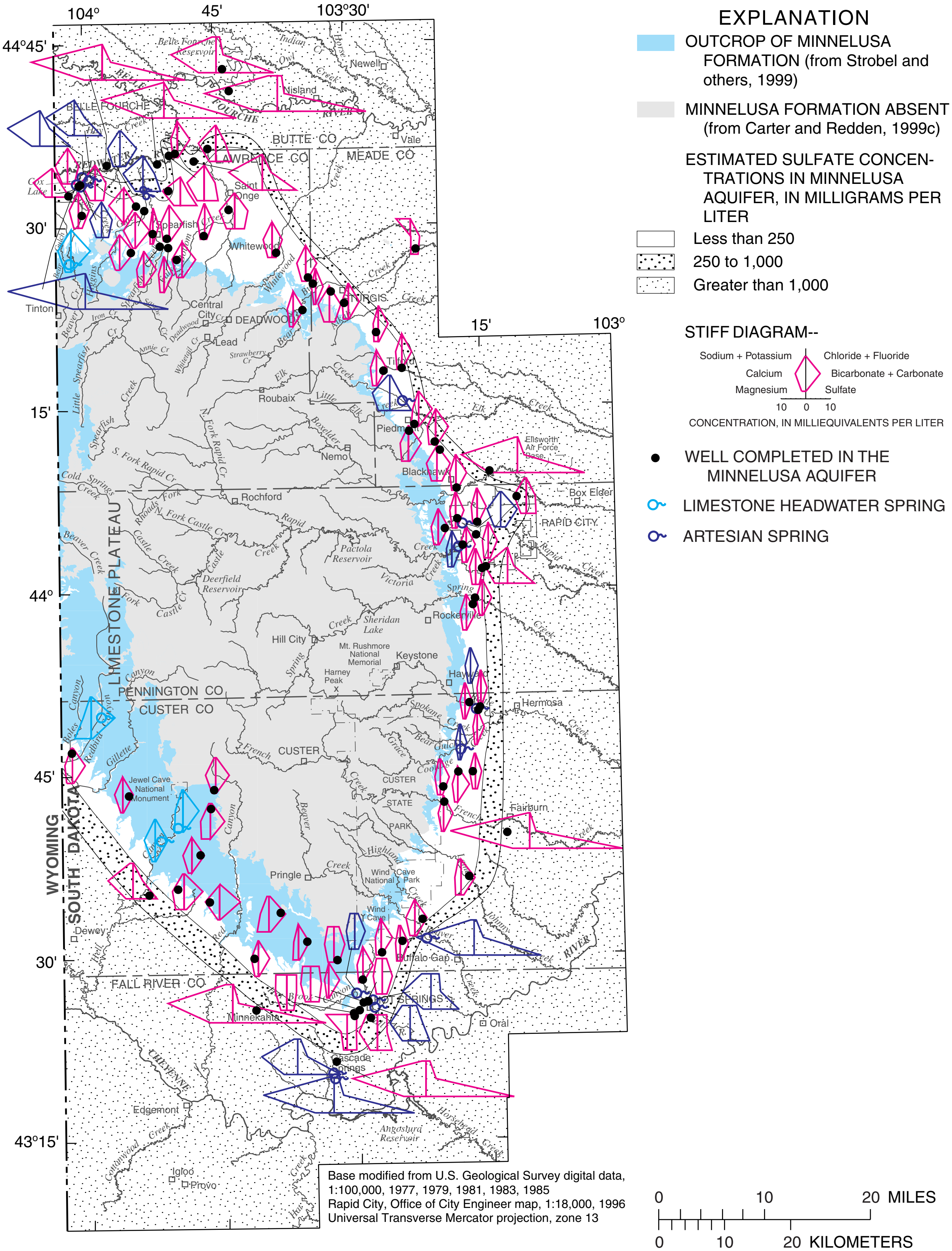


Figure 35. Stiff diagrams (Stiff, 1951) showing the distribution of common-ion chemistry in the Minnelusa aquifer. Approximation location of anhydrite dissolution front showing transition between low and high sulfate concentrations also is shown (from Naus and others, 2001).

Sulfate concentrations in the Minnelusa aquifer are dependent on the amount of anhydrite present in the Minnelusa Formation. Near the outcrop, sulfate concentrations generally are low (less than 250 mg/L) because anhydrite has been removed by dissolution. An abrupt increase in sulfate concentrations occurs downgradient, where a transition zone surrounds the core of the Black Hills. This transition zone is the area within which the sulfate concentrations range from 250 to 1,000 mg/L (fig. 35) and marks an area of active removal of anhydrite by dissolution. Downgradient from the transition zone, sulfate concentrations are greater than 1,000 mg/L, which delineates a zone in which thick anhydrite beds remain in the formation. The transition zone probably is shifting downgradient over geologic time as the anhydrite in the formation is dissolved (Kyllonen and Peter, 1987).

Figures 34 and 35 also show Stiff diagrams (Stiff, 1951) for artesian springs in the Black Hills area, most of which probably originate from the Madison and/or Minnelusa aquifers (Naus and others, 2001). Artesian springs with high sulfate concentrations probably are influenced by anhydrite in the Minnelusa Formation. Artesian springs with low sulfate concentrations occur only upgradient from the transition zone (fig. 35). Additional discussions regarding potential sources of artesian springs are presented in subsequent sections of the report.

Concentrations and variability of many trace elements are small in the major aquifers. Strontium generally has higher concentrations than other trace elements, but is not harmful. Similarly, barium, boron, iron, manganese, lithium, and zinc concentrations also may be high in comparison to other trace elements.

Most naturally occurring radionuclides in water are the result of radioactive decay of uranium-238, thorium-232, and uranium-235, with uranium-238 producing the greatest part of the radioactivity observed (Hem, 1985). In general, gross alpha-particle activity, gross-beta activity, and radium-226 concentrations, are higher in the Deadwood and Inyan Kara aquifers than in the Madison, Minnelusa, and Minnekahta aquifers.

In the Deadwood aquifer, more than 30 percent of the samples analyzed for radium-226 or radium-226 and radium-228 exceeded the MCL of 5 pCi/L for the combined radium-226 and radium-228 standard. Almost 90 percent of the samples from the Deadwood aquifer exceeded the proposed MCL of 300 pCi/L for radon in States without an active indoor air program; several of these samples (fig. 36) also exceeded the proposed MCL of 4,000 pCi/L for radon in States with an active indoor air program. Samples from the Deadwood aquifer have lower uranium concentrations relative to the other major aquifers, which may be due to the reducing conditions of the Deadwood aquifer (Rounds, 1991).

Uranium deposits have been mined in the Inyan Kara Group in the southern Black Hills. Uranium may be introduced into the Inyan Kara Group through upward leakage of water from the Minnelusa aquifer (Gott and others, 1974). As water in the Inyan Kara aquifer migrates downgradient, geochemical conditions favor the precipitation of uranium (Gott and others, 1974). Some water from the Inyan Kara aquifer, especially in the southern Black Hills, contains relatively high concentrations of radionuclides. Almost 20 percent of the samples collected from the Inyan Kara aquifer exceeded the MCL for the combined radium-226 and radium-228 standard; all but one of these samples exceeding the MCL were from wells in the southern Black Hills. About 4 percent of the samples exceeded the MCL for uranium; all these samples exceeding the MCL were from wells located in the southern Black Hills.

General Characteristics for Minor Aquifers

Water-quality characteristics were summarized by Williamson and Carter (2001) for various minor aquifers. The minor aquifers in the study area include the Newcastle aquifer and alluvial aquifers. Local aquifers do exist in the various semiconfining and confining units. Water-quality data also were summarized for several of these local aquifers, which included the Spearfish, Sundance, Morrison, Graneros, and Pierre aquifers.

Relations between dissolved solids and specific conductance concentrations are presented in figure 37 for the minor aquifers with sufficient data, which include the Sundance, Morrison, Newcastle, and alluvial aquifers. The r^2 values are consistently high, indicating strong correlations for these aquifers.

Water in many of the minor aquifers can be very hard and high in dissolved solids concentrations. Most samples from the Sundance aquifer indicate slightly saline water. Sulfate concentrations also can be high in the minor aquifers, such as the Spearfish aquifer where high sulfate concentrations can result from dissolution of gypsum. Both dissolved solids and sulfate concentrations are low in the Newcastle aquifer. A variety of water types can occur within and among the minor

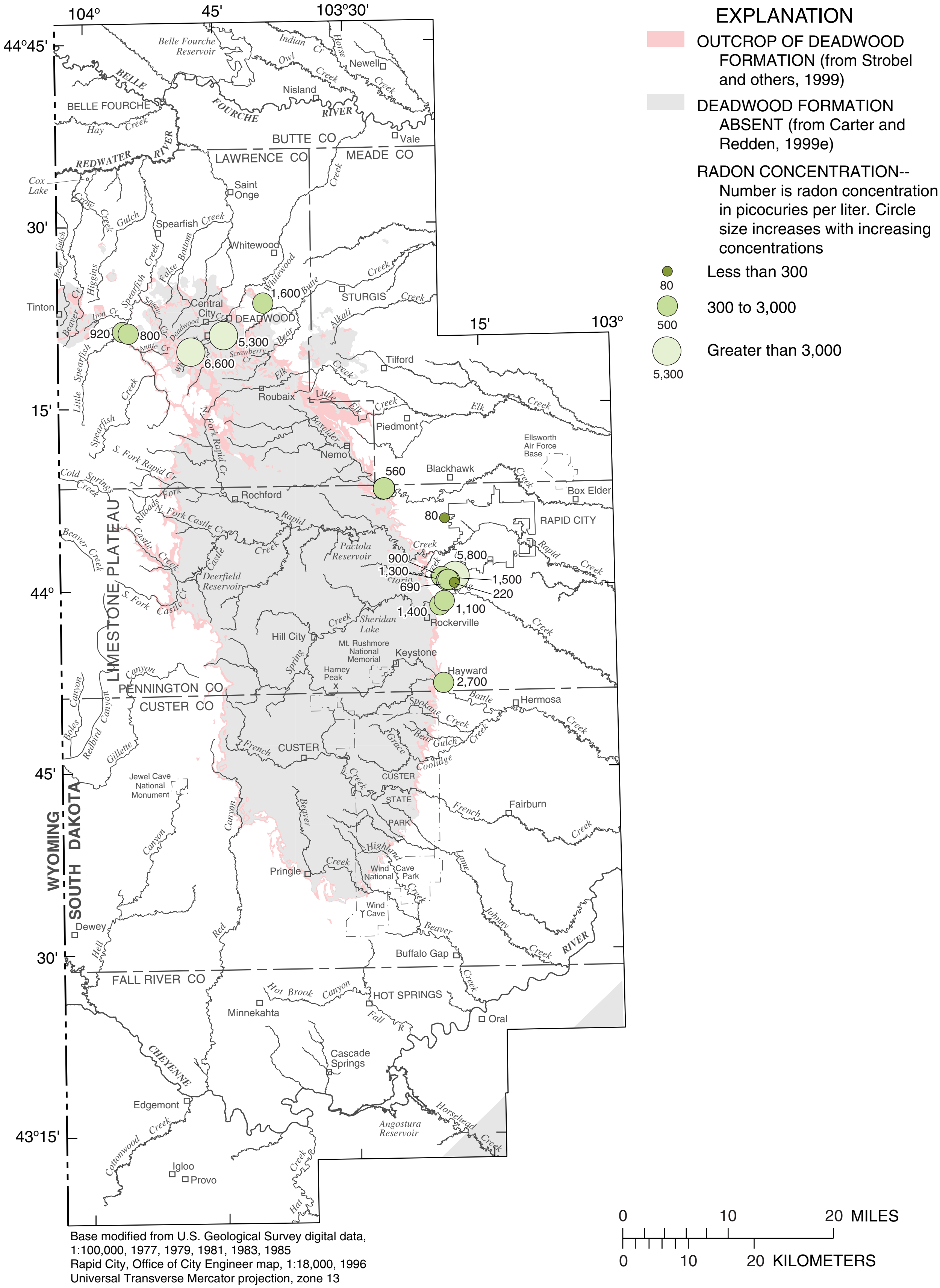


Figure 36. Distribution of radon concentrations in the Deadwood aquifer (from Williamson and Carter, 2001).

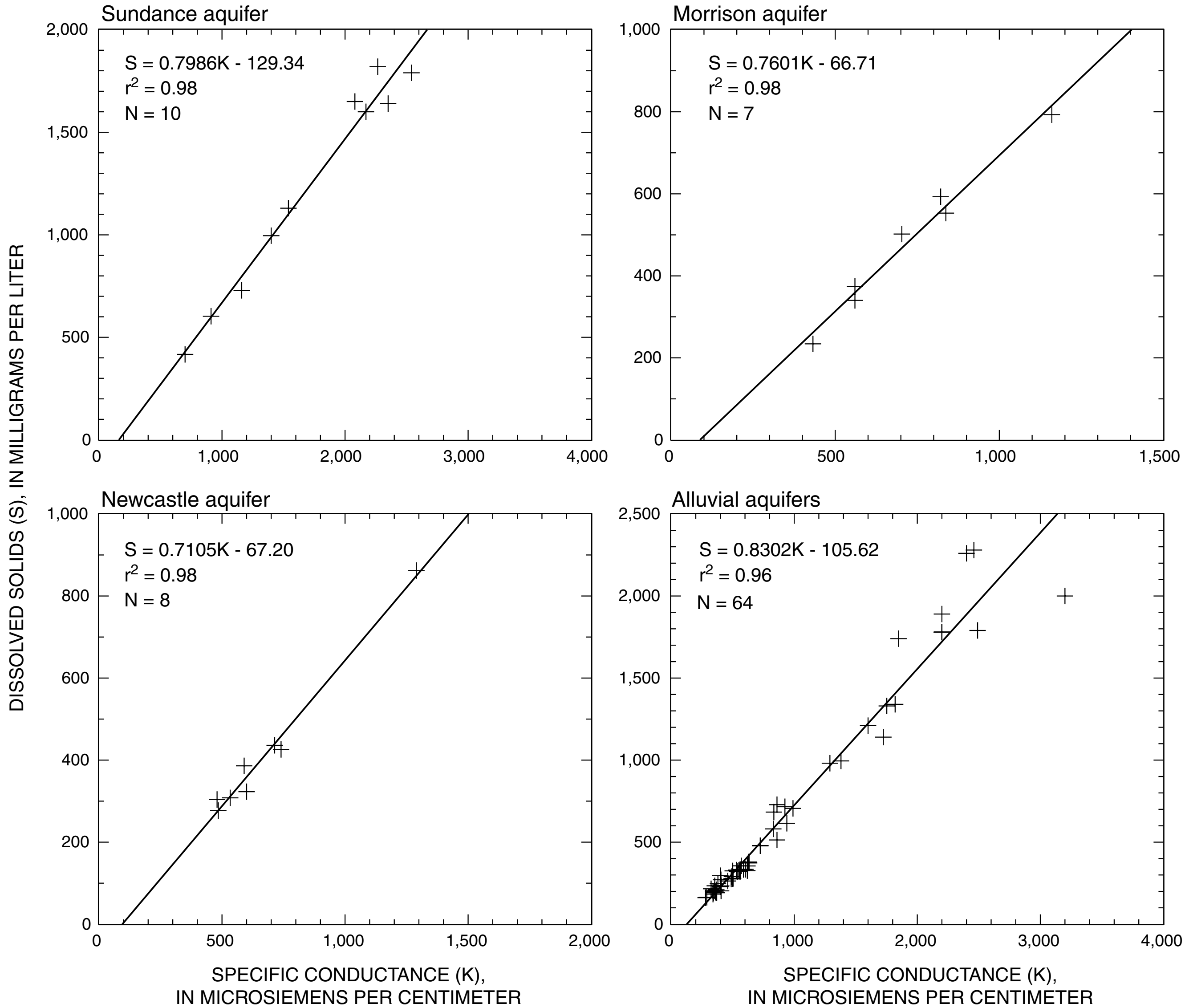


Figure 37. Relations between dissolved solids and specific conductance for the minor aquifers.

aquifers. In general, the dominance of sodium and sulfate increases with increasing amounts of shale present in the formations due to the large cation-exchange capabilities of clay minerals (generally sodium concentrations increase) and due to the reduced circulation of water through the shale (Hem, 1985). The dominance of calcium, magnesium, and bicarbonate increases with increasing amounts of sandstone (where calcium carbonate commonly is the cementing material) and carbonate rocks present in the geologic units. The Sundance aquifer has the highest selenium concentrations of all aquifers considered in this report.

Concentrations of common ions in alluvial aquifers generally increase with increasing distance from the core of the Black Hills, which is largely due to contact of the water with underlying geologic units and to the composition of alluvial deposits. Wells completed in alluvial deposits that do not overlie Cretaceous shales generally yield fresh water of a calcium bicarbonate or calcium magnesium bicarbonate type. Wells that are completed in alluvial deposits that overlie the Cretaceous shales generally yield slightly saline water in which sodium and/or sulfate is dominant. Water from alluvial aquifers may be high in uranium concentrations, especially in the southern Black Hills. About 17 percent of the samples exceeded the proposed MCL for uranium, and all samples exceeding this MCL were from wells in the southern Black Hills.

Susceptibility to Contamination

The Black Hills Hydrology Study focused primarily on determination of natural water-quality characteristics, and investigation of contamination potential was not an objective of the study. The susceptibility of the aquifers to contamination in the study area is an important issue, however, and can be addressed to some extent.

Nitrite plus nitrate concentrations for various aquifers (fig. 38) can provide a general indication of possible human influence. Although nitrogen is essential for plant growth, high concentrations of nitrite plus nitrate can cause excessive plant growth and can be harmful to livestock and humans. Excessive concentrations of nitrite plus nitrate in drinking water are a health concern for pregnant women, children, and the elderly (may cause methemoglobinemia (blue-baby syndrome) in small children). Nitrite plus nitrate in ground water can originate from natural processes or as contamination from nitrogen sources, such as fertilizers and sewage, on the land surface or in the soil zone. Nitrite plus nitrate concentrations for most samples in the Black Hills area generally are low (fig. 38); however, samples approaching or exceeding the national nitrate background concentration of 2.0 mg/L (U.S. Geological Survey, 1999) may provide indications of possible human influence in a variety of land-use settings. The extreme values for nitrite plus nitrate in figure 38 are unusually high and may reflect poor well construction and surface contamination as opposed to aquifer conditions.

The potential for contamination of ground water in the Black Hills area can be large because many aquifer outcrops can be subject to various forms of land development. Rapid ground-water velocities also are possible in many aquifers because of high secondary permeability. Contamination of ground water by septic tanks has been documented for wells in the Blackhawk, Piedmont, and Sturgis areas (Bartlett and West Engineers, Inc., 1998).

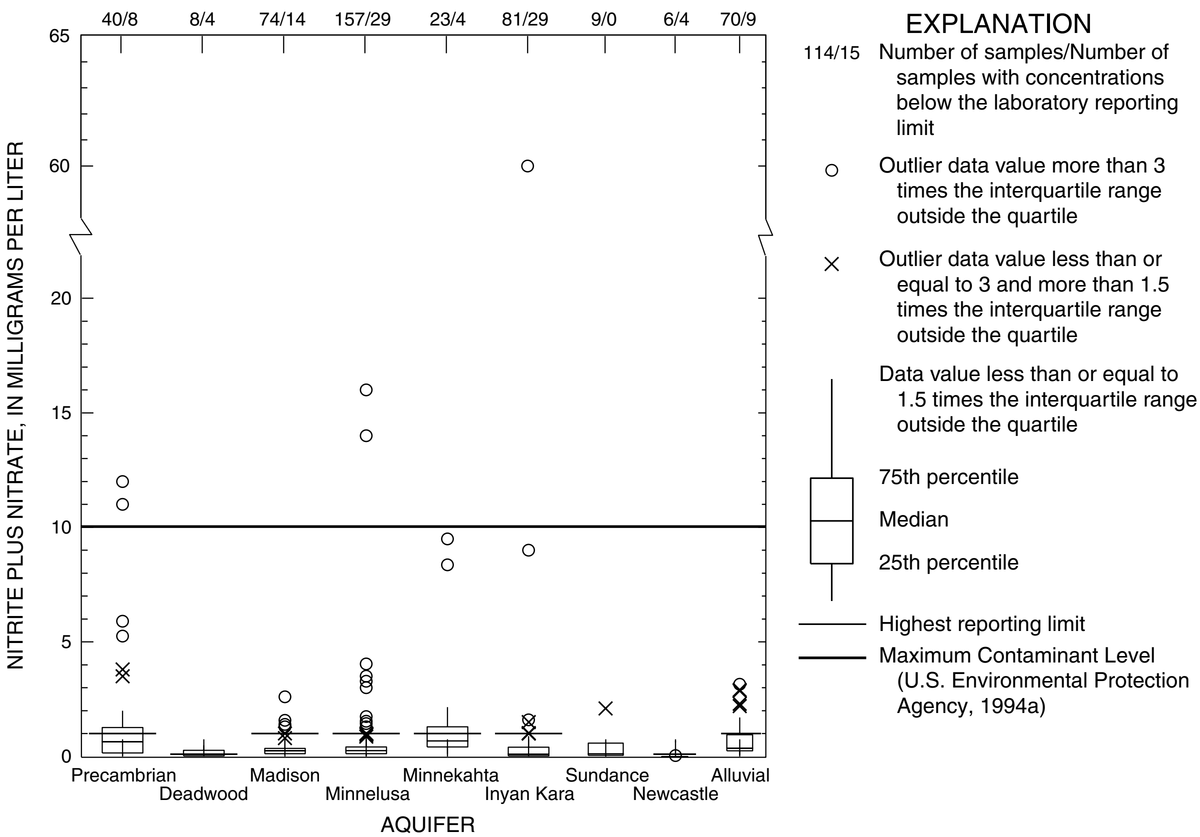


Figure 38. Boxplots of concentrations of nitrite plus nitrate for selected aquifers (modified from Williamson and Carter, 2001).

Maps showing sensitivity of ground water to contamination were produced by Putnam (2000) for Lawrence County and by Davis and others (1994) for the Rapid Creek Basin. The most sensitive hydrogeologic units are limestones, unconsolidated sands and gravels, and sandstones (Putnam, 2000). The least sensitive units include shales or units with interbedded shales. The Madison, Minnelusa, and Minnekahta aquifers are especially sensitive to contamination because of high secondary permeability and potential for streamflow recharge.

Summary Relative to Water Use

Concentrations of various constituents exceeding SMCL's and MCL's affect the use of water in some areas for many aquifers within the study area. Most concentrations exceeding standards are for various SMCL's and generally affect the water only aesthetically. Radionuclide concentrations can be high in some of the major aquifers, especially in the Deadwood and Inyan Kara aquifers, and may preclude the use of water in some areas. Hard water may require special treatment for certain uses. Other factors, such as the sodium adsorption ratio (SAR) and specific conductance, affect irrigation use.

The general suitability of ground water for irrigation in the study area can be determined by using the South Dakota irrigation-water diagram (fig. 39). The diagram is based on South Dakota irrigation-water standards (revised January 7, 1982) and shows the State's water-quality and soil-texture requirements for the issuance of an irrigation permit. The adjusted SAR for each aquifer was calculated according to Koch (1983) from the mean concentrations of calcium, magnesium, sodium, and bicarbonate for each aquifer. Water from all aquifers, with the exceptions of the Pierre and Sundance aquifers, generally is suitable for irrigation, but may not be in specific instances if either the specific conductance or the SAR is high.

High concentrations of iron and manganese occasionally can hamper the use of water from the Precambrian aquifer. None of the reported samples from the Precambrian aquifer exceeded drinking-water standards for radionuclides.

The principal deterrents to use of water from the Deadwood aquifer are high concentrations of radionuclides, including radium-226 and radon. In addition, concentrations of iron and manganese can be high.

Water from the Madison aquifer can contain high concentrations of iron and manganese that may deter its use. Water from the Madison aquifer is hard to very hard and may require special treatment for certain uses. In downgradient wells (generally deeper than 2,000 ft), concentrations of dissolved solids and sulfate also may deter use from this aquifer. Hot water from deep wells and in the Hot Springs area, may not be desirable for some uses. Radionuclide concentrations in the Madison aquifer generally are acceptable.

The principal properties or constituents that may hamper the use of water from the Minnelusa aquifer include hardness and high concentrations of iron and manganese. Generally, downgradient wells (generally deeper than 1,000 ft) also have high concentrations of dissolved solids and sulfate. Hot water, from deep wells, may not be desirable for some uses. Arsenic concentrations in the Minnelusa aquifer exceed the revised MCL of 10 μg/L in some locations. Only a few samples exceeded the MCL's for various radionuclides.

Samples from the Minnekahta aquifer are available only from shallow wells near the outcrop. Water from the Minnekahta aquifer is harder than that from any of the other major aquifers in the study area, and may require special treatment for certain uses. Water generally is suitable for all water uses; few samples exceeded SMCL's and no samples available for this study from the Minnekahta aquifer exceeded drinking-water standards for any radionuclides.

The use of water from the Inyan Kara aquifer may be hampered by high concentrations of dissolved solids, iron, sulfate, and manganese. In the southern Black Hills, radium-226 and uranium concentrations also may preclude its use. Hard water from wells located on or near the outcrop of the Inyan Kara Group may require special treatment.

The use of water from the minor aquifers (Spearfish, Sundance, Morrison, Pierre, Graneros, Newcastle, and alluvial aquifers) may be hampered by hardness and concentrations of dissolved solids and sulfate. Concentrations of radionuclides, with the exception of uranium, generally are at acceptable levels in samples from the minor aquifers. Selenium concentrations in some places are an additional deterrent to the use of water from the Sundance aquifer.

Water from alluvial aquifers generally is very hard and may require special treatment for certain uses. High concentrations of dissolved solids, sulfate, iron, and manganese may limit the use of water from alluvial aquifers that overlie the Cretaceous shales. In the southern Black Hills, uranium concentrations in alluvial aquifers can be high in many locations.

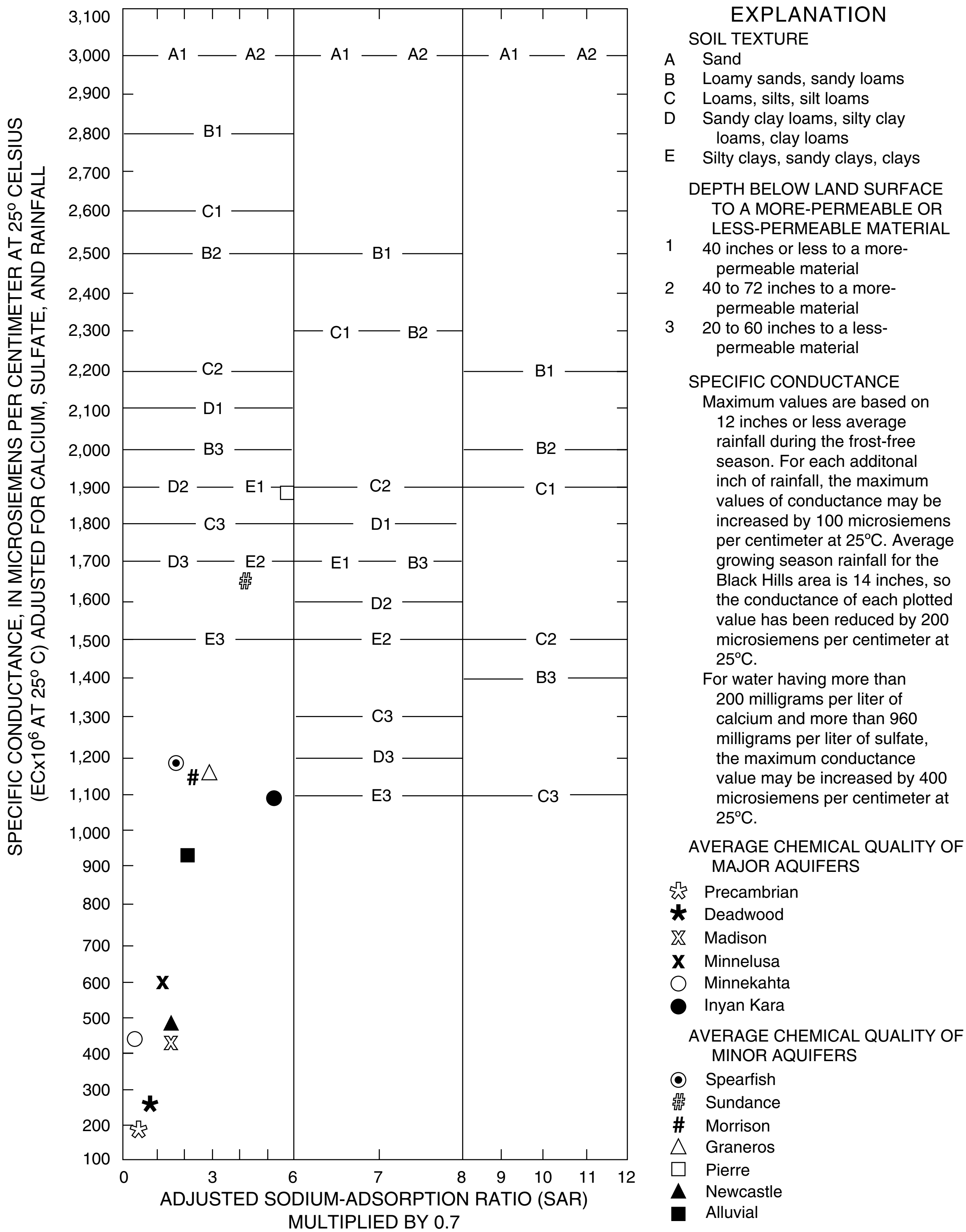


Figure 39. South Dakota irrigation-water classification diagram for selected aquifers (from Williamson and Carter, 2001). This diagram is based on South Dakota standards (revised Jan. 7, 1982) for maximum allowable specific conductance and adjusted sodium-adsorption-ratio values for which an irrigation permit can be issued for applying water under various soil-texture conditions. Water can be applied under all conditions at or above the plotted point, but not below it, provided other conditions as defined by the State Conservation Commission are met (from Koch, 1983).

Surface-Water Characteristics

Within this section, surface-water characteristics, including both streamflow and water-quality characteristics, are described. Surface-water characteristics can be affected by numerous physical variables such as topography, land cover, soil conditions, mineralogy, and ground-water conditions, all of which may be affected by geologic conditions. In addition, streamflow is affected by numerous climatic variables including timing, intensity, and amount of precipitation, as well as other variables affecting evaporative processes.

Streamflow Characteristics

Streamflow characteristics in the Black Hills area are highly affected by the hydrogeologic settings previously described (fig. 23). Streamflow characteristics described in this section include variability of streamflow, the response of streamflow to precipitation, and annual yield characteristics. More detailed discussions of these topics were presented by Driscoll and Carter (2001).

Streamflow Variability

A distinctive effect of hydrogeologic setting is on the timing and variability of streamflow, which results primarily from interactions between surface water and ground water. Locations of streamflow-gaging stations for basins representative of the five hydrogeologic settings were presented in figure 23. Site information and selected flow characteristics are summarized (by hydrogeologic setting) in table 5. One of the flow characteristics summarized is the "base flow index" (BFI), which represents the estimated percentage of average streamflow contributed by base flow, for any given gage. BFI's were determined with a computer program described by Wahl and Wahl (1995).

Table 5 also includes mean flow values for representative gages (for the periods of record shown) in cubic feet per second and mean values of annual basin yield, expressed in inches per unit area. Because basin yields are normalized, relative to surface drainage area, values are directly comparable among different gages. For example, the mean flow of 11.73 ft^3/s for Castle Creek (station 06409000) is about 2.7 times larger than the mean flow of 4.33 ft^3/s for Cold Springs Creek (station 06429500); however, the mean annual basin yield for Castle Creek (2.01 inches) is smaller than for Cold Springs Creek (3.10 inches).

The last flow characteristic summarized in table 5 is the coefficient of variation (standard deviation divided by mean) for annual basin yield, which provides a useful measure of annual flow variability. This statistic is directly comparable among different gages because the standard deviations are normalized relative to means. For example, standard deviations for Beaver Creek at Mallo Camp (06392900) and Rhoads Fork (06408700) are very different; however, coefficients of variation are nearly identical. A notable example is provided by two gages representative of artesian spring basins—Cascade Springs (06400497) and Cox Lake (06430540), which have anomalously large values for annual basin yield (orders of magnitude higher than annual precipitation) because of extremely large artesian springflow that occurs in very small drainages. Standard deviations for these sites are the largest in table 5; however, the coefficients of variation are the smallest, which is consistent with the BFI's, which are the largest in the table and are indicative of extremely large contributions from base flow.

Duration curves showing variability in daily flow are presented in figure 40 for selected basins. Streamflow variability is small for limestone headwater and artesian spring basins because streamflow consists almost entirely of base flow from spring discharge. For the individual limestone headwater basins, measured daily flows generally vary by less than an order of magnitude, indicating that direct runoff is very uncommon from outcrops of the Madison Limestone and Minnelusa Formation, which are the predominant outcrops for this setting. Streams in the crystalline core setting have large variability in daily flow. Loss zone and exterior settings have large flow variability and low-flow and zero-flow periods are common.

Relative variability of monthly and annual flow also is much smaller for basins representative of limestone headwater and artesian spring settings than for the other settings (figs. 41 and 42). Annual flow values are expressed as annual yield (fig. 42) for all hydrogeologic settings except the artesian spring setting, for which annual yield values can be unrealistically large (table 5), as previously discussed. Coefficients of variation for these settings are consistently smaller than for the other settings (table 5). BFI's are consistently larger, indicating large proportions of base flow for these settings. All measures considered indicate much higher flow variability for the other three settings.

Table 5. Summary of selected site information and flow characteristics for streamflow-gaging stations representative of hydrogeologic settings

[Modified from Driscoll and Carter (2001). --, not determined]

Station number	Station name	Drainage area (square miles)	Period of record used (water years)	Base flow index (percent)	Mean flow (cubic feet per second)	Annual basin yield		
						Mean (inches)	Standard deviation	Coefficient of variation (standard deviation/mean)
	Limestone Headwater Basins							
06392900	Beaver Creek at Mallo Camp, near Four Corners, WY[1]	10.3	1975-82, 1992-98	88.6	1.88	2.48	0.63	0.25
06408700	Rhoads Fork near Rochford	7.95	1983-98	98.7	5.47	9.34	2.48	.27
06409000	Castle Creek above Deerfield Reservoir, near Hill City	79.2	1949-98	87.1	11.73	2.01	.75	.37
06429500	Cold Springs Creek at Buckhorn, WY[1]	19.0	1975-82, 1992-98	91.4	4.33	3.10	.68	.22
06430770	Spearfish Creek near Lead	63.5	1989-98	[2]91.0	[2]25.43	[2]5.44	[2]2.59	[2].48
06430850	Little Spearfish Creek near Lead	25.8	1989-98	97.0	16.59	8.74	2.31	.26
	Crystalline Core Basins							
06402430	Beaver Creek near Pringle	45.8	1991-98	73.1	2.86	.85	.76	.89
06402995	French Creek above Stockade Lake, near Custer[3]	68.7	--	--	--	--	--	--
06403300	French Creek above Fairburn	105	1983-98	55.5	10.94	1.42	1.19	.84
06404000	Battle Creek near Keystone	58.0	1962-98	45.4	9.39	2.20	1.59	.72
06404800	Grace Coolidge Creek near Hayward[3]	7.48	--	--	--	--	--	--
06404998	Grace Coolidge Creek near Game Lodge, near Custer	25.2	1977-98	58.9	5.07	2.73	2.36	.86
06405800	Bear Gulch near Hayward	4.23	1990-98	41.1	1.48	4.75	2.76	.58
06406920	Spring Creek above Sheridan Lake, near Keystone[3]	127	--	--	--	--	--	--
06407500	Spring Creek near Keystone	163	1987-98	54.1	25.06	2.09	1.73	.83
06422500	Boxelder Creek near Nemo	96.0	1967-98	64.9	19.53	2.76	2.19	.79
06424000	Elk Creek near Roubaix	21.5	1992-98	61.1	13.42	8.48	4.08	.48
06430800	Annie Creek near Lead	3.55	1989-98	51.1	1.72	6.55	4.42	.67
06430898	Squaw Creek near Spearfish	6.95	1989-98	52.5	3.76	7.34	4.44	.60
06436156	Whitetail Creek at Lead	6.15	1989-98	63.0	4.79	10.57	6.01	.57
06437020	Bear Butte Creek near Deadwood[1]	16.6	1989-98	58.3	8.35	6.84	4.07	.60

Table 5. Summary of selected site information and flow characteristics for streamflow-gaging stations representative of hydrogeologic settings–Continued

[Modified from Driscoll and Carter (2001). --, not determined]

Station number	Station name	Drainage area (square miles)	Period of record used (water years)	Base flow index (percent)	Mean flow (cubic feet per second)	Annual basin yield		
						Mean (inches)	Standard deviation	Coefficient of variation (standard deviation/mean)
			Loss Zones Basins					
06408500	Spring Creek near Hermosa	199	1950-98	44.1	7.15	.49	.73	1.49
06423010	Boxelder Creek near Rapid City	128	1979-98	14.4	5.88	.62	1.23	1.98
			Artesian Spring Basins					
06392950	Stockade Beaver Creek near Newcastle, WY[1]	107	1975-82, 1992-98	93.5	12.15	1.54	0.23	0.15
06400497	Cascade Springs near Hot Springs	.47	1977-95	99.2	19.53	564	40.34	.07
06402000	Fall River at Hot Springs	137	1939-46, 1948-98	96.0	23.61	2.34	.25	.11
06402470	Beaver Creek above Buffalo Gap	111	1991-97	97.4	10.21	1.25	.25	.20
06412810	Cleghorn Springs at Rapid City[3]	--	--	--	--	--	--	--
06429905	Sand Creek near Ranch A, near Beulah, WY[1]	267	1977-83, 1992-98	95.1	22.58	1.15	.22	.19
06430532	Crow Creek near Beulah, WY	40.8	1993-98	92.6	40.68	13.5	1.13	.08
06430540	Cox Lake outlet near Beulah, WY	.07	1991-95	99.3	4.22	819	9.16	.01
			Exterior Basins					
06395000	Cheyenne River at Edgemont[3]	7,143	--	--	--	--	--	--
06400000	Hat Creek near Edgemont	1,044	1951-98	15.5	16.61	.22	.26	1.18
06400875	Horsehead Creek at Oelrichs[1]	187	1984-98	12.6	6.75	.49	.70	1.43
06433500	Hay Creek at Belle Fourche	121	1954-96	17.5	1.74	.20	.23	1.15
06436700	Indian Creek near Arpan[1]	315	1962-81	6.6	19.98	.86	.92	1.07
06436760	Horse Creek above Vale[3]	464	--	--	--	--	--	--
06437500	Bear Butte Creek near Sturgis[1]	192	1946-72	32.3	13.93	.99	1.04	1.05

1Site used only for analysis of streamflow characteristics.

[2]Flow characteristics affected by relatively consistent diversions of about 10 cubic feet per second.

[3]Site used only for analysis of water-quality characteristics.

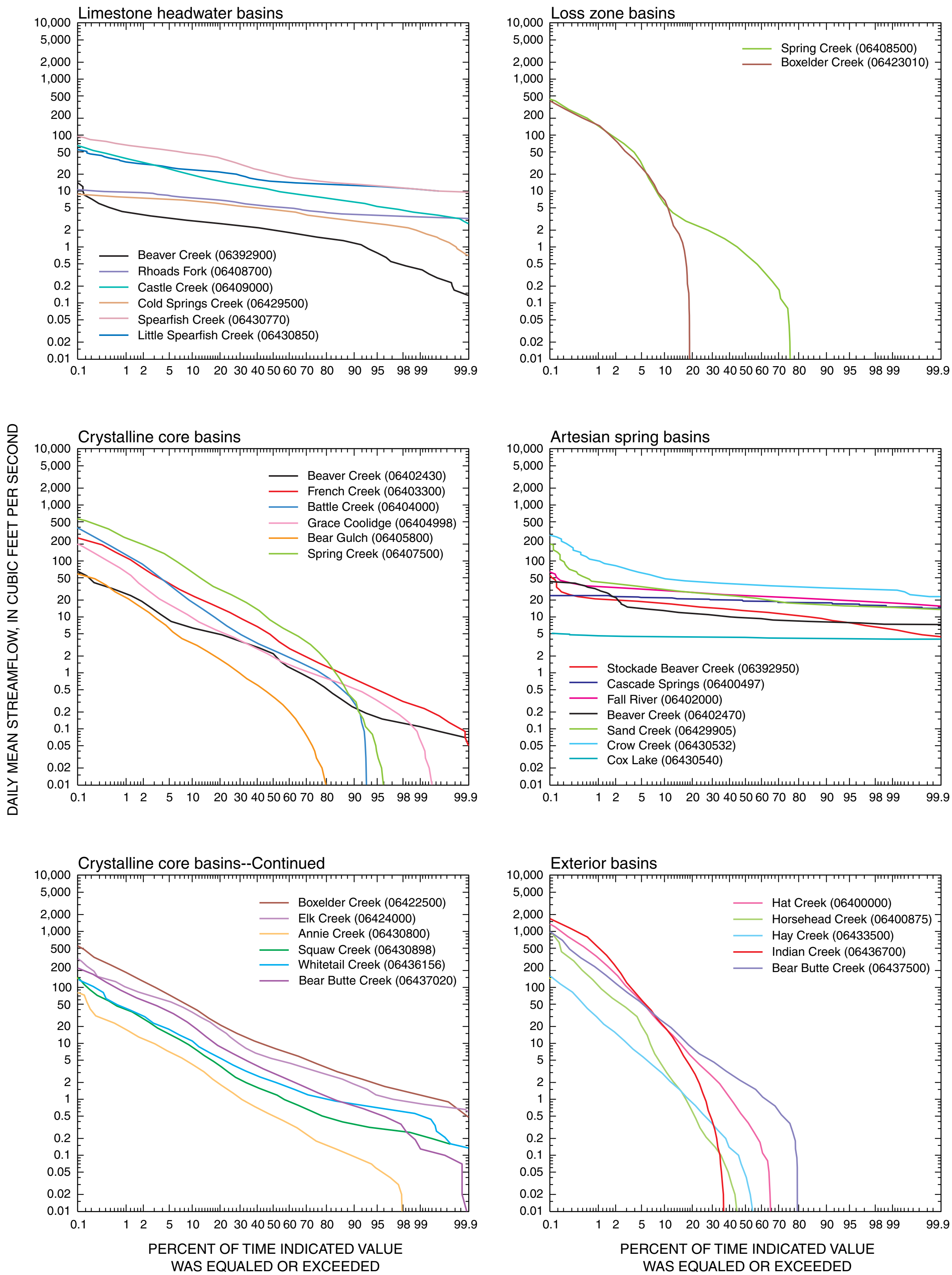


Figure 40. Duration curves of daily mean streamflow for basins representative of hydrogeologic settings (from Driscoll and Carter, 2001).

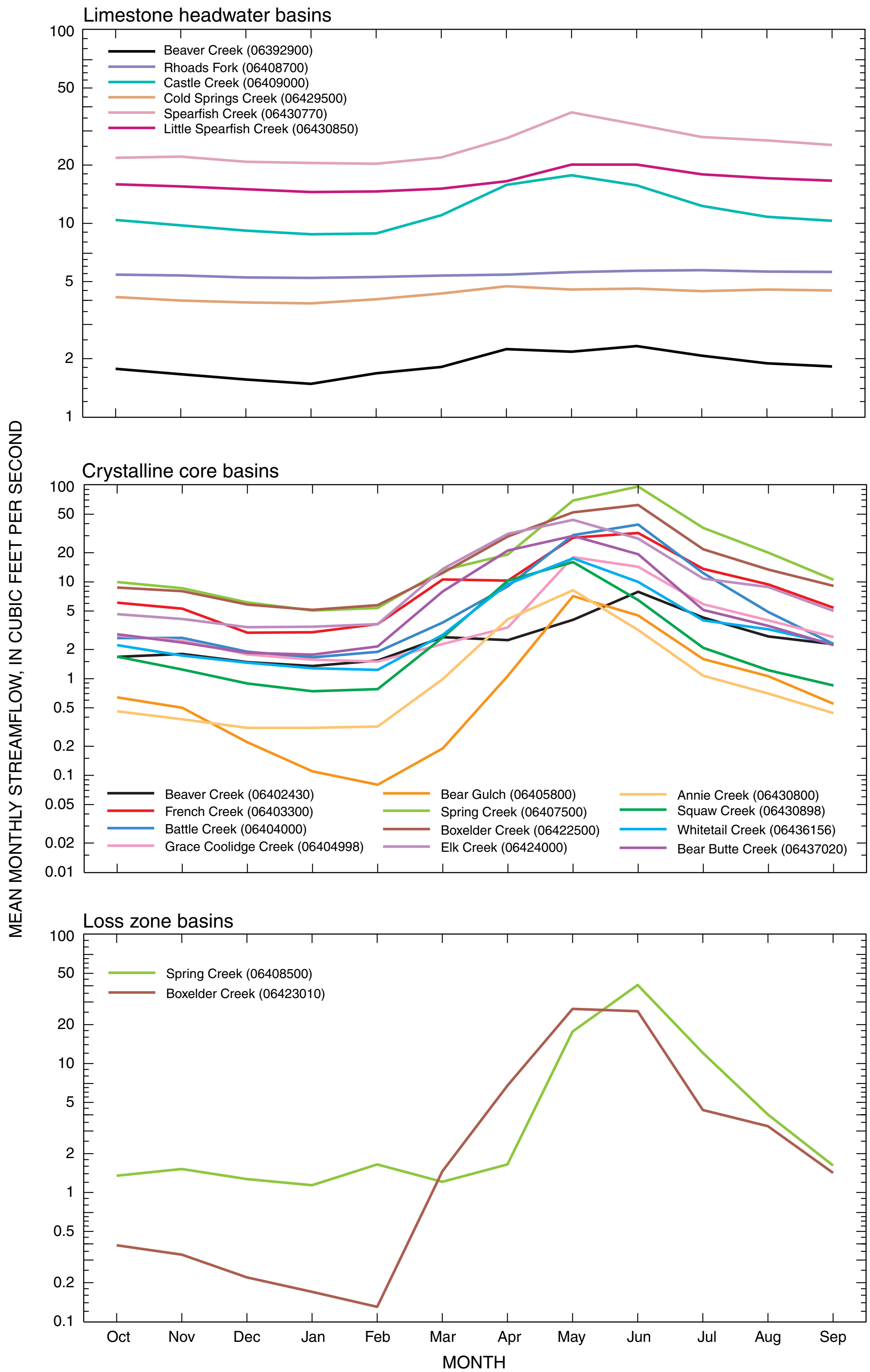


Figure 41. Mean monthly streamflow for basins representative of hydrogeologic settings (from Driscoll and Carter, 2001).

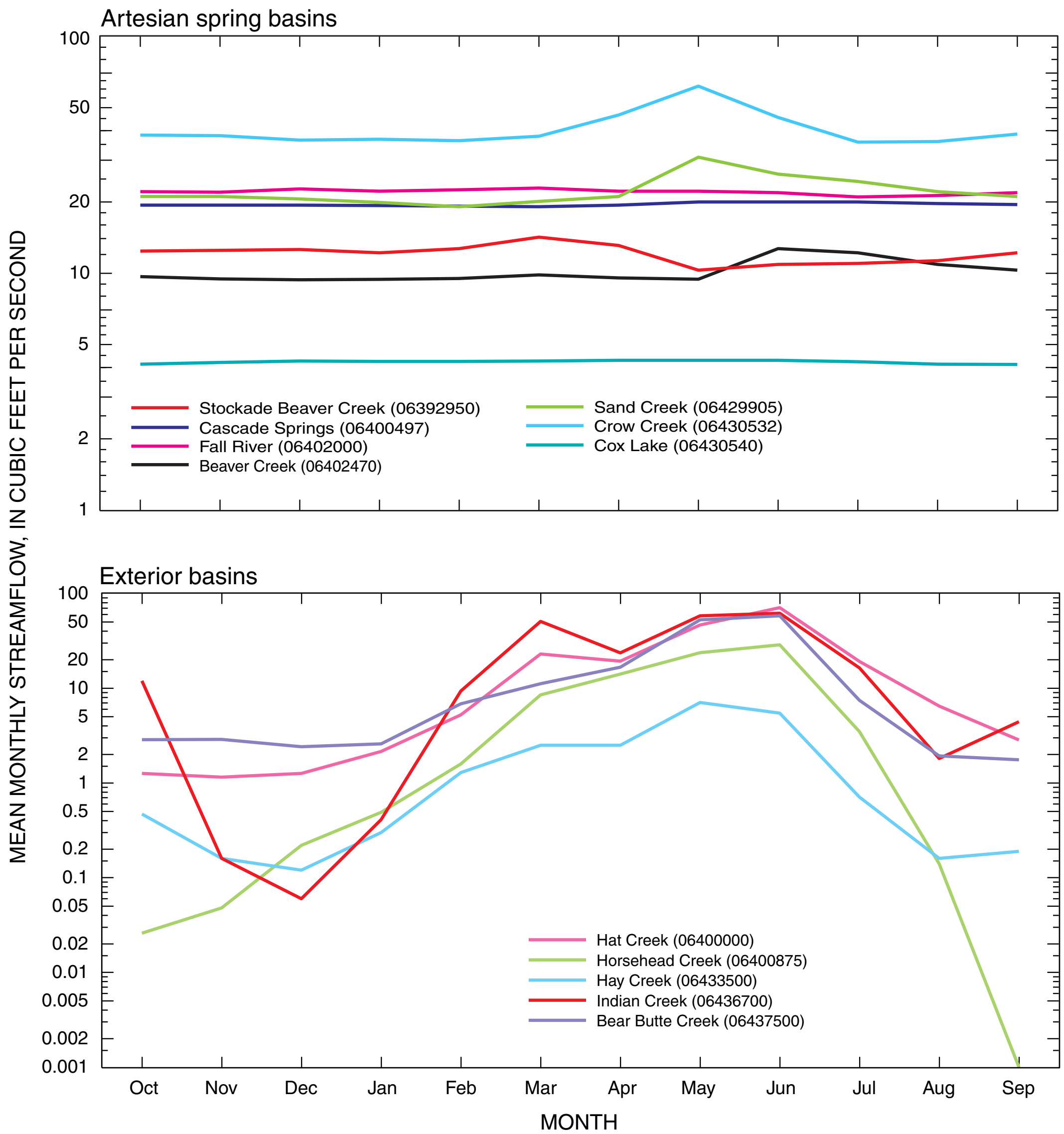
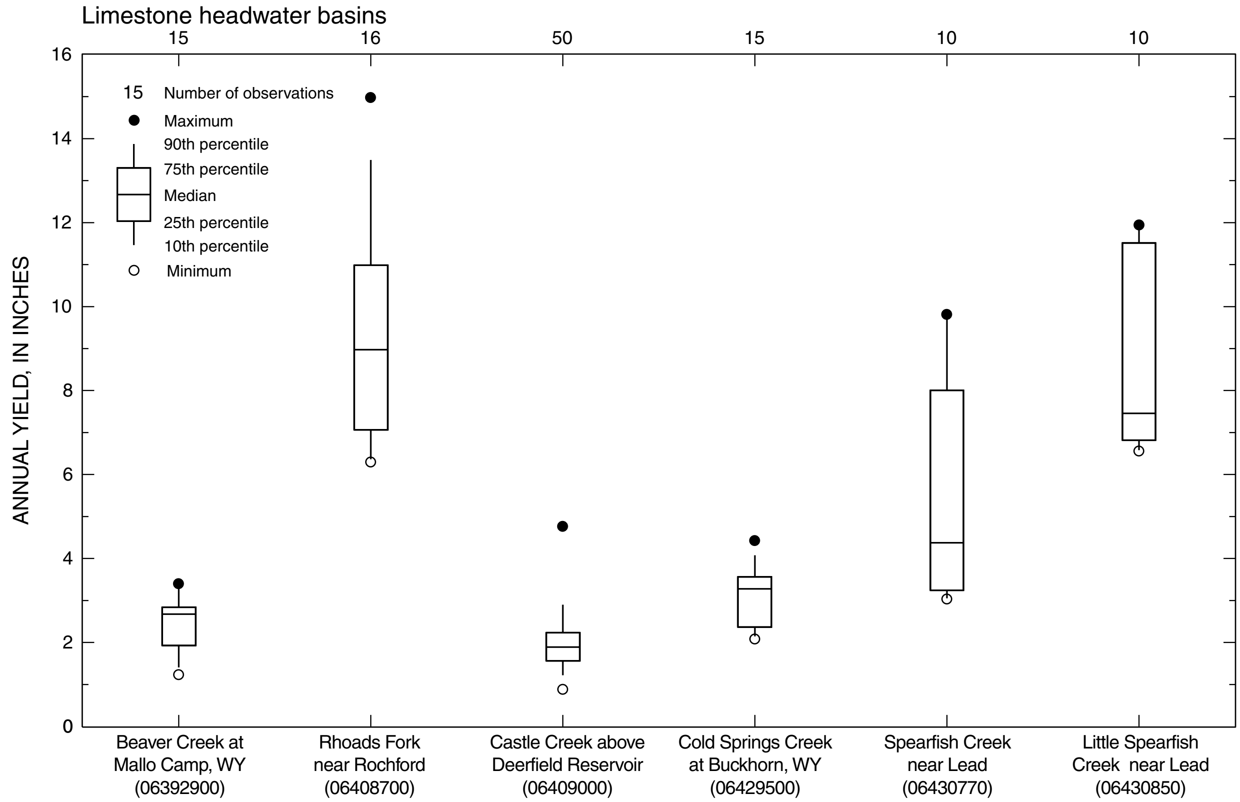


Figure 41. Mean monthly streamflow for basins representative of hydrogeologic settings (from Driscoll and Carter, 2001).—Continued



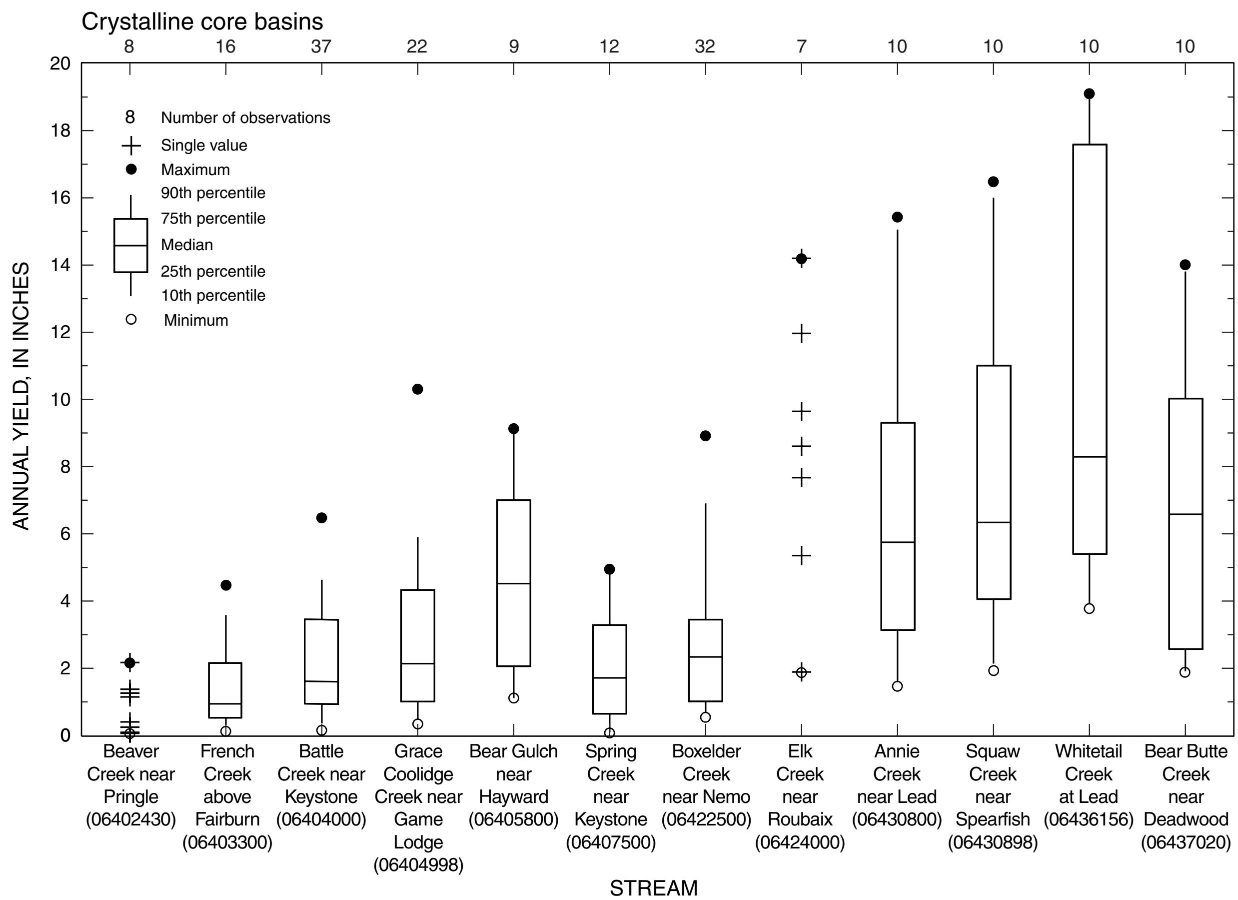


Figure 42. Distribution of annual yield for basins representative of hydrogeologic settings (from Driscoll and Carter, 2001).

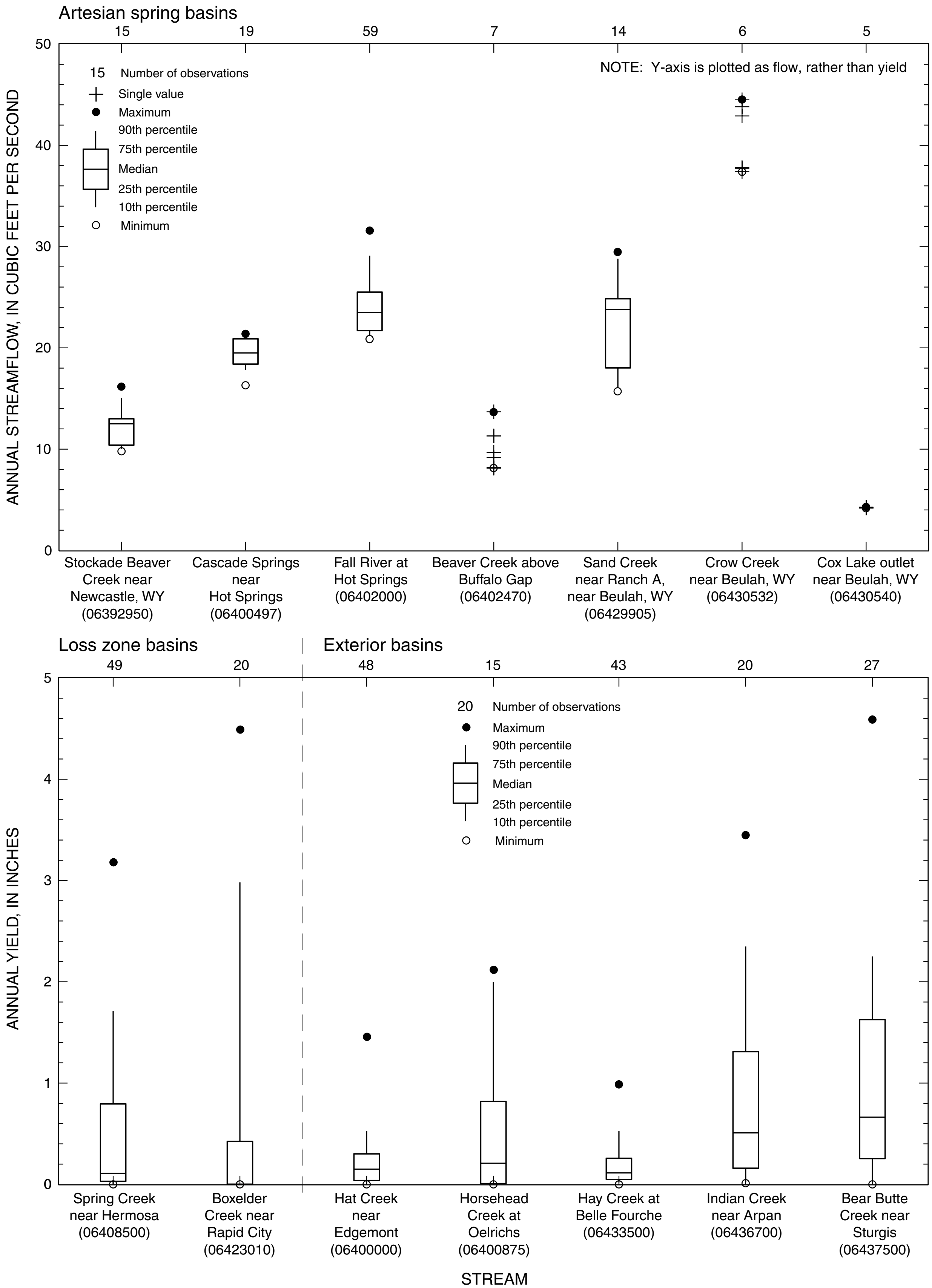


Figure 42. Distribution of annual yield for basins representative of hydrogeologic settings (from Driscoll and Carter, 2001).—Continued

BFI's for the crystalline core basins generally approach or slightly exceed 50 percent (table 5). Monthly flow characteristics (fig. 41), however, indicate a short-term response to precipitation patterns (fig. 8), which probably indicates a relatively large component of interflow contributing to base flow. This interpretation is supported by the general physical characteristics of the crystalline core basins, where large relief and steep planar surfaces provide conditions amenable to non-vertical flow components in the unsaturated zone. Ground-water discharge also contributes to streamflow; however, ground-water storage available for contribution to streamflow apparently is quickly depleted, as evidenced by the lower end of the range of annual yield values for the crystalline core basins (fig. 42). Daily flow values span two or more orders of magnitude for all crystalline core basins (fig. 40).

Few gages representative of the loss zone setting exist because sustained flow is uncommon downstream from outcrop areas where large streamflow losses provide recharge to the Madison and Minnelusa aquifers (Hortness and Driscoll, 1998). The only two representative loss zone gages (fig. 23) are located on Spring Creek (06408500) and Boxelder Creek (06423010). Annual basin yields for these gages are much smaller than for gages located upstream (stations 06407500 on Spring Creek and 06422500 on Boxelder Creek) and relative variability in flow is larger (table 5, figs. 40-42). Spring Creek does have relatively consistent base flow (table 5, BFI = 44 percent) from alluvial springs that occur a short distance upstream from the gage.

Seven representative gages for the artesian spring setting are considered (fig. 23), of which two (Cascade Springs and Cox Lake) are located in extremely small drainages with no influence from streamflow losses. Four of the gages are located in larger drainages downstream from loss zones, and one basin (Fall River, 06402000) heads predominantly within the loss zone setting (fig. 23). Monthly means (fig. 41) for Fall River show no apparent influence of flows through loss zones, in spite of storm flows that occasionally increase daily flows (fig. 40). Minor influence of flows through loss zones is apparent in both monthly and daily flow characteristics for the other four gages (figs. 40 and 41). The influence of minor irrigation diversions along Stockade Beaver Creek (06392950) during late spring and summer months also is apparent.

For the exterior setting, daily flows for representative gages vary by more than four orders of magnitude (fig. 40) and zero-flow conditions are common, which is consistent with BFI's that typically are small (table 5). Large variability in monthly and annual flows also is characteristic for the exterior setting (figs. 41 and 42). Annual basin yields also are smaller than for most other settings, which is consistent with smaller precipitation and larger evaporation rates at lower altitudes. Many of these sites also are affected by minor irrigation withdrawals.

Response to Precipitation

Streams representative of the various hydrogeologic settings generally have distinctive characteristics relative to responsiveness to precipitation, as described within this section. Methods used for determination of precipitation over drainage areas were described by Driscoll and Carter (2001), who provided detailed discussions regarding relations between streamflow and precipitation.

The limestone headwater basins generally have weak correlations between annual streamflow and precipitation, as summarized in table 6. The r^2 values are low and p-values indicate that the correlations are not statistically significant (>0.05) for most of the representative basins, which is consistent with minimal variability in daily (fig. 40) and monthly (fig. 41) flow. Correlations with annual streamflow improve when "moving-average" precipitation (annual precipitation averaged over multiple years) is considered as the explanatory variable. Regression information is summarized in table 6 for the number of years of moving-average precipitation for which r^2 values are maximized for each basin.

The regression equation (table 6) for Castle Creek (station 06409000) probably is the most reliable, in spite of an associated r^2 value that is relatively low, primarily because the length of record is the longest (table 5). High r^2 values for several basins probably result primarily from relatively short periods of record; thus, associated regression equations for these stations may not be representative of long-term conditions. The p-values generally indicate strong statistical significance, however, which provides confidence that long-term precipitation patterns are much more important than short-term patterns for explaining streamflow variability in the limestone headwater setting. This concept is consistent with the hydrogeologic setting, where streamflow is dominated by headwater springflow.

Table 6. Summary of regression information for limestone headwater basins

[Regression information (from Driscoll and Carter, 2001) is provided for streamflow as a function of annual precipitation and as a function of moving average precipitation over a specified number of years. Int, intercept; <, less than]

Station number	Station name	Annual precipitation		Moving average precipitation				
		r^2	p-value	Number of years	r^2	p-value	Slope	Int
06392900	Beaver Creek at Mallo Camp	0.01	0.668	11	0.24	0.063	0.211	-2.78
06408700	Rhoads Fork	.16	.123	9	.93	<.010	.658	-9.12
06409000	Castle Creek	.31	<.010	3	.58	<.010	1.043	-10.70
06429500	Cold Springs Creek	.01	.800	11	.70	<.010	.722	11.65
06430770	Spearfish Creek near Lead	.72	<.010	7	.99	<.010	3.858	-68.63
06430850	Little Spearfish Creek	.53	.017	7	.93	<.010	1.450	-19.32

Graphs showing relations between annual streamflow and precipitation for crystalline core basins are presented in figure 43. Each graph includes a linear regression line, along with the corresponding equation and r^2 value. All of the slopes are highly significant; thus, p-values are not shown. The r^2 values range from 0.52 for Beaver Creek (06402430) to 0.87 for Bear Gulch (06405800), and are much higher as a group than for the limestone headwater basins (table 6), which is consistent with larger variability in flow characteristics (figs. 40-42).

An exponential regression curve, along with the corresponding equation and r^2 value, also is shown on each graph in figure 43. All of the exponential equations would predict small, positive streamflow for zero precipitation (which is not realistic), but avoid prediction of negative streamflow in the lower range of typical annual precipitation, which is indicated for many of the linear regression equations.

Each graph in figure 43 also includes a curve labeled "runoff efficiency prediction," which is derived from linear regression equations of runoff efficiency as a function of precipitation. Runoff efficiency (the ratio of annual basin yield to precipitation) represents the percentage of annual precipitation returned as streamflow. Runoff efficiency regression lines for the 12 representative crystalline core basins are shown in figure 44; regression equations were presented by Driscoll and Carter (2001). Figure 44 indicates that within each basin, runoff efficiency increases with increasing annual precipitation, and that basins with higher precipitation generally have higher efficiencies.

The runoff efficiency predictions (fig. 43) are derived by substituting values for annual precipitation into the runoff efficiency regression equations. Runoff efficiency predictions are unrealistic (slightly negative) for very low precipitation values, but are consistently positive for the measured ranges of precipitation and also closely resemble the linear regression equations (streamflow versus precipitation) through this range.

Relations between streamflow and precipitation for the two loss-zone basins are presented in figure 45. It is apparent that low-flow and zero-flow years are common, with substantial flows occurring only when upstream flows are sufficiently large to sustain flow through loss zones. A power equation and associated r^2 value are shown for each basin, which provide reasonable fits for the nonlinear data.

Regression statistics (annual streamflow versus precipitation) for artesian spring basins are summarized in table 7. Regression equations, which are not meaningful because of low r^2 values and p-values greater than 0.05, are not provided. Weak correlations are consistent with small variability in flow characteristics (figs. 40-42) associated with ground-water discharge and with long ground-water residence times. Naus and others (2001) concluded that large proportions of springflow for several of the representative artesian springs have residence times exceeding 50 years.

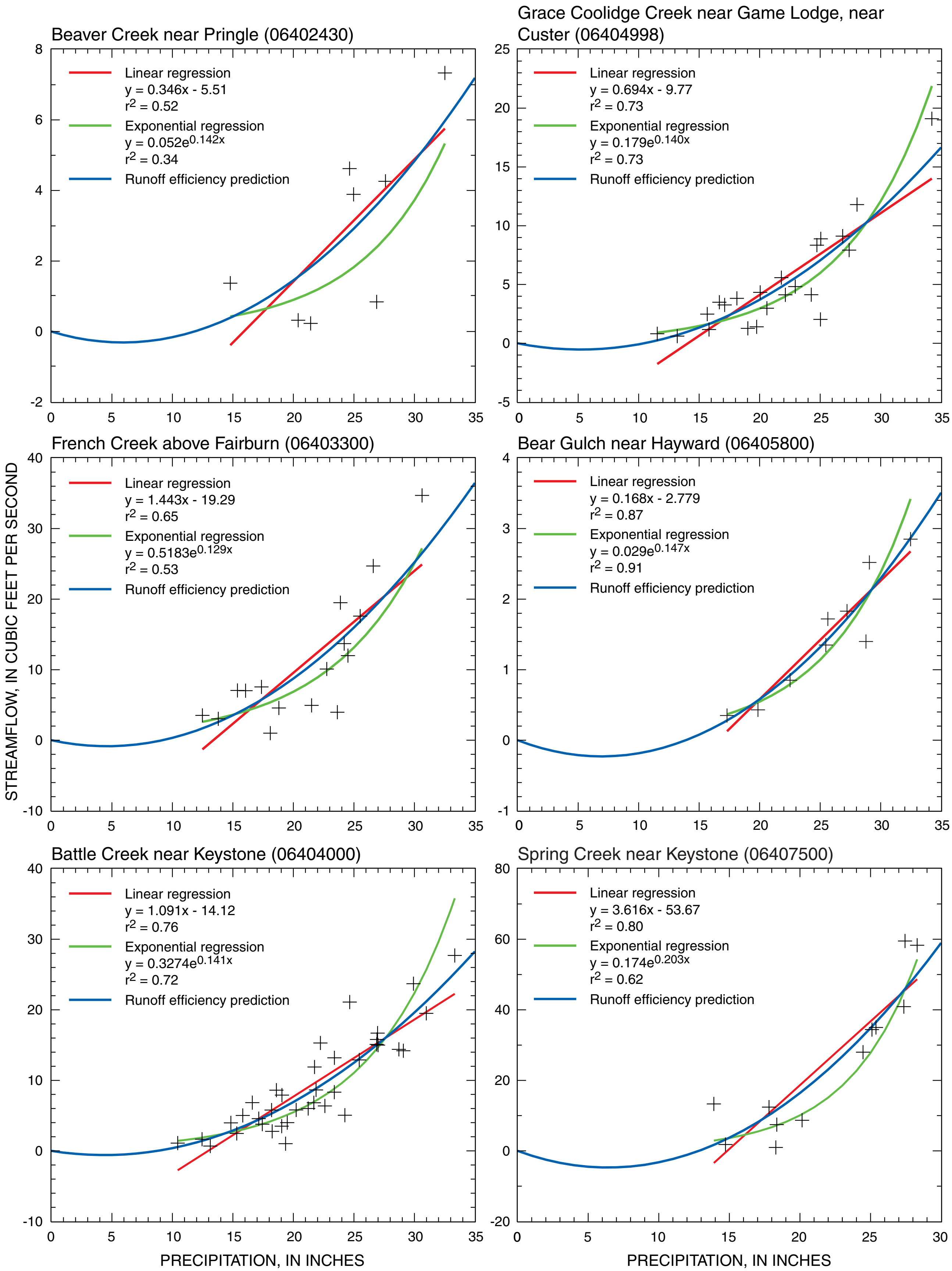


Figure 43. Relations between annual streamflow and precipitation for crystalline core basins (from Driscoll and Carter, 2001).

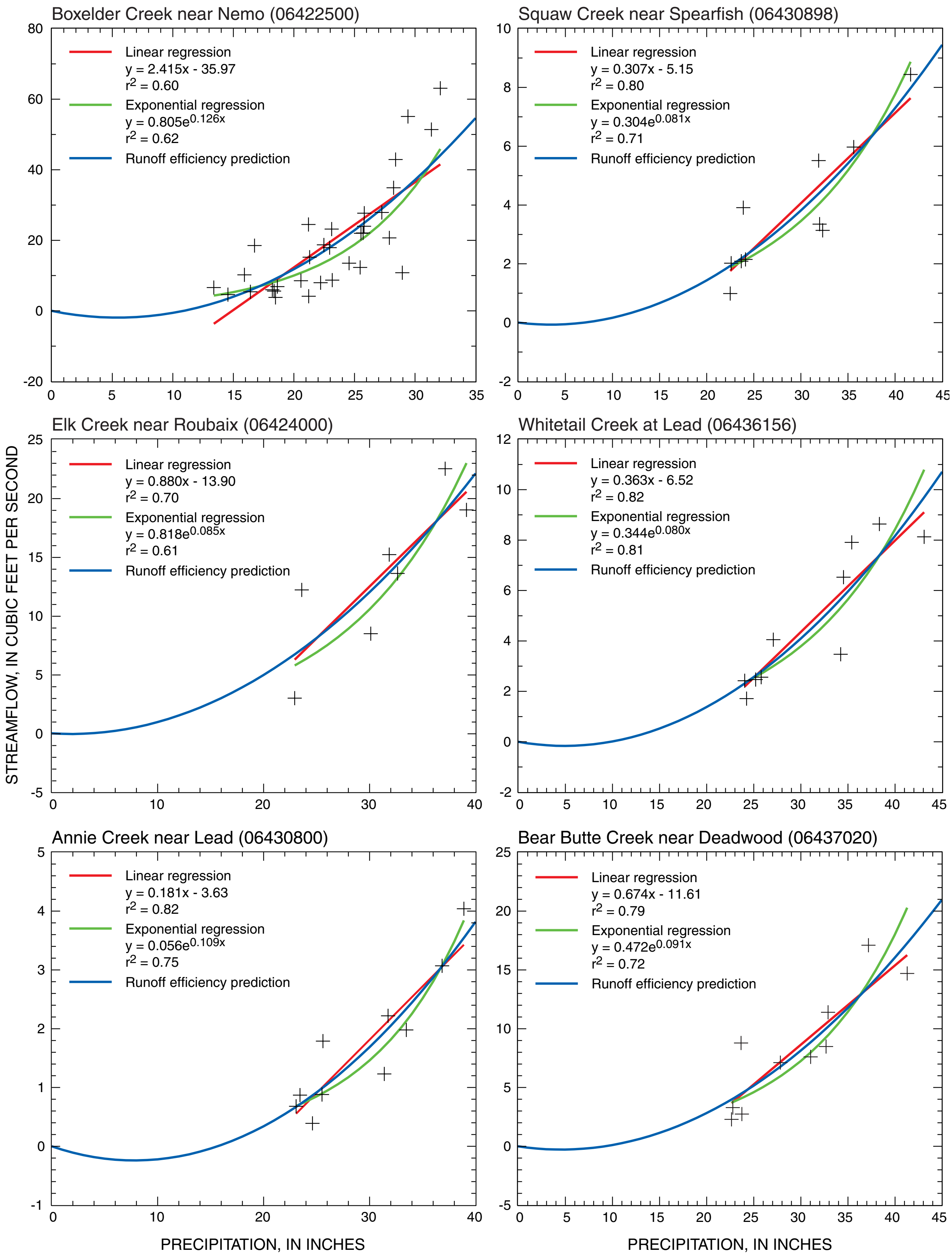


Figure 43. Relations between annual streamflow and precipitation for crystalline core basins (from Driscoll and Carter, 2001).—Continued

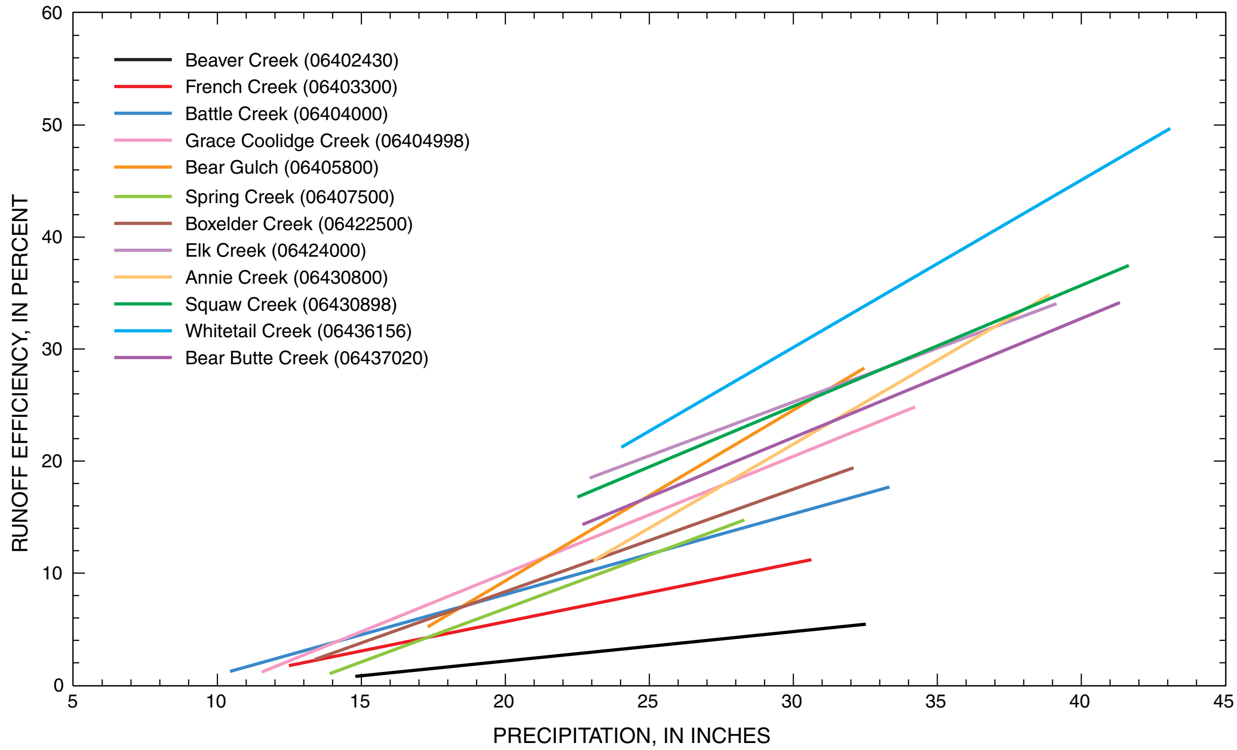


Figure 44. Relations between annual runoff efficiency and precipitation for crystalline core basins (from Driscoll and Carter, 2001).

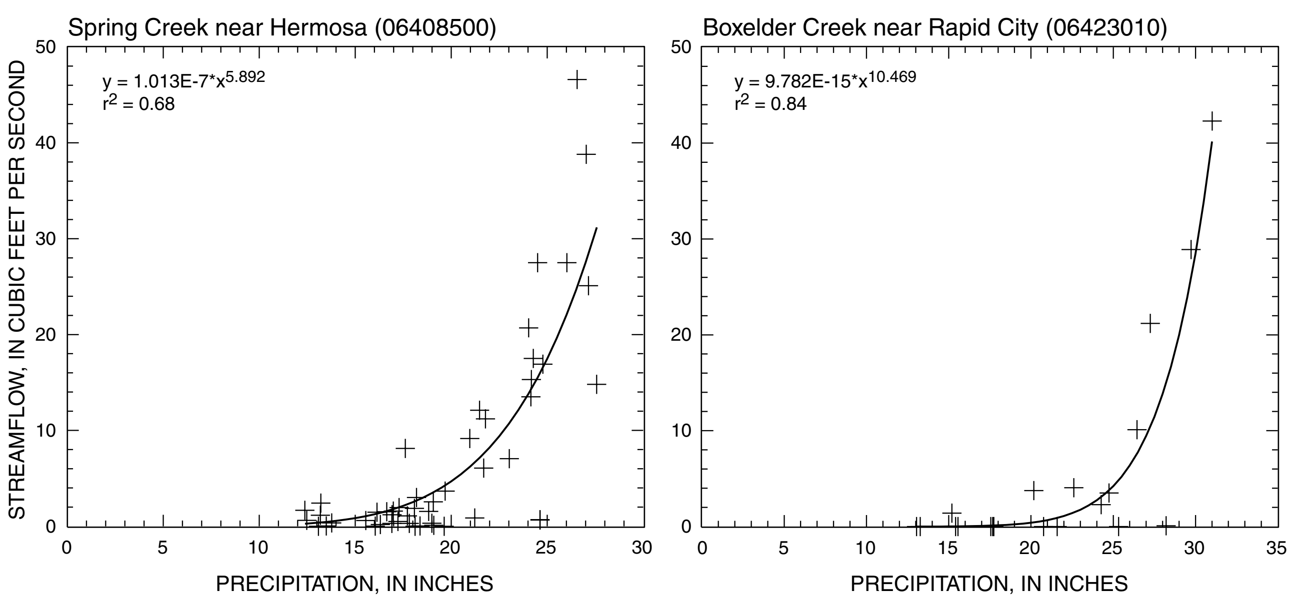


Figure 45. Relations between annual streamflow and precipitation for loss zone basins (from Driscoll and Carter, 2001).

Table 7. Summary of regression information for artesian spring basins

[Regression information (from Driscoll and Carter, 2001) is provided for streamflow as a function of annual precipitation]

Station number	Station name	Annual precipitation	
		r^2	p-value
06392950	Stockade Beaver Creek	0.16	0.135
06400497	Cascade Springs	.07	.289
06402000	Fall River	.003	.660
06402470	Beaver Creek above Buffalo Gap	.49	.079
06429905	Sand Creek	.04	.481
06430532	Crow Creek	.39	.185
06430540	Cox Lake	.55	.152

Driscoll and Carter (2001) identified a distinctive temporal trend in streamflow for the Fall River, which is composed almost entirely of artesian springflow. Peterlin (1990) investigated possible causes for declining streamflow that occurred during about 1940-70 (fig. 46), but did not conclusively determine causes. Wet climatic conditions during the 1990's have resulted in increased streamflow.

Relations between annual flow and precipitation for representative exterior basins are presented in figure 47. The p-values indicate that all correlations are statistically significant; however, the r^2 values generally are weak, relative to r^2 values for linear regressions for the crystalline core basins (fig. 43). A probable explanation is that crystalline core basins generally have larger base-flow components than exterior basins (table 5), which apparently are strongly influenced by annual precipitation amounts. In contrast, exterior basins are dominated by direct runoff, which is more responsive to event-oriented factors such as precipitation intensity.

Relations between annual runoff efficiency and precipitation for exterior basins are shown in figure 48. Runoff efficiencies generally increase with increasing precipitation, but efficiencies generally are lower than for the crystalline core basins (fig. 44) because of generally lower precipitation, increased evaporation potential, and minor irrigation withdrawals.

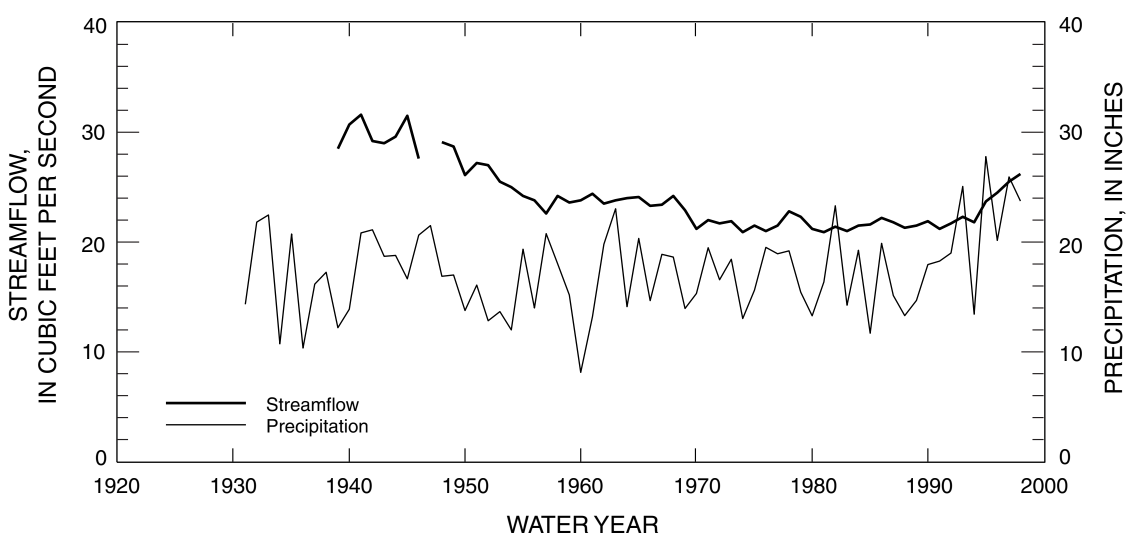


Figure 46. Long-term trends in annual streamflow for station 06402000 (Fall River near Hot Springs), relative to annual precipitation.

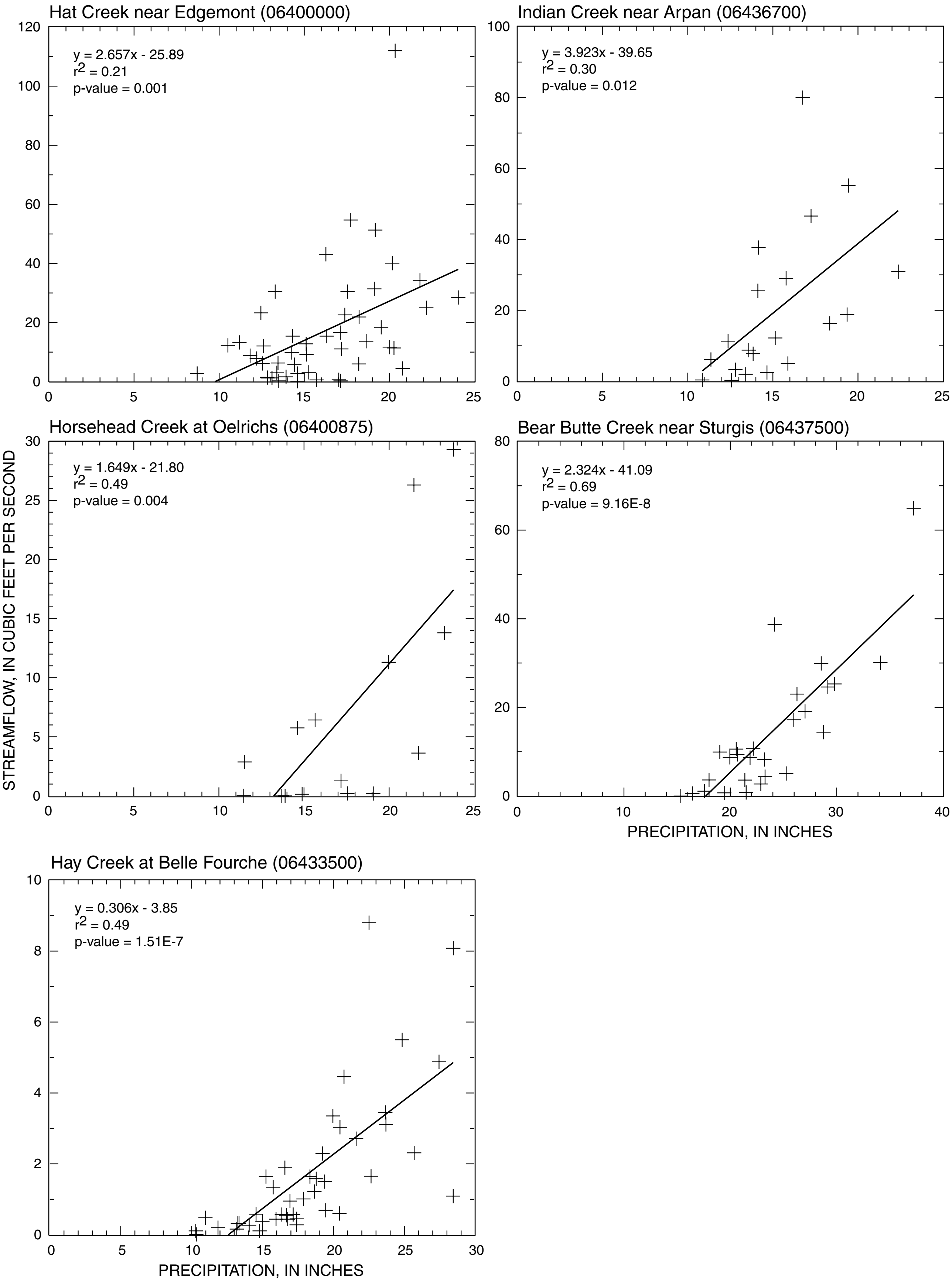


Figure 47. Relations between annual streamflow and precipitation for exterior basins (from Driscoll and Carter, 2001).

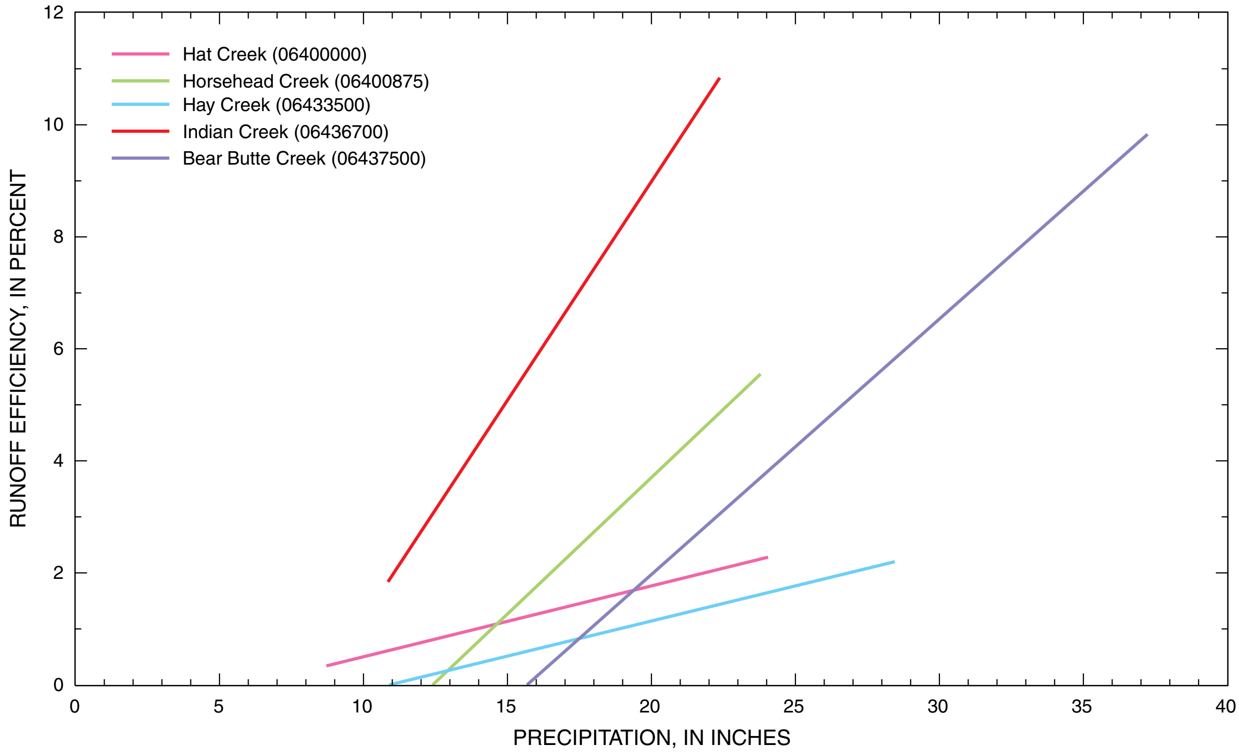


Figure 48. Relations between annual runoff efficiency and precipitation for exterior basins (from Driscoll and Carter, 2001).

Annual Yield

Annual yield characteristics are highly variable throughout the study area, primarily because of orographic effects, which influence both precipitation and evapotranspiration. Selected information for gages used for analysis of basin yield is presented in table 8. With the exception of site 2 (station 06395000, Cheyenne River), all of the sites considered are representative gages for either the limestone headwater, crystalline core, or exterior hydrogeologic settings (table 5). Two of the representative gages from these settings (stations 06405800, Bear Gulch and 06436700, Indian Creek) are excluded because annual yields may not be representative of areal conditions (Driscoll and Carter, 2001). All of the loss zone and artesian spring gages also are excluded.

Mean annual basin yields that are based on surface drainage areas for periods of measured record for selected gages are shown in figure 49. The largest yields occur in high-altitude areas of the northern Black Hills that receive large annual precipitation (fig. 4).

Large differences in annual yields are apparent for several of the limestone headwater basins, which results from incongruences between contributing ground- and surface-water areas. Mean annual yields for the four limestone headwater basins in South Dakota (sites 10, 11, 15, and 17; fig. 49) were estimated by Carter, Driscoll, Hamade, and Jarrell (2001) based on contributing ground-water areas. The contributing ground-water areas (fig. 50) were delineated by Jarrell (2000), based primarily on the structural orientation of the underlying Ordovician and Cambrian rocks. For the two limestone headwater basins in Wyoming (sites 1 and 14), relatively low yields indicate that contributing ground-water areas probably are smaller than the associated surface-water areas; however, estimates of contributing areas are not available.

Table 8. Summary of information used in analysis of yield characteristics

[From Driscoll and Carter (2001). --, not applicable]

Site number (fig. 49)	Station number	Station name	Period of record (water years)	Contributing area (square miles)		Mean annual yield for period of record (inches)		Mean annual yield efficiency[3] 1950-98 (percent)	
				Surface water	Ground water[1]	Surface water	Ground water[2]	Surface water	Ground water[2]
1	06392900	Beaver Creek at Mallo Camp	1975-82, 1992-98	10.3	([4])	2.48	--	[5]10.6	--
2	06395000	Cheyenne River	1947-98	7,143	--	.15	--	[6].9	--
3	06400000	Hat Creek	1951-98	1,044	--	.22	--	1.3	--
4	06400875	Horsehead Creek	1984-98	187	--	.49	--	2.1	--
5	06402430	Beaver Creek near Pringle	1991-98	45.8	--	.85	--	1.8	--
6	06403300	French Creek	1983-98	105	--	1.42	--	5.4	--
7	06404000	Battle Creek	1962-98	58.0	--	2.20	--	8.3	--
8	06404998	Grace Coolidge	1977-98	25.2	--	2.73	--	9.9	--
9	06407500	Spring Creek	1987-98	163	--	2.09	--	6.7	--
10	06408700	Rhoads Fork	1983-98	7.95	13.1	9.34	5.67	[5]41.8	[5]25.4
11	06409000	Castle Creek	1948-98	79.2	41.7	2.01	3.82	[6]9.3	[6]17.7
12	06422500	Boxelder Creek	1967-98	96.0	--	2.76	--	10.8	--
13	06424000	Elk Creek	1992-98	21.5	--	8.48	--	21.5	--
14	06429500	Cold Springs Creek	1975-82, 1992-98	19.0	([4])	3.10	--	[5]13.1	--
15	06430770	Spearfish Creek	1989-98	63.5	50.8	[7]7.58	9.48	[5,7]25.1	[5,7]31.4
16	06430800	Annie Creek	1989-98	3.55	--	6.55	--	16.4	--
17	06430850	Little Spearfish Creek	1989-98	25.8	25.4	8.74	8.88	[5]31.8	[5]32.3
18	06430898	Squaw Creek	1989-98	6.95	--	7.34	--	21.5	--
19	06433500	Hay Creek	1954-96	121	--	.20	--	1.0	--
20	06436156	Whitetail Creek	1989-98	6.15	--	10.57	--	27.2	--
21	06437020	Bear Butte Creek near Deadwood	1989-98	16.6	--	6.84	--	18.7	--
22	06437500	Bear Butte Creek near Sturgis	1946-72	[8]120	--	1.58	--	6.0	--

[1]Estimate of contributing ground-water area from Carter, Driscoll, Hamade, and Jarrell (2001).
[2]Yield estimates, where applicable, adjusted based on contributing ground-water area.
[3]Estimated using relations between runoff efficiency and precipitation from Carter, Driscoll, and Hamade (2001), unless otherwise noted.
[4]Contributing areas for surface water and ground water probably not congruent; however, no estimates available.
[5]Estimated using average runoff efficiency for the available period of record.
[6]Period of record sufficient for computation of yield efficiency.
[7]A flow of 10 cubic feet per second has been added to the measured streamflow to account for diverted flow.
[8]Approximate drainage area below loss zone. Actual drainage area is 192 square miles.

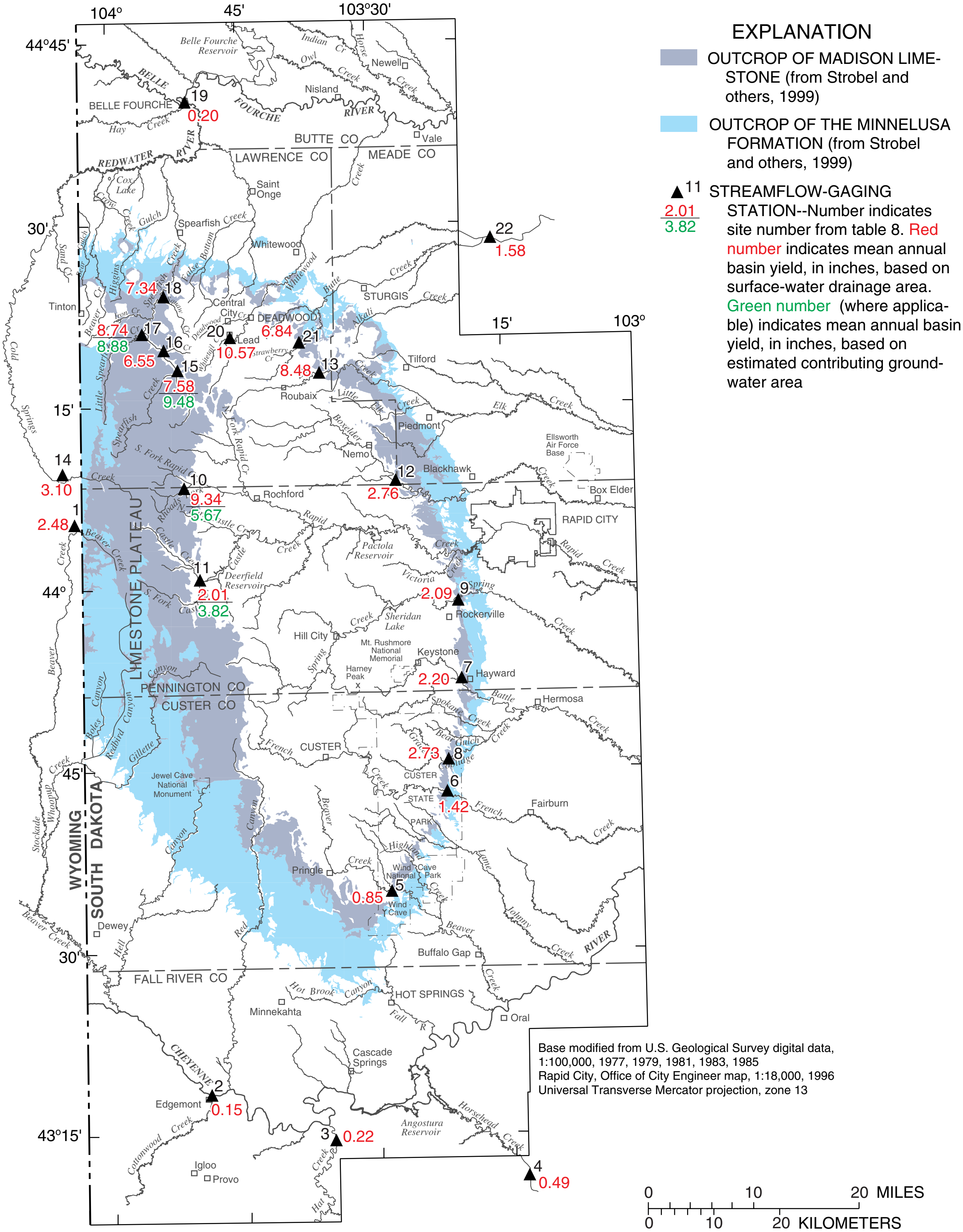


Figure 49. Basin yields for selected streamflow-gaging stations. For some stations, basin yields that are based on contributing ground-water areas estimated by Jarrell (2000) also are shown. Basin yields are for periods of record, which are not the same for all stations.

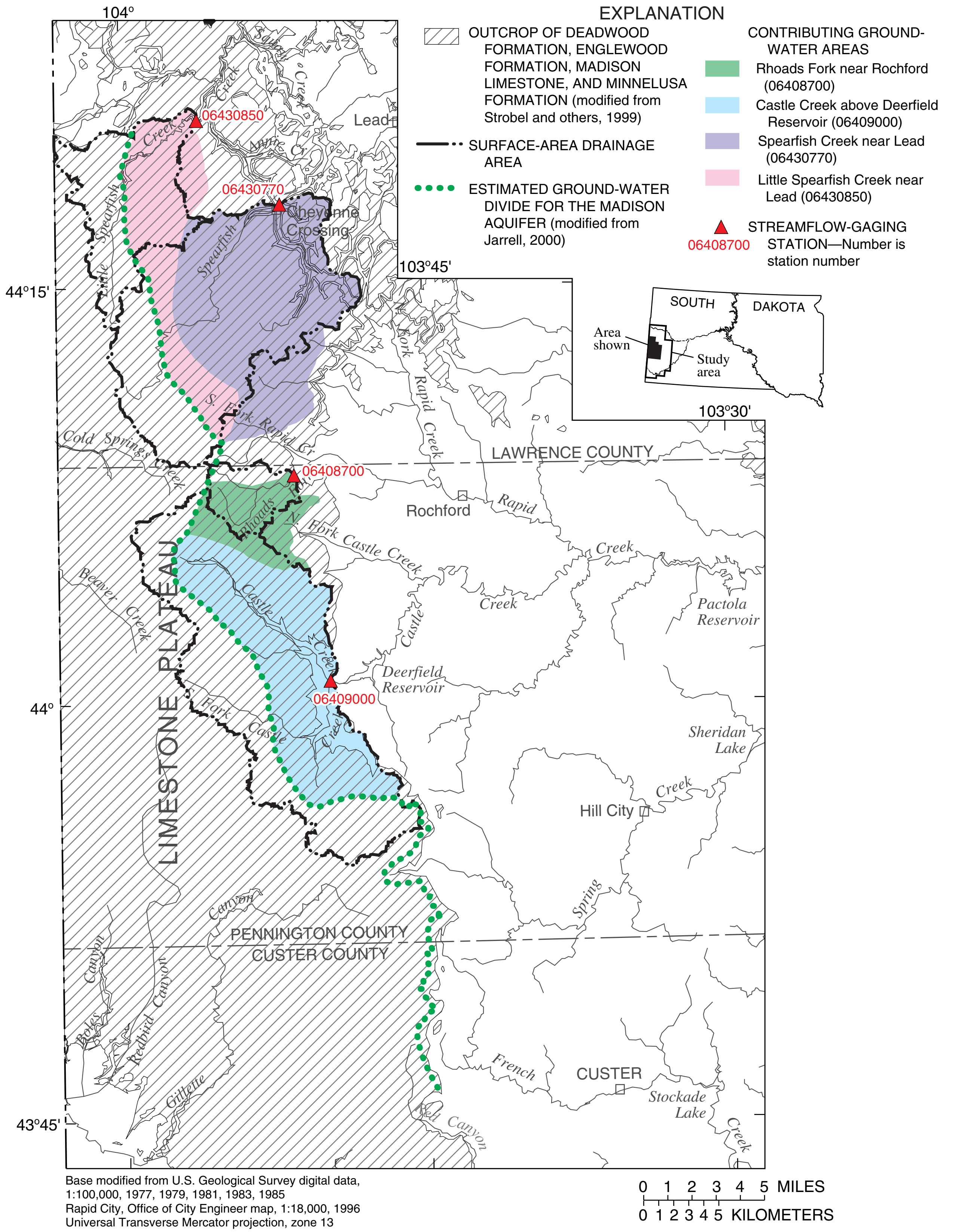


Figure 50. Comparison between surface-drainage areas and contributing ground-water areas for streamflow-gaging stations in Limestone Plateau area (modified from Jarrell, 2000). Streamflow in the basins shown generally is dominated by ground-water discharge of headwater springs. Recharge occurring in areas west of the ground-water divide does not contribute to headwater springflow east of the divide.

The approximate location of a ground-water divide that was identified by Jarrell (2000) also is shown in figure 50. This divide coincides with the western extent of the contributing ground-water areas for the four gaging stations that are shown. West of the ground-water divide, infiltration of precipitation results in ground-water recharge that is assumed to flow to the west, contributing to regional flowpaths in the Madison and Minnelusa aquifers that wrap around the northern or southern flanks of the uplift (fig. 17). East of the divide, recharge is assumed to contribute to headwater springflow along the eastern flank of the Limestone Plateau.

The ground-water divide extends about 10 mi south of the Castle Creek Basin and approximately coincides with the western extent of the Spring and French Creek drainage areas in this vicinity. The ground-water divide is not defined south of this point because the surface drainages contribute to Red Canyon, which flows to the south and provides streamflow recharge to the Madison and Minnelusa aquifers along the western flank of the uplift. Westerly ground-water flow directions are not possible immediately north of the ground-water divide because the Madison and Minnelusa aquifers are absent in the vicinity of Tertiary intrusive units (fig. 14).

After adjusting for contributing ground-water areas, annual yields for the limestone headwater basins (table 8; fig. 49) generally are consistent with a pattern of increasing yields corresponding with increasing annual precipitation (fig. 4). Adjusted yields for limestone headwater basins, which are dominated by ground-water discharge, also are generally similar to yields for nearby streams that are dominated by surface influences. These similarities were used by Carter, Driscoll, and Hamade (2001) in developing a method for estimating precipitation recharge to the Madison and Minnelusa aquifers. An important initial assumption was that in areas of comparable precipitation, evapotranspiration in outcrops of the Madison and Minnelusa Formations is similar to evapotranspiration for crystalline core settings, where recharge to regional flow systems is considered negligible. A further assumption was made that direct runoff is negligible for Madison and Minnelusa outcrops, which is supported by the daily flow characteristics for the limestone headwater setting. These assumptions resulted in a concept that streamflow yield in the crystalline core setting can be used as a surrogate for the efficiency of precipitation recharge to the Madison and Minnelusa aquifers. This concept is schematically illustrated in figure 51.

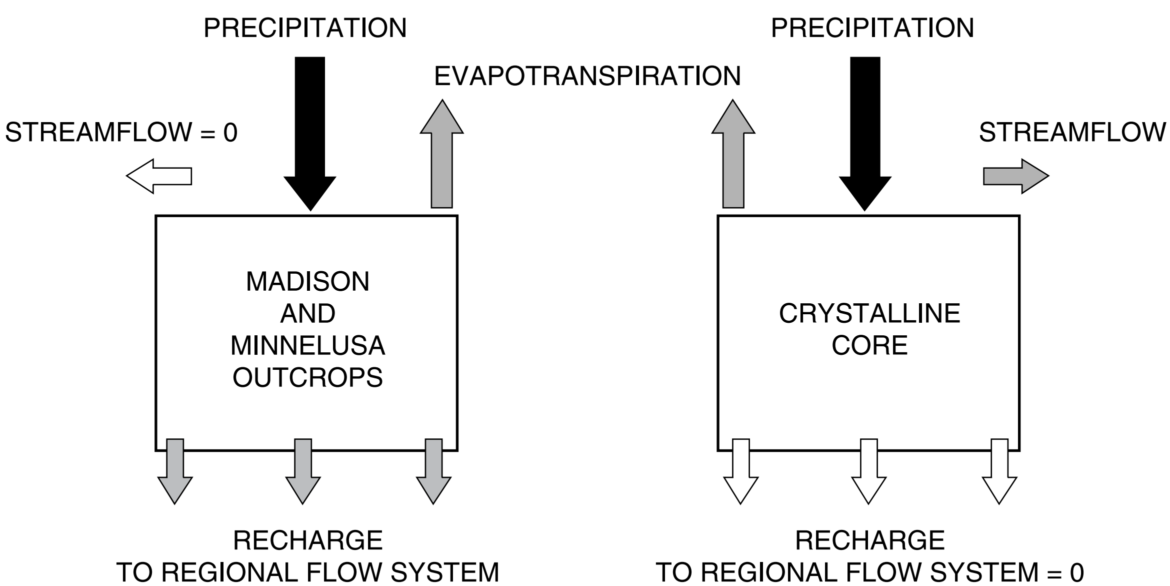


Figure 51. Schematic diagram illustrating recharge and streamflow characteristics for selected outcrop types (from Carter, Driscoll, and Hamade, 2001).

Carter, Driscoll, and Hamade (2001) used estimates of average runoff efficiencies for 1950-98 to develop a map of generalized yield efficiency for the study area (fig. 52). Where applicable, estimated yield efficiencies shown in figure 52 are representative of estimated yield efficiencies for the contributing ground-water areas. For basins where contributing surface- and ground-water areas are assumed to be congruent, yield efficiency is considered equivalent to runoff efficiency. For areas where direct runoff is negligible, yield efficiency is considered equivalent to the efficiency of precipitation recharge. For many gages, estimation of average yield efficiencies for this period required extrapolation of incomplete streamflow records (table 5) using precipitation records. Records were extrapolated to compensate for bias resulting from short-term records for many gages that are skewed towards wet climatic conditions during the 1990's. Yield efficiencies for most of the limestone headwater gages are simply averages for the available periods of record, because relations between streamflow and precipitation for this setting generally are very weak or unrealistic.

Carter, Driscoll, and Hamade (2001) also considered precipitation patterns and topography in contouring yield efficiencies, which provide a reasonable fit with calculated efficiencies (fig. 52). Estimates of contributing areas are not available for the two limestone headwater gages in Wyoming (sites 1 and 14); thus, yield efficiencies could not be adjusted. For Annie Creek (site 16), the calculated yield efficiency (16.4 percent) is lower than for other nearby streams, which may result from extensive mining operations that utilize substantial quantities of water through evaporation for heap-leach processes. For Hay Creek (site 19), the calculated yield efficiency (1.0 percent) is notably lower than the mapped contours, which probably results from precipitation recharge to outcrops of the Inyan Kara Group (fig. 14).

Carter, Driscoll, and Hamade (2001) used relations between yield efficiency and precipitation in developing a GIS algorithm for systematically estimating annual recharge from infiltration of precipitation, based on annual precipitation on outcrop areas. Linear regression and best-fit exponential equations were determined for 11 basins, which include all of the representative crystalline basins (table 5) except Bear Gulch. Exponential equations were in the form of:

$$YE_{annual} = \left[\frac{P_{annual}}{P_{average}}\right]^n \times YE_{average} \qquad (1)$$

where

YE_{annual} = annual yield efficiency, in percent;
P_{annual} = annual precipitation, in inches;
$P_{average}$ = average annual precipitation for 1950-98, in inches;
$YE_{average}$ = average annual yield efficiency for 1950-98, in percent; and
n = exponent.

Best-fit exponents ranged from 1.1 for Elk Creek to 2.5 for Spring Creek. An exponent of 1.6 was chosen as best representing the range of best-fit exponents (Carter, Driscoll, and Hamade, 2001), which allowed a systematic approach to estimation of annual recharge. Scatter plots with the linear regression lines, best-fit exponential curves, and exponential curves using an exponent of 1.6 are shown in figure 53. The three methods provide very similar results through the mid-range of measured precipitation values, with the largest differences occurring for the upper part of the range.

The spatial distribution of average annual yield potential for the Black Hills area is shown in figure 54. Average annual recharge from infiltration of precipitation on outcrops of the Madison Limestone and Minnelusa Formation is shown as an example. Estimates were derived by Carter, Driscoll, and Hamade (2001) using a GIS algorithm that compared digital grids (1,000-by-1,000 meters, including outcrop areas in Wyoming) for annual precipitation, average annual precipitation (fig. 4), and average annual yield efficiency (fig. 53). Annual recharge rates for individual grid cells ranged from 0.4 inch at the southern extremity of the outcrops to 8.7 inches in the northern Black Hills. Although this "yield-efficiency algorithm" was developed initially for estimating precipitation recharge for the Madison and Minnelusa aquifers, applications for estimating streamflow yield and recharge for other aquifers also are appropriate and are used later in this report.

Water Quality

This section summarizes water-quality characteristics for surface water within the study area. More detailed discussions are presented by Williamson and Carter (2001). Standards and criteria that apply to surface waters are presented in the following section, after which common-ion characteristics, anthropogenic effects on water quality, and additional factors relative to in-stream standards are discussed.

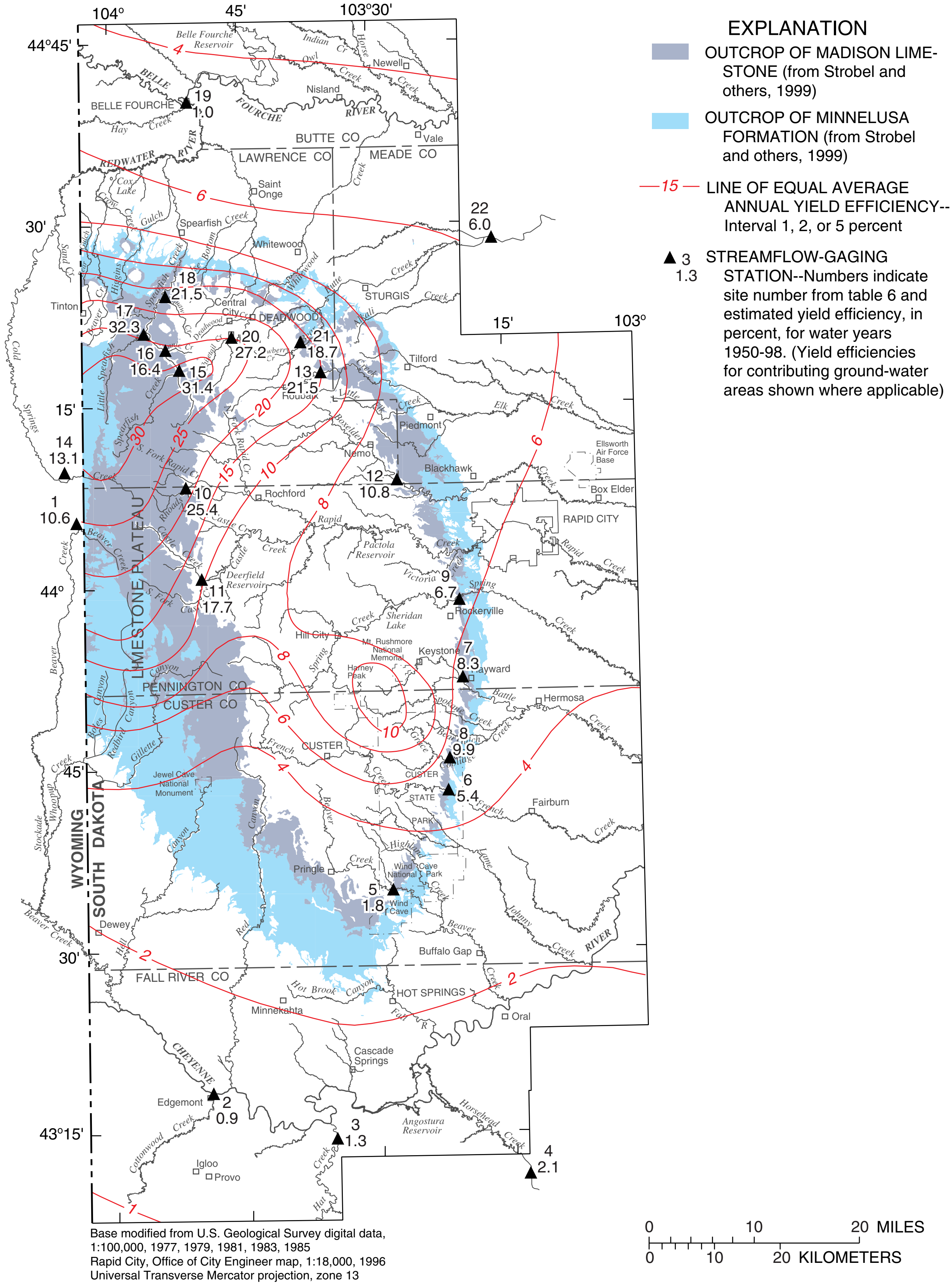


Figure 52. Generalized average annual yield efficiency (in percent of annual precipitation), water years 1950-98 (from Carter, Driscoll, and Hamade, 2001).

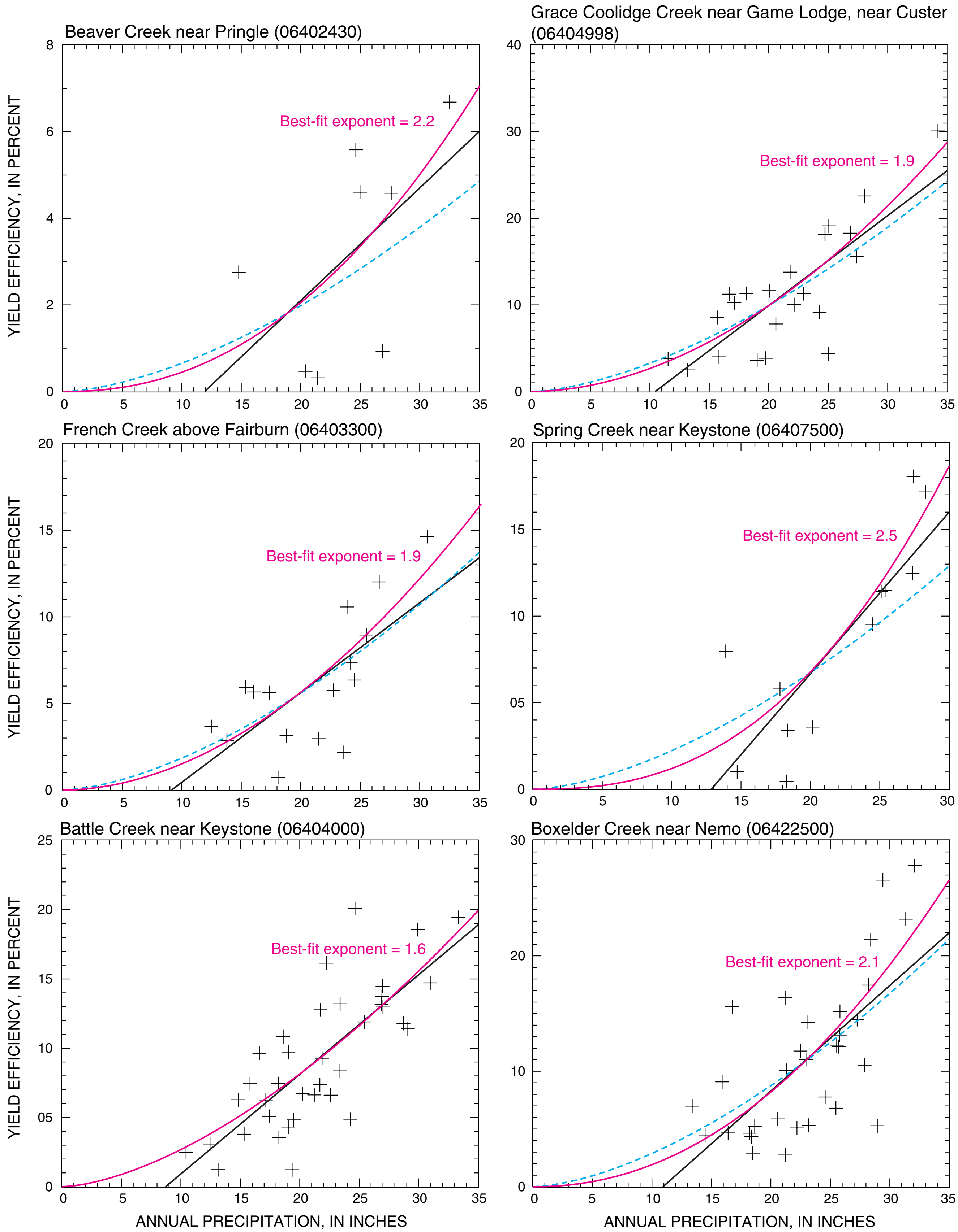


Figure 53. Relations between yield efficiency and precipitation for selected streamflow-gaging stations (modified from Carter, Driscoll, and Hamade, 2001).

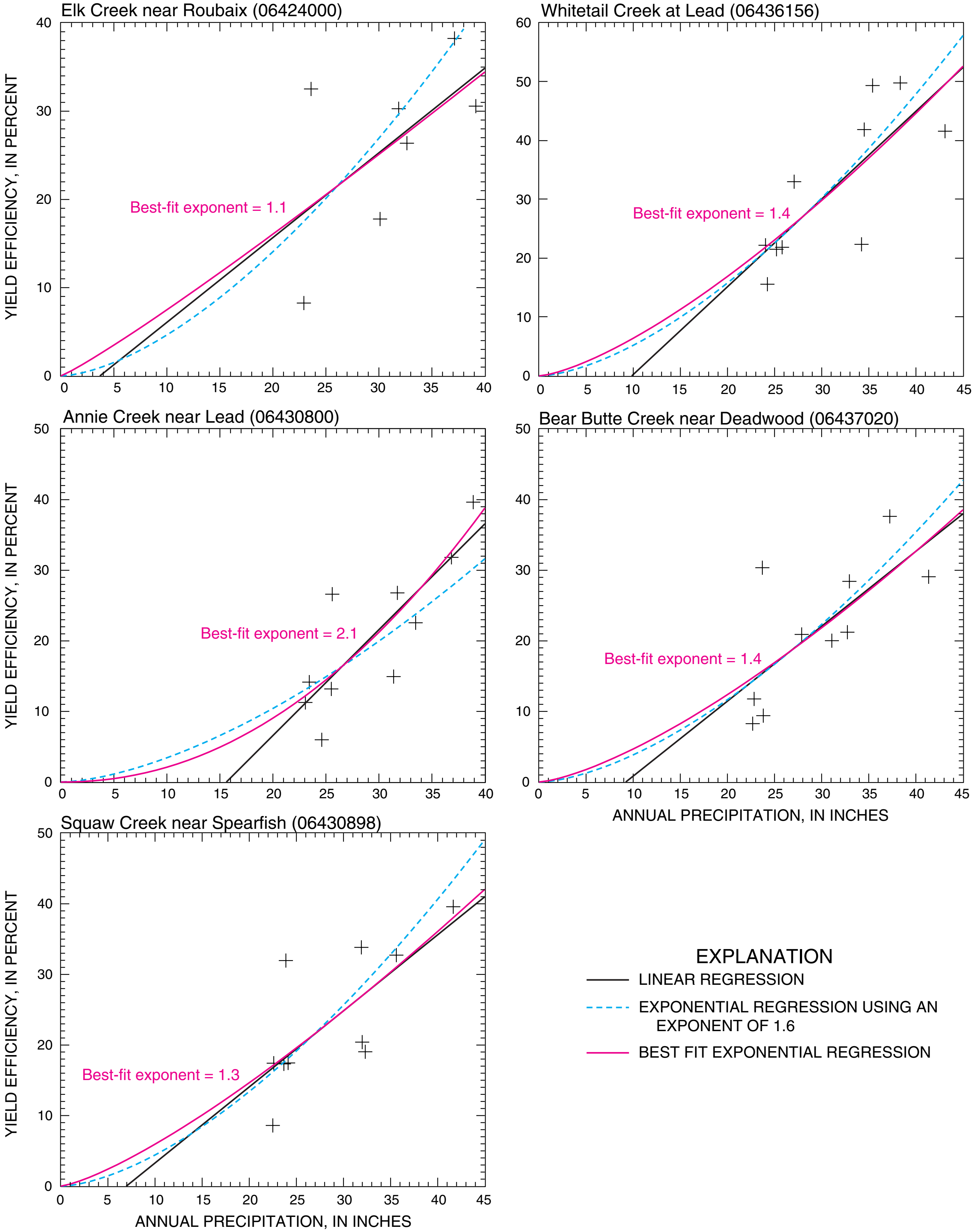
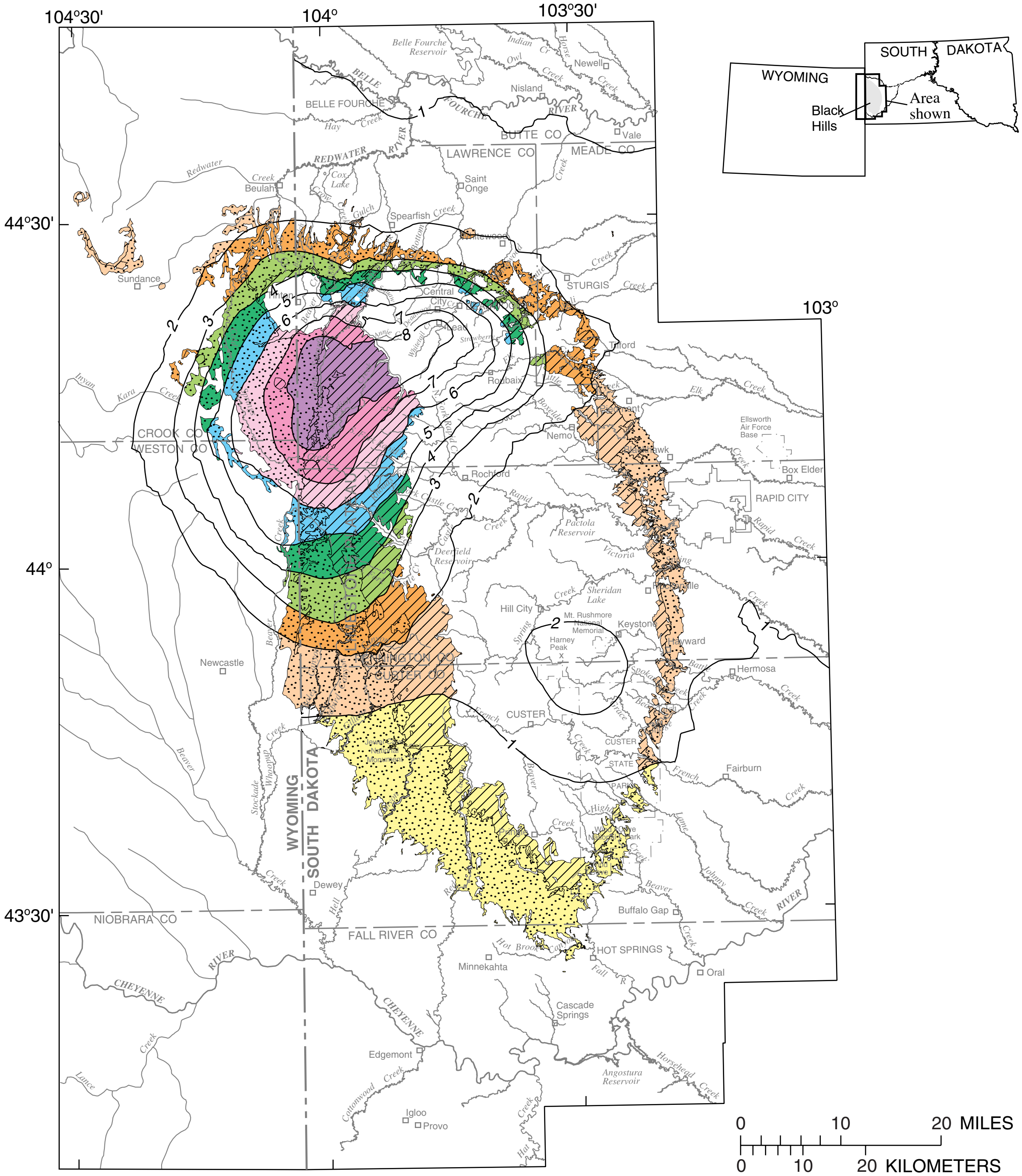


Figure 53. Relations between yield efficiency and precipitation for selected streamflow-gaging stations (modified from Carter, Driscoll, and Hamade, 2001).—Continued



Base modified from U.S. Geological Survey digital data, 1:100,000, 1977, 1979, 1981, 1983, 1985
Rapid City, Office of City Engineer map, 1:18,000, 1996; Universal Transverse Mercator projection, zone 13

EXPLANATION

CONNECTED OUTCROP OF MADISON LIMESTONE FOR WHICH PRECIPITATION RECHARGE IS PRESCRIBED (modified from Strobel and others, 1999; DeWitt and others, 1989)

CONNECTED OUTCROP OF MINNELUSA FORMATION FOR WHICH PRECIPITATION RECHARGE IS PRESCRIBED (modified from Strobel and others, 1999; DeWitt and others, 1989)

—2— LINE OF EQUAL YIELD POTENTIAL--Number indicates average annual yield potential. Interval 1 inch

AVERAGE ANNUAL RECHARGE, IN INCHES

- Less than 1
- 1 to 2
- 2 to 3
- 3 to 4
- 4 to 5
- 5 to 6
- 6 to 7
- 7 to 8
- 8 to 9

Figure 54. Estimated annual yield potential for the Black Hills area, water years 1950-98 (from Carter, Driscoll, and Hamade, 2001). Average annual recharge from precipitation on outcrops of the Madison Limestone and Minnelusa Formation is shown as an example.

Standards and Criteria

Most water-quality standards for surface water are determined by the applicable beneficial-use criteria (table 9) established for individual stream reaches. All streams in South Dakota are designated for the beneficial uses of irrigation and of wildlife propagation and stock watering. Additional beneficial uses are assigned to stream segments as applicable. Designated beneficial uses for specific stream segments are indicated by South Dakota Department of Environment and Natural Resources (2001a). Aquatic-life criteria are estimates of the highest constituent concentrations that aquatic life can be exposed to without adverse effects. The chronic criteria are based on extended exposures, and acute criteria are based on very short-term exposures. The aquatic criteria for several trace elements vary with stream hardness, as shown in figure 55. Drinking-water standards apply only to finished waters that are used for public consumption. These standards do provide another useful basis for comparison, however, and were presented previously in table 4.

Common-ion Chemistry

Common-ion chemistry of surface water in the study area is highly influenced by geology. Within this section, distinctive water-quality characteristics are described for the hydrogeologic settings (limestone headwater, crystalline core, artesian spring, and exterior settings) that were discussed previously (fig. 23). The loss zone setting is not characterized because the primary hydrologic influence is on streamflow, not water quality. Site information for sampling sites representative of the hydrogeologic settings was presented previously in table 5.

Specific conductance can provide an excellent indication of dissolved solids concentrations in surface water (fig. 56). Results of regression analyses (dissolved solids versus specific conductance) for samples available for sites representative of hydrogeologic settings are presented in table 10. Correlations between dissolved solids and specific conductance are strong for all of the hydrogeologic settings except the limestone headwater setting, for which variability in dissolved solids concentrations is small. Table 10 also includes regression information for all available samples for surface-water sites within the study area considered by Williamson and Carter (2001), which include numerous sites in addition to those that are representative of the hydrogeologic settings. The equation provided for all samples generally should provide reasonable estimates of dissolved solids concentrations (based on specific conductance) for surface water within the study area, regardless of hydrogeologic conditions. This equation would be most appropriate for locations where water quality is influenced by multiple hydrogeologic settings, which is a common circumstance within the study area.

Relations between specific conductance and streamflow for the hydrogeologic settings are shown in figure 57. Because of the large number of sites representative of the crystalline core setting, only five sites are shown for this setting. Measurements of specific conductance are abundant because measurements typically are obtained at USGS streamflow-gaging stations when streamflow is measured.

Sites representative of the crystalline core and exterior settings have large variability in streamflow and exhibit generally inverse relations between specific conductance and streamflow (fig. 57). Specific conductance values (and dissolved solids concentrations) generally decrease with increasing streamflow, primarily because of dilution by direct runoff.

Streamflow variability generally is small for the limestone headwater and artesian spring settings (fig. 57) because of dominance by ground-water discharge. Specific conductance values are nearly identical for all of the limestone headwater sites, where the primary geologic influences are outcrops of the Madison Limestone or outcrops of the Minnelusa Formation in areas where anhydrite beds have been removed by dissolution over geologic time.

Specific conductance characteristics are fairly distinct for each of the artesian spring sites (fig. 57). The lowest specific conductance values generally are for Cleghorn Springs, which is located within an outcrop of the Minnelusa Formation (fig. 35). The other artesian spring sites reflect larger influence from anhydrite beds within the Minnelusa Formation or possibly from dissolution of sulfate within other overlying geologic units. Minor influences (reduced specific conductance values) from occasional runoff events are apparent for several sites.

Table 9. Surface-water-quality standards for selected physical properties and constituents

[All constituents in milligrams per liter unless otherwise noted. µS/cm, microsiemens per centimeter at 25 degrees Celsius; µg/L, micrograms per liter; °F, degrees Fahrenheit; °C, degrees Celsius; ≥, greater than or equal to; --, no data available]

Property or constituent	Beneficial-use criteria[1]										Aquatic-life criteria for fisheries[1] (acute/chronic) (µg/L)
	Domestic water supply (mean/daily maximum)	Coldwater permanent fisheries	Coldwater marginal fisheries	Warmwater permanent fisheries	Warmwater semi-permanent fisheries	Warmwater marginal fisheries	Immersion waters	Limited contact waters	Wildlife propagation and stock-watering waters	Irrigation waters	
Specific conductance (µS/cm)	--	--	--	--	--	--	--	--	[2]4,000/7,000	[2]2,500/4,375	--
pH range (standard units)	6.5-9.0	6.6-8.6	6.5-8.8	6.5-9	6.5-9	6.5-9	--	--	6.0-9.5	--	--
Temperature °F/°C (maximum)	--	65/18.3	75/24	80/27	90/32	90/32	--	--	--	--	--
Dissolved oxygen (minimum)	--	≥6.0 ≥7 during spawning	≥5.0	≥5.0	≥5.0	≥4.0	≥5.0	≥5.0	--	--	--
Total dissolved solids	[2]1,000/1,750	--	--	--	--	--	--	--	[2]2,500/4,375	--	--
Total suspended solids	--	[2]30/53	[2]90/158	[2]90/158	[2]90/158	[2]150/263	--	--	--	--	--
Sodium-adsorption ratio	--	--	--	--	--	--	--	--	--	10	--
Chloride	[2]250/438	[2]100/175	--	--	--	--	--	--	--	--	--
Fluoride	4.0	--	--	--	--	--	--	--	--	--	--
Sulfate	[2]500/875	--	--	--	--	--	--	--	--	--	--
Nitrate (as N)	10	--	--	--	--	--	--	--	[2]50/88	--	--
Un-ionized ammonia (as N)	--	0.02	0.02	0.04	0.04	0.05	--	--	--	--	--
Cyanide (free)	--	--	--	--	--	--	--	--	--	--	22/5.2
Dissolved arsenic	--	--	--	--	--	--	--	--	--	--	360/190
Dissolved barium	1.0	--	--	--	--	--	--	--	--	--	--
Dissolved cadmium	--	--	--	--	--	--	--	--	--	--	[3]3.7/[3]1.0
Dissolved copper	--	--	--	--	--	--	--	--	--	--	[3]17/[3]11
Dissolved lead	--	--	--	--	--	--	--	--	--	--	[3]65/[3]2.5
Dissolved mercury	--	--	--	--	--	--	--	--	--	--	2.1/[4]0.012
Dissolved selenium	--	--	--	--	--	--	--	--	--	--	20/5
Dissolved zinc	--	--	--	--	--	--	---	--	--	--	[3]110/[3]100

[1]South Dakota Department of Environment and Natural Resources (2001a).
[2]30-day average/daily maximum.
[3]Hardness-dependent criteria; value given is an example based on hardness of 100 milligrams per liter as $CaCO_3$.
[4]Chronic criteria based on total recoverable concentration.

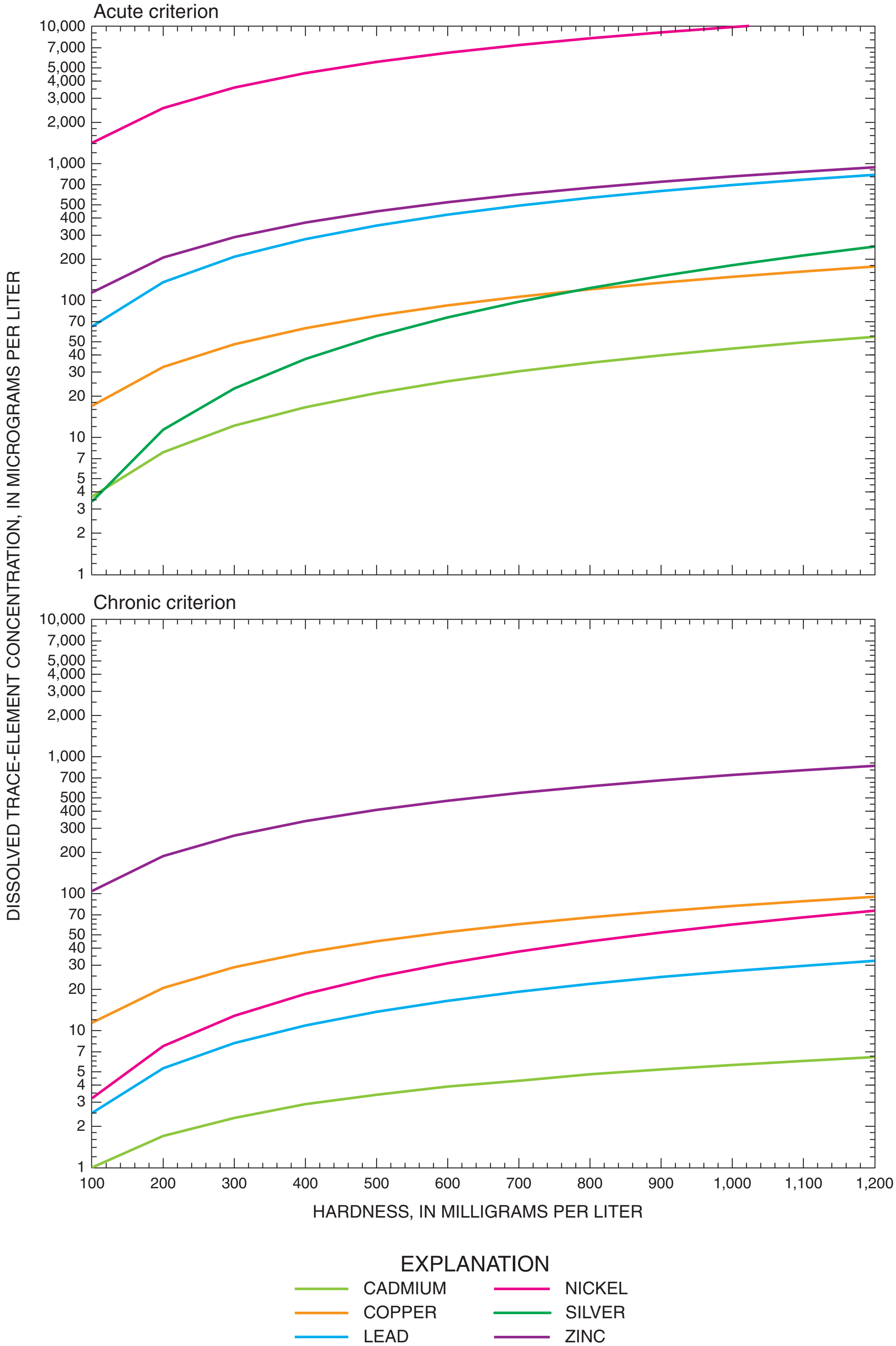


Figure 55. Relations between hardness and freshwater aquatic-life standards for acute and chronic toxicity of selected trace elements (South Dakota Department of Environment and Natural Resources, 1998).

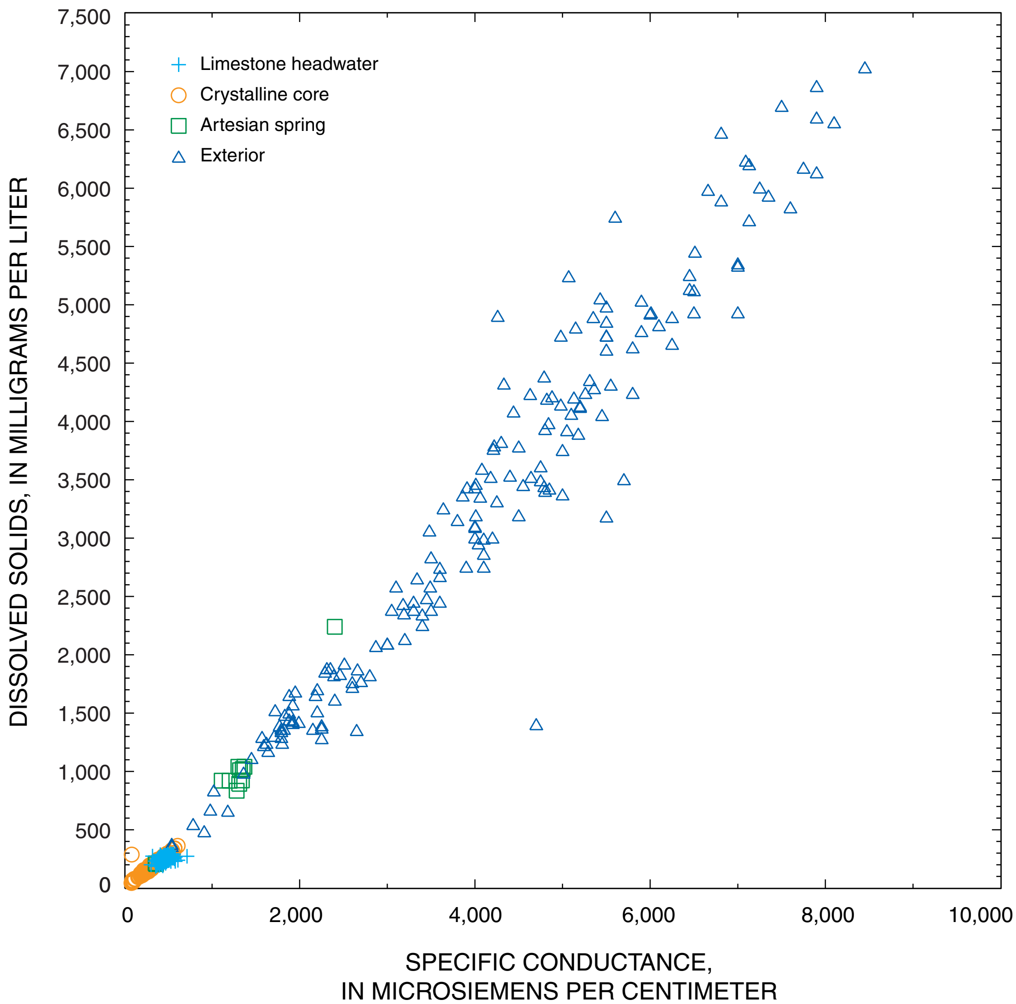


Figure 56. Relations between dissolved solids and specific conductance by hydrogeologic settings (from Williamson and Carter, 2001).

Table 10. Summary of regression information (dissolved solids versus specific conductance), by hydrogeologic setting

[*S* = dissolved solids, in milligrams per liter; *K* = specific conductance, in microsiemens per centimeter]

Hydrogeologic setting	Equation of line	r^2	Number of samples
Headwater spring	$S = 0.21K + 158.16$	0.2437	261
Crystalline core	$S = 0.55K + 15.83$	.8914	136
Artesian spring	$S = 0.93K - 194.22$	.9614	13
Exterior	$S = 0.85K - 249.50$	.9676	174
All[1]	$S = 0.86K - 131.14$	.9692	[1]2,355

[1]Includes numerous surface-water samples in addition to those available for sites representative of hydrogeologic settings.

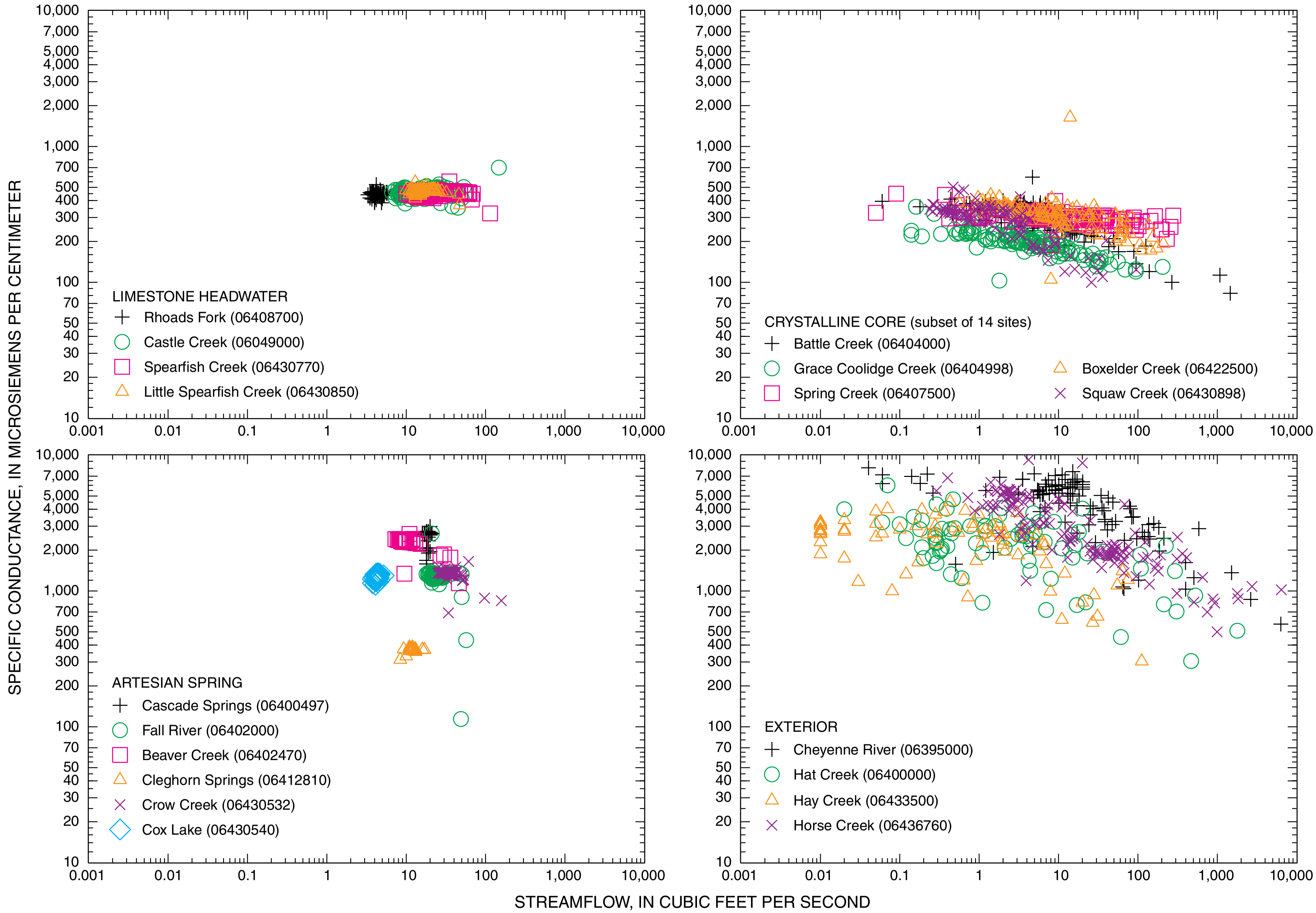


Figure 57. Relations between specific conductance and streamflow for selected sampling sites by hydrogeologic setting (from Williamson and Carter, 2001).

Trilinear diagrams showing ionic proportions (by hydrogeologic setting) for available samples for representative sites are shown in figure 58. Median concentrations for each setting are shown in figure 59. Proportions of common ions are similar for some sites representative of the limestone headwater and crystalline core settings (fig. 58); however, many of the crystalline core sites have smaller proportions of bicarbonate and larger proportions of chloride than limestone headwater sites. The limestone headwater and crystalline core settings have the lowest median concentrations of common ions (fig. 59), which is consistent with low concentrations of dissolved solids, relative to the artesian spring and exterior settings (fig. 56).

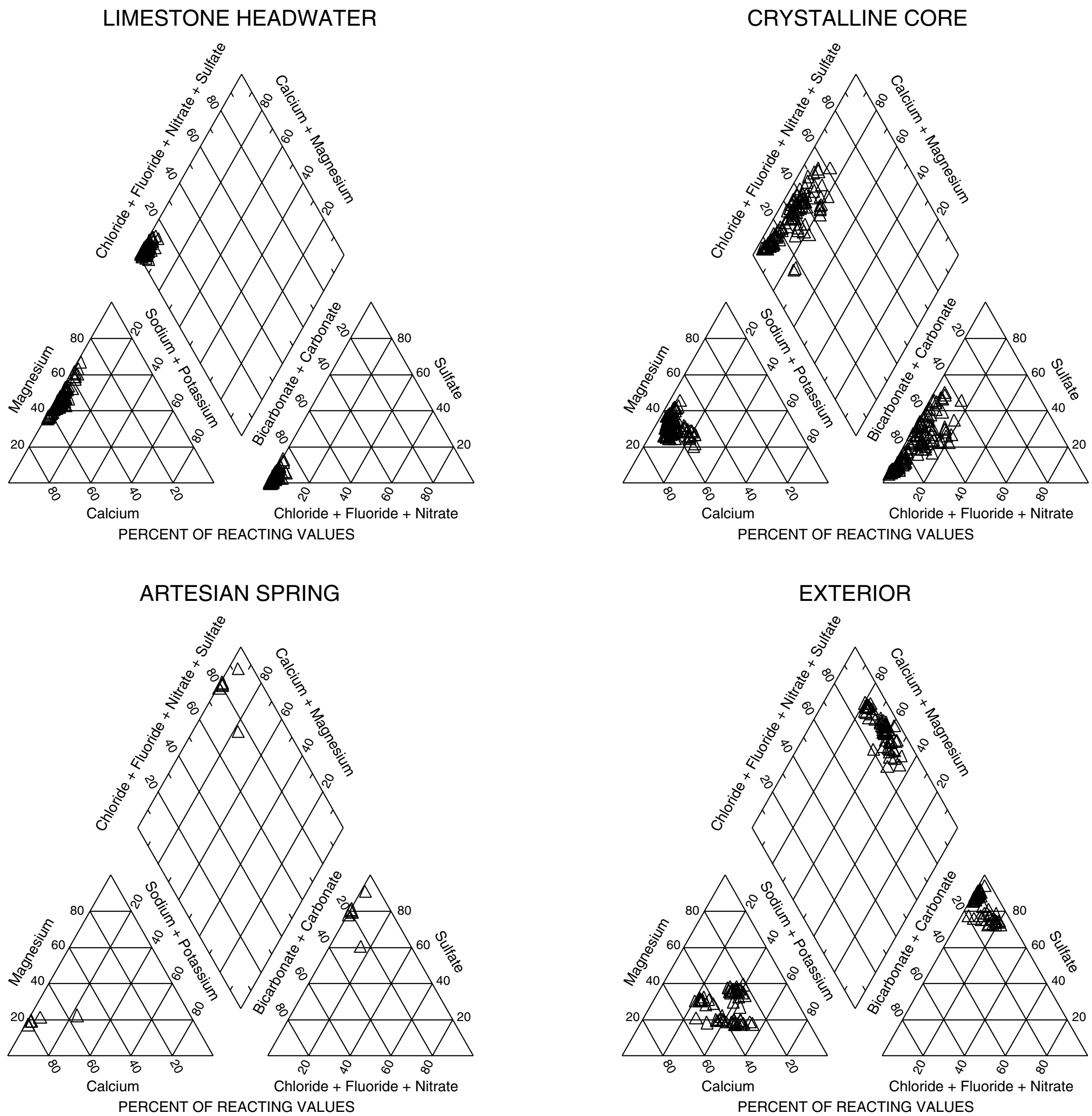


Figure 58. Trilinear diagrams showing proportional concentrations of common ions by hydrogeologic setting (from Williamson and Carter, 2001).

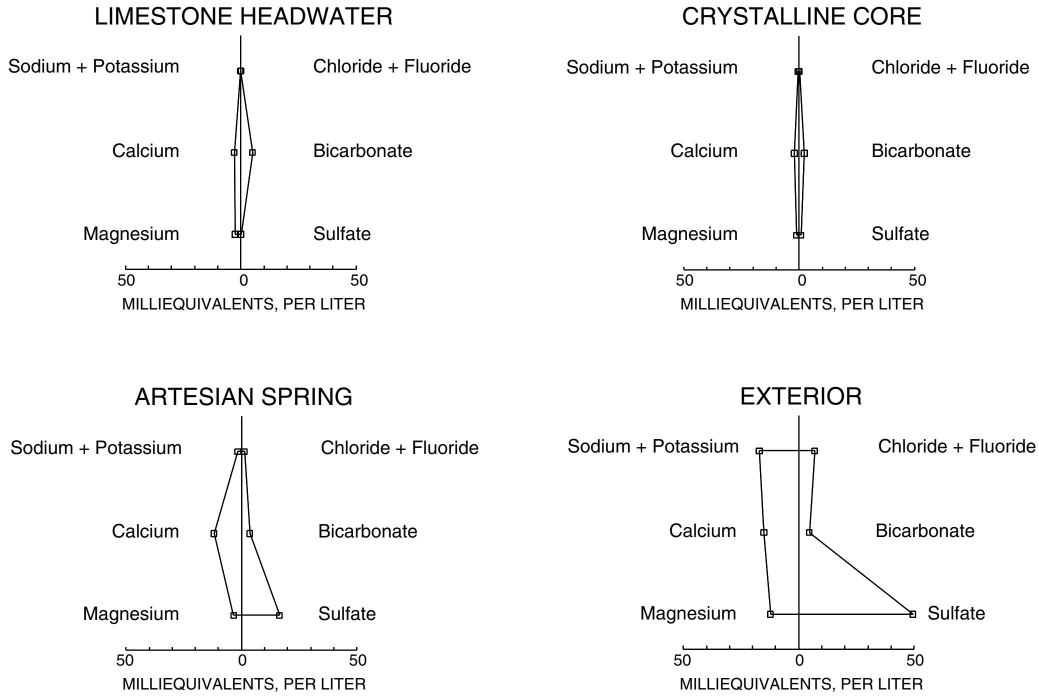


Figure 59. Stiff diagrams (Stiff, 1951) showing median concentrations by hydrogeologic setting (from Williamson and Carter, 2001).

Concentrations of dissolved solids for most sites representative of the artesian spring setting are higher than for the limestone headwater and crystalline core sites (fig. 56). The representative artesian spring sites generally have calcium sulfate water types (figs. 58 and 59), resulting from increased sulfate concentrations. Most water discharged by artesian springs probably originates primarily from the Madison and/or Minnelusa aquifers (Naus and others, 2001); however, high sulfate concentrations probably result primarily from rock/water interactions in the Minnelusa Formation. Other potential sulfate sources may be in contact with shale confining units such as the Spearfish Formation. Several artesian springs with low sulfate concentrations are located upgradient from the anhydrite transition zone in the Minnelusa aquifer (fig. 35).

Exterior sites generally have sodium calcium magnesium sulfate type waters (figs. 58 and 59), with high concentrations of dissolved solids (fig. 56), relative to other sites. Although no stream reaches within the exterior setting are classified for domestic water supply, most samples for exterior sites have dissolved solids concentrations that exceed the 30-day average and maximum criteria of 1,000 and 1,750 mg/L, respectively (table 9). The criteria for wildlife propagation and stock watering also are exceeded for many samples.

Median concentrations of sodium, chloride, magnesium, and sulfate are much higher for the exterior sites than for sites representative of the other settings (fig. 59). Specific conductance values for many of the exterior samples (fig. 56) exceed the criteria for irrigation waters (table 10); however, no samples for this setting exceed the criterion for sodium-adsorption ratio (Williamson and Carter, 2001). The highest chloride concentrations occur in the Cheyenne River at Edgemont, where numerous samples exceed the SMCL of 250 mg/L. Figure 60 presents a spatial distribution of median sulfate concentrations for the study area, including numerous sites in addition to those that are representative of the hydrogeologic settings. The highest sulfate concentrations are for sites within the exterior hydrogeologic setting, which results from contact with outcrops of Cretaceous shales within this area (figs. 9 and 14). Several streams with large influence from artesian springs also have relatively high sulfate concentrations.

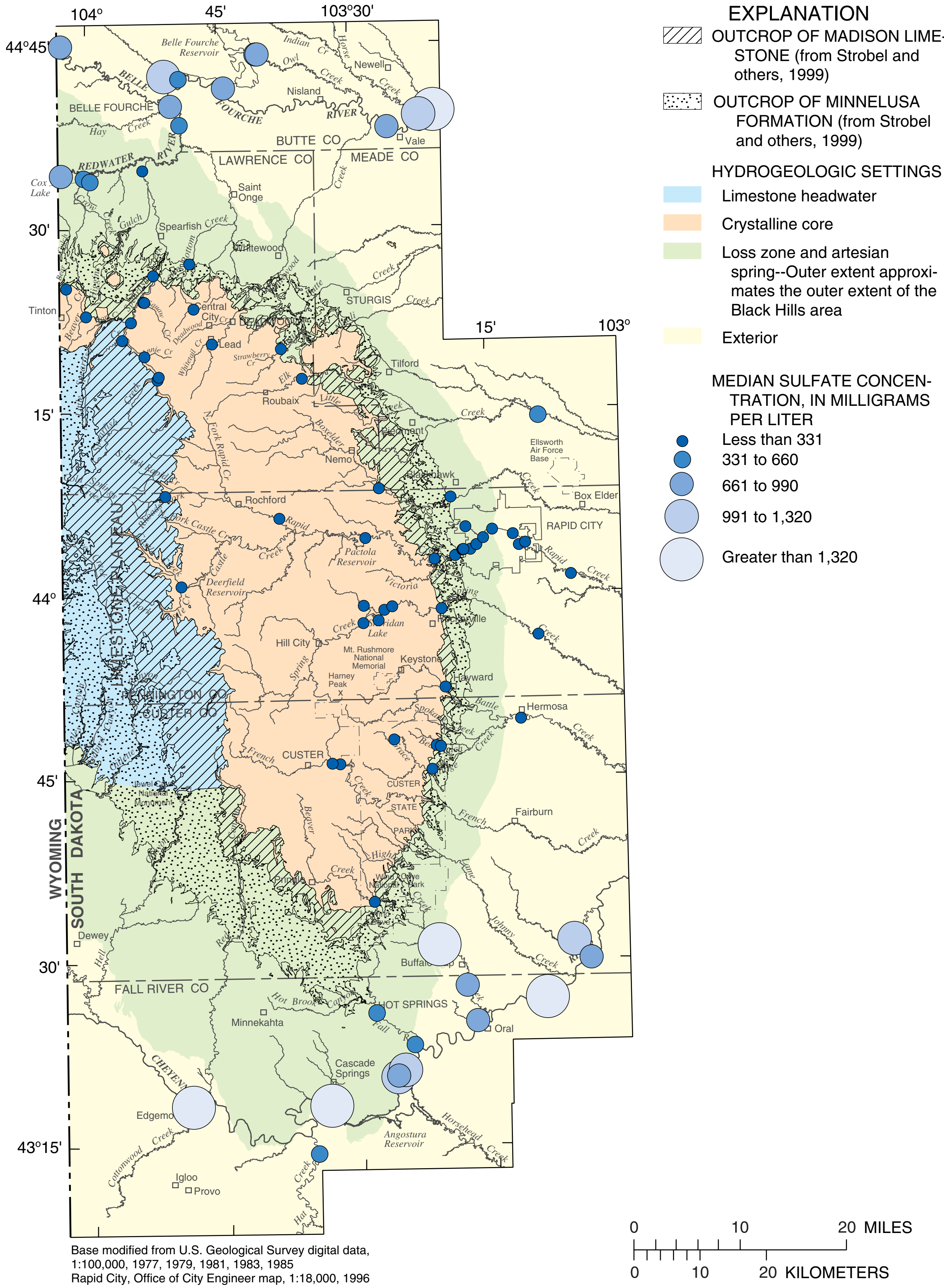


Figure 60. Spatial distribution of median sulfate concentrations in surface water (from Williamson and Carter, 2001).

Anthropogenic Effects

Various human activities, including agriculture, mining, and urban/suburban development, have potential to influence the water quality of streams within the study area. In many cases, anthropogenic influences on water quality cannot necessarily be distinguished from naturally occurring influences. Water-quality sampling for the Black Hills Hydrology Study was performed primarily for the purpose of identifying baseline water-quality conditions; extensive investigations to evaluate effects of anthropogenic influences were not performed as part of the study. Much of the information summarized in this section was obtained from results of site-specific studies by previous investigators. A detailed summary of previous investigations is provided in Williamson and Carter (2001).

Extensive studies of potential effects of irrigation drainage were performed under the National Irrigation Water Quality Program (NIWQP) during the late 1980's by Greene and others (1990) for the Angostura Reclamation Unit and by Roddy and others (1991) for the Belle Fourche Reclamation Project. Concentrations of pesticides in water, bottom sediment, and biota were less than laboratory reporting limits for most samples collected by these investigators. Results of follow-up sampling efforts were presented by Sando and others (2001), who concluded that for both irrigation projects, available data were insufficient to confidently quantify potential increases in dissolved solids loading resulting from irrigation operations.

Selenium, which can be concentrated by irrigation operations, was a special concern of NIWQP because of toxicity and prevalence in irrigation areas in the western United States. Selenium is abundant in the Cretaceous shales surrounding the Black Hills area (figs. 9 and 14), which is reflected in the maximum selenium concentrations measured at selected sites within the study area (fig. 61). Selenium concentrations below the laboratory reporting limit of 1 μg/L have been reported for numerous additional sites (not shown in fig. 61), most of which are located within the limestone headwater, crystalline core, or artesian spring settings. Sando and others (2001) concluded that selenium routinely occurs in concentrations that could be problematic, both upstream and downstream from both irrigation projects. Increased selenium loading resulting from irrigation operations was not discernible for the Angostura Reclamation Unit; however, increased loading probably occurs from return flows in Horse Creek, within the Belle Fourche Reclamation Project.

Effects of mining activity on water quality can be difficult to quantify because natural influences in mineralized areas may have similar effects on water quality. Mining activities generally have been most extensive in the northern Black Hills, primarily in the vicinity of Tertiary intrusive units (fig. 14) where the largest deposits of gold ores tend to be located.

One potential effect of mining is acid-mine drainage, which can result from weathering and oxidation of sulfide minerals that occur in some locations. Torve (1991) examined pH conditions and metals concentrations near abandoned mines along Deadwood Creek and False Bottom Creek. Measured pH values for sites (fig. 62) upstream, immediately downstream, and farther downstream of the mines are presented in figure 63. The pH values are depressed in the immediate vicinity of the mines, but increase farther downstream as a result of the buffering effect of carbonate minerals. Similar areas of low pH also exist in naturally occurring bog-iron areas, which can be found in various locations in the Black Hills area (Luza, 1969; Rice, 1970).

Low pH conditions can cause increased solubility of various metals, which can be abundant in mineralized areas where mining has occurred. Torve (1991) documented elevated concentrations of dissolved cadmium, copper, and zinc that approached or exceeded aquatic-life criteria near mines along Deadwood Creek and False Bottom Creek. Decreased concentrations of dissolved metals were documented farther downstream, where pH levels increased. Approximately 900 abandoned or inactive mines exist throughout the Black Hills area (South Dakota Department of Environment and Natural Resources, 2001b); similar pH and trace-metal conditions occur at a limited number of these locations (Rahn and others, 1996).

Acid-mine drainage has been a problem at the Gilt Edge mine (a large, open-pit, heap-leach-recovery gold mine that was operated during 1986-98), which was listed as a Superfund site by USEPA in 2000 (South Dakota Department of Environment and Natural Resources, 2001b). The mine area is drained by Strawberry Creek, which is a tributary to Bear Butte Creek. Copper concentrations have frequently exceeded aquatic standards in Bear Butte Creek (fig. 64). In addition, dissolved solids concentrations have increased since 1994 in Bear Butte Creek downstream of the confluence with Strawberry Creek, primarily from increases in sodium and sulfate concentrations (Williamson and Carter, 2001).

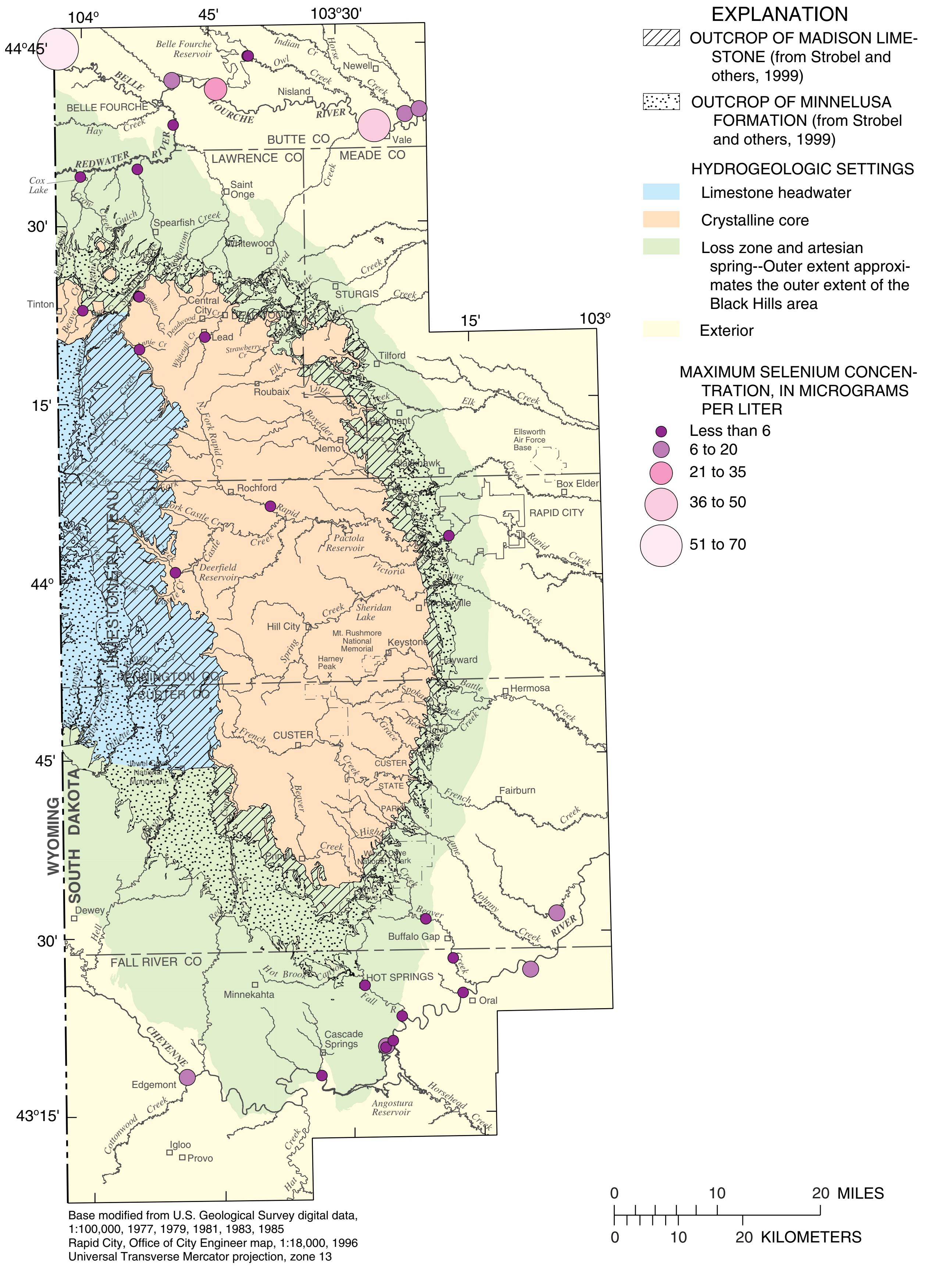


Figure 61. Spatial distribution of maximum selenium concentrations in surface water (from Williamson and Carter, 2001).

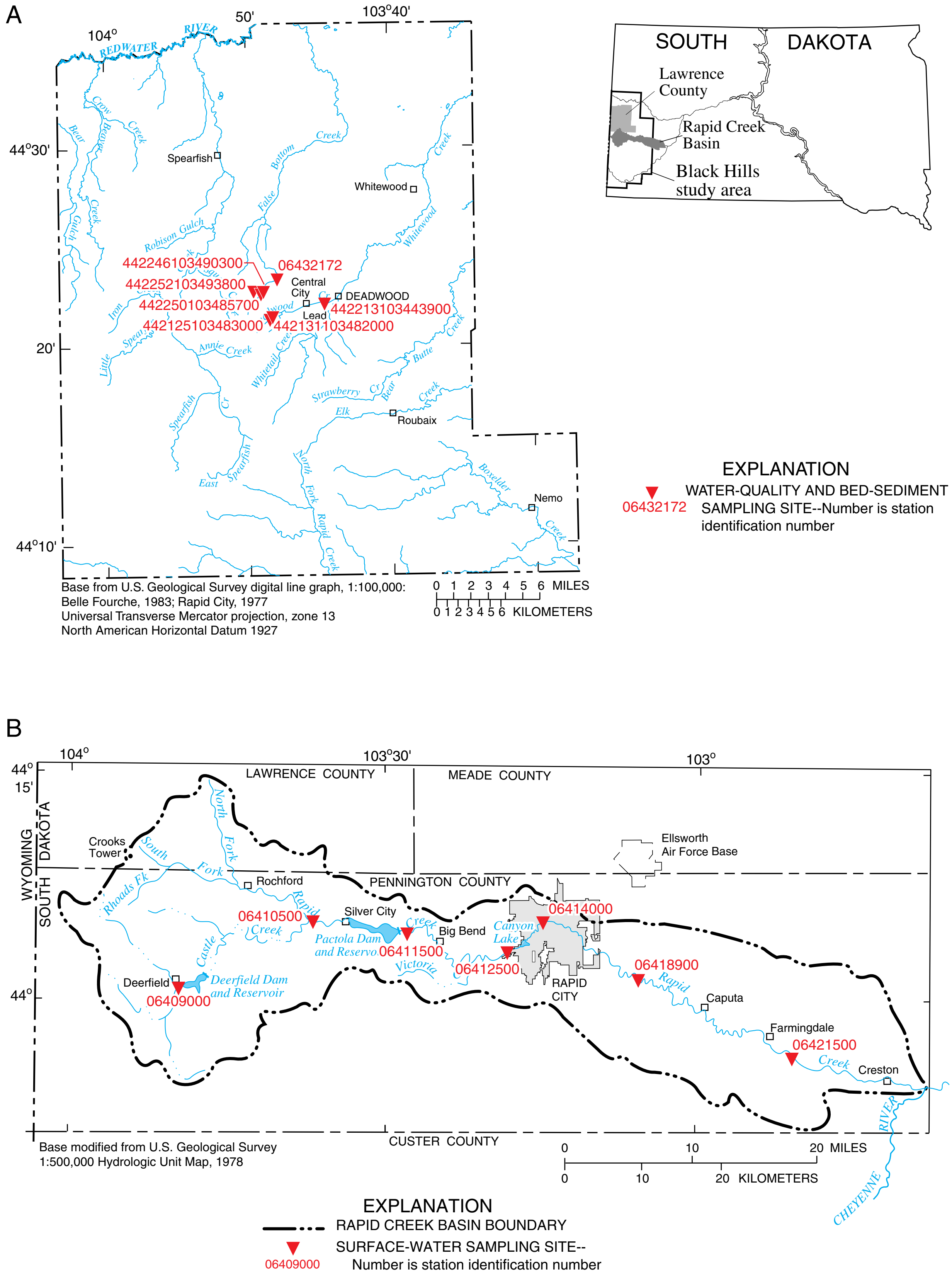


Figure 62. Locations of selected water-quality sampling sites used for analysis of anthropogenic effects in (A) Lawrence County and (B) Rapid Creek Basin.

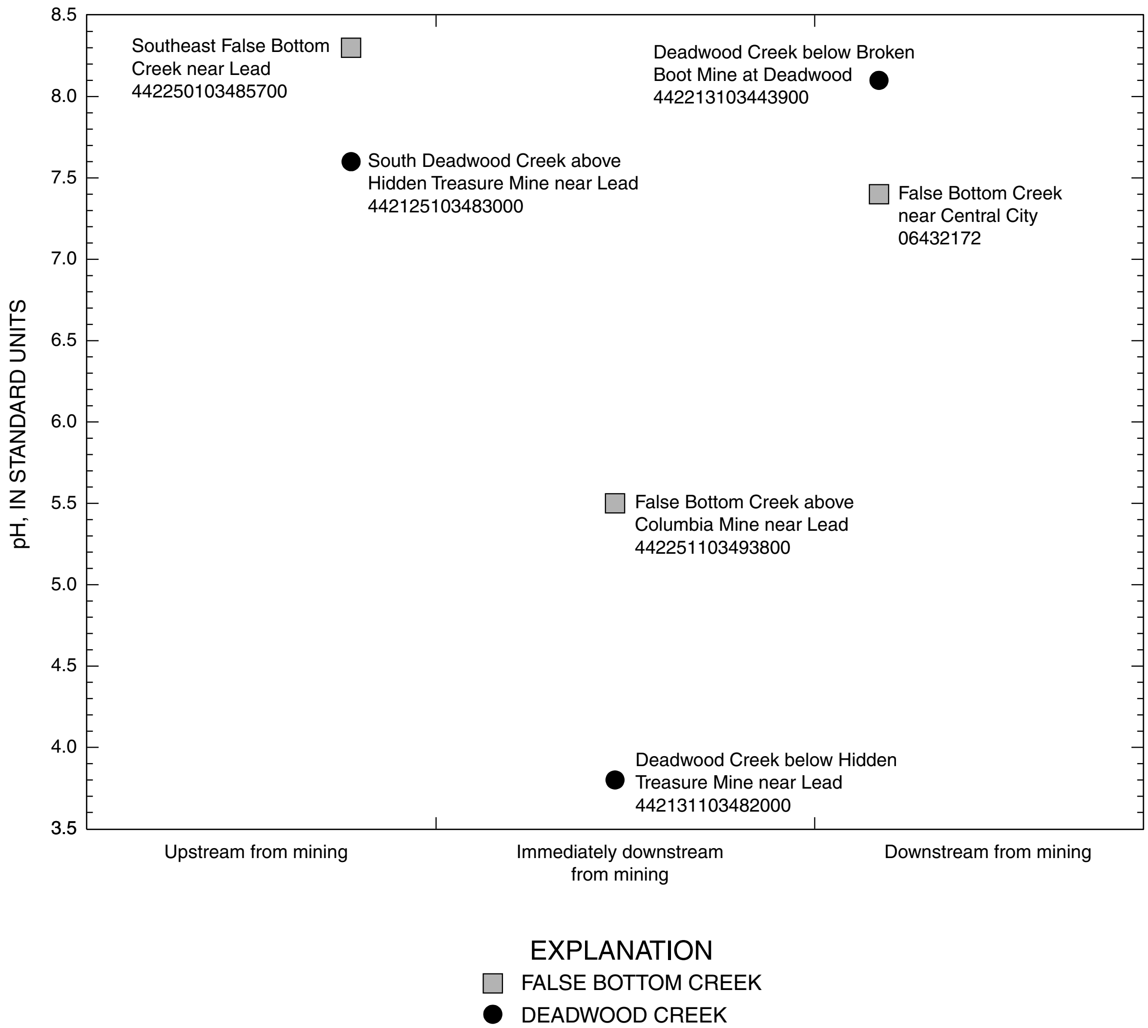


Figure 63. Downstream progression of pH for selected streams influenced by acid-mine drainage (from Williamson and Carter, 2001).

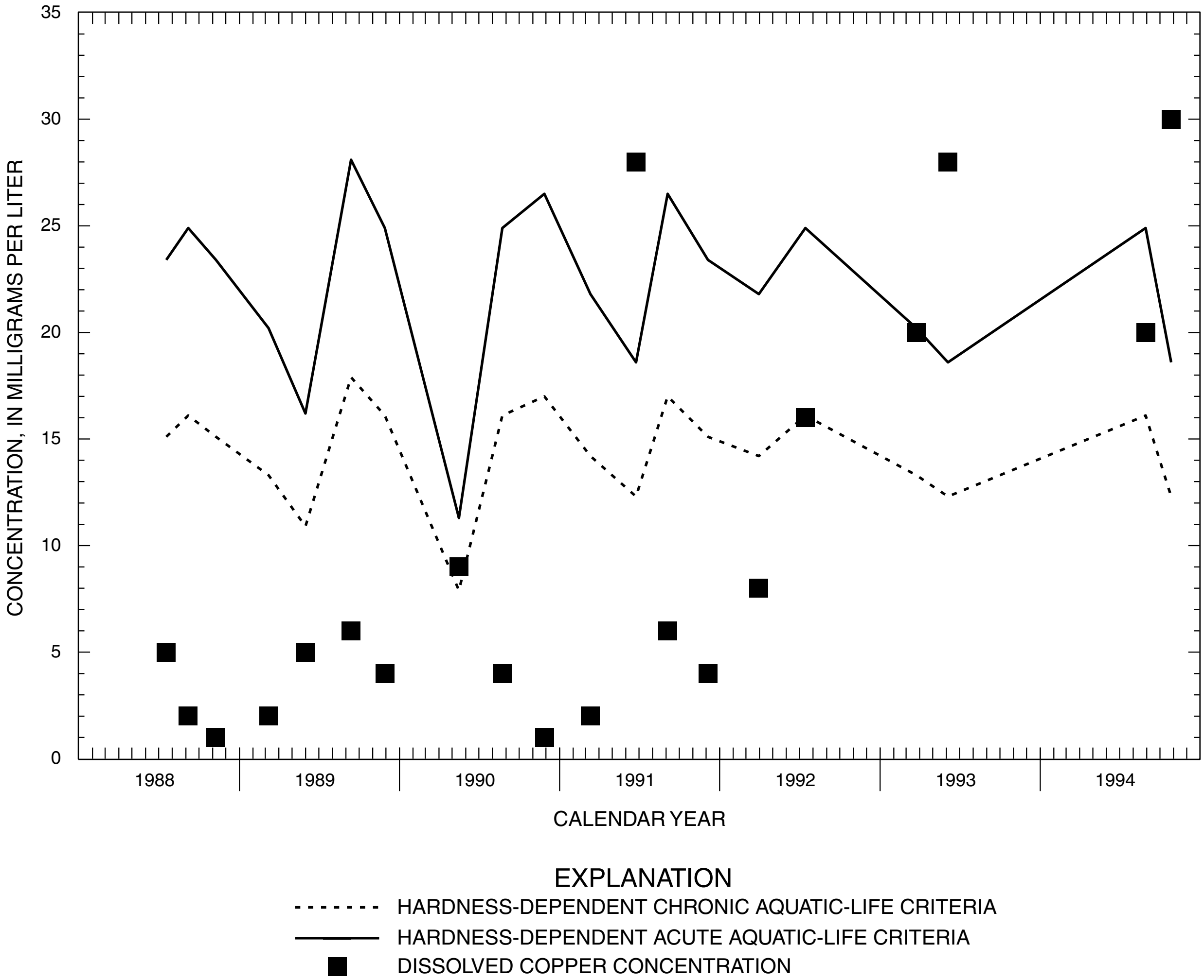


Figure 64. Comparison of dissolved copper concentrations to hardness-dependent chronic and acute aquatic-life criteria for Bear Butte Creek near Deadwood (06437020) (from Williamson and Carter, 2001).

Acid-mine drainage from the Richmond Hill Mine resulted in depressed pH levels in headwater reaches of Squaw Creek (Durkin, 1996). This condition resulted in a need for substantial reclamation efforts in closing the mine, which consisted of capping large deposits of sulfide minerals to prevent exposure to air and water.

The spatial distribution of maximum arsenic concentrations for the study area is shown in figure 65. None of the samples have exceeded the aquatic-life criteria (table 9); however, concentrations exceeding the revised MCL of 10 µg/L (table 4) have been measured at seven sites. One concentration exceeding 50 µg/L has been measured. Concentrations below the laboratory reporting limit of 1 µg/L have been reported for numerous additional sites (not shown in fig. 65). Arsenic commonly is associated with gold ores, with most of the elevated concentrations occurring within the crystalline core setting in locations where mining activities have occurred. Arsenic also is relatively abundant in the Cretaceous shales surrounding the Black Hills area, with concentrations above 5 µg/L measured at upstream locations on the Cheyenne and Belle Fourche Rivers.

Potential effects of mining operations on arsenic concentrations generally cannot be quantified because background samples (prior to mining activities) are not available. Elevated arsenic concentrations in Whitewood Creek and downstream reaches of the Belle Fourche River have been attributed to long-term discharge of mine tailings, which resulted in designation of an 18-mi reach as a Superfund site by USEPA in 1983 (Goddard, 1989). Other constituents of concern along Whitewood Creek include cadmium, copper, cyanide, iron, manganese, mercury, and silver. The largest concentrations of dissolved arsenic typically occur during storm-flow recessions and result from seepage from tailings deposits in the adjacent flood plain. Concentrations of total arsenic exceeding 100 µg/L are common during storm flows when sediments are mobilized.

A variety of factors influencing the solubility of arsenic have been identified by previous investigators. Arsenic is most soluble within a pH range of 8.1 to 8.6; however, adsorption/desorption processes also can affect the mobility of arsenic (Fuller and Davis, 1989; Fuller and others, 1993). The abundance of calcite and ferrihydrite in bed sediments was identified by Williamson and Hayes (2000) as important factors in controlling adsorption/desorption of arsenic. Solubility of arsenic is increased by abundance of calcite because of associated increases in pH. Solubility of arsenic is reduced by abundance of ferrihydrite, which provides iron surfaces for adsorption of arsenic. The highest concentrations of dissolved arsenic typically occur in streams with limited quantities of ferrihydrite in bed sediments. Photosynthesis and respiration by aquatic vegetation within a stream can cause diurnal changes in pH, resulting in fluctuations in arsenic concentrations (Fuller and Davis, 1989).

Arsenic concentrations in Annie Creek that consistently approached the former MCL of 50 µg/L resulted in USEPA designation of a 5-acre site on the National Priorities List (Driscoll and Hayes, 1995). A failed tailings impoundment from an abandoned mine had resulted in large deposits of arsenic-rich tailings along the stream channel. Natural geology and the Annie Creek Mine, which occupies much of the Annie Creek drainage basin, also may contribute to elevated arsenic concentrations in Annie Creek. An important influence is small concentrations of ferrihydrite in bed sediments (Williamson and Hayes, 2000), which limits adsorption/coprecipitation on ferrihydrite and contributes to high concentrations of dissolved arsenic.

Elevated concentrations of nitrite plus nitrate in Annie Creek, with two samples exceeding the MCL of 10 µg/L in 1995 and 1996, were noted by Williamson and Carter (2001). The concentrations probably result from the breakdown of blasting agents (ammonium nitrate mixed with diesel fuel) and cyanide, which is used in the heap-leach extraction process. Two denitrification facilities were installed within the Annie Creek Basin in 1997 and have been effective in reducing nitrite plus nitrate concentrations (South Dakota Department of Environment and Natural Resources, 1998).

Elevated nitrite plus nitrate concentrations also can result from various other influences including fertilizers, animal and human waste, and natural processes that occur within stream riparian systems. Nitrite plus nitrate concentrations for sites representative of hydrogeologic settings are shown in figure 66. In this figure, Annie Creek is separated from other crystalline core sites because of the substantial influence from mining operations, as previously discussed. Concentrations for most sites representative of individual hydrogeologic settings are below 1 µg/L, which probably is indicative of conditions with little human influence. Numerous samples for sites that are not representative of individual hydrogeologic settings are included in an “other sites” category. Most of the elevated concentrations for this category were collected prior to 1980 from Horse Creek near Vale (Williamson and Carter, 2001), which is influenced by irrigation return flows.

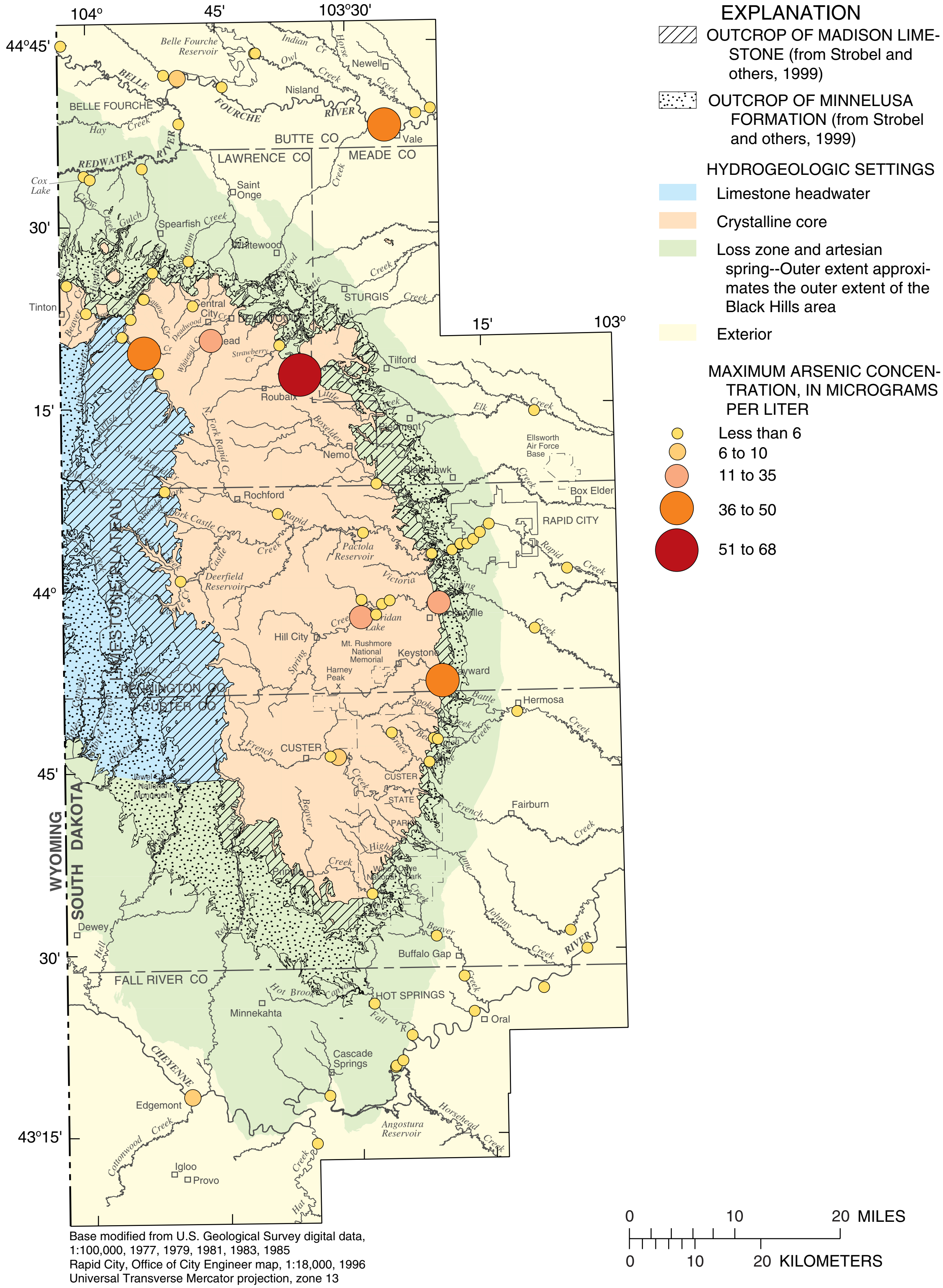


Figure 65. Spatial distribution of maximum arsenic concentrations in surface water (modified from Williamson and Carter, 2001).

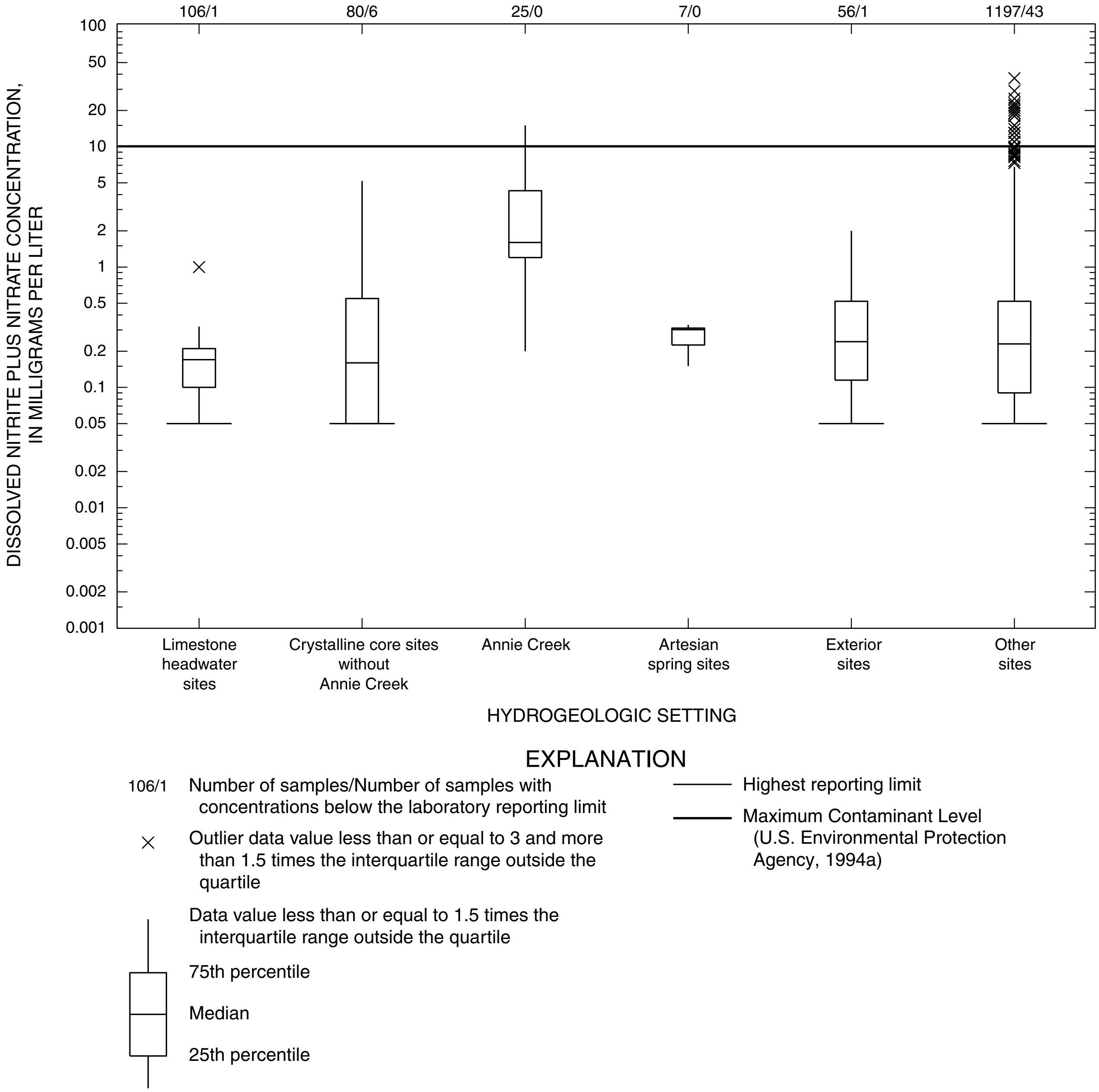


Figure 66. Boxplots of concentrations of dissolved nitrite plus nitrate by hydrogeologic setting, with Annie Creek separated from other crystalline core sites.

Nitrite plus nitrate concentrations for selected sites within the Rapid Creek Basin (fig. 62) are presented in figure 67. Concentrations for the farthest site upstream (06409000, Castle Creek above Deerfield) generally are slightly higher than for the next two sites downstream. One possible nitrogen source, based on land-use activities, is cattle grazing upstream from the farthest site upstream. Rapid Creek at Rapid City (06414000) shows minor influences of nitrite plus nitrate from various possible urban sources. Concentrations at Rapid Creek near Farmingdale (06421500) increase markedly, reflecting various possible influences including non-point urban/suburban sources, effluent discharge from the Rapid City municipal wastewater treatment plant, and various possible influences from agricultural activities.

Numerous constituents in addition to nitrite plus nitrate can influence water quality in urban environments. Williamson and others (1996) documented increased concentrations of various trace metals contributed to Rapid Creek by the Rapid City municipal wastewater treatment plant. Goddard and others (1989) summarized data collected during a USEPA National Urban Runoff Program study in the Rapid City area. In an interpretation of this data set, Harms and others (1983) documented increases in numerous constituents in Rapid Creek associated with urban runoff.

Additional Factors Relative to In-Stream Standards

Beneficial uses for various stream reaches can be influenced by several other factors in addition to those that have been previously discussed. Detailed discussions of low pH conditions associated with acid-mine drainage were presented previously. Low pH values below the desirable range (table 9) also can occur naturally, especially in areas with bog-iron deposits (Luza, 1969). Occurrences generally are in limited stream reaches within the crystalline core setting. Occasional occurrences of pH values exceeding the desirable range have been noted in the limestone headwater and crystalline core settings (Williamson and Carter, 2001).

Beneficial uses for some stream reaches also can be limited by temperature and dissolved oxygen concentrations. In-stream temperatures exceeding beneficial-use criteria for fisheries (table 9) occasionally occur, especially during low-flow conditions. Concentrations of dissolved oxygen below desirable ranges also can occur occasionally, especially during low-flow conditions and when elevated temperatures reduce oxygen solubility. Low concentrations of dissolved oxygen associated with urban runoff also have been documented (Harms and others, 1983).

Concentrations of suspended solids exceeding beneficial-use criteria are relatively common in the Black Hills area. The largest (and most frequent) exceedances typically occur in the exterior setting, where fine-grained particulates associated with clay soils are easily mobilized during high-flow conditions. High concentrations of suspended solids also typically occur during urban runoff events (Harms and others, 1983).

HYDROLOGIC BUDGETS

The quantification of various hydrologic budget components is important for managing and understanding the water resources in the Black Hills area. This section contains summaries of hydrologic budgets that were developed by previous investigators for ground water, surface water, and the combined ground-water/surface-water system. A detailed hydrologic budget for the Madison and Minnelusa aquifers in the Black Hills of South Dakota and Wyoming for 1987-96 was presented by Carter, Driscoll, Hamade, and Jarrell (2001). Detailed hydrologic data sets were available for this period, during which changes in ground-water storage were considered negligible. Basic hydrologic budgets for the Black Hills of South Dakota for 1950-98 were presented by Driscoll and Carter (2001) for: (1) the Madison and Minnelusa aquifers; (2) other bedrock aquifers; (3) surface water; and (4) the combined ground-water/surface-water system. This timeframe includes an extended period of particularly dry climatic conditions that occurred during the 1950’s (fig. 8) and the particularly wet conditions that occurred during the middle to late 1990’s.

All hydrologic budgets were developed from the following basic continuity equation, which states that for any designated volume:

$$\Sigma Inflows - \Sigma Outflows = \Delta Storage \qquad (2)$$

where:

$\Sigma Inflows$ = sum of inflows;

$\Sigma Outflows$ = sum of outflows; and

$\Delta Storage$ = change in storage.

Thus, a positive $\Delta Storage$ results when inflows exceed outflows.

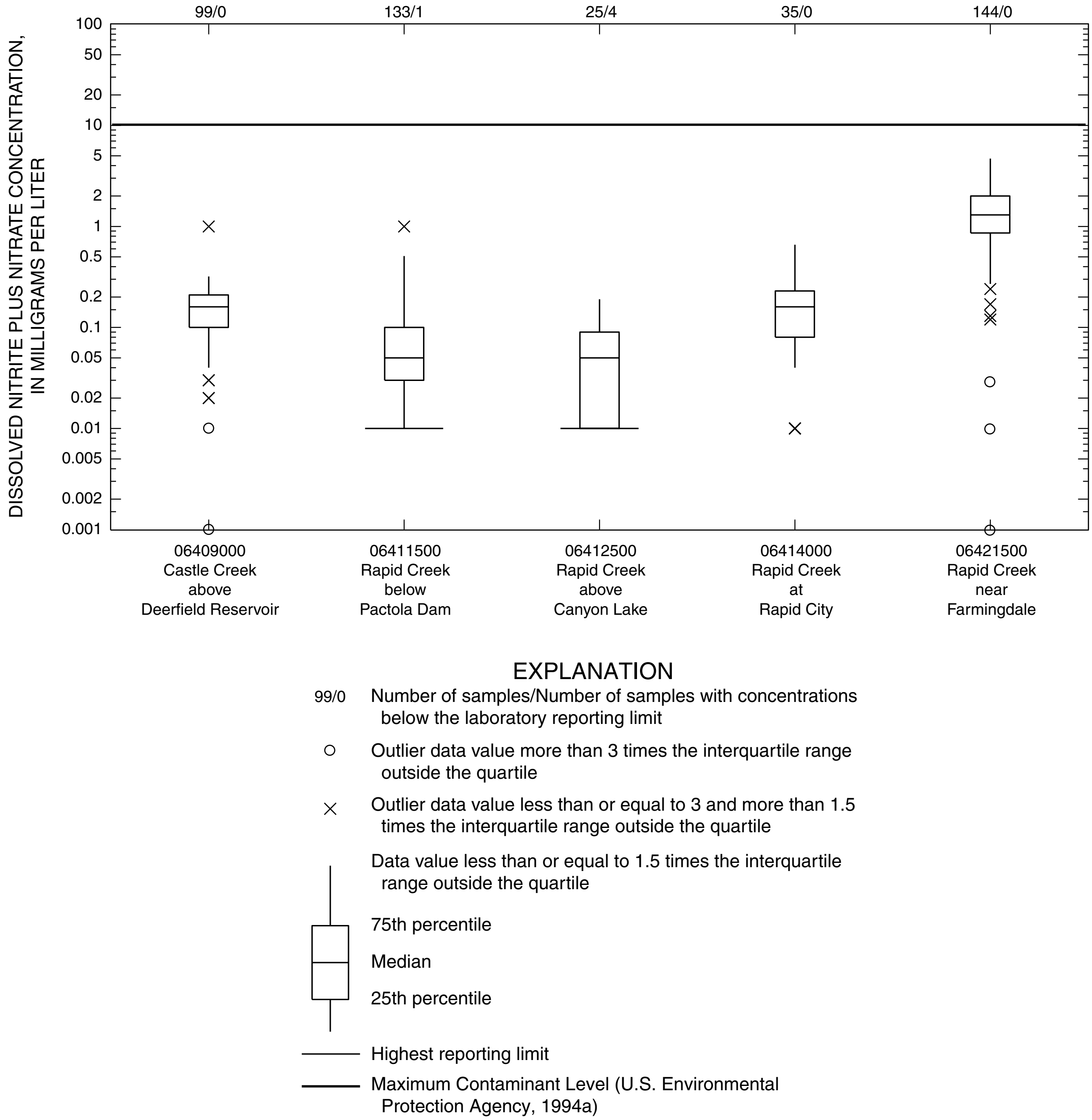


Figure 67. Boxplots of concentrations of dissolved nitrite plus nitrate within the Rapid Creek Basin.

Methods for Estimating Basin Yield and Recharge

Total yield, which is considered to be the sum of runoff plus recharge, was estimated over selected areas using the "yield-efficiency algorithm" discussed in a previous section. Estimates of precipitation, total yield, and evapotranspiration for the entire study area are presented in table 11. For 1950-98, precipitation averaged 18.98 inches or about 5.2 million acre-ft/yr. This is equivalent to an average flow rate of about 7,240 ft^3/s. Of this amount, total yield is estimated as about 440,600 acre-ft/yr (about 608 ft^3/s), which is equivalent to about 1.59 inches over the study area. Thus, evapotranspiration is estimated as 17.39 in/yr, which accounts for about 92 percent of annual precipitation.

Total yield was apportioned between precipitation recharge and runoff on the basis of recharge factors for each aquifer (table 12). With the exception of localized aquifers in the crystalline core, as discussed later, recharge was estimated by multiplying the total yield by the recharge factor, which is the fraction of total yield estimated to be recharge for a particular unit. The remainder of total yield (if any) was assumed to contribute to runoff from the outcrop area. Of the total average annual precipitation in the study area (table 11), runoff accounts for about 4.9 percent (352 ft^3/s), and precipitation recharge accounts for about 3.5 percent (256 ft^3/s). Thus, only a small fraction of the water originating from precipitation recharges the aquifers by infiltrating in outcrop areas.

Table 11. Estimates of average precipitation, precipitation recharge, runoff, total yield, and evapotranspiration for the study area, water years 1950-98

[Modified from Driscoll and Carter (2000)]

Units	Precipitation	Precipitation recharge	Runoff	Total yield (Precipitation recharge + runoff)	Evapotranspiration
Acre-feet per year	5,245,400	185,500	255,100	440,600	4,804,800
Cubic feet per second	7,240.4	256	352	608.2	6,632.2
Inches per year	18.98	0.67	0.92	1.59	17.39

Table 12. Recharge factors and outcrop areas for bedrock aquifers

[From Driscoll and Carter (2001). --, not applicable]

Aquifer unit	Recharge factor[1]	Outcrop area (acres)
Localized aquifers in crystalline core area (Precambrian/Tertiary/Other[2])	--	616,800
Deadwood	0.80	66,200
Madison	1.00	292,600
Minnelusa	1.00	300,000
Minnekahta	1.00	72,100
Inyan Kara	.80	219,700
Jurassic-sequence semiconfining unit	.40	75,800
Cretaceous-sequence confining unit	.05	716,100

[1]Fraction of total yield estimated to result in recharge, with remainder (if any) assumed to contribute to runoff.

[2]Other consists of other units within the crystalline core area, including: (1) isolated outcrops of the Deadwood Formation, Madison Limestone, and Minnelusa Formation, and Minnekahta Limestone above the loss zones; and (2) unconsolidated sedimentary deposits.

As previously discussed, direct runoff from outcrops of the Madison Limestone and Minnelusa Formation seldom occurs. Thus, all precipitation on these outcrops that is not evapotranspirated was assumed to recharge the aquifers; hence, recharge factors for these aquifers were assumed to be 1.00. The recharge factor for the Minnekahta aquifer also was assumed to be 1.00, based on similar formation properties between the Minnekahta Limestone and Madison Limestone. Recharge factors for the Inyan Kara and Deadwood aquifers were assumed to be 0.80 because these formations contain more shale layers than the Madison, Minnelusa, and Minnekahta Formations. The Sundance aquifer within the Jurassic-sequence semiconfining unit is a productive aquifer, but only constitutes about one-half of the outcrop area of the total unit. Thus, a recharge efficiency of 0.40 (one-half of 0.80) was assumed for the entire Jurassic-sequence semiconfining unit. Likewise, the Newcastle Sandstone contains a productive aquifer within the Cretaceous-sequence confining unit; however, the Newcastle Sandstone constitutes only a small portion of the total unit in outcrop area. Thus, a recharge factor of 0.05 was assumed for the entire Cretaceous-sequence confining unit.

Recharge does occur to numerous localized aquifers within the crystalline core area, especially where extensive fractures and weathered zones are present. These aquifers are not considered regional, however, as indicated by the fact that wells constructed in Precambrian rocks in western South Dakota outside of the Black Hills have not encountered measurable amounts of ground water (Rahn, 1985). Thus, regional ground-water flow in the Precambrian rocks was assumed to be negligible although some flow may occur in the upper weathered zone. Recharge to localized aquifers in the crystalline core area was assumed equal to well withdrawals from this unit. Actual recharge to the crystalline core aquifers must be much larger than this estimate to accommodate ground-water discharge that contributes to base flow of many streams. Recharge conditions are highly transient and have large spatial variability; thus, quantification was not attempted.

Within the crystalline core area, numerous erosional remnants of sedimentary outcrops occur that are "isolated" from regional ground-water flow systems as described in a previous section (fig. 14). Precipitation recharge was prescribed only for "connected" outcrops and was not prescribed for isolated outcrops. Infiltration of precipitation on isolated outcrops was assumed to contribute to streamflow, which eventually has potential to provide streamflow recharge to the Madison and Minnelusa aquifers. Additional methods beyond identification of isolated and connected outcrop areas were used in quantifying precipitation recharge for the Deadwood aquifer as described by Driscoll and Carter (2001). Additional details regarding precipitation recharge are discussed in the following sections.

Ground-Water Budgets

Ground-water budgets were developed for five major, sedimentary bedrock aquifers within the study area (Deadwood, Madison, Minnelusa, Minnekahta, and Inyan Kara aquifers) and for additional minor aquifers within the Jurassic-sequence semiconfining unit and Cretaceous-sequence confining unit. A budget also was developed for localized aquifers within the crystalline core area, which is dominated by Precambrian igneous and metamorphic rocks, but also includes Tertiary igneous rocks, erosional remnants of various sedimentary rocks, and minor, unconsolidated sedimentary deposits. These localized aquifers are subsequently referred to as the crystalline core aquifers. A combined budget was developed for the Madison and Minnelusa aquifers because most of the budget components cannot be quantified individually for these two aquifers; this budget is presented prior to budgets for the other bedrock aquifers.

Long-term budgets were developed for the period 1950-98, during which changes in ground-water storage were assumed to be negligible. Various components in equation 1 for ground-water budgets are schematically illustrated in figure 16 for the Madison aquifer. Inflows may include recharge, vertical leakage from adjacent aquifers, and lateral ground-water inflows across the study area boundary. Recharge occurs at or near land surface and includes infiltration of precipitation on outcrops of the bedrock units and streamflow recharge, which occurs where streams cross the outcrops. Streamflow recharge was considered an inflow component only for the Madison and Minnelusa aquifers. Although the Minnekahta aquifer also receives limited recharge from streamflow losses, this recharge probably is very small relative to streamflow recharge to the Madison and Minnelusa aquifers and could not be quantified.

Outflows include springflow, well withdrawals, vertical leakage to adjacent aquifers, and lateral ground-water outflow across the study area boundary. Springflow includes headwater springs and artesian springs. Headwater springs, which generally occur near the base of the Madison Limestone in the Limestone Plateau area, were considered an outflow component for only the Deadwood, Madison, and Minnelusa aquifers. Artesian springs, which constitute a form of leakage but are treated as a separate component because of magnitude and measurability, were considered an outflow component for only the Madison and Minnelusa aquifers.

Vertical leakage to and from adjacent aquifers, which is difficult to quantify and cannot be distinguished from lateral ground-water inflows or outflows across the study area boundaries, probably is small relative to other budget components in most cases. Thus, for budget purposes, leakage was included with ground-water inflows and outflows. Assuming that Δ*Storage* is equal to zero, the sum of the inflows is equal to the sum of the outflows, and the hydrologic budget equation can be rewritten as:

$$\text{Ground-water}_{\text{outflow}} - \text{Ground-water}_{\text{inflow}}$$
$$= \text{Recharge} - \text{Headwater springflow}$$
$$- \text{Artesian springflow} - \text{Well withdrawals} \qquad (3)$$

The terms on the right side of equation 3 generally can be quantified more accurately than the terms on the left. Therefore, net ground-water flow (outflow minus inflow) from the study area can be calculated as the residual, given estimates for the other budget components.

Budgets for Madison and Minnelusa Aquifers

Recent investigations have provided extensive information regarding various budget components for the Madison and Minnelusa aquifers, which are presented in table 13. Many of the budget components were based on hydrologic measurements with relatively short periods of record. Precipitation records indicate that prolonged drought conditions occurred prior to many available hydrologic records. Thus, Carter, Driscoll, and Hamade (2001) estimated recharge for the Madison and Minnelusa aquifer for 1931-98 (fig. 68) to address particularly dry conditions that occurred during the 1930's and 1950's. Estimates of streamflow recharge, precipitation recharge, and combined recharge from both sources (for the expanded area that includes part of Wyoming) are shown in figure 68. The 1931-98 averages are smaller than the 1950-98 averages (table 13) because of prolonged droughts during the 1930's and 1950's. Streamflow recharge is relatively steady; however, precipitation recharge is highly variable, depending on annual precipitation amounts.

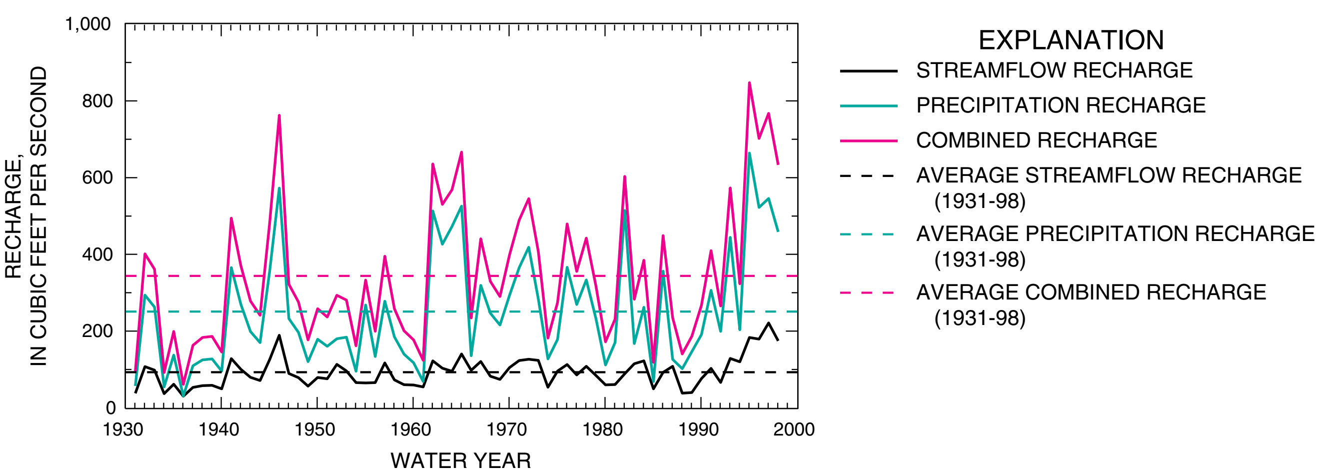


Figure 68. Annual recharge to the Madison and Minnelusa aquifers, water years 1931-98, in the Black Hills of South Dakota and Wyoming (from Carter, Driscoll, and Hamade, 2001).

Table 13. Estimated annual hydrologic budget components for the Madison and Minnelusa aquifers, water years 1931-98, for the Black Hills of South Dakota and Wyoming

[All estimates in cubic feet per second. --, not determined]

Water year	Recharge[1]			Headwater springflow[2]	Artesian springflow[3]	Well withdrawals[3]	Net ground-water outflow[3]
	Streamflow	Precipitation	Total				
1931	38.17	57.37	95.53	3.6	--	--	--
1932	107.61	293.82	401.44	76.7	--	--	--
1933	98.50	262.78	361.28	63.6	--	--	--
1934	37.38	54.70	92.08	4.5	--	--	--
1935	61.71	137.54	199.25	28.6	--	--	--
1936	30.45	31.08	61.53	1.0	--	--	--
1937	53.55	109.75	163.30	16.4	--	--	--
1938	58.12	125.31	183.44	21.4	--	--	--
1939	58.78	127.53	186.31	24.0	--	--	--
1940	49.57	96.18	145.75	13.5	--	--	--
1941	128.70	365.63	494.34	99.7	--	--	--
1942	100.57	269.84	370.41	65.4	--	--	--
1943	79.75	198.96	278.72	48.5	--	--	--
1944	71.33	170.29	241.62	36.2	--	--	--
1945	125.98	356.35	482.33	102.1	--	--	--
1946	189.51	572.68	762.19	190.8	--	--	--
1947	89.69	232.79	322.47	56.5	--	--	--
1948	79.14	196.87	276.01	45.9	--	--	--
1949	56.72	120.53	177.24	20.1	--	--	--
1950	79.50	178.87	258.36	40.0	--	--	--
1951	76.09	160.75	236.84	30.5	--	--	--
1952	113.52	180.03	293.55	41.7	--	--	--
1953	96.62	184.32	280.94	45.8	--	--	--
1954	66.10	95.61	161.71	16.9	--	--	--
1955	65.04	268.06	333.09	71.0	--	--	--
1956	65.90	134.06	199.96	25.2	--	--	--
1957	117.12	278.05	395.17	70.7	--	--	--
1958	73.20	185.27	258.47	39.8	--	--	--
1959	60.53	140.36	200.89	26.2	--	--	--
1960	59.57	117.59	177.16	26.7	--	--	--
1961	54.97	68.88	123.85	7.0	--	--	--
1962	122.52	513.23	635.75	158.7	--	--	--
1963	103.64	426.54	530.18	128.5	--	--	--
1964	95.48	472.86	568.33	157.9	--	--	--
1965	140.80	525.80	666.60	161.7	--	--	--
1966	98.23	136.11	234.33	21.1	--	--	--
1967	121.00	319.45	440.45	83.3	--	--	--

Table 13. Estimated annual hydrologic budget components for the Madison and Minnelusa aquifers, water years 1931-98, for the Black Hills of South Dakota and Wyoming–Continued

[All estimates in cubic feet per second. --, not determined]

Water year	Recharge[1]			Headwater springflow[2]	Artesian springflow[3]	Well withdrawals[3]	Net ground-water outflow[3]
	Streamflow	Precipitation	Total				
1968	82.87	246.91	329.78	62.2	--	--	--
1969	74.24	215.90	290.14	50.6	--	--	--
1970	105.19	293.58	398.77	78.7	--	--	--
1971	123.68	365.41	489.09	101.4	--	--	--
1972	126.93	418.46	545.40	119.9	--	--	--
1973	123.78	283.41	407.18	69.5	--	--	--
1974	54.09	127.82	181.92	23.3	--	--	--
1975	96.06	178.43	274.49	37.5	--	--	--
1976	113.01	366.44	479.45	101.1	--	--	--
1977	86.23	269.50	355.73	69.1	--	--	--
1978	108.65	333.69	442.34	88.7	--	--	--
1979	84.96	233.26	318.22	54.8	--	--	--
1980	60.17	112.06	172.23	18.0	--	--	--
1981	60.88	170.50	231.38	32.4	--	--	--
1982	89.00	514.20	603.20	158.7	--	--	--
1983	115.39	167.59	282.97	37.7	--	--	--
1984	122.53	262.19	384.72	63.2	--	--	--
1985	49.88	68.91	118.79	8.0	--	--	--
1986	92.52	356.64	449.17	90.8	--	--	--
1987	108.41	126.33	234.73	21.1	199.6	28.3	100.4
1988	38.38	102.37	140.74	16.2	186.2	28.3	100.4
1989	40.36	146.66	187.01	26.6	163.4	28.3	100.4
1990	76.27	190.95	267.22	38.7	168.1	28.3	100.4
1991	103.11	306.66	409.77	74.9	170.6	28.3	100.4
1992	66.30	199.31	265.61	48.5	165.9	28.3	100.4
1993	128.83	444.35	573.18	127.2	173.9	28.3	100.4
1994	120.16	203.50	323.65	49.3	193.6	28.3	100.4
1995	183.57	663.81	847.38	214.6	221.5	28.3	100.4
1996	179.48	522.32	701.80	158.2	245.9	28.3	100.4
1997	221.55	545.83	767.38	192.9	--	--	--
1998	174.77	458.38	633.15	152.8	--	--	--
1931-98 average	93.18	250.90	344.08	65.5	--	--	--
1950-98 average	98.39	271.04	369.40	72.2	--	--	--
1987-96 average	104.49	290.63	395.11	77.5	188.9	28.3	100.4

[1]From Carter, Driscoll, and Hamade (2001).

[2]From Carter, Driscoll, Hamade, and Jarrell (2001). Estimates are based strictly on estimated annual recharge and are intended only for estimation of long-term averages of springflow.

[3]From Carter, Driscoll, Hamade, and Jarrell (2001).

Driscoll and Carter (2001) developed average budgets for the Madison and Minnelusa aquifer for 1950-98 for the Black Hills area of South Dakota (study area) and for the Black Hills area of South Dakota and Wyoming (expanded area) (table 14). The budget for the study area includes an estimate of headwater springflow (78 ft^3/s) that was obtained by applying estimates of precipitation recharge to the area east of the ground-water divide on the Limestone Plateau (fig. 50). “Net recharge” of 214 ft^3/s was calculated by subtracting headwater springflow from the sum of streamflow and precipitation recharge. A breakdown between wells and artesian springs (28 and 128 ft^3/s, respectively) also is provided.

Table 14 also provides a budget for 1950-98 for an expanded area that includes large outcrops areas in Wyoming (fig. 54). For the expanded area, average precipitation recharge (271 ft^3/s) is considerably larger than for the study area; however, average streamflow recharge in the expanded area (98 ft^3/s) is only slightly larger. Average headwater springflow (72 ft^3/s) is slightly smaller because measured average flows of about 6 ft^3/s for Beaver and Cold Springs Creek are excluded. Both streams originate as headwater springs in South Dakota; however, both streams are depleted by streamflow losses that provide recharge to the Minnelusa aquifer just downstream (west) of the Wyoming border. Artesian springflow (169 ft^3/s) is larger and primarily accounts for artesian springflow measured along Stockade Beaver Creek and Sand Creek (fig. 34). Average net ground-water outflow (100 ft^3/s) is larger and reflects additional recharge within the expanded area.

Table 14. Average hydrologic budgets for the Madison and Minnelusa aquifers

[From Carter, Driscoll, Hamade, and Jarrell (2001) and Driscoll and Carter (2001)]

Units	Streamflow recharge	Precipitation recharge	Headwater springflow	Net recharge	Well withdrawals	Artesian springflow	Net ground-water outflow from study area/expanded area
Black Hills of South Dakota (study area), water years 1950-98							
Acre-feet	66,600	144,900	56,500	155,000	20,300	92,800	41,900
Cubic feet per second	92	200	[1]78	214	[2]28	128	58
Black Hills of South Dakota and Wyoming (expanded area), water years 1950-98							
Acre-feet	71,000	196,300	52,200	215,100	20,300	122,400	72,400
Cubic feet per second	98	271	72	297	[2]28	169	100
Black Hills of South Dakota and Wyoming (expanded area), water years 1987-96							
Acre-feet	75,300	210,800	56,500	229,600	20,300	136,900	72,400
Cubic feet per second	104	291	[1]78	317	28	189	100

[1]Includes 6 cubic feet per second of discharge for Beaver Creek and Cold Springs Creek in South Dakota, which subsequently recharges Minnelusa aquifer a short distance downstream in Wyoming. Thus, this flow is treated as a discharge for South Dakota; however, discharge and recharge are offsetting when both South Dakota and Wyoming are considered.

[2]Identical estimate used for well withdrawals in both areas. Areas considered in Wyoming primarily are recharge areas, where well withdrawals are minor.

Precipitation recharge accounts for about 73 percent of the recharge in the expanded area. Artesian springflow is the single largest discharge component and accounts for about 46 percent of the total outflows for the expanded area. Headwater springflow accounts for about 20 percent of the total outflow for the expanded area. Thus, most of the total outflow from the Madison and Minnelusa aquifers is from springflow, which then provides flow to area streams. Ground water flowing out of the expanded area accounts for about 27 percent of the total outflows. Well withdrawals account for about 8 percent of the total outflows.

Carter, Driscoll, Hamade, and Jarrell (2001) computed an average hydrologic budget for 1987-96 for the Black Hills of South Dakota and Wyoming (expanded area) (table 14). For this period, change in storage was assumed equal to zero based on well hydrographs. Total springflow (including headwater and artesian springflow) averaged 267 ft^3/s, which constitutes about 68 percent of estimated recharge; artesian springflow of 189 ft^3/s was the single largest outflow component.

Hydrologic budgets also were quantified by Carter, Driscoll, Hamade, and Jarrell (2001) for nine subareas (figs. 69 and 70) for periods of decreasing storage (1987-92) and increasing storage (1993-96), with changes in storage assumed equal but opposite (table 15). For most subareas, net ground-water outflow exceeds inflow, and ranges from 5.9 ft^3/s in the area east of Rapid City (subarea 4) to 48.6 ft^3/s along the southwestern flanks of the Black Hills (subarea 9). Net ground-water inflow exceeds outflow for two subareas where the discharge of large artesian springs exceeds estimated recharge within subareas (subareas 7 and 8).

Detailed budgets developed for the nine subareas included estimates of transmissivity and flow components for the individual aquifers at specific flow zones (figs. 69 and 70). The net outflow and inflows from the subarea budgets were used with Darcy's Law to estimate transmissivity and flow across exterior flow zones corresponding with parts of the study area boundary and interior flow zones between subareas. For estimation purposes, it was assumed that transmissivities of the Madison and Minnelusa aquifers are equal in corresponding flow zones. The resulting transmissivity estimates ranged from about 100 to 7,400 ft^2/d, which are consistent with estimates from other previous investigations (table 2). The highest transmissivity estimates are for areas in the northern and southwestern parts of the study area, and the lowest transmissivity estimates are along the eastern study area boundary. Because the transmissivity estimates presented by Carter, Driscoll, Hamade, and Jarrell (2001) are averages over large areas, much larger spatial variability in actual transmissivities can be expected.

The large changes in storage (table 15) for subarea 2 (about 69,000 acre-ft) result primarily from a large net outflow rate and a large differential in streamflow recharge rates between the dry and wet periods. The large storage change is consistent with large water-level fluctuations for observation wells in this subarea. In comparison, changes in storage are much smaller for subareas 3 and 5, both of which have much smaller differentials in both streamflow and precipitation recharge rates. Both of these subareas are influenced by artesian springs with highly variable discharge rates relative to most other artesian springs. Water-level data for wells in subarea 5 are sparse; however, hydrographs for wells in subarea 3 show relatively small water-level fluctuations.

Budgets for subareas in the southern Black Hills are consistent with geochemical interpretations (Naus and others, 2001), which indicate long flowpaths along the western and southwestern flanks contributing to large artesian springs in the area, as discussed in a following section. The average discharge of Beaver Creek Spring in subarea 7 exceeds estimated recharge for this subarea and probably is influenced by outflow from subarea 8, which is consistent with geochemical information. Similarly, discharge of artesian springs in subarea 8 is much larger than recharge in this subarea. Outflow from subarea 9 is a probable source for inflow to subarea 8, which again is substantiated by geochemical information.

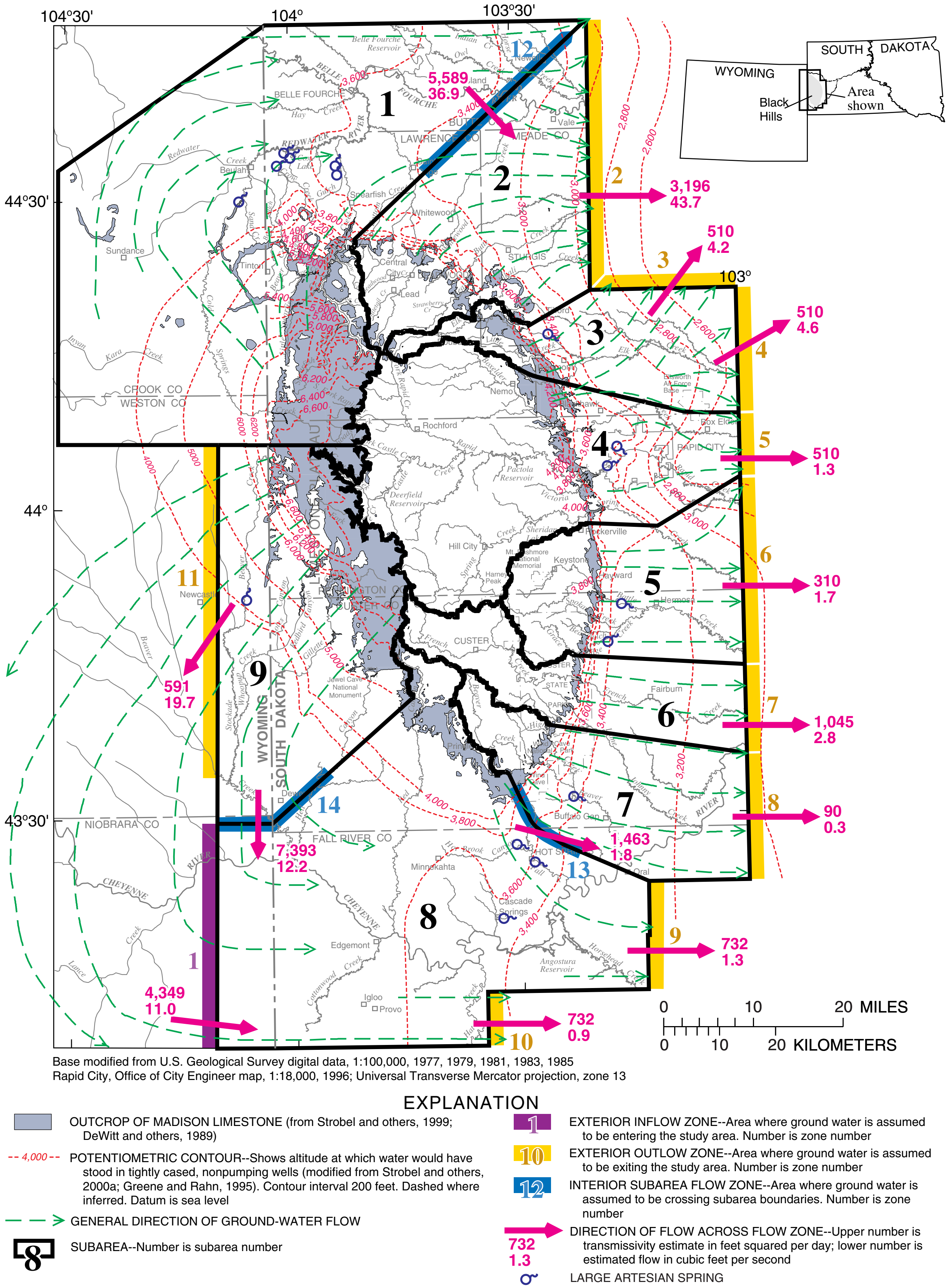
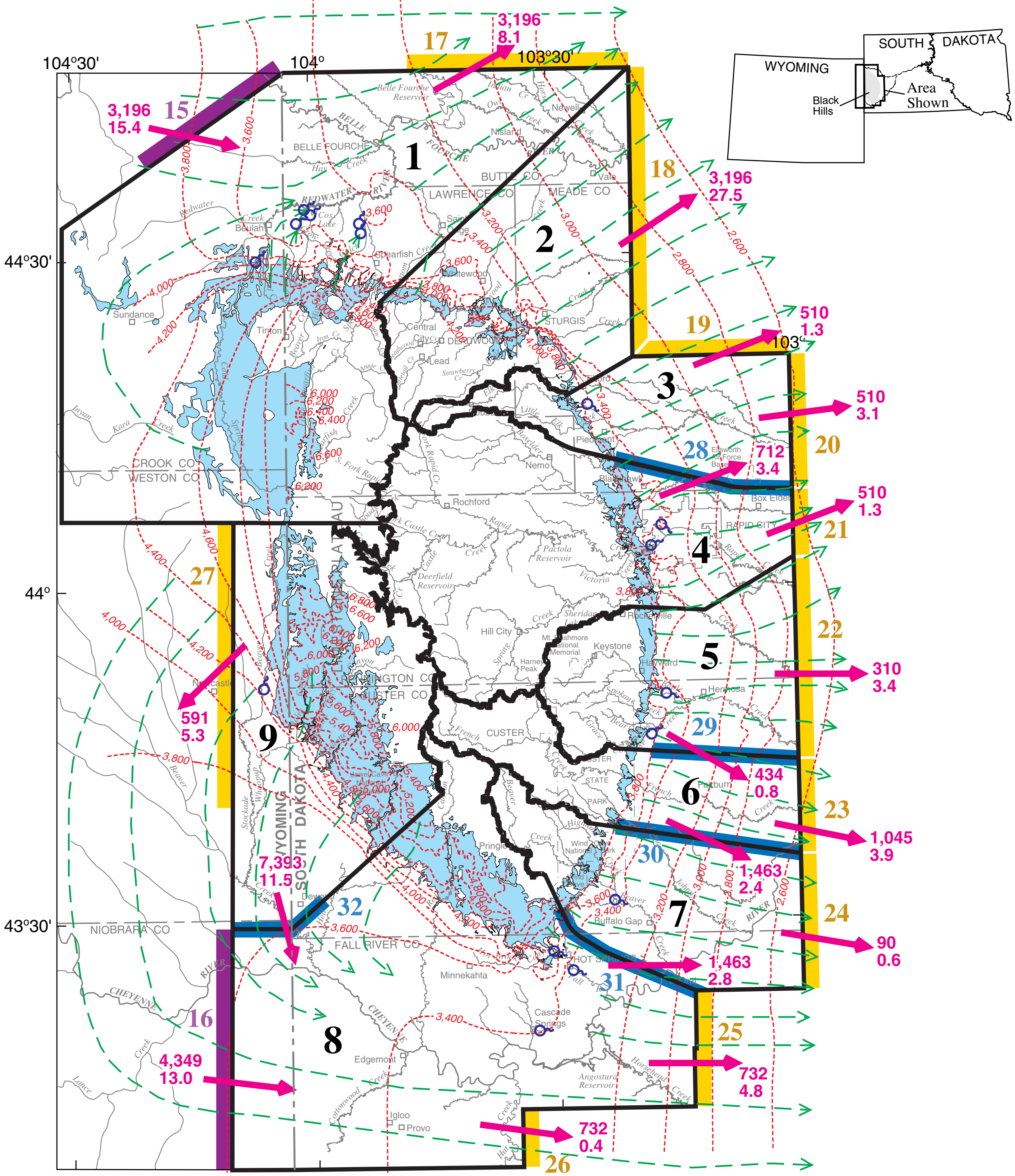


Figure 69. Subareas, generalized ground-water flow directions, and flow zones for the Madison aquifer. Estimated transmissivities and flow components for flow zones also are shown (from Carter, Driscoll, Hamade, and Jarrell, 2001).



Base modified from U.S. Geological Survey digital data, 1:100,000, 1977, 1979, 1981, 1983, 1985
Rapid City, Office of City Engineer map, 1:18,000, 1996; Universal Transverse Mercator projection, zone 13

0 10 20 MILES
0 10 20 KILOMETERS

EXPLANATION

OUTCROP OF MINNELUSA FORMATION (from Strobel and others, 1999; DeWitt and others, 1989)

4,000 POTENTIOMETRIC CONTOUR--Shows altitude at which water would have stood in tightly cased, nonpumping wells (modified from Strobel and others, 2000b; Downie and Dinwiddie, 1988). Contour interval 200 feet, where appropriate.Dashed where inferred. Datum is sea level

GENERAL DIRECTION OF GROUND-WATER FLOW

8 SUBAREA--Number is subarea number

15 EXTERIOR INFLOW ZONE--Area where ground water is assumed to be entering the study area. Number is zone number

25 EXTERIOR OUTLOW ZONE--Area where ground water is assumed to be exiting the study area. Number is zone number

30 INTERIOR FLOW ZONE--Area where ground water is assumed to be crossing subarea boundaries. Number is zone number

510 3.1 DIRECTION OF FLOW ACROSS FLOW ZONE--Upper number is transmissivity estimate in feet squared per day; lower number is estimated flow in cubic feet per second

LARGE ARTESIAN SPRING

Figure 70. Subareas, generalized ground-water flow directions, and flow zones for the Minnelusa aquifer. Estimated transmissivities and flow components for flow zones also are shown (from Carter, Driscoll, Hamade, and Jarrell, 2001).

Table 15. Hydrologic budgets, by subareas, for the Madison and Minnelusa aquifers in the Black Hills area, water years 1987-96

[From Carter, Driscoll, Hamade, and Jarrell (2001). ft³/s, cubic feet per second; acre-ft, acre-feet]

Water year	Inflows (ft³/s)			Outflows (ft³/s)			Sum (ft³/s)		Change in storage	
	Stream-flow recharge	Precipita-tion recharge	Net ground-water inflow	Artesian spring-flow	Well with-drawals	Net ground-water outflow	Inflows	Outflows	ft³/s	acre-ft
					Subarea 1					
1987	8.6	64.8	0.0	101.8	11.0	29.6	73.4	142.4	-69.0	-49,961
1988	5.9	53.9	.0	94.8	12.6	29.6	59.8	137.0	-77.2	-55,898
1989	6.1	65.5	.0	75.5	11.7	29.6	71.6	116.8	-45.2	-32,728
1990	7.5	83.9	.0	78.8	12.2	29.6	91.4	120.6	-29.2	-21,143
1991	8.5	122.7	.0	81.4	10.9	29.6	131.2	121.9	9.3	6,734
1992	6.6	86.5	.0	76.0	11.3	29.6	93.1	116.9	-23.8	-17,233
Average/sum[1] 1987-92	7.2	79.6	.0	84.7	11.6	29.6	86.8	125.9	-39.2	[1]-170,228
1993	9.8	170.2	.0	80.6	9.1	29.6	180.0	119.3	60.7	43,951
1994	11.3	100.7	.0	93.6	10.6	29.6	112.0	133.8	-21.8	-15,785
1995	15.9	260.0	.0	103.9	9.4	29.6	275.9	142.9	133.0	96,301
1996	16.1	202.9	.0	116.7	9.6	29.6	219.0	155.9	63.1	45,689
Average/sum[1] 1993-96	13.3	183.5	.0	98.7	9.7	29.6	196.7	138.0	58.8	[1]170,156
Average/sum[1] 1987-96	9.6	121.1	.0	90.3	10.8	29.6	130.7	130.8	.0	[1]-72
					Subarea 2					
1987	27.5	6.1	.0	.0	3.7	34.2	33.6	37.9	-4.3	-3,113
1988	5.8	5.9	.0	.0	4.2	34.2	11.7	38.4	-26.7	-19,333
1989	9.2	6.2	.0	.0	3.8	34.2	15.4	38.0	-22.6	-16,364
1990	15.5	5.8	.0	.0	4.1	34.2	21.3	38.3	-17.0	-12,309
1991	23.8	8.9	.0	.0	3.7	34.2	32.7	37.9	-5.2	-3,765
1992	12.3	5.7	.0	.0	3.8	34.2	18.0	38.0	-20.0	-14,481
Average/sum[1] 1987-92	15.7	6.4	.0	.0	3.9	34.2	22.1	38.1	-16.0	[1]-69,366
1993	32.9	16.0	.0	.0	3.1	34.2	48.9	37.3	11.6	8,399
1994	35.4	5.9	.0	.0	3.6	34.2	41.3	37.8	3.5	2,534
1995	58.5	31.7	.0	.0	3.2	34.2	90.2	37.4	52.8	38,231
1996	49.7	16.0	.0	.0	3.3	34.2	65.7	37.5	28.2	20,419
Average/sum[1] 1993-96	44.1	17.4	.0	.0	3.3	34.2	61.5	37.5	24.0	[1]69,583
Average/sum[1] 1987-96	27.1	10.8	.0	.0	3.7	34.2	37.9	37.9	.0	[1]217

Table 15. Hydrologic budgets, by subareas, for the Madison and Minnelusa aquifers in the Black Hills area, water years 1987-96–Continued

[From Carter, Driscoll, Hamade, and Jarrell (2001). ft^3/s, cubic feet per second; acre-ft, acre-feet]

Water year	Inflows (ft^3/s)			Outflows (ft^3/s)			Sum (ft^3/s)		Change in storage	
	Stream-flow recharge	Precipita-tion recharge	Net ground-water inflow	Artesian spring-flow	Well with-drawals	Net ground-water outflow	Inflows	Outflows	ft^3/s	acre-ft
Subarea 3										
1987	10.6	1.6	0.0	1.6	0.6	9.8	12.2	12.0	0.2	145
1988	2.5	1.3	.0	.3	.6	9.8	3.8	10.7	-6.9	-4,996
1989	2.7	2.1	.0	.0	.6	9.8	4.8	10.4	-5.6	-4,055
1990	8.6	2.1	.0	.0	.6	9.8	10.7	10.4	.3	217
1991	9.1	4.6	.0	.1	.6	9.8	13.7	10.5	3.2	2,317
1992	5.6	1.8	.0	.0	.6	9.8	7.4	10.4	-3.0	-2,172
Average/sum[1] 1987-92	6.5	2.3	.0	.3	.6	9.8	8.8	10.7	-2.0	[1]-8,544
1993	10.0	5.3	.0	.1	.6	9.8	15.3	10.5	4.8	3,476
1994	10.9	1.7	.0	.4	.6	9.8	12.6	10.8	1.8	1,303
1995	12.0	10.3	.0	8.6	.6	9.8	22.3	19.0	3.3	2,389
1996	13.9	6.3	.0	8.2	.6	9.8	20.2	18.6	1.6	1,159
Average/sum[1] 1993-96	11.7	5.9	.0	4.3	.6	9.8	17.6	14.7	2.9	[1]8,327
Average/sum[1] 1987-96	8.6	3.7	.0	1.9	.6	9.8	12.3	12.3	.0	[1]-217
Subarea 4										
1987	33.3	2.7	.0	25.6	4.3	5.9	36.0	35.8	.2	145
1988	17.3	1.0	.0	25.6	4.2	5.9	18.3	35.7	-17.4	-12,599
1989	15.6	3.9	.0	25.6	3.4	5.9	19.5	34.9	-15.4	-11,151
1990	24.1	3.7	.0	25.6	5.0	5.9	27.8	36.5	-8.7	-6,299
1991	33.7	10.3	.0	25.6	6.7	5.9	44.0	38.2	5.8	4,200
1992	25.9	3.2	.0	25.6	10.5	5.9	29.1	42.0	-12.9	-9,340
Average/sum[1] 1987-92	25.0	4.1	.0	25.6	5.7	5.9	29.1	37.2	-8.1	[1]-35,045
1993	43.3	8.4	.0	25.8	10.2	5.9	51.7	41.9	9.8	7,096
1994	40.3	1.4	.0	25.6	11.1	5.9	41.7	42.6	-0.9	-652
1995	47.5	12.3	.0	27.1	9.1	5.9	59.8	42.1	17.7	12,816
1996	55.8	9.6	.0	30.1	7.6	5.9	65.4	43.6	21.8	15,785
Average/sum[1] 1993-96	46.7	7.9	.0	27.2	9.5	5.9	54.7	42.6	12.1	[1]35,045
Average/sum[1] 1987-96	33.7	5.7	.0	26.2	7.2	5.9	39.3	39.3	.0	[1]0

Table 15. Hydrologic budgets, by subareas, for the Madison and Minnelusa aquifers in the Black Hills area, water years 1987-96–Continued

[From Carter, Driscoll, Hamade, and Jarrell (2001). ft^3/s, cubic feet per second; acre-ft, acre-feet]

Water year	Inflows (ft^3/s)			Outflows (ft^3/s)			Sum (ft^3/s)		Change in storage	
	Stream-flow recharge	Precipita-tion recharge	Net ground-water inflow	Artesian spring-flow	Well with-drawals	Net ground-water outflow	Inflows	Outflows	ft^3/s	acre-ft
				Subarea 5						
1987	12.8	1.3	0.0	6.2	1.6	6.0	14.1	13.8	0.3	217
1988	2.4	.6	.0	2.4	1.6	6.0	3.0	10.0	-7.0	-5,068
1989	3.1	2.2	.0	1.2	1.6	6.0	5.3	8.8	-3.5	-2,534
1990	10.6	3.6	.0	1.6	1.6	6.0	14.2	9.2	5.0	3,620
1991	12.9	4.5	.0	3.5	1.6	6.0	17.4	11.1	6.3	4,562
1992	7.9	2.0	.0	4.5	1.6	6.0	9.9	12.1	-2.2	-1,593
Average/sum[1] 1987-92	8.3	2.4	.0	3.2	1.6	6.0	10.7	10.8	-0.2	[1]-796
1993	17.3	4.9	.0	7.1	1.6	6.0	22.2	14.7	7.5	5,430
1994	10.2	1.3	.0	10.8	1.7	6.0	11.5	18.5	-7.0	-5,068
1995	18.7	7.2	.0	11.3	1.6	6.0	25.9	18.9	7.0	5,068
1996	18.0	4.7	.0	21.0	1.7	6.0	22.7	28.7	-6.0	-4,344
Average/sum[1] 1993-96	16.1	4.5	.0	12.6	1.7	6.0	20.6	20.2	.4	[1]1,086
Average/sum[1] 1987-96	11.4	3.2	.0	7.0	1.6	6.0	14.6	14.6	.0	[1]290
				Subarea 6						
1987	7.8	.3	.0	.0	.0	8.3	8.1	8.3	-0.2	-145
1988	2.7	.2	.0	.0	.0	8.3	2.9	8.3	-5.4	-3,910
1989	1.6	.6	.0	.0	.0	8.3	2.2	8.3	-6.1	-4,417
1990	5.1	.8	.0	.0	.0	8.3	5.9	8.3	-2.4	-1,738
1991	7.8	.9	.0	.0	.0	8.3	8.7	8.3	.4	290
1992	5.5	.7	.0	.0	.0	8.3	6.2	8.3	-2.1	-1,521
Average/sum[1] 1987-92	5.1	.6	.0	.0	.0	8.3	5.7	8.3	-2.6	[1]-11,440
1993	9.2	1.3	.0	.0	.0	8.3	10.5	8.3	2.2	1,593
1994	7.4	.3	.0	.0	.0	8.3	7.7	8.3	-0.6	-434
1995	13.0	2.5	.0	.0	.0	8.3	15.5	8.3	7.2	5,213
1996	13.9	1.5	.0	.0	.0	8.3	15.4	8.3	7.1	5,141
Average/sum[1] 1993-96	10.9	1.4	.0	.0	.0	8.3	12.3	8.3	4.0	[1]11,513
Average/sum[1] 1987-96	7.4	.9	.0	.0	.0	8.3	8.3	8.3	.0	[1]72

Table 15. Hydrologic budgets, by subareas, for the Madison and Minnelusa aquifers in the Black Hills area, water years 1987-96–Continued

[From Carter, Driscoll, Hamade, and Jarrell (2001). ft³/s, cubic feet per second; acre-ft, acre-feet]

Water year	Inflows (ft³/s)			Outflows (ft³/s)			Sum (ft³/s)		Change in storage	
	Stream-flow recharge	Precipita-tion recharge	Net ground-water inflow	Artesian spring-flow	Well with-drawals	Net ground-water outflow	Inflows	Outflows	ft³/s	acre-ft
					Subarea 7					
1987	2.0	0.6	6.1	10.0	0.1	0.0	8.7	10.1	-1.4	-1,014
1988	.4	.4	6.1	10.0	.1	.0	6.9	10.1	-3.2	-2,317
1989	.4	.6	6.1	9.0	.1	.0	7.1	9.1	-2.0	-1,448
1990	1.0	.9	6.1	9.0	.1	.0	8.0	9.1	-1.1	-796
1991	1.6	1.1	6.1	8.1	.1	.0	8.8	8.2	.6	434
1992	1.0	1.2	6.1	8.0	.1	.0	8.3	8.1	.2	145
Average/sum[1] 1987-92	1.1	.8	6.1	9.0	.1	.0	8.0	9.1	-1.2	[1]-4,996
1993	2.2	2.1	6.1	9.1	.1	.0	10.4	9.2	1.2	869
1994	2.3	.6	6.1	9.5	.1	.0	9.0	9.6	-0.6	-434
1995	6.1	4.2	6.1	12.5	.1	.0	16.4	12.6	3.8	2,751
1996	6.1	1.2	6.1	11.1	.1	.0	13.4	11.2	2.2	1,593
Average/sum[1] 1993-96	4.2	2.0	6.1	10.6	.1	.0	12.3	10.7	1.7	[1]4,779
Average/sum[1] 1987-96	2.3	1.3	6.1	9.6	.1	.0	9.7	9.7	.0	[1]-217
					Subarea 8					
1987	5.9	3.2	35.6	45.4	1.8	.0	44.7	47.2	-2.5	-1,810
1988	1.5	2.2	35.6	44.1	1.8	.0	39.3	45.9	-6.6	-4,779
1989	1.6	3.8	35.6	43.1	1.8	.0	41.0	44.9	-3.9	-2,824
1990	3.9	5.4	35.6	44.1	1.9	.0	44.9	46.0	-1.1	-796
1991	5.5	5.7	35.6	43.1	1.8	.0	46.8	44.9	1.9	1,376
1992	1.6	5.2	35.6	42.9	1.8	.0	42.4	44.7	-2.3	-1,665
Average/sum[1] 1987-92	3.3	4.3	35.6	43.8	1.8	.0	43.2	45.6	-2.4	[1]-10,499
1993	4.1	10.8	35.6	42.1	1.8	.0	50.5	43.9	6.6	4,779
1994	2.4	2.3	35.6	44.0	1.9	.0	40.3	45.9	-5.6	-4,055
1995	11.9	14.6	35.6	46.3	1.9	.0	62.1	48.2	13.9	10,065
1996	6.0	7.5	35.6	47.5	1.8	.0	49.1	49.3	-0.2	-145
Average/sum[1] 1993-96	6.1	8.8	35.6	45.0	1.9	.0	50.5	46.8	3.7	[1]10,644
Average/sum[1] 1987-96	4.4	6.1	35.6	44.3	1.8	.0	46.1	46.1	.0	[1]145

Table 15. Hydrologic budgets, by subareas, for the Madison and Minnelusa aquifers in the Black Hills area, water years 1987-96–Continued

[From Carter, Driscoll, Hamade, and Jarrell (2001). ft^3/s, cubic feet per second; acre-ft, acre-feet]

Water year	Inflows (ft^3/s)			Outflows (ft^3/s)			Sum (ft^3/s)		Change in storage	
	Stream-flow recharge	Precipita-tion recharge	Net ground-water inflow	Artesian spring-flow	Well with-drawals	Net ground-water outflow	Inflows	Outflows	ft^3/s	acre-ft
				Subarea 9						
1987	0.0	24.6	0.0	9.0	2.2	48.6	24.6	59.8	-35.2	-25,487
1988	.0	20.7	.0	9.0	2.2	48.6	20.7	59.8	-39.1	-28,311
1989	.0	35.3	.0	9.0	2.2	48.6	35.3	59.8	-24.5	-17,740
1990	.0	46.1	.0	9.0	2.2	48.6	46.1	59.8	-13.7	-9,920
1991	.0	73.1	.0	8.8	2.2	48.6	73.1	59.6	13.5	9,775
1992	.0	44.6	.0	8.9	2.2	48.6	44.6	59.7	-15.1	-10,933
Average/sum[1] 1987-92	.0	40.7	.0	9.0	2.2	48.6	40.7	59.8	-19.0	[1]-82,616
1993	.0	98.2	.0	9.1	2.2	48.6	98.2	59.9	38.3	27,732
1994	.0	40.2	.0	9.7	2.3	48.6	40.2	60.6	-20.4	-14,771
1995	.0	106.4	.0	11.8	2.2	48.6	106.4	62.6	43.8	31,714
1996	.0	114.3	.0	11.3	2.3	48.6	114.3	62.2	52.1	37,724
Average/sum[1] 1993-96	.0	89.8	.0	10.5	2.3	48.6	89.8	61.3	28.5	[1]82,399
Average/sum[1] 1987-96	.0	60.4	.0	9.6	2.2	48.6	60.4	60.4	.0	[1]-217

[1]Sum used for change in storage in acre-feet.

Budgets for Other Bedrock Aquifers

Recharge estimates for the other bedrock aquifers consist only of precipitation recharge, which was derived by Driscoll and Carter (2001) using the yield-efficiency algorithm. Total yield, which is the sum of runoff plus recharge, was computed by applying the yield-efficiency algorithm to the estimates of precipitation on outcrops of the various bedrock formations.

Individual ground-water budgets and an overall budget for all bedrock aquifers in the study area are presented in table 16. The Madison and Minnelusa aquifers, which have the largest outcrop areas of the major aquifers in the study area, dominate the overall ground-water budgets. In contrast, runoff from aquifer outcrops is dominated by the crystalline core area, with negligible runoff from outcrops of the Madison, Minnelusa, or Minnekahta aquifers. Combined recharge for all bedrock aquifers was estimated as 348 ft^3/s, of which about 84 percent is recharge to the Madison and Minnelusa aquifers. Total well withdrawals and springflow account for 259 ft^3/s, of which about 90 percent is from the Madison and Minnelusa aquifers. Net ground-water outflow from the study area was calculated as 89 ft^3/s and ranges from zero (assumed) for the crystalline core aquifers to about 58 ft^3/s (65 percent) for the Madison and Minnelusa aquifers. For the Deadwood aquifer, well withdrawals and springflow were estimated as 14 ft^3/s, which consists primarily of spring discharge in headwater areas of about 13 ft^3/s. For the Madison and Minnelusa aquifers, springflow of 206 ft^3/s is much larger than well withdrawals of 28 ft^3/s. For all other aquifers, springflow is small and was neglected; only well withdrawals are listed in table 16.

Table 16. Average ground-water budgets for bedrock aquifers, water years 1950-98

[From Driscoll and Carter (2001). --, no data]

Units	Precipitation	Evapotran-spiration	Total yield	Runoff	Precipitation recharge	Streamflow recharge	Total recharge	Well with-drawals and springflow[1]	Net study area outflow
Crystalline Core (Precambrian, Tertiary, and Other Minor Units)									
Acre-feet per year	1,084,500	964,200	120,300	116,700	3,600	0	3,600	3,600	0.0
Cubic feet per second	1,497	1,331	166	161	5	0	5	5	0
Inches per year	21.10	18.76	2.34	2.27	0.07	0	0.07	--	--
Deadwood									
Acre-feet per year	128,200	110,100	18,100	3,600	14,500	0	14,500	10,100	4,400
Cubic feet per second	177	152	25	5	20	0	20	[1]14	6
Inches per year	23.24	19.96	3.28	0.65	2.63	0	2.63	--	--
Madison and Minnelusa									
Acre-feet per year	1,021,500	876,600	144,900	0	144,900	66,600	211,500	169,500	41,900
Cubic feet per second	1,410	1,210	200	0	200	92	292	[1]234	58
Inches per year	20.69	17.76	2.93	0	2.93	1.35	4.28	--	--
Minnekahta									
Acre-feet per year	120,300	113,800	6,500	0	6,500	0	6,500	700	5,800
Cubic feet per second	166	157	9	0	9	0	9	1	8
Inches per year	20.02	18.94	1.08	0	1.08	0	1.08	--	--
Inyan Kara									
Acre-feet per year	326,700	312,200	14,500	2,900	11,600	0	11,600	1,400	10,200
Cubic feet per second	451	431	20	4	16	0	16	2	14
Inches per year	17.84	17.05	0.79	0.16	0.63	0	0.63	--	--
Jurassic-Sequence Semiconfining Unit									
Acre-feet per year	115,900	110,000	5,800	3,600	2,200	0	2,200	700	1,500
Cubic feet per second	160	152	8	5	3	0	3	1	2
Inches per year	18.35	17.43	0.92	0.57	0.35	0	0.35	--	--
Cretaceous-Sequence Confining Unit									
Acre-feet per year	1,028,700	980,900	47,800	45,600	2,200	0	2,200	1,400	800
Cubic feet per second	1,420	1,354	66	63	3	0	3	2	1
Inches per year	17.24	16.44	0.80	0.76	0.04	0	0.04	--	--
Overall Budget for Bedrock Aquifers									
Acre-feet per year	3,825,900	3,468,000	357,900	172,400	185,500	66,600	252,100	187,600	64,600
Cubic feet per second	5,281	4,787	494	238	256	92	348	259	89
Inches per year	19.46	17.64	1.82	0.88	0.94	1.35	1.28	--	--

[1]Includes estimated springflow of 13 cubic feet per second for Deadwood aquifer and 206 cubic feet per second for Madison and Minnelusa aquifers. For other aquifers, springflow is considered negligible and estimates include only well withdrawals.

Surface-Water Budgets

Surface-water budgets were developed by Driscoll and Carter (2001) by consideration of stream channels within various specified areas, for which the basic continuity equation (eq. 2) was applied. Inflows considered included stream channels crossing boundaries for specified areas and net tributary flows generated within specified areas. Because net tributary flows (flows less depletions) were considered, flow depletions such as streamflow losses or diversions were not included as outflows. Storage changes for the four large Bureau of Reclamation reservoirs (Angostura, Deerfield, Pactola, and Belle Fourche) located within the study area were considered, with records of storage changes (positive change reflects increased storage) derived primarily from Miller and Driscoll (1998). Large storage increases occurred during 1950-98 for Angostura Reservoir (completed during 1950), Pactola Reservoir (completed during 1956), and Belle Fourche Reservoir, which had very low storage during 1950.

Average surface-water budgets for 1950-98 are provided in table 17. Inflows to the study area averaged about 106 and 146 ft^3/s in the Cheyenne and Belle Fourche River drainages, respectively, with combined inflows averaging about 252 ft^3/s. Net tributary flows generated within the study area were estimated as about 201 and 107 ft^3/s in the Cheyenne and Belle Fourche River drainages, respectively, with combined tributary flows of about 308 ft^3/s. Considering storage increases of about 7 ft^3/s in Bureau of Reclamation reservoirs, total outflows from the study area were estimated as about 554 ft^3/s, with outflows of about 303 ft^3/s for the Cheyenne River drainage and 251 ft^3/s for the Belle Fourche River drainage.

The primary surface-water outflows occur along the eastern side of the study area, of which the Cheyenne and Belle Fourche Rivers constitute the largest percentage of outflows (Driscoll and Carter, 2001). Estimated outflows include combined flows of about 6 ft^3/s in the Beaver Creek and Cold Springs Creek drainages, which drain a portion of the Limestone Plateau area, with flow to the west into Wyoming. Flow in both of these streams is lost to the Minnelusa Formation a short distance downstream of the Wyoming border.

Combined Ground-Water and Surface-Water Budgets

Combined average ground- and surface-water budgets are presented in figure 71A, which includes a detailed budget that shows complex ground- and surface-water interactions, and in figure 71B, which shows a more simplified budget. These budgets also are used to show consumptive uses of water that occur within the study area. Total consumptive use within the study area from both ground-water and surface-water resources was estimated by Driscoll and Carter (2001) as 218 ft^3/s, which includes well withdrawals (40 ft^3/s), reservoir evaporation (38 ft^3/s), and consumptive withdrawals from streams (140 ft^3/s). Consumptive uses consist primarily of consumptive irrigation withdrawals, which do not include unconsumed irrigation return flows. Most well withdrawals are consumed; however, in some locations (such as Rapid City), some portion of municipal well withdrawals may be unconsumed and returned to streams via wastewater treatment effluent.

Table 17. Average surface-water budgets for study area, water years 1950-98

[From Driscoll and Carter (2001). All in cubic feet per second]

Basin	Study area inflows	+ Study area tributaries	- Change in storage	= Study area outflows
Cheyenne River	105.8	201.2	4.5	302.5
Belle Fourche River	146.4	107.3	2.7	251.0
Combined	252.2	308.5	7.2	553.5

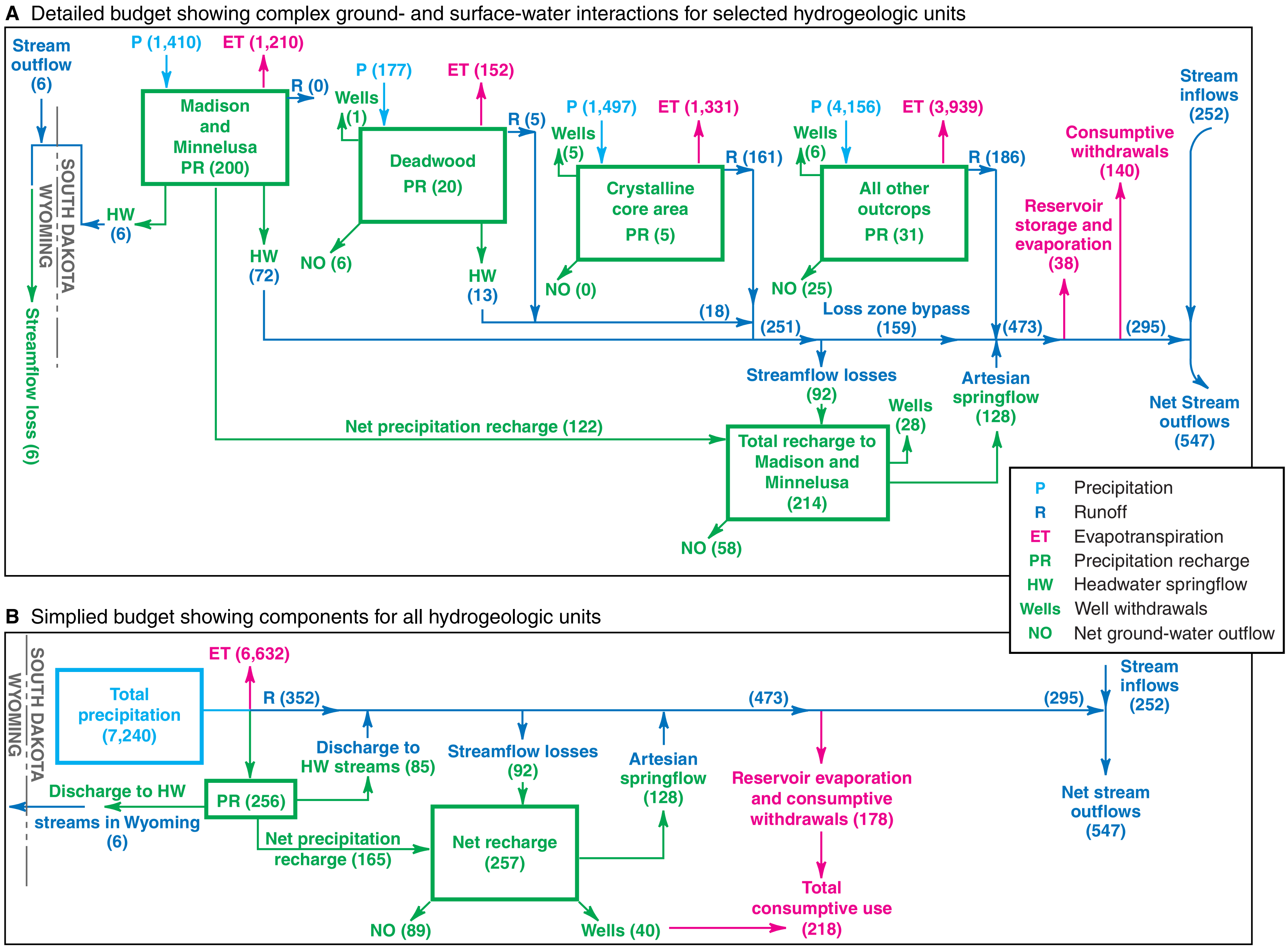


Figure 71. Schematic diagram showing average hydrologic budget components for study area, water years 1950-98 (from Driscoll and Carter, 2001). All values in cubic feet per second.

A schematic diagram is presented as figure 72 that shows the progression of average streamflow relative to surface geology and streamflow depletions. Streamflow upstream from loss zones to the Madison and Minnelusa aquifers averages about 251 ft^3/s, which consists of headwater springflow from the Madison and Minnelusa aquifers (72 ft^3/s) and Deadwood aquifer (13 ft^3/s) in the Limestone Plateau area, and of runoff from the Deadwood Formation (5 ft^3/s) and crystalline core area (161 ft^3/s). Streamflow losses to the Madison and Minnelusa aquifers average 92 ft^3/s; thus, combined streamflow downstream from loss zones (loss-zone bypass) averages about 159 ft^3/s.

Artesian springflow (128 ft^3/s) and runoff from outcrops beyond the Madison Limestone and Minnelusa Formation (186 ft^3/s) provide additional streamflow beyond the loss zones (fig. 72). Thus, average streamflow prior to major depletions, which result primarily from irrigation operations, is about 473 ft^3/s. Reservoir evaporation and consumptive withdrawals of 178 ft^3/s reduce average tributary flows to the Cheyenne and Belle Fourche Rivers to 295 ft^3/s. The tributary flows of 295 ft^3/s in figure 72 differ from the study area tributary flows of 308 ft^3/s in table 12 by the reservoir storage change (7 ft^3/s) and by combined flows for Beaver and Cold Springs Creeks (6 ft^3/s). The flows of Beaver and Cold Springs Creeks do not contribute to flows of the Cheyenne and Belle Fourche Rivers because streamflow losses occur a short distance downstream from the Wyoming border, as previously discussed.

Evaluation of Hydrologic Budgets

Numerous assumptions and estimates have been made in developing budgets for the complex hydrology of the Black Hills area. This section summarizes evaluations of budget components by Driscoll and Carter (2001), who concluded that methods used have provided reasonable budget estimates.

Hydrologic budgets for the Madison and Minnelusa aquifers are especially important because these aquifers dominate the overall ground-water budgets for the Black Hills area and heavily influence the surface-water budget. Estimates of streamflow recharge, which are based largely on measured values, are considered

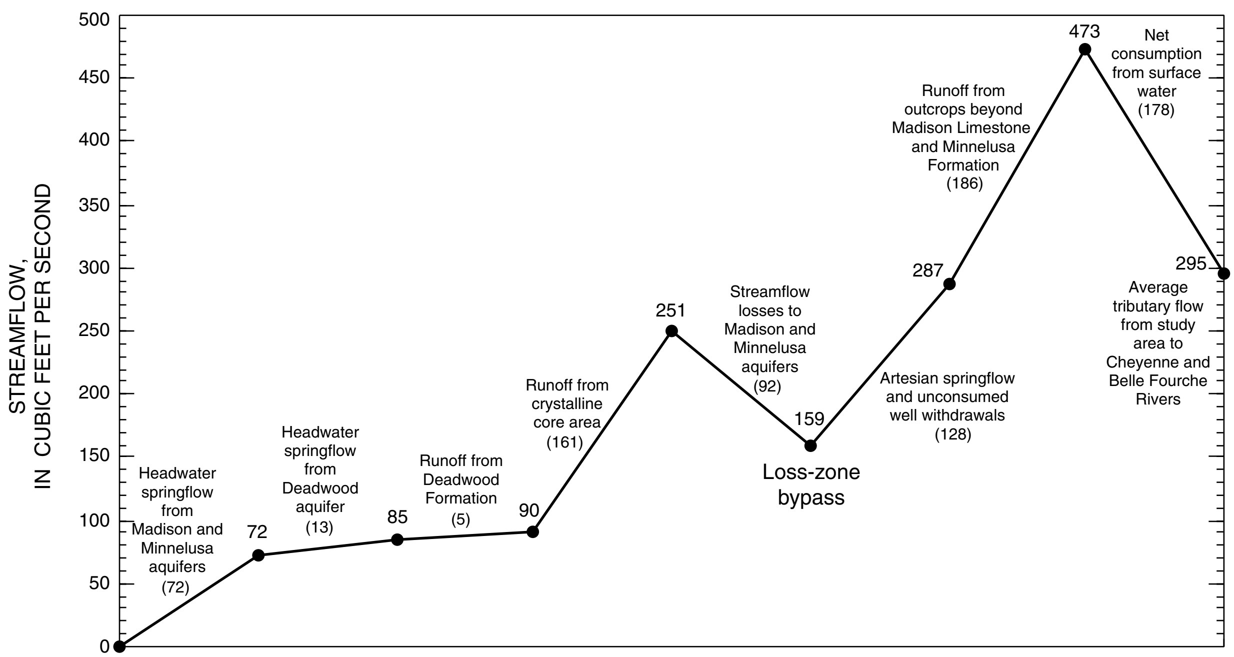


Figure 72. Schematic showing generalized average streamflow (water years 1950-98) relative to surface geology and depletions (from Discroll and Carter, 2001).

more accurate than estimates of precipitation recharge, which have relatively large uncertainty associated with use of the yield-efficiency algorithm (and the assumptions on which the procedure is based). Short-term estimates for artesian springflow, which are based primarily on measured values, have relatively small uncertainty; however, extrapolation for longer term budgets (1950-98) introduces additional uncertainty. Uncertainties are larger for estimates of headwater springflow, which are based on yield potential for inferred areas contributing to ground-water discharge. Uncertainties are small for well withdrawals; thus, most of the uncertainties for estimates of net ground-water outflow from the study area are related to uncertainties for estimates of precipitation recharge. Detailed water-budget analyses for subareas (figs. 69 and 70) provided confidence that estimates for all water-budget components are reasonable.

Budgets for other aquifers are based primarily on estimates of precipitation recharge, which were derived using the yield-efficiency algorithm. The assumed “recharge factors” used to apportion overall yield potential between runoff and recharge are another source of potential error. Considerable evidence exists that direct runoff is uncommon from outcrops of the Madison and Minnelusa aquifers; however, information regarding other aquifer outcrops is sparse.

The yield-efficiency algorithm also was used extensively in developing surface-water budgets and in estimating consumptive withdrawals for the study area. An analysis of streamflow depletion from streamflow losses indicated that estimates of total basin yield from the crystalline core area were reasonable. Evaluations of consumptive withdrawal estimates indicated that the yield-efficiency algorithm also performed well for areas beyond the Madison/Minnelusa outcrop band. Systematic biases in yield estimates undoubtedly would occur for various localized applications; however, evaluations performed have provided confidence that the algorithm systematically produces reasonable and reproducible estimates of basin yield from the spatial distribution of annual precipitation. Readers are cautioned that because of inherent, unexplained variability between annual yield and precipitation, estimates for individual years that are based on this algorithm have a relatively high level of uncertainty. Uncertainties associated with long-term estimates are much smaller, however.

MADISON AND MINNELUSA FLOW SYSTEM

A major focus of the Black Hills Hydrology Study has been to obtain a better understanding of flow systems within the Madison and Minnelusa aquifers, which are extremely complex due to heterogeneity and anisotropy related to karst features and fractures and to interactions between the aquifers and surface-water resources. A variety of information has been considered in evaluating flowpaths, mixing conditions, and interactions. Much of the relevant background information, such as potentiometric-surface maps, hydrographs for colocated Madison and Minnelusa wells, major-ion chemistry, and hydrologic budgets, has been presented in previous sections. Background information regarding isotopes is presented in the following section.

Isotope Information

Various isotopes were used by Naus and others (2001) in evaluating the Madison and Minnelusa flow systems. The stable isotopes of oxygen (^{18}O and ^{16}O) and hydrogen (^{2}H, deuterium; and ^{1}H) were used to evaluate ground-water flowpaths, recharge areas, and mixing conditions. The radioisotope tritium (^{3}H) provided additional information for evaluation of mixing conditions and ground-water ages. Background information for these isotopes and their distributions in the Black Hills area are discussed in the following sections.

Background Information and Composition of Recharge Water

This section presents background information regarding stable isotopes and tritium, which generally are not affected by interactions between minerals and ground water. The stable isotopes of oxygen and hydrogen are useful as flowpath tracers because of distinctive patterns in the Black Hills area resulting from meteorological influences. Tritium, which is subject to decay over time, is useful for age dating because large temporal variations of concentrations in precipitation have resulted from atmospheric testing of thermonuclear bombs during the 1950’s and 1960’s.

Stable isotope values are given in “delta notation,” which compares the ratio between heavy and light isotopes of a sample to that of a reference standard. Delta values are expressed as a difference, in parts per thousand, or per mil (‰), from the reference standard. For example, the oxygen isotope value of a sample written in delta notation is:

$$\delta^{18}O_{sample} = \frac{{}^{18}O/{}^{16}O_{sample} - {}^{18}O/{}^{16}O_{standard}}{{}^{18}O/{}^{16}O_{standard}}$$

$$\times 1,000\text{ ‰ VSMOW}.$$

A sample with a δ value of -20 ‰ is depleted by 20 parts per thousand (2 percent) in the heavier isotope of the element relative to the standard. In this report, $\delta^{18}O$ ($^{18}O/^{16}O$) and δD (deuterium/hydrogen) values are reported in per mil relative to Vienna Standard Mean Ocean Water (VSMOW) and are described as lighter and heavier in relation to each other. The lighter values are more negative relative to the heavier values, which are less negative.

Distinct isotopic signatures can result from isotope fractionation, which results from the loss of water vapor from a cooling air mass as it passes from its oceanic source over continents. As air masses rise to higher altitudes, lower temperatures and the subsequent formation of precipitation cause fractionation to occur within the cloud, and ^{18}O and deuterium (D) are partitioned preferentially into the rain or snow. The heavy isotopes thus are distilled from the vapor, which is progressively depleted in ^{18}O and D (Clark and Fritz, 1997). A linear relation exists between $\delta^{18}O$ and δD for samples collected in the Black Hills area (fig. 73); thus, subsequent discussions and illustrations refer only to $\delta^{18}O$ for simplicity.

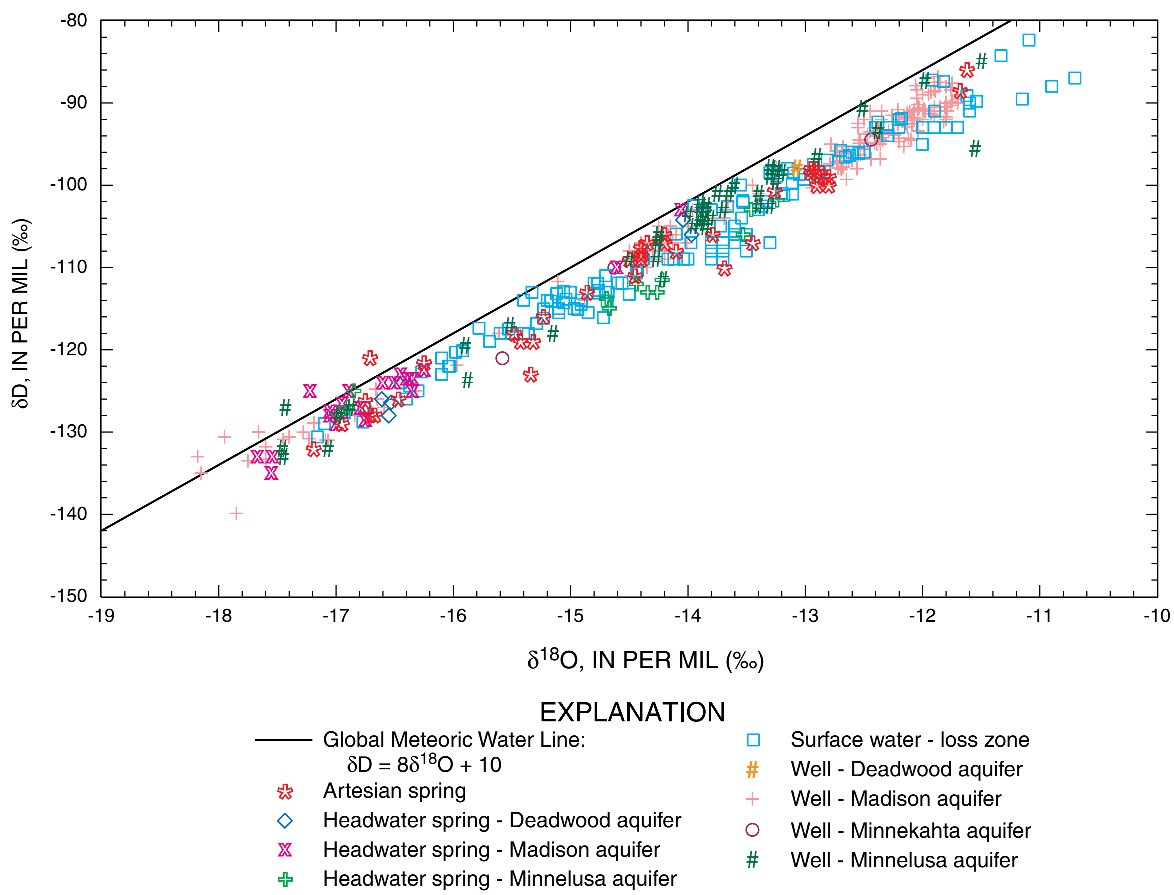


Figure 73. Relation between $\delta^{18}O$ and δD in Black Hills samples in comparison to the Global Meteoric Water Line (Craig, 1961) (from Naus and others, 2001).

A generalized distribution of $\delta^{18}O$ values for near-recharge areas is shown in figure 74. Sampling sites and values from which contours were derived also are shown in figure 74. Sites considered generally are located in or near recharge areas and probably are affected primarily by localized flowpaths.

In near-recharge areas, $\delta^{18}O$ values are influenced primarily by orography and storm patterns. Precipitation in the northern Black Hills generally is isotopically lighter than in the south because of relatively high altitudes and the influence of Pacific storms that are isotopically depleted in crossing the Rocky Mountains. The generally lower altitudes in the southern Black Hills, combined with warm weather patterns from the south-southeast, result in precipitation that is isotopically heavier than in the north (Back and others, 1983; Busby and others, 1983; Greene, 1997). The resulting distribution of isotopes in near-recharge areas of the Black Hills serves as a natural tracer for ground-water flowpaths.

Temporal variability in $\delta^{18}O$ values for selected surface-water and ground-water sites is shown in figure 75; locations for these sites were provided by Naus and others (2001, p. 42-44). Temporal variability for ground-water samples generally is small relative to surface-water samples (fig. 75). The temporal variability in $\delta^{18}O$ values for selected loss-zone streams (Spring Creek, Rapid Creek, and Boxelder Creek) (figure 75A) is due to seasonal variability in isotopic composition of precipitation. Data sets for selected headwater springs (Rhoads Fork and Castle Creek) are somewhat limited, but indicate less variability because of mixing associated with ground-water storage. Thus, for the wells and headwater springs shown in figure 74, variability in $\delta^{18}O$ was assumed by Naus and others (2001) to be small and values were considered representative of average isotopic composition in near-recharge areas.

Tritium, which beta-decays to ^{3}He with a half-life of 12.43 years (Clark and Fritz, 1997), is produced naturally in small concentrations by cosmic radiation in the stratosphere. Naturally occurring background concentrations of tritium in continental precipitation are estimated to range from 1 to 20 TU (tritium units), depending on location (Michel, 1989). One TU is defined as one ^{3}H atom per 10^{18} atoms of hydrogen, which is equivalent to 3.19 pCi/L (picocuries per liter) in water (International Atomic Energy Agency, 1981). Because of nuclear-bomb testing during the 1950’s and 1960’s and a subsequent treaty limiting such tests, tritium concentrations in atmospheric water increased sharply in 1953, peaked in 1963, and then declined. Current sources of tritium, such as nuclear power production, contribute to atmospheric tritium concentrations that are slightly higher than background concentrations prior to nuclear testing.

Numerous factors limit capabilities for age-dating of ground-water. One important factor is imprecise data for tritium concentrations in precipitation for the Black Hills area. Estimates by Naus and others (2001) are shown in figure 76, along with decay curves for selected 12-year increments that approximate the half-life decay of tritium. Estimates were derived primarily from data by Michel (1989) that were based on measurements at Ottawa, Canada (fig. 77) and other remote locations. Seasonal variability in tritium concentrations in precipitation (fig. 77) is another limiting factor. The composition of recharge water also can be affected by streamflow recharge, which is inherently older than precipitation. Especially large age differences can occur for streams such as Spearfish Creek or Rapid Creek, where large proportions of flow may originate from discharge of headwater springs.

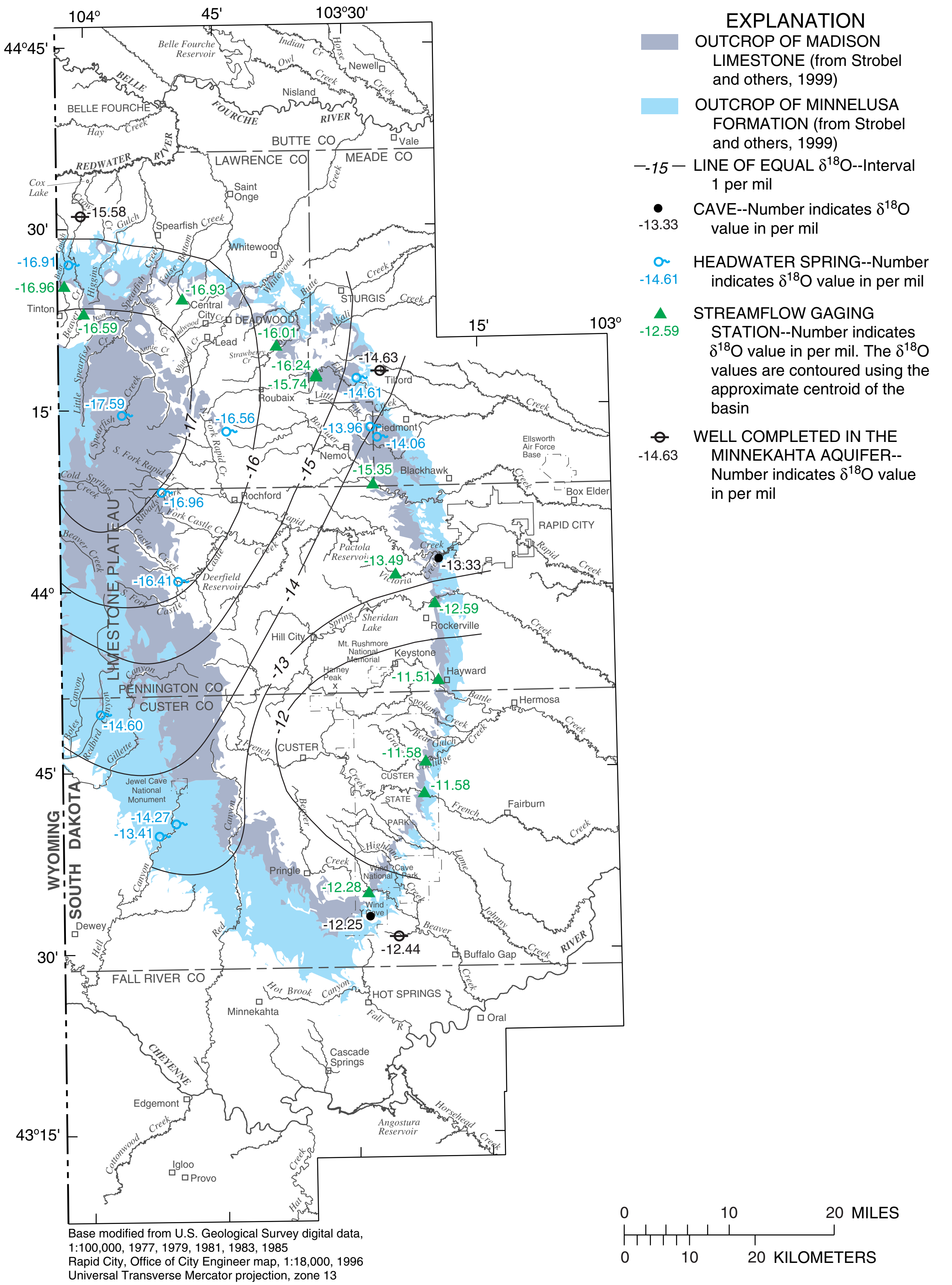


Figure 74. Generalized distribution of $\delta^{18}O$ in surface water and ground water in near-recharge areas (from Naus and others, 2001).

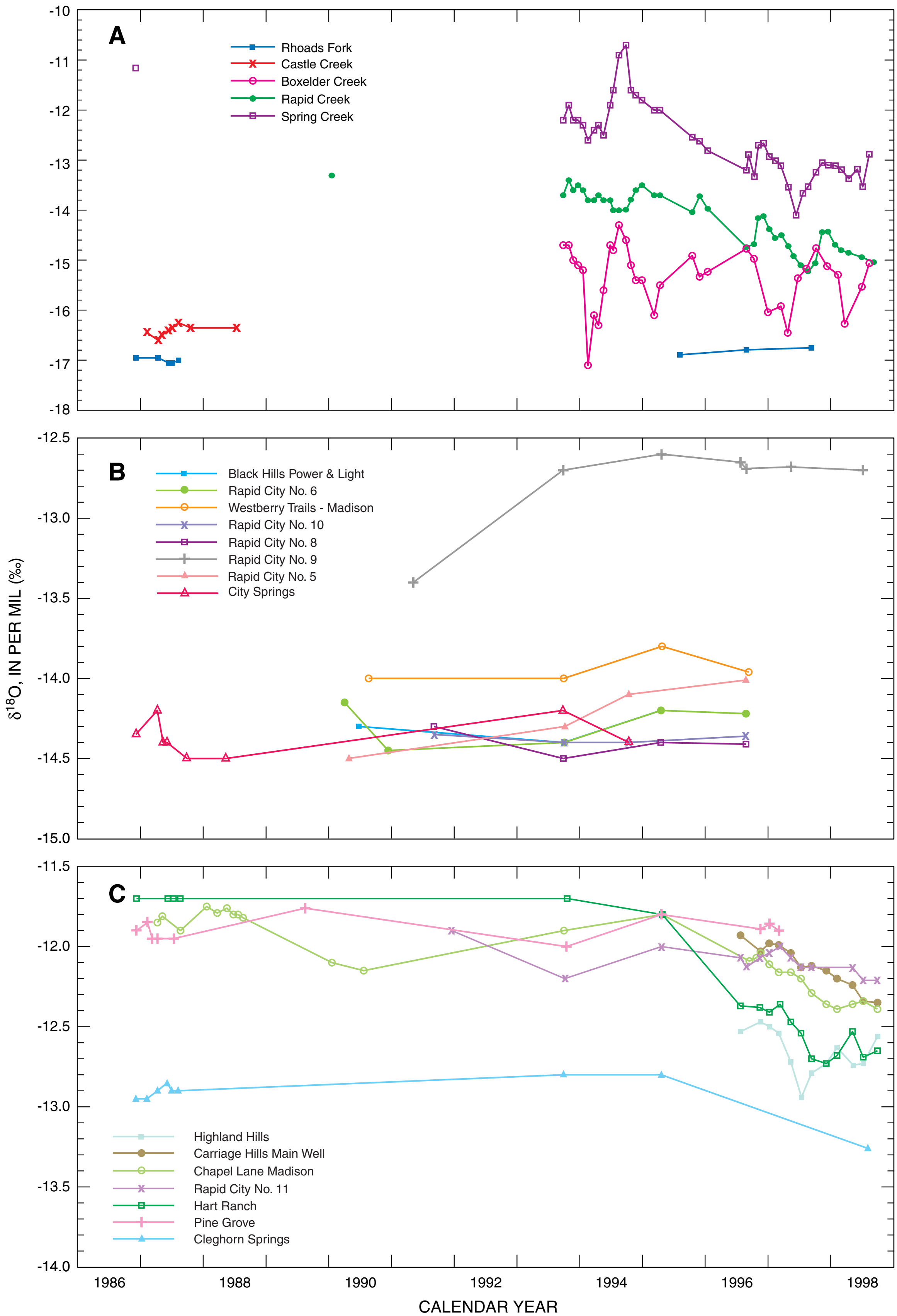


Figure 75. Temporal variation of $\delta^{18}O$ for selected sites (from Naus and others, 2001). Graph A shows selected loss-zone streams and headwater springs. Graphs B and C show selected wells and artesian springs.

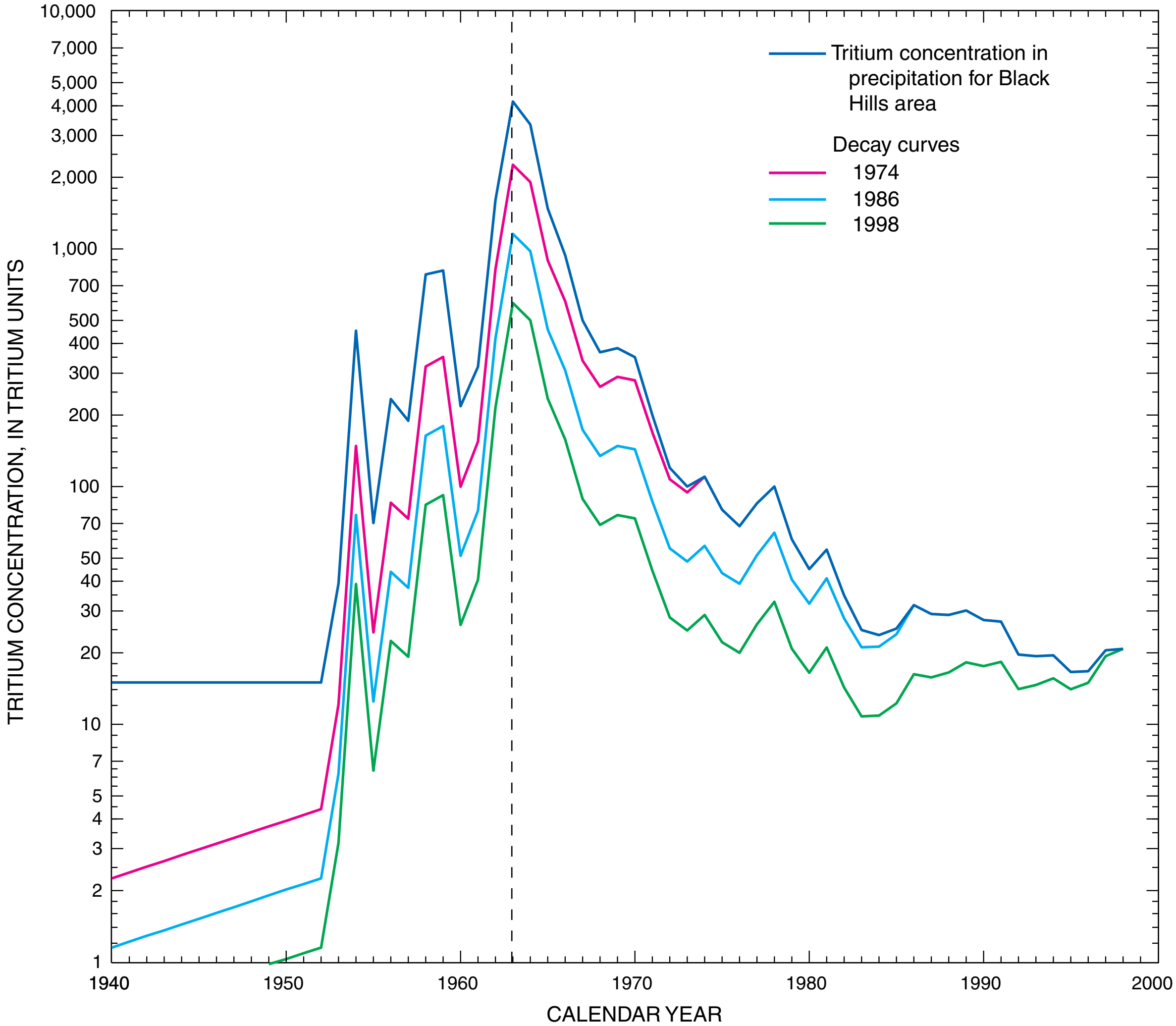


Figure 76. Estimated tritium concentrations in precipitation for Black Hills area and decay curves for selected years. Decay curves depict decayed tritium concentrations for selected sampling years. Maximum tritium concentrations of about 4,200 tritium units occurred in about 1963. Tritium has a half-life of about 12.43 years and decay curves are presented for selected 12-year increments that approximate this half-life. Using 1963 as an example, the tritium concentration in a sample collected in 1974 containing water recharged in 1963 would be equal to about 2,200 tritium units. The tritium concentration would have decayed by almost one-half to 1,100 tritium units for a sample collected 12 years later in 1986, and again by about one-half to about 600 tritium units for a sample collected in 1998.

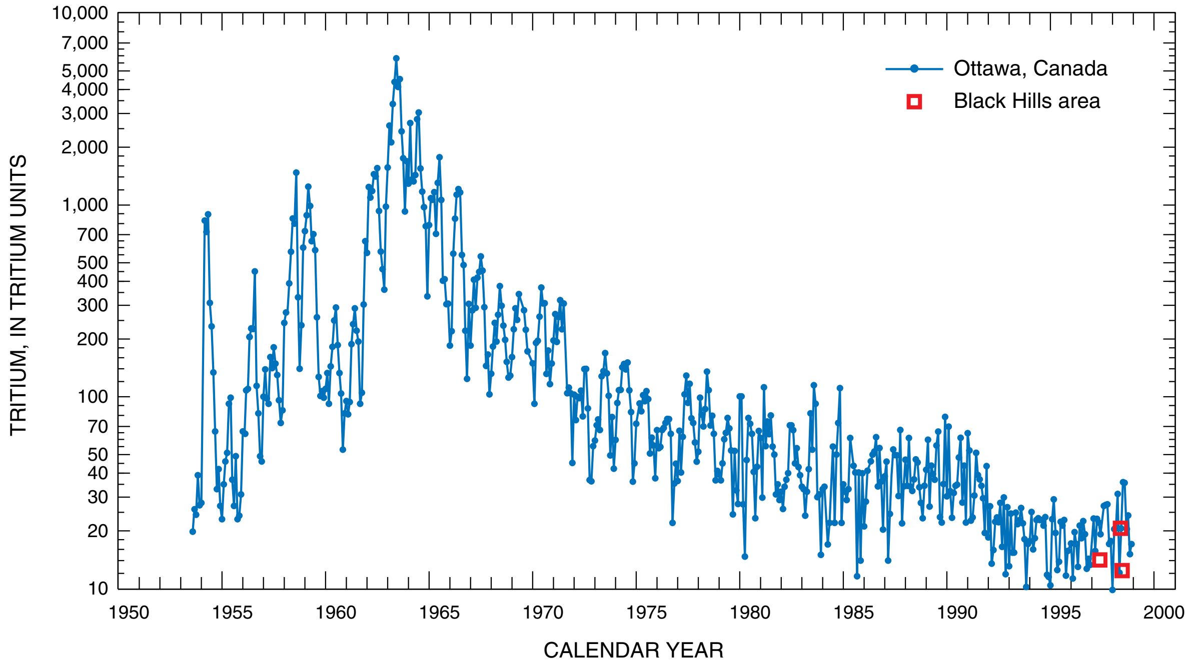


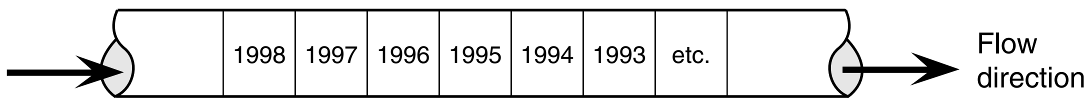
Figure 77. Monthly tritium concentrations in precipitation at Ottawa, Canada. Samples collected in Black Hills area of South Dakota also are shown (from Naus and others, 2001).

Another important consideration for age dating is ground-water mixing conditions, which can be highly variable because of large heterogeneity within the Madison and Minnelusa aquifers. Naus and others (2001) presented three simplified conceptual mixing models (fig. 78) that were used in evaluating ground-water ages and mixing conditions for the Black Hills area.

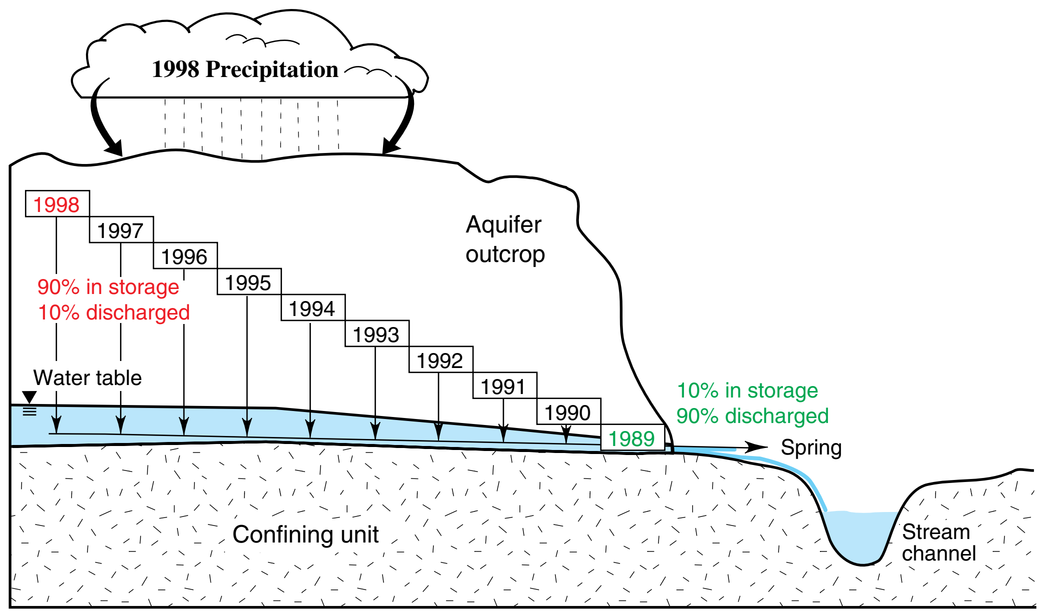
The first conceptual model (fig. 78A) depicts slug flow (often termed pipe or piston flow). The decay curves presented in figure 76 would be applicable for slug-flow conditions. For the Madison and Minnelusa aquifers, a slug-flow model could approximate ground-water flow conditions in dual-porosity settings if the dominant flow proportions are in continuous fractures and solution openings, with minimal contributions from the low-porosity matrix, dead-end fractures, or discontinuous solution openings.

The second conceptual model (fig. 78B) depicts the “immediate-arrival” mixing model, which generally is applicable for locations with water-table conditions within outcrop areas, such as headwater springs. For this scenario, water recharged during a given year is mixed with equal proportions of water recharged during previous years. For the hypothetical spring with a maximum traveltime of about 10 years that is shown in figure 78B, 10 percent of the water recharged during the current year is discharged as springflow during that same year. The remaining 90 percent of the water discharged is composed of equal proportions of water recharged during each of the previous 9 years.

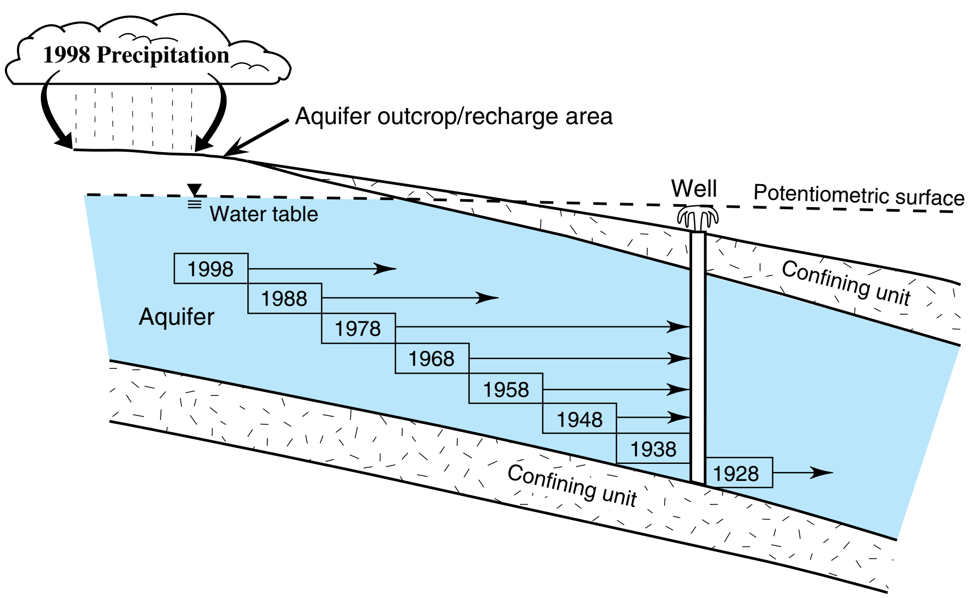
The third conceptual model (fig. 78C) depicts the “time-delay” mixing model, which assumes a delay time before any recharge water reaches a discharge point. This model generally is appropriate where an upper confining unit is present and wells or springs are located some distance from outcrops areas, which is applicable for many locations around the periphery of the Black Hills, especially where confined conditions occur. For example, the hypothetical artesian well shown in figure 78C withdraws a mixture of water that was recharged during a 50-year period from 1929 to 1978. The minimum traveltime (delay time) in this case is 20 years; in other words, the earliest arrival of recharge water is delayed by about 20 years before reaching the discharge point. The maximum traveltime is 70 years.



A Slug flow or pipe flow - negligible mixing with delayed arival



B Hypothetical water-table spring with maximum traveltime of 10 years - thorough mixing with immediate arrival



C Well completed in artesian aquifer at considerable distance from recharge area - thorough mixing with delayed arrival

Figure 78. Schematic diagrams illustrating mixing models for age dating for various ground-water flow conditions (from Naus and others, 2001).

The immediate-arrival and time-delay mixing models are based on various assumptions including equal annual recharge from year to year. Routine violation of this assumption (fig. 68) is one limitation for these conceptual models. A larger limitation is the nonhomogeneous hydraulic characteristics that commonly occur and result in nonuniform mixing conditions within the Madison and Minnelusa aquifers. Given the large range of hydraulic characteristics, the simplified models cannot address all of the complex mixing and flow conditions that occur. The models do, however, provide a mechanism by which finite numerical age estimates can be derived for water samples.

Decay-curve families for these two mixing models for various sample-collection years were presented by Naus and others (2001). Example curves for a 1995 sample-collection date, which was a midpoint for the main sampling period for the study, are presented in figure 79. The graph includes a family of curves depicting minimum traveltimes, or delay times, in 4-year increments. The 0-year delay curve in figure 79 is applicable for the immediate-arrival mixing model (fig. 78B), and the other curves are applicable for the time-delay mixing model (fig. 78C).

Using a sample concentration of 50 TU as an example, the immediate-arrival model (0-year delay curve) indicates two possible solutions, including equal annual recharge during about 1964-95 or 1930-95 (fig. 79). Two solutions also are possible for time-delay scenarios with delay times less than about 24 years. Using the 20-year delay curve for the same example, the concentration would indicate recharge during about 1969-75 or 1915-75. Because multiple solutions are possible for most sample concentrations, two or more samples usually are necessary for making general estimations of ground-water age.

Isotope Distributions and General Considerations

Distributions for $\delta^{18}O$ and tritium are presented in this section. Various general considerations associated with the isotope distributions also are discussed.

Distributions of $\delta^{18}O$ values in Madison and Minnelusa wells and selected springs are shown in figure 80. Samples from sites considered representative of the isotopic composition of recharge in the study area were presented earlier in figure 74 and are excluded from figure 80. Distributions of stable isotopes generally are consistent with spatial patterns in recharge areas (fig. 74), with isotopically lighter precipitation generally occurring at higher altitudes and latitudes.

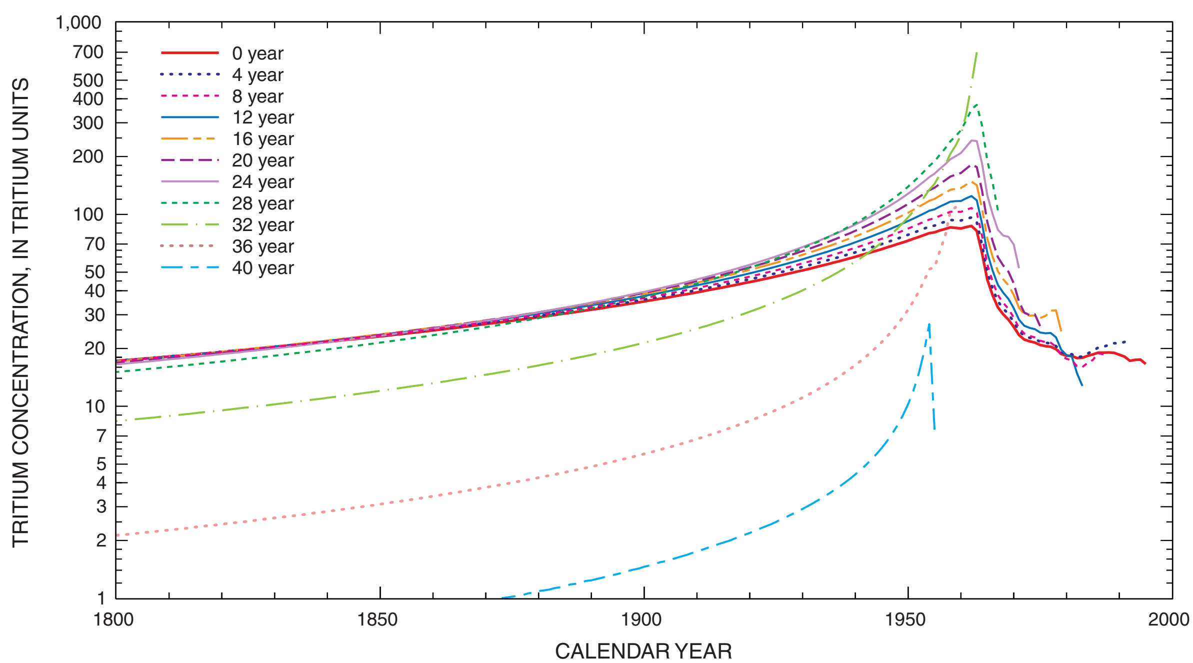


Figure 79. Decay-curve family for delayed-arrival mixing model for a 1995 sampling date (from Naus and others, 2001). Each curve shows average decayed tritium concentrations, for hypothetical mixes over time, for specified delay times that are provided in 4-year increments.

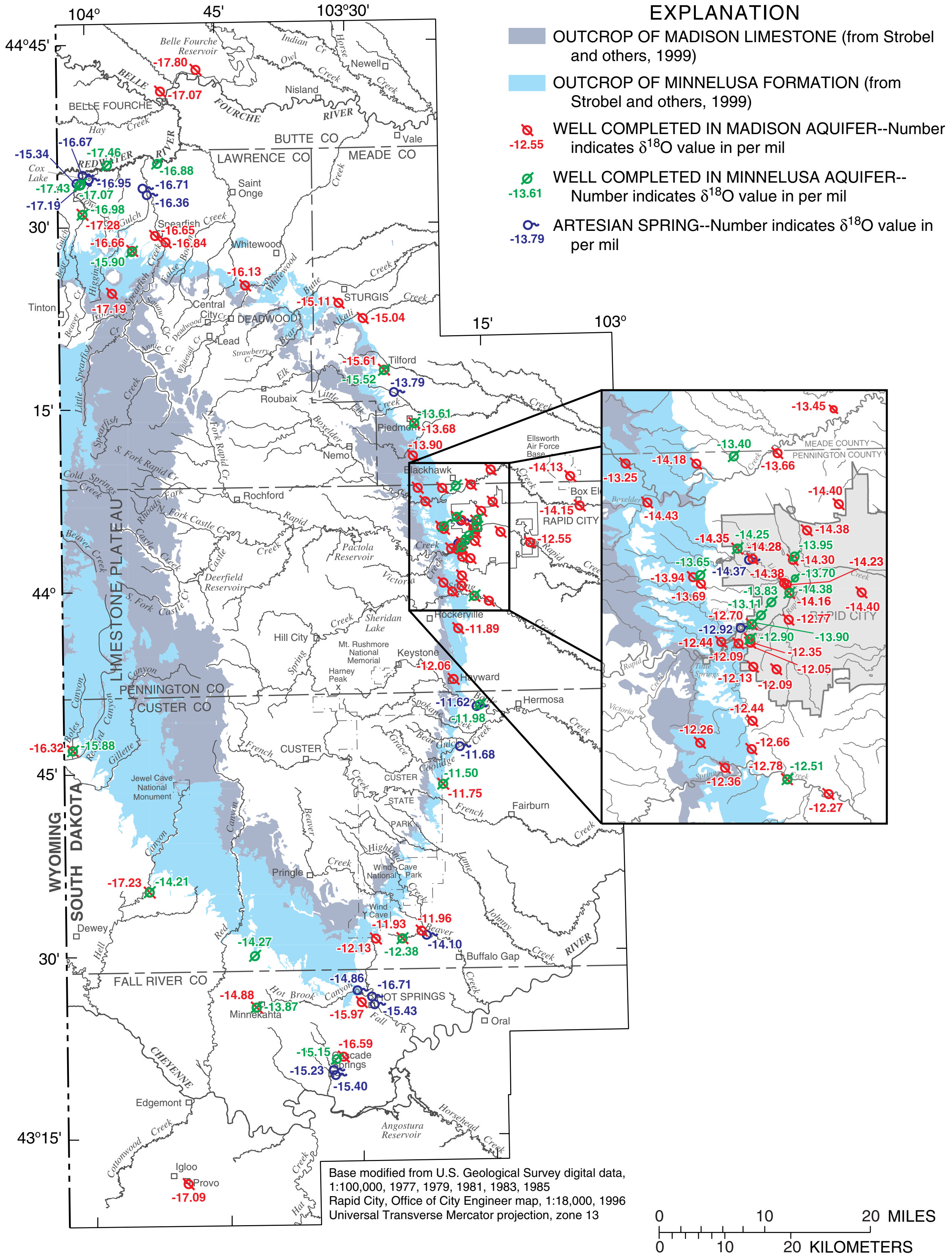


Figure 80. Distribution of $\delta^{18}O$ in samples from selected Madison and Minnelusa wells and springs in the Black Hills area (modified from Naus and others, 2001). Sampling dates are through 1998, with mean values shown for sites with multiple samples.

The effects of recharge altitude are apparent in some areas. The $\delta^{18}O$ values are lighter in the Madison aquifer than in the Minnelusa aquifer for 10 of 13 well pairs shown in figure 80 because the Madison aquifer generally is recharged at a higher altitude than the Minnelusa aquifer. The Madison aquifer also is influenced by preferentially larger volumes of isotopically light streamflow recharge, relative to the Minnelusa aquifer.

The $\delta^{18}O$ values for samples from a Madison/Minnelusa well pair at Tilford (fig. 80) are somewhat lighter than in nearby outcrop areas (fig. 74) and are distinctively lighter than for samples from a Madison/Minnelusa well pair at Piedmont and for the artesian spring between Tilford and Piedmont. The spring and the Piedmont wells probably are influenced predominantly by recharge on nearby outcrops, whereas the Tilford wells probably are influenced predominantly by streamflow recharge from Elk Creek.

In the southern Black Hills, the $\delta^{18}O$ values for samples from sites near Battle, Grace Coolidge, and French Creeks (fig. 80) are isotopically heavier than for any other part of the Black Hills area and are similar to values for nearby streamflow loss zones (fig. 74). Along the southern and southwestern flanks of the uplift, $\delta^{18}O$ values for samples from most wells and springs (fig. 80) are much lighter than estimated values for near-recharge areas immediately nearby (fig. 74), indicating either recharge areas to the northwest or possible influence of regional flow from the west, as discussed in subsequent sections of this report.

Distributions of tritium in Madison and Minnelusa wells and selected springs are shown in figure 81. Large spatial variability in concentrations occurs near outcrop areas, which reflects large variability in mixing conditions and aquifer characteristics (heterogeneity). Most samples from wells that are far removed from outcrop areas have low, or nondetectable (<0.3 TU) tritium concentrations. Concentrations noted as <0.3 TU are equivalent to about <1.0 pCi/L, which is the method reporting limit (MRL) for most of the laboratory analyses that have been performed. Samples reported as <0.3 TU are assumed to be composed primarily of water recharged prior to initial influence of nuclear testing during the early 1950's (pre-bomb water), regardless of mixing conditions (Naus and others, 2001).

Samples with tritium concentrations that equal or slightly exceed the MRL also are dominated by pre-bomb water, but probably are showing the presence of some proportion of modern water (recharged since the early 1950's). For concentrations between 0.3 and 1.0 TU, the detection of modern water is fairly certain, from an analytical standpoint, and would indicate either: (1) the initial arrival of modern water for slug-flow conditions; or (2) at least some proportion of modern water for all mixing conditions.

Given the uncertainty in estimation of tritium concentrations in recent precipitation (since about 1992) for the Black Hills area, concentrations as low as 10 TU (fig. 76) may be possible for recently recharged water. Thus, for samples with tritium concentrations between about 1 and 5 TU, dominant proportions of pre-bomb water generally can be assumed. For concentrations greater than about 5 TU, it is difficult to make generalizations because of numerous possible mixing scenarios; however, for all mixing conditions, the probability of dominant proportions of modern water increases with increasing tritium concentrations.

Boxplots showing the distribution of tritium in samples collected during 1990-98 from wells, headwater springs, artesian springs, and streams upstream from loss zones in the Black Hills area are presented in figure 82. Concentrations for samples from streams upstream of loss zones generally are comparable with estimated concentrations for precipitation in the Black Hills area since about 1985 (fig. 76). As a group, headwater springs have the highest tritium concentrations, which generally indicates relatively large proportions of modern water. Potential age ranges for selected headwater springs were estimated by Naus and others (2001) using the immediate-arrival mixing model. Concentrations for samples generally are lower from wells than artesian springs, which probably tend to develop near preferential flowpaths that may be further enhanced by dissolution activity and thus are associated with relatively faster traveltimes than flowpaths to wells.

Tritium distributions for the ground-water sites (fig. 82) provide evidence that the mixing models illustrated in figure 78 have general applicability. The lower end of the range of tritium concentrations for headwater springs is much higher than for the wells and artesian springs, which is consistent with the concept of an immediate-arrival mixing model in outcrop areas for the Madison and Minnelusa aquifers. The lower end of the range of concentrations for wells and artesian springs is near zero, which supports applicability of the time-delay model in areas where an upper confining layer is present.

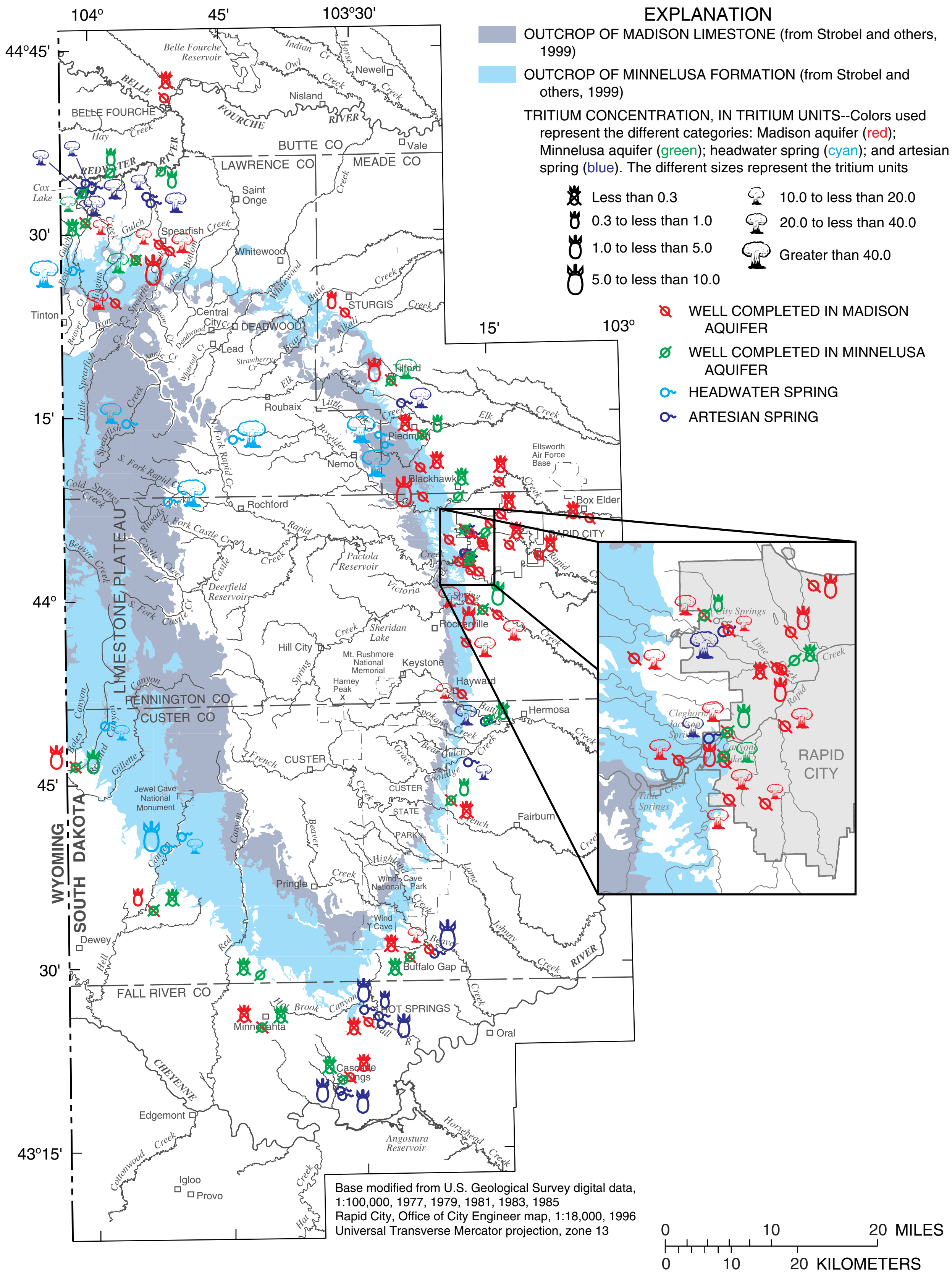


Figure 81. Tritium occurrence for selected sample sites in Black Hills area (modified from Naus and others, 2001). Sites considered include only those sampled during the 1990's with the most recent concentration shown for sites with multiple samples. Higher tritium concentrations indicate larger proportions of modern water.

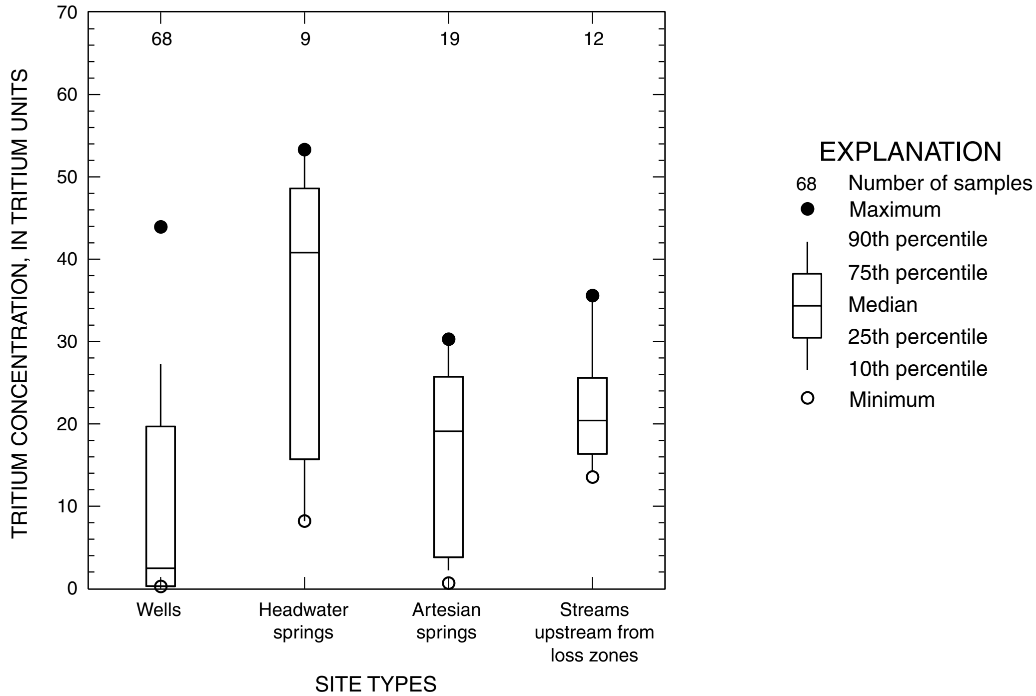


Figure 82. Boxplots of tritium concentrations for selected ground-water and surface-water samples collected during 1980-98 in the Black Hills area (from Naus and others, 2001).

The tritium distributions also provide evidence that purely slug-flow conditions probably are uncommon within the Madison and Minnelusa aquifers. For slug-flow conditions, water recharged between about 1962 and 1970 would have tritium concentrations of about 70 TU to as much as several hundred TU for all sample dates prior to 1998 (fig. 76). Tritium concentrations for the 96 ground-water samples collected since 1990, however, are uniformly less than 70 TU (fig. 82).

Flowpaths, Ages, and Mixing Conditions

A summary of interpretations regarding flowpaths, ages, and mixing conditions for the Madison and Minnelusa aquifers is presented in this section. Discussions include the Rapid City area and the northern and southern Black Hills. Regional flowpath considerations are included in discussions of the northern and southern Black Hills.

Rapid City Area

Extensive data sets of isotope information are available for the Rapid City area. In addition, dye testing in this area has provided useful information regarding flowpaths, relative ages of ground water, and mixing conditions. Evaluation of flowpaths and mixing conditions for the Rapid City area is especially complicated because: (1) recharge generally is dominated by streamflow losses, rather than precipitation on outcrop areas (Carter, Driscoll, and Hamade, 2001); and (2) large and variable withdrawals from the Madison and Minnelusa aquifers have occurred. Municipal production from the Madison aquifer has increased substantially since about 1990 (Anderson and others, 1999).

Dye testing has consisted of dye injection at a swallow hole in the loss zone of Boxelder Creek, with dye recovery documented at locations indicated in figure 83, many of which are located in the Rapid Creek drainage basin. The time required (in days) for the earliest arrival of dye from the dye injection point on Boxelder Creek to reach the various locations is shown in figure 83.

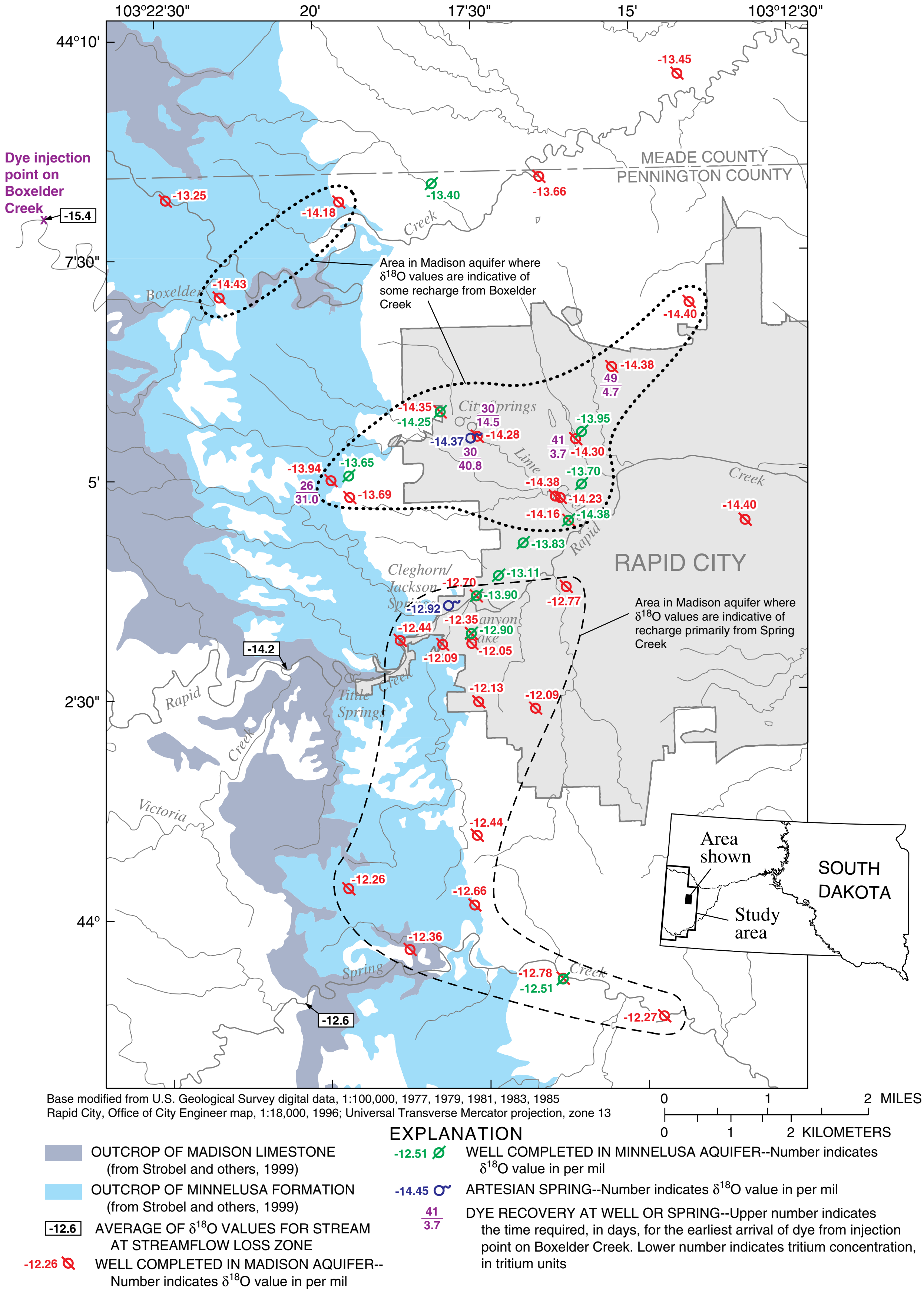


Figure 83. Concentrations of $\delta^{18}O$ in Madison and Minnelusa aquifers in Rapid City area (modified from Naus and others, 2001). Average $\delta^{18}O$ values for streamflow loss zones also are shown. In addition, traveltimes and tritium concentrations for sites at which dye recovery was reported by Greene (1999) are shown.

A distinct division of stable isotope values is apparent for the Madison aquifer along Rapid Creek (fig. 83). Areas south of Rapid Creek reflect isotopic composition similar to Spring Creek (fig. 83), which is isotopically heavier than Rapid Creek and Boxelder Creek (fig. 75). Sites north of Rapid Creek generally reflect isotopically lighter composition similar to that of Rapid Creek and Boxelder Creek, indicating a combination of recharge from these areas. Areas that probably have some influence from water recharged in Boxelder Creek also are shown in figure 83. The $\delta^{18}O$ values for samples from Minnelusa wells in the Rapid City area show a general gradation from north to south of lighter to heavier values, with no distinct division, which probably indicates larger influence from precipitation recharge, than for the Madison aquifer in this area.

Results of 1993 dye testing in the Rapid City area (Greene, 1999), in combination with tritium samples, provide relatively definitive age-dating information for several sites. A tritium concentration of 58.1 TU was measured at the dye-injection location along Boxelder Creek. Two of the wells with dye recovery (within less than 50 days of injection) had detectable, but low, tritium values (3.7 and 4.7 TU; fig. 83), which generally would be considered indicative of large proportions of pre-bomb water. This inference is reasonable, but somewhat misleading because the small proportions of modern water in the mix consist of extremely modern water. Unequal mixing of about 5 to 10 percent very recent water (tritium concentration of about 58 TU) with 90 to 95 percent pre-bomb water (0 TU) was suggested by Naus and others (2001) as a viable mix for these sites. This scenario is a noteworthy example of unequal mixing conditions that can occur in a dual-porosity system.

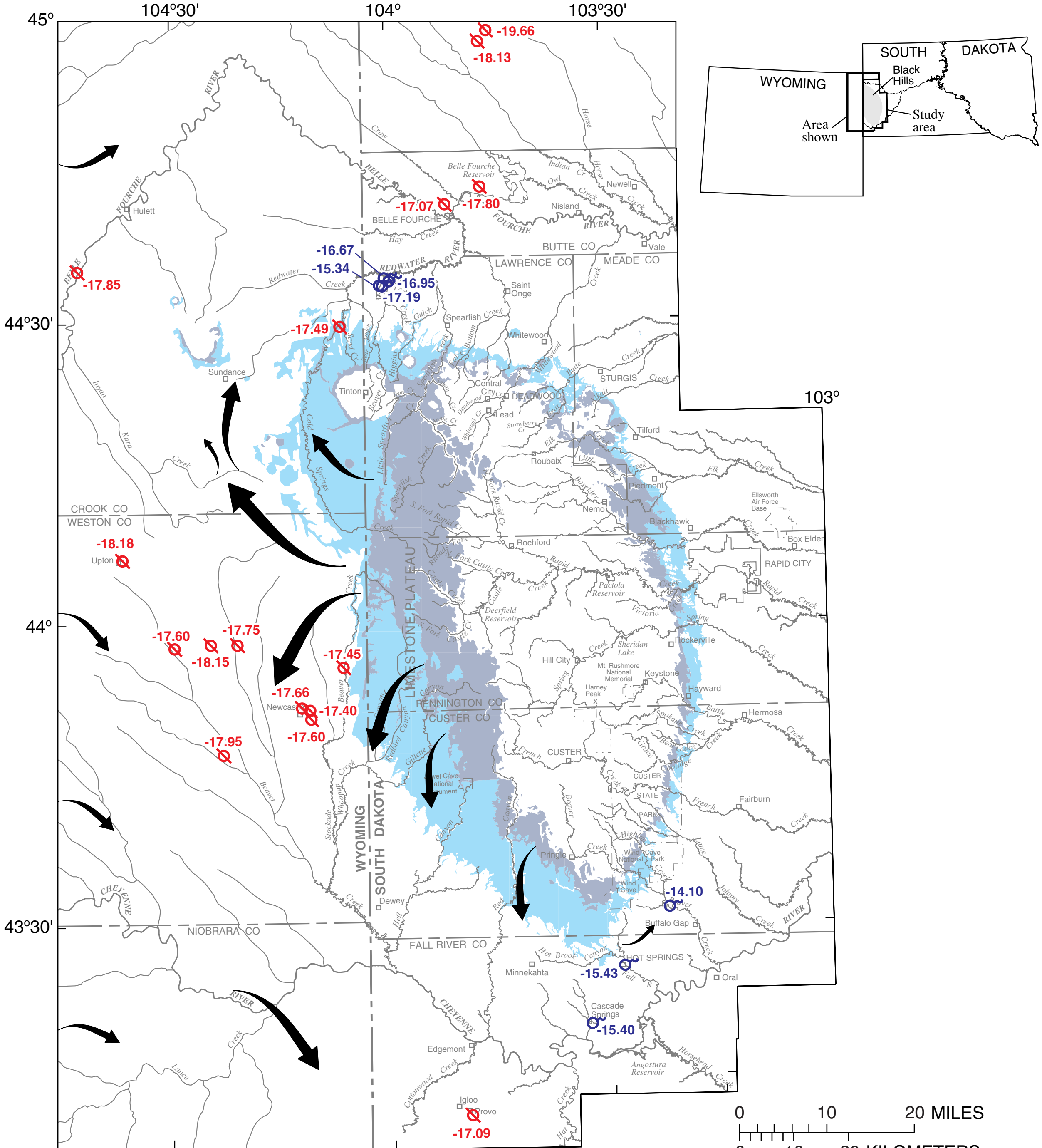
Time-series $\delta^{18}O$ and tritium data provide useful information regarding mixing conditions and general ages for numerous other sampling sites in the Rapid City area. Sites with minimal variability in $\delta^{18}O$ values generally have tritium data indicative of dominant proportions of pre-bomb water, reflecting generally thorough mixing conditions. A number of sites, however, showed response to temporal $\delta^{18}O$ trends in streamflow recharge (fig. 75), with associated tritium data generally indicating relatively large proportions of modern recharge. Water from several large Madison production wells located along Rapid Creek had changes in $\delta^{18}O$ values indicative of changes in capture zones associated with recent production. Potential changes in ground-water age characteristics at several wells also were indicated by tritium concentrations for sequential samples.

Northern Black Hills

Naus and others (2001) examined the possible influence of regional flow components from the west for the Madison aquifer in the northern part of the study area. Samples from two Madison wells just north of the study area (fig. 84) show influence of regional flow, with $\delta^{18}O$ values (-18.13 ‰ and -19.66 ‰) that are much lighter than any of the values in the Black Hills recharge areas (fig. 74). Busby and others (1983) and Plummer and others (1990) noted samples from Madison wells near recharge areas in Wyoming near the Bighorn Mountains with $\delta^{18}O$ values as light as -18.5 ‰ and near the Laramie Mountains as light as -19.25 ‰. Concentrations of common ions for samples from the two wells just north of the study area (Busby and others, 1991) are approximately an order of magnitude higher than for samples from wells along the northwestern flank of the Black Hills and are consistent with a regional flowpath trending northeasterly from the Bighorn Mountains in Wyoming (fig. 17).

The $\delta^{18}O$ value of -17.80 ‰ for a well sample northeast of Belle Fourche (fig. 84) indicates possible minor influence from regional flow; however, the value of -17.07 ‰ for the well sample at Belle Fourche provides no indication of regional influence. Ion concentrations for the well sample northeast of Belle Fourche are slightly higher than for the well sample at Belle Fourche (fig. 34), but are much lower than concentrations for samples from the two wells just north of the study area (Busby and others, 1991). Ion concentrations also are low for a sample from a well just north of Nisland (fig. 34). Thus, Naus and others (2001) concluded that regional flowpaths in the Madison aquifer are largely deflected to the north of the study area. Regional $\delta^{18}O$ values are not available for the Minnelusa aquifer; however, values for Minnelusa wells along the northwestern flank of the uplift (fig. 84) are similar to those in near-recharge areas to the south and west (fig. 74).

Several large artesian springs located along the northern axis of the uplift are a major hydrologic feature of the northern Black Hills area. These springs are a major discharge area for the Madison and Minnelusa aquifers (Klemp, 1995), with cumulative discharge of all artesian springs along the northern flank estimated as 90 ft^3/s for 1987-96 (Carter, Driscoll, Hamade, and



Base modified from U.S. Geological Survey digital data, 1:100,000, 1977, 1979, 1981, 1983, 1985
Rapid City, Office of City Engineer map, 1:18,000, 1996; Universal Transverse Mercator projection, zone 13

EXPLANATION

OUTCROP OF MADISON LIMESTONE (from Strobel and others, 1999; DeWitt and others, 1989)

OUTCROP OF MINNELUSA FORMATION (from Strobel and others, 1999; DeWitt and others, 1989)

GENERALIZED FLOWPATH

-17.09 WELL COMPLETED IN MADISON AQUIFER--Number indicates $\delta^{18}O$ value in per mil

-15.40 ARTESIAN SPRING--Number indicates $\delta^{18}O$ value in per mil

Figure 84. Distribution of $\delta^{18}O$ in selected Madison wells and springs and generalized flowpaths, based on $\delta^{18}O$ values, in the Black Hills of South Dakota and Wyoming (from Naus and others, 2001).

Jarrell, 2001). The $\delta^{18}O$ values (fig. 80) and tritium concentrations (fig. 81) for most of these springs are quite similar, which probably indicates generally thorough mixing conditions resulting from large discharges (and associated large recharge areas). The tritium concentrations are consistently indicative of modern recharge, which does not necessarily preclude regional flow contributions, but confirms Black Hills recharge as the primary recharge component. Multiple tritium samples are available for several springs and wells along the northern flank of the uplift. General evaluations of potential age ranges derived using the time-delay mixing model were discussed by Naus and others (2001).

Results of water-budget analyses for the Madison and Minnelusa aquifers (figs. 69 and 70, table 15) support the geochemical interpretation of minimal influence from regional flowpaths in the northern part of the study area. For subarea 1, which includes the large artesian spring area along the northern axis, ground-water outflow is estimated to exceed inflow by about 30 ft^3/s. This analysis does not necessarily preclude a regional ground-water flow component in the northern part of the study area, but does support the conclusion of dominance by recharge in the Black Hills area.

Southern Black Hills

Naus and others (2001) also evaluated the possible influence of regional flow components for the Madison aquifer in the southern part of the study area. Light $\delta^{18}O$ values from samples for several sites west of the study area in Weston County (fig. 84) are indicative of a possible transition zone between regional and areal flowpaths. Ion concentrations for samples from wells in this area (Busby and others, 1991) generally showed little, if any, influence of regional flow, however. High ion concentrations for samples from several Madison wells in the southwestern corner of the study area (fig. 34) indicate possible regional influence, but cannot necessarily be distinguished from basinward increases in constituent concentrations. Farther north, regional influence probably is minor or negligible.

Several large artesian springs located along the southern axis of the Black Hills uplift comprise another major discharge area for the Madison and Minnelusa aquifers (Whalen, 1994; Hayes, 1999). For subareas 7 and 8 (figs. 69 and 70), artesian spring discharge averaged about 54 ft^3/s during 1987-96 (table 15). Combined streamflow and precipitation recharge for these subareas averaged only about 14 ft^3/s; thus, much of the recharge for the springs probably occurs in subarea 9, where recharge averaged about 60 ft^3/s. Considering the three subareas (7, 8, and 9) collectively, ground-water outflow is estimated to exceed inflow by about 7 ft^3/s. This analysis does not necessarily preclude a regional ground-water flow component in the southern part of the study area, but does indicate dominance by recharge in the Black Hills area.

Dominance by recharge in the Black Hills area also is indicated by the $\delta^{18}O$ values for the large springs (fig. 84), which are notably heavier than reported regional values (Busby and others, 1983; Plummer and others, 1990) and essentially preclude substantial influence from regional flowpaths. This conclusion is supported by ion concentrations for the springs (fig. 34 and 35), which are insufficient to be indicative of regional flow. High sulfate concentrations for several of the springs reflect influence from dissolution of anhydrite within the Minnelusa Formation.

Low, but detectable, tritium concentrations in the large springs (fig. 81) confirm the influence of recharge from within the study area, but indicate relatively long traveltimes. This is consistent with the $\delta^{18}O$ values, which indicate potential recharge areas extending along the entire southwestern flank of the uplift. Generally low tritium concentrations for large springs in the southern Black Hills, relative to concentrations for springs along the northern axis, may be influenced by generally smaller recharge rates in the southern Black Hills.

The $\delta^{18}O$ value for Beaver Creek Spring (-14.10 ‰; figs. 80 and 84) located just northwest of Buffalo Gap is much lighter than estimated values for nearby outcrop areas and samples from nearby wells, which indicates a possible flowpath extending from the general Hot Springs area. This interpretation also is supported by a low tritium value for Beaver Creek Spring (fig. 81), which indicates generally long traveltimes, and by the water-budget analysis, which indicates a substantial flow component to the spring from west of the uplift axis (figs. 69 and 70, table 15).

Tritium concentrations for artesian springs along Battle and Grace Coolidge Creeks indicate younger water than for several nearby wells (fig. 81), which is consistent with development of artesian springs along preferential flowpaths. The discharges of these springs (which are located in subarea 5) are highly variable (table 15), and probably are influenced primarily by variability in streamflow recharge, which dominates recharge in this subarea.

Interactions Between Aquifers

Geologic conditions facilitate hydraulic connection between the Madison and Minnelusa aquifers. Confining layers in the lower portion of the Minnelusa Formation probably are influenced by paleokarst features such as caverns and sinkholes in the upper Madison Limestone. Extensive fracturing and solution activity have contributed to enhanced secondary porosity in both formations and decreased competency of the confining layers. Potential exists for downward leakage (from Minnelusa to Madison) in recharge areas where the aquifers are unconfined (water-table conditions), and for either upward or downward leakage, depending on direction of hydraulic gradient, where confined conditions exist. Hydraulic connections probably occur at various artesian springs, many of which discharge stratigraphically within or slightly above the Minnelusa Formation, but may include the Madison aquifer as a source (Whalen, 1994; Klemp, 1995; Hayes, 1999). Naus and others (2001) evaluated potential interactions through analysis of hydraulic and geochemical information for well pairs and artesian springs.

Hydrographs for Madison and Minnelusa well pairs were presented previously in figs. 26-29 and potential interactions were discussed in the section "Comparisons between Madison and Minnelusa aquifers." Hydraulic connection for most colocated wells cannot be confirmed or refuted because aquifer testing has not been performed. Of the well pairs with generally similar hydrographs (fig. 27), hydraulic connection has been confirmed by aquifer testing (Greene, 1993) only for the City Quarry wells (fig. 27D). These wells are located near City Springs, where aquifer connection has been confirmed by dye testing (Greene, 1999). Aquifer testing (Greene and others, 1999) provided no indication of hydraulic connection in the vicinity of the Spearfish Golf Course wells (fig. 27A). Hydrographs for many other colocated wells indicate distinct hydraulic separation and no evidence of direct hydraulic connection between the aquifers.

Although some exchange of water must occur in locations where hydraulic head differences occur, Naus and others (2001) concluded from examination of geochemical information for colocated wells that general leakage between the Madison and Minnelusa aquifers probably does not result in areally extensive mixing in most locations. The most notable differences in ion chemistry (figs. 34 and 35) are in concentrations of sulfate, which increase with increasing distance from recharge areas in the Minnelusa aquifer but do not provide definitive information regarding aquifer mixing. Although $\delta^{18}O$ values for samples from colocated Madison and Minnelusa wells (fig. 80) reflect generally higher recharge altitudes for the Madison aquifer than for the Minnelusa aquifer, information regarding aquifer mixing is inconclusive. Similarly, tritium concentrations for well pairs (fig. 81) do not provide conclusive information.

Substantial interactions between the Madison and Minnelusa aquifers can occur at artesian springs, many of which have large discharges. Combined discharge of all artesian springs within the Black Hills area of South Dakota and Wyoming was estimated by Driscoll and Carter (2001) as 169 ft^3/s for 1950-98, which represents about 46 percent of average recharge to both aquifers (table 14). Numerous investigators have identified the Madison and Minnelusa aquifers as probable sources for artesian springs in the Black Hills, based on hydraulic properties and geochemical characteristics. The Minnekahta aquifer also may be a contributing source in locations where the Minnekahta Limestone is present. The underlying Deadwood aquifer also cannot be discounted as a possible source.

Locations of artesian springs are shown on the previously presented potentiometric-surface maps of the Madison and Minnelusa aquifers (figs. 19 and 20). Hydraulic heads in these aquifers at major spring locations were estimated from potentiometric-surface maps and are summarized in table 18, along with the approximate land-surface altitudes near the springs.

Several hydraulic possibilities exist for interactions between the Madison and Minnelusa aquifers at spring locations, including: (1) water originates only from the Minnelusa aquifer, with no contribution from the underlying Madison aquifer; (2) water originates entirely from the Madison aquifer and passes through the Minnelusa Formation, with little interaction; (3) water originates entirely from the Madison aquifer, part of which discharges at the surface and part of which recharges the Minnelusa aquifer; and (4) water originating from both aquifers contributes to springflow. For cases where the Madison aquifer contributes to springflow, leakage to the Minnelusa aquifer could consist of either focused leakage in the immediate vicinity of the spring-discharge point or general leakage in upgradient locations.

Table 18. Selected hydraulic and geochemical information for major artesian springs

[Modified from Naus and others (2001). ft^3/s, cubic feet per second; mg/L, milligrams per liter; <, less than; ≈, approximately equal]

Name	Approximate discharge (ft^3/s)	Altitude of land surface (feet above sea level)	Hydraulic head (feet above sea level)		Sulfate (mg/L)	Estimated spring source from previous studies
			Madison aquifer	Minnelusa aquifer		
Higgins Gulch	5-10	3,405	3,490	3,550	110	Mostly Madison[1]
Old Spearfish Hatchery	≈5	3,405	3,500	3,550	340	70 percent Madison, 30 percent Minnelusa[1]
NcNenny Rearing Pond	≈1	3,400	3,720	3,580	130	Mostly Madison[1]
Mirror Lake	≈1	3,410	3,720	3,580	1,600	50 percent Madison, 50 percent Minnelusa[1]
Cox Lake	≈5	3,415	3,705	3,580	545	Mostly Madison[1]
Crow Creek	30-50	3,355	3,710	3,560	580	--
Elk Creek	0-20	3,450	3,450	3,450	420	--
City Springs	0-5	3,440	3,450	3,450	98	--
Cleghorn Springs	20-25	3,380	3,420	3,380	25	--
Battle Creek	1-10	3,540	3,540	3,540	19	--
Grace Coolidge Creek	0-20	3,650	3,650	3,650	11	--
Beaver Creek Spring	10-15	3,460	3,480	3,480	1,300	Mostly Madison with dissolved Minnelusa minerals[2]
Hot Brook Spring	<5	3,625	3,700	3,625	76	Mostly Madison[2]
Evans Plunge Spring	<5	3,465	3,610	3,420	540	Mostly Madison[2]
Fall River	20-30	3,415	3,580	3,360	400	--
Cool Spring	≈2	3,450	3,505	3,450	830	--
Cascade Spring	18-22	3,440	3,495	3,450	1,500	Mostly Madison with dissolved Minnelusa minerals[3]

[1]Estimated by Klemp (1995).
[2]Estimated by Whalen (1994).
[3]Estimated by Hayes (1999).

Precise quantification of relative contributions from source aquifers to individual springs is not necessarily possible; however, the Minnelusa aquifer probably can be discounted as a primary source for springs located where the Minnelusa Formation is exposed, which generally precludes confined conditions in the Minnelusa aquifer. This setting exists at Cleghorn/Jackson Springs in western Rapid City and at Hot Brook Spring located just northwest of Hot Springs (fig. 20). Other artesian springs listed in table 18 occur in locations where the Minnelusa Formation is confined by overlying units and confined conditions are assumed for both the Madison and Minnelusa aquifers. Mapped hydraulic heads at most artesian spring locations are higher in the Madison aquifer than the Minnelusa aquifer (table 18, figs. 19 and 20), which could indicate higher potential for contributions to springflow from the Madison aquifer, relative to contributions from the Minnelusa aquifer. An alternative line of reasoning may be plausible, however. Higher hydraulic head in the Madison aquifer also indicates relatively competent confinement by the overlying Minnelusa Formation, which could imply larger contributions from the Minnelusa aquifer. Thus, generalities regarding dominant contributions to artesian springflow cannot be inferred from comparisons of hydraulic head.

Discharge characteristics for artesian springs (table 18) probably are affected by hydraulic conditions at spring locations. Flow variability is minimal for many artesian springs (Miller and Driscoll, 1998; Anderson and others, 1999; U.S. Geological Survey, 2000), including McNenny Rearing Pond, Cox Lake, Crow Creek, Cleghorn Springs, Beaver Creek Springs, Fall River, and Cascade Springs. Hydraulic heads at these sites generally are substantially above land surface for one or both of the two aquifers. In contrast, discharge is much more variable for Elk Creek, City Springs, Battle Creek, and Grace Coolidge Creek. At these sites, mapped hydraulic heads in both aquifers are approximately coincident with land-surface altitude.

Contributions from the Madison and Minnelusa aquifers to individual artesian springs cannot necessarily be quantified precisely (Naus and others, 2001) because of geochemical similarities between the aquifers (figs. 35 and 36). For some springs, high sulfate concentrations (table 18) could indicate Minnelusa influence, but may result from dissolution of Minnelusa minerals by water from the Madison aquifer. Previous investigators (Whalen, 1994; Klemp, 1995; and Hayes, 1999) used geochemical modeling to estimate contributions of the Madison and Minnelusa aquifers to selected springs; generalized results are summarized in table 18. The Madison aquifer generally was identified as the primary source, with variable contributions from the Minnelusa aquifer, or chemical influences resulting from residence time within the Minnelusa Formation.

Saturation indices reported by Naus and others (2001) indicate that the Madison aquifer is undersaturated with respect to gypsum, even at the highest sulfate concentrations. Generally higher hydraulic head in the Madison aquifer, in combination with gypsum undersaturation, was concluded by Naus and others (2001) to be a primary mechanism driving interactions with the Minnelusa aquifer, in areas where confined conditions exist. Upward leakage from the Madison aquifer probably contributes to general dissolution of anhydrite deposits and development of breccia pipes in the Minnelusa aquifer. Breccia development may be especially prevalent in locations where the competency of intervening confining layers has been decreased by fracturing or by depositional influences in the Minnelusa Formation resulting from paleokarstification of the Madison Limestone.

Hayes (1999) hypothesized that upward leakage from the Madison aquifer was contributing to ongoing development of breccia pipes at Cascade Springs. Hayes (1999) noted that breccia pipes commonly occur in the upper Minnelusa Formation, but very few have been observed in the lower part of the formation. Networks of interbedded breccia layers and short, vertical breccia dikes do occur in the lower Minnelusa Formation, however, as schematically illustrated in figure 85.

Development of breccia pipes probably contributes to enhanced vertical hydraulic conductivity in the Minnelusa aquifer. Breccia pipes are a likely pathway for upward movement of large quantities of water through the Minnelusa aquifer at artesian spring locations. Dissolution processes are an important factor in a self-perpetuating process associated with development of preferential flowpaths and artesian springs. Preferential flowpaths initially develop in locations with large secondary porosity and associated hydraulic conductivity, with ongoing enhancement resulting from dissolution activity.

Hayes (1999) further hypothesized that many exposed breccia pipes of the upper Minnelusa Formation probably are the throats of abandoned artesian springs. An outward (downgradient) shifting of locations of artesian springs probably has occurred as upgradient spring-discharge points are abandoned and new ones are occupied, keeping pace with regional erosion over geologic time (Hayes, 1999). In response, hydraulic heads in the Madison and Minnelusa aquifers have declined over geologic time, as indicated by exposed breccia pipes located upgradient from Cascade Springs (Hayes, 1999). Further supporting evidence is provided by Ford and others (1993), who concluded that water-level declines of more than 300 ft have occurred in the Madison aquifer during the last 350,000 years, based on geochemical data for Wind Cave.

Ground water discharging from the Madison aquifer at artesian springs was referred to as "rejected recharge" by Huntoon (1985), who hypothesized that recharge is rejected as transmissivity decreases with distance from upgradient recharge areas. This hypothesis is consistent with decreasing potential for large secondary porosity with increasing distance from the uplift, which results from: (1) decreased deformation and associated fracturing of rocks; and (2) decreasing potential for dissolution enhancement associated with increasing basinward concentrations of dissolved constituents.

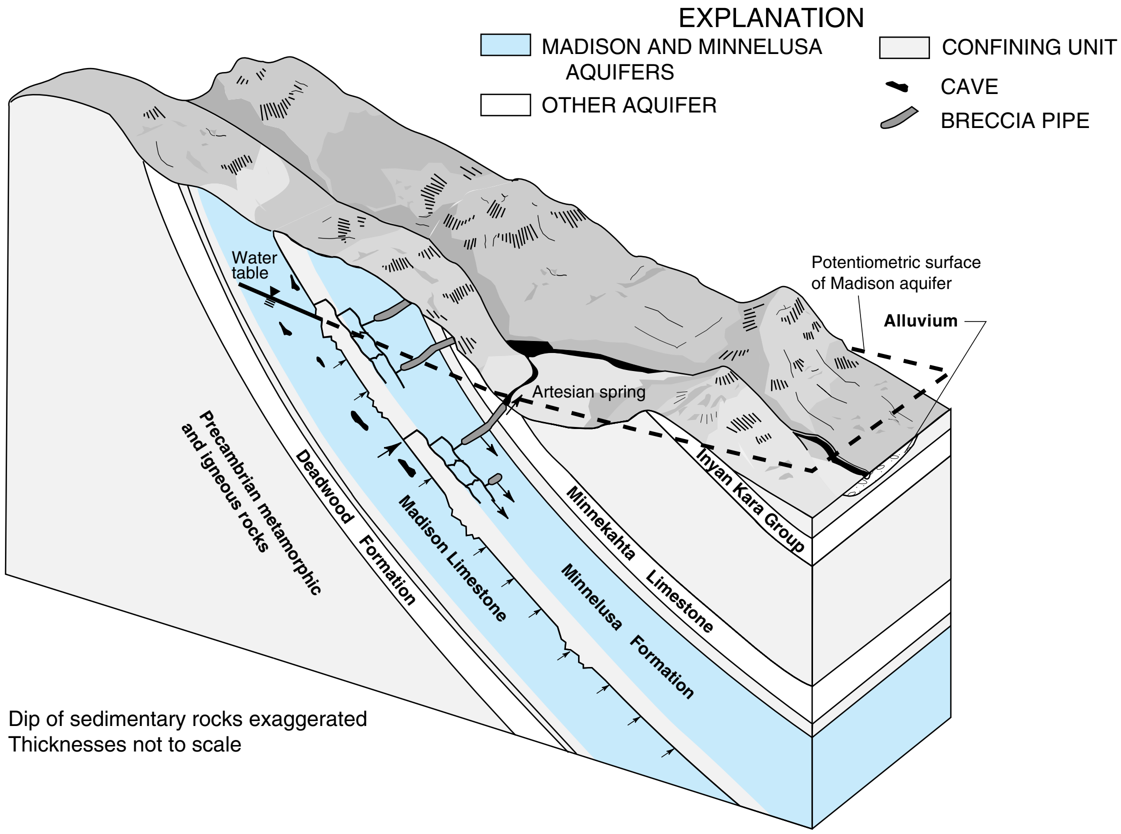


Figure 85. Schematic showing breccia pipes and caves in relation to the hydrogeologic setting of the Black Hills area (modified from Hayes, 1999). Breccia pipes contribute to secondary porosity in the Madison and Minnelusa aquifers and may result from upward leakage from the Madison aquifer, creating conduits for artesian springs. Arrows show general areal leakage and focused leakage at breccia pipes.

Artesian springs are essentially a "pressure-relief" mechanism that influence the upper limit for hydraulic head. As previously noted, springs located where hydraulic head is substantially above land surface generally have relatively stable discharge. In locations where hydraulic head is near land surface, however, discharge characteristics generally are more variable, with springflow increasing in response to increasing water levels.

Considering all available information, it was concluded by Naus and others (2001) that interactions between the Madison and Minnelusa aquifers are an important factor governing the hydraulic behavior of the two aquifers. The exchange of water resulting from general, areal leakage probably is small, relative to that which occurs near artesian springs; however, both processes probably contribute to the control of hydraulic heads in the Black Hills area. Interactions between the aquifers are an important consideration regarding the hydraulic behavior of the Madison and Minnelusa aquifer system, which has a major influence on the overall hydrology of the Black Hills area, as discussed in the following section.

Influence on Overall Hydrology of Black Hills Area

The Madison and Minnelusa aquifers strongly influence the hydrology of the Black Hills area and have important water-management implications. The ground-water hydrology of the area is dominated by the Madison and Minnelusa aquifers, which receive about 84 percent of estimated recharge to all bedrock aquifers in the study area. The Madison and Minnelusa aquifers are the most heavily used aquifers in the study area, with ongoing development occurring to meet steadily expanding needs for ground-water resources. Surface-water resources, which also are heavily used in the study area, are influenced by interactions with these aquifers.

Headwater springflow, which originates primarily from the Madison and Minnelusa aquifers in the Limestone Plateau area, provides a reliable source of baseflow in several streams. Relatively stable discharge is an important characteristic of the limestone headwater springs, which generally are influenced primarily by long-term, rather than short-term, climatic conditions. Thus, adequate baseflow for sustaining aquatic populations often can be maintained during extended periods of dry conditions in streams with substantial influence from headwater springflow. A general absence of direct runoff is another distinguishing characteristic of areas with extensive outcrops of the Madison Limestone and Minnelusa Formation.

Streamflow losses that provide recharge to the Madison and Minnelusa aquifers are another important influence on the hydrology of the Black Hills area. Only two streams in the study area maintain perennial flow through loss zones—Whitewood Creek, in which the loss zone probably has been sealed by mine tailings (Hortness and Driscoll, 1998), and Rapid Creek. Spearfish Creek also would maintain perennial flow under natural conditions; however, flow is diverted around the loss zone for hydroelectric power generation. Utilization of surface- and ground-water resources upstream from loss zones has potential to reduce recharge to the Madison and Minnelusa aquifers. In some streams, discharge of limestone headwater springs can provide a relatively consistent source of streamflow for subsequent recharge in loss zones.

Streamflow loss zones create potential for direct introduction of contaminants into the Madison and Minnelusa aquifers, which can have small filtration capacity and large potential for rapid transport of contaminants because of large secondary permeability. Similar concerns exist for all recharge areas of these aquifers, regardless of proximity to streamflow loss zones. The Madison and Minnelusa aquifers are among the most sensitive of all hydrogeologic units in the study area, relative to contamination potential (Davis and others, 1994; Putnam, 2000).

Artesian springflow from the Madison and Minnelusa aquifers is an especially important influence on surface-water resources in the study area. Artesian springflow provides large and consistent baseflow for many streams in exterior areas, where zero-flow conditions commonly occur in streams that are not influenced by artesian springs. This baseflow is especially important in maintaining adequate flow for aquatic populations and for various water-supply needs including stock water, municipal supply, and irrigation. In many locations irrigation withdrawals are made directly from streams. In addition, artesian springs provide large and consistent baseflow upstream from Angostura and Belle Fourche Reservoirs, which also are heavily used for recreational purposes.

As previously mentioned, artesian springs are essentially a pressure-relief mechanism that influence the upper limit for hydraulic head in the Madison and Minnelusa aquifers. These aquifers have a “maximum-sustainable equilibrium” water level, which is controlled by the discharge of artesian springs, with springflow increasing in response to increasing recharge and water levels. Conversely, artesian springflow decreases in response to decreasing water levels. As discussed previously, hydraulic heads are approximately coincident with land-surface altitudes near some artesian springs (table 18). Thus, large-scale well withdrawals near some artesian springs could diminish springflow before large-scale declines in water levels would occur. However, to some extent, the large recharge potential of the Madison and Minnelusa aquifers may be sufficient to replenish springflow and water levels relatively quickly during episodic periods of prolonged wet conditions.

Additional insights regarding potential effects of large-scale development of the Madison and Minnelusa aquifers can be obtained by reviewing the water budget for these aquifers for the entire Black Hills area (including both South Dakota and Wyoming) that was presented previously in table 14. For this budget, the dominant outflow component is artesian springflow of 169 ft^3/s, which is larger than combined well withdrawals (28 ft^3/s) and ground-water outflow from the study area (100 ft^3/s). Hypothetically, artesian springflow could be replaced by increased well withdrawals without causing substantial changes in hydraulic gradient or associated ground-water outflow from the study area (figs. 69 and 70).

The previous interpretations could provide insights regarding future management of ground-water resources and potential influences on surface-water resources. Many large springs currently have relatively stable discharge, whereas others, including both small and large springs, have large variability in discharge characteristics. Large-scale development of the Madison and Minnelusa aquifers probably has the potential to influence the balance of this dynamic "plumbing system." Potential effects would be most pronounced during prolonged drought periods, which would result in decreased recharge and increased demand. Although discharges of many artesian springs respond slowly to changes in recharge conditions (both decreases and increases), other springs respond more quickly. Similarly, responses to large-scale increases in well withdrawals probably would be distinct for particular withdrawal scenarios.

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