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.

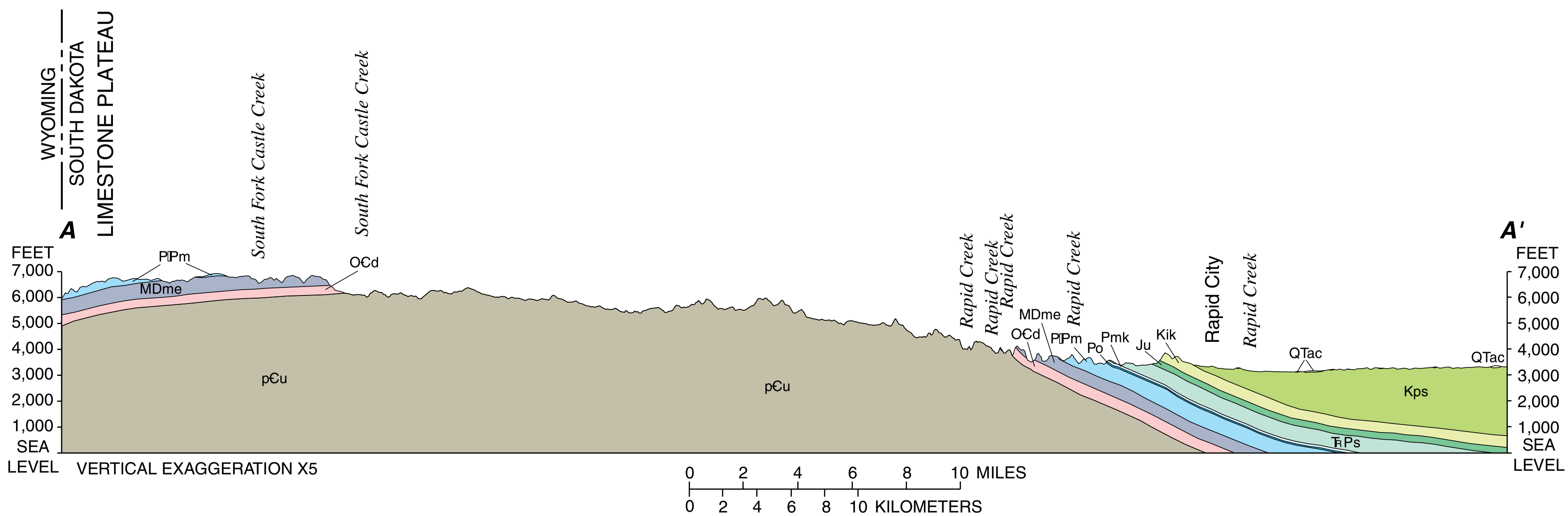


Figure 15. Geologic cross section A-A′ (modified from Strobel and others, 1999). Location of section is shown in figure 14. Abbreviations for stratigraphic intervals are explained in figure 9.

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).

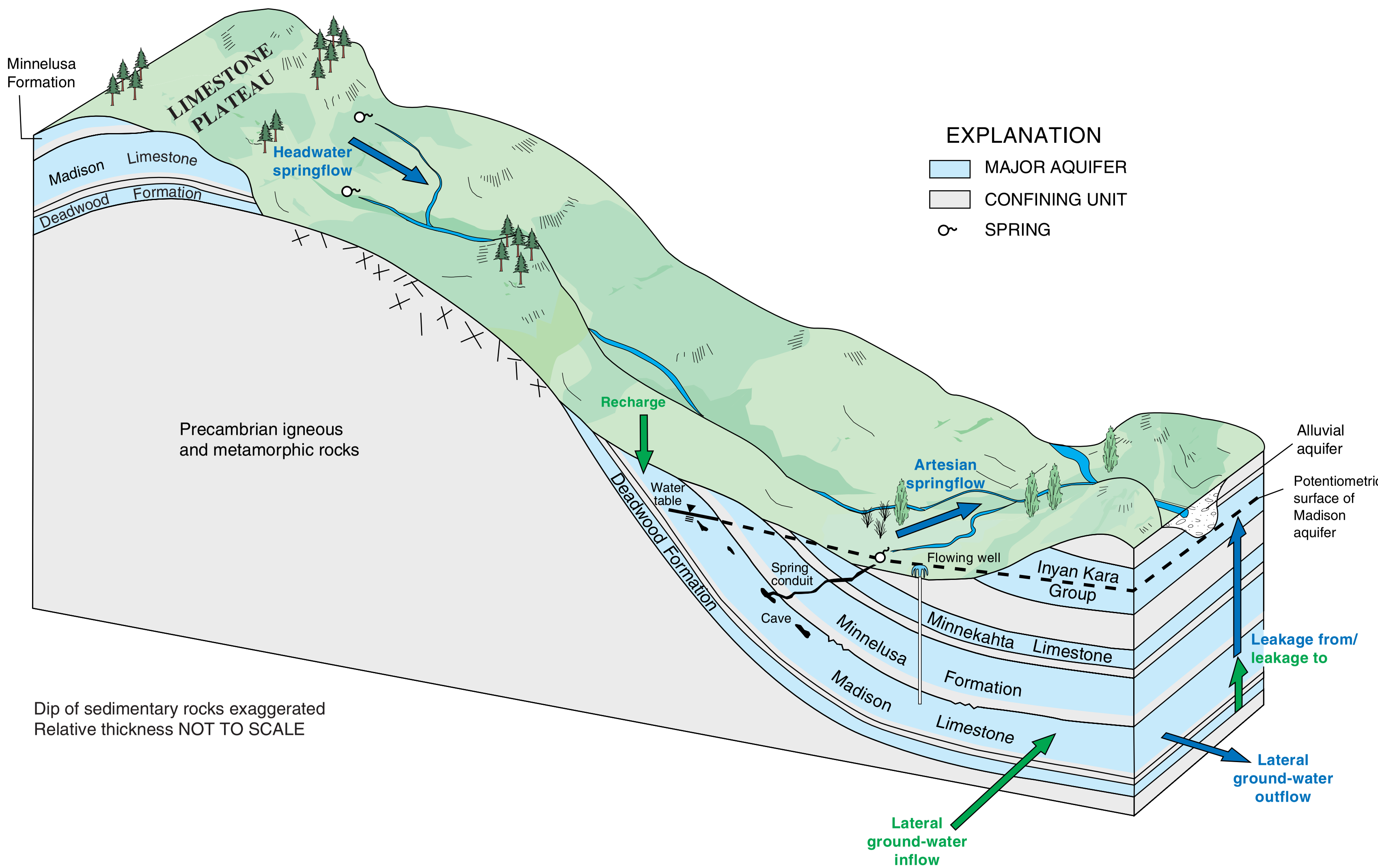


Figure 16. Schematic showing simplified hydrogeologic setting of the Black Hills area. Schematic generally corresponds with geologic cross section shown in figure 15. Components considered for hydrologic budget of the Madison aquifer also are shown with inflow components shown in green and outflow components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

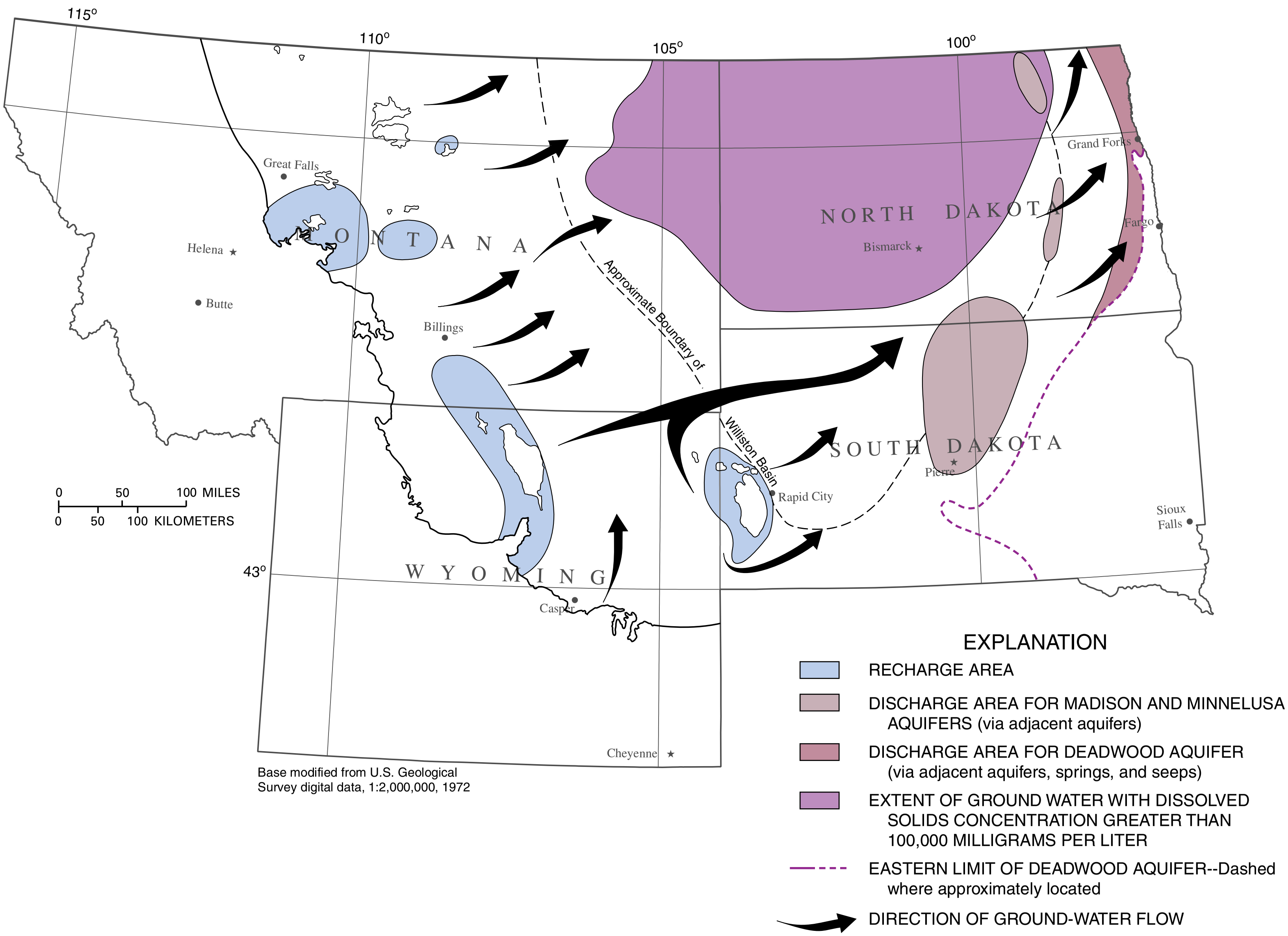


Figure 17. General direction of ground-water flow in regional aquifer system within Paleozoic aquifer units (modified from Downey and Dinwiddie, 1988; Whitehead, 1996).

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.

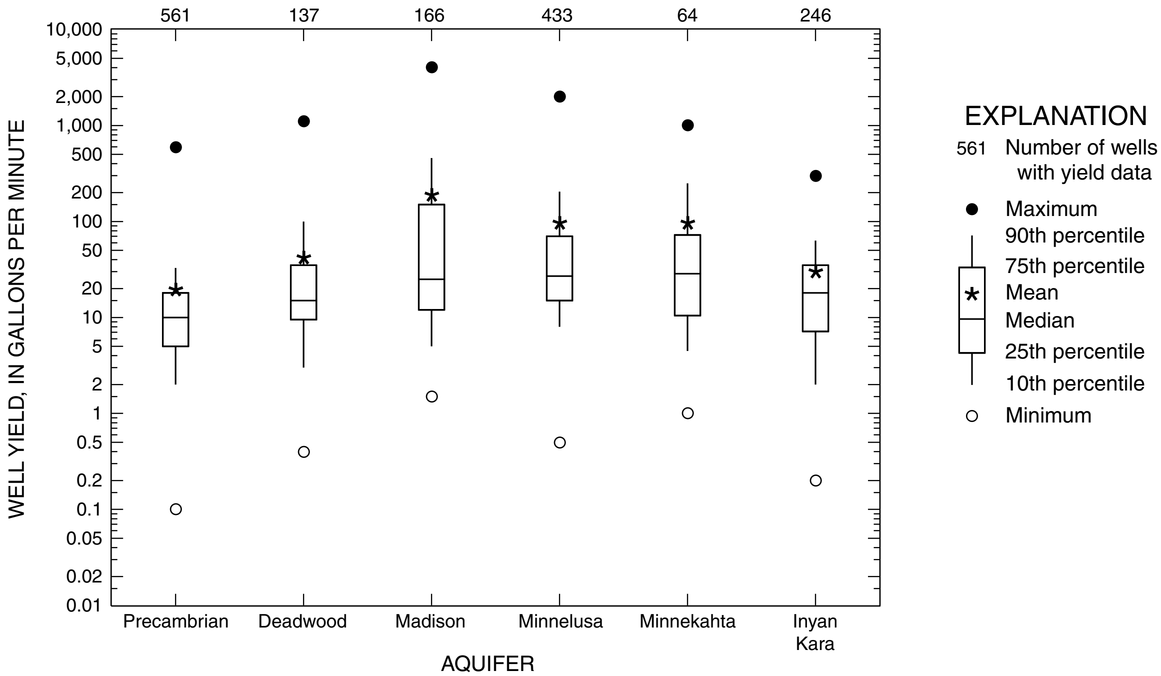


Figure 18. Boxplots showing distribution of well yields from selected aquifers (data obtained from U.S. Geological Survey Ground Water Site Inventory database).

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	2.4×10^{-5} - 1.9	--	--	--	Crook County, Wyoming
Woodward-Clyde Consultants, 1980	--	3,000	2×10^{-4} - 3×10^{-4}	--	Eastern Wyoming, western South Dakota
Blankennagel and others, 1981	--	5,090	2×10^{-5}	--	Yellowstone County, Montana
Downey, 1984	--	250 - 3,500	--	--	Montana, North Dakota, South Dakota, Wyoming
Plummer and others, 1990	--	--	1.12×10^{-6} - 3×10^{-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	9.0×10^{-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	3×10^{-4} - 1×10^{-3}	--	Spearfish area
Carter, Driscoll, Hamade, and Jarrell, 2001	--	100 - 7,400	--	--	Black Hills area
Minnelusa aquifer					
Blankennagel and others, 1977	$<2.4 \times 10^{-5}$ - 1.4	--	--	--	Crook County, Wyoming
Pakkong, 1979	--	880	--	--	Boulder Park area, South Dakota
Woodward-Clyde Consultants, 1980	--	30 - 300	6.6×10^{-5} - 2.0×10^{-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	5.0×10^{-9} - 7.4×10^{-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	2.1×10^{-5} - 2.5×10^{-5}	--	Wall area, South Dakota
Boggs and Jenkins, 1980	--	50 - 190	1.4×10^{-5} - 1.0×10^{-4}	--	Northwestern Fall River County
Bredehoeft and others, 1983	8.3	--	1.0×10^{-5}	--	South Dakota
Rahn, 1985	--	--	--	0.26/0.17	Western South Dakota
Kyllonen and Peter, 1987	--	0.86 - 6,000	--	--	Northern Black Hills

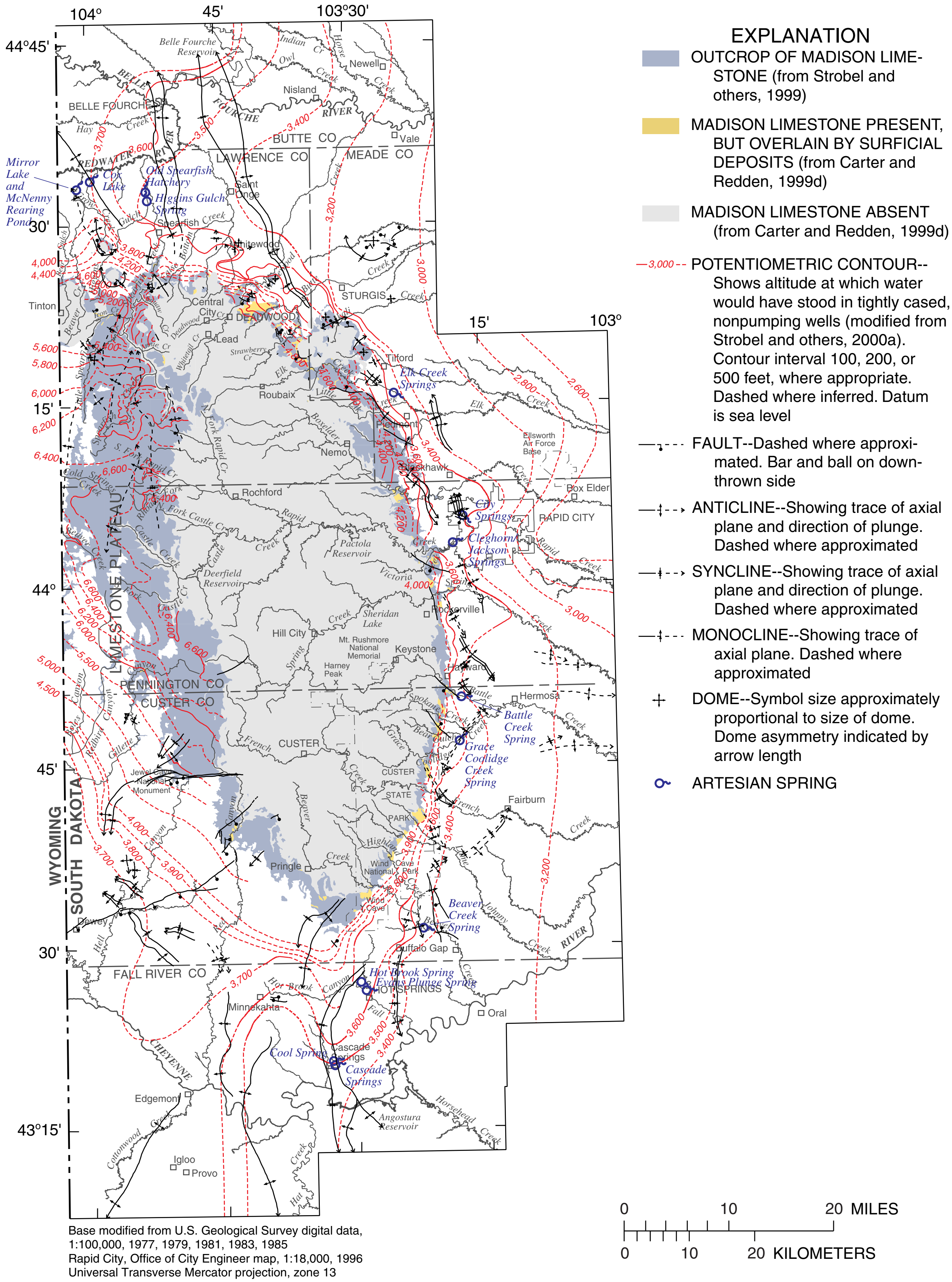


Figure 19. Potentiometric surface of the Madison aquifer and locations of major artesian springs.

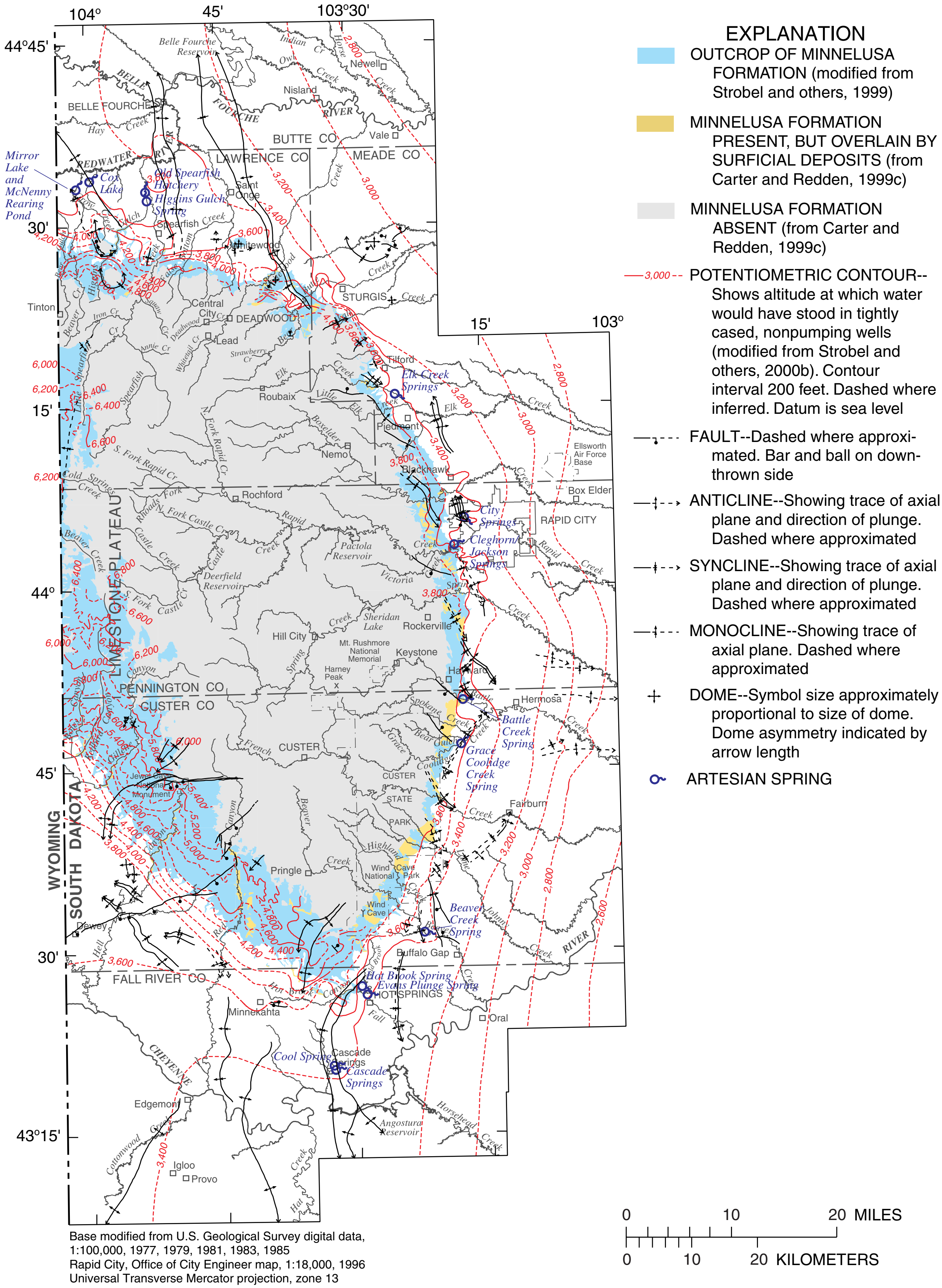


Figure 20. Potentiometric surface of the Minnelusa aquifer and locations of major artesian springs.

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).

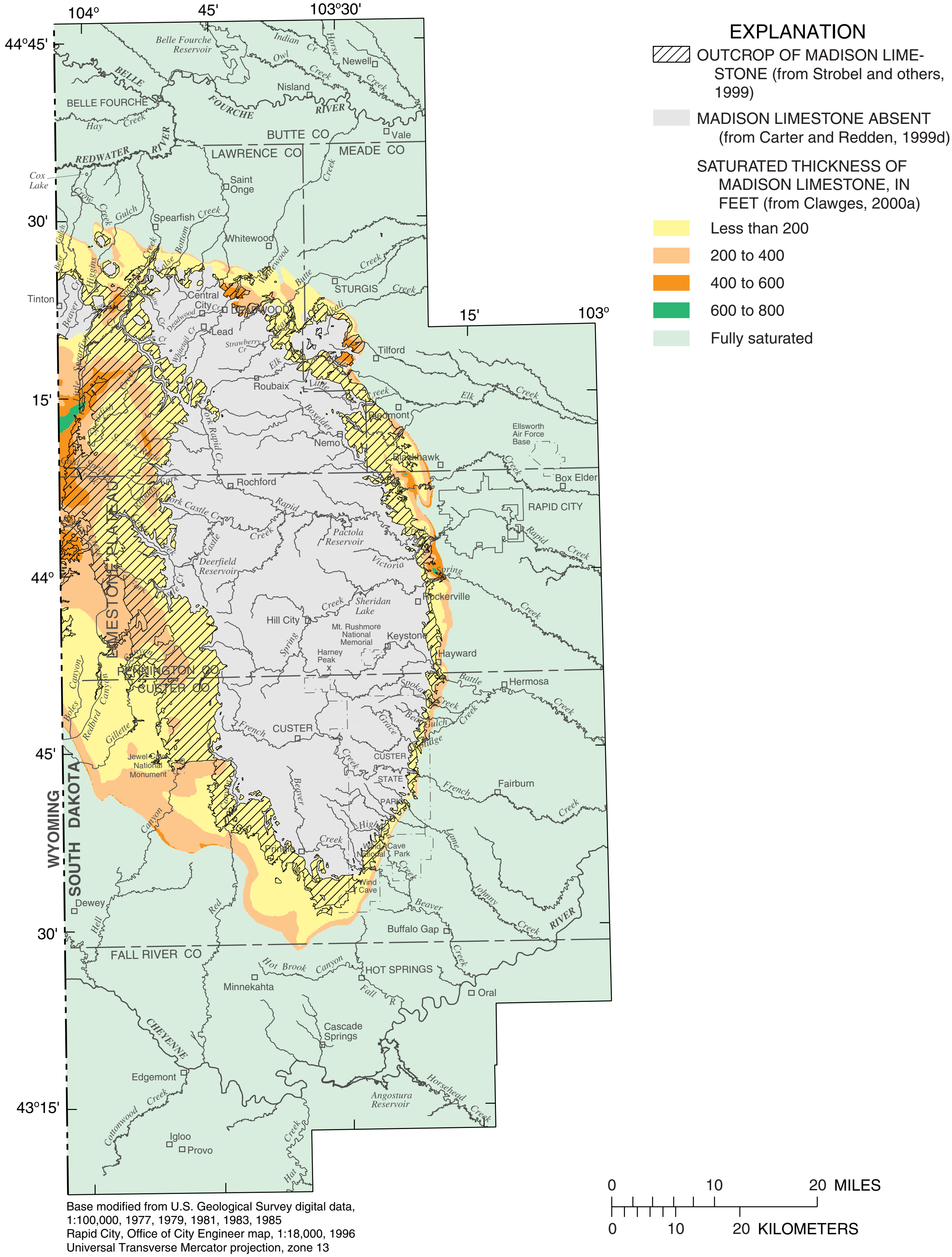


Figure 21. Saturated thickness of the Madison aquifer.

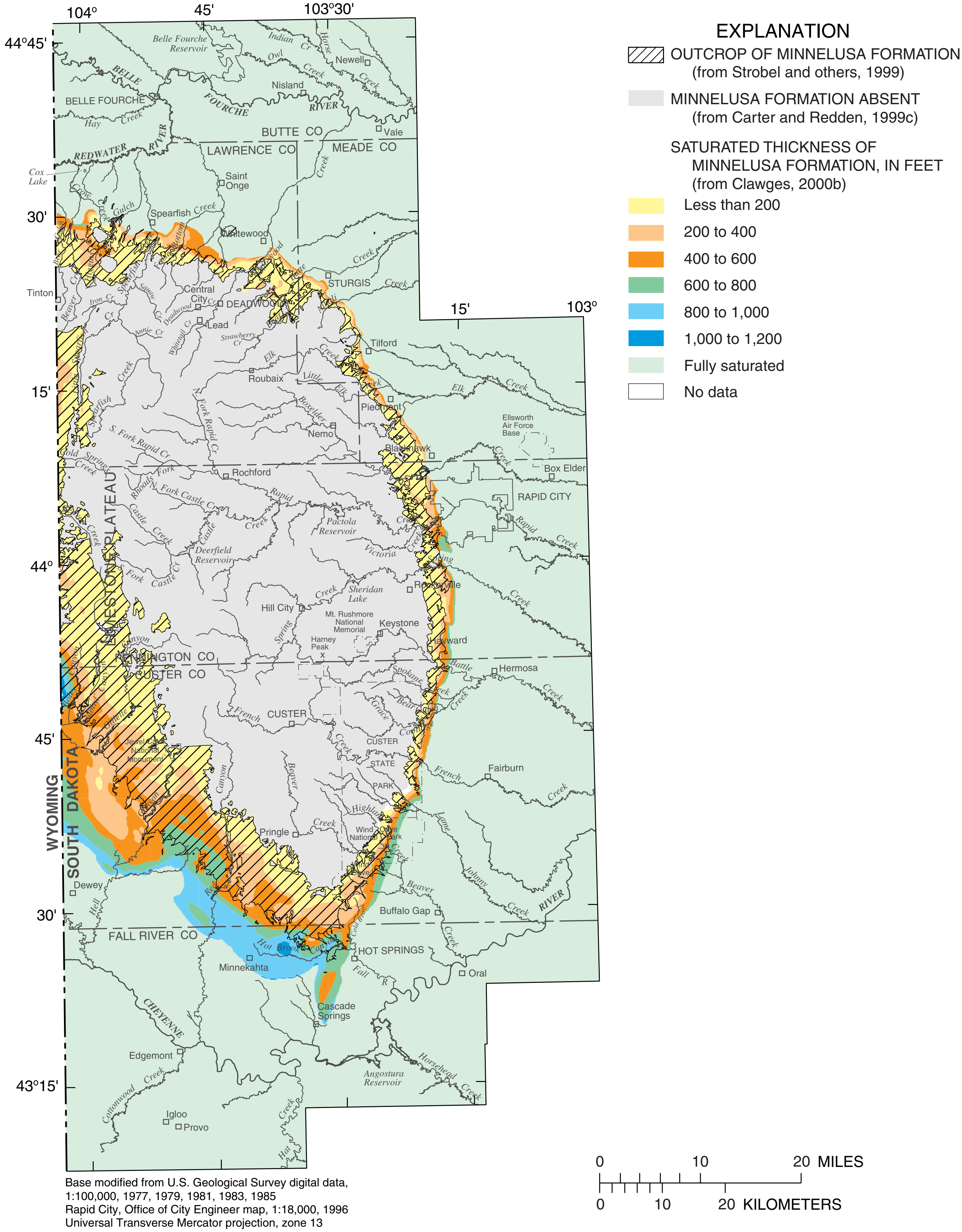


Figure 22. Saturated thickness of the Minnelusa aquifer.

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.

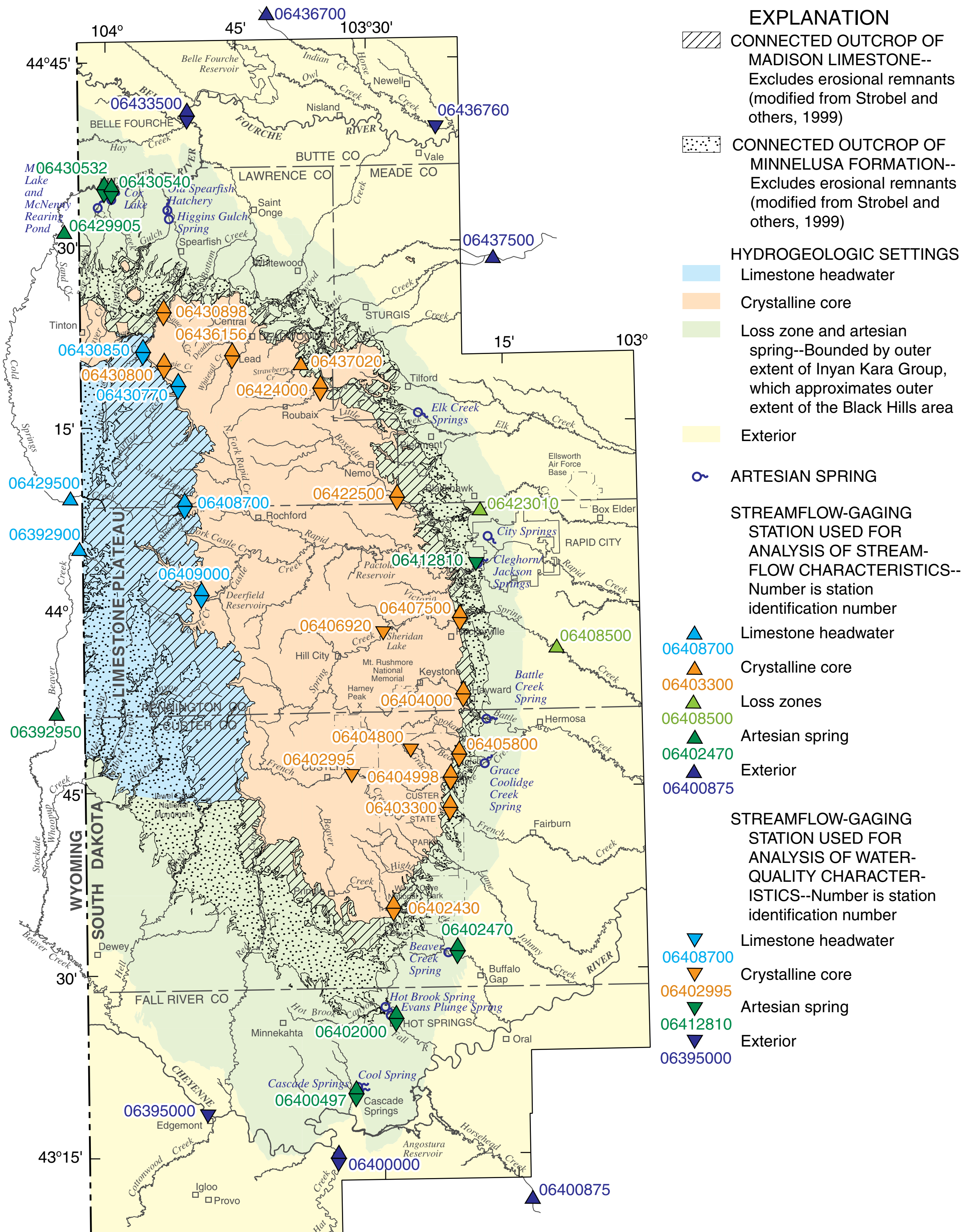
The crystalline core setting consists primarily of igneous and metamorphic Precambrian rocks within the central part of the Black Hills, but also includes numerous Tertiary intrusives in the northern Black Hills (fig. 14). Unconsolidated Quaternary and Tertiary deposits also occur in various locations. 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).



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 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.

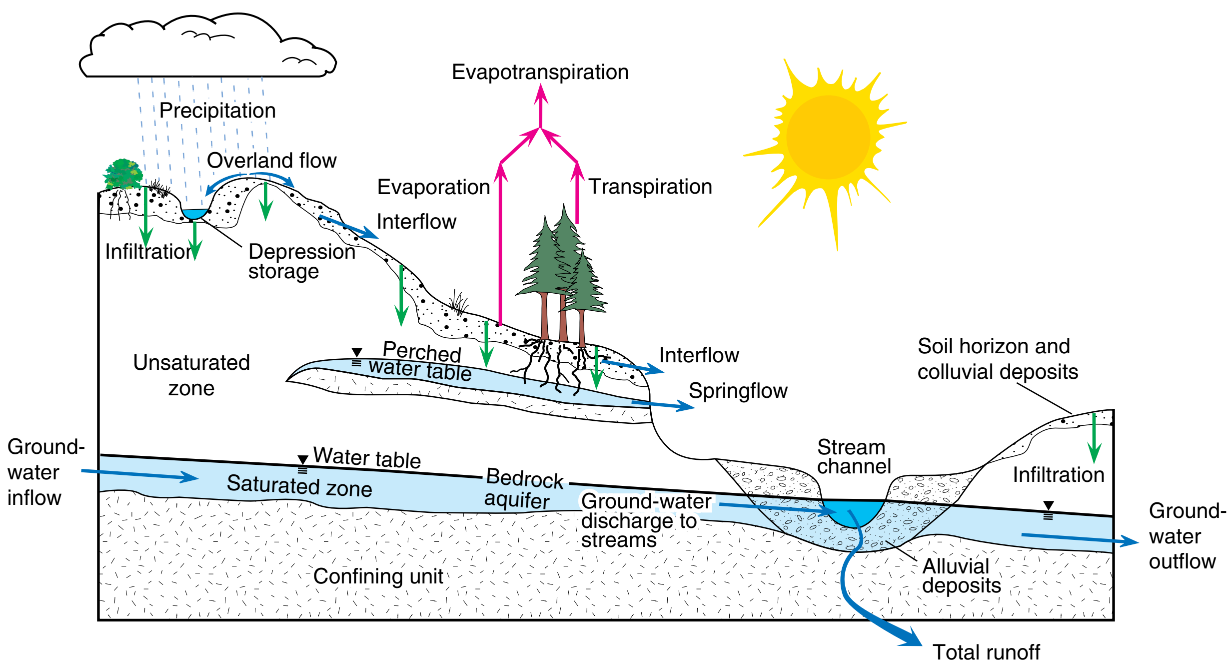


Figure 24. Schematic diagram illustrating hydrologic processes (modified from Driscoll and Carter, 2001).

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.

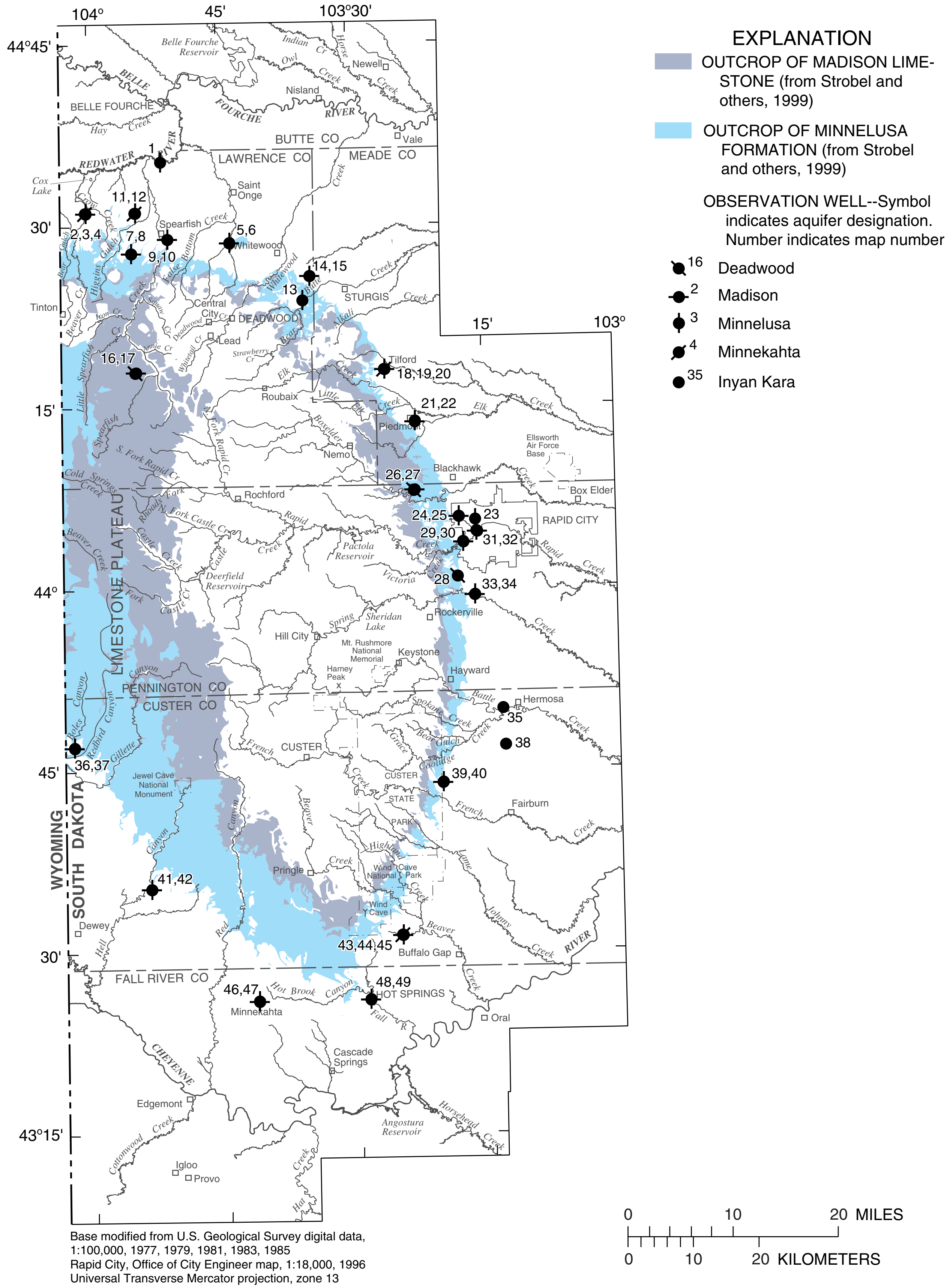


Figure 25. Location of observation wells for which hydrographs are presented.

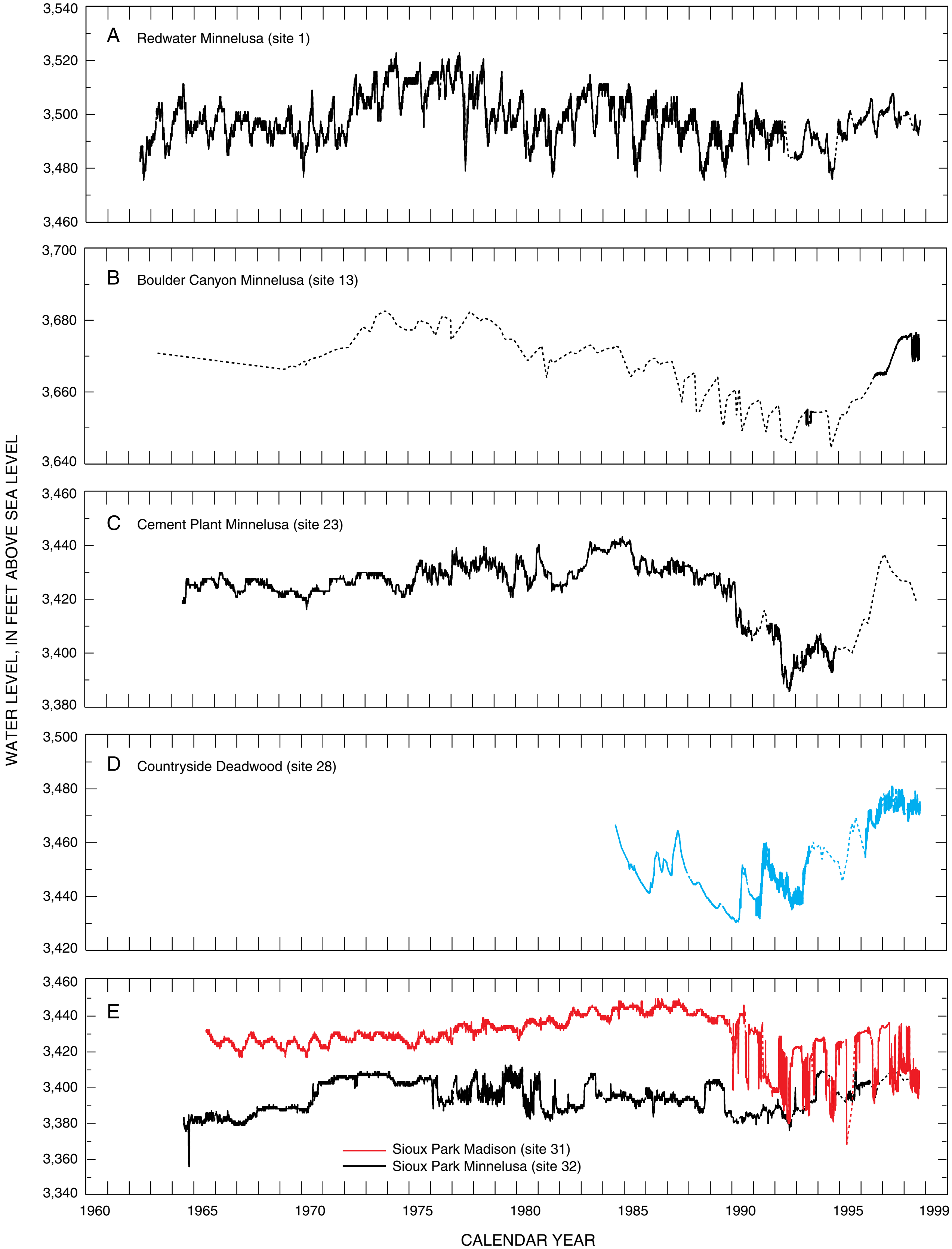


Figure 26. Hydrographs illustrating temporal trends in ground-water levels.

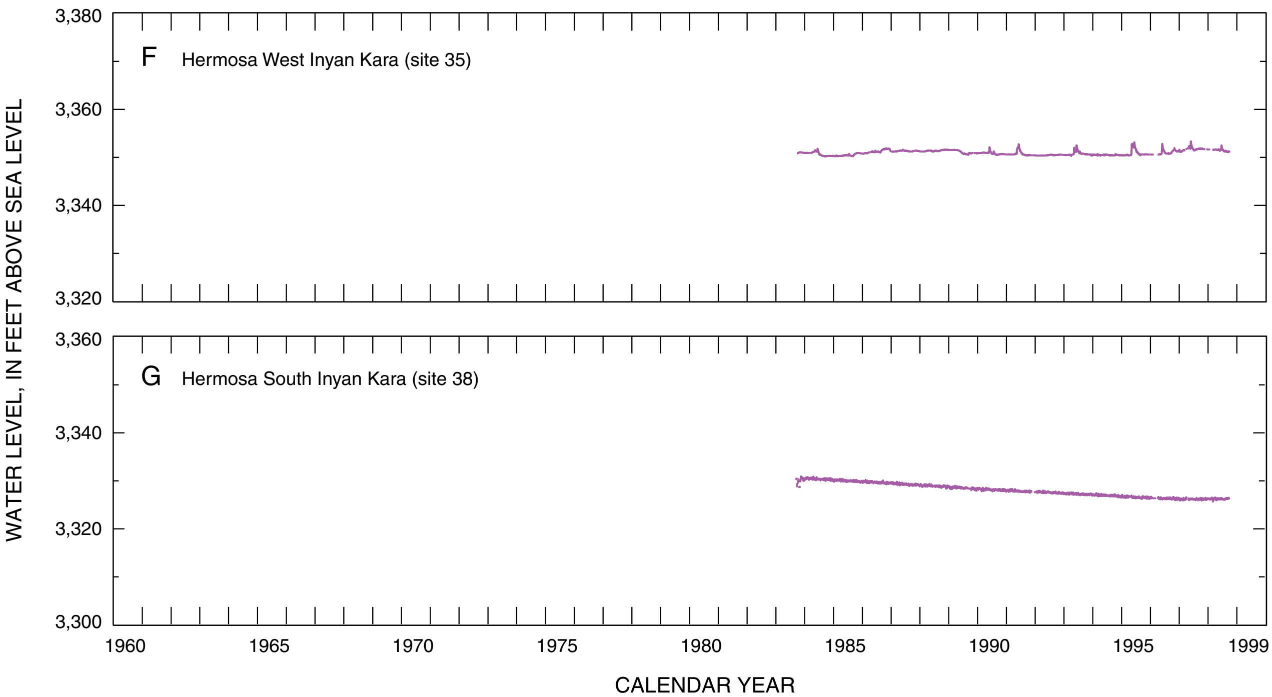


Figure 26. Hydrographs illustrating temporal trends in ground-water levels.—Continued

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.

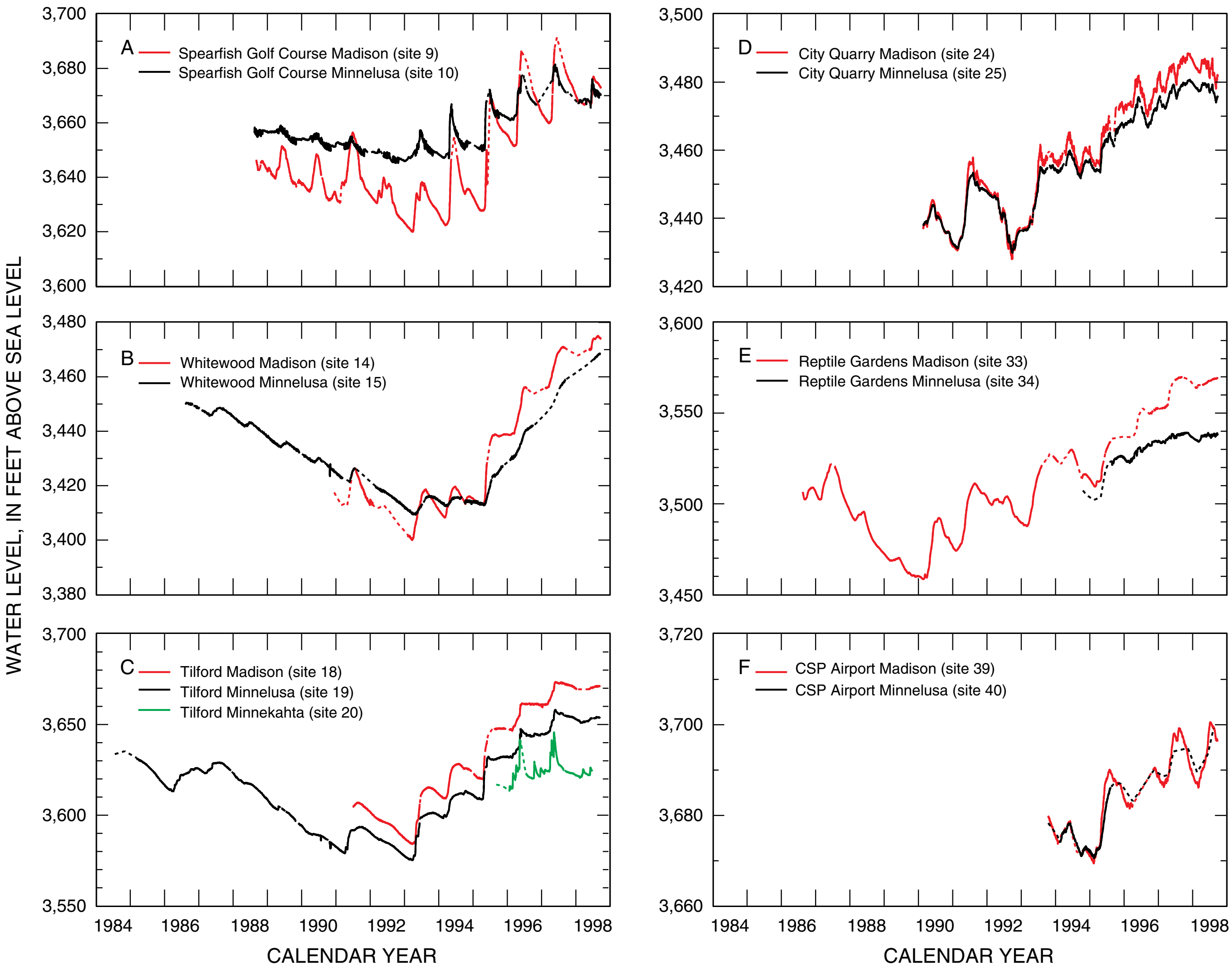


Figure 27. Hydrographs illustrating general similarities in water levels for some colocated Madison/Minnelusa wells with confined conditions.

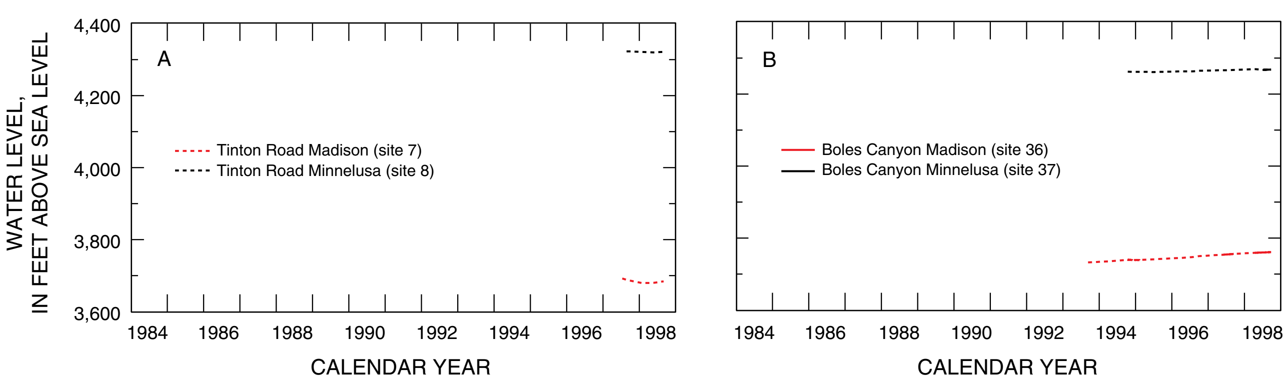


Figure 28. Hydrographs illustrating distinct hydraulic separation for two Madison/Minnelusa well pairs with unconfined conditions.

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.

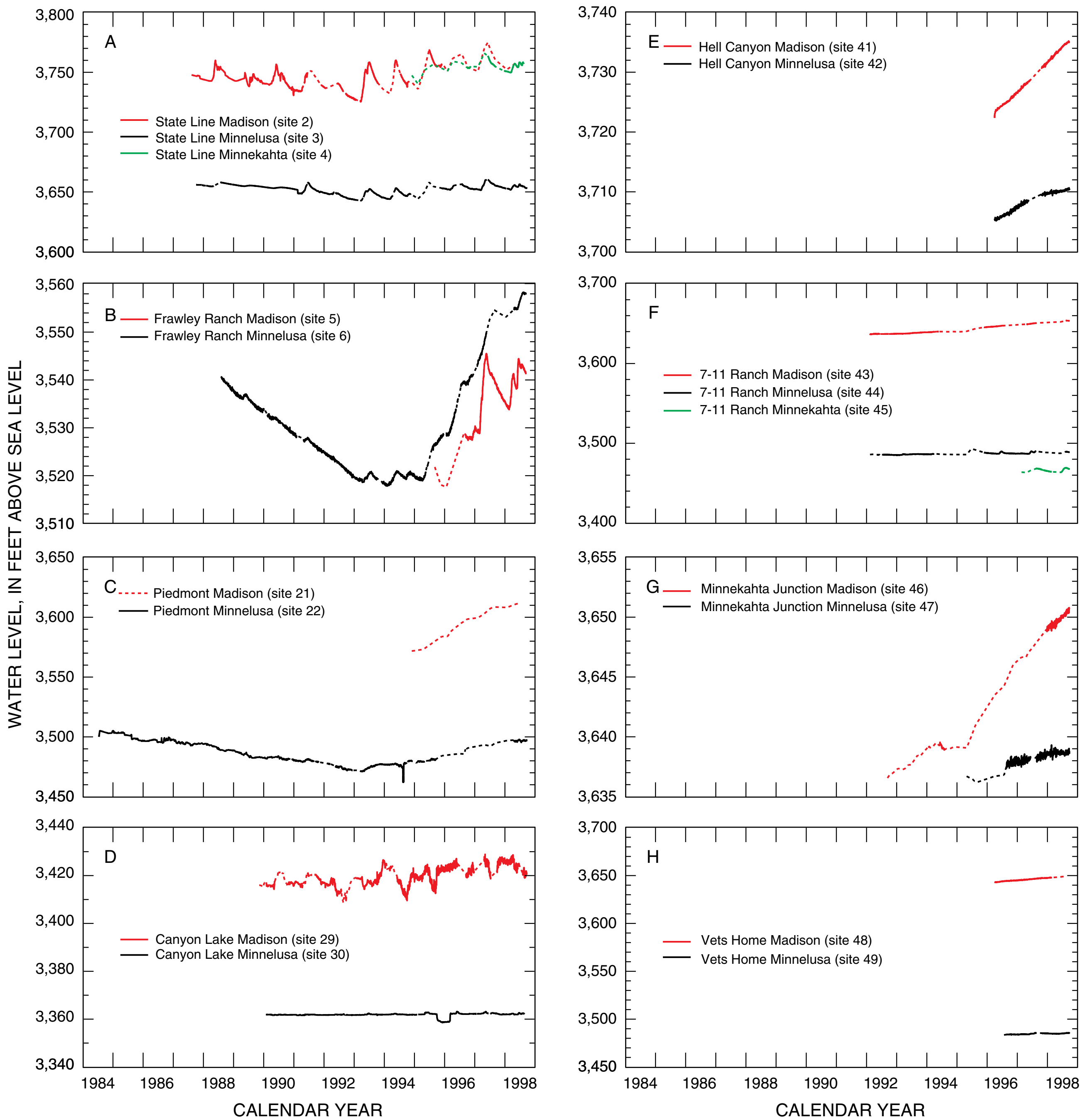


Figure 29. Hydrographs illustrating generally separated water levels for some colocated Madison/Minnelusa wells.

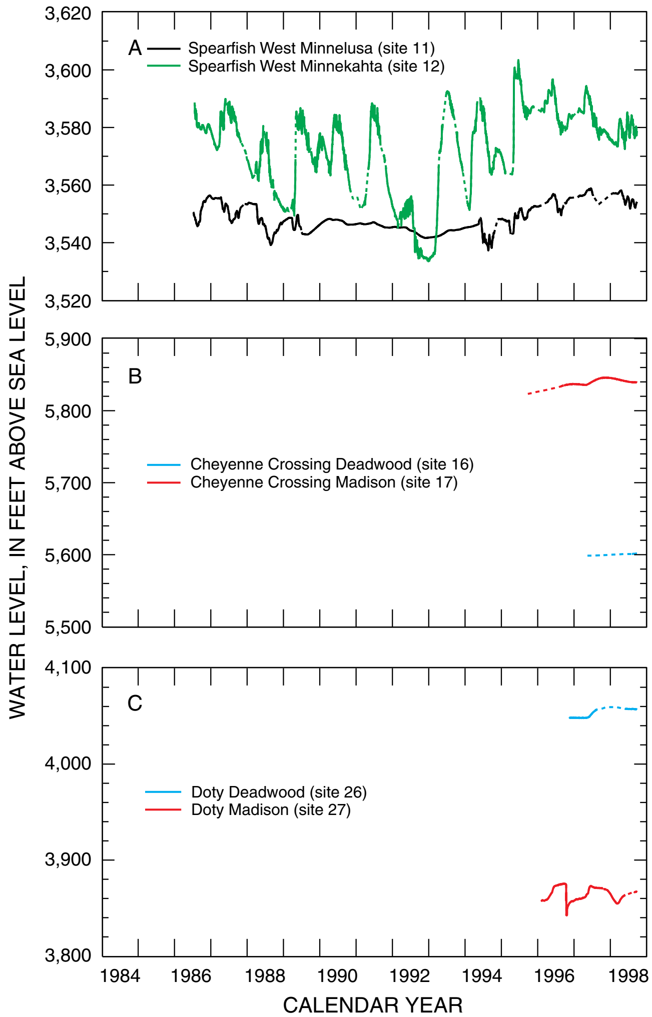


Figure 30. Hydrographs for colocated Minnelusa/Minnekahta and Deadwood/Madison wells.

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.