Geologic Setting

The oldest geologic units in the study area are the Precambrian crystalline (metamorphic and igneous) rocks (fig. 2), which form a basement under the Paleozoic, Mesozoic, and Cenozoic rocks and sediments. The 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, but are composed mostly of igneous rocks or metasedimentary rocks, such as schists and graywackes. The Paleozoic and Mesozoic rocks were deposited as nearly horizontal beds. Subsequent uplift during the Laramide orogeny and related erosion exposed the Precambrian rocks in the crystalline core of the Black Hills, with the Paleozoic and Mesozoic sedimentary rocks exposed in roughly concentric rings around the core. Deformation during the Laramide orogeny contributed to the numerous fractures, folds, and other features present throughout the Black Hills. Tertiary intrusive activity also contributed to rock fracturing in the northern Black Hills where numerous intrusions exist.

Surrounding the crystalline core is a layered series of sedimentary rocks (fig. 3) including 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 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) (fig. 4). Following are descriptions for Paleozoic bedrock formations in the Black Hills, which includes the Madison Limestone, Minnelusa Formation, and stratigraphically adjacent units.

The oldest sedimentary formation 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 the 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 above 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). Local variations in thickness are due largely to the karst topography that developed before the deposition of the overlying formations (DeWitt and others, 1986). Because the Madison Limestone was exposed to erosion and karstification for millions of years, the formation is unconformably overlain by the Minnelusa Formation.

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 lower 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 near the outcrop areas, forming collapse features filled with breccia (Braddock, 1963). The thickness of the Minnelusa Formation in the study area increases from north to south and ranges from 375 ft near Belle Fourche to 1,175 ft near Edgemont (Carter and Redden, 1999c). Along the northeastern part of the central Black Hills, there is little anhydrite in the subsurface due to a change in the depositional environment. On the south and southwest side of the study area, there is a considerable increase in thickness of clastic units as well as a thick section of anhydrite. 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.

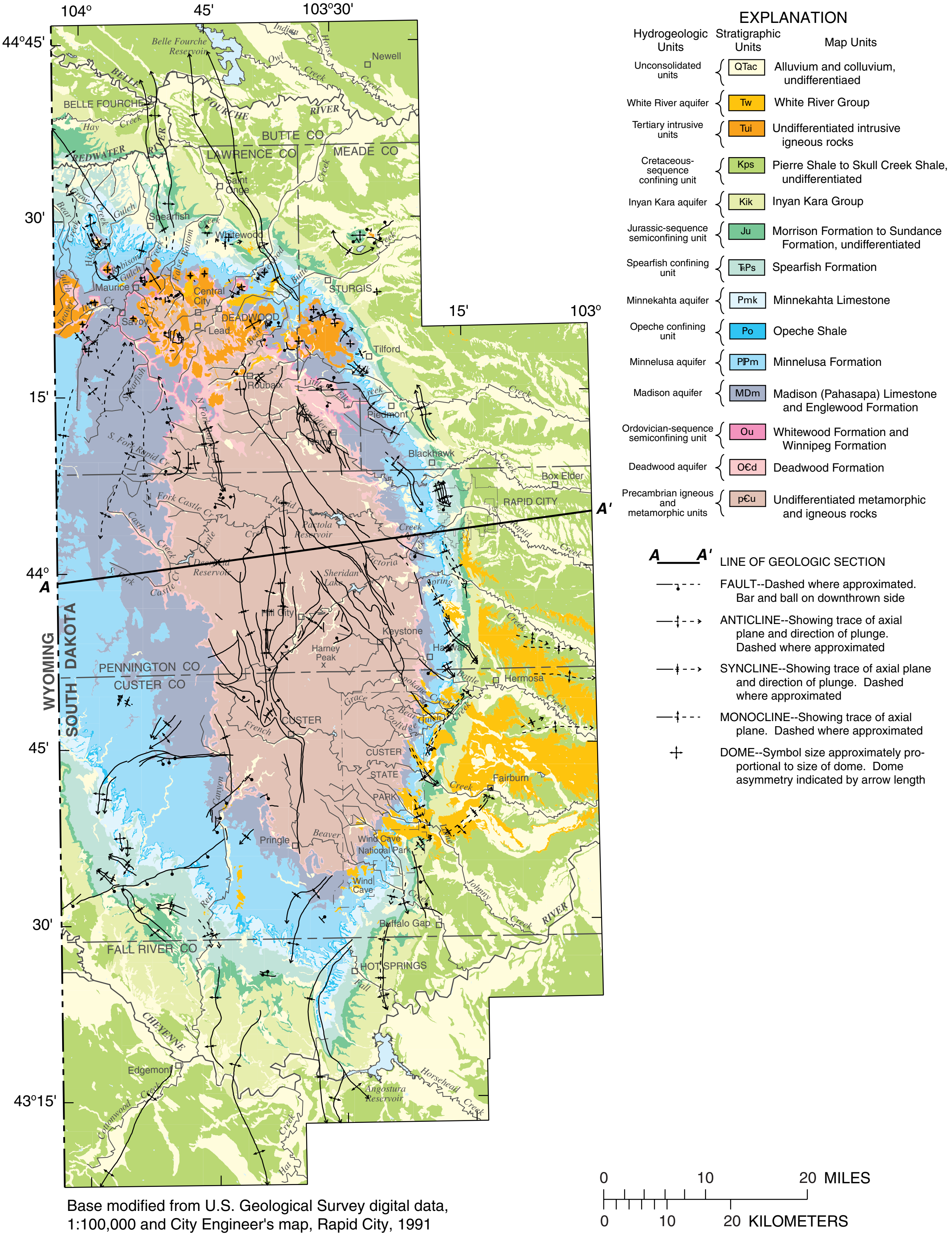
ERATHEM	SYSTEM	ABBREVIATION FOR STRATIGRAPHIC INTERVAL	GEOLOGIC UNIT			THICKNESS IN FEET	DESCRIPTION
CENOZOIC	QUATERNARY & TERTIARY (?)	QTu		UNDIFFERENTIATED SANDS AND GRAVELS		0-50	Sand, gravel, and boulders
		Tw		WHITE RIVER GROUP		0-300	Light colored clays with sandstone channel fillings and local limestone lenses.
	TERTIARY[1]	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		[2]80-300	Impure chalk and calcareous shale.
				CARLILE SHALE	Turner Sandy Member Wall Creek Member	[2]350-750	Light-gray shale with numerous large concretions and sandy layers. Dark-gray shale
				GREENHORN FORMATION		(25-30) (200-350)	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.
				MOWRY SHALE		125-230	Light-gray siliceous shale. Fish scales and thin layers of bentonite.
				MUDDY SANDSTONE	NEWCASTLE	20-150	Brown to light yellow and white sandstone.
				SKULL CREEK SHALE		150-270	Dark gray to black siliceous shale.
		Kik	INYAN KARA GROUP	FALL RIVER FORMATION		10-200	Massive to slabby sandstone.
				LAKOTA FM	Fuson Shale Minnewaste Limestone Chilson Member	10-190 0-25 25-485	Coarse gray to buff cross-bedded conglomeratic sandstone, interbedded with buff, red, and gray clay, especially toward top. Local fine-grained limestone.
	JURASSIC	Ju		MORRISON FORMATION		0-220	Green to maroon shale. Thin sandstone.
				UNKPAPA SS		0-225	Massive fine-grained sandstone.
				SUNDANCE FORMATION	Redwater Member Lak Member Hulett Member Stockade Beaver Mem. Canyon Spr Member	250-450	Greenish-gray shale, thin limestone lenses. Glauconltic sandstone; red sandstone near middle.
				GYPSUM SPRING FORMATION		0-45	Red siltstone, gypsum, and limestone.
	TRIASSIC	TRPs		SPEARFISH FORMATION	Goose Egg Equivalent	250-800	Red sandy shale, soft red sandstone and siltstone with gypsum and thin limestone layers. Gypsum locally near the base.
PALEOZOIC	PERMIAN	Pmk		MINNEKAHTA LIMESTONE		[2]25-65	Thin to medium-bedded fine-grained, purplish gray laminated limestone.
		Po		OPECHE SHALE		[2]25-150	Red shale and sandstone.
	PENNSYLVANIAN	PIPm		MINNELUSA FORMATION		[2]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	MDm		MADISON (PAHASAPA) LIMESTONE		[2]250-1,000	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		[2]0-235	Buff dolomite and limestone.
				WINNIPEG FORMATION		[2]0-150	Green shale with siltstone.
	CAMBRIAN	OЄd		DEADWOOD FORMATION		[2]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 METAMORPHIC AND IGNEOUS ROCKS			Schist, slate, quartzite, and arkosic grit. Intruded by diorite, metamorphosed to amphibolite, and by granite and pegmatite.

[1] Also may include intrusive igneous rocks

[2] Modified based on drill-hole data

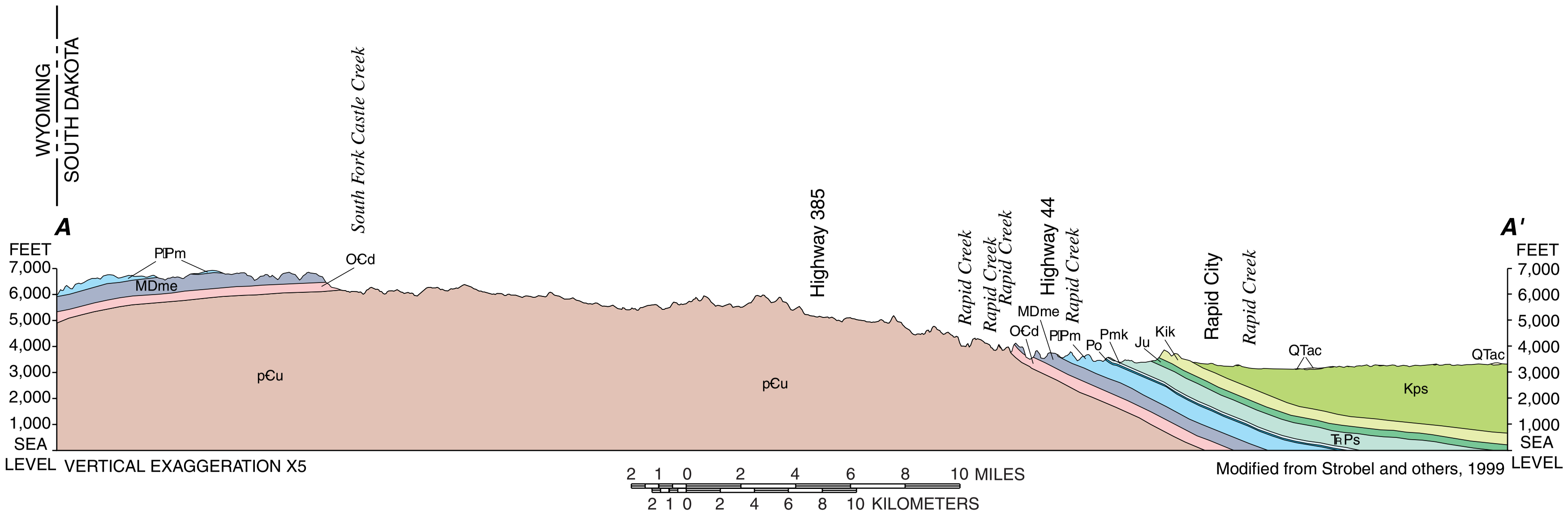
Modified from information furnished by the Department of Geology and Geological Engineering, South Dakota School of Mines and Technology (written commun., January 1994)

Figure 2. Stratigraphic section for the Black Hills.



Base modified from U.S. Geological Survey digital data, 1:100,000 and City Engineer's map, Rapid City, 1991

Figure 3. Distribution of hydrogeologic units in the Black Hills area (modified from Strobel and others, 1999).



Modified from Strobel and others, 1999

Figure 4. Geologic section A-A' (Location of section is shown in figure 3. Abbreviations for stratigraphic intervals are explained in figure 2.).

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

Hydrologic Setting

The Precambrian basement rocks generally have low permeability and form the lower confining unit for the series of sedimentary aquifers in the Black Hills area. Localized aquifers occur in Precambrian rocks in many locations in the central core of the Black Hills, where enhanced secondary permeability results from weathering and fracturing. In these aquifers, water-table (unconfined) conditions generally prevail and land-surface topography can strongly control ground-water flow directions. Many wells completed in the Precambrian rocks are located along stream channels.

Many of the sedimentary formations 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 units or by relatively impermeable layers within the individual formations. Extremely variable leakage can occur between these aquifers (Peter, 1985; Greene, 1993).

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 the 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. Where present, the Whitewood and Winnipeg Formations act as a semi-confining unit overlying the Deadwood aquifer (Strobel and others, 1999). These units 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 Strobel and others (1999) included as part of the Madison aquifer.

The Madison aquifer generally occurs within the karstic upper part of the Madison Limestone; however, Strobel and others (1999) included the entire Madison Limestone and the Englewood Formation in their delineation of the aquifer. Numerous fractures and solution openings in the Madison Limestone provide extensive secondary porosity in the aquifer. The Madison aquifer receives significant recharge from streamflow losses and precipitation on the outcrop. The Madison aquifer is confined by low permeability layers in the overlying Minnelusa Formation.

The Minnelusa aquifer occurs within the thin layers of sandstone, dolomite, and anhydrite in the lower portion of the Minnelusa Formation and sandstone and gypsum in the upper portion. The Minnelusa aquifer has primary porosity in the sandstone units and secondary porosity from fracturing and collapse breccia associated with dissolution of interbedded evaporites. 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, which is preferentially recharged because of its upgradient location. The Minnelusa aquifer is confined by the overlying Opeche Shale.

The Minnekahta aquifer, which overlies the Opeche Shale, typically is very permeable, but is limited in amount of yield by the aquifer thickness. The Minnekahta aquifer receives significant recharge from precipitation and limited recharge from streamflow losses on the outcrop. The overlying Spearfish Formation acts as a confining unit to the aquifer.

Within the Mesozoic rock interval, the Inyan Kara aquifer is used extensively. Aquifers in various other formations 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 aquifers (Swenson, 1968; Gott and others, 1974). As much as 4,000 ft of Cretaceous shales act as the upper confining layer to aquifers in the Mesozoic rock interval.

Artesian (confined) conditions generally exist within the aforementioned aquifers, where an upper confining layer is present. Under artesian 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 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. The hydrogeologic setting of the Black Hills area is schematically illustrated in figure 5.

Streamflow within the study area is affected by both topography and geology. The base flow of most streams in the Black Hills originates in the higher elevations, where relatively large precipitation and small evapotranspiration result in more water being available for springflow and streamflow. Numerous streams have significant headwater springs originating from the Paleozoic carbonate rocks along the "Limestone Plateau" (fig. 1) on the western side of the study area. This area is a large discharge zone for aquifers in the Paleozoic rock interval, especially for the Madison aquifer. The headwater springs provide significant base flow for several streams that flow across the crystalline core.

Most streams generally lose all or part of their flow as they cross the outcrop 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, and limited losses probably also occur within the outcrop of the Minnekahta Limestone (Hortness and Driscoll, 1998). 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).

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

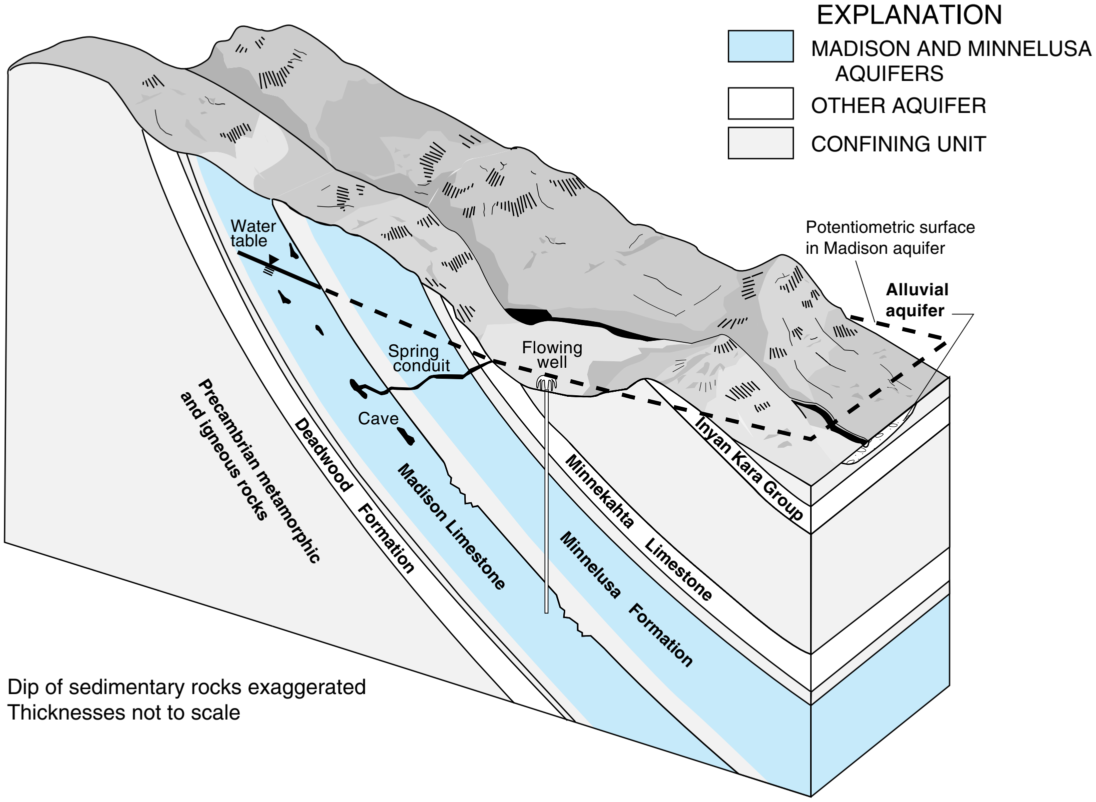


Figure 5. Schematic showing simplified hydrogeologic setting of the Black Hills area.

Department of Environment and Natural Resources has provided support and extensive technical assistance to the study. In addition, the authors acknowledge the technical assistance from many faculty and students at the South Dakota School of Mines and Technology.

RECHARGE PROCESSES AND GENERAL METHODS FOR QUANTIFYING RECHARGE

This section describes processes affecting recharge to the Madison and Minnelusa aquifers and provides an overview of the general methods used to quantify recharge. An overview of previous investigations regarding recharge to the Madison and Minnelusa aquifers also is provided.

Previous Investigations

Numerous previous investigators have studied recharge to the Madison and Minnelusa aquifers. Most of the previous investigations have focused on streamflow losses. Losses from local Black Hills streams to outcrops of various sedimentary formations were first noted by Dodge (1876), although it was then believed that most losses occurred to the Minnelusa Formation and overlying sandstone units (Newton and Jenney, 1880). Streamflow losses for various Black Hills streams were estimated by Brown (1944), Crooks (1968), Rahn and Gries (1973), Peter (1985), and Greene (1997). The most comprehensive study of streamflow losses in the Black Hills area was by Hortness and Driscoll (1998), who documented losses for 24 streams based on extensive measurements and analyses of streamflow records.

Cox (1962) estimated recharge for the Minnelusa aquifer in the northern Black Hills as 2 inches from infiltration of precipitation. Minimum precipitation recharge for the Madison and Minnelusa aquifers was estimated by Rahn and Gries (1973) to range from 0.6 in/yr in the southern Black Hills to 6.8 in/yr in the northern Black Hills. Peter (1985) estimated that between 1 and 2 inches of the annual precipitation becomes recharge to the Madison and Minnelusa aquifers in the Rapid City area. Annual recharge to the Madison aquifer on the western flanks of the Black Hills in the Limestone Plateau area was estimated to be 6.8 inches (Downey, 1986).

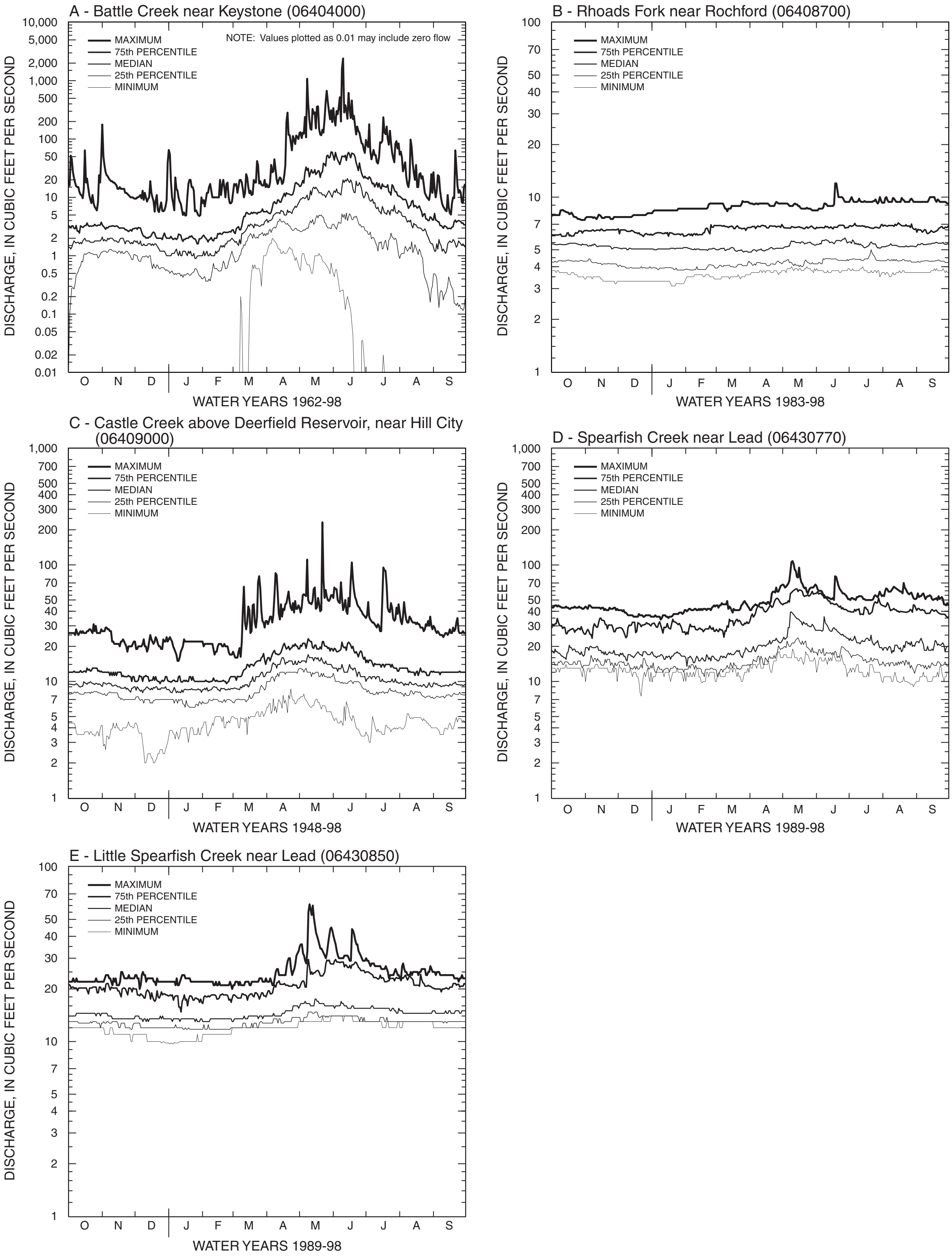
Recharge Processes

As discussed, many previous investigations have addressed quantification of streamflow loss rates. These investigations have provided various insights regarding the processes affecting recharge to the Madison and Minnelusa aquifers. One very important factor is the potential for extremely large secondary porosity within these aquifers, which is evidenced by the large infiltration rates that are associated with dramatic streamflow losses that can be as large as tens of cubic feet per second for some stream reaches (Hortness and Driscoll, 1998). Large secondary porosity and associated infiltration rates also are consistent with the physical nature of both formations, which commonly have fractures and solution features in outcrop sections. The Madison Limestone is especially prone to solution openings, as exemplified by large caves such as Wind Cave and Jewel Cave, which are two of the largest caves in the world.

The fact that both the Madison and Minnelusa aquifers have large secondary porosity in some locations does not necessarily imply that infiltration rates will be uniformly large in all outcrop sections. Both aquifers are prone to large heterogeneity, or variability in aquifer characteristics (Cox, 1962; Greene, 1993; Greene and Rahn, 1995), as evidenced by the extremely large range in well yields that can occur. This is visually apparent in many locations in caves within the Madison Limestone, where rates of cave drip can be very small in the ceilings of man-size passageways (Wiles, 1992).

Rates of recharge resulting from infiltration of precipitation on outcrops can be highly affected by conditions in the soil horizon. Much of the precipitation that occurs is eventually returned to the atmosphere though evaporation and transpiration (evapotranspiration). Recharge can occur only when water infiltrates to sufficient depth to escape the root zone. Thus, recharge rates can be affected by infiltration rates, along with thicknesses and associated storage capacities of overlying soils, which can be highly variable.

A perspective on the infiltration capacity of the Madison and Minnelusa aquifers on a watershed scale can be obtained by examination of streamflow information for selected gaging stations. Duration hydrographs are presented in figure 6 for four streamflow-gaging stations (graphs B through E) that are located in or near the Limestone Plateau area, which is dominated by large outcrop areas of the Madison Limestone and


A - Battle Creek near Keystone (06404000)

B - Rhoads Fork near Rochford (06408700)

C - Castle Creek above Deerfield Reservoir, near Hill City (06409000)

D - Spearfish Creek near Lead (06430770)

E - Little Spearfish Creek near Lead (06430850)

Figure 6. Daily-duration hydrographs for selected gaging stations.

Minnelusa Formation (fig. 1). Locations of gaging stations are shown in figure 7. Flow at these sites is dominated by base flow originating from ground-water discharge from the Madison and Minnelusa aquifers. For comparison, a duration hydrograph also is presented for a gaging station on Battle Creek (graph A in fig. 6), the drainage area of which is dominated by Precambrian igneous and metamorphic rocks. Flow in Battle Creek is highly variable and responsive to short-term climatic conditions, indicating dominance from surface-water flow components relative to ground-water flow components. Additional discussions of differences in flow characteristics for different hydrogeologic settings were presented by Miller and Driscoll (1998).

An important observation from examination of the duration hydrographs is that direct surface runoff from outcrops of the Madison Limestone and Minnelusa Formation is very unusual. Surface runoff is virtually nonexistent for Rhoads Fork (graph B), for which the surface drainage area is comprised almost entirely of Madison Limestone outcrops. The entire range in variability in daily flow for this site falls easily within one order of magnitude, compared with a range spanning in excess of five orders of magnitude for Battle Creek. Increasingly larger components of surface runoff are apparent for graphs E, D, and C, respectively, which can be attributed to increasingly larger percentages of outcrops other than the Madison Limestone and Minnelusa Formation within these drainage basins (figs. 3 and 7).

The preceding discussions are used as the basis of an assumption that direct surface runoff from the Madison Limestone and Minnelusa Formation is almost nonexistent and can be neglected for many purposes associated with calculation of recharge to these aquifers. This assumption is very important in developing methods for quantification of recharge from direct precipitation, as discussed in the following section.

General Methods for Quantifying Recharge

Quantifying recharge to the Madison and Minnelusa aquifers requires methods for quantification of both streamflow recharge and precipitation recharge, as discussed in this section. Various considerations regarding areas and uncertainties associated with recharge estimates also are discussed.

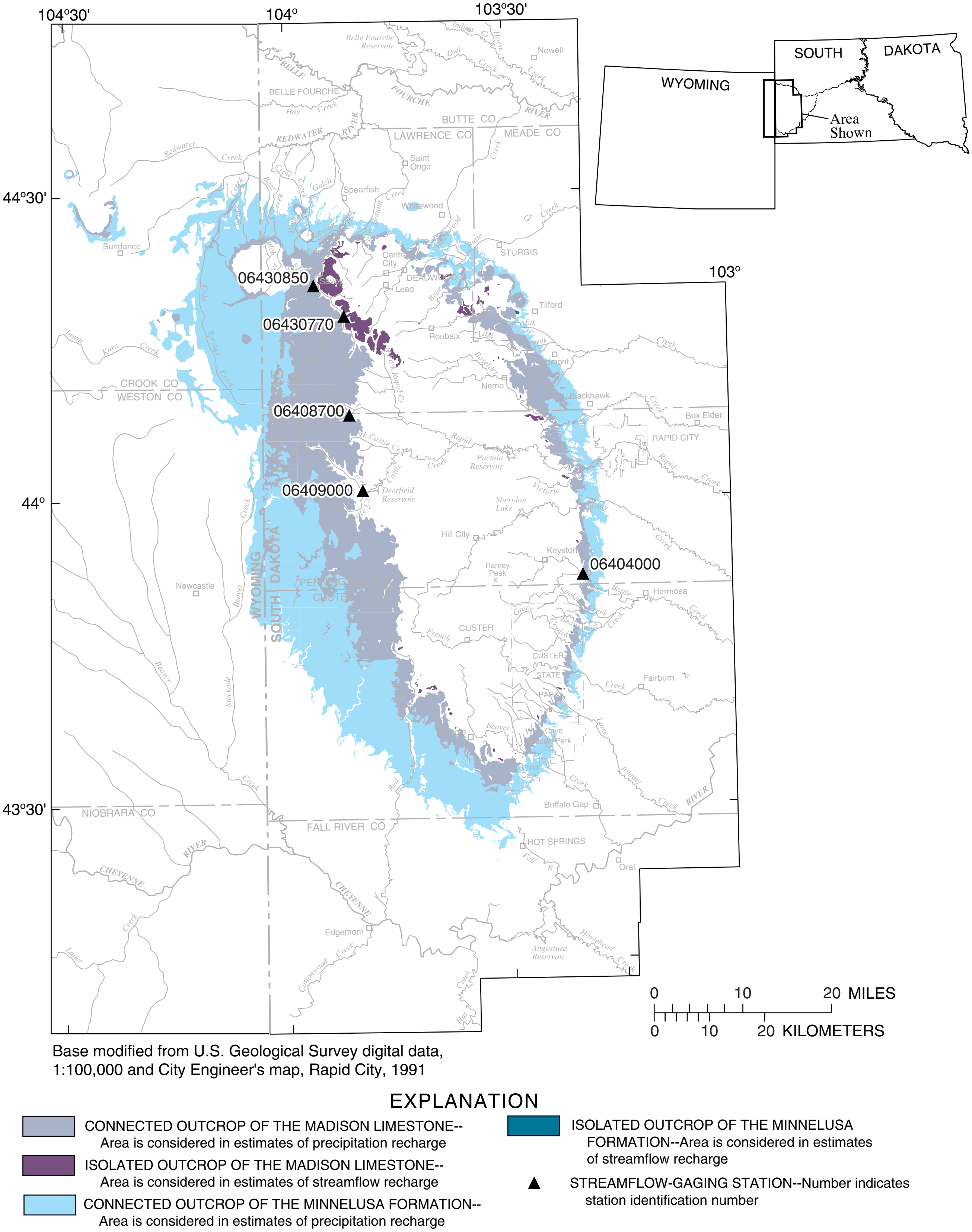
Annual recharge estimates are made for water years 1931-98, which corresponds with a period for which precipitation records have been compiled for the Black Hills area (Driscoll, Hamade, and Kenner, 2000). All recharge estimates provided in this report are by water year, which represents the period from October 1 through September 30, and all discussions of timeframes refer to water years, rather than calender years, unless noted otherwise.

Considerations Regarding Recharge Areas

Because outcrops of the Madison Limestone and Minnelusa Formation are not entirely continuous throughout the study area, quantifying precipitation recharge requires identification of outcrop areas where effective recharge occurs. Outcrops that are considered “isolated” from the regional ground-water flow system (erosional remnants) are identified in figure 7. Recharge that occurs in isolated outcrops does not directly join the regional ground-water flow system because these outcrops are not hydraulically connected to a regional aquifer. Thus, for subsequent calculations, precipitation recharge is prescribed only for the “connected” outcrops of the Madison Limestone and Minnelusa Formation.

Subsequent calculations of streamflow recharge require determination of drainage areas contributing to streamflow loss zones that occur within outcrop areas of the Madison Limestone and Minnelusa Formation. For these calculations, isolated outcrops of the Madison Limestone and Minnelusa Formation are included as drainage areas contributing to loss zones. Direct runoff from the isolated outcrops probably is uncommon; however, these areas generally contribute base flow to streams upstream from loss zones. Several small basins upstream from loss zones contain minor connected outcrops that are subtracted from the drainage areas contributing to streamflow loss zones.

Isolated outcrop areas were determined from hydrogeologic and structure-contour maps of the study area (DeWitt and others, 1989; Carter and Redden, 1999c, 1999d; Strobel and others, 1999) and are identified in figure 7. Outcrop areas generally are considered isolated where surrounded by outcrops of an older formation or by Tertiary intrusives because recharge would not be able to move laterally without eventually being discharged at the contact with the older formation or intrusive. An exception to this criterion is that outcrops of the Minnelusa Formation that are surrounded by outcrops of the Madison Limestone are considered connected, rather than isolated.



Base modified from U.S. Geological Survey digital data, 1:100,000 and City Engineer's map, Rapid City, 1991

Figure 7. Connected and isolated outcrop areas of the Madison Limestone and Minnelusa Formation for recharge considerations (geology modified from Strobel and others, 1999; DeWitt and others, 1989). Location of streamflow-gaging stations for which duration hydrographs also are presented.

Recharge estimates presented in this report consists of "regional recharge," which refers to recharge to outcrops connected to the regional flow system. Precipitation recharge to isolated outcrops of the Madison Limestone and Minnelusa Formation is excluded because most of this recharge is ultimately discharged as base flow to streams, which may be subsequently recharged in loss zones located farther downstream. The term "regional recharge" is used primarily because of considerations regarding large headwater springs located mainly along Rapid Creek and Spearfish Creek and recharged in the Limestone Plateau area (fig. 1). Some of this water from headwater springs also contributes to subsequent streamflow recharge farther downstream; however, two important distinctions exist between infiltration of precipitation on the Limestone Plateau area and on isolated outcrops. First, the water in the Limestone Plateau area is part of the regional flow system recharged in the continuous part of the formation outcrops prior to discharge at headwater springs; hence the term regional recharge. Second, much of the discharge from the headwater springs in Rapid Creek and Spearfish Creek does not necessarily contribute to subsequent streamflow recharge. Streamflow losses in these streams are small, relative to the drainage areas, and streamflow generated from other areas generally is sufficient to satisfy the loss thresholds.

Methods for Quantifying Streamflow Recharge

The Madison and Minnelusa aquifers receive relatively consistent recharge from area streams, which generally lose flow crossing the formation outcrops. During periods of base flow, most streams generally lose their entire flow as they cross these outcrops (loss zones), up to "threshold" rates that are unique for each stream. Hortness and Driscoll (1998) concluded that loss thresholds for individual streams generally are relatively constant, without measurable effects from flow rate or duration of flow through loss zones. Minor variability in apparent loss rates was attributed to localized springflow within loss reaches.

Estimates of streamflow recharge are based, when possible, on loss thresholds that were determined by Hortness and Driscoll (1998) for 24 area streams. This constitutes the majority of drainage areas that provide streamflow recharge to the Madison and Minnelusa aquifers. Some of the loss thresholds determined by Hortness and Driscoll (1998) were based on measurement sites that do not include the entire drainage area above the outcrops. Therefore, some of the thresholds are adjusted to account for additional, unmeasured flow from the additional minor drainage areas. Estimates of streamflow recharge exclude alluvial ground-water flow upstream from loss zones because alluvial flow could not be determined.

Some of the stream reaches measured by Hortness and Driscoll (1998) included outcrops of the Deadwood Formation or Minnekahta Limestone, primarily because of access considerations. Thus, some of the calculated loss thresholds may apply to these outcrops. Examination of additional information led to a conclusion by Hortness and Driscoll (1998) that losses to the Deadwood Formation generally are minimal. Losses to the Minnekahta Limestone were difficult to isolate from potential losses to extensive alluvial deposits that commonly occur near outcrops of the Minnekahta Limestone. For this report, all streamflow losses are assumed to recharge the Madison and Minnelusa aquifers, except those specifically identified by Hortness and Driscoll (1998) for other aquifers.

Estimates of streamflow recharge are developed for three types of drainage basins: (1) those with continuous-record streamflow-gaging stations, (2) those with only miscellaneous-record measurement sites; and (3) those with no available measurements (ungaged). Loss thresholds have not been determined for the ungaged basins, but were available from Hortness and Driscoll (1998) for the other two types of basins.

For the basins with continuous-record gaging stations, daily mean flows are available, and loss threshold values can be used along with daily flow records to calculate recharge rates. The general method for calculating recharge rates follows: (1) if the daily mean flow measured at the gaging station was less than the loss threshold rate, daily recharge to the Madison and/or Minnelusa aquifers was equal to the measured flow; or (2) if the measured flow was greater than or equal to the loss threshold rate, daily recharge to the aquifers was equal to the threshold rate. Calculated daily losses were aggregated to provide estimates of annual recharge.

For some streams, Hortness and Driscoll (1998) were able to quantify individual loss thresholds to the Madison and Minnelusa aquifers; thus, individual and combined recharge to the aquifers can be determined. For stations for which individual loss thresholds had been determined, the loss threshold for the Madison aquifer is applied first to daily mean flows, and any flow greater than this threshold then is applied to the loss threshold for the Minnelusa aquifer. Combined recharge rates are equal to the sum of the individual recharge rates of the Madison and Minnelusa aquifers.

Flows from selected continuous-record gaging stations are used to estimate daily flows for streams with miscellaneous-record measurement sites. The daily flow estimates are based strictly on the ratio of the drainage area for each basin, relative to the drainage area for a representative continuous-record gage. Daily losses are calculated in the same fashion as those for the continuous-record gaging stations, and annual recharge again is computed by aggregating daily losses.

The ungaged basins generally consist of small drainage areas with undetermined loss thresholds that are situated between larger basins for which loss thresholds have been determined. Hortness and Driscoll (1998) did not attempt to quantify loss thresholds for these small basins; however, field observations indicated that flow seldom occurs below the loss zone. Therefore, a simplifying assumption that 90 percent of runoff generated within these basins becomes recharge to the Madison and Minnelusa aquifers is made for estimating recharge from ungaged streams. Annual flows for ungaged basins are estimated strictly from annual flows for representative continuous-record gages, again using drainage-area ratios. Because the ungaged basins contain outcrops of the Deadwood Formation, which would receive precipitation recharge to the Deadwood aquifer, streamflow recharge to the Madison and Minnelusa aquifers is overestimated slightly. However, this slight overestimation is assumed to be equal to the alluvial ground-water flow upstream from loss zones that could not be determined.

All of the continuous-record gages used for direct calculation of daily losses have daily records at least for water years 1992-98, with the oldest records dating to 1962. A variety of regression methods are used to estimate streamflow back to 1950 for calculation of streamflow recharge, which requires utilization of gages with longer records. Estimates of streamflow recharge are further extended to 1931 using correlations with estimates of precipitation recharge. Additional details are provided in subsequent sections. An evaluation of uncertainties associated with recharge estimates also is provided.

Methods for Quantifying Precipitation Recharge

Recharge resulting from infiltration of direct precipitation can be a very difficult variable to quantify. Pan evaporation, which can be measured directly, might be useful in computing precipitation recharge. However, evaporation data are sparse and evaporation rates are quite variable in the study area, primarily because of differences in energy input resulting from differences in elevation and aspect (Wrage, 1994). Furthermore, pan evaporation exceeds precipitation for most parts of the Black Hills during all but the wettest years. Thus, evapotranspiration generally is limited by precipitation amounts and availability of soil moisture. Measured evapotranspiration rates of the Black Hills pine forest do not exist, and estimation of evapotranspiration generally involves extensive modeling efforts that require input of hourly climatic data (Fluke, 1996).

Development of the assumption that surface runoff from outcrops of the Madison Limestone and Minnelusa Formation is negligible (as discussed in a previous section) provides a simplified approach to quantifying precipitation recharge. By neglecting surface runoff, it can be assumed that all precipitation on outcrops of the Madison Limestone and Minnelusa Formation that is not evapotranspired becomes recharge, as schematically illustrated in figure 8.

Streamflow in drainage basins within the crystalline core of the Black Hills area can be used as an indirect measure of evapotranspiration. This concept also is schematically illustrated in figure 8. A similar approach was used by Anderson (1980) in three watersheds in the Sturgis area. Recharge does occur to numerous localized aquifers in fractured crystalline rocks, especially where extensive weathering has occurred in outcrop areas. However, these aquifers are not regional, 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). Therefore, regional ground-water flow in the crystalline rocks can reasonably be considered negligible.

Streamflow records are available for numerous drainage basins within the crystalline core area, which are appropriate for use in estimating basin yield. In the absence of a regional ground-water flow component, basin yield can be considered as the residual between precipitation and evapotranspiration, for periods sufficiently long to neglect change in storage. As discussed, localized aquifers are common in the fractured crystalline rocks, and streams draining these rocks generally

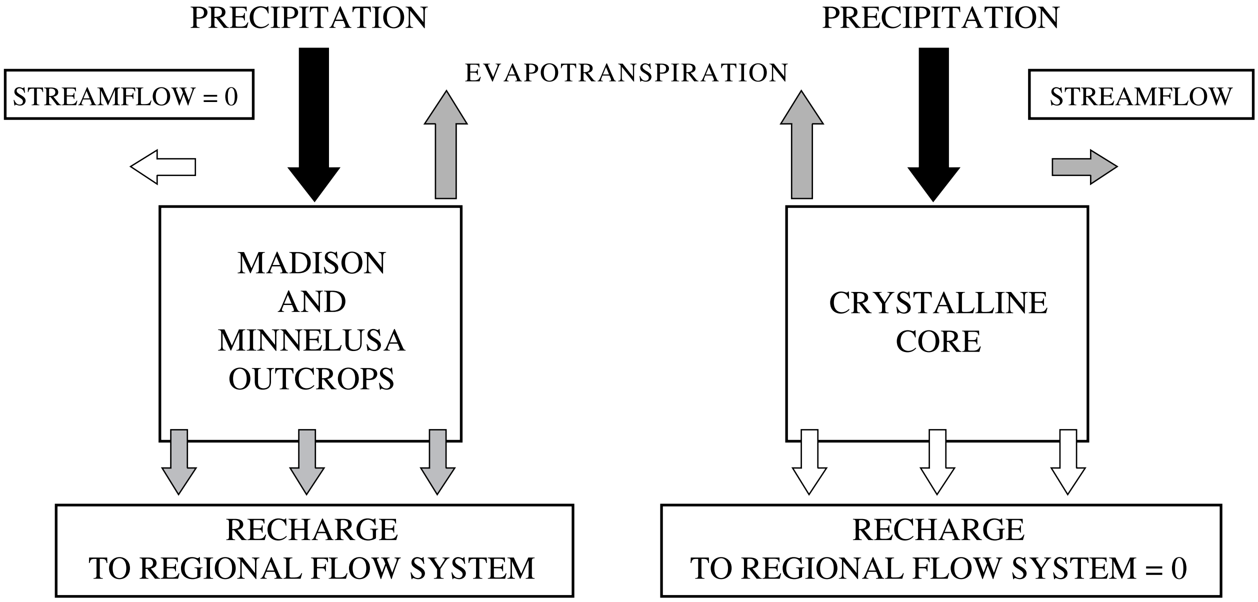


Figure 8. Schematic diagram illustrating recharge and streamflow characteristics for selected outcrop types.

have at least some component of base flow that can be attributed to ground-water discharge. However, the relatively minor ground-water components in these areas primarily reflect changes in storage in the crystalline rocks. Thus, streamflow (or basin yield) effectively represents the entire quantity of water not lost through evapotranspiration, which for the crystalline areas consists predominantly of runoff with a minor ground-water component.

In this report, basin yields are first normalized, relative to drainage area, by expressing in inches per unit of drainage area. Yields are further converted to yield efficiencies, by dividing by precipitation on contributing drainage areas. Relations between yield efficiency and precipitation are identified, which are developed for use in generically estimating annual yield for given areas, based on average yield efficiency and annual precipitation. The resulting annual yield is used as a surrogate for estimating annual recharge from infiltration of precipitation on outcrop areas of the Madison and Minnelusa aquifers. Additional details are provided in subsequent sections.

Uncertainties Associated with Recharge Estimates

There are a large number of uncertainties associated with the recharge estimates provided in this report. Most of the uncertainties cannot be accurately evaluated because of unknowns associated with the variables involved and the broad assumptions necessary in estimating recharge. It is possible, however, to provide a sense of the relative level of uncertainty associated with most of the methods used. Following are preliminary discussions of uncertainties associated with some of these methods. Additional discussions are provided in subsequent sections, where additional details regarding methods or results are available.

Uncertainties for estimates of streamflow recharge for the continuous-record gages probably are small, relative to other uncertainties, because uncertainties associated with measured flow records and the determination of loss thresholds are relatively small. Estimates of streamflow recharge for 1992-98 are better than estimates for earlier periods because more continuous-record gaging stations were in operation. Additional uncertainties are introduced when flow estimates are based on flow records for other gages, which is done for continuous-record gages outside of the period of record, miscellaneous-record measurement sites, and ungaged basins. Estimates for ungaged basins have additional uncertainty associated with the assumption that 90 percent of streamflow in these ungaged areas becomes recharge. This additional uncertainty is not particularly critical, however, because the ungaged basins constitute less than 10 percent of the drainage area contributing streamflow recharge to the Madison and Minnelusa aquifers, compared with about 80 percent for basins with continuous flow records. The largest uncertainties for streamflow recharge estimates are for 1931-50, when estimates are based on correlations with estimates of precipitation recharge.

Uncertainties associated with estimates of precipitation recharge result from: (1) the methods used and associated assumptions, which may be large and cannot be quantified (additional discussions of these uncertainties will be provided later in the report); and (2) measurement of precipitation. Uncertainties become progressively larger for earlier periods due to sparser precipitation data.

The methods that are used for estimating precipitation recharge provide a consistent, systematic approach that is based on precipitation measurements that have a relatively small level of uncertainty. Minor uncertainty is associated with the spatial distribution of measured precipitation; however, the method used (Driscoll, Hamade, and Kenner, 2000) is consistent and systematic, and probably introduces little bias. Thus, errors associated with the spatial distribution of precipitation probably are random and tend to cancel out over time.

Large uncertainties are associated with the approach that is used for generically estimating annual basin yield and yield efficiency, along with the assumption that yield efficiency is a reasonable surrogate for estimating recharge rates for the Madison and Minnelusa aquifers. There also is considerable potential for systematic bias associated with this assumption. A likely source of bias is that precipitation recharge to the Madison and Minnelusa aquifers may be consistently underestimated. An inherent assumption associated with the approach is that the amount of water escaping the root zone in the outcrops of the Madison Limestone and Minnelusa Formation is similar to that escaping the root zone in lower permeability settings such as the Precambrian rocks, where the ground-water component of streamflow is relatively small. Because of the large secondary porosities associated with outcrops of the Madison Limestone and Minnelusa Formation, it is likely that the amount of water escaping the root zone in these outcrops is larger than in other settings. Therefore, the recharge estimates presented in this report probably are conservative.

In general, the best recharge estimates are streamflow recharge values for 1992-98 that are calculated from measured loss thresholds and daily streamflow records for continuous-record gages. Estimates of streamflow recharge become progressively more uncertain for previous periods, as availability of streamflow records becomes sparser. The uncertainty associated with estimates of precipitation recharge generally is larger than for streamflow recharge. This does not necessarily imply that errors are large, but does recognize that potential for error is large. The uncertainty associated with estimates of precipitation recharge changes little over time and is influenced only by availability of precipitation measurement sites. Thus, uncertainties for combined recharge from streamflow and precipitation are subject to less change over time than estimates of streamflow recharge alone. Although recharge estimates are somewhat poorer for earlier periods, estimates for the 1930's and 1950's are especially important, because this is the driest period for which adequate precipitation data are available for hydrologic analysis.

As discussed, uncertainties associated with recharge estimates cannot be evaluated precisely at this time. Results of an initial water-budget analysis, which utilized the same general methods for estimation of recharge, were presented by Hamade (2000). These initial results indicate that recharge estimates are in a range that is compatible with other components of the water budget.

STREAMFLOW RECHARGE

Streamflow losses from area streams provide a consistent source of recharge to the Madison and Minnelusa aquifers. Streamflow records for 39 measurement sites (table 1 fig. 9) are considered in calculating streamflow recharge. One gage (06425500; site 22 in table 1) used in quantifying streamflow recharge is outside the study area and is shown in figure 1. Most of the gages are used for direct calculations of streamflow recharge. Several gages (sites 9, 15, 19, 22, 27, 28, 31, and 35) are used only in statistical correlations for extending streamflow records.

The streamflow measurement sites are used to delineate 13 drainage basins with continuous-record gages and 19 basins with miscellaneous measurement sites (fig. 10). In addition, 23 ungaged basins are delineated. Basins with continuous-record gages account for 78 percent of the study area, and basins with miscellaneous-record measurement sites account for 13 percent. The ungaged basins account for only 9 percent of the study area.

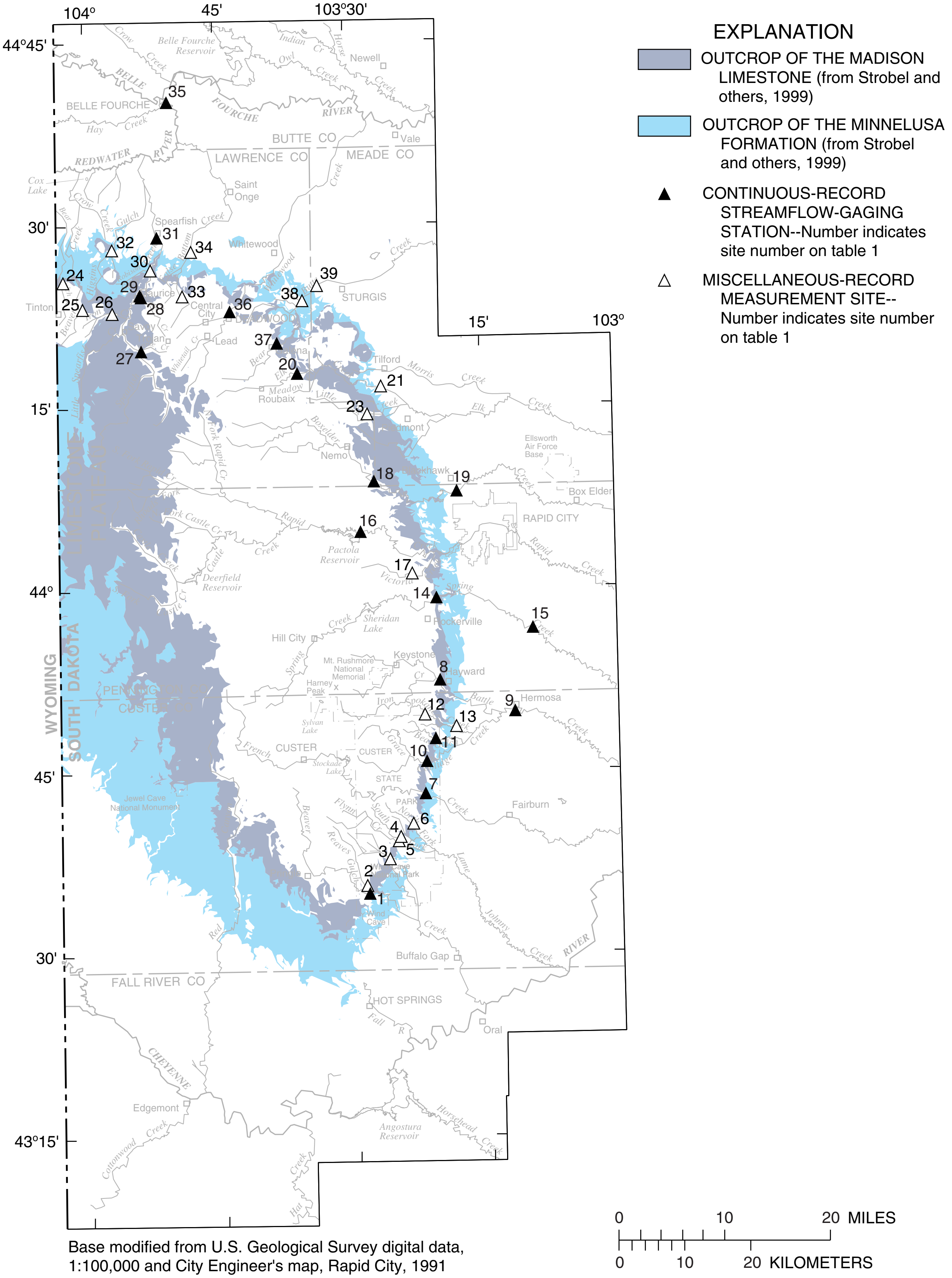


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Figure 9. Location of gaging stations used to estimate streamflow recharge.

Table 1. Summary of selected site information for gaging stations used in determining streamflow recharge

[Type of station: C, continuous-record; M, miscellaneous-record. --, none used]

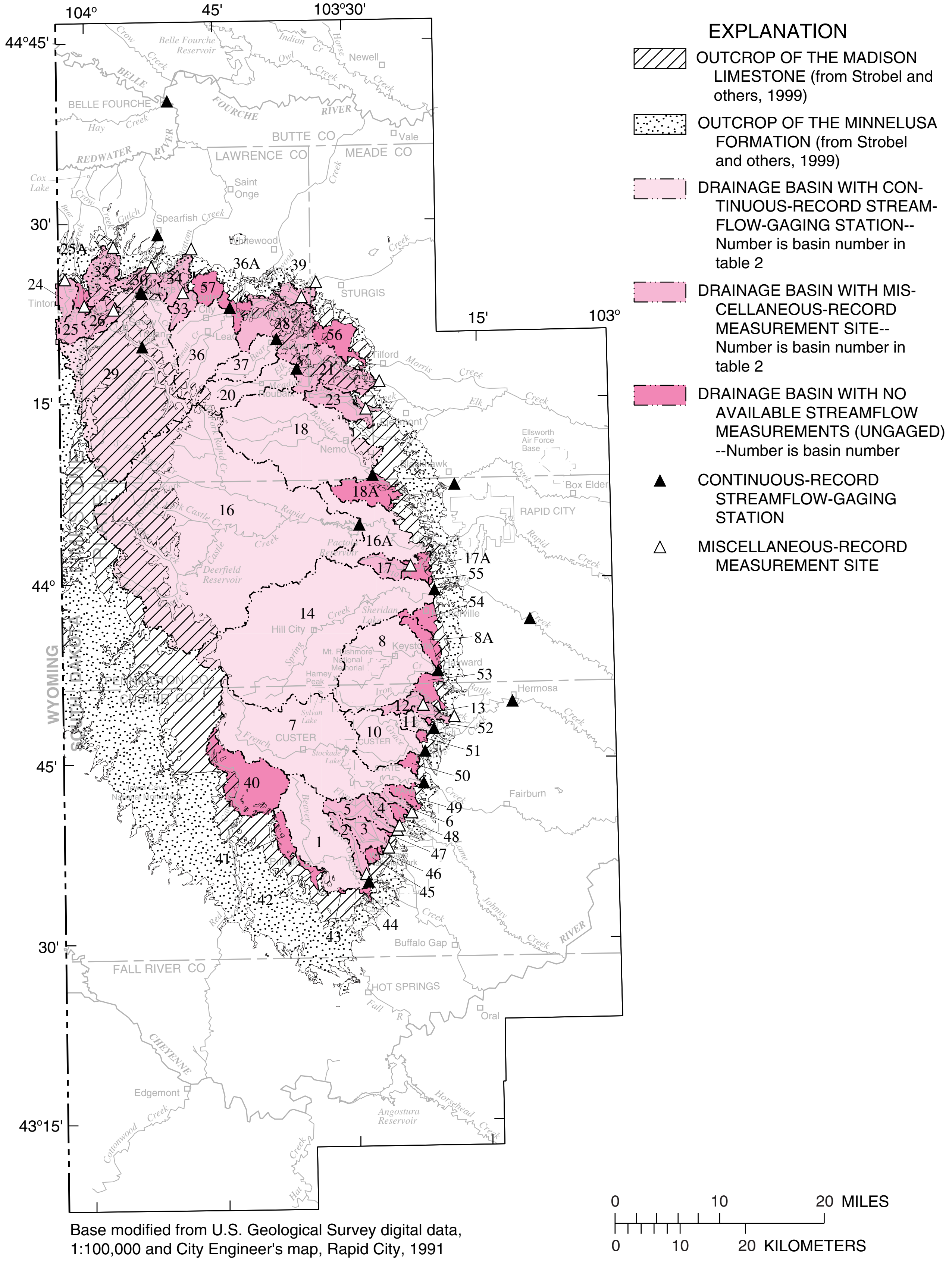
Site number	Station identification number	Station name	Latitude	Longitude	Type of station	Drainage area (square miles)	Period of record used (water years)
			(degrees, minutes, seconds)				
1	06402430	Beaver Creek near Pringle	43 34 53	103 28 34	C	45.8	1991-98
2	433532103284800	Reaves Gulch above Madison outcrop, near Pringle	43 35 32	103 28 48	M	6.86	--
3	433745103261900	Highland Creek above Madison outcrop, near Pringle	43 37 45	103 26 19	M	8.69	--
4	433930103250000	South Fork Lame Johnny Creek above Madison outcrop, near Fairburn	43 39 30	103 25 00	M	4.34	--
5	433910103251000	Flynn Creek above Madison outcrop, near Fairburn	43 39 10	103 25 10	M	10.3	--
6	434105103240200	North Fork Lame Johnny Creek above Madison outcrop, near Fairburn	43 41 05	103 24 02	M	2.80	--
7	06403300	French Creek above Fairburn	43 43 02	103 22 03	C	105	1983-98
8	06404000	Battle Creek near Keystone	43 52 21	103 20 10	C	58.0	1962-98
9	06406000	Battle Creek at Hermosa	43 49 41	103 11 44	C[1]	178	1950-98
10	06404998	Grace Coolidge Creek near Game Lodge, near Custer	43 45 40	103 21 49	C	25.2	1977-98
11	06405800	Bear Gulch near Hayward	43 47 31	103 20 49	C	4.23	1990-98
12	434929103215700	Spokane Creek above Madison outcrop, near Hayward	43 49 29	103 21 57	M	4.92	--
13	434800103174400	Spokane Creek below Madison outcrop, near Hayward	43 48 00	103 17 44	M	3.76	--
14	06407500	Spring Creek near Keystone	43 58 45	103 20 25	C	163	1987-98
15	06408500	Spring Creek near Hermosa	43 56 31	103 09 32	C[1]	199	1950-98
16	06411500	Rapid Creek below Pactola Dam	44 04 36	103 28 54	C	320	1946-98
17	440105103230700	Victoria Creek below Victoria Dam, near Rapid City	44 01 05	103 23 07	M	6.82	--
18	06422500	Boxelder Creek near Nemo	44 08 38	103 27 16	C	96.0	1967-98
19	06423010	Boxelder Creek near Rapid City	44 07 54	103 17 54	C	128	1978-98
20	06424000	Elk Creek near Roubaix	44 17 41	103 35 47	C	21.5	1992-98
21	441614103253300	Elk Creek at Minnekahta outcrop, near Tilford	44 16 14	103 25 33	M	23.8	--
22	06425500	Elk Creek near Elm Springs	44 14 54	102 30 10	C[1]	540	1950-98
23	441412103275600	Little Elk Creek below Dalton Lake, near Piedmont	44 14 12	103 27 56	M	11.39	--

Table 1. Summary of selected site information for gaging stations used in determining streamflow recharge–Continued

[Type of station: C, continuous-record; M, miscellaneous-record. --, none used]

Site number	Station identification number	Station name	Latitude (degrees, minutes, seconds)	Longitude (degrees, minutes, seconds)	Type of station	Drainage area (square miles)	Period of record used (water years)
24	06429920	Bear Gulch near Maurice	44 25 14	104 02 26	M	6.17	--
25	06430520	Beaver Creek near Maurice	44 22 57	104 00 13	M	6.86	--
26	442242103565400	Iron Creek below Sawmill Gulch, near Savoy	44 22 42	103 56 54	M	8.16	--
27	06430800	Annie Creek near Lead	44 19 37	103 53 38	C[1]	3.55	1989-98
28	06430898	Squaw Creek near Spearfish	44 24 04	103 53 35	C[1]	6.95	1989-98
29	06430900	Spearfish Creek above Spearfish	44 24 06	103 53 40	C	139	1989-97
30	06430950	Spearfish Creek below Robison Gulch, near Spearfish	44 26 14	103 52 32	M	8.44	--
31	06431500	Spearfish Creek at Spearfish	44 28 57	103 51 40	C	168	1947-98
32	442754103565000	Higgins Gulch below East Fork, near Spearfish	44 27 54	103 56 50	M	12.55	--
33	442405103485100	False Bottom Creek above Madison outcrop, near Central City	44 24 05	103 48 51	M	5.55	--
34	06432180	False Bottom Creek (below Minnelusa outcrop) near Spearfish	44 27 09	103 48 22	M	8.91	--
35	06433000	Redwater River above Belle Fourche	44 40 02	103 50 20	C[1]	920	1946-98
36	06436170	Whitewood Creek at Deadwood	44 22 48	103 43 25	C	40.6	1981-95
37	06437020	Bear Butte Creek near Deadwood	44 20 08	103 38 06	C	16.6	1989-98
38	442337103350600	Bear Butte Creek at Boulder Park, near Sturgis	44 23 37	103 35 06	M	32.23	--
39	442447103332800	Bear Butte Creek above Sturgis	44 24 47	103 33 28	M	5.59	--

[1] Continuous-record station used only for extension of streamflow records.



Base modified from U.S. Geological Survey digital data, 1:100,000 and City Engineer's map, Rapid City, 1991

Figure 10. Drainage basins for which streamflow recharge was estimated.