GEOHYDROLOGY AND WATER QUALITY OF THE INYAN KARA, MINNELUSA, AND MADISON AQUIFERS OF THE NORTHERN BLACK HILLS, SOUTH DAKOTA AND WYOMING, AND BEAR LODGE MOUNTAINS, WYOMING

By David P. Kyllonen and Kathy D. Peter

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CONVERSION FACTORS

For use of readers who prefer to use metric units, conversion factors for terms used in this report are listed below:

Multiply Inch-pound units	By	To obtain SI units
foot (ft)	0.3048	meter
foot per mile (ft/mi)	0.1894	meter per kilometer
foot per day (ft/d)	0.3048	meter per day
foot squared per day (ft^2/d)	0.0929	meter squared per day
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
gallon per minute (gal/min)	0.0630	liter per second
gallon per minute per foot [(gal/min)/ft]	0.2067	liter per second per meter
inch	25.40	millimeter
inch per year (in/yr)	25.40	millimeter per year
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer

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

$$^{\circ}F = 1.8(^{\circ}C) + 32$$

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ABSTRACT

The Inyan Kara, Minnelusa, and Madison aquifers are the principal sources of ground water in the northern Black Hills, South Dakota and Wyoming, and Bear Lodge Mountains, Wyoming. The Inyan Kara aquifer is composed of the sandstone beds of the Inyan Kara Group of Early Cretaceous age. The Inyan Kara Group is 81 to 475 feet thick. The Minnelusa aquifer is composed of sandstone beds in the upper 200 to 300 feet of the Minnelusa Formation of Pennsylvanian and Permian age, which is 340 to 800 feet thick. The Madison Limestone of Mississippian age is 400 to 850 feet thick and is an aquifer where solution features and fractures provide permeability. The aquifers are exposed in the Bear Lodge Mountains and the Black Hills and are about 3,000 to 5,000 feet below the land surface in the northeast corner of the study area. The direction of ground-water movement is from the outcrop area toward central South Dakota.

Recharge is by infiltration of precipitation and streamflow on the outcrop. Discharge is by springs and well withdrawals. Water leaks between aquifers through semiconfining layers, wells completed in more than one aquifer, and wells with deteriorating casing. The aquifers extend outside the study area and water flows into and out of the study area in the subsurface.

All three aquifers yield water to flowing wells in some part of the area. Measured and reported well yields in each of the three aquifers exceed 100 gallons per minute. A well open to the Minnelusa Formation and the upper part of the Madison Limestone yielded more than 2,000 gallons per minute.

Water from the Inyan Kara aquifer may require treatment for gross alpha radiation, iron, manganese, sulfate, and hardness before use in public water systems. In some areas the concentration of sodium and bicarbonate may affect the use of water for irrigation, depending on the soil type. Water from the Minnelusa aquifer in the northern one-half of the study area may require treatment for sulfate and hardness before use in public water systems. Water from the Madison aquifer in the northern one-half of the study area may require treatment of fluoride, gross alpha radiation, sulfate, and hardness before use in public water systems. Water from the Minnelusa and Madison aquifers in the southern one-half of the study area, though very hard (more than 180 milligrams per liter hardness as calcium carbonate), is suitable for public water systems and irrigation.

A digital model was used to improve the understanding of the hydrology. Flow between the Minnelusa and the Inyan Kara aquifers appears to be insignificant, based on the model results. The relationship between the Minnelusa, the Madison, and perhaps deeper aquifers is not understood. The model indicated there may be significant recharge to the Minnelusa and Madison aquifers by leakage between these two aquifers and perhaps deeper aquifers.

INTRODUCTION

Known and predicted demands by municipal and industrial users for water in the Black Hills have increased interest in the potential ground-water supplies in the area. Since settlement in the area began during the early 1900's, the major users of water have been municipalities, many of which depend entirely on ground-water supplies, and agriculture. Interest has increased in ground water in the Black Hills for industrial uses, such as coal slurry pipelines and coal gasification plants, as a result of expansion of coal development in northeastern Wyoming beginning in the 1970's. Lack of information on the aquifers in the Black Hills has hampered evaluation of potential impacts of future development. A better understanding of the aquifer system of the northern Black Hills would assist in planning data collection and further investigations.

Purpose and Scope

The purpose of this report is to describe the geohydrology and water quality of the three major aquifers--Inyan Kara, Minnelusa, and Madison--in the northern Black Hills and the Bear Lodge Mountains. The study area (fig. 1) is about 3,000 mi^2 and includes parts of Lawrence, Meade, and Butte Counties in South Dakota and Crook County in Wyoming.

There are three major physiographic features in the study area, the Black Hills, the Bear Lodge Mountains, and the plains underlain by the Williston Basin (fig. 1). Land surface altitude of the Black Hills in the study area ranges from about 4,000 to 7,064 ft above sea level, the highest point being Terry Peak, southwest of Lead. The altitude of the Bear Lodge Mountains ranges from about 3,400 to 6,656 ft above sea level, the highest peak being Warren Peak, northwest of Sundance. The northeast corner of the study area is part of the plains underlain by the Williston Basin where land surface altitude ranges from about 2,800 to 3,200 ft above sea level.

Most of the hydrologic units in the area are not areally extensive or generally capable of yielding large quantities of water to wells. The three aquifers studied are the most extensive and have had the most development. They are part or all of three geologic units: the Inyan Kara Group of Early Cretaceous age, the Minnelusa Formation of Pennsylvanian and Permian age, and the Madison Limestone of Mississippian age (table 1). The Deadwood Formation and the Whitewood Dolomite are believed to be aquifers in other parts of the Williston Basin, but little information is available in the study area to determine their hydrologic properties.

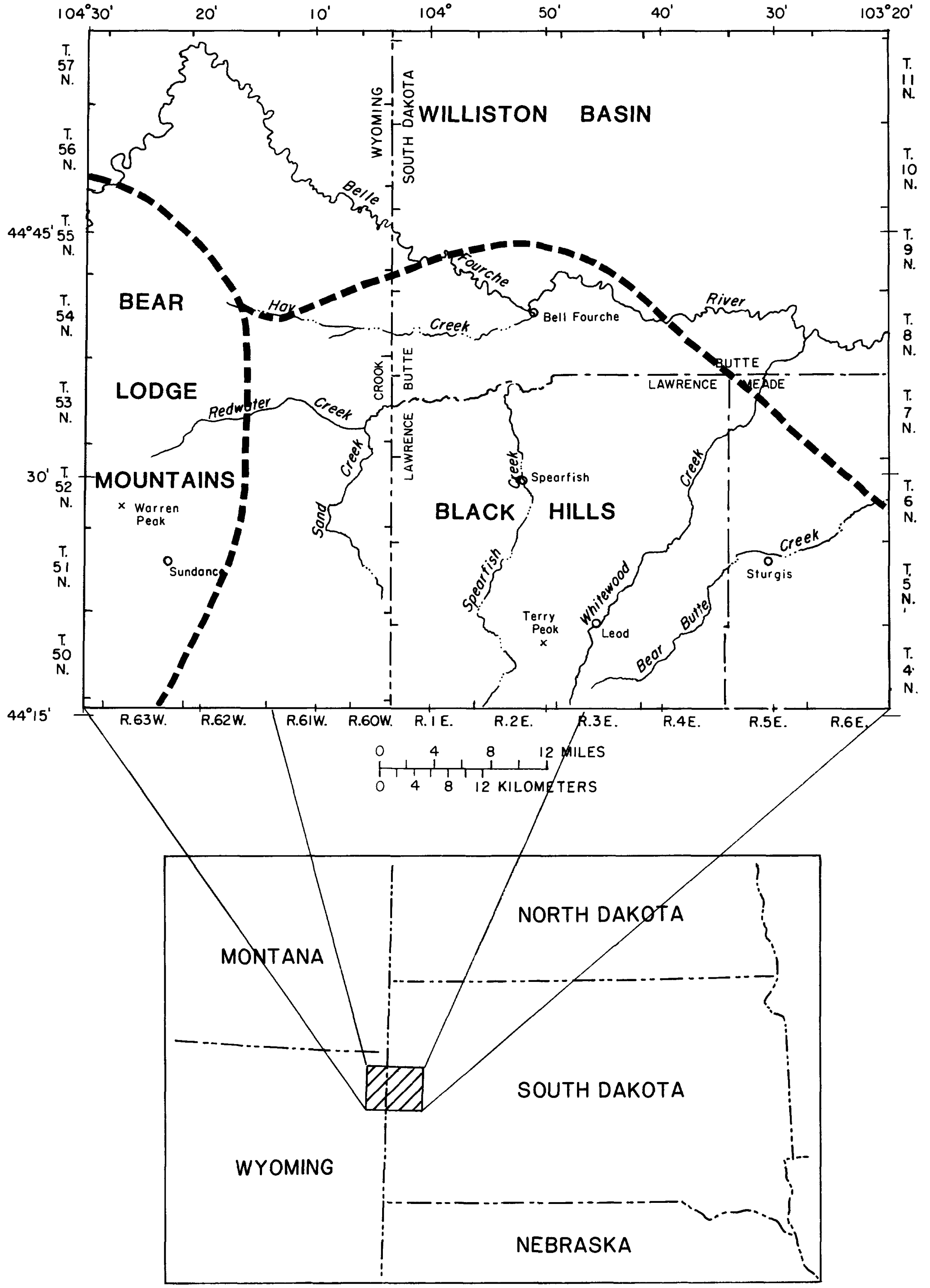


Figure 1.--Location of study area and major physiographic features.

Table 1.--Bedrock formations and aquifers (modified from Robinson, and others, 1964, p. 6-7; Cattermole, 1969 and 1972; Staatz, 1983, p. 6; Peterson, 1984, table 1; and Lisenbee, 1985)

Erathem	System	Series	Geologic unit	Thickness (feet)	Hydrologic unit (model unit)	Description
Cenozoic	Tertiary	Oligocene	White River Formation	0-150	White River aquifer	Sandstone, claystone, and siltstone. Gray and brown. Occurs in isolated outcrops.
		Eocene and Paleocene	Intrusive rocks			Alkalic plutons.
Mesozoic	Cretaceous	Upper	Pierre Shale	1,200-1,400	Confining layer	Shale, sandstone, marl, limestone, and bentonite. Gray and black.
			Niobrara Formation	120-225		
			Carlile Shale	600-800		
			Greenhorn Formation	70-370		
			Belle Fourche Shale	350-850		
		Lower	Mowry Shale	180-230		
			Newcastle Sandstone	0-95	Newcastle aquifer 1/	Sandstone, siltstone, shale, and claystone. The lateral equivalent of the Dakota Sandstone, a major aquifer in eastern South Dakota (Schoon, 1971), the Newcastle is too thin and fine-grained in this area to be a major aquifer.
			Skull Creek Shale	110-321	Confining layer	Shale; black or dark gray.
			Inyan Kara Group: Fall River Formation or Sandstone	10-200	Inyan Kara aquifer (Layer 1)	Sandstone, conglomeratic sandstone, shale, siltstone, and claystone. Coal beds near base locally. Gray, brown, yellow, black, white, red, and green. Thickness varies within short distances. Thin bedded to massive. The top is generally identifiable by the presence of brown sandstone, however, interbedded black shale in the Fall River Formation may be mistaken for the Skull Creek Shale.
			Inyan Kara Group: Lakota Formation	30-300		
	Jurassic	Upper	Morrison Formation / Unkpapa Sandstone 2/	20-260 / 0-150	Confining layer	Shale, siltstone, sandstone, limestone, and massive gypsum beds. Red, green, gray, yellow, pink, and maroon. Parts of these geologic units are aquifers locally, in particular the sandstone members of the Sundance Formation. The upper contact of the Morrison Formation is often difficult to identify because thin, discontinuous, gray sandstone beds in the Morrison are similar to those in the Lakota Formation. The Minnekahta Limestone is particularly easy to identify in the subsurface by its lithology and color; it is gray or pink limestone which stands out in comparison to the red or maroon shales in the overlying and underlying units.
		Middle	Sundance Formation	145-530		
			Gypsum Spring Formation 3/	0-201		
	Triassic		Spearfish Formation	450-845		
	Permian		Minnekahta Limestone	25-65		

			Opeche Formation or Shale	50-199		
Paleozoic	Pennsylvanian		Minnelusa Formation	340-800	Minnelusa aquifer (Layer 2)	Sandstone, limestone, dolomite, and shale. Local gypsum and anhydrite. Yellow, white, brown, purple, red, and gray. Thin to thick bedding. Top is generally identifiable by change from the maroon and red shale and sandstone beds of the Opeche Formation to the yellow, red, and white sandstone beds of the Minnelusa Formation.
					Confining layer	Limestone, dolomite, sandstone and shale. Gray, red, salmon, purple, and white. Thin to thick bedding. Red clayey shale commonly at base of the formation.
	Mississippian		Madison Limestone or Pahasapa Limestone	400-850	Madison aquifer (Layer 3)	Limestone, locally dolomitic. Gray or buff. Generally massive, cavernous in upper part. Permeability from fractures and solution features. Distinguishable from the Minnelusa by the buff color and lack of bedding.
	Devonian		Englewood Formation	40-75	Confining beds	Limestone, dolomite, and shale. Pink, buff, and green. The Whitewood Dolomite is the stratigraphic equivalent of the more extensive Red River Formation. The Englewood Formation appears to be, based on the location of springs and the reaches of stream infiltration, the hydrologic base of the Madison aquifer. The Englewood Formation is identified by its distinctive pink-colored limestone beds.
	Silurian		Not present	0		
	Ordovician		Whitewood Dolomite or Limestone or Red River Formation	50-427		
			Winnipeg Formation	60-179		
	Cambrian		Deadwood Formation	300-500	Deadwood aquifer[1/]	Sandstone, shaly limestone, and shale. White, red, brown, green, and gray. Missing in some areas where replaced by intrusives.
Precambrian			Basement rocks undifferentiated		Base of hydrologic system	Igneous and metamorphic rocks.

1/ Minor source of supply in the study area.
2/ Only in the southern and eastern Black Hills.
3/ Only in the northern and western Black Hills.

Site-Numbering System

The location of wells and springs used in this report are numbered according to the Federal system of land subdivision used in western South Dakota and eastern Wyoming (fig. 2). The first number designates the township and direction (north/south) from the respective parallel in each state. The second number indicates the range and direction (east/west) from the Black Hills Meridian. The third number indicates the section. A section is then divided into quarters (160 acres). These can then be further divided into quarter-quarter sections (40 acres), quarter-quarter-quarter sections (10 acres), and quarter-quarter-quarter-quarter sections (2.5 acres). The number of letters indicates the accuracy of each spring or well location. These section subdivisions are lettered A, B, C, and D in a counter-clockwise direction beginning with A in the northeast quarter, the largest quarter listed first. For example, spring 6N2E15BDD is in the southeast quarter of the southeast quarter of the northwest quarter of section 15 in Township 6 north, Range 2 east. If more than one site is located in a particular subdivision, consecutive numbers are assigned beginning with one. If a section does not measure one-mile square, it is handled as if it were a full section using the southeast section corner as the reference point for subdivision.

Methods of Investigation

To describe the geohydrology and water quality, information was compiled on depth of aquifer tops and bottoms, water levels, hydraulic conductivity, transmissivity, storage coefficient, well yields, and chemical analyses. New observation wells and streamflow-gaging stations were not installed as part of the study. Records of approximately 500 wells were compiled from files of the U.S. Geological Survey in Rapid City, South Dakota, the South Dakota Department of Water and Natural Resources in Pierre, and the Office of the Wyoming State Engineer in Cheyenne. Depths to formation tops and bottoms were interpreted from lithological and geophysical logs. The altitude of the top of the Minnelusa Formation was mapped because it is the most extensive of the three geologic units for which sufficient data were available to enable a reasonably reliable map to be prepared. Thickness maps of the Minnelusa and Madison aquifers were prepared. A total of 656 water-level measurements were used to construct the potentiometric-surface maps of the three aquifers. Historical changes of water levels were evaluated using measurements of the U.S. Geological Survey observation-well network reported by Bradford (1981) and measurements reported by N. H. Darton (1909a and 1918) and Davis and others (1961). Descriptions and locations of the 21 observation wells in the U.S. Geological Survey network are shown in table 2 and figure 3. All but four of these observation wells are also used for water supply. There are no aquifer tests in the study area, therefore, estimates of hydraulic conductivity, transmissivity, and storage coefficient of the aquifers were compiled from previous regional studies. A summary of well yields from the three aquifers was prepared from U.S. Geological Survey Records. Chemical analyses from U.S. Geological Survey records were compiled and compared to criteria established by the U.S. Environmental Protection Agency (1975-80, 1979).

Estimates of recharge and discharge were made using measurements of streamflow, discharge from major springs, and precipitation (table 2). The U.S. Geological Survey maintained 15 streamflow-gaging sites in the study area during 1982 and 1983 (fig. 3) (U.S. Geological Survey, 1976-84), of which eight were used in this study to evaluate changes in streamflow in the outcrop area (table 2). Additional measurements of stream and spring flows made by the U.S. Geological Survey and by P. H. Rahn and J. P. Gries (South Dakota School of Mines and Technology, Rapid City, South Dakota,

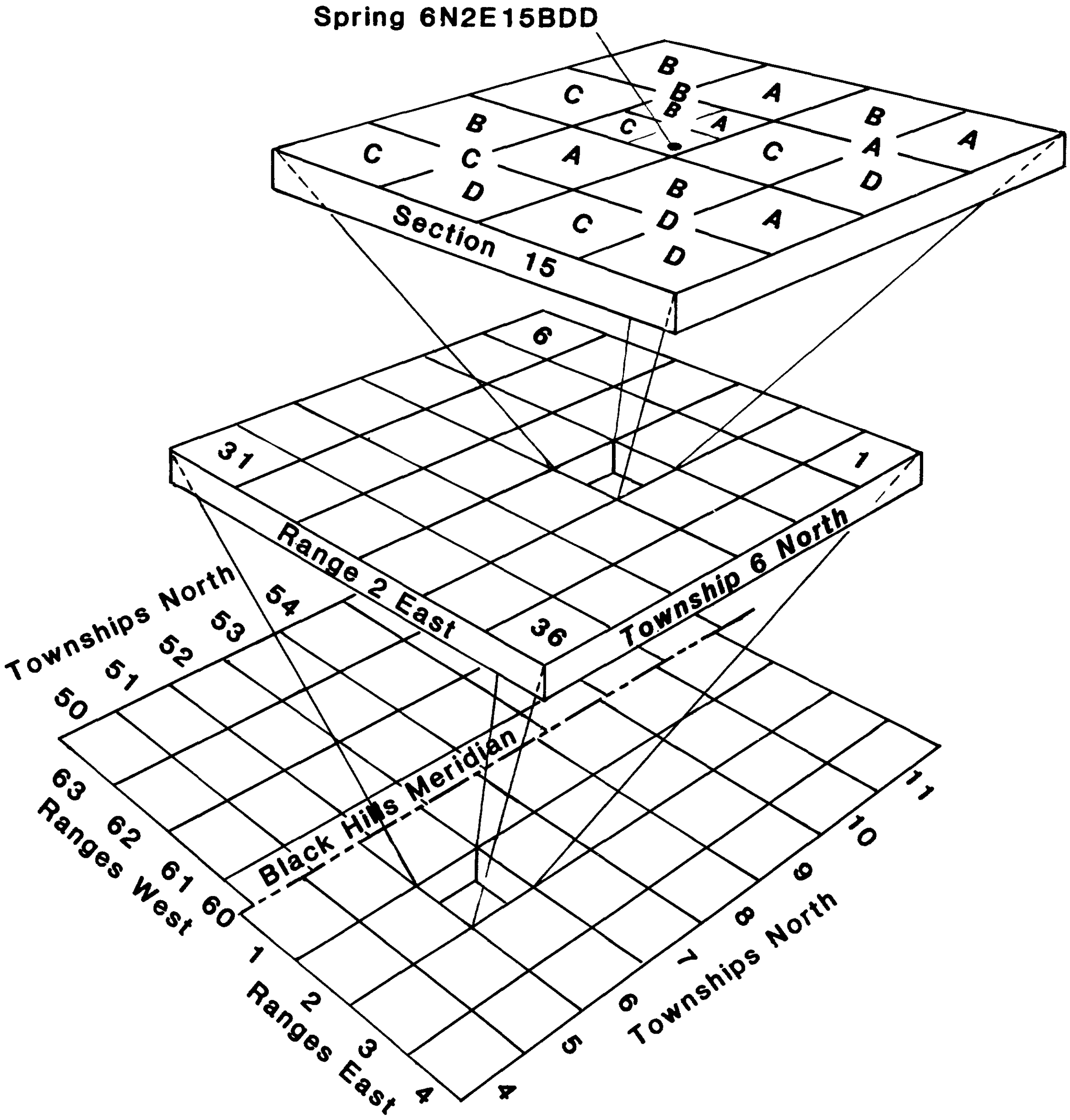


Figure 2.--Site-numbering system.

Table 2.--Selected observation well, streamflow, spring, and precipitation records

SELECTED OBSERVATION WELLS (Bradford, 1981)

Location	Aquifer	Period of record	Highest water level above or below land surface[1] (feet)	Lowest water level above or below land surface[1] (feet)
		Butte County		
7N1E11ABD	Minnelusa	1967-78	134.8	87.9
7N1E12AA	Minnelusa	1962-78	155.6	116.4
8N3E 2BDBC	Minnelusa	1961-72	199.8	26.5
8N3E12ACCC	Inyan Kara	1970-76	.18	-19.17
8N3E12DBBD	Inyan Kara	1970-78	44.8	7.9
8N3E31ACA	Minnelusa	1962-69	202.7	68.4
8N3E33CCB	Minnelusa	1961-79	219.2	124.6
		Lawrence County		
6N2E23BBBA	Minnelusa	1956-80	-36.07	-84.05
7N1E14CCDD2	Minnelusa	1975-80	33.7	15.4
7N1E20AAD	Minnelusa	1966-80	69.14	54.2
7N1E21BBC	Minnelusa	1960-80	74.2	52.2
7N1E26ACD	Minnelusa	1960-80	-19.49	-34.25
7N1E29BBA	Minnelusa	1976-80	46.2	-45.16
7N1E30BDA2	Minnekahta	1952-80	25.6	13.6
7N2E10BAD	Minnelusa	1963-80	302.8	218.6
7N2E19ACC	Minnelusa	1963-67	91.5	75.2
7N2E26BCDA	Minnelusa	1963-80	143.2	112.7
7N3E 7AABA	Minnelusa	1962-79	195.9	112.4
		Meade County		
4N6E19AABB	Minnelusa	1984	-4.40	--
6N5E 21DABA	Inyan Kara	1959-80	-1.01	-12.54
6N5E22DDBC	Inyan Kara	1957-80	60.3	43.30

Table 2.--Selected observation well, streamflow, spring, and precipitation records--Continued

SELECTED STREAM GAGES (U.S. Geological Survey, 1984)

Location	Period of record	Average discharge (cubic feet per second)	Maximum discharge (cubic feet per second)	Minimum discharge (cubic feet per second)
Sand Creek near Ranch A, near Beulah, Wyo.	October 1976 to September 1983 (discontinued)	24.1	514	14
Redwater Creek at Wyoming-South Dakota State line	June 1954 to September 1983[2/]	36.0	2,440	0
Spearfish Creek at Spearfish, S. Dak.	October 1946 to September 1983	52.7	4,240	0
Redwater River above Belle Fourche, S. Dak.	November 1945 to September 1983	137	16,400	0
Hay Creek at Belle Fourche, S. Dak.	October 1953 to September 1983	1.55	930	0
Whitewood Creek at Deadwood, S. Dak.	October 1981 to September 1983	--	2,660	3.5
Whitewood Creek above Whitewood, S. Dak.	October 1982 to September 1983	--	684	11
Whitewood Creek near Whitewood, S. Dak.	October 1981 to September 1983	--	3,050	4.0

SELECTED SPRINGS

Location	Name	Average discharge (cubic feet per second)	Number of measurements
52N61W24A	Sand Creek[3/]	23.81	14 (Rahn and Gries, 1973)
7N1E21BBC	Unnamed	17.47	1 (Rahn and Gries, 1973)

Table 2.--Selected observation well, streamflow, spring, and precipitation records--Continued

SELECTED PRECIPITATION GAGES

(U.S. Department of Commerce, 1980-83, 1981-83)

Station	Period of record (years)	Normal precipitation 4/ (inches)
Newell	63	14.39
Belle Fourche	75	14.87
Spearfish	94	21.06
Fort Meade	68	19.18
Deadwood	64	28.48
Lead 1E	75	28.65
Colony	68	14.36
Alva 5SE	37	22.85
Sundance	59	16.94

1/ (-) below land surface.

2/ Interrupted period of record April 1929 to September 1931, and February 1936 to July 1937 (published as "near Beulah, Wyo.").

3/ The stream gage at Sand Creek near Ranch A, near Beulah, Wyoming, measures stream discharge below this spring.

4/ Normal is defined by the National Oceanic and Atmospheric Administration as the average precipitation over the period 1951-80. The normals for National Weather Service localities are adjusted so as to be representative for the current observation site (U.S. Department of Commerce, 1983).

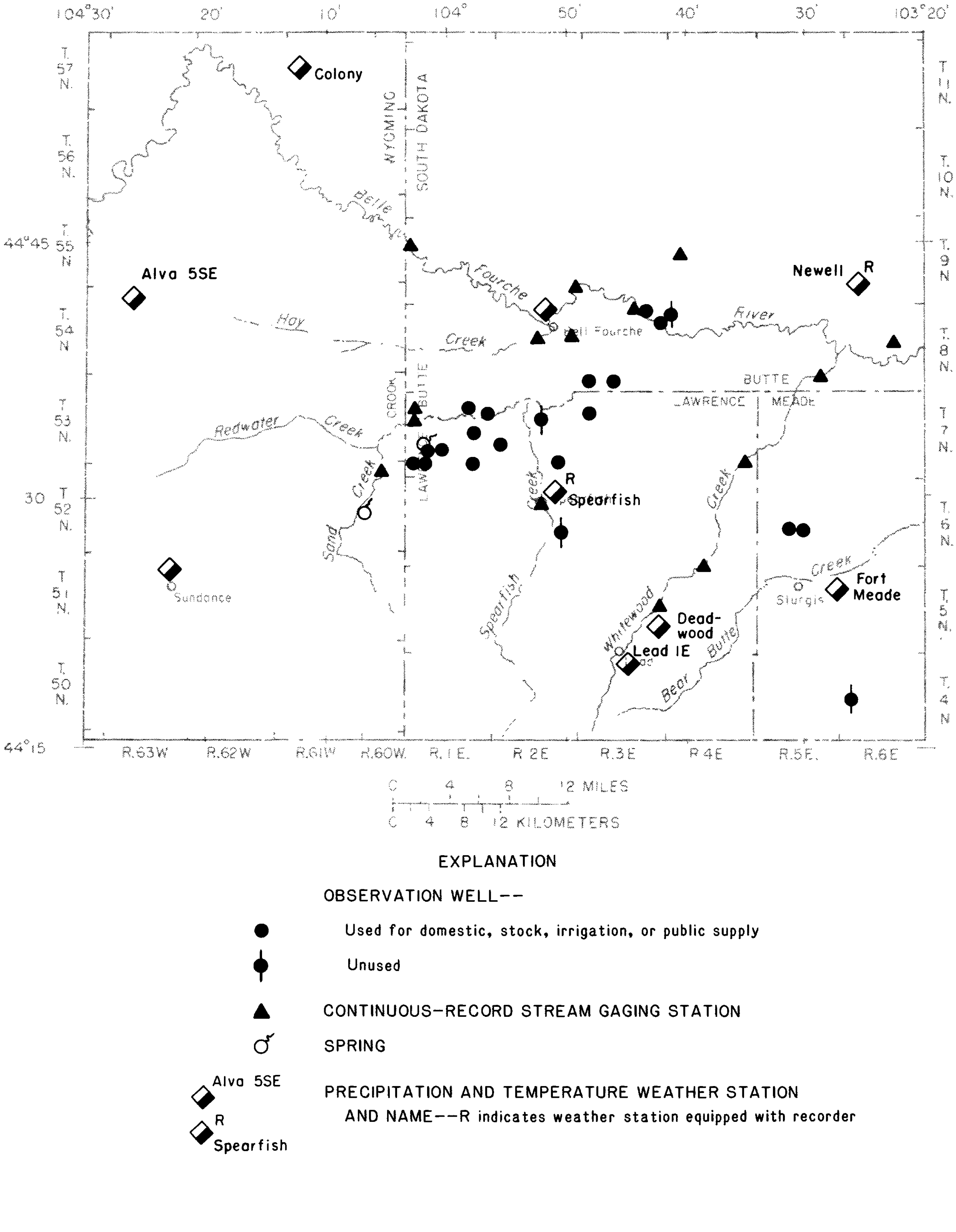


Figure 3.--Location of selected observation wells, streamflow-gaging stations, springs, and weather stations. (Modified from Bradford, 1981; U.S. Department of Commerce, 1980-83 and 1981-83; and U.S. Geological Survey, 1982.)

personal commun., 1983; and Rahn and Gries, 1973) also were used. There are nine precipitation-gaging sites maintained by the National Weather Service, U.S. Department of Commerce in the study area (fig. 3) (U.S. Department of Commerce, 1980-83 and 1981-83).

A digital model was used to test the interpretation of the geohydrology, such as the distribution of transmissivity, recharge and discharge rates, and the relation between the three aquifers. The digital model was developed from the hydrologic data gathered for this study and uses a finite-difference approximation of the mathematical equations which numerically simulate the flow of ground water through the aquifers. Potentiometric levels for each aquifer were computed using the digital model and compared to the measured potentiometric levels.

Acknowledgments

The South Dakota Department of Water and Natural Resources, well drillers, municipal officials, the U.S. Forest Service, the Office of the Wyoming State Engineer, and well owners provided well information. Perry H. Rahn and J. P. Gries of the South Dakota School of Mines and Technology provided miscellaneous stream and spring flow measurements. Perry H. Rahn assisted in a dye test of Whitewood Creek. Alvis L. Lisenbee, also of the South Dakota School of Mines and Technology, provided preliminary copies of his tectonic maps of the Black Hills for use in interpretation of the geologic structure.

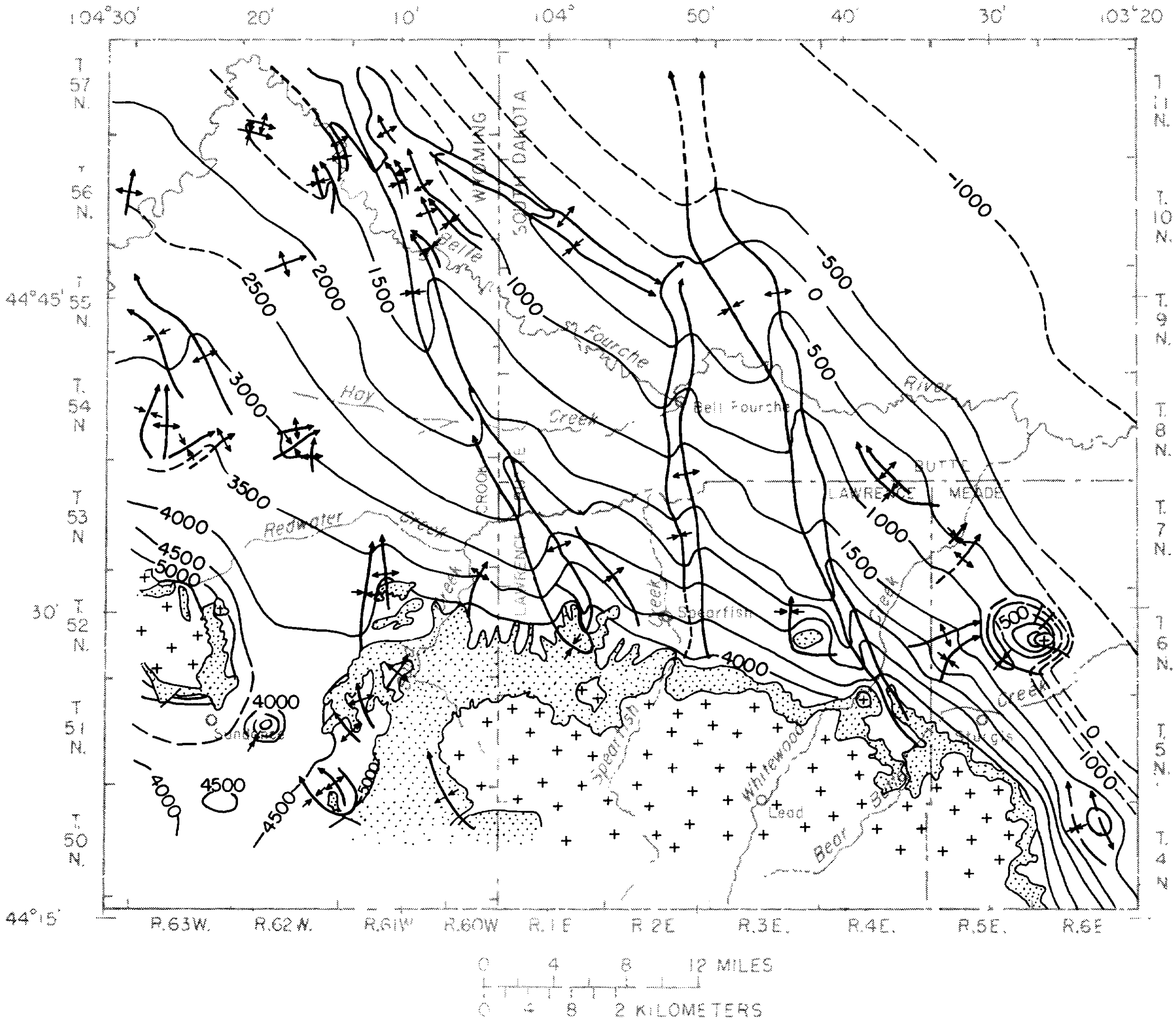
GEOHYDROLOGY

Geologic Setting

The Black Hills and Bear Lodge Mountains were formed by the Black Hills uplift and are geologically similar. The study area consists of isolated groups of mountains, generally formed by intrusives of Tertiary age, and ridges, plateaus, and upland valleys formed by eroded outcrops of sedimentary rocks. Precambrian metamorphic rocks are exposed in the Black Hills near the southern boundary of the study area. The geologic structure of the sedimentary units is illustrated by the structure-contour map of the Minnelusa Formation (fig. 4). The sedimentary rocks generally dip away from the Black Hills (fig. 5), toward the northeast and the Williston Basin, or, in the area south of the Bear Lodge Mountains, toward the southwest and the Powder River Basin outside the study area. The Inyan Kara, Minnelusa, and Madison aquifers are deepest in the Williston Basin in the northeast corner of the study area, where they are from 3,000 to 5,000 ft below land surface. The Williston Basin comprises plains of low relief, buttes, and badlands formed by erosion of shale and limestone. The shale, as much as 2,000 ft thick in the northeastern part of the study area, confines the Inyan Kara aquifer.

Folding, shown in figure 4, occurs in all sedimentary rock units older than the Tertiary White River Formation. The dominant trend of the folds is to the north and northwest. Many folds range from 20 to 40 mi in length and extend beyond the study area.

The structural features of the northern Black Hills and Bear Lodge Mountains may effect the hydrology. Secondary permeability caused by solution and fracturing is probably enhanced in folds. Detailed examination of geologic maps shows several small (less than 10 mi in length) folds. There is little subsurface data to accurately map these



EXPLANATION

OUTCROP OF MINNELUSA FORMATION

MINNELUSA FORMATION NOT PRESENT

—2500— STRUCTURE CONTOUR --Shows altitude of the top of the Minnelusa Formation. Dashed where approximately located. Contour interval 500 feet. Datum is sea level

ANTICLINE--Showing trace crestal plane and direction of plunge. Dashed where approximately located

SYNCLINE--Showing trace of trough plane and direction of plunge. Dashed where approximately located

MONOCLINE--Showing trace. Dashed where approximately located

FAULT

Figure 4.--Major geologic structures and altitude of the top of the Minnelusa Formation. (Structure modified from Lisenbee, 1985, and DeWitt, 1973. Geology modified from Darton, 1905; Darton and Paige, 1925; and Staatz, 1983, pl. 1).

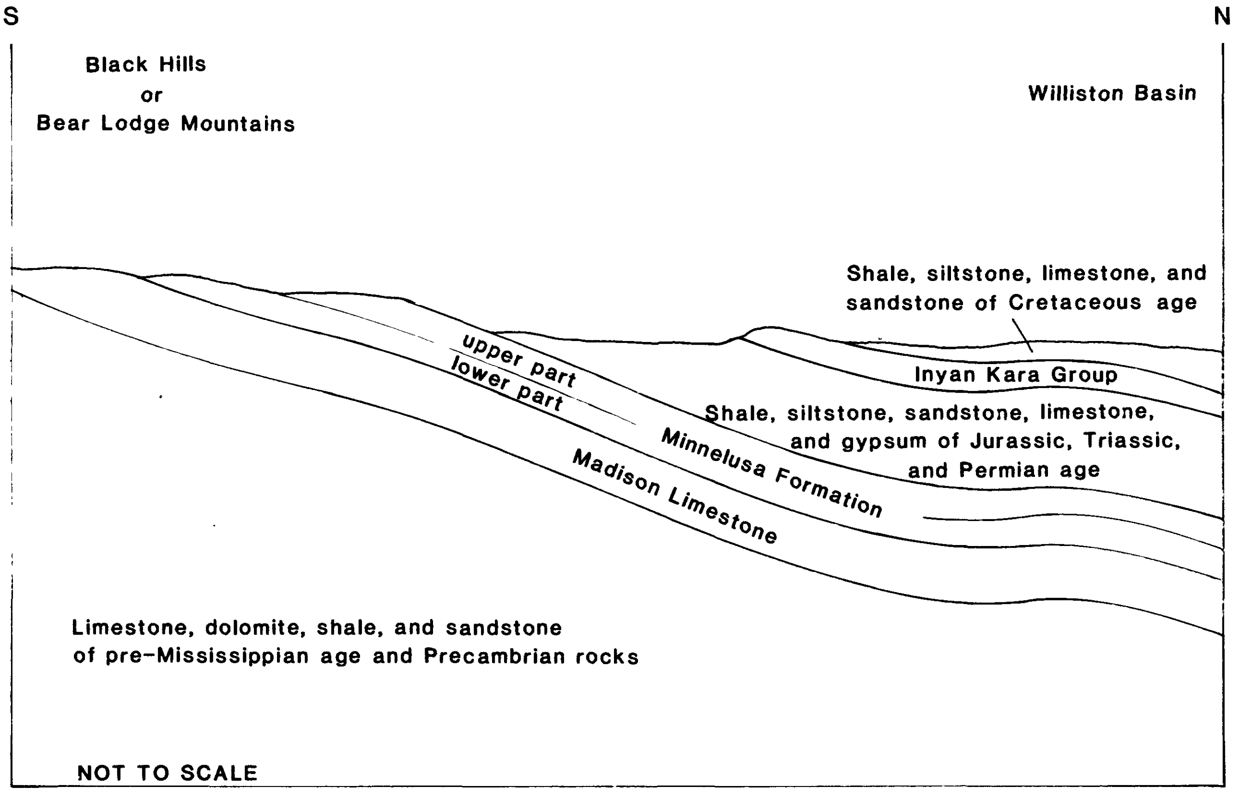


Figure 5.--Generalized geologic section.

small folds at depth. Of the large folds, the northern terminus of the Belle Fourche anticline, where it intersects the Colony anticline and is nearest the Whitewood anticline (fig. 6), may be of particular interest because the known large yields of the few wells penetrating the Madison in this area of deformation indicate there may be increased permeability from dissolution of the limestone along fractures in the folds. The surface geology shows an anticline-syncline pair north of this area in T. 10 and 11 N., R. 2 and 3 E., that appears to be an extension of the Whitewood anticline-syncline pair, however, there is no subsurface data to support this conclusion.

Aquifer Characteristics

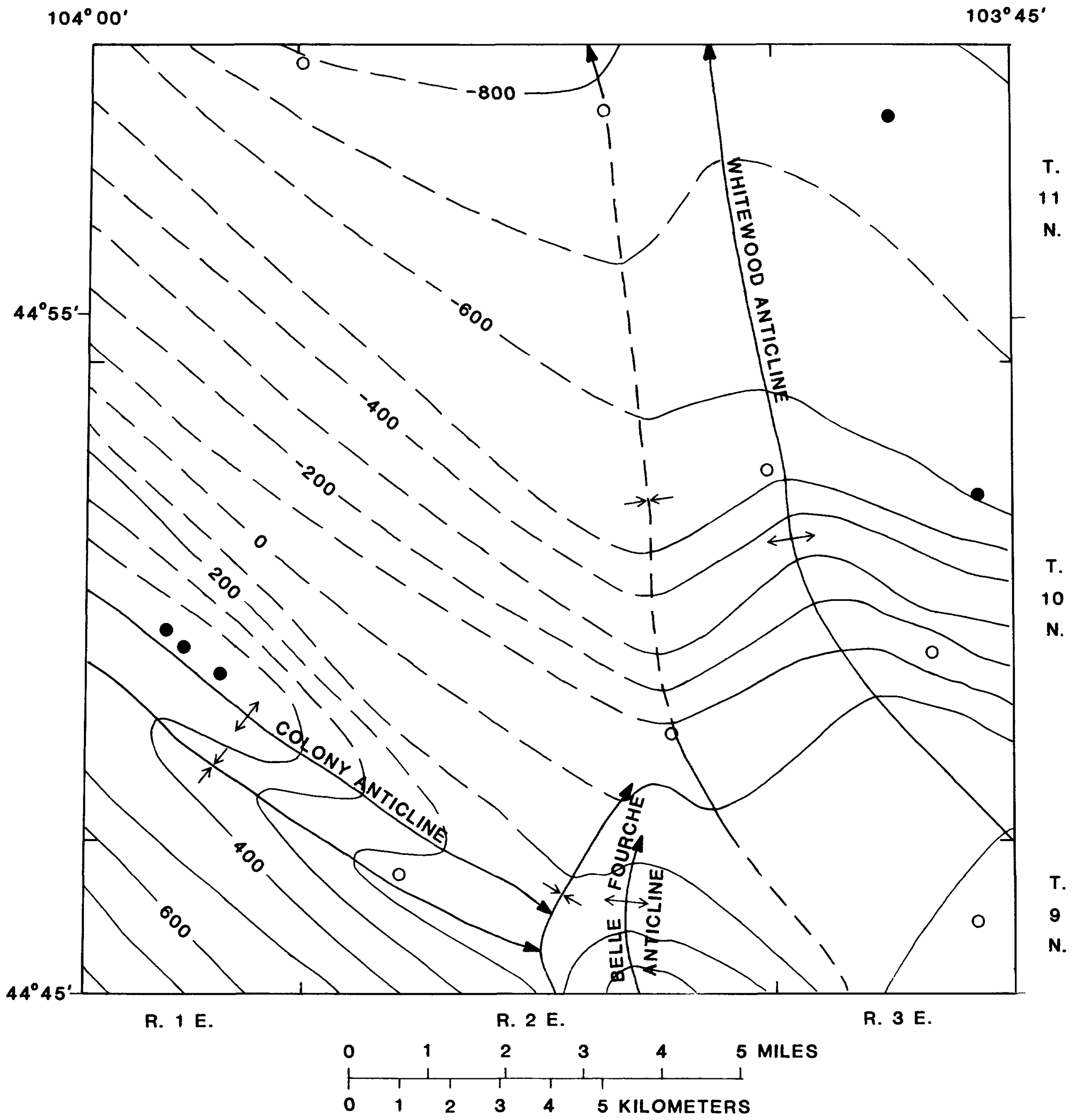
Inyan Kara Aquifer

The area of the outcrop of the Lower Cretaceous Inyan Kara Group, comprising the Fall River and Lakota Formations, is approximately 145 mi^2. The outcrop forms a ridge along the northeastern periphery of the Black Hills and a narrow plateau in the northern Bear Lodge Mountains. It is buried at depths exceeding 3,000 ft in the northeastern portion of the study area. The top of the Inyan Kara Group, where present, is about 1,500 ft higher, or nearer land surface, than the top of the Minnelusa (shown in fig. 4). The Inyan Kara Group is 81 to 475 ft thick in the study area. The erratic thickness of the Inyan Kara is caused mainly by variations in thickness of the Lakota Formation. The Lakota Formation "changes thickness within short distances and is characterized by complexly interfingering beds of conspiciously crossbedded sandstone, conglomeratic sandstone, and variegated claystone" (Robinson and others, 1964, p. 23). The Fall River Formation "has a fairly uniform thickness and is characterized by relatively persistent beds of evenly bedded sandstone, siltstone, and dark-gray shale" (Robinson and others, 1964, p. 23).

The sandstones in the Inyan Kara Group form the Inyan Kara aquifer. For this study, it was assumed the overlying Skull Creek Shale of Early Cretaceous age (table 1), in combination with other, younger, Cretaceous shales, is a confining layer for the Inyan Kara aquifer.

The Fall River Formation has been mistakenly called the Dakota Formation, as in Darton and Paige (1925), and locally the term "Dakota" is still used informally by well owners and some drillers. The geologic nomenclature presently accepted by the U.S. Geological Survey, shown in table 1, is based on investigations by Waage (1950).

The potentiometric surface of the Inyan Kara aquifer shows that the general direction of ground-water movement is to the northeast (fig. 7). With the exception of the outcrop area, most wells in the Inyan Kara will flow. The gradient is about 80 ft/mi near the outcrop and about 20 ft/mi elsewhere. The steeper gradient represents significant vertical flow in the aquifer, which is assumed to be insignificant when constructing potentiometric maps using wells that only partially penetrate the aquifer. Three-hundred fifty water-level measurements were used in constructing the potentiometric-surface map for the Inyan Kara aquifer. The wells were distributed throughout the study area except in the northeastern part, where there were few wells. There are four observation wells in the study area (table 2). The hydrographs for these wells are shown in figure 8 and their locations are included in figure 3. The two wells in Butte County, in T. 8 N., R. 3 E., Sec. 12, showed that the potentiometric surface fluctuated at least 20 ft and possibly more than 37 ft between 1970 and 1978. The two observation wells in Meade County show that the potentiometric surface declined more than 15 ft between 1957 and 1984 in T. 6 N., R. 5 E., Sec. 21 and 22.



EXPLANATION

-200 — STRUCTURE CONTOUR--Shows altitude of the top of the Minnelusa Formation. Dashed where approximately located. Contour interval 100 feet. Datum is sea level

ANTICLINE--Showing trace of crestal plane and direction of plunge. Dashed where approximately located

SYNCLINE--Showing trace of trough plane and direction of plunge. Dashed where approximately located

WELL OR TEST HOLE--

● Penetrating the Minnelusa Formation

○ Not penetrating the Minnelusa Formatinon

Figure 6.--Altitude of the top of the Minnelusa Formation in the vicinity of three anticlines. (Structures modified from Lisenbee, 1985.)

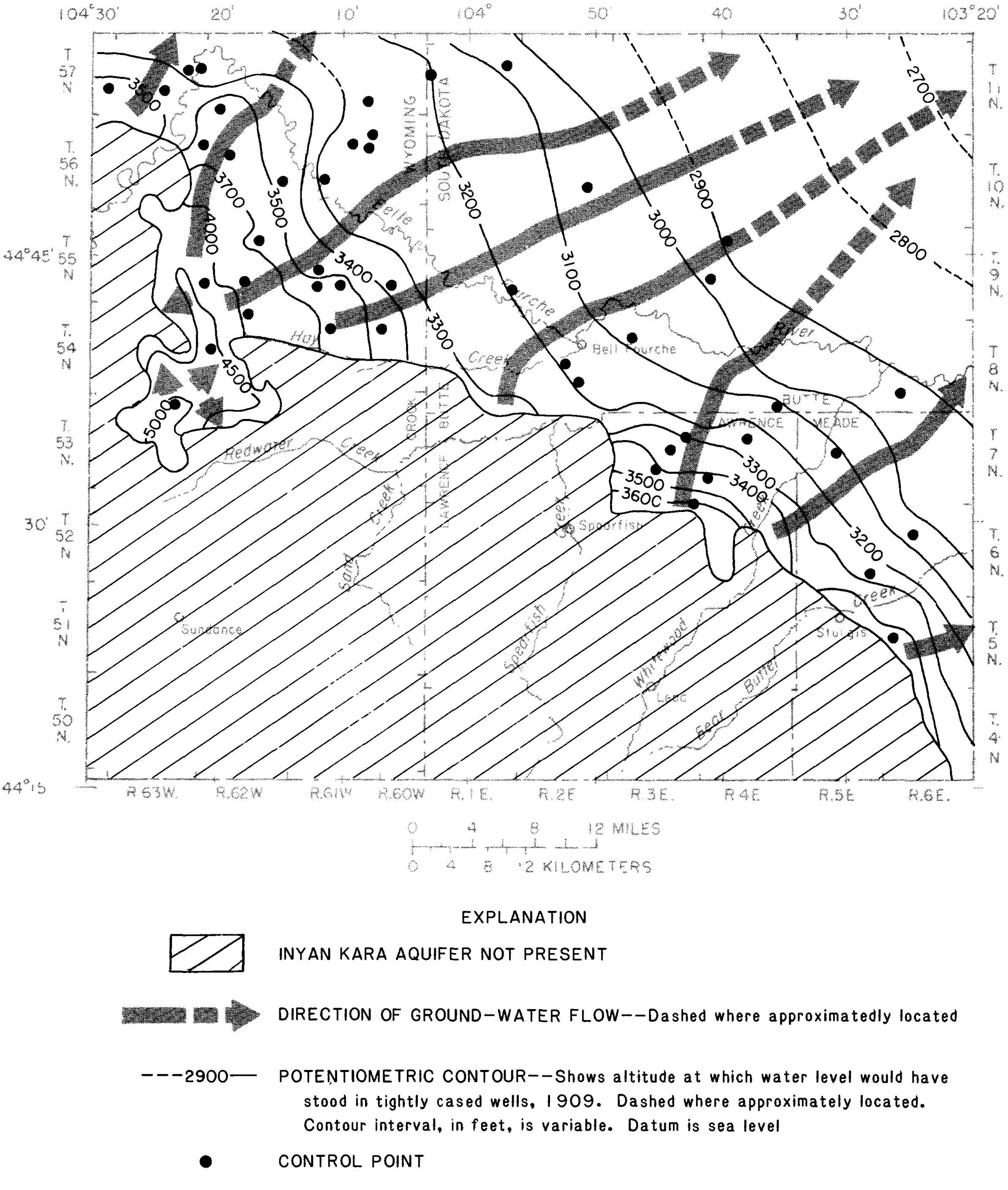


Figure 7.--Potentiometric surface and selected flowpaths of the Inyan Kara aquifer (1909). (Area where the aquifer not present modified from Darton, 1905; Darton and O'Harra, 1905; Darton and Paige, 1925; Robinson and others, 1964, p. 1; and Staatz, 1983, pl. 1.)

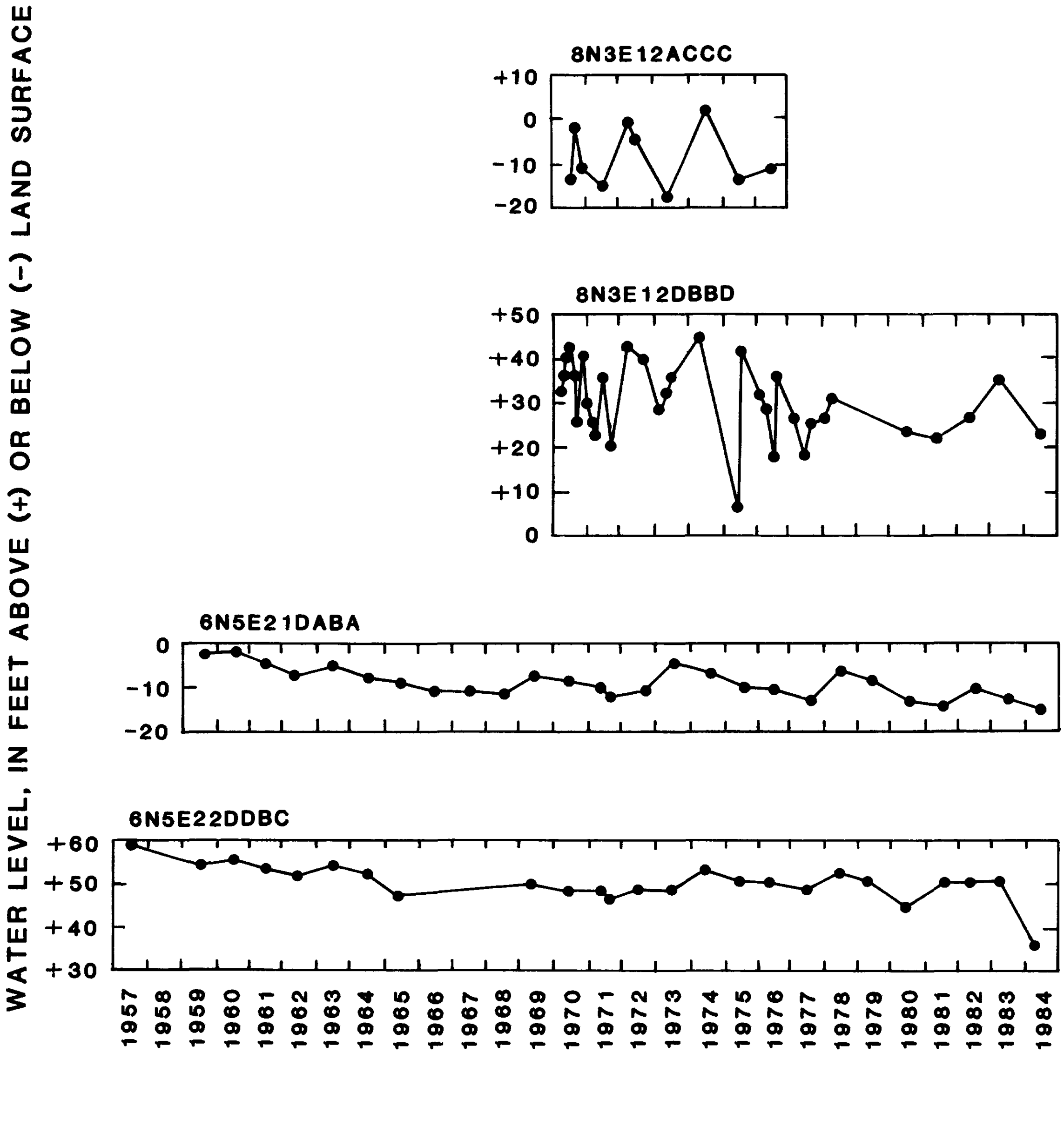


Figure 8.--Hydrographs of four observation wells in the Inyan Kara aquifer.

It is not possible to describe regional changes in the potentiometric surface without long-term observation wells distributed throughout the area. However, in some areas, comparisons of current water levels with historic water levels give an indication of the magnitude of decline of the potentiometric surface in those locations. The potentiometric surface of the Inyan Kara aquifer near Belle Fourche has declined since development began in the early 1900's (fig. 8). Measurements made before 1909 (Darton, 1909a and 1909b) of water levels in 10 wells within a 5-mi radius of Belle Fourche showed the potentiometric surface to be about 3,150 ft above sea level in 1909. Measurements made in 1982 and 1983 of numerous wells in the same area show the potentiometric surface to be less than 3,100 ft above sea level, indicating a decline of about 50 ft. In 1909 the water levels in two wells, 9N4E18DD and 9N4E19CA, indicated the potentiometric surface to be more than 3,025 ft above sea level (Darton, 1909b). The potentiometric surface in 1982 and 1983 in this area, based on measurements in newer wells, is 2,900 to 3,000 ft above sea level. This indicates water levels have declined as much as 125 ft in T. 9 N., R. 4 E. in response to withdrawals, particularly by uncontrolled flowing wells.

Small local-flow systems characterize the Inyan Kara aquifer near the outcrop area where water from precipitation infiltrates the aquifer, moves a short distance, and discharges at seeps and springs. The largest of these local systems is near the outcrop southwest of Hay Creek and northwest of Redwater Creek. This part of the aquifer, about 30 mi^2 in area, does not affect the regional flow system.

Hydraulic conductivity, transmissivity, and storage coefficient of the Inyan Kara aquifer have been estimated by previous investigators. Estimates of hydraulic conductivity from these previous investigations (table 3) range from 0 to 100 ft/d. Transmissivity estimates range from 50 ft^2/d (for the Fall River Formation alone) to 3,800 ft^2/d (table 3). Storage coefficient estimates range from 1×10^{-5} to 1×10^{-4} (for the Lakota Formation alone). Specific yield generally ranges from 0.1 to 0.3 (Lohman, 1979, p. 53-54) for unconfined aquifers. Presumably the specific yield of the Inyan Kara aquifer is within this range where it is an unconfined aquifer.

The principal use of water from the Inyan Kara aquifer in the study area is for domestic and stock supply. The Inyan Kara reportedly yields from 1 to 300 gal/min to wells. Discharges from flowing wells generally are less than 30 gal/min.

Table 3.—Estimates of hydraulic conductivity, transmissivity, storage coefficient, and vertical hydraulic conductivity from previous investigations and this study

Source	Hydraulic conductivity (feet per day)	Transmissivity (feet squared per day)	Storage coefficient (dimensionless)	Vertical hydraulic conductivity (feet per day)	Data source or method
Inyan Kara aquifer					
Niven (1967, p. 8)	0-100	—	—	—	Gas permeameter measurements of core taken parallel and perpendicular to bedding.
Miller and Rahn (1974, p. 230)	0.944	178	—	—	Aquifer test.
Gries and others (1976, p. 2, 7-9)	1.26	250 to 580	2.1×10^{-5} to 2.5×10^{-5}	—	Aquifer test.
Boggs and Jenkins (1980, p. 31)	—	50 (Fall River Formation) 190 (Lakota Formation)	1.4×10^{-5} (Fall River Fm.) 1.0×10^{-4} (Lakota Fm.)	— —	Aquifer test.
Bredehoeft and others (1983, p. 20)	8.3	—	1×10^{-5}	—	Model input.
H. L. Case III, U.S. Geological Survey Albuquerque, N. Mex., written commun., 1982	1.0	—	—	—	Model input.
	1.02	—	—	—	Specific capacity method of Meyer (1963).
J. S. Downey, U.S. Geological Survey, Denver, Colo., written commun., 1982	—	1,800 to 3,800	—	—	Model input.
This investigation	—	0.86 to 6,000	—	—	Model input.
Confining layer overlying Minnelusa aquifer					
H. L. Case III, U.S. Geological Survey, Albuquerque, N. Mex., written commun., 1982	—	—	—	9×10^{-8}	Model input.
J. S. Downey, U.S. Geological Survey, Denver, Colo., written commun., 1982	—	—	—	2.5×10^{-8}	Model input.
This investigation	—	—	—	1.1×10^{-7}	Model input.
Minnelusa aquifer					
Blankennagel and others (1977, p.50)	less than 2.4×10^{-5} to 1.4	—	—	—	Permeability test of core.

Pakkong (1979, p. 41)	—	880	—	—	Aquifer test.
Woodward-Clyde Consultants (1980, p. 4-12)	—	30 to 300	6.6×10^{-5} to 2.0×10^{-4}	—	Aquifer tests, flow and specific capacity data, permeability data, and lithologic considerations.
J. S. Downey, U.S. Geological Survey, Denver, Colo., written commun., 1982	—	700 to 1,000	—	—	Model input.
This investigation	—	0.86 to 8,600	—	—	
		Confining layer overlying Madison aquifer			
Blankennagel and others (1977, p. 50-51)	—	—	—	less than 2.4×10^{-5} to .01	Permeability test of core.
Downey (1982, p. 74)	—	—	—	5.0×10^{-7} to 7.0×10^{-7}	Model input.
This investigation	—	—	—	3.4×10^{-6} to 3.4×10^{-4}	Model input.
		Madison aquifer			
Konikow (1976, p. 41)	—	860 to 2,200	—	—	Flownet analysis and model, includes correction for temperature variation.
Miller (1976, p. 25)	—	0.01 to 5,400	—	—	Drill-stem tests in southeastern Montana.
Blankennagel and others (1977, p. 52-53)	2.4×10^{-5} to 1.9	—	—	—	Permeability test of core.
Woodward-Clyde Consultants (1980, p. 4-13)	—	3,000	2×10^{-4} to 3×10^{-4}	—	Aquifer tests, long-term response of aquifer to pumping in western Black Hills region, and model.
Blankennagel and others (1981, p. 50)	—	5,090	2×10^{-5}	—	Step-drawdown tests.
Downey (1982, p. 61)	—	250 to 1,500	—	—	The range given here is for northern Black Hills part of Downey's model of the northern Great Plains. A correction for temperature variation is included.
This report	—	4.3 to 8,600	—	—	Model input.

Minnelusa Aquifer

The Pennysivanian and Permian Minnelusa Formation is exposed over approximately 80 mi^2. The outcrop forms rolling hills and a high plateau in the Black Hills and southern Bear Lodge Mountains. Most of the outcrop area is in the Black Hills. The Minnelusa Formation is buried at depths exceeding 4,500 ft in the northeastern portion of the study area. The altitude of the top of the Minnelusa Formation is shown in figure 4. Thickness of the Minnelusa Formation ranges from about 340 to 800 ft, as shown in figure 9. The Formation is thinnest in the outcrop area, in the north-central part of the study area between the Whitewood and Colony anticlines, and where it has been cut by intrusives. In some areas, intrusives have entirely replaced the Minnelusa Formation. The upper part of the Minnelusa Formation is about 200 ft thick and is composed of sandstone, limestone, dolomite, and shale with local deposits of anhydrite and gypsum. The lower part of the Minnelusa Formation is similar in lithology to the upper part but has more limestone and dolomite. Dissolution and transport of the anhydrite and gypsum in the outcrop has resulted in brecciation of parts of the Minnelusa.

The sandstones in the upper part of the Minnelusa Formation form the Minnelusa aquifer. The hydraulic heads in the Minnelusa aquifer near the outcrop are generally 50 ft lower than the hydraulic heads in the Madison aquifer at the same place. The hydraulic head in the Minnelusa aquifer is more than 400 ft higher than that in the Madison aquifer east of the Black Hills, outside the study area, as measured in a well in the city of Box Elder. In most of the study area, the head relationship of the Minnelusa and Madison aquifers is not known. The lower part of the Minnelusa Formation, which has a smaller hydraulic conductivity than the upper part of the Minnelusa Formation (J. S. Downey, U.S. Geological Survey, Denver, Colo., written commun., 1982), was interpreted to be a leaky confining layer because of the difference in head in the Minnelusa and the Madison aquifers. The lower part of the Minnelusa Formation is a confining layer overlying the Madison aquifer. Fracturing from folding and brecciation near the outcrop may have increased the permeability of the lower part of the Minnelusa a considerable, but unknown, amount. The confining layer overlying the Minnelusa aquifer is about 1,300 ft thick and composed of the Permian, Triassic, and Jurassic-age shales and siltstones with some interbedded sandstone, limestone, and gypsum. The rock units in the confining layer include in descending order the Morrison Formation or Unkpapa Sandstone, Sundance, Gypsum Spring, Spearfish Formations, Minnekahta Limestone, and Opech Formation (table 1).

The general direction of ground-water movement in the Minnelusa aquifer is to the east and northeast (fig. 10). Wells flow in most of the study area in South Dakota and in approximately the northern one-half of Crook County, Wyoming, that is in the study area. The gradient of the potentiometric surface is about 120 ft/mi near the outcrop and about 10 to 40 ft/mi elsewhere. The steep gradient implies there may be significant vertical flow in the aquifer. The gradient is difficult to interpret using observation wells that may only penetrate the top portion of the aquifer. The potentiometric-surface map was constructed using 175 water-level measurements. Most of the wells are located in the southern one-half of the study area, close to the outcrop, with the remainder of the wells distributed sparsely in the northern one-half of the study area where the depth to the aquifer is greater.

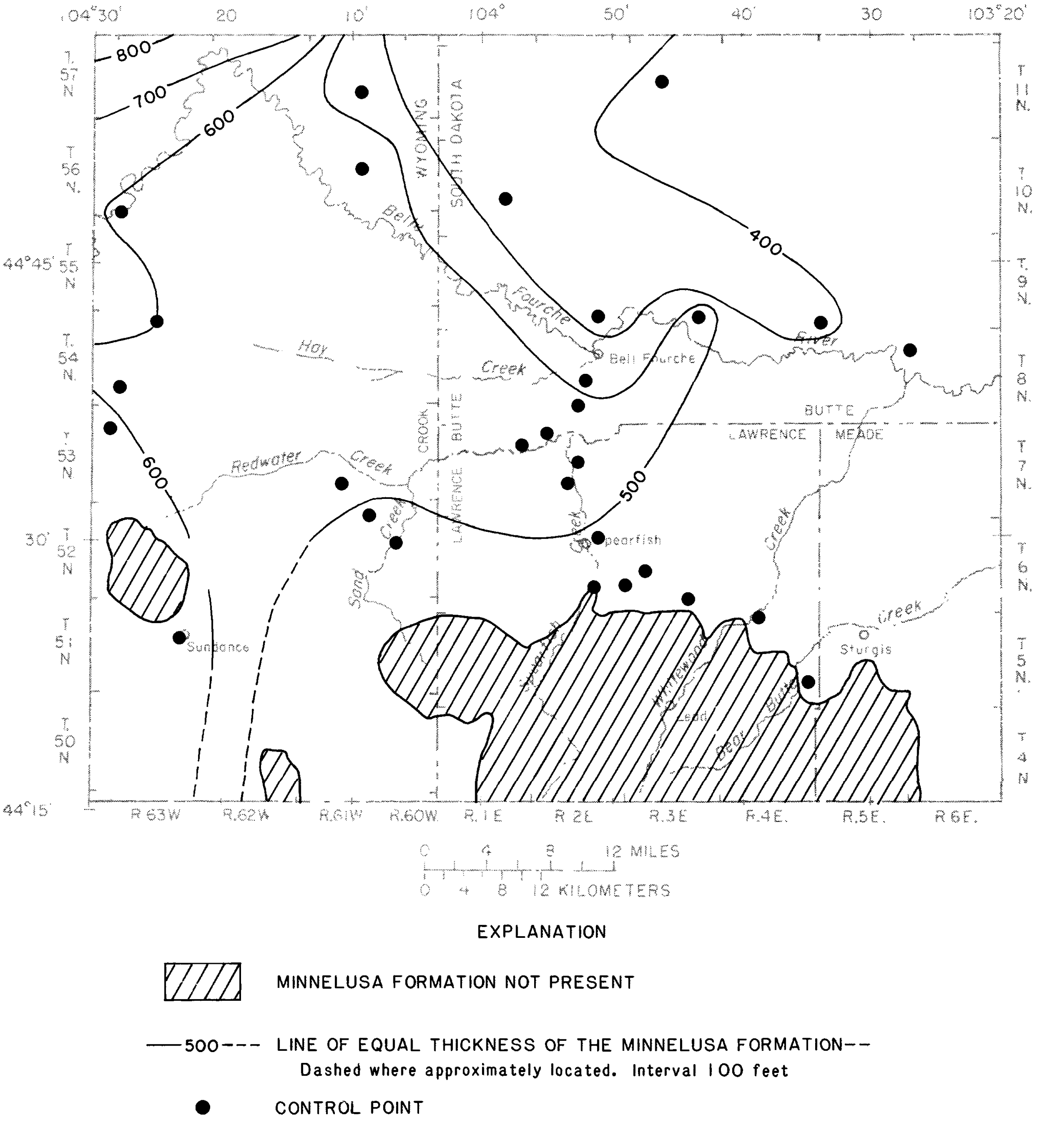
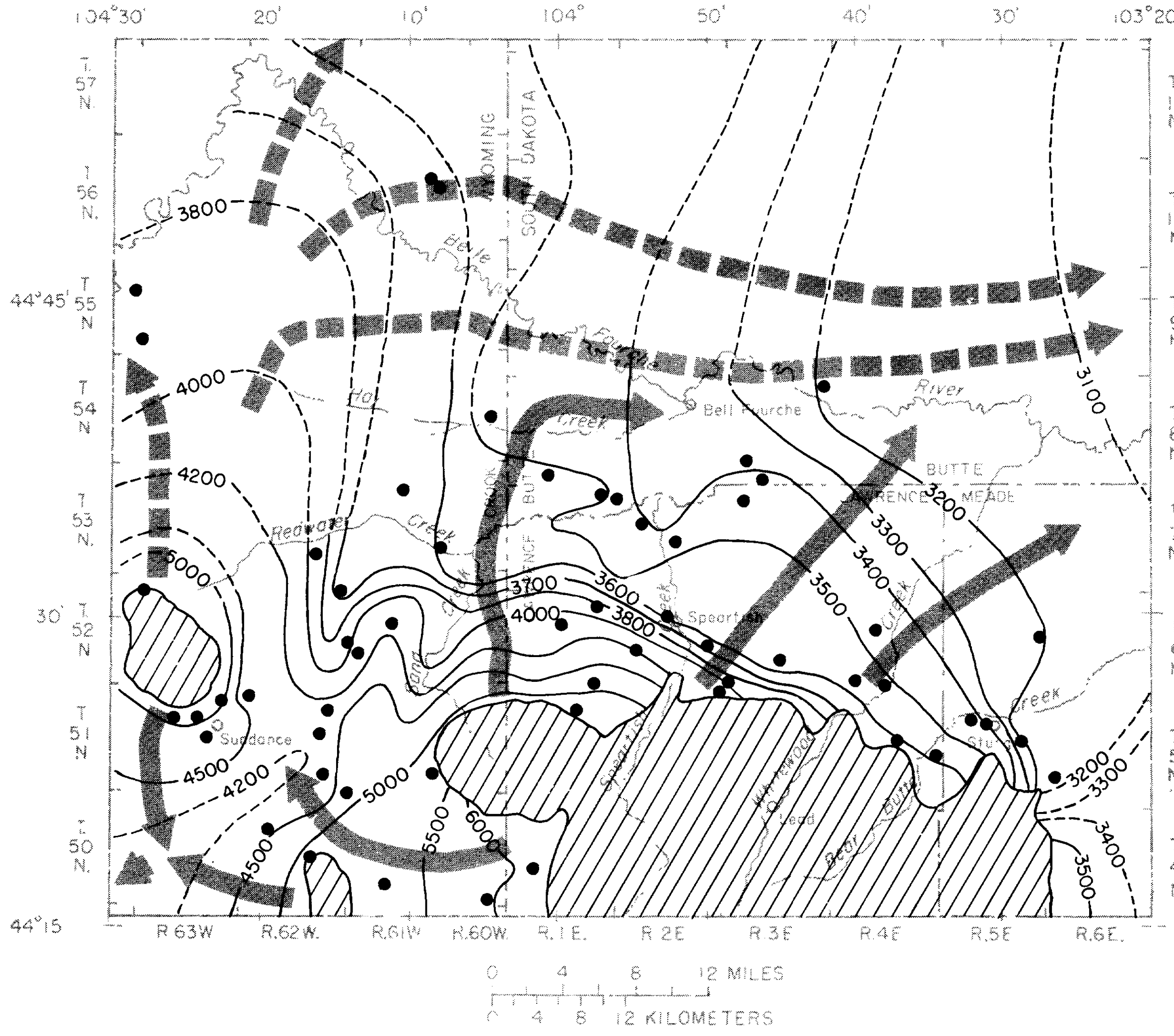


Figure 9.--Thickness of the Minnelusa Formation. (Area where the aquifer not present modified from Darton, 1905; Darton and Paige, 1925; and Staatz, 1983, pl. 1. Thicknesses from unpublished geologic logs on file at the U.S. Geological Survey office, Rapid City, S. Dak.; Darton, 1909b, p. 24; and Darton and Paige, 1925, p. 8.)



EXPLANATION

MINNELUSA AQUIFER NOT PRESENT

DIRECTION OF GROUND-WATER FLOW--Dashed where approximately located

——3600--- POTENTIOMETRIC CONTOUR--Shows altitude at which water level would have stood in tightly cased wells, 1960. Dashed where approximately located. Contour interval, in feet, is variable. Datum is sea level

● CONTROL POINT

Figure 10.--Potentiometric surface and selected flowpaths of the Minnelusa aquifer (1960). (Area where the aquifer not present modified from Darton, 1905; Darton and Paige, 1925; Staatz, 1983, pl. 1.)

Water levels in wells in the Minnelusa aquifer have declined as much as 99 ft in the vicinity of Spearfish. A 415-ft deep observation well about 1.5 mi south of town, 6N2E23BBBA, was flowing in 1907, with a water level of 15 ft above land surface (Davis and others, 1961, p. 98). By 1956, the water level had declined to 74 ft below land surface. From 1956 to 1980, the water level fluctuated from 36 to 84 ft below land surface (Bradford, 1981, p. 57), but did not show a declining trend (fig. 11). In the 1930's, a Spearfish Fish Hatchery spring, 6N2E15BDD, which probably was fed by the Minnelusa and Madison aquifers, stopped flowing (Allan Sandoval, U.S. Fish and Wildlife Service, Denver, Colo., personal commun., 1984). The system probably has established local equilibrium, as indicated by relatively constant water levels since 1950.

The decline in head probably was caused by reduced recharge to the Madison and Minnelusa aquifers in the early 20th century. Most of the flow of Spearfish Creek has been diverted since the turn of the century around the outcrop of the Minnelusa Formation and the underlying Madison Limestone and piped directly to the town of Spearfish, where the water is used for city supply and generation of electricity. Part of the water is returned to the creek in town and maintains the base flow. Most of the flow in creeks in the Black Hills infiltrates the Madison and Minnelusa outcrops and provides recharge to the aquifers. During a period of high flow on Spearfish Creek in May 1984, when the creek flowed its entire length, discharge decreased 29 ft^3/s across a reach underlain by gravels and the Madison aquifer. Discharge decreased 14 ft^3/s across a reach underlain by gravels and the Minnelusa aquifer. It is not known what part of this loss was recharging the Madison and Minnelusa aquifers, however, it is certain that in its undiverted state, streamflow in Spearfish Creek was recharging the aquifers in some amount. The average discharge of Spearfish Creek at Spearfish, below the return flows, for 37 years is 52.7 ft^3/s (U.S. Geological Survey, 1984, p. 82). Other than during peak flows in the spring, this discharge represents the aforementioned return flows. Thus, the average flow above the diversions is probably also about 50 ft^3/s. If there were no diversion the creek probably would flow perennially for most or all of the reach underlain by the Madison and Minnelusa aquifers and recharge these aquifers. However, if there were no diversion, streamflow would be less downstream of these outcrops. The diversions have reduced recharge to the aquifers and the potentiometric surface has declined, as shown by the change in the water level in the well in 6N2E23BBBA (fig. 11). The lowered heads dried up the spring at the fish hatchery and reduced outflow from the area and leakage to other aquifers some unknown amount, thus discharge from the aquifer was reduced.

Hydraulic conductivity and transmissivity of the Minnelusa aquifer and the vertical hydraulic conductivity of the overlying confining layer, have been estimated by previous investigators (table 3). Estimates of hydraulic conductivity by Blankennagel and others (1977, p. 50) range from less than 2.4×10^{-5} to 1.4. Estimates of transmissivity values range from 30 to 1,000 ft^2/d (table 3). The storage coefficient in the study area was estimated to be 6.6×10^{-5} to 2.0×10^{-4} by Woodward-Clyde Consultants (1980, p. 4-12). Estimates of vertical hydraulic-conductivity values of this confining layer range from 2.5×10^{-8} to 9.0×10^{-8} ft/d (table 3). Presumably the specific yield of the Minnelusa aquifer ranges from 0.1 to 0.3 where it is unconfined (Lohman, 1979, p. 53-54).

The principal use of water from the Minnelusa aquifer in the study area is for municipal, domestic, and stock supply. Reportedly, the Minnelusa aquifer yields in excess of 1,000 gal/min to wells in some areas. Well 7N1E11ABD flowed about 1,300 gal/min in 1967. From 1967 to 1978, the water level fluctuated between 87.9 and 134.8 ft above land surface (Bradford, 1981, p. 19) as shown by the hydrograph in

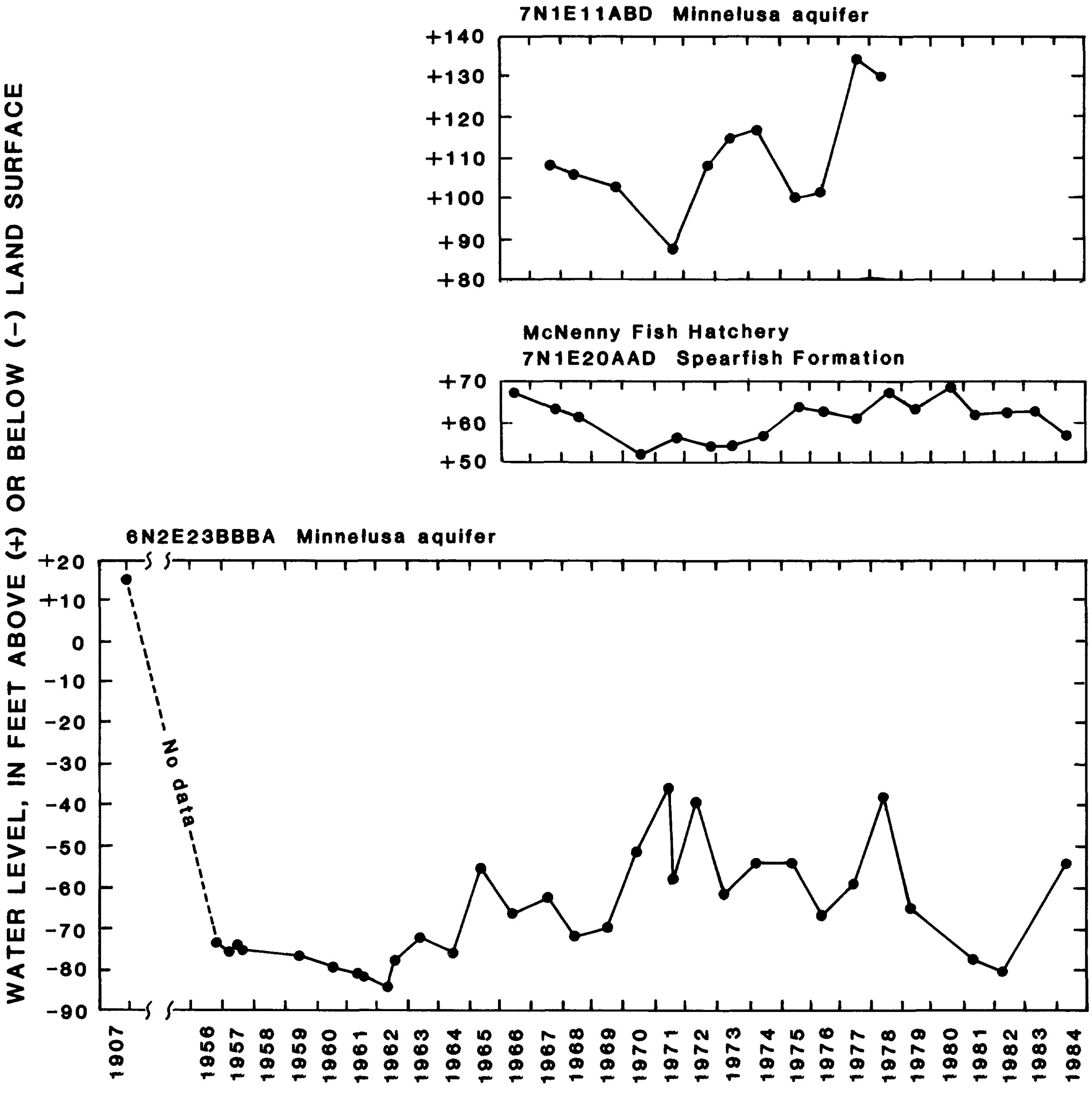


Figure 11.--Hydrographs of three observation wells.

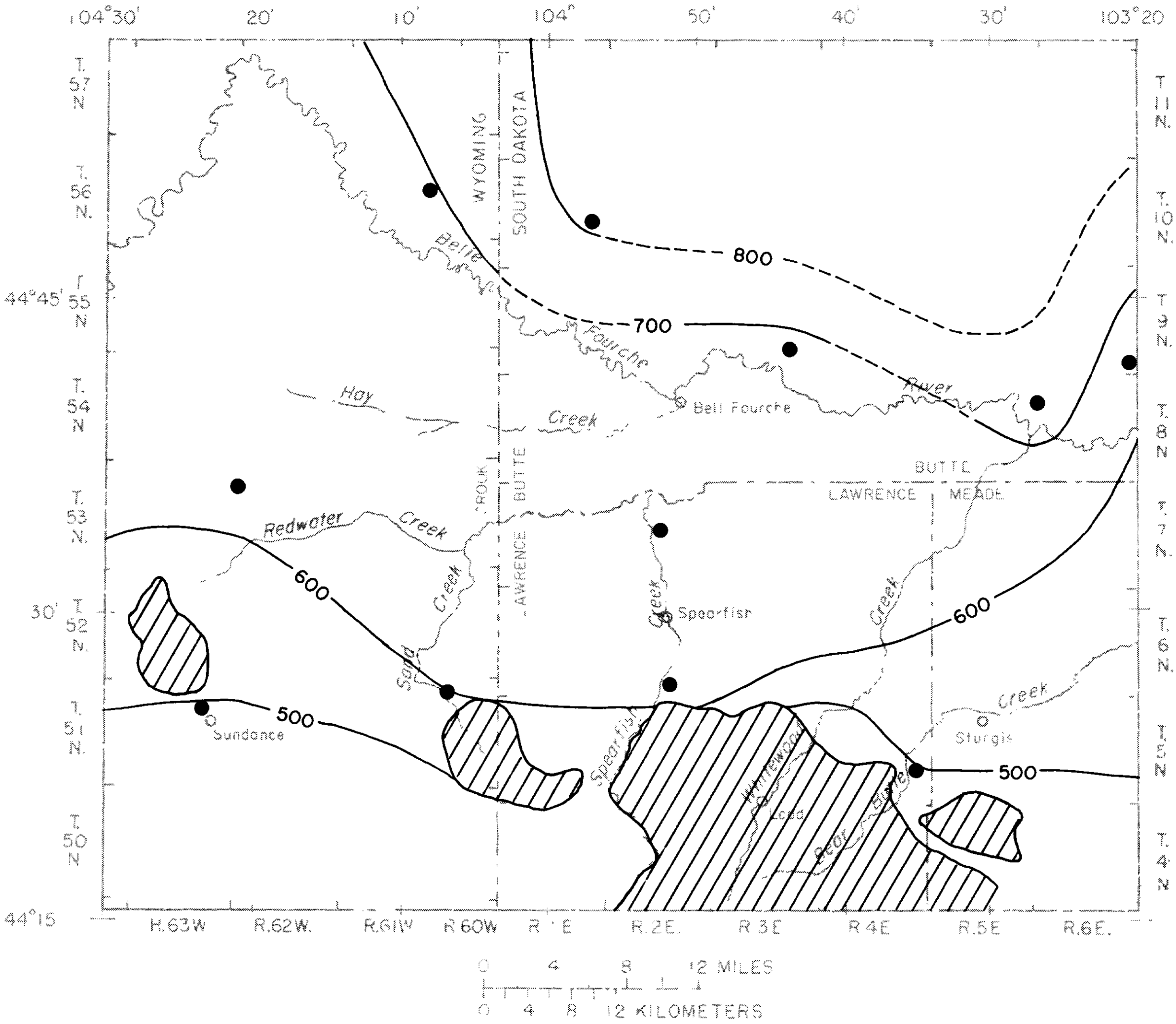
figure 11, with no significant trend. Well 7N2E21AABC is reported to flow 1,000 gal/min. Well 7N1E20AAD, at the McNenny Fish Hatchery, flowed, when drilled, 1,600 gal/min. This well is completed in the Spearfish Formation, however, the water supplying the well and nearby springs probably leaks upward from the Minnelusa aquifer. There are similar springs, issuing from the Spearfish Formation or the Minnekahta Limestone, in other areas of the Black Hills. It is known, from dye tests and detailed potentiometric maps, that those springs are fed by leakage from the Minnelusa and the Madison aquifers (Rahn and Gries, 1973). When the McNenny Fish Hatchery well was completed, discharge from one of the nearby springs decreased from 600 to 200 gal/min (unpublished U.S. Geological Survey records, Rapid City, S. Dak., Subdistrict office). Though this well has a large discharge, the aquifer system adjusted by decreasing spring discharge, and the system appears to have returned to equilibrium, as indicated by the water levels in this McNenny Fish Hatchery well. As shown in figure 11, the water levels fluctuated between 54.2 and 69.1 ft above land surface, showing no consistent decline, from 1966 to 1984 (Bradford, 1981, p. 57, and unpublished U.S. Geological Survey records, Rapid City, S. Dak., Subdistrict office), indicating the system locally has returned to equilibrium.

The high yields of some wells completed in the Minnelusa aquifer, the presence of springs fed by the Minnelusa, and the relative stability of water levels in areas of development indicate the Minnelusa has the potential for additional development. As shown by the effect of reduced recharge from Spearfish Creek on the water levels in Spearfish and by the effect of the McNenny Fish Hatchery well or spring discharge, increased well withdrawals will affect the area. For example, water levels will decline and spring discharges may decrease measurably. Additional investigations would assist the planning and evaluation of increased ground-water withdrawals from the Minnelusa.

Madison Aquifer

The Mississippian Madison Limestone outcrop forms a high plateau covering approximately 35 mi^2 in the Black Hills and southern Bear Lodge Mountains. It is buried about 5,000 ft below land surface in the northeastern part of the study area. The top of the Madison Limestone is about 370 to 800 ft lower than the top of the Minnelusa Formation (fig. 5). The Madison Limestone thickness ranges from 500 ft in the southernmost part of the area to 850 ft in the northeastern corner of the study area (fig. 12). It is thinner than 500 ft in the outcrop and where it has been cut by intrusives. The Madison is an aquifer where fractures or solution features have increased the permeability of the limestone and dolomite beds.

The limestone, dolomite, and shale beds of the lower part of the Minnelusa Formation (table 1) act as a leaky confining layer separating the Minnelusa and Madison aquifers. It is assumed that the underlying confining bed for the Madison aquifer consists of the Englewood Formation of Devonian and Mississippian age, a dolomitic siltstone that appears to have a lower vertical hydraulic conductivity than the Madison aquifer (Joe S. Downey, U.S. Geological Survey, Denver, Colo., written commun., 1982). There are no wells completed solely in the Englewood Formation so this assumption can not be tested, however, typically springs discharge at the base of the Madison Limestone cliffs at the upper contact of the Englewood, and streams appear to lose little or no flow as they cross the Englewood, though they lose some or all of their flow as they cross the Madison Limestone outcrop.



EXPLANATION

MADISON LIMESTONE NOT PRESENT

——800--- LINE OF EQUAL THICKNESS OF THE MADISON LIMESTONE-- Dashed where approximately located. Interval 100 feet

● CONTROL POINT

Figure 12.--Thickness of the Madison Limestone. (Area where the formation is not present modified from Darton, 1905; Darton and Paige, 1925; and Staatz, 1983, pl. 1. Thicknesses from unpublished geologic logs on file at the U.S. Geological Survey office, Rapid City, S. Dak.; Darton and Paige, 1925, p. 7-8; and Darton, 1905, p. 2).

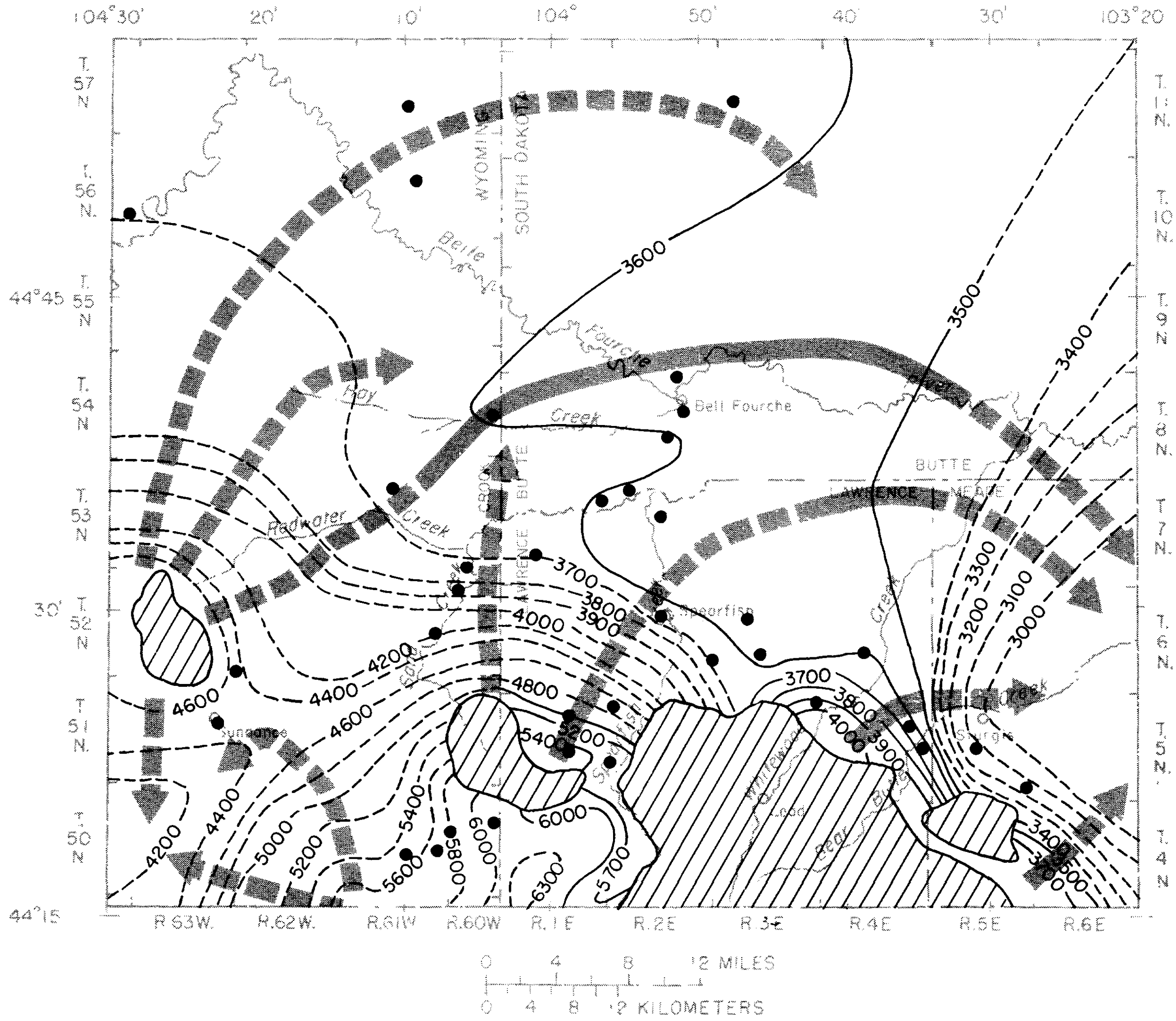
The potentiometric-surface map of the Madison aquifer shows the general direction of ground-water movement to be to the east and southeast (fig. 13). The potentiometric contours are based on forty water-level measurements, mostly located near the outcrop. The map is not as reliable as the potentiometric-surface maps for the Inyan Kara and Minnelusa aquifers because of the smaller number and poorer distribution of data points. The ground-water gradient is about 80 ft/mi near the outcrop and about 10 to 30 ft/mi elsewhere, except along the southeastern edge of the study area, where the gradient steepens to about 80 ft/mi. Wells flow in most of the study area in South Dakota and in approximately the northern one-half of Crook County, Wyoming.

There has been a reduction in recharge from infiltration of Spearfish Creek streamflow on the Madison aquifer outcrop because of diversion of flow. As a result, the potentiometric suface of the Madison aquifer has probably declined in the vicinity of Spearfish. The amount of reduction is not known because there are no observation wells in the Madison in this area.

Hydraulic conductivity and transmissivity of the Madison aquifer and vertical hydraulic-conductivity estimates for the overlying and underlying confining beds have been estimated by previous investigators (table 3). Estimates of hydraulic conductivity by Blankennagel and others (1977, p. 52-53) range from 2.4×10^{-5} to 1.9 ft/d. Estimates of transmissivity range from 0.01 to 5,400 ft^2/d in the study area (table 3). The wide range of values would be expected for an aquifer composed of fractured limestone. Storage coefficient was estimated to be 2×10^{-4} to 3×10^{-4} by Woodward-Clyde Consultants (1980, p. 4-13) and 2×10^{-5} by Blankennagel and others (1981, p. 50). Vertical hydraulic-conductivity estimates vary from 5.0×10^{-7} to 0.034 ft/d for the confining bed overlying the Madison aquifer. Downey (1982, p. 73) estimated the vertical hydraulic conductivity of the underlying confining bed to be from about 5×10^{-7} to 15×10^{-6} ft/d in the study area. Presumably the specific yield of the Madison aquifer where it is unconfined ranges from 0.1 to 0.3 (Lohman, 1979, p. 53-59).

The principal use of water from the Madison aquifer in the study area is for municipal supply. The Madison aquifer yields in excess of 500 gal/min to wells in some parts of the study area. The municipal well in Belle Fourche flowed 650 gal/min when completed in 1981. The static water level was 404 ft above the top of the casing, thus the specific capacity is about 1.6 (gal/min)/ft. The well with the greatest yield in the study area was drilled for oil exploration in 9N3E27ADBD. A discharge of 2,000 gal/min and a static water level of 590 ft above land surface were measured in 1952, thus the specific capacity is about 3.4 (gal/min)/ft. At the time, the well was open to both the Minnelusa aquifer and the upper 350 ft of the Madison aquifer (J. P. Gries, South Dakota School of Mines and Technology, Rapid City, S. Dak., written commun.). It is not known what portion of the discharge came from each aquifer. This well is on the Whitewood anticline and the high yield is caused by increased permeability due to fracturing and dissolution on the fold. The Belle Fourche municipal well is in the Belle Fourche syncline and though there may be less open fracturing in a syncline, there may be sufficient solution of the rock material to have increased permeability and, thus, increased well yield.

The large yields of wells in the Madison aquifer, particularly near geologic structures, suggest the Madison has the potential for additional development. However, development may effect other water supplies hydraulically related to the Madison. As demonstrated by the relationship of a reduction in streamflow in Spearfish Creek on the Madison aquifer outcrop and the drying up of the spring in Spearfish, the Madison may be a source of recharge for shallower aquifers and springs. The relationship between



EXPLANATION

MADISON AQUIFER NOT PRESENT

DIRECTION OF GROUND-WATER FLOW--Dashed where approximatly located

——3600--- POTENTIOMETRIC CONTOUR--Shows altitude at which water level would have stood in tightly cased wells, 1983. Dashed where approximately located. Contour interval, in feet is variable. Datum is sea level

● CONTROL POINT

Figure 13.--Potentiometric surface and selected flowpaths of the Madison aquifer (1983). (Area where the aquifer not present modified from Darton, 1905; Darton and Paige, 1925; and Staatz, 1983, pl. 1.)

the Minnelusa and Madison aquifers is not well understood, partly because the potentiometric surface of the Madison is not mapped adequately to evaluate head relationships. Further investigations of the Madison aquifer, including installation of observation wells, would aid the evaluation of the potential supply of the Madison aquifer.

WATER QUALITY

Water quality is determined by the chemical and physical properties of water that effect its suitability for a particular use. Water that is unsuitable for use in a public water system because of its quality could be acceptable for another use, such as irrigation or a certain industrial process. Criteria for water to be used in public water systems have been established by the U.S. Environmental Protection Agency (1975-80, 1979).

The National Interim Primary Drinking Water Regulations (U.S. Environmental Protection Agency, 1975-80) are mandatory criteria for public water systems establishing the maximum permissible levels of constituents known to be either toxic or injurious to health. The constituents listed under the heading of "mandatory (primary)" in table 4 are in this category. All may be toxic to humans at levels in excess of the maximum permissible level established by the U.S. Enviromental Protection Agency, except fluoride. Excessive concentration of fluoride produces objectionable dental fluorosis, or mottling of teeth. Excessive concentrations of nitrate in water, though not toxic to adults, can cause methemoglobinemia or blue baby disease if ingested directly or indirectly by infants.

The National Secondary Drinking Water Regulations (U.S. Environmental Protection Agency, 1979) are recommended criteria for public water systems. The maximum recommended levels of the constituents, listed under the heading of "recommended (secondary)" in table 4, were established because excessive concentrations are undesirable, although not toxic. Excessive concentration of chloride, copper, iron, or zinc imparts an undesirable taste to water. Iron or manganese in excessive concentrations stains porcelain and laundry. If the pH of water is outside the recommended range, it can interfere with water treatment. If the pH is too low, it can be corrosive to pipes. Excessive concentration of sulfate is undesirable because it can have a laxative effect on people not accustomed to drinking the water. Water with a concentration of dissolved solids exceeding the maximum recommended level generally has a concentration of a specific constituent, such as sulfate, which exceeds the maximum permissible or recommended level.

A summary of water-quality analyses for samples from the Inyan Kara, Minnelusa, and Madison aquifers as related to maximum permissible and recommended levels established by the U.S. Environmental Protection Agency (1975-80, 1979) is presented in table 4. Table 4 also shows data for hardness, sodium, bicarbonate, and uranium. Although no maximum permissible or recommended levels have been established for these constituents, they may affect the use of the water for public water systems or irrigation. Constituents which exceed the maximum permissible or recommended levels in at least one analysis are fluoride, selenium, gross alpha radiation, dissolved solids, iron, manganese, and sulfate. Though the concentration of dissolved solids in water from the Inyan Kara, Minnelusa, and Madison aquifers exceeds the maximum recommended level in some areas, the actual deterrent to use is the concentrations of specific constituents, discussed individually later in the report.

Table 4.--Concentrations of selected constituents and values of properties that may affect use of water from the bedrock aquifers

[Concentrations in micrograms per liter except as indicated. <1, less than 1]

Constituent	Maximum level allowed for community water systems 1/	Range of concentrations or values (number of analyses given in parentheses; concentrations exceeding the maximum primary or secondary drinking-water standards levels are underlined) Inyan Kara aquifer		Minnelusa aquifer		Madison aquifer	
Mandatory (Primary)							
Arsenic	50	< 1	(11)	2-5	(6)	< 1-4	(7)
Barium	1,000	0-100	(6)	30-300	(4)	50-300	(7)
Cadmium	10	0-2	(6)	1-2	(4)	0-3	(8)
Chromium	50	0	(6)	0-20	(4)	0-20	(7)
Fluoride, milligrams per liter	2/ 2.0-2.2	0-1.5	(93)	0-2.6	(106)	0.1-3.6	(19)
Lead	50	0-15	(6)	0-4	(4)	0-29	(8)
Mercury	2	<0.1	(11)	0-.2	(6)	0-.3	(6)
Nitrate as nitrogen (N), milligrams per liter	10	<.1-.6	(11)	0-1.3	(9)	0.09-.13	(2)
Selenium	10	0-23	(11)	1-4	(6)	1.0-8	(7)
Silver	50	—		—		0	(1)
Radium-226, picocuries per liter	3/ 5	0.28-2.9	(11)	0.1-1.2	(4)	1.3	(1)
Gross alpha radiation, picocuries per liter as natural uranium	4/ 15	5.6-23	(6)	3.3-5.6	(4)	3-140	(6)
Recommended (Secondary)							
Chloride, milligrams per liter	250	<0.1-170	(87)	0-61	(127)	0-67	(36)
Copper	1,000	0.0-2	(6)	0-14	(4)	0-7	(7)
Dissolved solids, milligrams per liter	500	238-2,040	(87)	128-3,160	(113)	190-2,750	(24)
Iron	300	20-2,400	(14)	10-740	(8)	10-7,900	(8)
Manganese	50	20-140	(8)	0-1	(5)	1-150	(7)
pH, standard units	6.5-8.5	6.8-8.5	(75)	6.8-8.2	(114)	6.8-8.4	(30)
Sulfate, milligrams per liter	250	8.0-1,200	(89)	0-2,000	(129)	4.8-1,700	(38)
Zinc	5,000	20-60	(6)	5-590	(4)	20-840	(7)
Constituents which may affect use but for which there is no limit							
Hardness as $CaCO_3$, milligrams per liter		1-810	(86)	8-4,700	(129)	7-1,800	(39)
Sodium, milligrams per liter		1-540	(79)	0.05-300	(122)	0.2-45	(37)
Bicarbonate, milligrams per liter		170-460	(54)	130-366	(53)	110-390	(23)
Uranium		0.4-12	(5)	2.5-10	(2)	3.7	(1)

1/ U.S. Environmental Protection Agency, 1975-80 and 1979.
2/ The maximum level allowed for fluoride depends on the annual average of the maximum daily air temperature at the location of the consumer. The annual average of the maximum daily air temperature in the study area ranges from 12.7 to 16.3 °C (U.S. Dept. of Commerce, 1983).
3/ Maximum permissible level is 5 picocuries per liter for radium-226 and radium-228 combined.
4/ Limit is for gross alpha particle activity including radium-226 but excluding radon (a gas produced by radium-226 decay) and uranium.

Water temperature affects use but there is no maximum permissible or recommended temperature. The available heat stored in the ground water may be used for heating. Alternatively, ground water can also be used for cooling. Water temperatures vary from about 7°C to as much as 56°C. Water temperature at shallow depths is usually at about the average annual air temperature of the area. As the water moves downgradient, to deeper parts of an aquifer, it is warmed by the naturally occurring heat of the earth. In general, the deeper the well, the hotter the water. A 4,858-ft deep flowing water well in T. 12 N., R. 3 E. had a reported bottom hole temperature of 56°C (unpublished U.S. Geological Survey records, Rapid City, S. Dak.). The average annual air temperature of the area is about 8°C (U.S. Department of Commerce, 1982). Therefore, the geothermal gradient at this well was about 1.0°C per 100 ft, which is typical for the Black Hills area. Water from deep wells may require some kind of cooling before use in public water systems.

Inyan Kara Aquifer

The maximum permissible or recommended levels allowed for public water systems for selenium, gross alpha radiation, dissolved solids, iron, manganese, and sulfate are exceeded in water from the Inyan Kara aquifer in some parts of the study area. There were less than 12 analyses for most of the primary constituents. These constituents rarely have excessive concentrations in naturally occurring ground water and are not known to generally exceed the maximum permissible or recommended levels in water from the Inyan Kara aquifer in South Dakota.

Only one of the water analyses, from 11 wells, for selenium showed a concentration exceeding the maximum permissible level. In most areas the concentration of selenium in water from the Inyan Kara aquifer is less than the maximum permissible level allowed for public water systems.

In parts of the study area, the level of gross alpha radiation in water from the Inyan Kara aquifer exceeds the maximum permissible level allowed for public water systems. However, the measurements are of gross alpha radiation from all sources, including uranium, and the maximum permissible level is established for radiation excluding uranium (U.S. Environmental Protection Agency, 1975-80). The Inyan Kara Group is a uranium-bearing rock unit (Robinson and others, 1964, p. 121) and water in the Inyan Kara aquifer does have dissolved uranium and radium-226, a product of the decay of uranium. There is no maximum permissible or recommended level for uranium established by the U.S. Environmental Protection Agency. All of the 11 analyses for radium-226 were less than the maximum permissible level established for radium-226 and radium-228, a product of the decay of thorium, combined. Though there are no analyses of water from the Inyan Kara aquifer in the northern Black Hills, water from this aquifer is not known to have a measurable concentration of radium-228 in the eastern and southern Black Hills. The concentration of radium-226 does, however, exceed 5 picocuries per liter in parts of those areas. It is possible the excessive gross alpha radiation in the study area is the result of uranium decay and not radium-226 decay. Additional sampling, particularly in the Bear Lodge Mountains, would increase the understanding of radioactivity in ground water in the Black Hills and Bear Lodge Mountains.

The concentration of dissolved solids is undesirably large, more than 500 mg/L (milligrams per liter), in some areas. However, in areas not near the outcrop, most of the dissolved ions in water in the Inyan Kara are sodium and bicarbonate. In and near the outcrop, water with dissolved-solids concentration exceeding 500 mg/L generally has sulfate concentration exceeding 250 mg/L.

The concentrations of iron and manganese exceed the maximum recommended levels in parts of the study area and in other parts of the Black Hills. Metal well casing may increase the concentration of iron in water from wells and it is difficult to determine whether the origin of the iron in ground water is in the aquifer or in the well; however, there is an undesirable iron concentration in water from the Inyan Kara aquifer in other areas of western South Dakota, and the probable cause is the oxidation and solution of iron minerals in the rock.

The concentration of dissolved sulfate exceeds the maximum recommended level, generally in or near the outcrop. The source of dissolved sulfate in water in the Inyan Kara aquifer may be leakage of water from underlying gypsiferous units, the Gypsum Spring and Spearfish Formations, or oxidation of sulfide minerals in the Inyan Kara aquifer. The areas of undesirable sulfate concentration are not sufficiently delineated by the available data to quantify the distance from the outcrop where the water probably is acceptable.

Water from the Inyan Kara aquifer is hard (hardness 120 to 180 mg/L as calcium carbonate) to very hard (hardness greater than 180 mg/L as calcium carbonate) near the outcrop and soft east and northeast of the outcrop in the Williston Basin. In areas where the water is soft, the concentration of sodium and bicarbonate may affect the use of water for irrigation, depending on the soil type.

The principal deterrents to the use of water from the Inyan Kara aquifer are excessive gross alpha radiation and concentrations of iron, manganese, sulfate, and hardness. In some areas, water to be used for public water systems may require treatment or dilution.

Minnelusa Aquifer

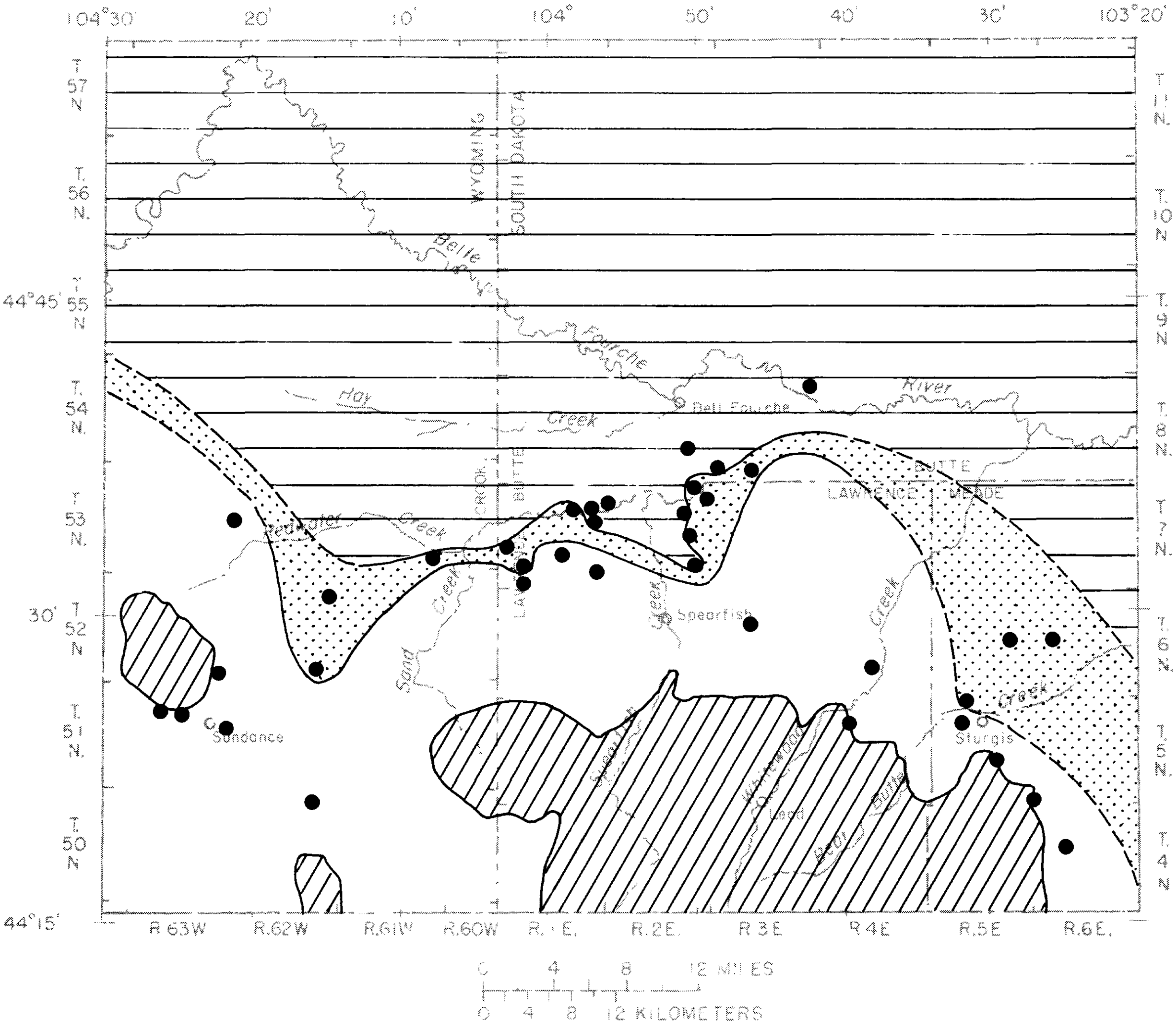
The maximum permissible or recommended levels allowed for public water systems for fluoride, dissolved solids, iron, and sulfate are exceeded in water from the Minnelusa aquifer in some parts of the study area. There are fewer than six analyses for most of the primary constituents; however, most of these constituents are not known to have excessive concentrations in water from the Minnelusa aquifer anywhere in South Dakota.

Of the 106 fluoride analyses, two analyses of water from well 8N3E2BDBC showed concentrations of fluoride exceeding the maximum permissible level. Generally the concentration of fluoride is less than 0.5 mg/L.

The concentration of dissolved solids in water from the Minnelusa aquifer in the northern one-half of the study area exceeds 500 mg/L. The sulfate concentration also is undesirably large in this area.

Iron concentration in water from the Minnelusa aquifer in the Black Hills is generally less than the maximum recommended level. The one analysis, of the eight in the study area, which had an undesirable concentration of iron was taken from a well with iron casing, which was suspected to have been corroding.

Sulfate concentration in water from the Minnelusa aquifer in approximately the northern one-half of the study area exceeds the maximum recommended level of 250 mg/L (U.S. Environmental Protection Agency, 1979) (fig. 14). The Minnelusa Formation has beds of anhydrite, a relatively soluble calcium-sulfate mineral, which is



EXPLANATION

MINNELUSA AQUIFER NOT PRESENT

AREA WHERE CONCENTRATION OF DISSOLVED SULFATE EXCEEDS 1,000 MILLIGRAMS PER LITER IN WATER FROM THE MINNELUSA AQUIFER--Boundary is dashed where approximately located

TRANSITIONAL AREA WHERE CONCENTRATION OF DISSOLVED SULFATE IS LESS THAN 1,000 MILLIGRAMS PER LITER AND GREATER THAN 250 MILLIGRAMS PER LITER--Bonundry is dashed where approximately located

AREA WHERE CONCENTRATION OF DISSOLVED SULFATE IS LESS THAN 250 MILLIGRAMS PER LITER

WELL FOR WHICH THERE IS A SULFATE ANALYSIS

Figure 14.--Area where the concentration of sulfate in water from the Minnelusa aquifer may affect its use. (Area where the aquifer is not present from Darton, 1905; Darton and Paige, 1925; and Staatz, 1983, pl. 1.)

the anhydrous form of gypsum. The anhydrite has been dissolved and removed near the Minnelusa Formation outcrop. As a result, water from the Minnelusa aquifer near the outcrop has low sulfate concentration (less than 250 mg/L). Across the middle of the study area, there is a transitional zone, approximately 2 to 10 mi wide, within which the concentration of dissolved sulfate in water from the Minnelusa aquifer ranges from 250 to 1,000 mg/L (fig. 14). The variability of concentration within this zone probably is caused by differences in the direction and rate of water movement, the presence of anhydrite beds, and whether the water is saturated, undersaturated, or at equilibrium with anhydrite. This transitional zone is probably moving slowly downgradient to the north and east as the anhydrite is dissolved. If the direction of water movement were reversed because of drawdowns from increased water development in the low sulfate area, the saline water would probably move towards the area of development.

Water from the Minnelusa aquifer is generally hard (hardness 120 to 180 mg/L as calcium carbonate) to very hard (hardness greater than 180 mg/L as calcium carbonate). Hardness in the Minnelusa aquifer mainly is caused by the solution of anhydrite, which adds calcium, as well as sulfate, to the water in the aquifer. The solution of limestone and dolomite beds and calcite cement in the sandstone beds also adds calcium to the water. The softest water in the study area is south and west of the transitional zone (fig. 11).

The principal deterrents to the use of water from the Minnelusa aquifer are sulfate and hardness. Water from the Minnelusa aquifer in the southern one-half of the study area, south of the transitional zone shown in figure 14, generally is suitable for public water systems and irrigation use without treatment.

Madison Aquifer

The maximum permissible or recommended levels allowed for public water systems for fluoride, gross alpha radiation, dissolved solids, iron, manganese, and sulfate are exceeded in water from the Madison aquifer in some parts of the study area. Only a few analyses of water from the Madison are available--eight or less analyses for most of the primary constituents. Concentrations of most dissolved constituents in water from the Madison aquifer in other areas of South Dakota are not known to exceed the maximum permissible or recommended levels allowed for public water systems.

Fluoride concentration is less than 0.5 mg/L in the aquifer in most of the study area. It exceeds the maximum permissible level in the northern part of the study area, within about 10 mi of the northern boundary.

The level of gross alpha radiation in water from the Madison aquifer exceeds the maximum permissible level allowed for public water systems in part of the study area. The highest concentration is in the north, based on analyses of water from two wells in T. 12 N., R. 3 E. The concentration of radium-226 in water from the Madison aquifer in other parts of South Dakota also exceeds the maximum permissible levels and it is likely that where the gross alpha radiation is excessive within the study area, the radium-226 concentration is also large. There are insufficient data to determine the distribution of radium-226 or gross alpha radiation in the study area, but in other parts of the Black Hills the water from the Madison aquifer near the outcrop has acceptable concentrations of radium-226 and gross alpha radiation.

In the northern part of the study area, furthest from the outcrop, the concentration of dissolved solids is undesirably large, more than 500 mg/L, in water from the

Madison aquifer. Sulfate, calcium, and bicarbonate, which are products of the dissolution of anhydrite, limestone, and dolomite, are the main dissolved ions.

The largest concentrations of iron and manganese are also in the northern part of the study area. The analyses with large concentrations of iron and manganese were for samples taken from wells with steel casings, therefore, the metal concentration may be caused by corrosion. In most of the study area, the concentrations of iron and manganese are probably acceptable, as is typical for the Madison aquifer in other parts of the Black Hills.

Sulfate concentration exceeds the maximum recommended level in water from the Madison aquifer in the northern part of the study area. The larger sulfate concentration may be caused by solution of anhydrite in the Madison Limestone, leakage of water from the Minnelusa aquifer to the Madison aquifer through the lower part of the Minnelusa Formation, or leakage of water from the Minnelusa aquifer into the well. Subordinate amounts of anhydrite have been identified in core samples of the Madison Limestone equivalents, and anhydrite nodules are reported to occur sporadically throughout the Madison Limestone (Thayer, 1981, p. 13). However, from examination of logs of the wells in the area, there appears to be less anhydrite in the Madison Limestone than the Minnelusa Formation. It is possible some of the wells reported to be open in the Madison aquifer are also open to the Minnelusa aquifer and water samples represent a mixture of water from both aquifers.

Water from the Madison aquifer is generally hard (hardness 120 to 180 mg/L as calcium carbonate) to very hard (hardness greater than 180 mg/L as calcium carbonate). Because the Madison Limestone consists mainly of limestone and dolomite, minerals containing calcium and magnesium, solution of the rock material increases hardness. The largest concentrations of hardness, however, are in the areas where sulfate concentration is large, which indicates calcium from anhydrite solution may be the main cause of hardness in the Madison aquifer in the northern study area.

The principal deterrents of the use of water from the Madison aquifer are fluoride, gross alpha radiation, sulfate, and hardness, in the northern part of the study area. In the southern one-half of the study area, water quality generally is suitable for public water systems and irrigation, though the water is hard to very hard.

COMPUTER SIMULATION

Description

A digital model was constructed to quickly determine whether the estimates of aquifer parameters, recharge, discharge, and water levels developed in this and previous studies presented, or could be easily modified to present, a consistent conceptual model of the flow system.

The digital model used a computer program developed by McDonald and Harbaugh (1984) which uses finite-difference numerical methods for simulation of three-dimensional ground-water flow in aquifers. It is called a "quasi-three-dimensional" model because it uses vertical conductance to simulate leakage between two aquifers or layers. Vertical conductance is defined as the vertical hydraulic conductivity divided by the thickness of the confining layer (McDonald and Harbaugh, 1984, p. 16-17). The rate of leakage between two aquifers is the difference in the altitudes of the potentiometric surfaces of the aquifers multiplied by the conductance of the confining layer separating them. Input to the digital model describes the aquifers as confined, unconfined, or a combination, as homogeneous or heterogeneous, as isotropic or anisotropic, and as having irregular or regular boundaries, based on the conceptual model. The boundary conditions are defined as part of the input data. Recharge and discharge rates, such as leakage from confining beds, stream gains and losses, precipitation, evapotranspiration, and spring and well discharges are also included in the input data.

The digital model used requires that the study area be divided into a grid network. Each rectangle in the network is called a cell and each cell has a node at the center. The digital model assigns a no-flow boundary around the perimeter of the model area (McDonald and Harbaugh, 1984, p. 41). Three different types of boundary conditions were used by the digital model (McDonald and Harbaugh, 1984, p. 41). These are as follows:

(1) No-flow boundary: No flow can occur across the boundary of the cell. Cells where the aquifer is missing and the perimeter of the model area were modeled as no-flow boundaries.

(2) Constant-head boundary: The hydraulic head in the cell is held constant throughout the simulation. Flow may occur to or from a constant-head cell. Cells in which a perennial stream crosses the outcrop of an aquifer were modeled as constant-head boundaries.

(3) Specified-flow boundary: A finite rate of withdrawal or recharge is assigned to the cell. Cells where subsurface flow into or out of the model is known to occur were modeled with specified-flow boundaries. These included most of the outermost cells, cells containing springs, and cells representing Spearfish Creek where it would have recharged the Minnelusa and Madison layers prior to development.

The digital model generates potentiometric levels from the data input from a conceptual model. The computed hydraulic heads are then compared to the observed hydraulic heads. If the data input represents the aquifer system accurately, the computed potentiometric levels will be comparable to the measured potentiometric levels, within the error of measurement. If the computed potentiometric levels do not fit the observed potentiometric levels, the conceptual model may be incorrect or inaccurate. By varying the input data within a reasonable range or altering the boundary conditions and computing new potentiometric levels, the model is essentially "rebuilt," until the computed and measured potentiometric levels are comparable. If the model cannot be modified in such a way that the computed and measured potentiometric levels are similar, then the concept of the system may be inadequate or incorrect, that is, some or all of the components of the conceptual model are incorrect; and the conceptual model is not internally consistent.

Application

Conceptual Model

A conceptual model of the aquifer system was developed from analyses of the characteristics of the three sedimentary aquifers discussed in earlier sections. In summary, the aquifer system consists of three aquifers, the Inyan Kara, Minnelusa, and Madison, which are separated by confining beds (fig. 15). The Inyan Kara and Minnelusa

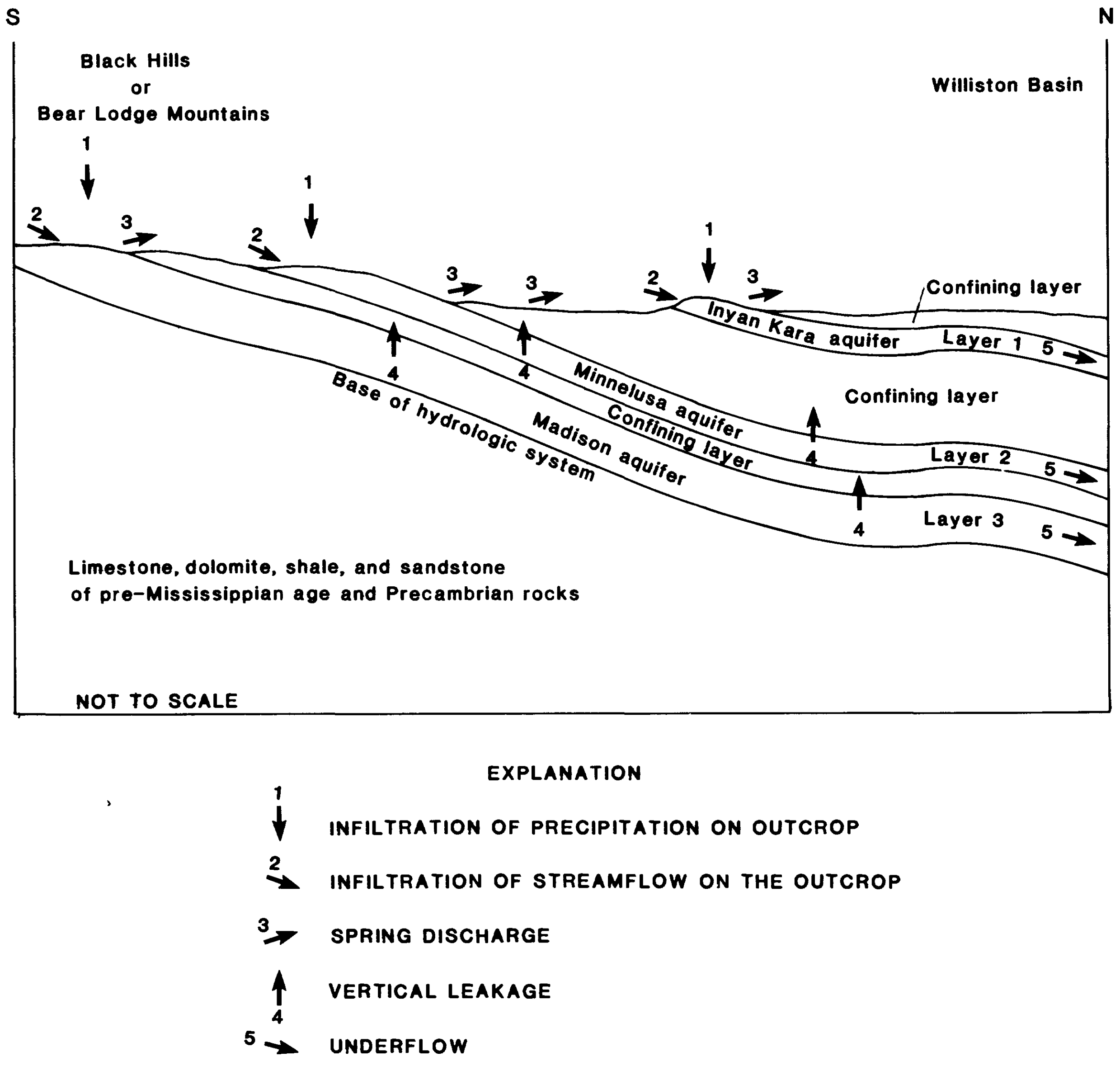


Figure 15.--Generalized geologic section showing hypothesized hydrologic system and forms of recharge and discharge.

aquifers are composed of sandstone and shale and the Madison aquifer is composed of fractured limestone and dolomite. Shale, siltstone, gypsum, and unfractured limestones and dolomites form the confining beds. The potentiometric-surface maps (figs. 7, 10, and 13) indicate that the water generally moves from the outcrop areas to the north and east and thence out of the study area. Recharge to the aquifers occurs as infiltration of precipitation and streamflow (fig. 15). Discharge occurs as springs (fig. 15). Water leaks between aquifers through confining layers, wells completed in more than one aquifer, and wells with deteriorating casing. The aquifers extend outside the study area and water flows into and out of the study area in the subsurface.

Model Development

Some simplifying assumptions are always necessary in constructing a digital ground-water model of the real hydrologic system with all its complexities. Other assumptions stem from the theory behind the governing equations of ground-water flow on which the digital model is based. The initial simplying assumptions used in this study to simulate the real system include:

(1) The aquifers are isotropic.

(2) Recharge is instantaneous and constant with time.

(3) Point discharge from springs occurs at a constant rate.

(4) Horizontal components of flow in the confining layers between aquifers and vertical components of flow in the aquifers are negligible.

(5) The potentiometric surfaces do not change significantly over long periods of time, that is, they are in a state of equilibrium, or steady state.

(6) Negligible leakage occurs through the confining layer overlying the Inyan Kara aquifer.

(7) Negligible leakage occurs through the confining layer underlying the Madison aquifer.

(8) Transmissivity varies over the areal extent of each aquifer within the ranges determined by previous investigations.

As required for the numerical method used by McDonald and Harbaugh (1983), the part of the study area to be simulated was divided into a three-dimensional grid network. The network had 3 layers, representing the 3 aquifers, and a grid consisting of 39 rows and 47 columns (fig. 16). There were 1,833 cells (and nodes) in each layer and 3 layers, for a total of 5,499 cells (and nodes) in the model. The dimensions of the cells in the horizontal plane, shown in figure 16, varied from 1,666 by 2,500 ft in the southwestern corner of the model area to 12,500 by 16,666 ft in the northeastern corner of the model area. The cells are dimensionless vertically.

Each node was assigned values representing the average values of the physical properties of the system in that node. The finite-difference grid was superimposed on the structure-contour map and each of the three potentiometric-surface maps. Each node was assigned a value for the altitude of the top of each aquifer using the structure contour and thickness maps. Each node was also assigned values for the aquifer

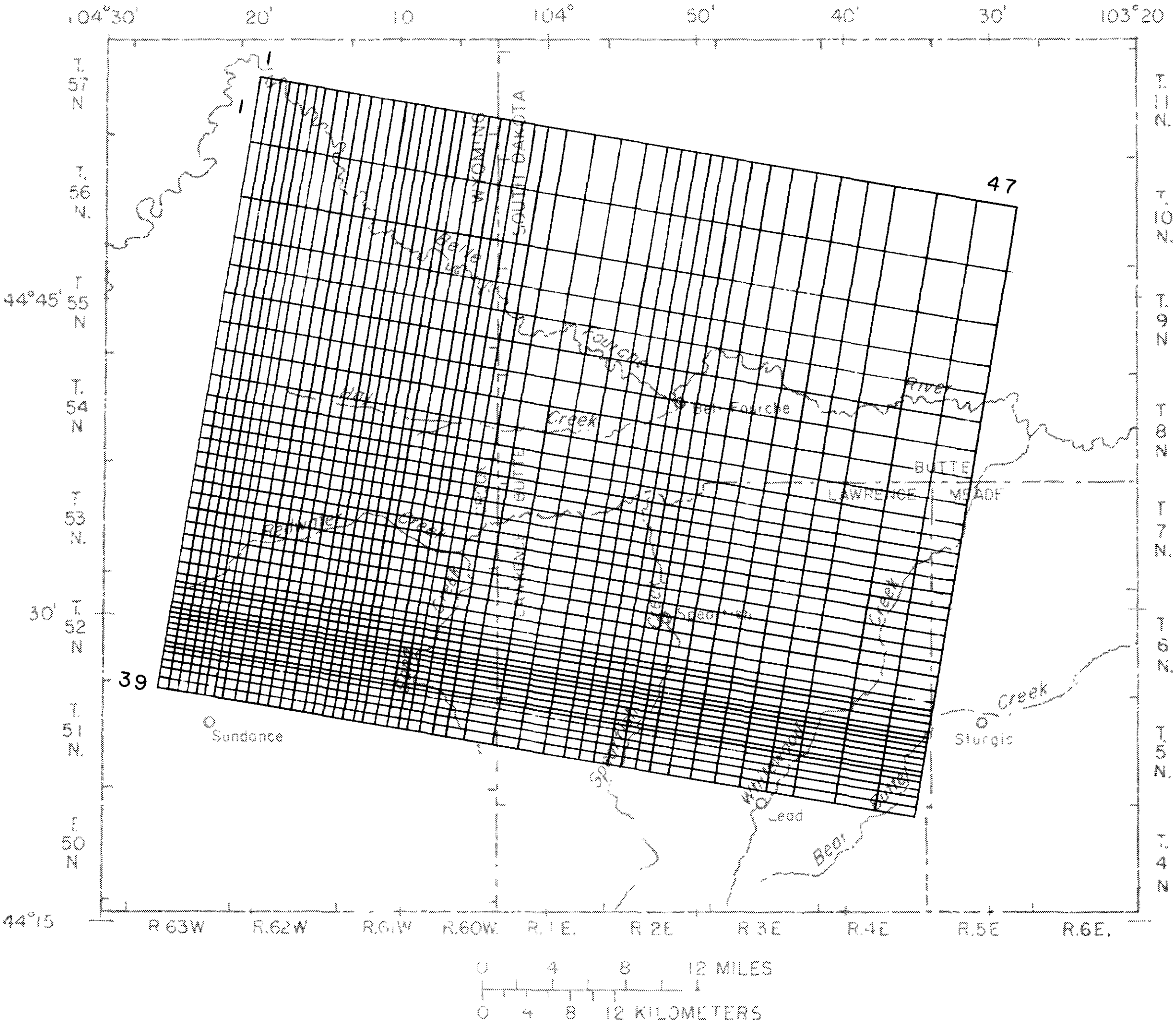


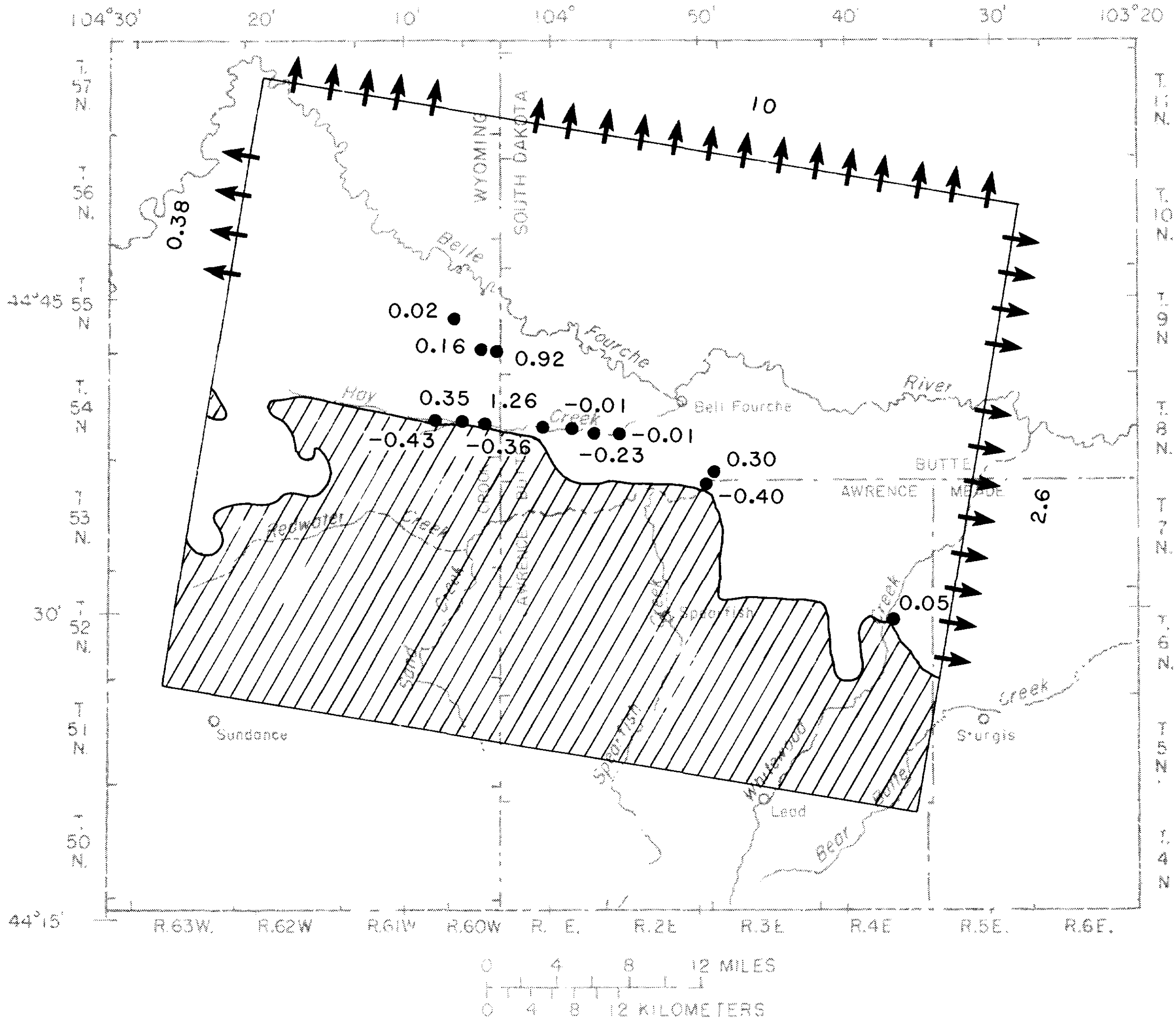
Figure 16.--Finite-difference grid used in model.

transmissivities and confining bed vertical hydraulic conductances, based on values from previous investigations (table 3). These assigned values of transmissivity and vertical hydraulic conductance were varied by trial and error in an effort to simulate the measured water levels.

To minimize boundary effects, the grid was oriented (fig. 16) to take advantage of as many identifiable boundaries as possible. No-flow boundaries were used for most of the southern boundary in each layer of the digital model because the aquifers are missing in the south (figs. 17, 18, and 19). Much of the western boundary can be simulated as a no-flow boundary because the ground-water flow is approximately parallel to the finite-difference grid border in each layer (figs. 7, 10, and 13) and therefore there is no subsurface flow into or out of the model area. Ground water enters or exists in the subsurface along the northern boundary for each layer, therefore, specified-flow boundaries have been used in the digital model along the northern edge. Similarly, the eastern boundary is a specified-flow boundary in each layer where ground water is flowing out of the model area in the subsurface. The specified flow was calculated by multiplying the estimated gradient, measured from the contoured potentiometric maps (figs. 7, 10, and 13), at each cell by the cell width and the transmissivity assigned to that cell. When transmissivity was changed during model development, the specified flow was recalculated. The specified-flow boundary was changed to a constant-head boundary in later runs to test the sensitivity of the model to this boundary. Nodes representing reaches of perennial streams flowing across the outcrop of each aquifer were represented by constant-head nodes (figs. 17, 18, and 19). The model calculated the rate of flow into and out of the constant-head nodes and these rates were compared to streamflows, if known.

Recharge from precipitation was estimated from evapotranspiration estimates from previous investigations (Rahn and Gries, 1973) and from precipitation records (U.S. Department of Commerce, 1980-83 and 1981-83). Evapotranspiration is estimated to be about 15 in/yr in the study area, based on hydrologic budget analyses (Rahn and Gries, 1973, p. 17). The average annual precipitation is about 16 in/yr on the outcrop of the Inyan Kara aquifer and it is estimated that about 1 in/yr recharges the aquifer, distributed over the outcrop. Precipitation on the higher parts of the study area averages about 22 in/yr on the outcrop area of the Minnelusa and Madison aquifers. The estimated recharge rate to both the Minnelusa and Madison aquifers is approximately 7 in/yr, distributed over the outcrop. An error in the estimate for the rate of recharge will change the component of the volumetric water budget from precipitation proportionately; that is, a recharge rate of 2 in/yr would double the volume of water from precipitation recharging the Inyan Kara aquifer. The estimated volumetric rate of recharge from precipitation to the Inyan Kara aquifer is 11 ft^3/s, to the Minnelusa aquifer is 41 ft^3/s, and to the Madison aquifer is 16 ft^3/s. Distributed recharge was replaced by constant heads in the outcrop in later runs to test the significance of these estimated recharge rates.

The recharge rates used in the model were only approximations. More accurate estimates could be made using a network of continuous-record precipitation gages and streamflow gages to calculate rainfall-runoff relationships. Investigations on relatively impermeable terrane may provide estimates of evapotranspiration.



EXPLANATION

NO-FLOW BOUNDARY--Aquifer not present

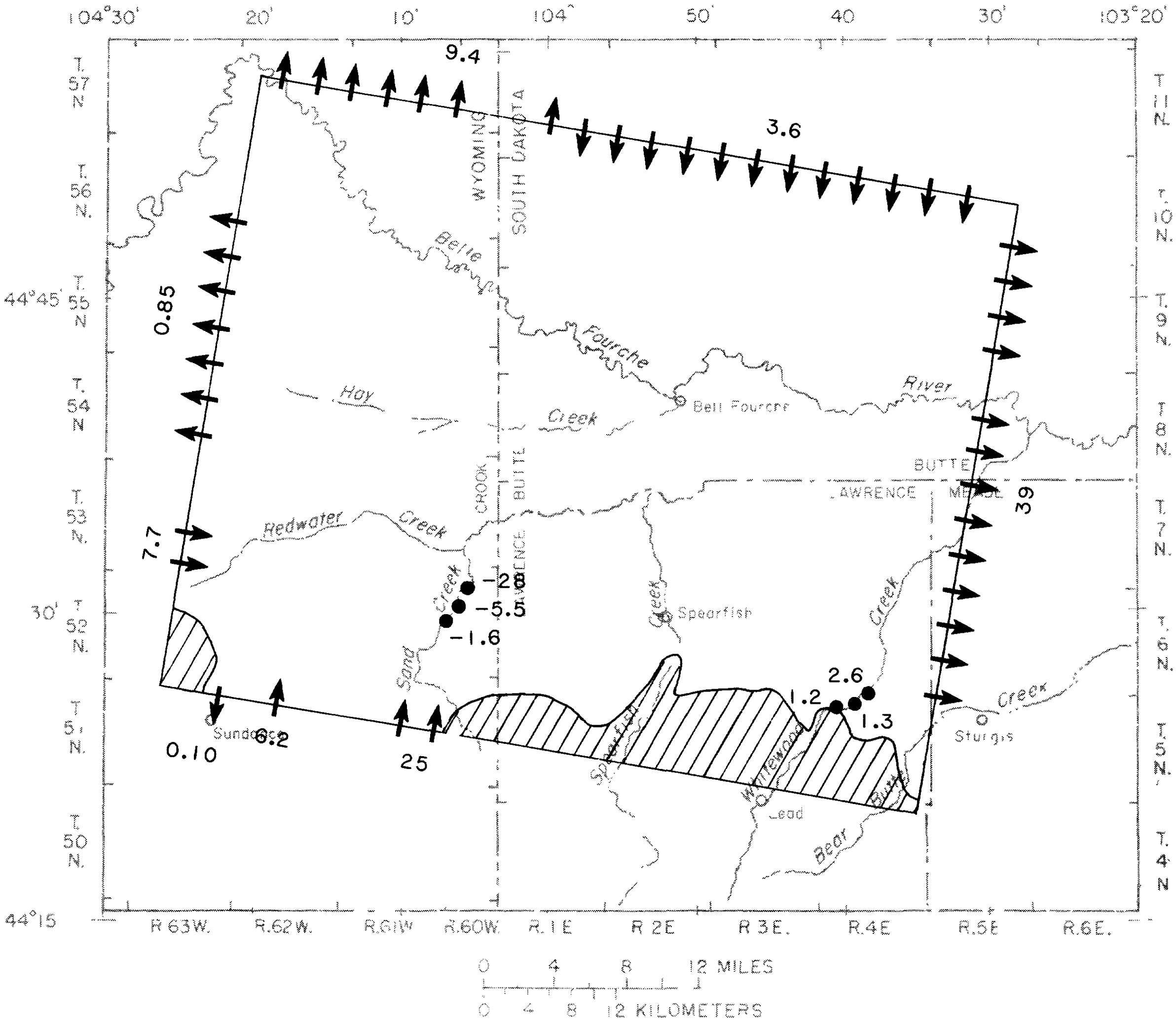
10

SPECIFIED FLOW BOUNDARY--Number indicates rate in cubic feet per second. Arrows show direction of water movement, into or out of the model area

-0.43

CONSTANT-HEAD NODE--Number indicates rate in cubic feet per second of simulated recharge or discharge. Positive number indicates recharge from streamflow infiltration and negative number indicates discharge as springs

Figure 17.--Recharge and discharge rates for specified-flow and constant-head nodes for the Inyan Kara aquifer (layer 1).



EXPLANATION

NO-FLOW BOUNDARY--Aquifer not present

9.4

SPECIFIED FLOW BOUNDARY--Number indicates rate in cubic feet per second. Arrows show direction of water movement, into or out of the model area

-1.6

CONSTANT-HEAD NODE--Number indicates rate in cubic feet per second of simulated recharge or discharge. Positive number indicates recharge from streamflow infiltration and negative number indicates discharge as springs

Figure 18.--Recharge and discharge rates for specified-flow and constant-head nodes for the Minnelusa aquifer (layer 2).

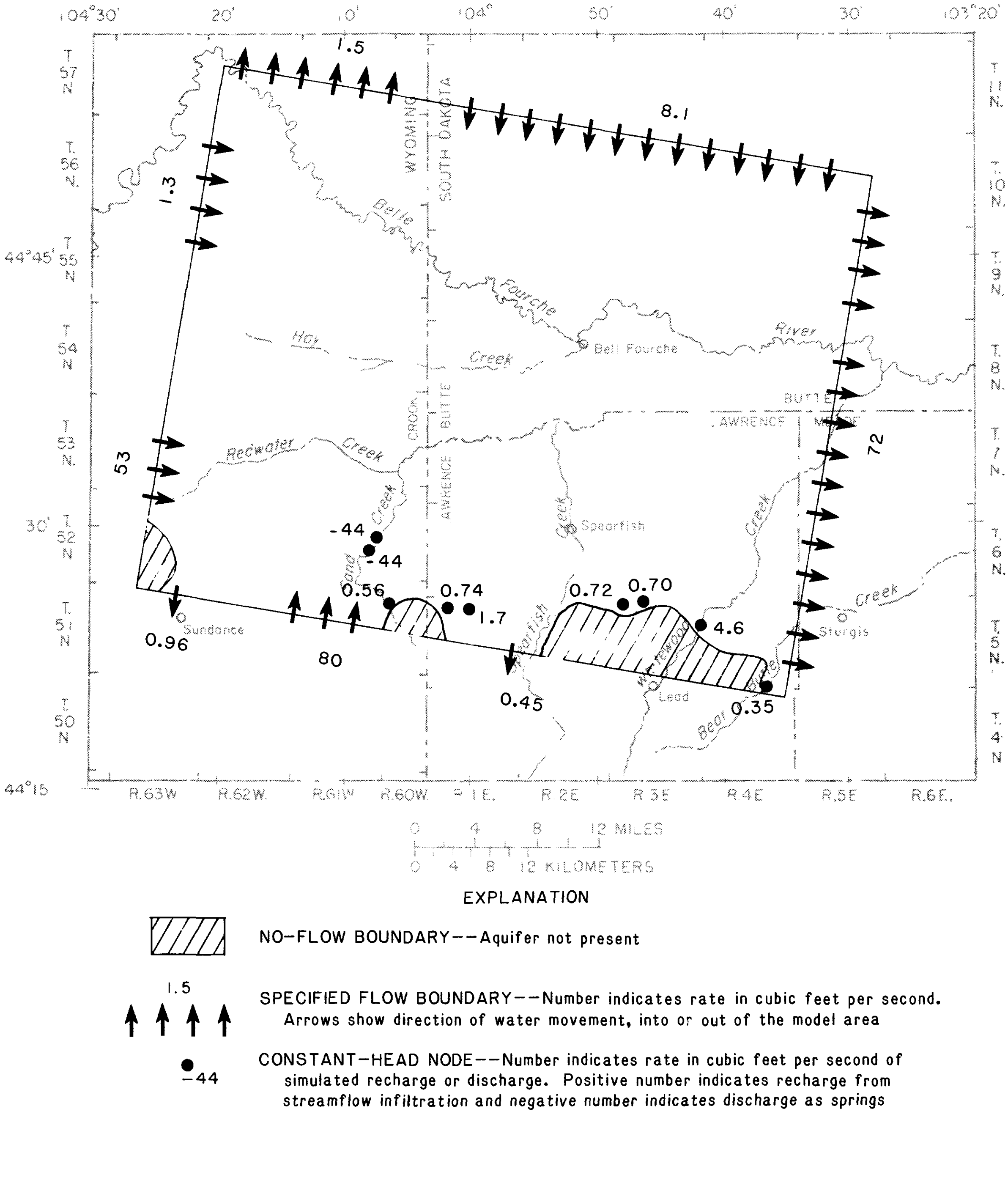


Figure 19.--Recharge and discharge rates for specified-flow and constant-head nodes for the Madison aquifer (layer 3).

Model Results

Attempts to calibrate the digital model were made by varying transmissivity and vertical hydraulic conductivity. The transmissivities and vertical conductivities were varied, based on the values determined by previous investigations (table 3), and the boundary fluxes adjusted accordingly, until the computed water levels could not be further improved. Computed potentiometric-surface maps were contoured from the final values generated by the digital model, and compared visually with the maps of the measured potentiometric surfaces. Measured water levels in wells were compared with the water levels computed for the cells containing the wells. The differences between the measured and computed water levels were summed and the algebraic mean difference and the absolute mean difference were calculated. The flow into and out of the constant head nodes was compared to measured streamflow, spring discharge, and estimates of recharge from precipitation.

A hydrologic budget for the model area was estimated. Shown in table 5, values for precipitation, spring discharge, recharge from Spearfish Creek, and boundary fluxes (inflow and outflow) were estimated and input to the model. The model calculated leakage between the layers and flow into or out of rivers and streams (constant-head nodes). Because inflow and outflow were calculated using gradients from approximated potentiometric maps (figs. 7, 10, and 13) and estimates of transmissivity, there may be considerable error in these boundary flux calculations.

Inyan Kara Aquifer

Comparison of the computed potentiometric-surface map of the Inyan Kara aquifer (fig. 20) with the conceptualized potentiometric-surface map drawn from observed water levels (fig. 7) is an indication of the accuracy of the model. The difference between the computed and the conceptualized potentiometric surfaces ranged from 0 to about 130 ft. The difference was less than 30 ft in most of the model area. Generally, in areas where the conceptualized potentiometric surface is not well known (few or no water-level measurements), the differences were greater than 100 ft. The absolute value of the difference between computed and measured water levels ranged from 0 to 272 ft. The algebraic mean difference was 21 ft and the absolute mean difference was 65 ft (table 6). The general direction of ground-water flow is about the same in the observed and computed system.

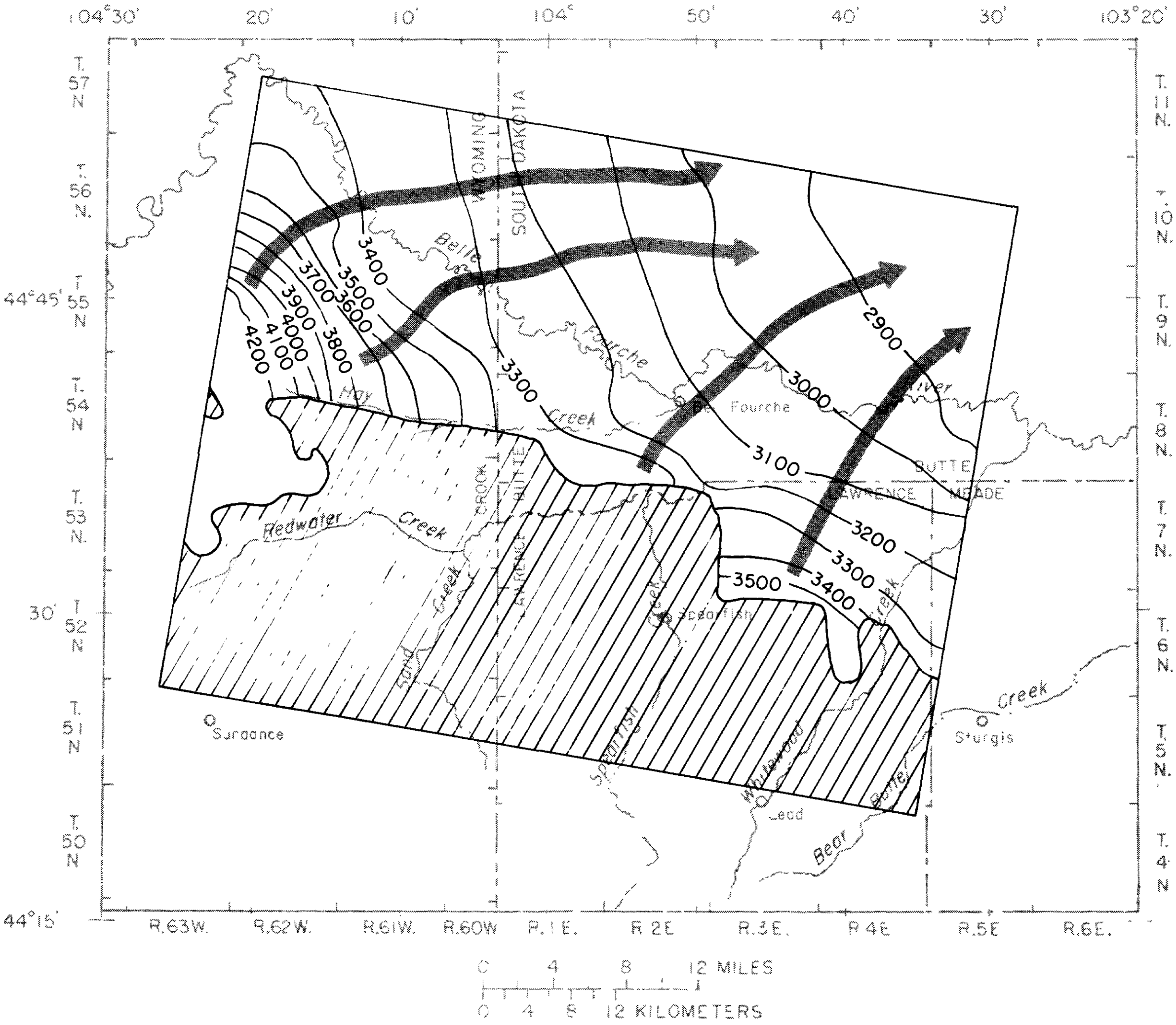
The rates of recharge to and discharge from the Inyan Kara aquifer calculated by the model for constant-head nodes and specific-flow boundaries, shown in figure 17 and table 5, appear reasonable, based on records of streamflow and the potentiometric maps. The flow into and out of constant-head nodes representing the perennial reaches of Hay Creek and Spearfish Creek (fig. 17) are less than the average flows on those creeks (table 2), and the flow into the node on Whitewood Creek is less than the minimum flow of Whitewood Creek (table 2).

The areal distribution of transmissivity values for the Inyan Kara aquifer which best simulated the aquifer system ranged from 0.86 to 6,000 ft^2/d. The range of transmissivity values used in this model exceeded that of previous investigators. The highest transmissivity occurs in the northwest corner of the study area. The transmissivity values are lowest in the southeast corner of the study area. The area of higher transmissivity in the northwest may be caused by a facies difference within the aquifer, such as thicker, more numerous, or coarse-grained sandstones, or by a more permeable area, perhaps caused by removal of cement within the sandstone by ground-

Table 5.--Hydrologic budget of the model area

	Computed values (cubic feet per second)
Inyan Kara aquifer	
Recharge	
Precipitation 1/	12
Rivers and streams 2/	3.2
Leakage 3/ (from Minnelusa aquifer)	.01
Inflow	.0
Discharge	
Springs 1/	.08
Rivers and streams 2/	1.6
Leakage (to Minnelusa aquifer)	.0008
Outflow	13
Minnelusa aquifer	
Recharge	
Precipitation 1/	41
Rivers and streams 2/	8.8
Leakage (from Madison aquifer)	13
Leakage (from Inyan Kara aquifer)	.0008
Inflow	43
Discharge	
Springs 1/	18
Rivers and streams 2/	36
Leakage (to Madison aquifer)	1.8
Leakage (to Inyan Kara aquifer)	.01
Outflow	49
Madison aquifer	
Recharge	
Precipitation 1/	16
Rivers and streams 2/	14
Leakage 4/ (from below)	.0
Leakage (from Minnelusa aquifer)	1.8
Inflow	140
Discharge	
Springs 1/	.04
Rivers and streams 2/	87
Leakage 4/ (to below)	.0
Leakage (to Minnelusa aquifer)	13
Outflow	75

1/ Estimated values input to the model.
2/ Rivers simulated in model as constant-head nodes except for Spearfish Creek, which is simulated as recharging wells to the Minnelusa and Madison aquifers.
3/ Leakage to or from overlying confining bed assumed to be negligible.
4/ Leakage to or from underlying confining bed assumed to be negligible.



EXPLANATION

INYAN KARA AQUIFER NOT PRESENT

DIRECTION OF GROUND-WATER FLOW--As calculated by the model

—— 3600 —— POTENTIOMETRIC CONTOUR--As calculated by the model. Contour interval 100 feet. Datum is sea level

Figure 20.--Computed potentiometric surface and selected flowpaths of the Inyan Kara aquifer.

Table 6.--Comparison of measured and computed water levels for a range of transmissivities of the Inyan Kara aquifer and vertical conductances between the Inyan Kara and Minnelusa aquifers

	Steady-state model run, specified flow on outer boundary, specified recharge on the outcrop	Transmissivity of Inyan Kara aquifer		Vertical conductance between the Inyan Kara and Minelusa aquifers increased by an order of magnitude
		reduced by 25 percent	increased by 25 percent	
Inyan Kara aquifer				
Algabraic mean difference 1/ (feet)	21	48	5	23
Absolute mean difference 2/ (feet)	65	109	79	65
Maximum difference 3/ (feet)	-272	657	490	-270
Number of wells	141	141	141	141
Net flow into constant-head nodes (cubic feet per second)	3.1	3.0	3.4	3.1
Net flow out of constant-head nodes (cubic feet per second)	1.6	1.5	1.9	1.6
Minnelusa aquifer				
Algabraic mean difference 1/ (feet)	-10	16	16	16
Absolute mean difference 2/ (feet)	157	157	157	157
Maximum difference 3/ (feet)	-930	-894	-894	-894
Number of wells	116	116	116	116
Net flow into constant-head nodes (cubic feet per second)	3.8	3.3	3.3	3.3
Net flow out of constant-head nodes (cubic feet per second)	35.7	39.5	39.5	39.5
Madison aquifer				
Algabraic mean difference 1/ (feet)	-64	-98	98	-98
Absolute mean difference 2/ (feet)	319	314	314	314
Maximum difference 3/ (feet)	-1,136	-1,152	-1,152	-1,152
Number of wells	37	37	37	37
Net flow into constant-head nodes (cubic feet per second)	9.9	9.9	9.9	9.9
Net flow out of constant-head nodes (cubic feet per second)	87.2	83.3	83.3	83.3

1/ Derived by adding when computed value was higher than measured value (positive) and subtracting when computed value was lower than measured value (negative number).
2/ The absolute value of a number is the number without its associated sign. For example, the absolute value of 4 and -4 are the same.
3/ The value is negative when the maximum computed value is less than the corresponding measured value.

water solution. Another possibility is that sandstone beds in the Sundance Formation (table 1) are a significant aquifer in this area and are hydraulically connected to the Inyan Kara aquifer and therefore there would be more recharge from precipitation and a greater transmissivity because of greater thickness. There is insufficient information on the hydrologic properties of the Sundance Formation and its relationship to the Inyan Kara was not further evaluated as part of this study.

The model-calculated rate of leakage to the Minnelusa aquifer of 8.0×10^{-4} ft^3/s and from the Minnelusa aquifer of 0.01 ft^3/s indicate that flow between the Minnelusa and Inyan Kara aquifers is insignificant. The model is relatively insensitive to changes in vertical hydraulic conductance between these two layers, therefore, the calculated values may be in considerable error.

The sensitivity of the model to changes in transmissivity of the Inyan Kara aquifer was tested by changing the transmissivity by 25 percent (table 6). An increase of 25 percent increased the absolute mean difference from 65 to 109 ft. A decrease in the transmissivity of 25 percent increased the absolute mean difference from 65 to 79 ft, though the algebraic mean difference improved, decreasing from 21 to 5 ft.

Increasing the vertical hydraulic conductance between the Minnelusa and the Inyan Kara aquifers by an order of magnitude had little effect on the computed water levels. The differences between the measured and computed water levels changed by 3 ft or less (table 6).

The assumed boundary conditions were tested by successively changing the outer boundary of each layer to constant-head conditions, then also changing the outcrop nodes of each layer to constant-head conditions. The mean difference between the measured and computed water levels of the Inyan Kara aquifer changed 10 ft or less (table 7) for these different boundary conditions. The total flow out of the constant head nodes increased from 1.6 to 14.5 ft^3/s when the outer boundary was changed to constant-head conditions. The difference is comparable to specified-flow rate of 13 ft^3/s. Changing the outcrop to constant-head conditions increased the total flow into the constant-head nodes from 3.3 to 14 ft^3/s. The specified recharge rate was 12 ft^3/s, comparable to the increase of 11 ft^3/s. Changes in the Inyan Kara aquifer had little or no effect on the other layers (table 6).

Table 7.--Comparison of measured and computed water levels for a range of boundary conditions

	Steady-state model run, specified flow on outer boundary, specified recharge on the outcrop	Constant-head nodes in outcrop of the Inyan Kara Minnelusa, and Madison aquifers	Constant-head nodes in outer boundaries of:			Constant-head nodes in all outer boundaries. Constant-head nodes in outcrop of:		
			Madison aquifer	Minnelusa and Madison aquifers	Inyan Kara, Minnelusa, and Madison aquifers	Madison aquifer	Minnelusa and Madison aquifers	Inyan Kara, Minnelusa, and Madison aquifers
Inyan Kara aquifer								
Algebraic mean difference [1] (feet)	21	31	21	21	22	22	22	31
Absolute mean difference [2] (feet)	65	58	65	65	64	64	64	59
Maximum difference [3] (feet)	-272	177	-272	-272	-262	-262	-262	194
Number of wells	141	87	141	141	127	127	127	77
Net flow into constant-head nodes (cubic feet per second)	3.1	14.1	3.2	3.2	3.3	3.3	3.3	14.2
Net flow out of constant-head nodes (cubic feet per second)	1.6	1.0	1.6	1.6	14.5	14.5	14.6	14.2
Minnelusa aquifer								
Algebraic mean difference [1] (feet)	-10	150	-17	-105	-105	-97	106	106
Absolute mean difference [2] (feet)	157	189	156	168	168	166	156	156
Maximum difference [3] (feet)	-930	441	-957	-1,158	-1,158	-1,148	-415	-415
Number of wells	116	95	116	114	114	114	95	95
Net flow into constant-head nodes (cubic feet per second)	3.8	235	3.4	36.6	36.6	34.6	282	282
Net flow out of constant-head nodes (cubic feet per second)	35.7	198	31.4	73.6	73.6	76.0	271	271
Madison aquifer								
Algabraic mean difference [1] (feet)	-64	169	-65	-74	-74	130	140	140
Absolute mean difference [2] (feet)	319	250	351	349	349	190	196	197
Maximum difference [3] (feet)	-1,136	-624	-1,245	-1,253	-1,253	-660	-624	-624
Number of wells	37	26	32	32	32	22	22	22
Net flow into constant-head nodes (cubic feet per second)	9.9	148	72.2	73.4	73.4	231	226	226
Net flow out of constant-head nodes (cubic feet per second)	87.2	102	85.8	83.8	83.8	216	225	225

1/ Derived by adding when computed value was higher than measured value (positive) and subtracting when computed value was lower than measured value (negative number).
2/ The absolute value of a number is the number without its associated sign. For example, the absolute value of 4 and -4 are the same.
3/ The value is negative when the maximum computed value is less than the corresponding measured value.

Minnelusa Aquifer

The computed potentiometric surface of the Minnelusa aquifer (fig. 21) does not compare well with the potentiometric-surface map drawn from observed water levels (fig. 10). The computed potentiometric surface is 100 to 200 ft lower than the conceptualized potentiometric surface in most of the model area. In some areas differences exceed 500 ft. The absolute value of the differences between the measured and computed water levels range from 2 to 930 ft. The algebraic mean difference was -10 ft and the absolute mean difference was 157 ft (table 7). The hydraulic gradient of the computed surface is generally less than the gradient of the measured surface.

The computed water levels near Spearfish are as much as 500 ft lower than the measured water levels. The hydrologic properties of the lower part of the Minnelusa Formation are not well understood. Alternative models in which the vertical conductance between the Madison and the Minnelusa aquifers was increased did not converge. A smaller scale model of the Spearfish area using multiple layers for the Minnelusa Formation could test the hypothesis that the Madison aquifer is leaking significantly greater amounts of water to the Minnelusa aquifer than indicated by this investigation. However, the lack of water-level measurements for the Madison aquifer and information on the lower part of the Minnelusa Formation in the Spearfish area will limit the verification of the results.

The rates of recharge to and discharge from the Minnelusa aquifer calculated by the model for constant-head nodes representing Sand Creek and Whitewood Creek, shown in figure 18, are not all reasonable based on streamflow records. The average flow of Sand Creek is 24.1 ft^3/s (table 2). The calculated rate of discharge from the Minnelusa aquifer to the constant-head node representing Sand Creek was 35 ft^3/s (fig. 18), about 10 ft^3/s more than the average. The minimum measured discharge on Whitewood Creek above Whitewood (table 2) exceeds the calculated rate of discharge of the Minnelusa aquifer to the constant-head node representing Whitewood Creek (fig. 18).

The areal distribution of transmissivity values for the Minnelusa aquifer which best simulated the aquifer system ranged from 0.86 to 8,600 ft^2/d (table 3), however, the error of these estimates is not known. The transmissivities are generally lower in and near the outcrop areas. Values for transmissivity and for vertical hydraulic conductivity of the confining layer which separates the Minnelusa and Madison aquifers were assumed to be larger along the geologic structures because it was assumed the permeability would be greater in areas where the rocks are more likely to be fractured (Lattman and Parizek, 1964, p. 78; Siddiqui and Parizek, 1971, p. 1,303). Increasing the transmissivity and vertical hydraulic conductivity along the structures improved the results.

Changes in the boundary conditions affected the computed water levels, generally increasing the algebraic mean difference between the measured and computed levels (table 7). Making the outcrop and the outer boundary in the Minnelusa and Madison aquifers constant-head nodes increased the algebraic mean difference of the measured and computed water levels in the Minnelusa aquifer from -10 to 106 ft, though the absolute mean difference became slightly smaller, 157 to 156 ft. However, the flow into constant-head nodes increased to 282 ft^3/s; most of this increase was in nodes on the outcrop. With an outcrop area of 80 mi^2 and a precipitation rate of about 22 in/yr, the total potential annual recharge available to the Minnelusa aquifer from precipitation is about 130 ft^3/s, if evapotranspiration and runoff were negligible, an extreme

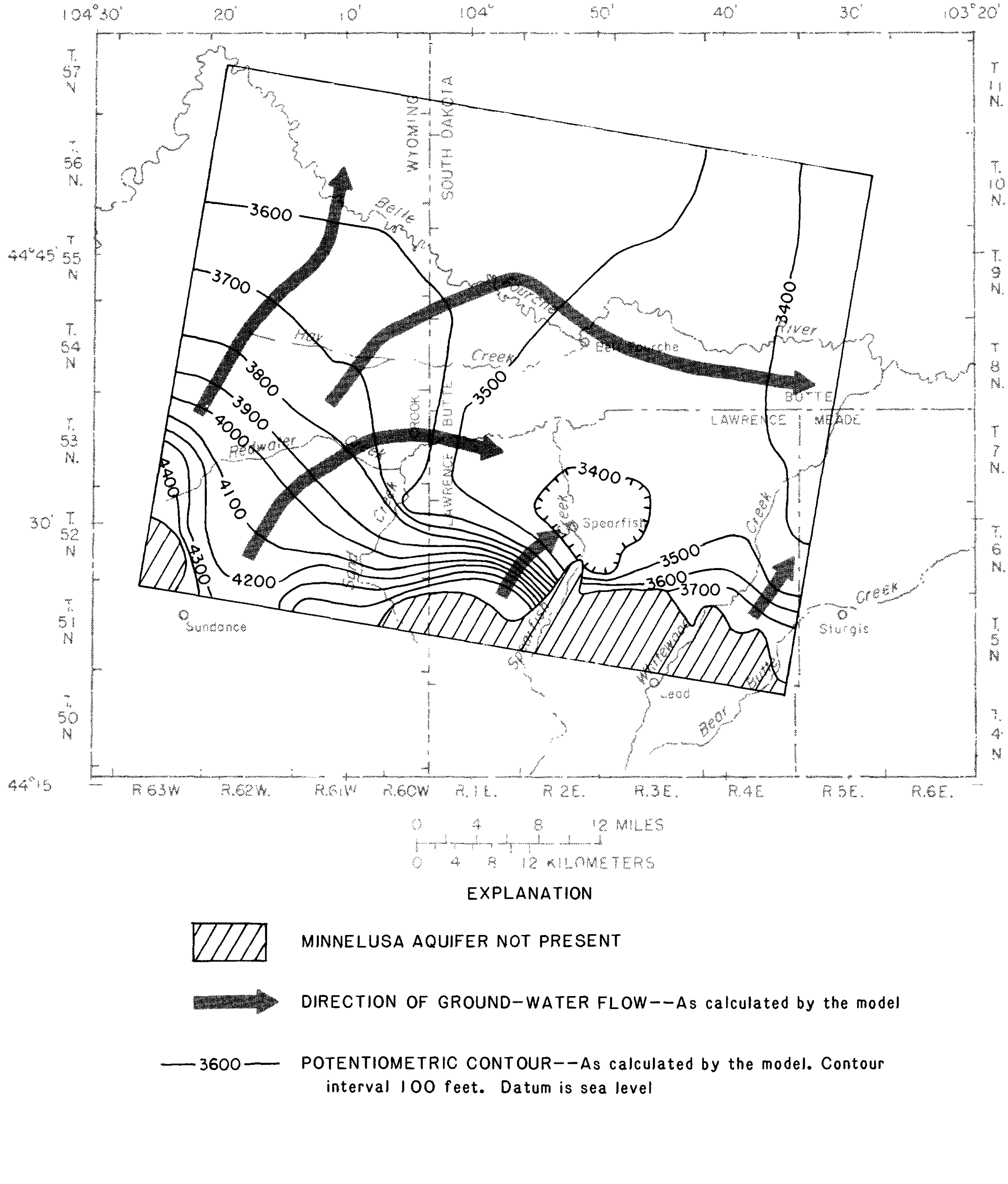


Figure 21.--Computed potentiometric surface and selected flowpaths of the Minnelusa aquifer.

and unrealistic situation. More likely recharge from precipitation is much less. The model does not adequately simulate all the recharge to the Minnelusa, which suggests more leakage from other units, such as the Madison than assumed. Furthermore, the error in the boundary fluxes (fig. 18 and table 5) is unknown. Without more reliable potentiometric data, the gradients used in calculating inflow and outflow are gross approximations.

Madison Aquifer

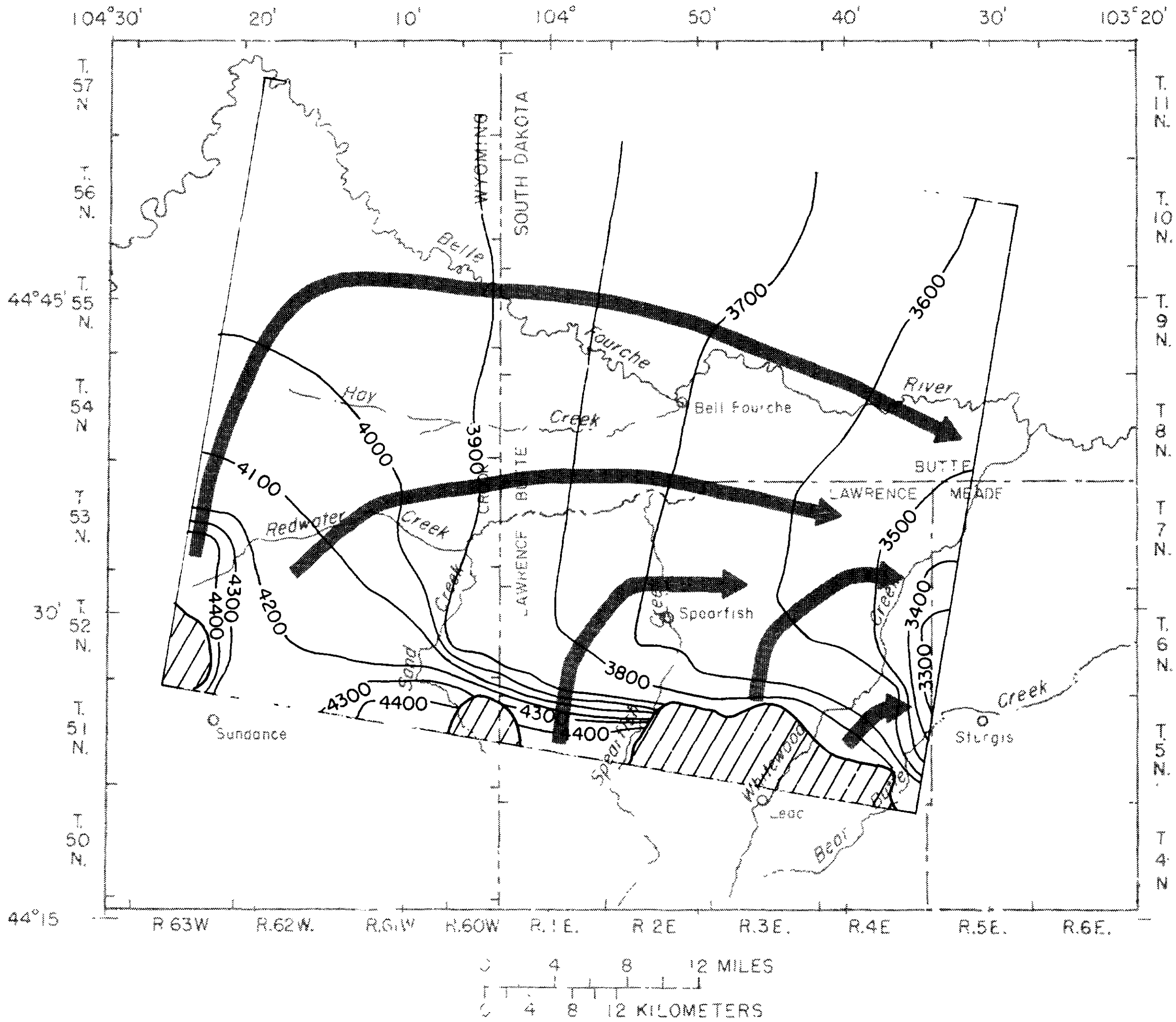
The computed potentiometric-surface map of the Madison aquifer (fig. 22) does not compare well with the potentiometric-surface map (fig. 13) drawn from measured water levels. The computed potentiometric surface is about 100 to 200 ft higher than the conceptualized potentimetric surface in most of the model area, except in and near the outcrop, where it is about 100 to 200 ft lower. The absolute value of the differences between measured and computed water levels range from 25 to 1,136 ft. The algebraic mean difference was -64 ft and the absolute mean difference was 319 ft. The general ground-water flowpaths are about the same in the conceptualized and computed systems. The largest differences are in the outcrop where the measured water levels are spring altitudes. These springs may not be representative of the regional flow system. There are no nearby water wells to confirm the Madison is saturated below the springs and no independent water-level measurements to show flow is essentially horizontal in this area. As a result, the estimated boundary flux near the outcrop, southwest of Sand Creek, may be in considerable error.

It was assumed that the Whitewood Dolomite and Deadwood Formation are not regionally extensive aquifers. They may actually be water-bearing and leakage between the Minnelusa aquifer, Madison aquifer, Whitewood Dolomite, and the Deadwood Formation may be more significant than assumed by this model.

The rates of recharge to and discharge from the Madison aquifer calculated by the model for constant-head nodes, shown in figure 19, do not appear reasonable for Sand Creek, but are within reason for Whitewood Creek. The computed discharge for the constant-head nodes representing Sand Creek (fig. 19) exceed the average measured discharge (table 2) by about 64 ft^3/s. The computed discharge to the constant-head nodes representing Whitewood Creek is 4.6 ft^3/s, about the measured minimum creek discharge (table 2).

The transmissivity values for the Madison aquifer which best simulated the digital aquifer system ranged from 4.3 to 8,600 ft^2/d over the area. The transmissivity was highest in the western portion of the study area where the aquifer is confined. Transmissivity is lower in and near the outcrop where the aquifer is believed to have a small saturated thickness, or may even be "dry," as evidenced by the dry streambeds in much of the outcrop areas.

Changing the conditions in the Madison aquifer to constant head in the outcrop and along the outer boundary generally improved the computed water levels, lessening the mean absolute difference from the measured water levels from 314 to 190 ft (table 7). However, the resultant recharge to the outcrop was far in excess of what was reasonable. With an outcrop area of 35 mi^2 in the study area and a precipitation rate of about 22 in/yr, the total potential annual recharge available to the Madison from precipitation is about 57 ft^3/s, if evapotranspiration and runoff were negligible, an extreme and unrealistic situation. Interaction with other aquifers is a possible source of additional recharge.



EXPLANATION

MADISON AQUIFER NOT PRESENT

DIRECTION OF GROUND-WATER FLOW--As calculated by the model

——3600—— POTENTIOMETRIC CONTOUR--As calculated by the model. Contour interval 100 feet. Datum is sea level

Figure 22.--Computed potentiometric surface and selected flowpaths of the Madison aquifer.

SUMMARY AND CONCLUSIONS

There are three major aquifers in the northern Black Hills and Bear Lodge Mountains and they are part or all of three geologic units: the Inyan Kara Group of Early Cretaceous age, the Minnelusa Formation of Pennsylvanian and Permian age, and the Madison Limestone (or Pahasapa Limestone) of Mississippian age. This study describes the geohydrology and water quality of these three aquifers in an area of about 3,000 square miles in South Dakota and Wyoming.

The area of the Inyan Kara Group outcrop is approximately 145 square miles in the northeastern Black Hills and northern Bear Lodge Mountains. It is buried at depths exceeding 3,000 feet below land surface in the northeastern part of the study area. The Inyan Kara Group comprises the Fall River and Lakota Formations. It is composed of sandstones, which form the aquifer, interbedded with shales and siltstones and is about 81 to 475 feet thick. The potentiometric surface of the Inyan Kara aquifer (fig. 7) shows that the general direction of flow is to the northeast and wells will flow in most areas other than in the outcrop. Near Belle Fourche, the potentiometric surface has declined as much as 125 feet by development since the early 1900's. Estimates of hydraulic conductivity from previous investigations range from 0 to 100 feet per day (table 3). Estimates of transmissivity from this and previous investigations range from 0.86 to 6,000 feet squared per day (table 3). Storage coefficients from previous investigations range from 1×10^{-5} for the entire aquifer to 1×10^{-4} for the Lakota Formation alone (table 3). Well yields range from 1 to 300 gallons per minute. The principal deterrents to the use of water from the Inyan Kara aquifer are locally excessive gross alpha radiation, concentrations of iron, manganese, sulfate, and hardness. In some areas water to be used for public water systems may require treatment or dilution.

Most of the outcrop of the Minnelusa Formation, which has a total area of about 80 square miles, is in the Black Hills. It is buried at depths exceeding 4,500 feet below land surface in the northeastern portion of the study area. Thickness of the Minnelusa Formation varies from about 340 to 800 feet. The upper part of the Minnelusa Formation, which forms an aquifer in this area, is composed of sandstone, limestone, dolomite, and shale with local deposits of anhydrite and gypsum and is about 200 feet thick in the study area. The lower part of the Minnelusa Formation is similar in lithology, but has more limestone and dolomite and is assumed, based on water levels, to be a confining or leaky confining layer. The confining layer on top of the Minnelusa is about 1,300 feet thick and composed of shales and siltstones, with some interbedded sandstone, limestone, and gypsum. The general direction of water movement in the Minnelusa aquifer is to the east and northeast. Wells flow in most of the study area in South Dakota and in approximately the northern one-half of Crook County, Wyoming, that is in the study area. Water levels in the vicinity of Spearfish, South Dakota, declined as much as 99 feet between 1907 and 1956, probably as a result of the diversion of flow in Spearfish Creek that resulted in reduced recharge to the aquifer. Since 1956 the water levels have fluctuated but have not shown a declining trend, as shown by a hydrograph in figure 10. Estimates of transmissivity of the Minnelusa aquifer from this and previous investigations range from 0.86 to 8,600 feet squared per day (table 3). Storage coefficient was estimated by Woodward-Clyde Consultants (1980, p. 4-12) to range from 6.6×10^{-5} to 2.0×10^{-4}. The Minnelusa aquifer yields in excess of 1,000 gallons per minute to wells in some areas. The principal deterrents to the use of the water from the Minnelusa aquifer are sulfate and hardness. Water from the Minnelusa aquifer in the southern one-half of the study area, south of the transitional zone shown in figure 14, generally is suitable for public water systems and irrigation use without treatment.

The Madison Limestone outcrop area is about 35 square miles, mostly in the Black Hills, and is buried at depths exceeding 5,000 feet below land surface in the northeastern part of the study area. The Madison Limestone consists of massive dolomite as well as limestone beds and is about 370 to 850 feet thick. It is an aquifer where fractures or solution features have increased the permeability. The general direction of water movement in the Madison aquifer is to the east and southeast. Wells flow in most of the study area in South Dakota and in approximately the northern one-half of Crook County, Wyoming. Estimates of transmissivity of the Madison aquifer from this and previous studies range from 0.01 to 8,600 feet squared per day (table 3). Storage coefficient was estimated by Woodward-Clyde Consultants (1980, p. 4-13) to range from 2×10^{-4} to 3×10^{-4} and 2×10^{-5} by Blankennagel and others (1981, p. 50). The Madison aquifer yields in excess of 500 gallons per minute to wells. A yield of 2,000 gallons per minute and a water level of 590 feet above land surface was reported in 1952 for a well located on the Whitewood anticline and open to both the Minnelusa aquifer and the upper 350 feet of the Madison aquifer. The aquifers may be more fractured and more permeable on the fold, hence the high yield. The principal deterrents to the use of water from the Madison are fluoride, gross alpha radiation, and hardness in the northern part of the study area. In the southern one-half of the study area, water quality generally is suitable for public water systems and irrigation, though hard (hardness 120 to 180 milligrams per liter as calcium carbonate) to very hard (hardness greater than 180 milligrams per liter as calcium carbonate).

A digital model, which uses a computer program to solve the equations for ground-water flow, was used to investigate the hydrology of the three aquifers. The model improved the overall understanding of the hydrology by testing the interpretation of the data and the interpretation of the ground-water hydrology. The model, which was "quasi-three-dimensional" and used a program by McDonald and Harbaugh (1984), simulated the system as three aquifers, in which flow is predominantly horizontal, separated by leaky confining beds, in which flow is predominantly vertical.

The absolute value of the differences between the measured water levels of the Inyan Kara aquifer and the water levels computed by the model range from 0 to 272 feet, with an absolute mean difference of 65 feet. Recharge to the outcrop was simulated as about 12 cubic feet per second. Approximately 13 cubic feet per second was simulated to flow out of the model area within the Inyan Kara aquifer. Flow between the Minnelusa and the Inyan Kara aquifers appears to be insignificant.

The absolute value of the differences between the measured water levels of the Minnelusa aquifer and water levels computed by the model ranged from 2 to 930 feet, with an absolute mean difference of 157 feet. The computed water levels near Spearfish are as much as 500 feet lower than the measured water levels. Recharge to the outcrop was estimated to be about 41 cubic feet per second. Unsuccessful attempts to improve the model by changing the boundary conditions to allow increased recharge on the outcrop indicated that there may be significant recharge from another source, such as leakage from the Madison.

The absolute value of the differences between the measured water levels of the Madison aquifer and water levels computed by the model ranged from 25 to 1,136 feet. The absolute mean difference was 319 feet. The largest differences are in the outcrop, where springs, assumed to represent discharge from a regional flow system, may be separated from the regional system by an unsaturated zone or may represent significant vertical movement within local flow systems. Recharge to the Madison outcrop was simulated as 16 cubic feet per second, however, changing the boundary conditions to allow increased recharge on the outcrop indicated there may be interaction with other aquifers, perhaps the underlying Whitewood Dolomite and Deadwood Formation.

The model results indicated the relationship between the Minnelusa, the Madison, and perhaps deeper aquifers is not understood. Further investigations of the hydrologic properties of the lower part of the Minnelusa Formation and the gradient between the Madison and Minnelusa aquifers would help clarify this relationship, which appears to be more complicated than assumed by this and previous investigations. The assumption that there is no interaction between the Madison aquifer and deeper aquifers, probably in the Whitewood Dolomite and Deadwood Formation, may not be valid. Testing of these deeper units to determine their geohydrologic properties and their relationship to the Madison would be useful to a complete evaluation of potential ground-water supplies in the area. Recharge to the Madison and Minnelusa aquifers is poorly understood. Quantifying recharge probably will require a better understanding of the interaction between the two aquifers.

Diversion of water in Spearfish Creek around aquifer outcrops has reduced recharge to the Minnelusa and Madison aquifers in the twentieth century. The potentiometric surface of the Minnelusa aquifer is known to have declined since installation of the diversion. The potentiometric surface of the Madison aquifer probably also has declined. A spring in Spearfish ceased to flow in the 1930's, probably in response to the lowering of the potentiometric surfaces. This series of events--diversion of flow, the consequent drawdown, and reduction in spring discharge--illustrates the close relationship of ground water and surface water in the study area.

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