GEOHYDROLOGIC EVALUATION OF THE UPPER PART OF THE MESAVERDE GROUP, NORTHWESTERN COLORADO

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CONVERSION FACTORS

Inch-pound units in this report may be converted to metric (International System) units by using the following conversion factors:

Multiply	*By*	*To obtain*
acre	0.4047	hectare
cubic foot per second (ft^3/s)	0.028317	cubic meter per second
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot per year (ft/yr)	0.3048	meter per year
per foot (ft^{-1})	3.281	per meter
foot squared (ft^2)	0.0929	meter squared
foot squared per day (ft^2/d)	0.0929	meter squared per day
gallon per minute (gal/min)	0.06309	liter per second
inch (in.)	2.540	centimeter
inch per foot (in/ft)	8.333	centimeter per meter
inch per year (in/yr)	2.540	centimeter per year
inch squared per pound (in^2/lb)	0.1450	$kilopascal^{-1}$
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer
ton, short (t)	0.9078	metric ton

Temperature in degree Fahrenheit (°F) may be converted to degree Celsius (°C) by use of the following equation:

$$°C = 5/9(°F-32)$$

Temperature in degree Celsius (°C) may be converted to degree Fahrenheit (°F) by use of the following equation:

$$°F = 9/5(°C)+32.$$

The following terms and abbreviations also are used in this report:
- microsiemens per centimeter at 25 degrees Celsius (µS/cm)
- milligrams per liter (mg/L)
- millidarcys (mD)

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

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ABSTRACT

Coal mining in Routt and Moffat Counties of northwestern Colorado has produced large areas of spoils and disturbed land that have the potential of degrading the surface and ground-water quality of the region. This investigation of the geology and hydrology of the bedrock aquifers in the area was undertaken to define the important characteristics of the hydrologic system and to evaluate the future impacts of mining on water quality.

Regional aquifers in the Trout Creek Sandstone Member of the Iles Formation and Twentymile Sandstone Member of the Williams Fork Formation and an important local aquifer are the principal water-yielding units in the 2,000-foot-thick sequence of shale, sandstone, and coal underlying the study area. The structural complexity of the region, coupled with rugged topography, cause the irregular outcrop of the aquifer units, primarily on the back slopes of the cuestas and elevated limbs of several anticlines. The aquifers are recharged by infiltration of precipitation in the elevated outcrops. Ground water generally moves at rates of 1 to 30 feet per year toward topographically low areas in Twentymile Park and the valleys of the Yampa River and its local tributaries. Discharge occurs by upward leakage through confining layers, lateral flow to stream valleys on low-lying outcrops, and evapotranspiration.

Solute-transport modeling indicates that movement of poor quality water from spoil aquifers will not significantly degrade the water quality in the bedrock aquifers. Mining primarily will affect surface-water quality through direct discharge of poor quality water into the streams from springs and seeps that develop in the spoil.

INTRODUCTION

Large reserves of bituminous to subbituminous coal are present in the upper members of the Cretaceous Mesaverde Group in northwestern Colorado (pl. 1). In the Williams Fork Mountains of Routt and Moffat Counties, coal production increased by 260 percent from 1970 to 1980, at a time when total coal production in the United States increased by about 50 percent. Three large open-pit mines and several smaller mines in Routt County produced 4 to 7 million tons of coal per year from 1980 to 1986. Past mining activities in the county have produced in excess of 9,000 acres of mine spoils and disturbed land. The areal extent of these areas can be expected to increase in size as mining continues. Mine spoil and disturbed land have the potential to degrade ground-water and surface-water quality by providing increased potential for leaching of soluble minerals.

Private industry, Federal, State, and local regulatory agencies, and the general public are faced with growing needs for hydrologic information pertaining to the natural environment of coal producing regions and the effects of mining-imposed changes on the environment. A study by the U.S. Geological Survey, in cooperation with the Colorado Department of Natural Resources, Mined Land Reclamation Division, the U.S. Office of Surface Mining Reclamation and Enforcement, and the U.S. Bureau of Land Management was done to meet such needs in Routt and Moffat Counties through an investigation of the geology and hydrology of the Williams Fork Mountain coal region (fig. 1). The study involved a detailed investigation of the ground-water hydrology of the eastern part of the area, where coal has been mined for almost a century and for which geohydrologic data are prevelant, and a more general overview of the geology and hydrology of the western part of the area, where mining has not been extensive and for which geohydrologic data are sparse.

The objectives of the more detailed investigation of the eastern part of the area include:

1. Defining the extent, thickness, lateral continuity, and structural configuration of the principal bedrock aquifers;
2. Mapping aquifer characteristics, potentiometric surfaces, and dissolved-solids concentrations in the principal bedrock aquifers;
3. Estimating the water budget and the rate and direction of ground-water movement for the area;
4. Defining dominant water-chemistry composition, dissolved-solids concentrations, and principal geochemical mechanisms; and
5. Estimating the effects of mining activities on ground-water levels and dissolved-solids concentrations in the bedrock aquifers by use of mathematical models of the aquifers.

Objectives of the general overview of the western part of the area include:

1. Defining the extent, thickness, and lateral continuity of the principal bedrock aquifers;
2. Determining the general hydrologic relations between components of the hydrologic system; and
3. Determining general directions of ground-water flow.

Purpose and Scope

This report describes the characteristics of the hydrologic system in the study area. The hydrologic characteristics are based on hydrologic data that consisted of approximately 400 lithologic or geophysical well logs, 2,400 water-level measurements made in cased wells, 1,600 chemical analyses of ground- and surface-water samples, and other published or unpublished documents, maps, and tables. Some of the data are proprietary and confidential. The majority of the data pertain to the eastern part of the study area. The availability of data affects the hydrologic interpretations that can be made and is the principal reason for the differences in study objectives for the eastern and western parts of the area. The hydrologic characteristics of the eastern part of the study area were corroborated and better defined by use of mathematical models of the ground-water flow and solute-transport systems.

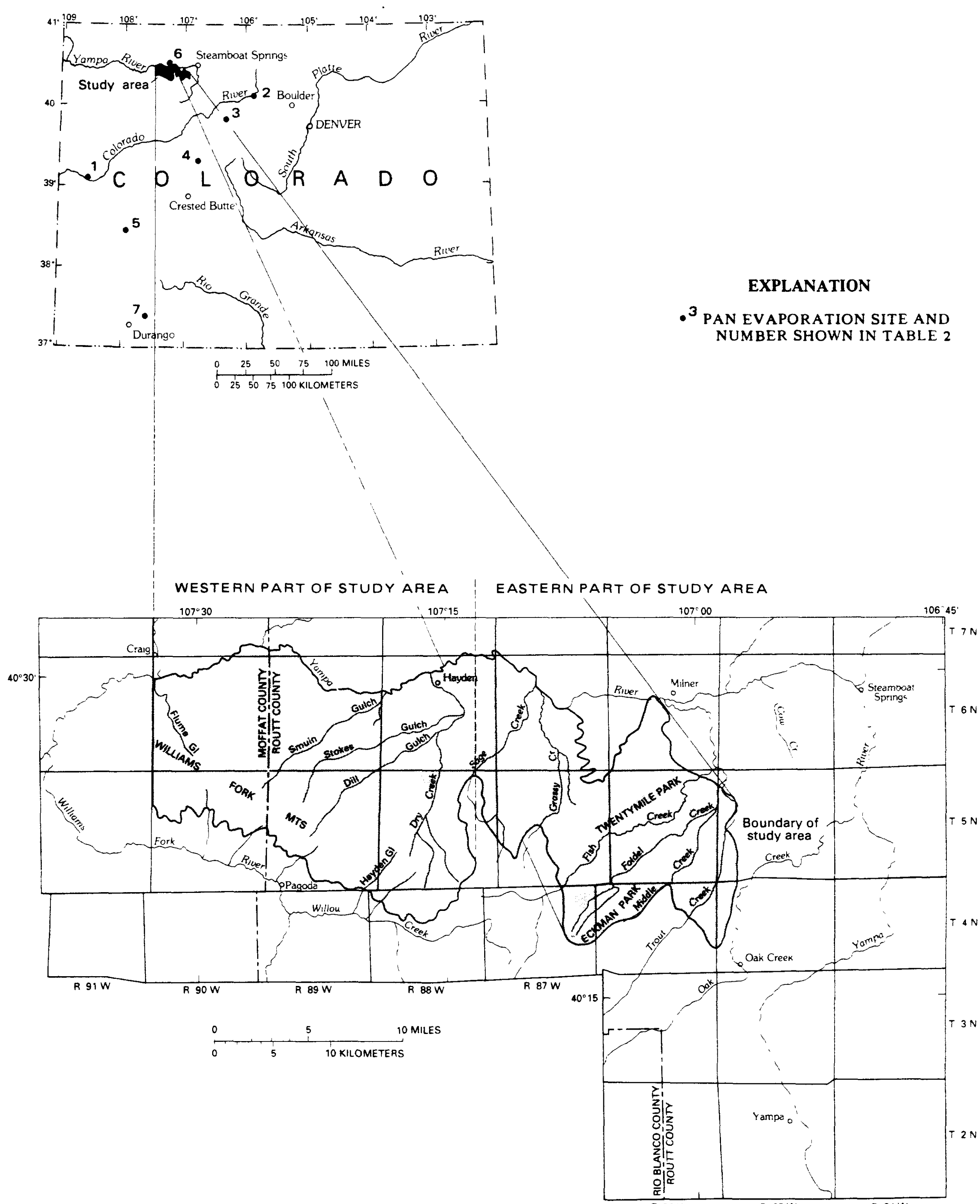


Figure 1.--Location of the study area.

Location

The 280-mi^2 study area is located in Routt and Moffat Counties in northwestern Colorado (fig. 1). The area is east of Craig and is bounded on the north by the Yampa River and on the northeast and south by the outcrop of the Trout Creek Sandstone Member of the Iles Formation. The Williams Fork Mountains have altitudes of more than 8,300 ft and extend from south of Craig to the southern margins of Twentymile Park (a broad intermountain valley in the eastern part of the area). The study area is drained by numerous ephemeral or discontinuous perennial streams. Trout Creek and its tributaries, Fish, Foidel, and Middle Creeks, are the principal perennial streams in the area.

Previous Research

Previous research within the area generally concerned evaluation of coal, oil, and gas reserves. Extensive coal reserves in the Williams Fork Mountains have attracted the attention of geologists since the 19th century. Coal investigations in the Williams Fork Mountains through the early 1920's are described in Bass and others (1955):

> The general region was traversed and mapped geologically by S.F. Emmons (1887), geologist with the 40th parallel survey in 1872, 4 years before Colorado was granted statehood. A geologic description, including a map, is given in his report on the region. Four years later the region was visited by C.A. White (1878 and 1889), a geologist with the Hayden survey. Topographic and geologic maps and descriptions, which are contained in reports of that survey, call attention to the extensive coal deposits.
>
> In the late eighties and early nineties, rumors that a railroad would be built into this region stimulated exploration, immigration, and settlement. Geologists and mining engineers employed by the proposed Denver, Northwestern Pacific (later the Moffat) Railroad investigated the resources of the area. From 1886 to 1905 several articles about coal in the area were published. These included papers by F.F. Chisholm (1886), L.S. Storrs (1902, p. 435-436), G.C. Hewett (1889, p. 376), R.C. Hills (1893, p. 354-358), H.F. Parsons and C.A. Liddell (1903), and W. Weston (1904), 1909, and 1914). A geologic report describing the coal deposits of the area was published by the U.S. Geological Survey in 1906 (Fenneman and Gale, 1906). Exploitation of the coal on a relatively large scale followed the arrival of the railroad in 1906. The coal in and near Twentymile Park was described by Campbell (1923).

Following Campbell's report, little work pertaining to coal was done within the area until the mid-1950's and the publication of a U.S. Geological Survey Bulletin by Bass and others (1955). Later investigations of coal reserves include work done by Horn (1959), Miller (1975), and Ryer (1977). In 1977-78, the U.S. Geological Survey Conservation Division conducted an extensive drilling program and published geological and geophysical information pertaining to all the holes (Brownfield, 1978a, 1978b; Bronson, 1979).

In 1979-80, Dames and Moore prepared several quadrangle coal-resource maps that were published by the U.S. Geological Survey (Dames and Moore, 1979, 1980a-h).

Investigations of oil and gas reserves began in the 1920's with studies on anticlines in the area (Crawford and others, 1920; Willson, 1920; Collins, 1921). Later, Sears (1924) published a report on the geology and gas prospects in the area. Parts of the Williams Fork Mountains were included in oil and gas investigation maps by Bradley (1945) and Dyni (1966).

Numerous theses have been written about parts of the area, including the works of Willson and Collins mentioned above. Blackmer (1939), Beattie (1958), Kerr (1958), Kucera (1962), Lauman (1965), Buffler (1967), Masters (1967), and Kiteley (1980) all wrote geological theses pertaining to parts of the study area.

Examination of surface and subsurface hydrology did not begin until the mid-1970's. Brogden and Giles (1977) published a reconnaissance ground-water hydrology report about a large area of Routt and Moffat Counties, which included most of the study area. Hounslow and Fitzpatrick (1978) and McWhorter and others (1979) published reports containing hydrologic information collected within the area. A regional environmental impact statement (U.S. Department of the Interior, 1976) contained some regional hydrologic information, while several unpublished site-specific studies for permit applications examined the hydrology of areas likely to be affected directly by mining activity. Warner and Dale (1981) made the first attempt to model the area in order to predict effects of mining on ground-water quality; however, their results were compromised by lack of data.

Acknowledgments

Some of the data used in this work were provided to the U.S. Geological Survey by the Colorado Yampa Coal Co. (CYCC), Twentymile Coal Co., Pittsburg and Midway Coal Mining Co., Peabody Coal Co., U.S. Bureau of Land Management, and the State Engineers and Mined Land Reclamation Division offices of the Colorado Department of Natural Resources. The helpful assistance and cooperation of members of these organizations is gratefully acknowledged.

Most of the results of this study that pertain to the western part of the study area were compiled and developed by the coauthor in 1978-80 (Stewart, 1983) while he was employed by the U.S. Geological Survey. Stewart's work was done in cooperation with the U.S. Bureau of Land Management. In addition, Mr. Robert S. Williams and Dr. Keenan Lee aided in field-data collection or provided guidance and ideas invaluable to the completion of the report.

DESCRIPTION OF STUDY AREA

Topography

Topography in the study area ranges from gently sloping valleys to rugged mountains and vertical cliffs. The form of the land surface is greatly affected by the lithologic and structural setting of the local area. For example, Twentymile Park is underlain by easily eroded shales that overlie the axis of a syncline (pl. 1). As a result, the area is characterized by low-relief, gently rolling terrain (fig. 2). The regional surface slopes gently toward the east and from the valley margins to the center of the park, where altitudes are about 6,800 ft.

Figure 2.--Rolling topography of Twentymile Park.

A second topographic form occurs in the part of the study area southeast of Trout Creek and in Eckman Park. In these two locations, gently dipping beds form cuestas, which are cut by subparallel subsequent streams. The topography of the dip slopes of the cuestas is smooth and has low to moderate relief between streams that drain the slopes. In both locations, the dip slopes have regional surface gradients toward the northwest and have altitudes in excess of 8,000 ft. The eroded back slopes of the cuestas have much steeper gradients and shorter streams. To the north of Foidel Creek, the resistant Twentymile Sandstone Member and overlying members of the Williams Fork Formation are exposed in massive sandstone cliffs that exceed 300 ft in height (fig. 3).

Figure 3.--Outcrop of Twentymile Sandstone Member of the Williams Fork Formation north of Foidel Creek.

A third topographic form in the eastern part of the area results from the erosion of the surficial parts of several anticlines. The topography is characterized by deeply incised slopes and cliffs, which produce a rough, high-relief terrain. No regional topographic gradient is present; instead, the local gradient depends on the location of underlying folds. The roughest terrain of this type occurs on the eastern limbs of the Sage Creek, Fish Creek, and Tow Creek anticlines (pl. 1; fig. 4).

Topographic configurations west of Dry Creek result from erosion of regional cuestas. The cuestas are surficial configurations of the regional Sand Wash Basin structure. The topography is characterized by flat, low-gradient ridges separated by narrow steep-sided alluvial valleys. Relief increases toward the south where gradients are steepest in cuesta escarpment areas along the southwestern margin of the study area. Here, southward-flowing streams have cut several thousand feet into the cuesta, producing steep, narrow, valleys surrounded by cliffs. The regional gradient of the entire western dip slope area is to the northeast. The cuesta escarpment area trends toward the southwest, from the approximately 8,000-ft divide of the Williams Fork Mountains.

Population Distribution and Land Use

The study area is sparsely populated and relatively remote. The only towns near the area are Craig, Hayden, Milner, and Steamboat Springs, which are located along the Yampa River to the north of the area, and Oak Creek,

Figure 4.--Rugged topography and steeply dipping beds near the axis of the Sage Creek anticline.

which is located to the southeast of the area. These towns had a combined population of 15,880 in 1980 (U.S. Bureau of the Census, 1981). Within the area, population is limited to scattered ranches and farms; 82 percent of the land in the study area is privately owned. Only one paved county road traverses the area. Most transportation is by means of a sparse network of improved and unimproved dirt roads.

Vegetation type and density varies with altitude, topography, and slope aspect. In areas of lower altitude and minimal topographic relief, vegetation consists of sagebrush and meadow grasses. On higher, steeper slopes, sagebrush is replaced by mountain shrubs such as Gambel Oak, serviceberry, and snowberry. At still higher altitudes, and on the lower north-facing slopes, sparse to dense groves of aspen and conifers are present.

Much of the area is used for grazing of cattle and sheep; dryland farming is limited to part of the lower altitude grassland areas. Several large open-pit mines are operating in the study area. Numerous small open-pit or underground mines have been active in the past. Mining has produced about 7,000 acres of mine spoils and disturbed land in the eastern part of the study area. Spoils are regraded and revegetated at operating mines (fig. 5), but unaltered spoils still are present at a few long-abandoned mines. No organized recreational facilities occur in the area; however, big game hunting is popular during the fall, and several professional outfitters lease large tracts of ranchland for commercial deer and elk hunting.

Figure 5.--Spoil piles and regraded spoil at the Edna Mine south of Trout Creek.

Mineral and Energy Resources

Primary mineral and energy resources include oil, gas, and coal; coal is the dominant resource. The existence of these resources is the main reason for the many geological investigations undertaken in the area.

Oil and gas occur in the Tow Creek and Buck Peak Fields. The Buck Peak Field, first developed in 1956, is on the axis of the Buck Peak anticline, T. 6 N, R. 90 W. The southern part of the Tow Creek Field, in T. 6 N, R. 87 W, is within the study area. The Tow Creek Field, located on the Tow Creek anticline axis, was first developed in 1924 and has not been as productive as the Buck Peak Field (Donaldson and MacMillan, 1980).

Routt County contains the largest strippable and underground coal reserves in Colorado, estimated at 413 million and 3,826 million tons, respectively (Green and others, 1980). Coal was first mined in Routt County in the late 1880's, and production increased by several orders of magnitude after completion of the railroad into the area in 1906. Production remained relatively constant until the late 1940's when it began to decrease until the late 1950's (fig. 6). Marked recovery began in the early 1960's and continued through 1980 (Martin, 1980).

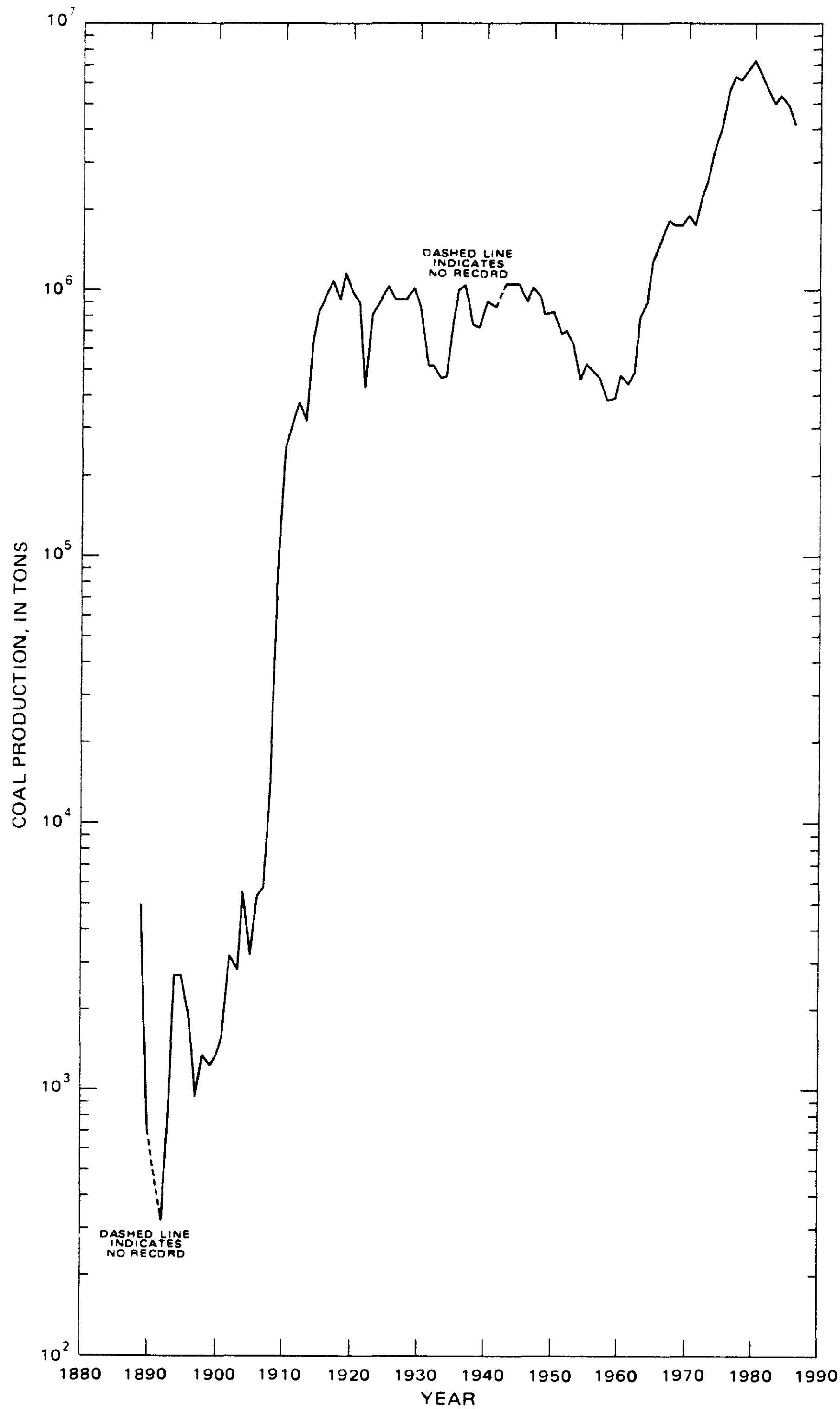


Figure 6.--Coal production in Routt County, 1889-1987.

REGIONAL GEOLOGIC SETTING

The Williams Fork Mountains are at the extreme southeastern end of the Sand Wash basin, one of several basins within Colorado that contain Cretaceous rocks. The Sand Wash basin is bordered on the east by the Park Range and on the south and west by the White River uplift and the Axial anticline (fig. 7).

Depositional History

Rocks of the Cretaceous Iles and Williams Fork Formations constitute the Mesaverde Group. These rocks and the overlying Lewis Shale were deposited during a 5-million-year timespan that began approximately 70 million years ago (Berman and others, 1980). Marine and nonmarine deposition occurred during two major regressive-transgressive phases extensive enough to move the strandline through the area. The first regressive phase began with the strandline situated 25 mi west of Craig, trending northeast to southwest (fig. 8, line 1). The seas regressed eastward out of the study area, and local deposition occurred under nonmarine deltaic conditions. A subsequent transgression moved the strandline back through the area, until the strandline was 10 mi west of Craig (fig. 8, line 3). A second regression moved the strandline back to the east, again resulting in nonmarine conditions prevailing in the study area (fig. 8, line 4). A final westward transgression resulted in the return of marine conditions and moved the strandline west of Craig (fig. 8, line 5).

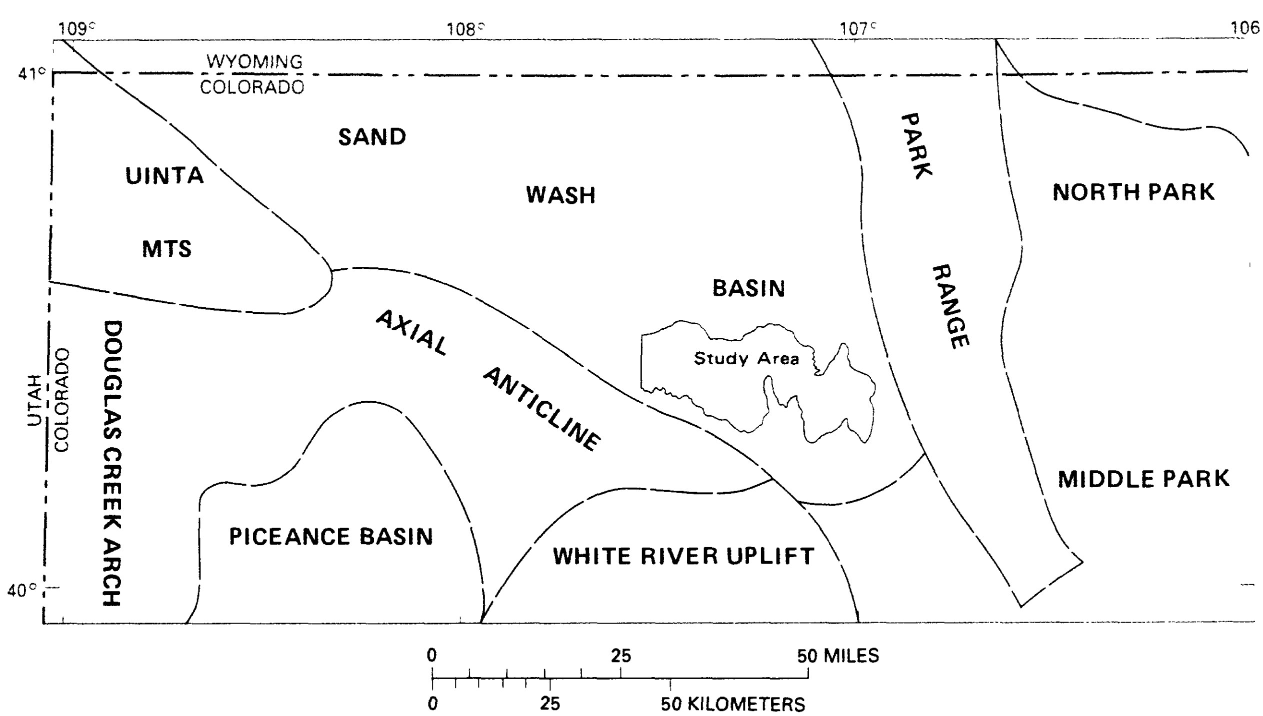


Figure 7.--Regional structural and physiographic setting.

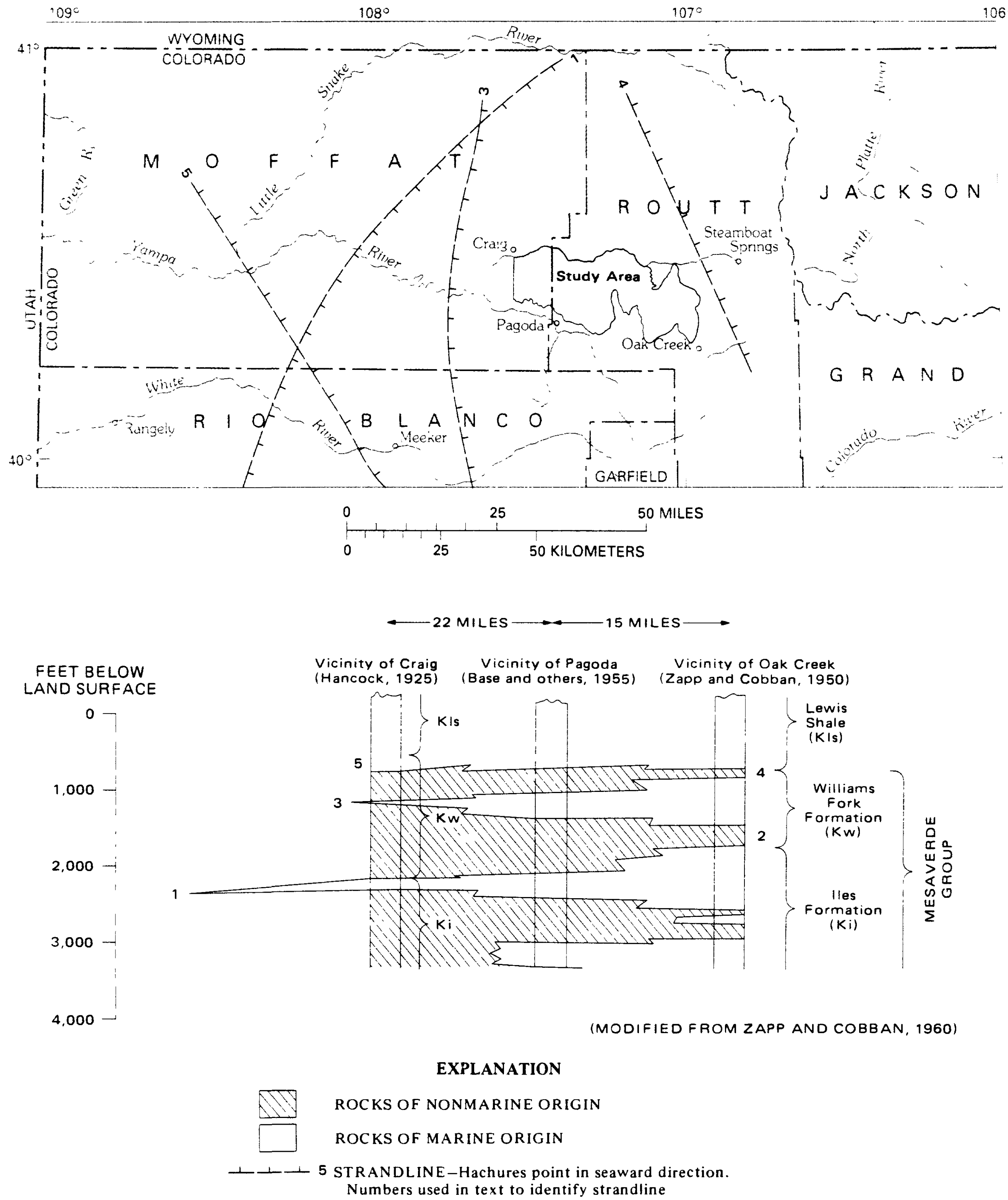


Figure 8.--Strandline boundaries for regressive and transgressive phases during deposition of study area formations.

The resulting stratigraphy has marine deposits that thicken toward the east and nonmarine deposits that thicken toward the west (fig. 8). The thick Trout Creek Sandstone Member of the Iles Formation and Twentymile Sandstone Member of the Williams Fork Formation were formed near the landward margins of the marine rocks at the regressive (upper) boundary of the marine sequence.

Stratigraphy

The multiple migrations of strandlines through the area resulted in stratigraphic relations that are complex and often poorly correlated. Sediments deposited during nonmarine conditions sometimes are of varied lithology, limited lateral continuity, and contain many facies changes. Also, numerous minor transgressive-regressive pulses during deposition produced local strandline migrations superimposed on the larger phases. The deposits are classified into two thick beach sandstones (the Trout Creek Sandstone Member of the Iles Formation and the Twentymile Sandstone Member of the Williams Fork Formation), three thick marine shales (those underlying the Trout Creek and Twentymile Sandstone Members, and the Lewis Shale), and several intervening sections that contain marine and nonmarine rock.

Iles Formation

Trout Creek Sandstone Member

The Trout Creek Sandstone Member is the upper part of the Iles Formation (pl. 1) and is the basal unit studied in this work (fig. 9). Type locality for the Trout Creek Sandstone Member is in the northeastern part of Twentymile Park along Trout Creek (Fenneman and Gale, 1906, p. 26). The unit thickness is fairly consistent, and this bed is considered the most reliable marker bed within the area (Bass and others, 1955, p. 155). The Trout Creek conformably overlies marine shales of the main body of the Iles Formation. The upper contact of the Trout Creek is conformable and very distinct and is the boundary between the Iles and Williams Fork Formations.

The Trout Creek Sandstone Member consists of massive, white to light-gray, moderately well-sorted, fine- to very fine-grained quartz arenite. The sandstone consists of about 90 percent subangular quartz and 10 percent black subangular chert. Individual sandstone grains are undeformed and have tangential grain-to-grain contacts, which indicates that little or no compaction has occurred. The few sedimentary structures present include trough cross-bedding and planar laminations (Ryer, 1977). Widely spaced fractures were present in some outcrops. Silica cementation normally is present but varies in amount at different locations. As a result, samples range from friable to well indurated; almost all surface samples are moderately to well indurated. Core samples generally are well indurated. Sandstone thicknesses reported in the literature seem to indicate a regional eastward thickening, from 75 ft at Pagoda (Konishi, 1959) to 132 ft in the vicinity of Oak Creek (Kucera, 1959); however, local variation in thickness is substantial. For example, the sandstone isolith map (fig. 10) indicates a sandstone thickness of less than

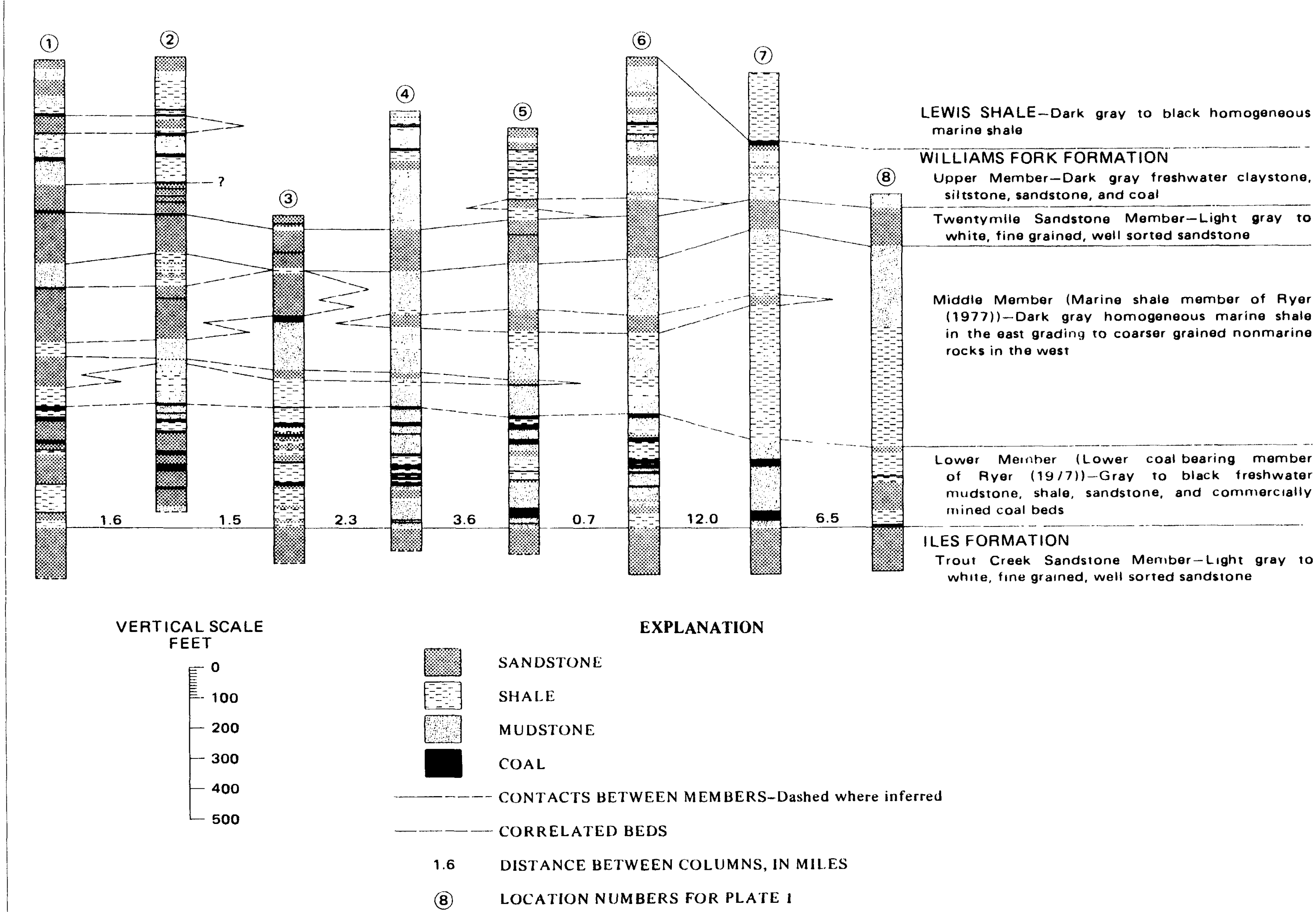


Figure 9.--Composite stratigraphic columns.

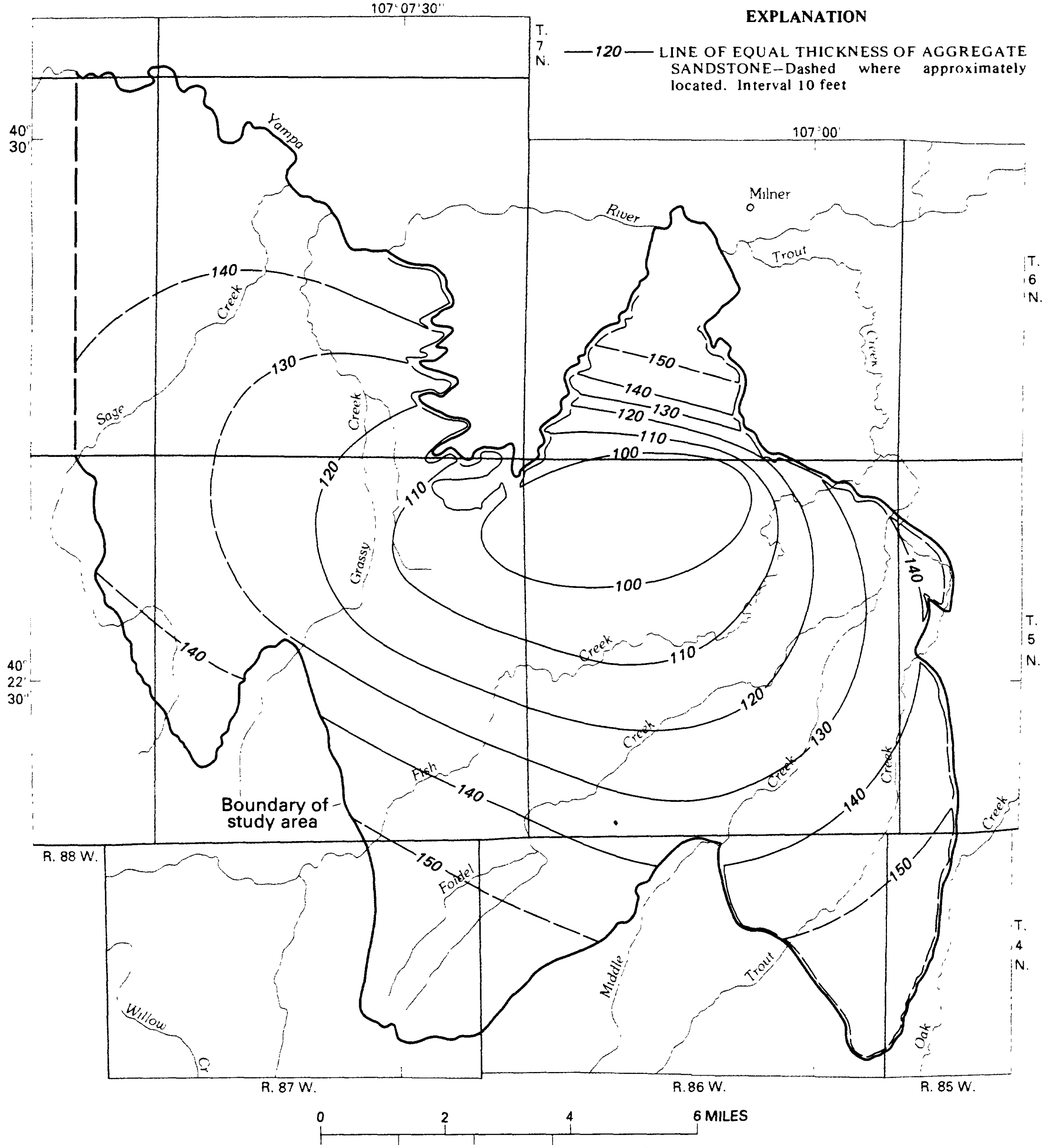


Figure 10.--Aggregate sandstone thickness of the Trout Creek Sandstone Member of the Iles Formation in the eastern part of the study area.

100 ft in the north-central part of Twentymile Park; thickness increases to 140 ft or more near the northern and southern outcrops. Data are inadequate to map sandstone thickness in the western part of the study area, and the regional trend in thickness is uncertain.

Williams Fork Formation

Most of the rocks exposed in the study area are part of the Williams Fork Formation (pl. 1). Rocks of the Williams Fork Formation were first named by Hancock (1925). The upper and lower formational contacts are conformable. The lower contact, with the Iles Formation, is distinct and is easily identified by the relatively coarse grain size and presence of black chert in the underlying Trout Creek Sandstone Member (Ryer, 1977). The upper contact, with the Lewis Shale, is transitional; the criteria used by Bass and others (1955) for separating the two formations are unknown. The sediments underlying the contact are nonmarine; the Lewis Shale is marine. The transitional zone between the two is about 10 ft thick, defining a relatively narrow zone in which to place the actual contact. The thickness of the Williams Fork Formation ranges from 1,100 ft at Mount Harris to 2,000 ft at the western study area boundary (Bass and others, 1955, p. 157). The increase in thickness occurs at the top of the formation where the formation thickens and the Lewis Shale thins. The Williams Fork Formation in the study area originally was classified in three segments (Bass and others, 1955); however, the fourfold classification used by Ryer (1977) is more representative and is used here. The four segments are the lower coal-bearing member (hereinafter referred to as the lower member), the middle shale member (hereinafter referred to as the middle member), the Twentymile Sandstone Member, and the upper member (fig. 9).

Lower member

The lower member contains extensive reserves of bituminous coal (Bass and others, 1955). The lower boundary is the distinct contact between the underlying Trout Creek Sandstone Member of the Iles Formation, a beach deposit, and the finer grained deposits of the lower member (Ryer, 1977). In the eastern part of the study area, the upper contact is between the nonmarine sandstones and mudstones of this member and the overlying marine shales. In the western part of the study area, where the overlying marine shale is absent, the contact is arbitrarily set approximately 50 ft above the uppermost thick coal seam (fig. 9). The dominant lithologies are gray to black siltstones, silty, fine-grained sandstones, and limey shales interbedded with coal seams. Toward the west, the section becomes sandier and coalbeds tend to be thinner and more numerous. The thickness of this member ranges from 300 ft in the east to 450 ft in the west, primarily because of facies changes across the area. In the eastern part of the area, data enable mapping, and the total sandstone thickness in this member ranges from 100 to 200 ft, thickening to the west (fig. 11). Shale thickness ranges from 100 to 200 ft, thickening to the east (fig. 12).

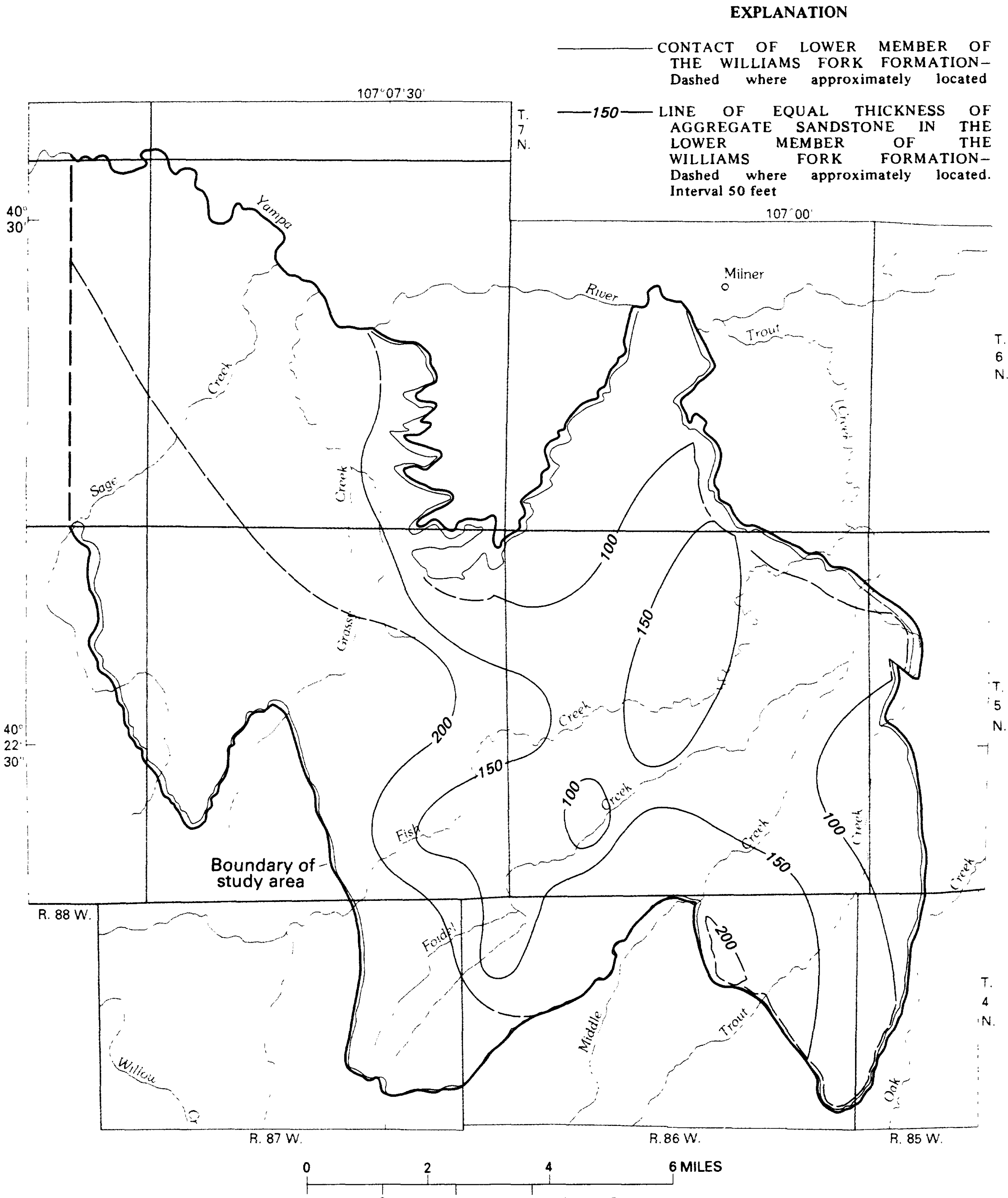


Figure 11.--Aggregate sandstone thickness of the lower member of the Williams Fork Formation in the eastern part of the study area.

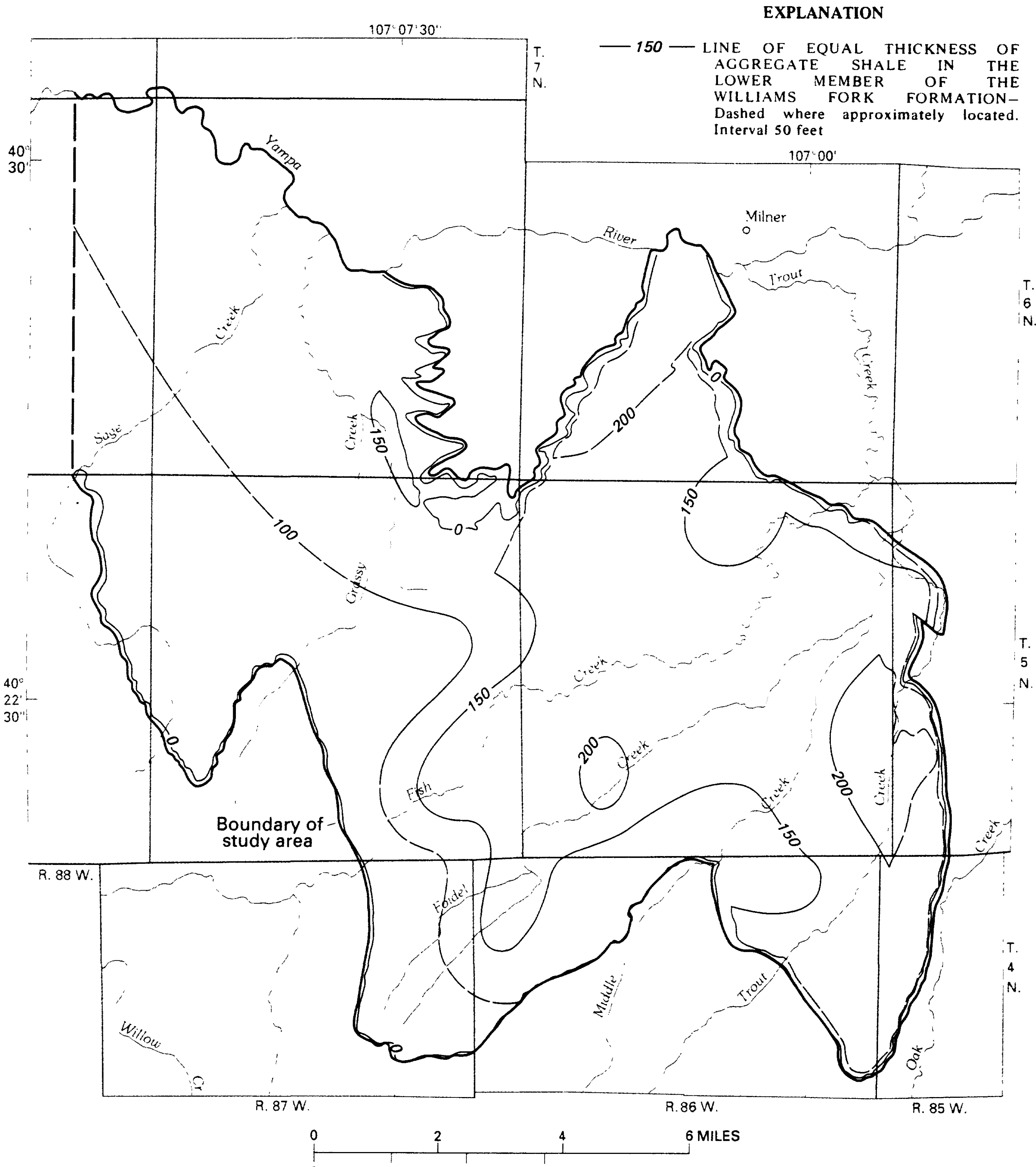


Figure 12.--Aggregate shale thickness of the lower member of the Williams Fork Formation in the eastern part of the study area.

Coal within this interval is mined extensively in the eastern part of the study area. Three seams--the Wolf Creek, the Wadge, and the Lennox--are extensive and continuous enough to have been named. The Wolf Creek coal is located 40 to 100 ft above the top of the Trout Creek Sandstone Member. This seam ranges from 0 to 18 ft in thickness over short horizontal distances and is of poor quality because of shale stringers. The Wolf Creek seam currently is not extensively mined. The Wadge coal seam lies 230 ft above the Trout Creek. It is a clean, continuous coalbed that ranges in thickness from 6 to 14 ft. This is the major source of coal at the three large operating open-pit mines and the one large underground mine in the eastern part of the study area. The Lennox seam is about 60 ft above the Wadge seam. It is about 4 ft thick and has been eroded away throughout most of the uplifted margin of the eastern area.

West of Hayden Gulch, coal seams in this member are thinner, more numerous, and generally not accessible by strip mining. Much less is known about these coals; therefore, correlation of the Wolf Creek, Wadge, and Lennox coals is not well defined west of Hayden Gulch.

Middle member

The middle member of the Williams Fork Formation is defined by the underlying contact with the lower member and an upper transitional, conformable contact with the overlying Twentymile Sandstone Member. Lithology of the middle member varies from marine shale in the eastern part of the study area to nonmarine gray siltstone, silty sandstone, and brown sandstone in the western part of the study area. There are few coal seams in this interval, and those present generally occur in the middle of the member in the far western part of the study area. Several sandstones 30 to 100 ft thick are present in the western part of the area. Thickness of this member ranges from 600 ft in the east to 450 ft in the west, primarily because of a facies change and stratigraphic climbing of the overlying Twentymile Sandstone Member. The middle member generally is 500 to 600 ft thick in the eastern part of the area (fig. 13). Outcrops of marine shale generally are less resistant than the outcrops of sandstone in the overlying and underlying units; the shales generally erode to form broad, gently sloping landforms. Shale thickness increases gradually across such outcrops.

Twentymile Sandstone Member

The Twentymile Sandstone Member, first named by Fenneman and Gale (1906), is very similar in appearance and origin to the Trout Creek Sandstone Member (Bass and others, 1955, p. 153); it is a white to light gray, moderately well-sorted, fine- to very fine-grained quartz arenite. The unit contains about 90 percent subangular quartz and 10 percent black, subangular chert and is moderately to well indurated. The cementing agent primarily is silica in the harder samples and clay in the softer samples. Tangential grain-to-grain contacts of outcrop samples indicate that little or no compaction has occurred. Thickness and character of the Twentymile Sandstone Member are more varied than in the underlying Trout Creek Sandstone Member. In the eastern part of the study area, the Twentymile Sandstone Member has an average thickness of

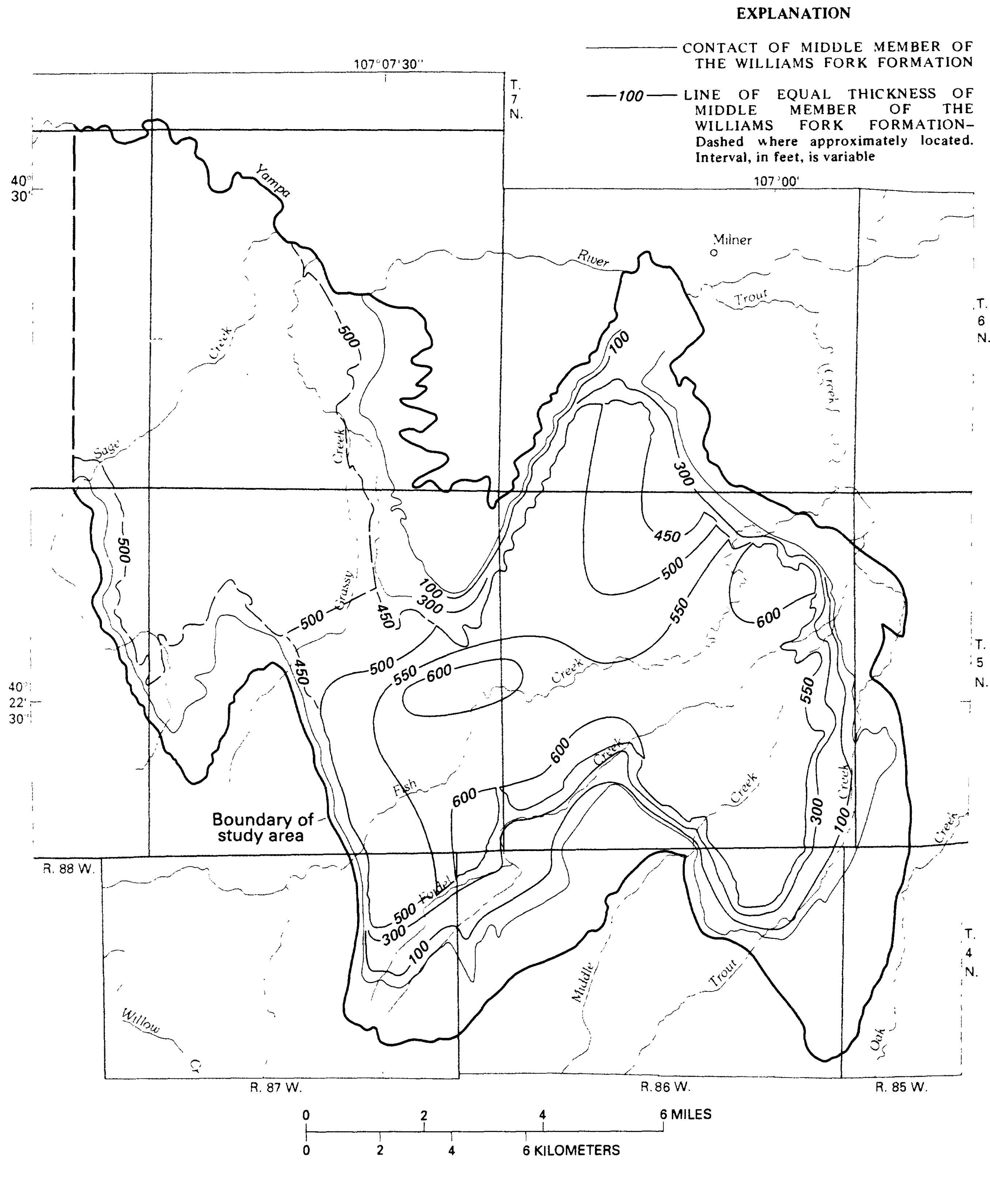


Figure 13.--Thickness of the middle member of the Williams Fork Formation in the eastern part of the study area.

about 100 ft and ranges in thickness from 80 to 180 feet. This large range and the seemingly random distribution of thickness preclude isopach mapping. The lower transitional contact between the beach sand of the Twentymile and the underlying marine shales is moderately well defined. The upper contact with the siltstones and fine-grained sandstones of the upper member is less well defined. In the western part of the area, the thickness is about 100 feet; however, rocks above and below the Twentymile Sandstone Member tend to be coarser silty sandstones or sandstones that produce poorly defined boundaries, particularly at the base.

Upper member

The upper member of the Williams Fork Formation includes all rocks between the top of the Twentymile Sandstone Member and the base of the Lewis Shale. Rocks in this member primarily are dark-gray mudstones, siltstones, and limey shales interbedded with sandstones 20 to 30 ft thick. Coal seams, some thick enough to be mined, occur near the top of the member in the east and from the base through the middle of the interval in the west. Thickness of the upper member increases from 300 ft in the east to 850 ft in the west. The combined thickness of the upper member and the Twentymile Sandstone Member is about 420 ft in Twentymile Park.

Lewis Shale

The Lewis Shale (pl. 1) is a dark-gray to black, homogeneous marine shale deposited during the last regional transgression (Zapp and Cobban, 1960). Erosional remnants of the lower part of the formation are located near the axis of the synclinal basin in Twentymile Park. A narrow outcrop of shale connects these exposures with the more extensive exposures located to the southeast of Hayden and Craig. The total thickness of the shale varies markedly throughout the area because of erosional thinning. In Twentymile Park, maximum thickness is about 700 ft; in the smaller synclinal basin to the west, a maximum thickness of about 500 ft is attained (fig. 14). The full stratigraphic thickness of the Lewis Shale is present only locally in the area east of Craig where the shale is conformably overlain by the Lance Formation. The shale attains a maximum thickness of about 2,300 ft in this area.

Lance Formation

The Lance Formation (pl. 1) is a transitional marine-deltaic sequence of interbedded gray shale and buff to tan, soft, fine-grained sandstone and a few coal beds (Bass and others, 1955). The only exposure of the formation in the study area occurs south of the Yampa River to the east of Craig, where it attains a maximum thickness of 300 to 400 ft.

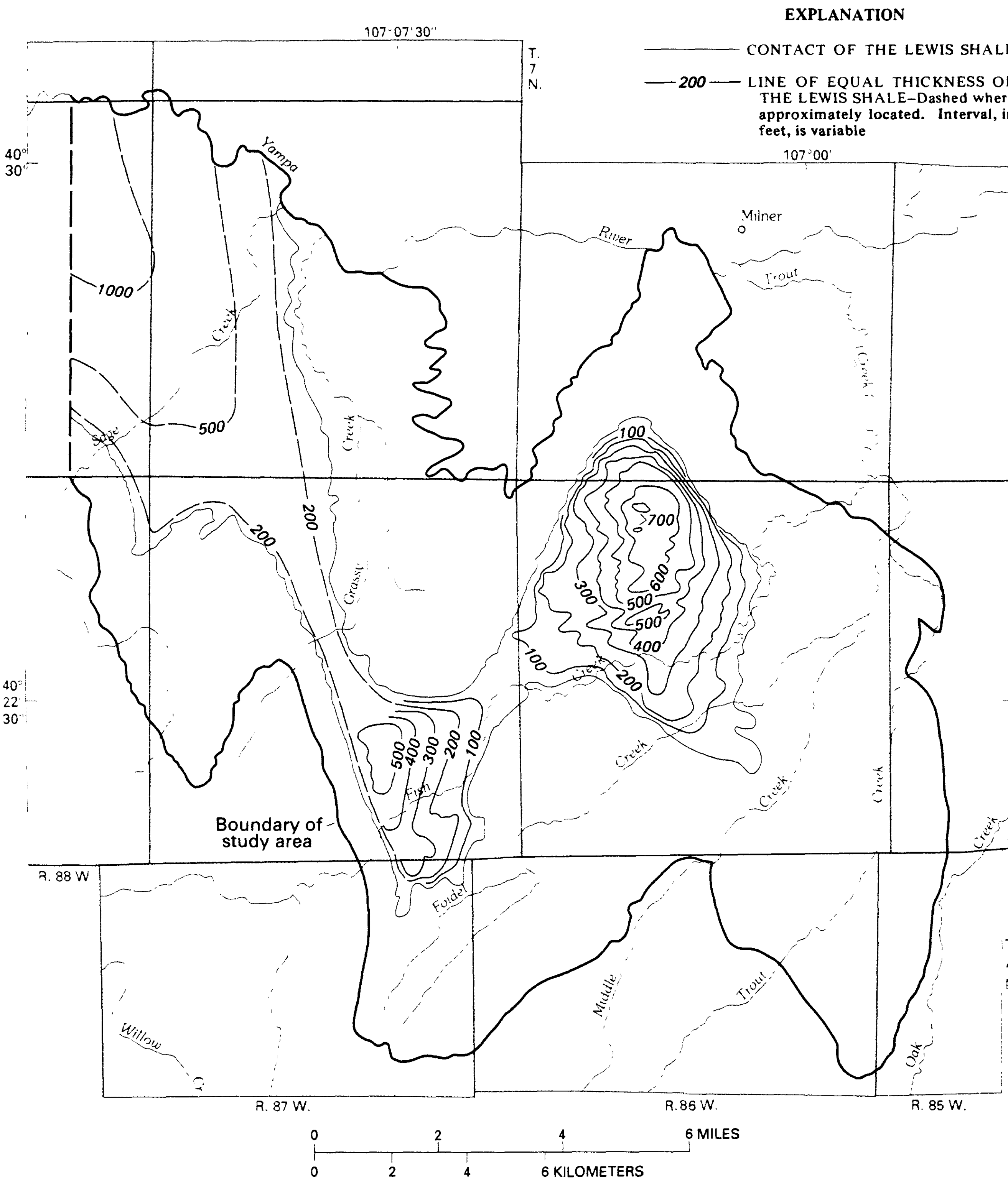


Figure 14.--Thickness of the Lewis Shale in the eastern part of the study area.

Structure and Faulting

Principal structural features within the study area are a result of the Laramide orogeny. This structural folding and mountain-building event began in Late Cretaceous time, 65 to 70 million years ago, and continued intermittently into late Eocene time (Tweto, 1980). The orogeny moved the epicontinental sea from Colorado for the final time. The resulting regional structure is shown in figure 7.

The major structure in the study area, the Hayden syncline (fig. 15), is the farthest southeastern extension of the Sand Wash basin. The Hayden syncline is located just east of Hayden, Colo. Smaller structures in the study area can be divided into eastern and western forms. These differing structural forms are important because they affect topography, vegetation, surface drainage, and ground-water movement.

The structural form of the study area from Hayden Gulch (about 10 mi southwest of Hayden) to the western boundary is basically a homocline dipping to the northeast at a 10 to 15° angle. One fold, the Buck Peak anticline (pl. 1), occurs in the far northwestern area. The structure of this anticline does not extend to the surface. The Buck Peak anticline axis trends northwest, parallel to regional strike, and oblique to minor fold axes in the west. Relief on this fold is estimated at 400 to 500 ft. A fault occurs just south of and parallel to the fold axis. Several smaller folds of similar alignment also are present.

The structural form of the area east of Hayden Gulch has a different origin and configuration. The primary tectonic feature affecting this region is the north-south trending Park Range (fig 7). Secondary structures, superimposed on the regional structure, complicate the structure in the eastern part of the study area.

Three generally north-south trending synclines are the principal secondary structures. The westernmost, here termed the Sage Creek syncline, is a northward-plunging asymmetrical syncline, underlying Sage Creek Reservoir. The asymmetry produces 50 to 60° dips and a northwestern strike in outcrops along the steeper western flank and 10 to 20° dips and a northeastern strike along the eastern flank (pl. 1). The second syncline seems to be a southward extension of the larger Hayden syncline. It also is northward plunging and asymmetrical. Outcrops on the western flank strike north to northwest and dip 50 to 60°; those on the eastern flank strike east to northeast and dip 10 to 25°. The Twentymile Park syncline is the largest and easternmost of the three synclines. It is a triple-plunging syncline that forms a small structural basin underlying Twentymile Park. The northward plunging southern limb is asymmetrical. Outcrops strike northeast and dip 20 to 35° along the eastern flank. The southward-plunging northern limb is symmetrical, although offset by faulting. Both flanks dip 10 to 35° to the southeast or southwest. The northernmost part of the syncline again plunges to the north, although structural features in this area are poorly defined.

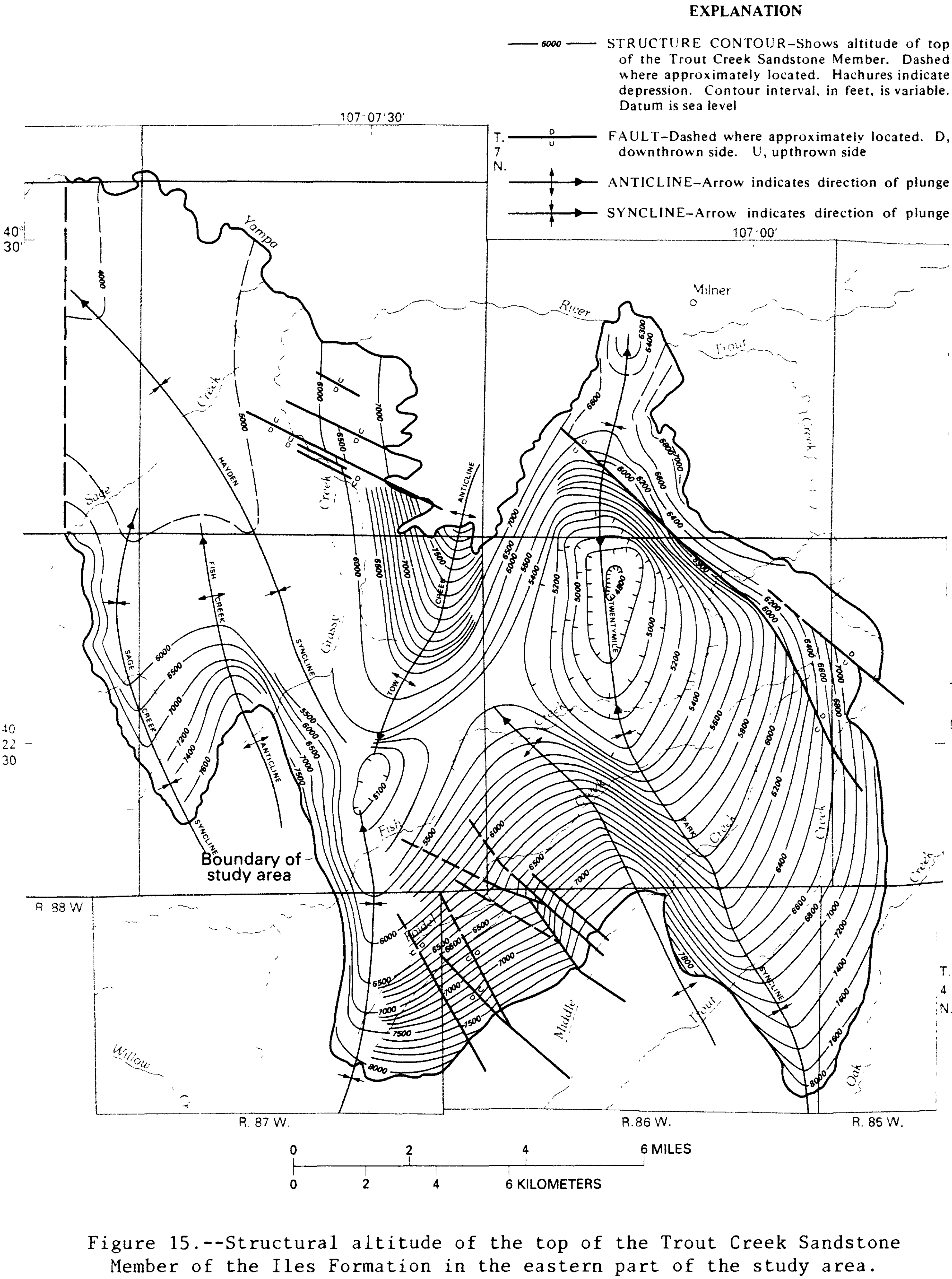


Figure 15.--Structural altitude of the top of the Trout Creek Sandstone Member of the Iles Formation in the eastern part of the study area.

Three principal anticlines occur in conjunction with the synclines in the eastern part of the area (pl. 1). The Tow Creek anticline plunges toward the southwest and is the largest of the four anticlines; it has 3,000 ft of vertical relief. The Tow Creek anticline has been stripped to its core in the Mancos Shale, which underlies the Mesaverde Group, hydraulically isolating its eastern and western flanks. The Sage Creek and Fish Creek anticlines are subparallel anticlines southeast of Hayden; both plunge northward. Of the two, the Sage Creek anticline is larger, tighter, and has more vertical relief. The eastern flanks of all three anticlines are much steeper than the western flanks. Outcrops on the eastern flanks commonly dip 30 to 60°; those on the western flanks commonly dip 10 to 20°. The steep-ended flanks resulted from compressive stresses produced by the north-south trending Park Range as it formed east of the study area.

Faults are more common east of Dry Creek. Although Bass and others (1955) mapped several surficial fault traces on the western flank of the Tow Creek anticline and to the northeast and south of Twentymile Park, many more faults are known to exist in the subsurface. Difficulty in identifying fault offset and orientation from lithologic or geophysical logs precluded most additional mapping. Numerous northwest-trending faults located south of Twentymile Park exhibit vertical offset of less than 100 ft, as measured in the dip slope south of Foidel Creek. Some of these offsets may result from strike-slip movement on the dip slope as indicated by slickenslides observed in coal mines in this area (Richard Tifft, Twentymile Coal Co., oral commun., 1985). Vertical offset ranges from 0 to 400 ft along the fault, or fault zone, located within the study area to the northeast of Twentymile Park. In addition to offset, faulting in this area has created an extensively fractured zone of rock within or between several fault planes that parallel the fault trace shown on plate 1.

Structural warping and faulting in the eastern part of the study area is indicated by the configuration and lateral extent of the bedrock formations. The top of the Trout Creek Sandstone Member has 3,200 ft of structural relief, between the trough of the Twentymile Park syncline and the southern outcrops of the formation (fig. 15). The basin underlying Twentymile Park contains two structural lows, one on the Twentymile syncline, the other at the southern end of the Tow Creek anticline. The combination of structure and topography produces an irregular, contorted outcrop line that delineates the limit of the water-yielding units considered in this study. The deformed and faulted structure of the Trout Creek Sandstone Member in the Iles Formation is expressed in the structure map of the base of the Twentymile Sandstone Member in the Williams Fork Formation (fig. 16). Structural relief on this surface exceeds 1,700 ft. The two structural low areas in the Trout Creek Sandstone Member also are evident in the structure of the base of the Lewis Shale (fig. 17). Maximum structural relief on the Lewis Shale is about 1,100 ft.

CLIMATIC CONDITIONS

All surface water and ground water in the study area is the result of precipitation. Changes in climatic conditions such as precipitation, temperature, wind, and evaporation can cause large and rapid changes in streamflow and more gradual changes in ground-water flow. The changes in

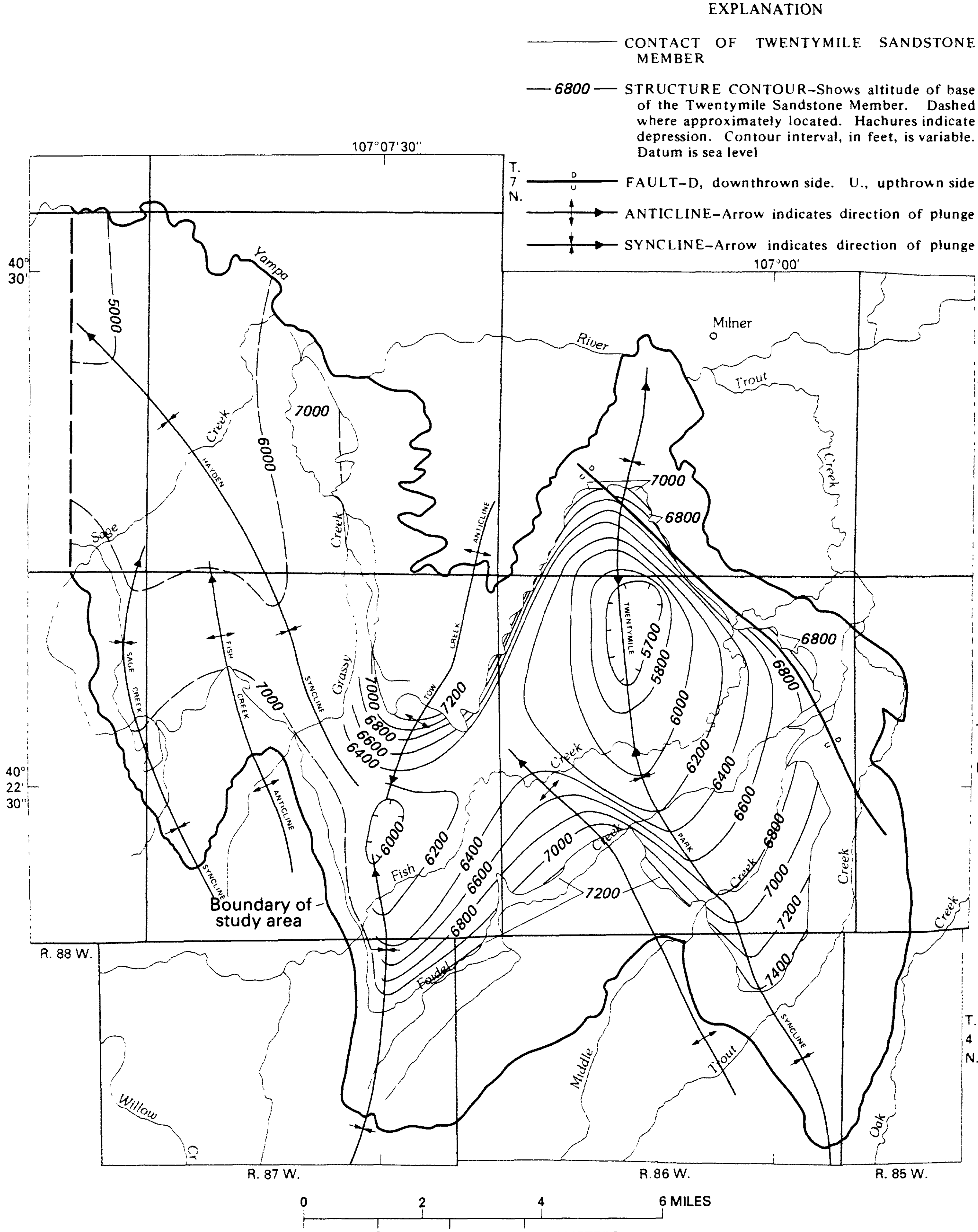


Figure 16.--Structural altitude of the base of the Twentymile Sandstone Member of the Williams Fork Formation in the eastern part of the study area.

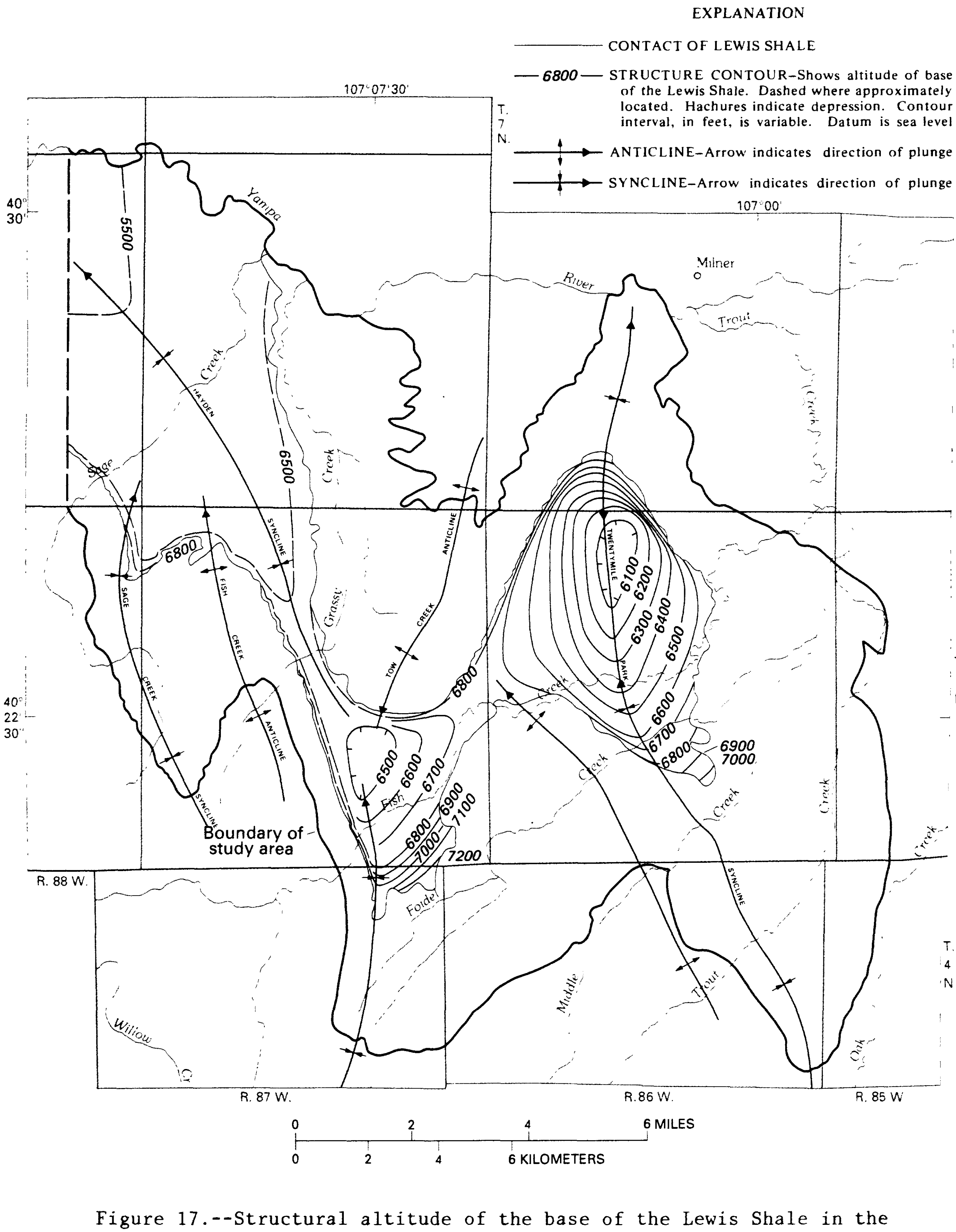


Figure 17.--Structural altitude of the base of the Lewis Shale in the eastern part of the study area.

ground-water flow primarily occur through changes in ground-water recharge. Climatic conditions affect ground-water recharge by means of changes in precipitation, evapotranspiration, vegetation, weathering, and landform and soil development. Principal climatic factors include precipitation, temperature, wind, and evaporation.

Precipitation

Precipitation on the western slope of the Rocky Mountains primarily is controlled by adiabatic cooling of eastward-tracking Pacific storm systems. As the systems gain altitude in crossing the mountains, the air cools and loses part of its moisture as rain and snow on the western slope and Continental Divide. Precipitation in the study area thus is correlated with altitude. Mean annual precipitation ranges from 13.8 in. at Craig (altitude 6,190 ft) to more than 46 in. near the crest of Quarry Mountain (altitude 8,200 ft) southwest of Steamboat Springs. The relations between precipitation and altitude (fig. 18) are based on data from 9 U.S. Weather Bureau (National Oceanic and Atmospheric Administration, 1890-1987) gages and 20 U.S. Geological Survey or privately operated gages (fig. 19; table 1). Periods of record ranged from 2 years to more than 90 years at Craig and Steamboat Springs. Monthly precipitation data were used to regress the shorter record stations in the eastern part of the study area against the longer record stations to better estimate the 90-year mean annual precipitation (table 1) in this area. Regressions of the 78-year mean annual precipitation at Hayden, Yampa, and Pyramid were not done because the mean for the 78-year period was not significantly different from the mean for the 90-year period. Stations located in the Williams Fork and Willow Creek drainage areas are outside the study area and did not correlate well with the distant longer record stations. As a result, the 18- to 50-year periods of record for the Williams Fork and Willow Creek stations were only used to estimate mean annual precipitation to the southwest of the study area.

The relations between precipitation and altitude for the drainage areas of the Williams Fork, Willow Creek, Grassy Creek, Trout Creek, Fish Creek, and Foidel Creek generally are similar, indicating that precipitation increases moderately with altitude in the southern and eastern parts of the study area. A much more rapid increase in precipitation with altitude occurs along the valley of the Yampa River west of Steamboat Springs. However, along the upper valley of the Yampa River southwest of Steamboat Springs, mean annual precipitation decreases with altitude. These marked differences in the precipitation patterns result from the complex interaction of storm movement and topography. Precipitation increases when topographic features such as the Williams Fork Mountains and the Yampa River valley enhance up-valley movement of storms. Cross-valley movement of storms may produce a rain shadow effect on the leeward slopes such as Twentymile Park, the upper reaches of the Yampa River, and Oak Creek. The resulting relations between precipitation and altitude range in slopes from 0.016 to -0.014 inches of precipitation per foot of altitude depending on the configuration and orientation of topography with respect to principal storm tracks.

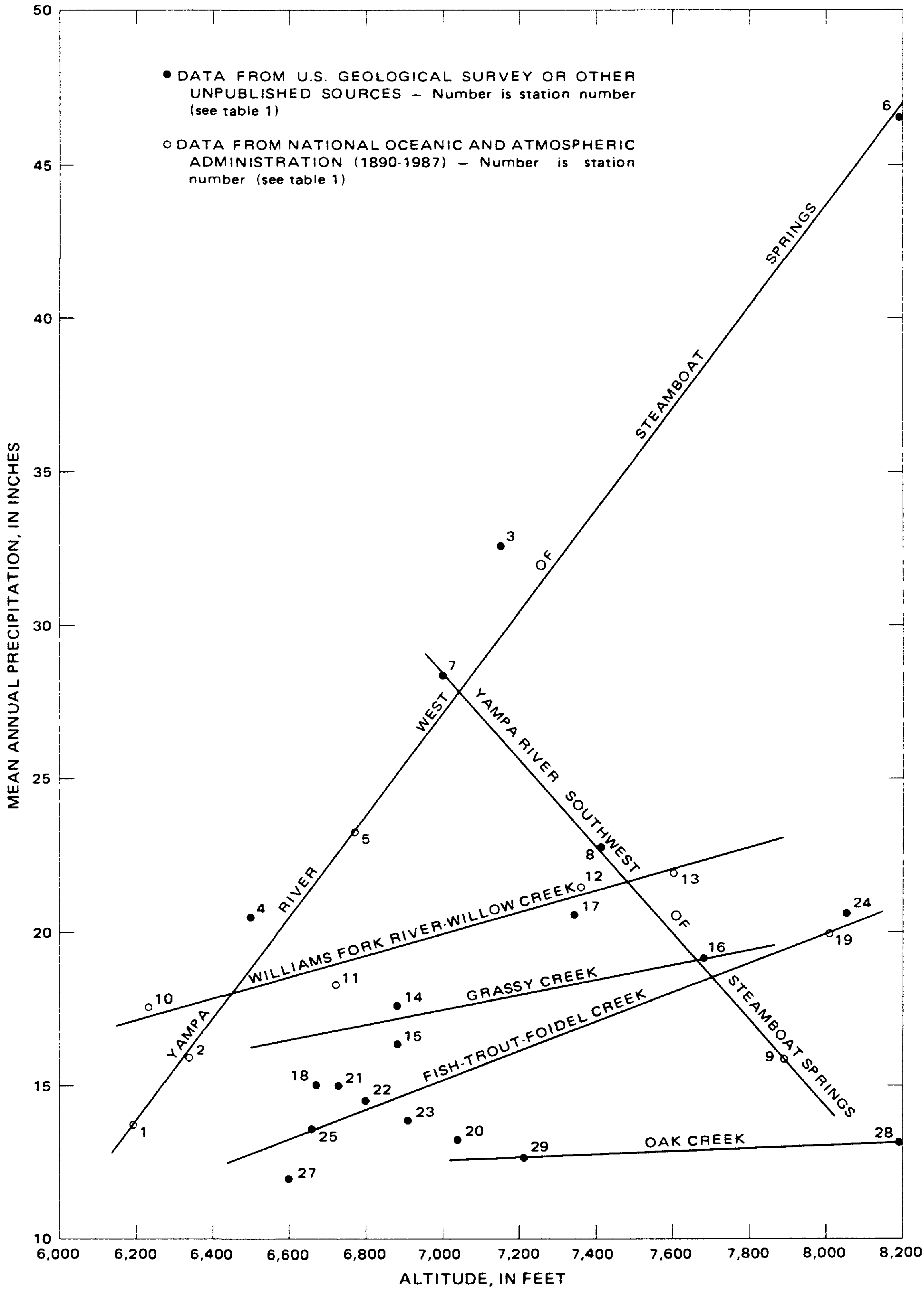


Figure 18.--Relation between mean annual precipitation and altitude.

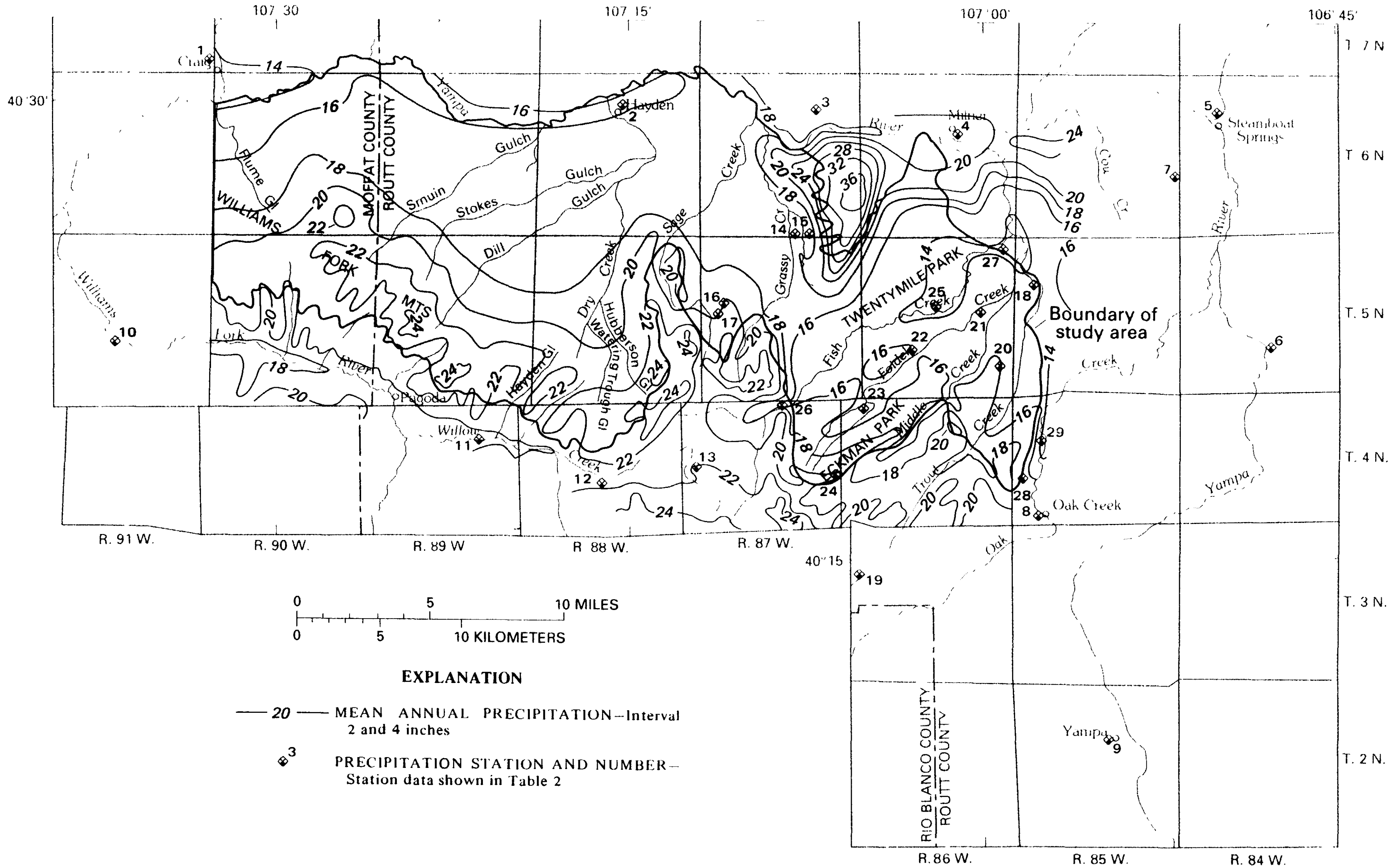


Figure 19.--Isohyetal map of the study area.

Table 1.--*Precipitation station index*

[NR, no regression]

Station number (figs. 18, 19)	Station name	Drainage area	Period of record	Regression station[1]	Regression correlation, R	Mean annual precipitation, (inches)
1	Craig	Yampa River	1894-1986	NR	NR	13.8
2	Hayden	(west of	1909-1986	NR	NR	15.9
3	Mt. Harris	Steamboat	1964-1966	5	([2])	32.6
4	Milner	Springs)	1964-1966	5	([2])	20.5
5	Steamboat Springs	|	1891-1986	NR	NR	23.3
6	Emerald	|	1964-1966	5	([2])	46.6
7	Catamount Lake	Yampa River (southwest	1983-1985	5	0.84	28.4
8	Oak Creek	of Steamboat	1964-1966	5	([2])	22.8
9	Yampa	Springs)	1909-1986	NR	NR	15.9
10	Hamilton	Williams Fork River	1936-1986	NR	NR	17.6
11	Pagoda	|	1890-1912	NR	NR	18.3
12	Willow Creek	Willow	1930-1948	NR	NR	21.5
13	Dunckley	Creek	1905-1909	11	0.66	22.0
14	Seneca M	Grassy Creek	1981-1985	2	0.84	17.6
15	Seneca L	|	1978-1983	2	0.89	16.4
16	Y-6	|	1980-1983	2	0.80	19.2
17	Y-1	Sage Creek	1980-1983	2	0.83	20.6
18	A	Trout Creek	1983-1985	2,5,9	0.84	15.0
19	Pyramid	|	1910-1986	NR	NR	20.0
20	Green	|	1980-1983	2,5,9	0.85	13.2
21	Lower Foidel	Foidel	1975-1981	2,5,9	0.84	15.0
22	2005	Creek	1982-1985	2,5,9	0.78	14.5
23	2001	|	1982-1985	2,5,9	0.71	13.9
24	Upper Foidel	|	1975-1981	2,5,9	0.78	20.6
25	1002	Fish Creek	1982-1983	2,5,9	0.78	13.6
26	[3]1001	|	1982-1983	2,5,9	0.80	12.6
27	Fish	|	1980-1981	2,5,9	0.90	12.0
28	Skyline	Oak Creek	1980-1985	2,5,9	0.91	13.2
29	Oak	|	1980-1982	2,5,9	0.83	12.6

[1]Mean of three monthly values was used for regression of three stations.
[2]Snow-course data, monthly correlation unavailable.
[3]Data not usable for figure 18.

The isohyetal map (fig. 19) for the area shows the distribution of mean annual precipitation. The map was developed using mean annual precipitation data and relations between precipitation and altitude shown in figure 18. Mean annual precipitation ranges from more than 36 in/yr on the crest of Mount Harris to less than 14 in/yr in Twentymile Park and in the Yampa Valley near Craig. Precipitation along the crest of the Williams Fork Mountains is estimated to range from 20 to 24 in/yr.

The mean monthly precipitation pattern varies from east to west across the study area. The mean monthly pattern for Steamboat Springs is characteristic of conditions in much of the western United States--greater precipitation in the winter, lesser precipitation in the summer. Precipitation patterns at Craig and Hayden are more characteristic of conditions in the study area; precipitation averages about 1 in/mo throughout the year (fig. 20). Orographic effects are pronounced at Steamboat Springs, producing greater winter snowfall than at Craig or Hayden.

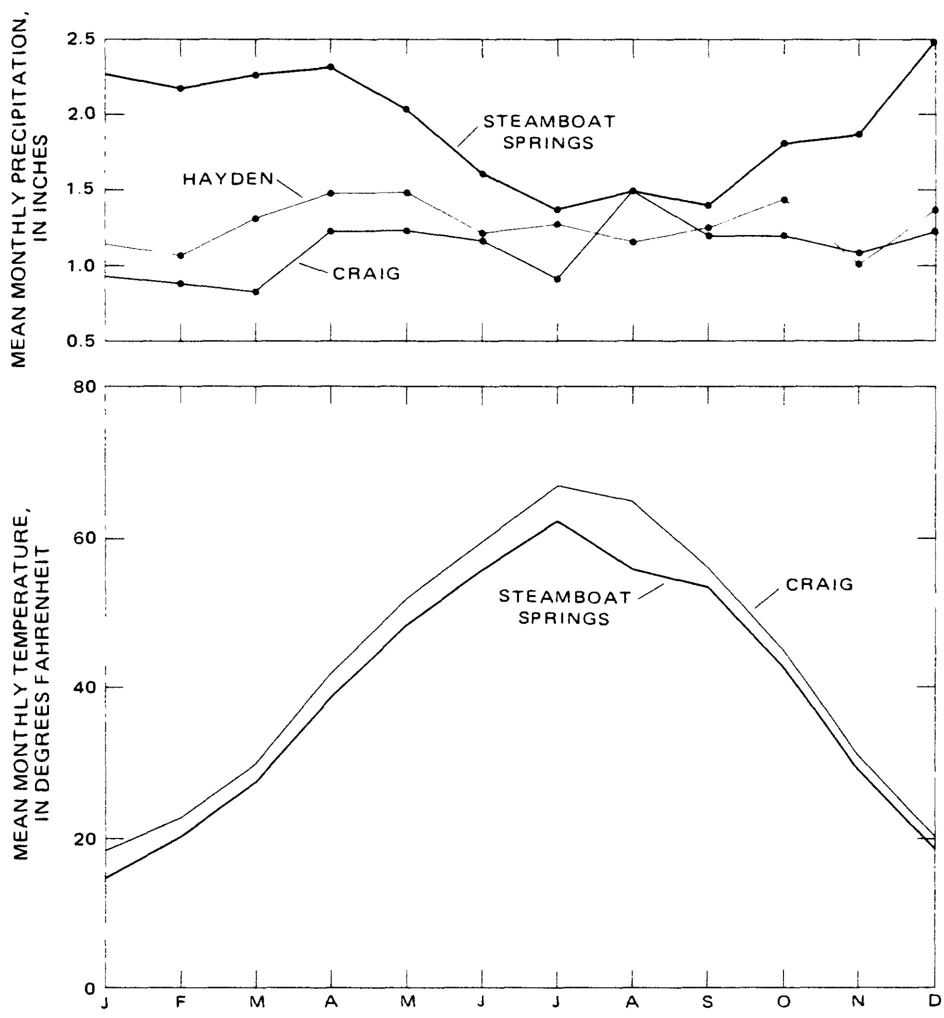


Figure 20.--Mean monthly precipitation and temperature distributions near the study area.

Temperature

Mean temperatures at Steamboat Springs and Craig have a strong seasonal correlation (fig. 20). Both curves have the same general shape but differ by 4 to 6 °F. This correlation indicates that factors that control temperature are more uniform in the area than factors that control precipitation.

The normally dry, cloudless conditions that occur at this altitude produce extreme seasonal and diurnal temperature fluctuations. Mean maximum daily temperatures in July range from 80 °F in Steamboat Springs to 85 °F in Craig. Mean minimum daily temperatures in January throughout the area are approximately -2 °F. Diurnal temperatures may fluctuate throughout a range of 40 °F or more at any time of the year.

Evaporation

Evaporation data for the study area are more limited than temperature or precipitation data. Only seven evaporation sites are maintained in the Colorado River watershed of western Colorado (table 2). Evaporation primarily is a function of available heat, solar insolation, humidity, and wind. At Hayden, the low humidity, intermittent winds, and small number of cloudy days result in a pan evaporation rate from May to October of about 42 in. (table 2). Using a pan coefficient of 0.7 (Kohler and others, 1959), lake evaporation is estimated to be about 29 in., well in excess of mean annual precipitation in most of the study area. Because information at Hayden is available only for May to October, annual evaporation actually is larger. This results in a precipitation-evaporation deficit, which greatly decreases the volume of water available to recharge the aquifers. No wind and humidity data are available for the study area. In general, the relative humidity is low, increasing only during thundershowers and snowstorms. Actual wind effects are unknown.

SURFACE-WATER HYDROLOGY

Surface-water-hydrology data are important to ground-water studies because knowledge of streamflow distribution and timing provides information about when and where recharge or discharge to streams may occur. Surface-water-chemistry data also provide information about ongoing surficial geochemical processes and about the chemical composition of discharging ground water.

Drainage Systems and Streamflow

Drainage systems and streamflow are affected by the origin and geographic location of the stream. The Yampa River and the Williams Fork are the two major streams that drain the study area. These streams are perennial throughout the area and have a mean annual flow of 1,100 ft^3/s (Yampa River at Hayden) and 44 ft^3/s (Williams Fork at Pagoda). The streams are located near the northern and southern periphery of the area and flow nearly due west across existing structural trends; both streams probably are antecedent and

Table 2.--*Pan-evaporation data from Colorado River basin sites*

[Values in inches except where noted; --, no data]

Site number (fig. 1)	Site name	Altitude (feet)	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	May-Oct. total	Source
1	Grand Junction	4,760	--	--	--	7.33	9.51	11.60	11.89	10.19	7.49	4.65	2.2	--	55.33	(1)
2	Grand Lake	8,288	--	--	--	--	6.18	8.14	8.34	7.14	5.44	3.63	--	--	38.87	(1)
3	Green Mountain Dam	7,740	--	--	--	--	5.71	6.48	6.95	6.12	4.60	3.23	--	--	33.09	(1)
4	Meredith	7,825	--	--	--	--	7.79	8.47	9.16	7.51	5.63	3.56	--	--	42.12	(1)
5	Montrose	5,785	1.53	1.37	3.13	5.45	7.36	9.07	8.87	7.61	5.57	3.42	1.74	1.52	41.90	(1)
6	Hayden	6,346	--	--	--	--	5.8	7.1	8.8	8.0	6.1	6.1	--	--	41.90	(2)
7	Vallecito Dam	5,010	--	--	2.33	3.93	5.41	6.54	6.45	5.51	3.97	3.06	1.80	--	30.94	(1)
														Mean	40.59	
														Standard deviation	7.9	

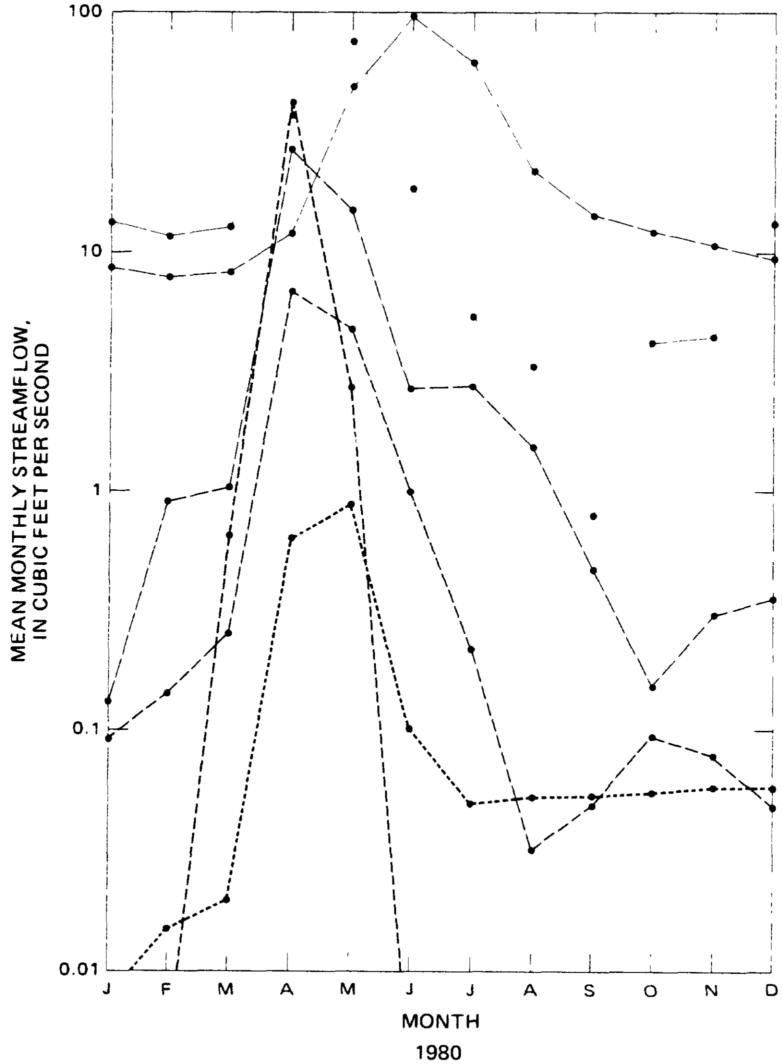
[1]Data from National Oceanic and Atmospheric Administration (1890-1987).
[2]Data from U.S. Department of the Interior (1976).

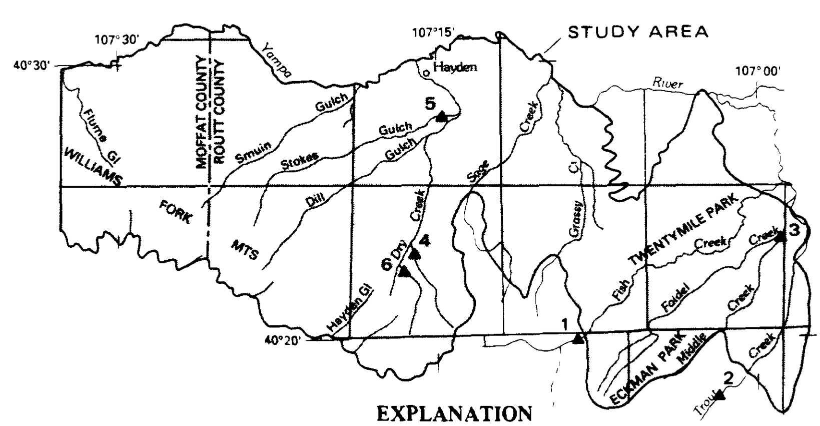
superposed (Hunt, 1969). Most of the smaller tributary streams follow structural trends, although some streams flow across the structure (pl. 1). Three tributary stream systems--western, central, and eastern--are unique because of differing structural settings.

The western system, which extends from the western study boundary to Hayden Gulch, is the simplest system. This area is drained by gulches that have formed on cuestas of the Mesaverde Group outcrop. Gullies are aligned subparallel to each other down the front and back of the cuesta. All western gulches begin along the cuesta ridges at altitudes less than 7,500 ft. Snowmelt runoff occurs only during the spring, generally rising, peaking, and receding within a few months (fig. 21, Stokes Gulch). This streamflow generally is small and occurs from March to July. Northward-draining gulches flow later in the year than do southward-draining gulches because of larger drainage areas, smaller gradients, and a northward aspect that delays snowmelt runoff. Gulches in the western area may provide recharge to the ground-water system only during the spring because they generally are dry by summer. Conversely, springs, seeps, and intermittent perennial base flow are evidence of ground-water discharge to some reaches of the gulches.

The central stream system drains the Sage Creek and Fish Creek anticline areas and the western side of the Tow Creek anticline (pl. 1). Within the area are three perennial streams--Dry Creek, Sage Creek, and Grassy Creek--in addition to numerous intermittent gulches. All streams originate in or near the study area, generally at altitudes less than 8,000 ft. Dry Creek is a subsequent stream draining the western flank of the Sage Creek anticline. Sage Creek drains the central and eastern parts of the Sage Creek anticline and flows across structural trends. Little streamflow data are available for these two streams. Two tributaries, Hubberson Gulch and Watering Trough Gulch, have 3 to 6 years of streamflow records that indicate ephemeral flow conditions. Grassy Creek drains most of the Fish Creek anticline and the western half of the Tow Creek anticline. Gain-loss measurements in Grassy Creek indicate that the upper reach of the creek gains flow from the outcrops of the Trout Creek Sandstone Member of the Iles Formation and from the Williams Fork Formation (fig. 22; table 3). Downstream from Grassy Creek station 4, which is at Routt County Road 29, the creek generally gains flow during the spring through summer months. The dryer climatic conditions during late summer and fall cause water levels in the alluvial aquifer and bedrock formations to decline, and this reach of Grassy Creek may lose flow during this time.

Streams in the eastern area differ from other streams in the study area. The eastern area is drained by four main streams--Fish Creek, Foidel Creek, Middle Creek, and Trout Creek--that converge south of Milner. Fish Creek, the northernmost stream, drains Dunckley Park, flows across structural trends in Fish Creek Canyon, and drains much of Twentymile Park. Fish Creek headwaters are above 10,000 ft on the northern side of the Dunckley Flat Tops. Mean annual flow at the gage in Fish Creek Canyon is about 13 ft^3/s. Partial records from four downstream gages indicate that Fish Creek is perennial, although base flow decreases downstream. Foidel Creek begins near Eckman Park at an altitude of about 7,600 ft and is the only creek in the eastern stream system that originates in the study area. The relatively small drainage area of Foidel Creek includes the southern part of Twentymile Park. Mean annual

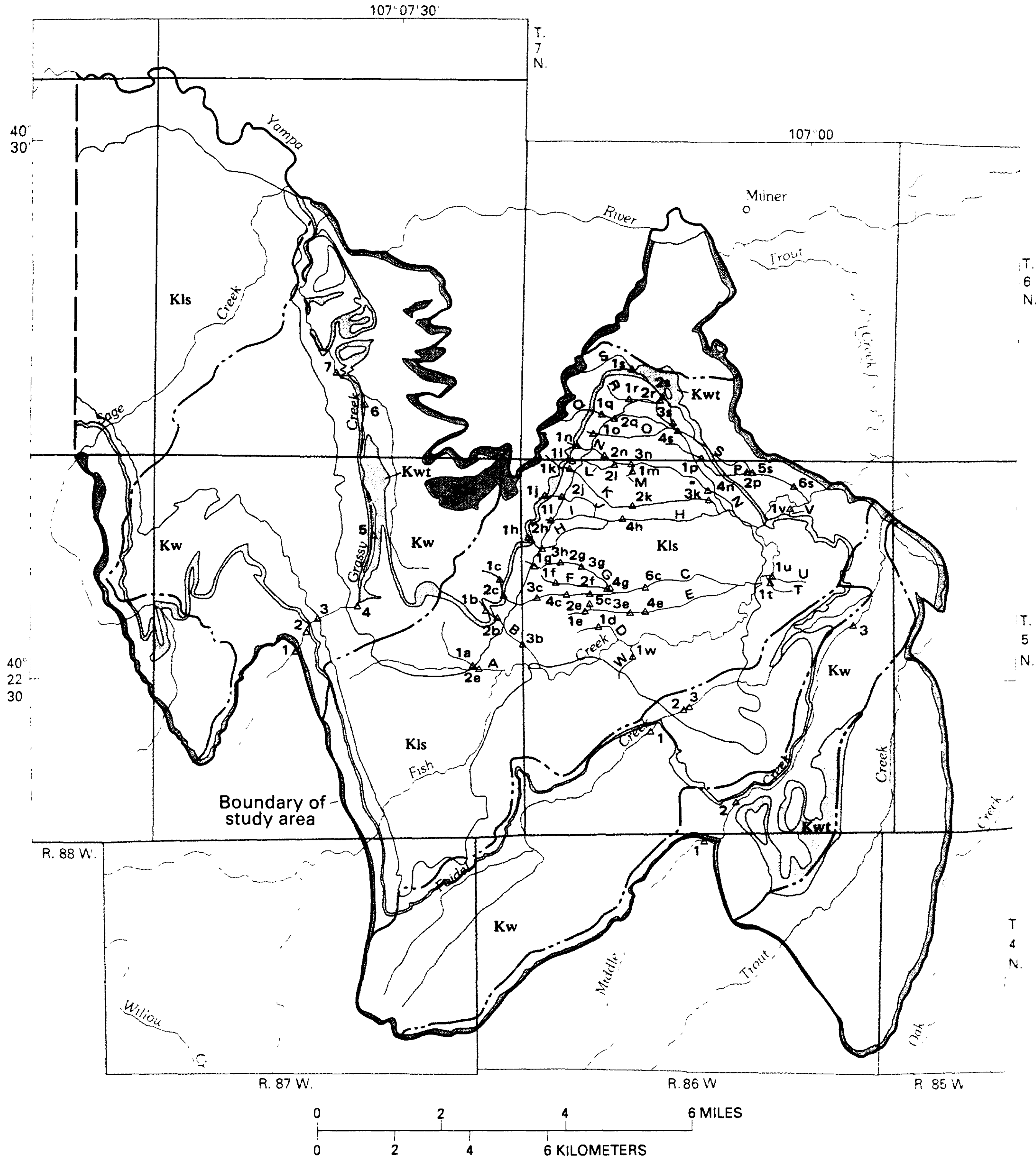




EXPLANATION

1 ———— FISH CREEK	5 -------- STOKES GULCH
2 ———— TROUT CREEK	6 ··········· WATERING TROUGH GULCH
3 ———— FOIDEL CREEK	▲4 STREAM-GAGING SITE AND NUMBER
4 ------- HUBBERSON GULCH	

Figure 21.--Hydrographs of representative streams in the study area.



EXPLANATION

OUTCROP OF BEDROCK GEOLOGIC UNITS

Kls — Lewis Shale

Kw — Williams Fork Formation

Kwt — Twentymile Sandstone Member of the Williams Fork Formation

Kis — Trout Creek Sandstone Member of the Iles Formation

DRAINAGE BOUNDARY OF PRINCIPAL STREAM

1 — PRINCIPAL STREAM—Number indicates streamflow measurement station listed in table 3

A 1a — TRIBUTARY STREAM—Capital letter is stream designation, number with lower case letter indicates streamflow measurement station listed in table 3

Figure 22.--Surface-water drainages and outcrops of bedrock geologic units in the eastern part of the study area.

Table 3.--*Hydrologic data from gain-loss measurements in streams near Twentymile Park*

[μS/cm, microsiemens per centimeter; °C, degrees Celsius; ft^3/s, cubic feet per second; CP, current-meter measurement of poor quality; F, 3-inch cutthroat flume; CG, current-meter measurement of good quality; V, timed volume measurement; --, no data; NA, not applicable]

Station (fig. 22)	Date	Flow-measuring device	Specific conductance (μS/cm)	Temperature (°C)	Flow (ft^3/s)	Change in flow (ft^3/s)	Remarks
				Grassy Creek			
1	06-25-86	CP	--	--	1.4	NA	
2	|	CP	--	--	1.4	0.0	
3	|	CP	--	--	1.4	0.0	
7	|	CP	--	--	4.4	+3.0	
1	07-22-86	F	950	12.5	0.42	NA	Above outcrop of Trout Creek Sandstone Member of Iles Formation.
3	|	F	950	12.5	0.54	+0.12	Below outcrop of Twentymile Sandstone Member of Williams Fork Formation.
4	|	F	--	--	0.55	+0.01	
5	|	F	900	15.0	0.66	+0.11	
6	|	F	850	18.5	0.64	-0.02	
1	09-15-86	F	1,000	13.0	0.25	NA	
3	|	F	950	10.0	0.30	+0.05	
4	|	F	1,000	12.5	0.24	-0.06	
5	|	F	900	14.5	0.22	-0.02	
6	|	F	850	17.5	0.25	+0.03	
1	|	F	1,000	14.5	0.24	NA	

Table 3.--*Hydrologic data from gain-loss measurements in streams near Twentymile Park*--Continued

Station (fig. 22)	Date	Flow-measuring device	Specific conductance (μS/cm)	Temperature (°C)	Flow (ft^3/s)	Change in flow (ft^3/s)	Remarks
				Foidel Creek			
1	06-26-86	CP	--	--	5	NA	
2		CG	--	--	4.85	NA	Change in flow not measurable.
1	09-17-86	F	2,600	16.0	0.63	NA	
3		F	2,600	15.5	0.71	+0.08	
				Middle Creek			
1	09-17-86	F	650	13.5	0.52	NA	Above Trout Creek Sandstone Member outcrop.
2		F	800	16.5	0.52	0.0	Below Trout Creek Sandstone Member outcrop.
3		F	850	16.0	0.54	+0.02	Below Twentymile Sandstone Member outcrop.
				Fish Creek unnamed tributaries--A and B			
(A) 1a	08-14-86	NA	--	--	Seep	NA	
2a		V	1,600	18.5	0.064	+0.064	
1a	09-15-86	NA	--	--	Seep	NA	
2a		V	--	--	0.062	+0.062	
(B) 1b	07-23-86	V	--	--	0.004	NA	Above Twentymile Sandstone Member outcrop.
2b		V	1,000	18.0	0.015	+0.011	Below Twentymile Sandstone Member outcrop.
3b		NA	--	--	0.0	-0.015	Stock pond intercepts flow.

Table 3.--*Hydrologic data from gain-loss measurements in streams near Twentymile Park*--Continued

Station (fig. 22)	Date	Flow-measuring device	Specific conductance (μS/cm)	Temperature (°C)	Flow (ft^3/s)	Change in flow (ft^3/s)	Remarks
			Fish Creek unnamed tributaries--A and B--Continued				
(B) 1b	09-18-86	NA	--	--	Seep	NA	
2b	|	V	--	--	0.005	+0.005	
3b	|	NA	--	--	0.0	-0.005	
			Fish Creek unnamed tributaries--C, D, and E				
(C) 1c	07-22-86	F	2,220	18.0	--	NA	Flume affected by submergence.
2c	|	F	2,200	15.0	0.10	NA	In Twentymile Sandstone Member outcrop.
3c	|	F	2,200	19.0	0.11	+0.01	Below Twentymile Sandstone Member outcrop.
4c	07-21-86	F	--	--	0.11	0.0	
5c	|	F	--	--	0.14	+0.03	
6c	|	F	--	--	0.12	-0.02	
3c	08-13-86	F	2,400	21.0	0.08	NA	
1c	09-17-86	F	2,200	14.0	0.06	NA	
2c	|	F	2,400	12.0	0.06	0.0	
3c	|	F	2,400	14.5	0.06	0.0	
4c	|	F	2,400	14.0	0.06	0.0	
5c	|	F	2,400	16.0	0.05	-0.01	
6c	|	F	2,800	13.0	0.04	-0.01	
(D) 1d	07-23-86	V	3,000	25.0	0.004	NA	
1d	08-14-86	NA	--	--	0.0	NA	
1d	09-17-86	NA	--	--	0.0	NA	

Table 3.--*Hydrologic data from gain-loss measurements in streams near Twentymile Park*--Continued

Station (fig. 22)	Date	Flow-measuring device	Specific conductance (μS/cm)	Temperature (°C)	Flow (ft^3/s)	Change in flow (ft^3/s)	Remarks
			Fish Creek unnamed tributaries--C, D, and E--Continued				
(E) 1e	07-21-86	NA	--	--	0.0	NA	
2e	|	V	900	16.0	0.006	NA	Discharge from flowing well tributary to creek E.
3e	07-23-86	V	8,000	27.5	0.010	+0.004	
4e	07-21-86	NA	--	--	Seep	-0.004	
1e	09-17-86	NA	--	--	0.0	NA	
2e	|	V	1,000	15.5	0.006	NA	Discharge from flowing well tributary to creek E.
3e	|	NA	--	--	0.0	-0.006	
4e	|	NA	--	--	0.0	NA	
			Fish Creek unnamed tributaries--F, G, H, and I				
(F) 1f	07-22-86	NA	--	--	0.0	NA	
2f	07-21-86	NA	--	--	0.0	NA	
1f	09-17-86	NA	--	--	0.0	NA	
(G) 2g	07-22-86	NA	--	--	0.0	NA	
3g	|	NA	--	--	0.0	NA	
4g	07-21-86	NA	--	--	0.0	NA	
1g	09-17-86	V	1,000	10.0	0.002	NA	Above stock pond.
2g	|	NA	--	--	0.0	-0.002	

Table 3.--*Hydrologic data from gain-loss measurements in streams near Twentymile Park*--Continued

Station (fig. 22)	Date	Flow-measuring device	Specific conductance (μS/cm)	Temperature (°C)	Flow (ft^3/s)	Change in flow (ft^3/s)	Remarks
			Fish Creek unnamed tributaries--F, G, H, and I--Continued				
(H) 3h	07-22-86	V	--	--	0.038	NA	Below Twentymile Sandstone Member outcrop.
4h	|	F	950	24.0	0.06	+0.02	
3h	08-14-86	V	1,500	21.0	0.011	NA	
4h	08-13-86	F	850	23.0	0.04	+0.03	
1h	09-17-86	V	850	8.0	0.010	NA	Above Twentymile Sandstone Member outcrop.
2h	|	V	2,400	10.0	0.004	NA	Spring discharging from Twentymile Sandstone Member.
3h	|	V	1,500	7.0	0.021	+0.011	Below Twentymile Sandstone Member outcrop.
4h	09-16-86	V	1,000	20.0	0.010	-0.011	
(I) 1i	07-22-86	NA	--	--	0.0	NA	
			Fish Creek unnamed tributaries--J, K, L, and M				
(J) 1j	07-22-86	NA	--	--	0.0	NA	Above Twentymile Sandstone Member outcrop.
2j	|	V	750	17.0	0.045	+0.045	Below Twentymile Sandstone Member outcrop.
2j	08-14-86	V	750	16.5	0.015	NA	
1j	09-17-86	NA	--	--	0.0	NA	
2j	|	V	720	7.0	0.010	+0.010	

Table 3.--*Hydrologic data from gain-loss measurements in streams near Twentymile Park*--Continued

Station (fig. 22)	Date	Flow-measuring device	Specific conductance (µS/cm)	Temperature (°C)	Flow (ft^3/s)	Change in flow (ft^3/s)	Remarks
			Fish Creek unnamed tributaries--J, K, L, and M--Continued				
(K) 2k	07-22-86	NA	--	--	Seep	NA	
3k	|	NA	--	--	0.0	NA	
2k	08-13-86	NA	--	--	0.0	NA	
1k	09-17-86	NA	--	--	0.0	NA	Below Twentymile Sandstone Member outcrop.
2k	09-16-86	NA	--	--	0.0	NA	
3k	|	NA	--	--	0.0	NA	
(L) 2l	07-22-86	V	--	--	0.008	NA	
2l	08-13-86	V	--	--	0.003	NA	
2l	09-16-86	NA	--	--	Seep	NA	
1l	09-17-86	NA	--	--	0.0	NA	
(M) 1m	07-22-86	V	500	27.0	0.018	NA	
1m	08-13-86	V	440	23.0	0.004	NA	
1m	09-16-86	V	450	18.0	0.004	NA	

Table 3.--*Hydrologic data from gain-loss measurements in streams near Twentymile Park*--Continued

Station (fig. 22)	Date	Flow-measuring device	Specific conductance (μS/cm)	Temperature (°C)	Flow (ft^3/s)	Change in flow (ft^3/s)	Remarks
			<u>Fish Creek unnamed tributaries--N, O, P, Q, and R</u>				
(N) 2n	07-22-86	NA	--	--	Seep	NA	
3n	|	V	1,700	24.0	0.004	NA	Below stock pond.
4n	|	NA	--	--	0.0	-0.004	Unmeasured tributary inflow occurs between stations 3n and 4n.
2n	08-13-86	NA	--	--	0.0	NA	
3n	|	NA	--	--	0.0	NA	
2n	09-16-86	NA	--	--	0.0	NA	
3n	|	NA	--	--	0.0	NA	
4n	|	NA	--	--	0.0	NA	
1n	09-17-86	NA	--	--	0.0	NA	
(O) 1o	09-17-86	NA	--	--	0.0	NA	
(P) 1p	08-13-86	NA	--	--	0.0	NA	
2p	|	NA	--	--	0.0	NA	
2p	09-16-86	NA	--	--	0.0	NA	
(Q) 1q	08-14-86	NA	--	--	--	NA	Stock pond discharge is by seepage.
2q	|	V	600	18.0	0.024	NA	Thunderstorm evening of 8-13-86.
2q	09-16-86	V	700	15.0	0.010	NA	

Table 3.--*Hydrologic data from gain-loss measurements in streams near Twentymile Park*--Continued

Station (fig. 22)	Date	Flow-measuring device	Specific conductance (μS/cm)	Temperature (°C)	Flow (ft^3/s)	Change in flow (ft^3/s)	Remarks
			Fish Creek unnamed tributaries--N, O, P, Q, and R--Continued				
(R) 1r	09-16-86	NA	--	--	0.0	NA	
2r	|	NA	--	--	0.0	NA	Springs discharge at R-S confluence.
			Fish Creek unnamed tributaries--S, T, U, V, and W				
(S) 1s	08-14-86	NA	--	--	Seep	NA	Thunderstorm evening of 8-13-86.
3s	|	F	650	23.0	0.20	+0.20	Thunderstorm evening of 8-13-86.
4s	|	F	--	--	0.21	+0.01	Tributary inflow occurs above station.
4s	08-13-86	F	580	21.0	0.06	NA	Prior to thunderstorm.
5s	08-12-86	F	660	15.0	0.08	+0.02	
6s	08-13-86	F	750	17.0	0.08	0.0	
1s	09-16-86	NA	--	--	0.0	NA	Beaver ponds between stations 1S and 2S.
2s	|	V	750	12.0	0.021	+0.021	Above springs at R-S confluence.
3s	|	F	650	14.0	0.04	+0.02	
4s	|	F	650	12.0	0.05	+0.01	Tributary inflow occurs above station.
5s	|	F	650	17.0	0.03	-0.02	
6s	|	F	850	17.0	0.04	+0.01	

Table 3.--*Hydrologic data from gain-loss measurements in streams near Twentymile Park*--Continued

Station (fig. 22)	Date	Flow-measuring device	Specific conductance (µS/cm)	Temperature (°C)	Flow (ft^3/s)	Change in flow (ft^3/s)	Remarks
Fish Creek unnamed tributaries--S, T, U, V, and W--Continued							
(T) 1t	08-12-86	NA	--	--	0.0	NA	
1t	09-18-86	NA	--	--	0.0	NA	
(U) 1u	08-12-86	V	--	--	0.006	NA	
1u	09-18-86	NA	--	--	0.0	NA	
(V) 1v	08-12-86	V	700	19.0	0.001	NA	
1v	09-18-86	NA	--	--	Seep	NA	
(W) 1w	08-14-86	V	1,450	19.5	0.003	NA	
1w	09-18-86	V	1,600	10.0	0.005	NA	

flow is 2.7 ft^3/s near the mouth of Foidel Creek. Base flow increases in the downstream reaches where the creek is perennial except during unusually dry years. An increase in streamflow of 0.08 ft^3/s was measured across the outcrop of Twentymile Sandstone Member (fig. 22, stations 1 and 3) on September 17, 1986 (table 3). Middle Creek flows into Foidel Creek in the eastern part of Twentymile Park. Middle Creek is similar to Foidel Creek but drains a larger area; its headwaters are above 8,400 ft in altitude. Runoff from Middle Creek peaks in the late spring and early summer. Mean annual flow is 4.4 ft^3/s near the mouth of Middle Creek. Streamflow measurements in Middle Creek on September 17, 1986, indicate minimal change in flow in a 5-mi reach of the creek between the outcrop of the Trout Creek Sandstone Member (station 1) and the downstream outcrop of the Twentymile Sandstone Member (fig. 22, station 3). Trout Creek is the largest stream draining the study area. From its headwaters at an altitude of 11,000 ft, it has a perennial base flow of 10 to 20 ft^3/s to its confluence with the Yampa River near Milner. Only the extreme southeastern part of the study area is drained by Trout Creek. Fish Creek and Middle Creek are confluent with Trout Creek near the eastern margin of the study area.

Streamflow gain-loss measurements made in numerous unnamed tributaries to Fish Creek (fig. 22) indicate that perennial flow occurs in some reaches of these streams. Most perennial flow is the result of ground-water discharge from the upstream outcrops of thick sandstone beds near the margins of the basin. Along the mountain front northwest of Twentymile Park water levels in the sandstones generally are above stream level. This is the result of recharge in the higher outcrops on either side of the stream valley. The resulting base flow may extend downstream beyond the mountain front onto the relatively impermeable strata of the Lewis Shale in Twentymile Park. In some streams, base flow may become tributary to Fish Creek, but, more commonly, the flow is lost to evapotranspiration along the channel or is captured in stock ponds. Most reaches of the tributary streams north of Fish Creek are ephemeral. In mid to late summer, the channels are dry or consist of alkali-encrusted desiccated mud or marsh. Although hydrostatic heads in the underlying bedrock aquifers may be 200 to 300 ft above land surface in parts of Twentymile Park, discharge from the aquifers to streamflow is not apparent except at a few uncontrolled flowing wells. Shale in the middle member of the Williams Fork Formation and in the Lewis Shale seems to form an effective confining layer that limits ground-water discharge to streamflow in Twentymile Park.

Springs are present throughout the study area and are an important source of surface water during low-flow periods. Discharge from most springs is diffuse and flows at a low rate. Springs are more prevalent in the western part of the study area, where they provide small quantities of water to intermittent streams in gulches.

Water Quality

Water quality in streams that drain the study area is affected markedly by the geologic materials within the drainage areas. Surface-water flow and water-quality data have been collected at 21 sites in or near the study area (Maura, 1982, 1985; Turk and Parker, 1982). Eighteen of these sites (fig. 23;

table 4) are on streams that drain stratigraphic intervals of (1) the lower Iles Formation and older rock units, (2) the Trout Creek Sandstone Member of the Iles Formation and the Williams Fork Formation, and (3) the Lewis Shale. Data from sites 1, 2, and 3 (table 4) represent runoff from geologic materials older than the Trout Creek Sandstone Member. These rocks generally are located at higher altitudes and outside of the study area. Water in streams that drain these geologic materials is a calcium bicarbonate type that generally has dissolved-solids concentrations ranging from 100 to 400 mg/L. This surface water is of better chemical quality than any other in the study area.

Data from sites 4 through 13 primarily represent runoff from the Trout Creek Sandstone Member and from the Williams Fork Formation. Rocks in this interval were deposited under a combination of marine, deltaic, and continental conditions. As a result, runoff is of a dissimilar chemical composition; generally, the water is either calcium magnesium bicarbonate or calcium magnesium sulfate. Three of the 10 sites in this group (sites 11, 12, and 13) are located downstream from large strip mines, and water quality may be affected by mine drainage. Dissolved-solids concentrations commonly range from 300 to 800 mg/L in streams unaffected by mining activities. At sites 11, 12, and 13, dissolved-solids concentration commonly range from 300 to 3,000 mg/L.

Streams that primarily drain the Lewis Shale or the shale units in the upper member of the Williams Fork Formation were sampled at sites 14 through 18. The marine sediments in these rock units markedly affect the surface-water chemistry. The streams in this group generally have a magnesium sodium sulfate water composition and dissolved-solids concentrations that commonly range from about 1,000 to about 8,000 mg/L.

GROUND-WATER HYDROLOGY

Lohman (1972) defines an aquifer as "...a formation...that contains sufficient saturated permeable material to yield significant quantities of water to wells and springs." "...Significant quantities of water..." in one region for one application may be insignificant in other regions or for other applications. The water-yielding units that are classified as aquifers in this study generally produce such small sustained yields (about 0-10 gal/min) that they would not be considered aquifers for many water-supply applications. However, these water-yielding units are the principal source of water in the local bedrock formations; they cause inflow to mines, and they supply usable volumes of water to the few stock or domestic wells in the area. Therefore, in this report, these water-yielding units are classified as aquifers.

Depositional Environments

Coal and associated deposits of the Iles and Williams Fork Formations developed in marine and deltaic plain environments located close to the shoreline (Weimer, 1976). Marine deposits of mudstone and shale generally are thick and homogeneous. These deposits have low permeability and are classified as regional confining beds. Near-shore marine deposits grade upward into massive transitional sandstones. These extensive sandstones

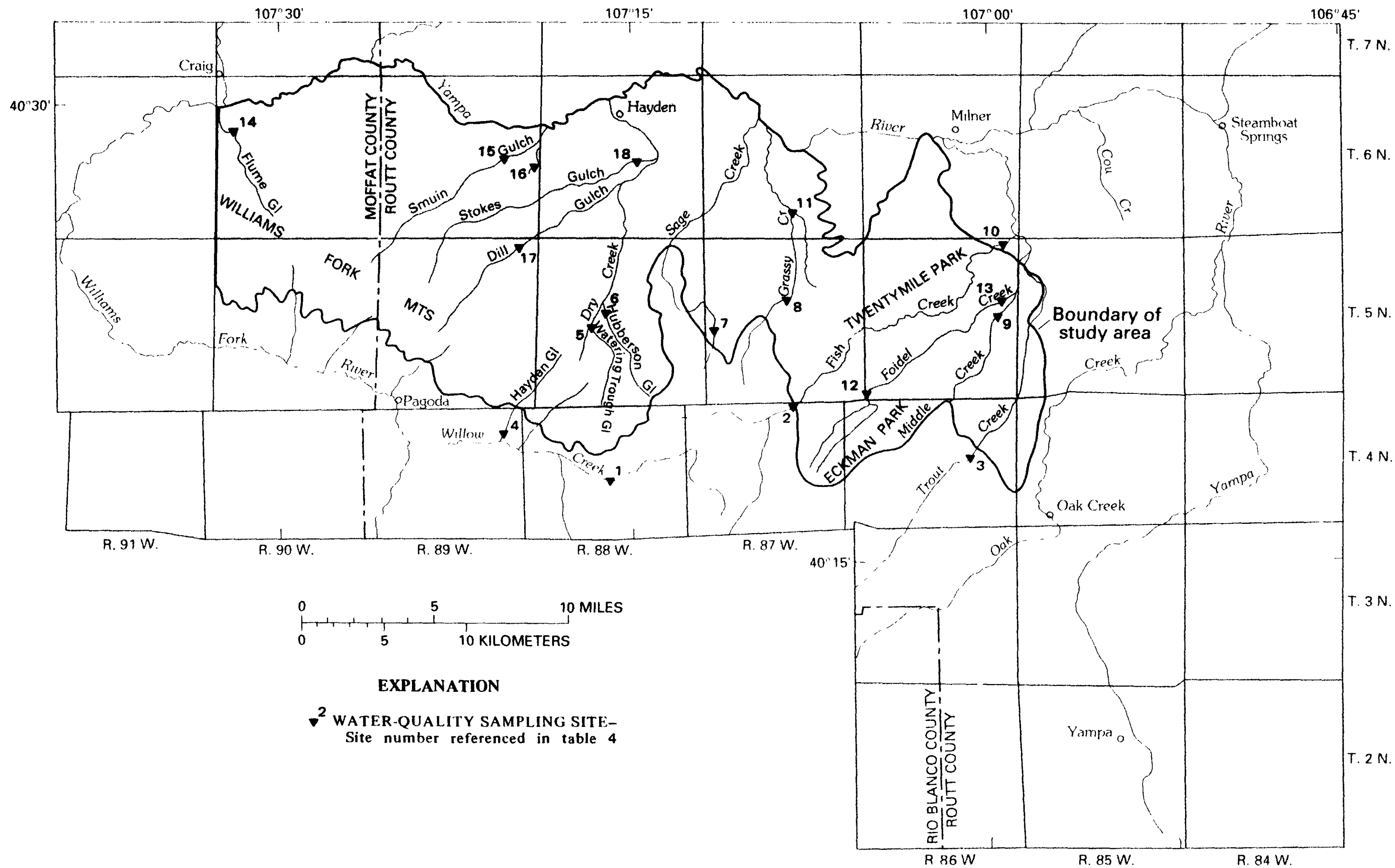


Figure 23.--Location of water-quality sampling sites at streams.

Table 4.--*Summary of surface-*

[μS/cm, microsiemens per centimeter; $\bar{x}$, mean; S, standard deviation;

Site number (fig. 24)	Stream	Station number	Number of analyses	Specific conductance (μS/cm)		pH (units)		Calcium	
				$\bar{x}$	S	$\bar{x}$	S	$\bar{x}$	S
1	Willow Creek near Dunckley	401747107161600	8	462	77	8.1	0.2	58	6
2	Fish Creek near Milner	09244100	8	513	153	8.2	0.2	55	15
3	Trout Creek near Oak Creek	401816107011000	7	184	61	8.0	0.2	23	8
4	Hayden Gulch near Pagoda	401913107204100	8	1,380	162	8.1	0.3	126	14
5	Watering Trough Gluch near Hayden	09244460	26	1,010	105	7.8	0.2	105	12
6	Hubberson Gulch near Hayden	09244464	28	1,010	429	7.9	0.3	116	34
7	Sage Creek near Hayden	09244415	13	616	227	8.1	0.3	78	22
8	Grassy Creek at Grassy Gap	402330107082000	7	864	760	7.9	0.3	65	26
9	Middle Creek near Oak Creek	09243700	50	620	155	8.0	0.4	68	13
10	Fish Creek at mouth near Milner	402530106585700	7	671	171	8.2	0.3	61	11
11	Grassy Creek near Mt. Harris	09244300	9	1,880	574	8.0	0.2	168	53
12	Foidel Creek near Oak Creek	09243800	33	862	294	7.7	0.4	97	27
13	Foidel Creek at mouth near Oak Creek	09243500	36	1,200	540	7.9	0.4	142	74
14	Flume Gulch near Craig	402911107323500	7	4,410	488	8.0	0.1	291	46
15	Smuin Gulch near Hayden	402829107193700	4	3,320	914	8.2	0.1	180	45
16	Smuin Tributary near Hayden	402845107185100	6	3,980	1,490	8.0	0.1	167	20
17	Dill Gulch near Hayden	402605107181500	3	5,540	1,830	8.2	0.1	203	15
18	Stokes Gulch near Hayden	09244470	9	4,120	2,930	8.0	0.5	170	88

water-chemistry data

values in milligrams per liter except where noted]

Magnesium		Sodium		Total alkalinity		Sulfate		Chloride		Dissolved solids	
$\bar{x}$	S	$\bar{x}$	S	$\bar{x}$	S	$\bar{x}$	S	$\bar{x}$	S	$\bar{x}$	S
22	6	15	4	214	47	51	23	3.6	1.1	293	49
33	15	16	6	213	68	86	37	3.0	2.4	338	113
8.0	2.6	3.5	0.6	87	31	6.1	3.0	0.71	0.46	106	33
95	13	59	9	317	19	471	67	13	2	972	112
57	7	39	7	344	29	228	34	11	2	664	77
76	22	58	23	347	44	389	147	13	9	837	282
40	14	16	6	228	57	158	75	5.9	2.3	449	146
35	14	27	9	204	85	154	59	5.7	1.8	424	162
29	7	30	11	220	55	117	37	4.4	1.4	395	97
37	10	41	21	193	48	190	89	4.9	2.3	463	137
116	43	124	59	278	114	816	274	29	13	1,450	459
48	17	37	19	291	77	236	106	9.5	9.6	599	207
69	40	66	33	228	70	507	385	11	5	937	562
341	36	426	72	373	113	2,390	353	86	42	3,770	515
258	86	360	178	365	31	1,720	741	66	36	2,820	1,090
253	138	533	268	465	67	1,830	972	120	53	3,200	1,420
513	123	777	280	497	68	3,330	1,040	140	30	5,270	1,530
438	277	806	462	221	92	3,280	1,960	100	73	5,060	3,030

predominantly are fine grained, well sorted, and permeable; thus, they form the regional aquifers in the area. The remaining rocks in the area primarily result from two nonmarine depositional environments--deltas and swamps. These various types of rocks may form either local aquifers or local confining layers. Distributary sandstones were deposited in deltaic distributary channels and are linear and vary in thickness and lateral continuity. Coals were formed in poorly drained bank deposits associated with distributary sands in a deltaic setting. The coals usually are variable in thickness and extent. Local aquifers are present in most of these units. Local confining layers, consisting of freshwater shales and mudstones, were formed in the low-energy environments of deltas and swamps. Thickness and lateral continuity of these deposits also are variable.

Regional Aquifers

Two lithologic units within the stratigraphic boundaries of the study area are classified as regional aquifers--the Trout Creek Sandstone Member of the upper Iles Formation and the Twentymile Sandstone Member of the Williams Fork Formation.

Trout Creek Aquifer

The Trout Creek aquifer is the lower of the two regional aquifers, generally occurring from 1,000 to 1,100 ft below the top of the Twentymile Sandstone Member (fig. 9). Thickness averages about 100 ft, with a range from 70 to 150 ft. The aquifer extends from the formational outcrops in the study area, into the subsurface to the west of the study area, and to the north of the Buck Peak anticline. The Yampa River forms a hydrologic boundary along the northern edge of the study area. The aquifer overlies about 300 ft of a marine shale that hydraulically isolates it from underlying formations. The upper aquifer boundary is poorly defined by nonmarine mudstones, thin, poorly developed coals, and silty sandstones, all of which can be classified as confining beds. The confining beds vary in thickness and lateral continuity and thus form a leaky confining layer.

Twentymile Aquifer

Physical characteristics of the Twentymile Sandstone Member are nearly identical to those of the Trout Creek Sandstone Member because of their similar depositional histories and environments. However, the Twentymile aquifer is less well defined by the boundaries of the geologic unit than is the Trout Creek aquifer. In the western part of the area, the Williams Fork Formation is much sandier than in the east, and the limits of the Twentymile aquifer are difficult to discern. In the eastern and western parts of the area, the Twentymile Sandstone Member is overlain by interbedded sandstone, coal, and shale of the upper member of the Williams Fork Formation. Because closely overlying and underlying sandstone and coal likely are in hydraulic connection with the Twentymile Sandstone Member, they are here considered to be part of the Twentymile aquifer. The aquifer thus extends from the base of the Lewis Shale to the top of the middle member in the eastern part of the area. In the central and western parts of the area the aquifer limits are poorly defined but include overlying and underlying hydraulically connected sandstone units. The middle member forms an underlying regional confining layer to the Twentymile aquifer. The Lewis Shale forms an overlying confining layer. Both units consist of as much as 600 ft of uniform marine shale in Twentymile Park. The Twentymile aquifer extends laterally from formational outcrops in the area to, and beyond, the hydrologic boundary of the Yampa River and beyond the western limit of the study area.

Local Aquifers

Local aquifers do not underlie the entire area but may have an important effect on the hydrology of some parts of the area. The aquifers are composed of discontinuous beds of coal or sandstone.

Coal Aquifers

Coal beds may form the most important aquifers in the area. Fracturing produces secondary permeability in the coal and can make a coal seam the most permeable bed in a specific area. More important, some coal aquifers are disrupted by mining, allowing aquifer water to come into direct contact with surface water or leachate from spoils.

The metamorphosed nature of coal makes it hydrologically similar to fractured crystalline materials. Limited data are available on fracturing in the local coal beds. In one area on the Fish Creek anticline, core samples indicated extensively fractured Wadge coal. These cleats primarily are conchoidal and oblique to subparallel lineations in the coal. No estimation of fracture density was made (Nancy Driver, U.S. Geological Survey, oral commun., 1980). Because of the limited data, no conclusions were reached on preferential fracture directions, nor was any attempt made to define the fracture pattern. The most likely patterns would be fractures parallel and perpendicular to the original bedding.

In the area east of Hayden Gulch, three coal seams in the lower member of the Williams Fork Formation and one coal seam in the upper member may be significant aquifers. The coal seams are, in ascending order, the Wolf Creek, Wadge, and Lennox coal of the lower member, and the Fish Creek coal of the upper member of the Williams Fork Formation. The Fish Creek coal seam is the only significant coal aquifer in the upper member of the Williams Fork Formation. Erosion has markedly decreased the areal extent of this coal; it occurs only beneath the Lewis Shale in the Twentymile Park area.

Much less information is available for the area west of Hayden Gulch. In general, the number and thickness of coal seams increases toward the west (Bass and others, 1955). A few isolated beds occur in the middle member of the Williams Fork Formation; however, these beds are difficult to correlate from drill hole to drill hole and probably are not laterally continuous. The lower member contains numerous thick seams, several of which correlate for a number of miles. The most widespread and most easily correlated seam west of Hayden Gulch is located 370 ft above the Trout Creek Sandstone Member. The bed is 10 ft thick and occurs in all drill holes and sections in that interval. The seam appears continuous through much of the central western area, but it eventually splits and cannot be correlated as it approaches the western boundary. A coal seam about 20 to 40 ft below the continuous coal seam also extends through part of the western area; it is about 5 to 10 ft thick and is not as continuous as the overlying seam. This particular seam is typical of the coal seams in the western area and correlates well for about 5 mi.

The upper member of the Williams Fork Formation contains coals that thicken appreciably toward the west. These seams are poorly correlated, indicating limited lateral continuity of coal beds. Most of these seams occur west of Hayden Gulch.

Thin Sandstone Aquifers

These local aquifers are not as important as regional aquifers or the coal aquifers; however, they can yield small quantities of water to wells. This type of aquifer consists of lenticular sandstone beds with a 40- to 60-ft maximum thickness. The aquifers generally are restricted to certain geographic localities and stratigraphic intervals (fig. 9).

Thin sandstone aquifers are most common in the west-central part of the study area. Here, two lenticular sandstone beds are located within the middle member of the Williams Fork Formation. These units are lithologically similar to the regional aquifers and consist of white to gray to light brown, moderately well-sorted, fine-grained quartz arenites that contain chert. The first local sandstone aquifer, 520 ft above the Trout Creek Sandstone Member, extends for about 12 mi and reaches a maximum thickness of 40 ft. The second sandstone aquifer was not entirely defined by drilling. This bed, about 700 ft above the Trout Creek Sandstone Member, seems to thicken to about 60 ft and extends a minimum of 9 mi (fig. 9). Local aquifers in the west central area are lenticular, reach a maximum thickness of about 50 ft, and extend for 10 to 20 mi. Fine-grained siltstone beds that overlie and underlie the sandstones form confining layers for these aquifers.

The thin sandstone and coal aquifers that are in the lower member of the Williams Fork Formation in the eastern part of the area seem to function as a single hydrologic unit and in this report are collectively referred to as the basal Williams Fork aquifer. This local aquifer consists of the three principal coal seams (Wolf Creek, Wadge, and Lennox) interbedded with shale and lenticular sandstone. The basal Williams Fork aquifer extends throughout the eastern part of the area, averages about 300 ft in thickness, and contains about 50 percent shale. The middle marine member of the Williams Fork Formation forms the overlying confining layer. Shale beds within and below the aquifer form a leaky confining layer between the basal Williams Fork aquifer and the underlying Trout Creek aquifer.

AQUIFER CHARACTERISTICS

The aquifer characteristics, hydraulic conductivity, transmissivity, porosity, specific storage, and storage coefficient are important in resource evaluations and modeling. In order to define these characteristics, aquifer tests were conducted in open holes and wells completed in a single interval, using pumping-well test and slug-test techniques. Laboratory analyses of rock samples collected from outcrops and drill cores also were used to define aquifer characteristics.

Methods of Determining Characteristics

Aquifer Tests

Pumping-well aquifer tests primarily were conducted for environmental impact evaluations at large strip mines in the eastern part of the area. These tests were done during the past 5 years by the U.S. Geological Survey and by private mining concerns. Results of 22 of these tests are listed in table 5. Locations of the wells tested are shown in figure 24. Hydraulic information is restricted to the coal-bearing zones, primarily the Wadge coal seam and rocks immediately above or below the coal; no information is available on the two regional aquifers. Information about storage coefficient was obtained at only a few wells because observation wells were not available at most of the pumping-well-test sites.

Transmissivity values from pumping-well tests are shown for the Wadge coal seam and its associated overburden and underburden in the lower member east of Hayden Gulch (table 5). Values range from 0.7 to 95 ft^2/d; the mean is 17 ft^2/d and the standard deviation is 20.6. Only one value, obtained from a well completed in an unknown thickness of aquifer northwest of Dry Creek, exceeds 50 ft^2/d.

All slug-type aquifer tests were conducted during the summer of 1980, primarily on wells drilled in 1976 and 1977. In all, 24 tests were successfully completed (table 6; fig. 24). Compared with pumping-well tests, the slug tests were done in a much wider combination of geographic and stratigraphic settings with a varied depth to the potentiometric surface. Aquifers were not heavily stressed by the slug test, and the resulting information is much less representative than the pumping-well test results. One significant figure was the assumed accuracy for these slug-test results.

Slug-test data were collected using a pressure transducer connected to a strip-chart recorder that had a resolution of one-tenth of a foot of hydraulic head. To simulate an instantaneous hydraulic-head change, a weighted, 20-ft-long, 1-inch-diameter pipe was used to displace water. After installation and calibration of the pressure transducer, the 1-inch-diameter pipe was inserted into the well, displacing water and causing a rise in head. Recovery to equilibrium was recorded on the strip chart. If the aquifer transmissivity value was small, only one recovery curve was generated. In a more transmissive aquifer, several insertion-removal cycles were measured to gather replicate information.

Table 5.--*Summary of data from pumping-well tests*

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

Well location[1]	Reported stratigraphic interval in Williams Fork Formation	Completion type	Trans-missivity (ft^2/d)	Saturated thickness (ft)	Hydraulic conduc-tivity (ft/d)	Storage coeffi-cient	Source[2]
4/86-7DAB	Wadge overburden	Open hole	8.3	47	0.2	--	2
4/86-18BBB	|	|	34	94	0.4	--	2
4/86-18BBB	|	|	43	73	0.6	--	2
4/86-18BCB	|	|	33	53	0.6	--	2
5/85-30BBD	Wadge underburden	Unknown	10.4	--	--	--	1
5/85-30CCC	Wadge coal and overburden	|	4.2	--	--	--	1
5/85-31CDA	|	|	1.1	--	--	--	1
5/85-13ABB	Lower member	Open hole	10.5	80	0.1	--	2
5/86-13ACC	|	|	3.7	157	0.02	--	2
5/86-29CDD	Wadge overburden	Single interval	4.3	11	0.4	--	2
5/86-29CDD	|	|	6.3	20	0.3	--	2
5/86-29CDD	Lower member	Open hole	8.6	20	0.4	--	2
5/86-29CDD	Wadge overburden	|	13.6	45	0.3	--	2
5/86-29CDD	|	|	13.6	41	0.3	--	2
5/86-32BBD	|	|	15.4	60	0.3	--	2
5/86-36DDB	Wadge coal and overburden	Unknown	0.7	--	--	--	1
5/87-11BDB	Wadge underburden	Open hole	22	151	0.2	--	3
5/87-19ABB	Lower member	Single interval	3.9	24	0.2	--	3
5/88-8CDC	Upper member	Open hole	95	--	--	--	4
6/87-34ACA	Wadge coal	Single interval	3.3	9	0.4	1×10^{-3}	3
6/87-34DDB	Wadge overburden	Open hole	16	135	0.1	--	3
6/88-33DBB	Upper member	|	33	--	--	--	4

[1]See figure 24 for well locations.

[2]Values reported in permitting documents submitted by 1, Pittsburg and Midway Coal Mining Co.; 2, Colorado Yampa Coal Co.; 3, Peabody Coal Co.; and 4, data from U.S. Geological Survey.

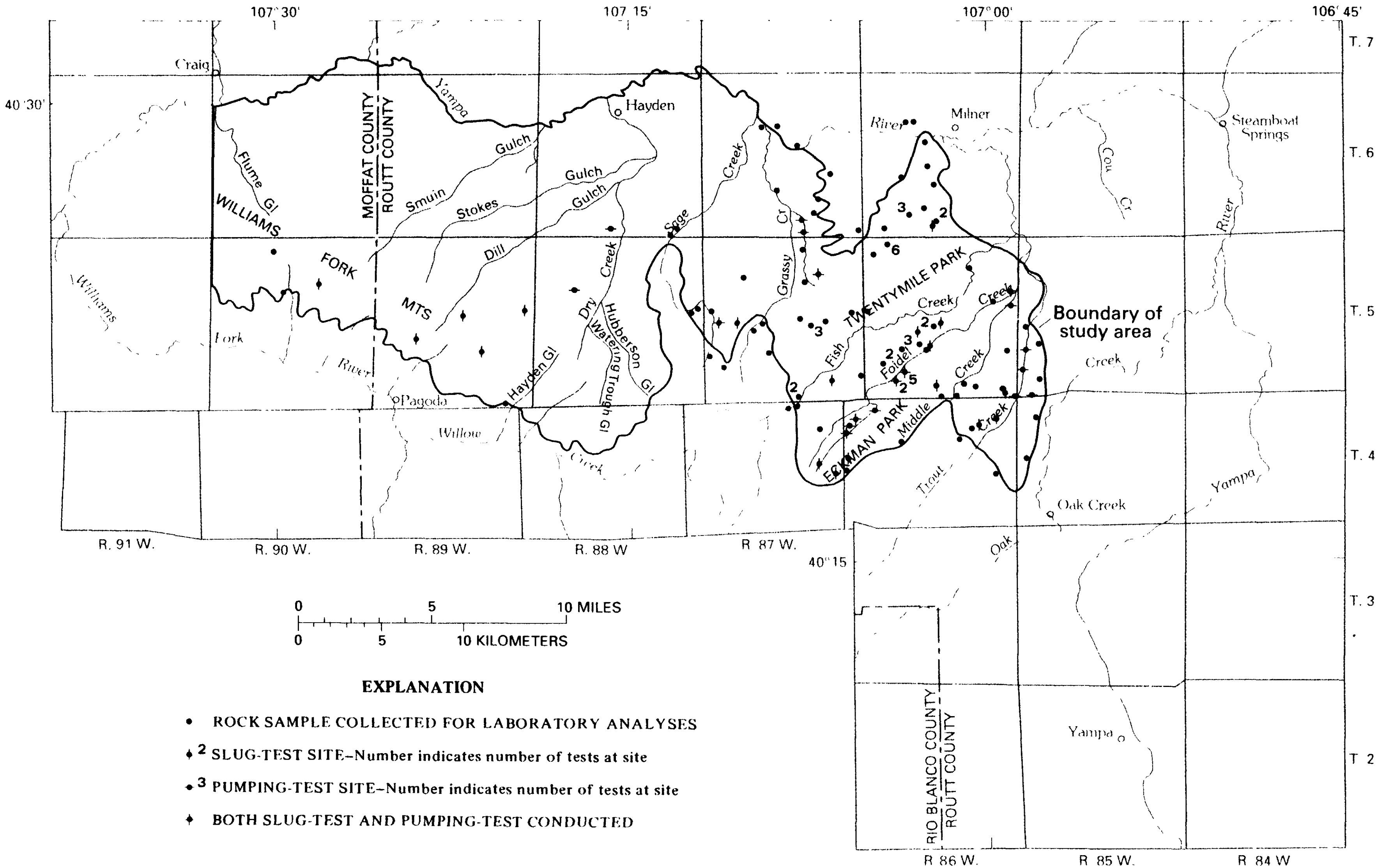


Figure 24.--Location of rock-sample, slug-test, and pumping-well test sites.

Table 6.--*Summary of data from slug tests in wells*

[ft^2/d, feet squared per day; ft, feet; ft/d, feet per day]

Well location[1]	Stratigraphic interval (members and coalbed are in Williams Fork or Iles Formations)	Trans-missivity (ft^2/d)	Saturated thicknes (ft)	Hydraulic conductivity (ft/d)
4/86-11DCC	Trout Creek Sandstone member and lower member	100-630	106	0.9-6.0
4/86-14AAA	|	3-7	143	0.02-0.05
4/86-18BBB	Lower member	210	94	2.0
4/86-18BCB	|	180-220	53	3.0-4.0
4/86-19BBD	|	1-3	44	0.02-0.07
4/86-24BCB	Wadge, overburden, lower member	4	260	0.02
4/87-24DBD	Wolf Creek coal and underburden	0.4	60	0.007
5/86-21AAA	Lewis Shale and upper member	80-250	10	8.0-25.0
5/86-21BCC	Lewis Shale	230-700	[2]188	1.0-4.0
5/86-21CDD	Lewis Shale and upper member	2.5	68	0.04
5/86-29CDD	Lower member	1.0	10	0.10
5/86-29CDD	|	5-30	45	0.1-0.7
5/86-32BBD	|	8-30	60	0.1-0.5
5/86-36CAC	Trout Creek Sandstone Member and lower member	1-4	165	0.006-0.02
5/87-11BDB	Wadge and underburden	0.4-3.0	151	0.003-0.02
5/87-19ABB	Lower member	30-90	24	1.0-4.0
5/87-20BBA	Wadge coal	0.2	10	0.02
5/89-13ACC	Upper member	70-210	50	1.0-4.0
5/89-15CAB	Upper member	6.0	128	0.05
5/89-20ACD	Twentymile Sandstone Member	50-130	112	0.4-1.0
5/89-23CCC	|	830	110	8.0
5/89-35ACA	Trout Creek Sandstone Member	960-2,800	304	3.0-9.0
5/90-11BCC	Twentymile Sandstone Member	1,000-3,000	210	5.0-10.0
6/86-33ADB	Trout Creek Sandstone Member and lower member	30-90	532	0.06-0.2

[1]See figure 24 for well locations.
[2]Shale interval shown for well completed only in weathered shale.

The resulting time-drawdown data were analyzed by one of two methods, depending on individual hydraulic conditions at each well. The first method, described by Cooper and others (1967), assumes a fully penetrating well in a homogeneous isotropic aquifer. The method is valid only in confined aquifers, which is a severe restriction. The solution involves a type-curve matching procedure similar to the Theis technique for pumping-test analysis. This procedure may be used for a recovering head resulting from either injection or removal of water. It yielded the best information about confined aquifers in wells that have sufficient water depth to allow the displacement pipe to be lowered beyond head fluctuation range. The procedure is sensitive to unconfined conditions and was not used in analysis of wells penetrating unconfined aquifers.

The second interpretive procedure is that of Bouwer and Rice (1976). It is based on the Theim equation and assumes the bailing of a well under homogeneous and isotropic conditions. Unlike the first procedure, the well need not fully penetrate the aquifer and, more importantly, the aquifer can be unconfined. The calculation technique is more complex than the Cooper method; however, no type-curve matching is needed. This procedure was used only for larger transmissivity tests in unconfined aquifers. Both procedures were used to interpret results of several tests. Results generally indicated agreement within at least one significant figure, the reporting accuracy for slug tests in this study.

The range in transmissivity listed for each slug test (table 6) results from the use of minimum and maximum values for well radii in the slug-test formulas. The open-hole completion wells contained no gravel packing, requiring the assumption that the maximum radius is the drilled-hole radius and the minimum radius is the inside-casing radius. The larger the transmissivity, the greater the resulting range between maximum and minimum values.

The overall transmissivity range for all slug-test wells was much greater than pumping-well-test range. There are two principal reasons for this. First, slug tests were conducted over a wider range of geological and geographical conditions. Second, slug tests displace a much smaller aquifer water volume, which produces transmissivity estimates of lesser accuracy. Many wells were completed as open holes. The aquifer penetrated by these wells varied in thickness, lithology, and in the degree of cementation and fracturing. The quantity of water removed or added for this test usually was limited to less than one well volume. The actual volume of aquifer tested is quite small, and localized irregularities do not average out as they do in the longer term pumping tests. These irregularities, particularly fracturing, may have an effect on the transmissivity near the well. Experimental error was minimized by use of an automated data-collection system and the use of only one person to perform the test and interpret the data.

The hydraulic conductivity of an aquifer is calculated by dividing the transmissivity by the saturated thickness. For a well completed in a single interval, the saturated thickness was assumed to equal the perforated interval. This thickness was used to calculate hydraulic conductivity for the single-interval wells listed in table 5. For wells completed as open holes, the water from all water-yielding intervals in the well is free to mix, regardless of perforation locations because water in the annulus is directly connected to water in the casing. The resulting transmissivity value from an aquifer test is an integrated average of all saturated intervals, which makes it impossible to distinguish between conductive and nonconductive saturated zones. In addition, most open-hole wells in the study area are not cased to the bottom of the drill hole, and few are sealed at the bottom of the casing; this may allow upward movement of water from intervals below the well casing. To simplify the calculation of hydraulic conductivity, it was assumed that the wells were sealed by collapsing at the first thick shale or mudstone below the casing, producing an impermeable seal between the well and uncased borehole below. Formational collapse could occur in cased areas containing shales; however, there is no data to document the occurrence or frequency of this condition. Therefore, it was assumed that no collapsing occurred in the cased interval. The validity of the above assumptions is unknown; therefore, these values should be used with caution. The hydraulic-conductivity values listed in tables 5 and 6 for open-hole completed wells are based on these assumptions.

Using the above assumptions, saturated aquifer thickness in open-hole completed wells was assumed to be the total thickness of all sandstones and coal in the cased interval below water level, regardless of perforated intervals. Assuming that the saturated thickness is limited to perforated intervals is incorrect because of the direct hydraulic connection between the water in the annulus and casing. Assuming the total saturated thickness of the well to be the aquifer thickness also is incorrect because of the smaller permeability values of interbedded fine-grained rocks.

Rock-Sample Analyses

The aquifer-test results provide minimal information about the aquifer characteristics of the regional aquifers. The characteristics of these aquifers in the eastern part of the area are of particular concern because determination of aquifer characteristics is requisite to successful simulation of the ground-water system. Rock samples were collected for laboratory analyses in an effort to better define the character of the regional aquifers.

Eighty-one rock samples (table 7) were collected from outcrops of the Twentymile Sandstone, Trout Creek Sandstone, and Tow Creek Sandstone Members. (The Tow Creek Sandstone Member is a potential aquifer in the middle part of the Iles Formation that subsequently was excluded from consideration in this study because of insignificant hydraulic connection with aquifers in the study area.) Twenty-two samples (table 7) also were collected from drill cores provided by the Twentymile Coal Co. The cores were obtained from depths of 301 to 1,432 ft in sandstone or siltstone of the Twentymile Sandstone Member, lower member of the Williams Fork Formation, and Trout Creek Sandstone Member. Physical characteristics of the regional aquifer samples were typical of the formational characteristics described in the "Stratigraphy" section of this report. All samples were intact, unfractured, and moderately to well indurated.

Table 7.--*Physical properties of sampled bedrock materials*

[ft, feet; g/cm³, grams per cubic centimeter; mD, millidarcys; ft/d, feet per day; mm, millimeters; ϕ, $-\log_2$ (d), where d is grain diameter measured in millimeters; Kwt, Twentymile Sandstone Member of Williams Fork Formation; Kws, siltstone bed in the lower member of Williams Fork Formation; Kwb, sandstone bed in the lower member of Williams Fork Formation; Kit, Trout Creek Sandstone Member of Iles Formation; Kio, Tow Creek Sandstone Member of Iles Formation; --, no data]

Sample location	Formation	Depth (ft)	Bulk density (g/cm³)	Porosity (percent)	Gas permeability (mD)	Hydraulic conductivity (ft/d)	Grain-size distribution (percent finer) Sieve size (mm) 1.0 (ϕ=0)	0.5 (ϕ=1)	0.25 (ϕ=2)	0.125 (ϕ=3)	0.0625 (ϕ=4)
					Williams Fork Formation						
4/86-14BAB	Kwt	0	2.16	20	4.2	1.7×10^{-3}	100	98.6	93.5	35.2	6.9
4/87-13BBA	Kwt	0	2.03	23	87	4.9×10^{-2}	100	99.0	95.1	35.3	7.9
5/86-6AAD	Kws	1,259	2.23	17	7.1	3.0×10^{-3}	100	97.8	92.1	82.3	13.7
5/86-6AAD	Kws	1,259	2.21	17	7.4	3.2×10^{-3}	--	--	--	--	--
5/86-6AAD	Kws	1,260	2.23	16	3.0	1.2×10^{-3}	--	--	--	--	--
5/86-6AAD	Kwb	1,375	2.25	15	1.4	4.9×10^{-4}	--	--	--	--	--
5/86-6AAD	Kwb	1,375	2.26	15	1.7	6.2×10^{-4}	--	--	--	--	--
5/86-6AAD	Kwb	1,375	2.25	16	2.2	8.3×10^{-4}	99.3	96.3	90.0	69.2	10.2
5/86-6CAA	Kwt	0	2.42	9.9	0.07	1.8×10^{-5}	--	--	--	--	--
5/86-10AAA	Kwt	0	2.21	16	18	8.5×10^{-3}	100	99.3	93.3	24.7	10.0
5/86-14ADC	Kwt	0	2.17	15	29	1.5×10^{-2}	100	99.4	96.9	90.8	24.4
5/86-18CBD	Kws	1,432	2.46	5.6	0.04	9.7×10^{-6}	--	--	--	--	--
5/86-21AAC	Kws	1,247	2.39	10	0.04	9.7×10^{-6}	--	--	--	--	--
5/86-21AAC	Kws	1,248	2.45	7.4	0.05	1.2×10^{-5}	--	--	--	--	--
5/86-21CCC	Kwt	0	--	--	--	--	99.8	99.4	94.6	26.6	8.9
5/86-25BAD	Kwt	0	2.00	25	219	1.4×10^{-1}	100	98.7	96.1	55.9	8.6
5/86-28BAB	Kwt	0	2.07	22	171	1.1×10^{-1}	100	99.6	93.6	17.3	6.9
[1]5/86-29BAA	Kwt	301	--	24	178	1.9×10^{-1}	--	--	--	--	--
5/86-29BAA	Kwt	302	2.00	24	162	9.7×10^{-2}	95.6	89.4	70.8	6.9	0.8
5/86-29BAA	Kwt	302	2.07	21	153	9.2×10^{-2}	100	98.9	94.4	33.0	12.3
5/86-30DBA	Kws	1,210	2.26	15	1.2	4.1×10^{-4}	--	--	--	--	--
5/86-30DBA	Kws	1,211	2.24	15	1.0	3.4×10^{-4}	--	--	--	--	--
5/86-34ABD	Kwt	0	2.03	24	244	1.6×10^{-1}	100	99.2	90.4	21.9	7.1
[1]5/86-35BCD	Kwt	0	2.05	23	49	5.1×10^{-2}	100	99.2	97.0	93.2	12.8
5/86-36CAB	Kwt	0	2.05	23	51	2.7×10^{-2}	99.1	97.1	92.7	68.1	9.6
[1]5/87-3ADC	Kwt	0	2.45	7.8	1.0	1.7×10^{-4}	100	97.8	75.5	30.8	14.8
5/87-8BDD	Kwt	0	2.41	10.0	0.5	1.6×10^{-4}	100	97.3	87.8	37.3	10.8
[1]5/87-10DAC	Kwt	0	1.89	29	243	3.1×10^{-1}	100	99.5	71.7	16.1	4.1
5/87-13DBC	Kwt	0	2.58	4.0	0.04	9.7×10^{-6}	100	77.1	59.7	46.1	11.8
[1]5/87-15DCC	Kwt	0	2.07	22	94	3.9×10^{-2}	100	99.3	68.2	19.9	5.1
5/87-18CBA	Kwt	0	2.07	22	24	1.2×10^{-2}	100	97.4	92.2	86.0	23.3
5/87-21BAB	Kwt	0	2.05	23	61	3.3×10^{-2}	100	99.6	92.6	24.7	10.0
5/87-23ABB	Kwt	0	2.15	20	16	7.3×10^{-3}	100	99.2	92.0	39.5	8.8
5/87-23BBC	Kws	1,237	2.30	14	2.2	8.3×10^{-4}	99.7	97.5	93.2	86.0	21.8
5/87-23BBC	Kws	1,238	2.31	13	1.0	3.4×10^{-4}	--	--	--	--	--
5/87-23BBC	Kws	1,238	2.31	13	1.4	5.0×10^{-4}	--	--	--	--	--
5/87-27BBC	Kwt	0	2.04	23	72	4.0×10^{-2}	100	98.7	96.4	89.0	24.1
5/87-34DCB	Kwt	0	--	--	--	--	100	99.6	96.7	54.7	13.4
5/87-34DCB	Kwt	0	1.99	25	243	1.6×10^{-1}	100	99.5	96.5	37.7	11.6
[1]5/87-36AAA	Kwt	0	2.00	24	432	6.9×10^{-1}	100	99.3	72.1	18.8	8.8
5/88-13ACD	Kwt	0	2.12	20	17	8.0×10^{-3}	100	97.8	93.8	78.0	15.5
5/88-30ACC	Kwt	0	--	--	--	--	100	99.6	93.6	32.6	12.5
5/90-4BDB	Kwt	0	--	--	--	--	99.9	98.6	90.2	73.4	30.0
6/86-21BDD	Kwt	0	1.95	27	38	1.9×10^{-2}	100	92.7	93.6	87.5	12.5
6/86-28CDC	Kwt	0	1.98	25	231	1.4×10^{-1}	100	98.9	93.7	23.1	7.5
6/86-31DAC	Kwt	0	2.06	22	45	2.4×10^{-2}	100	99.7	93.9	29.9	5.1
[1]6/86-33ADB	Kwb	641	2.14	19	18	1.3×10^{-2}	--	--	--	--	--
6/86-33ADB	Kwb	642	2.19	17	7.3	3.2×10^{-3}	--	--	--	--	--
6/87-9DDC	Kwt	0	--	--	--	--	100	99.2	91.2	71.8	38.3
6/87-9CDC	Kwt	0	2.26	16	1.1	3.8×10^{-4}	100	90.0	76.9	40.8	7.3
6/87-28ADB	Kwt	0	2.29	14	0.5	1.6×10^{-4}	100	97.9	89.8	37.2	12.7
6/88-35DAD	Kwt	0	2.26	15	0.7	2.3×10^{-4}	100	95.9	80.4	36.0	13.6

Table 7.--*Physical properties of sampled bedrock materials*--Continued

Sample location	Formation	Depth (ft)	Bulk density (g/cm^3)	Porosity (percent)	Gas permeability (mD)	Hydraulic conductivity (ft/d)	Grain-size distribution (percent finer) Sieve size (mm) 1.0 (ϕ=0)	0.5 (ϕ=1)	0.25 (ϕ=2)	0.125 (ϕ=3)	0.0625 (ϕ=4)
					Iles Formation						
4/85-7DDC	Kit	0	2.37	11	0.2	5.8×10^{-5}	98.1	86.7	72.6	54.5	9.6
5/85-19BDD	Kit	0	2.21	16	3.4	1.4×10^{-3}	100	99.2	95.2	60.0	6.4
[1]4/86-15DAB	Kit	0	2.10	21	56	1.8×10^{-2}	100	98.8	82.4	18.0	5.7
4/86-17DAB	Kit	0	2.11	20	13	6.1×10^{-3}	100	98.4	75.8	22.0	6.3
4/86-19CBC	Kit	0	1.98	25	79	4.4×10^{-2}	100	99.4	97.2	91.7	20.7
4/86-24CAA	Kit	0	2.11	19	15	6.8×10^{-3}	100	97.9	61.1	16.5	4.8
4/87-11BAB	Kit	0	--	--	--	--	100	99.7	96.9	89.0	17.7
4/87-11BCB	Kit	0	2.20	19	7.0	4.4×10^{-3}	100	99.1	94.1	38.1	15.2
5/85-19BCA	Kit	0	2.12	19	6.8	2.9×10^{-3}	100	98.3	93.2	67.4	15.5
5/85-31AAC	Kit	0	2.22	16	0.8	2.7×10^{-4}	100	91.7	78.3	66.8	30.1
5/86-33DDD	Kit	0	--	--	--	--	99.8	96.6	81.9	66.0	40.7
5/86-34CDC	Kit	0	2.68	2.2	0.02	4.4×10^{-6}	--	--	--	--	--
5/87-20ADD	Kit	0	2.22	17	5.8	2.4×10^{-3}	100	92.5	82.9	60.6	10.7
5/87-28ACB	Kit	0	2.07	22	14	6.6×10^{-3}	100	98.8	94.8	89.0	12.6
5/87-30BBD	Kit	0	2.11	21	89	5.1×10^{-2}	100	98.7	56.6	19.5	6.0
[1]5/87-30DDB	Kit	0	1.98	25	568	2.7×10^{-1}	100	99.9	97.4	28.4	5.4
5/88-13DBB	Kit	0	2.08	22	22	1.1×10^{-2}	100	98.1	93.9	63.9	15.6
5/89-36CCC	Kit	0	--	--	--	--	100	99.5	91.6	12.5	5.0
5/90-9DAC	Kit	0	--	--	--	--	99.9	98.6	88.9	62.7	26.5
[1]6/86-8DCB	Kit	0	2.01	26	35	1.5×10^{-2}	--	--	--	--	--
[1]6/86-8DDB	Kit	0	2.10	21	150	3.6×10^{-2}	100	97.0	71.5	15.3	5.6
6/86-16CAB	Kit	0	2.08	22	44	2.3×10^{-2}	100	97.5	93.4	89.2	21.4
6/86-20CDA	Kit	0	2.29	14	3.5	1.4×10^{-3}	100	99.2	94.7	44.4	17.7
6/86-28ABA	Kit	0	--	23	53	2.8×10^{-2}	100	97.8	93.3	73.7	7.6
[1]6/86-32ABD	Kit	1,151	2.28	14	2.3	1.2×10^{-3}	100	99.1	71.8	27.3	11.1
6/86-32ABD	Kit	1,152	2.26	15	4.8	1.9×10^{-3}	100	99.4	72.1	24.4	13.1
6/86-32ABD	Kit	1,153	2.25	15	9.9	4.5×10^{-3}	100	99.4	74.2	27.0	12.5
6/87-15DBB	Kit	0	2.09	21	15	7.0×10^{-3}	100	96.5	90.3	81.6	10.6
6/87-23DAD	Kit	0	2.13	20	22	1.1×10^{-2}	100	98.3	94.3	49.6	8.0
6/87-26CAA	Kit	0	2.16	19	13	6.1×10^{-3}	100	99.4	94.6	40.3	16.5
[1]6/87-35BBA	Kit	0	2.10	21	647	1.9×10^{-1}	100	99.1	57.8	6.7	2.8
6/87-36DAD	Kit	0	2.12	20	19	9.2×10^{-3}	100	96.8	91.9	86.7	17.4
6/88-35DDC	Kit	0	2.15	19	14	6.8×10^{-3}	100	95.6	89.6	83.8	44.1
4/85-8CAA	Kio	0	2.29	14	1.1	3.8×10^{-4}	100	97.6	89.0	42.9	10.2
4/85-19ADA	Kio	0	2.31	14	3.7	1.5×10^{-3}	100	96.8	87.4	41.1	10.5
4/85-19ADA	Kio	0	2.22	16	18	8.3×10^{-3}	100	98.2	72.0	24.7	9.2
4/85-30ACC	Kio	0	2.28	14	3.5	1.4×10^{-3}	100	96.3	72.8	21.3	5.8
4/85-31BAD	Kio	0	2.20	17	3.8	1.5×10^{-3}	100	98.3	85.3	33.6	7.5
4/85-31BBD	Kio	0	2.19	18	16	7.7×10^{-3}	100	97.7	91.5	49.8	11.6
4/86-22ACD	Kio	0	2.30	14	2.0	7.3×10^{-4}	100	96.4	88.5	72.0	14.2
4/86-23ACC	Kio	0	2.15	19	43	2.2×10^{-2}	100	99.8	91.2	20.0	3.9
4/86-23BAB	Kio	0	2.20	17	1.6	5.8×10^{-4}	100	99.3	77.7	24.6	5.2
4/86-28CCD	Kio	0	2.11	20	53	2.8×10^{-2}	100	99.5	76.7	24.7	8.4
4/87-10ACC	Kio	0	2.14	20	278	1.8×10^{-1}	100	99.1	84.9	21.3	7.8
4/87-34DBA	Kio	0	2.15	20	6.2	2.7×10^{-3}	100	96.5	89.2	55.6	12.5
5/85-20CAB	Kio	0	2.08	22	324	2.1×10^{-1}	100	99.0	64.7	11.1	4.5
5/86-1BAD	Kio	0	2.42	11	0.4	1.3×10^{-4}	100	98.9	92.5	29.2	9.9
5/88-25DAA	Kio	0	2.01	25	352	2.3×10^{-1}	100	98.3	92.8	35.6	8.9
6/86-23BCC	Kio	0	2.14	20	18	8.3×10^{-3}	100	99.1	94.0	43.6	15.0
6/86-25BAA	Kio	0	2.12	21	17	8.0×10^{-3}	100	98.5	94.6	82.8	15.5
6/86-25DBA	Kio	0	2.58	6.6	0.2	5.8×10^{-5}	100	96.9	87.8	72.2	24.9

[1]Data used in figure 25.

Plugs 1 in. in diameter and 1.25 in. long were cut from most samples for use in a helium gas expansion porosimeter, gas permeameter, water permeameter, and porometer. Most samples were analyzed for bulk density, porosity, and gas permeability. Grain-size distributions also were determined on a disaggregated part of each rock sample. Laboratory hydraulic-conductivity determinations were made on 14 samples in order to define a relation between gas permeability and hydraulic conductivity. This relation (fig. 25) was used to convert the determinations of gas permeability into estimates of hydraulic conductivity. The line of relation defined by the data in figure 25 is below the theoretical maximum (Klinkenberg relation; Klinkenberg, 1941) because clay in the sample reacts with water to decrease the permeability of the wetted sample.

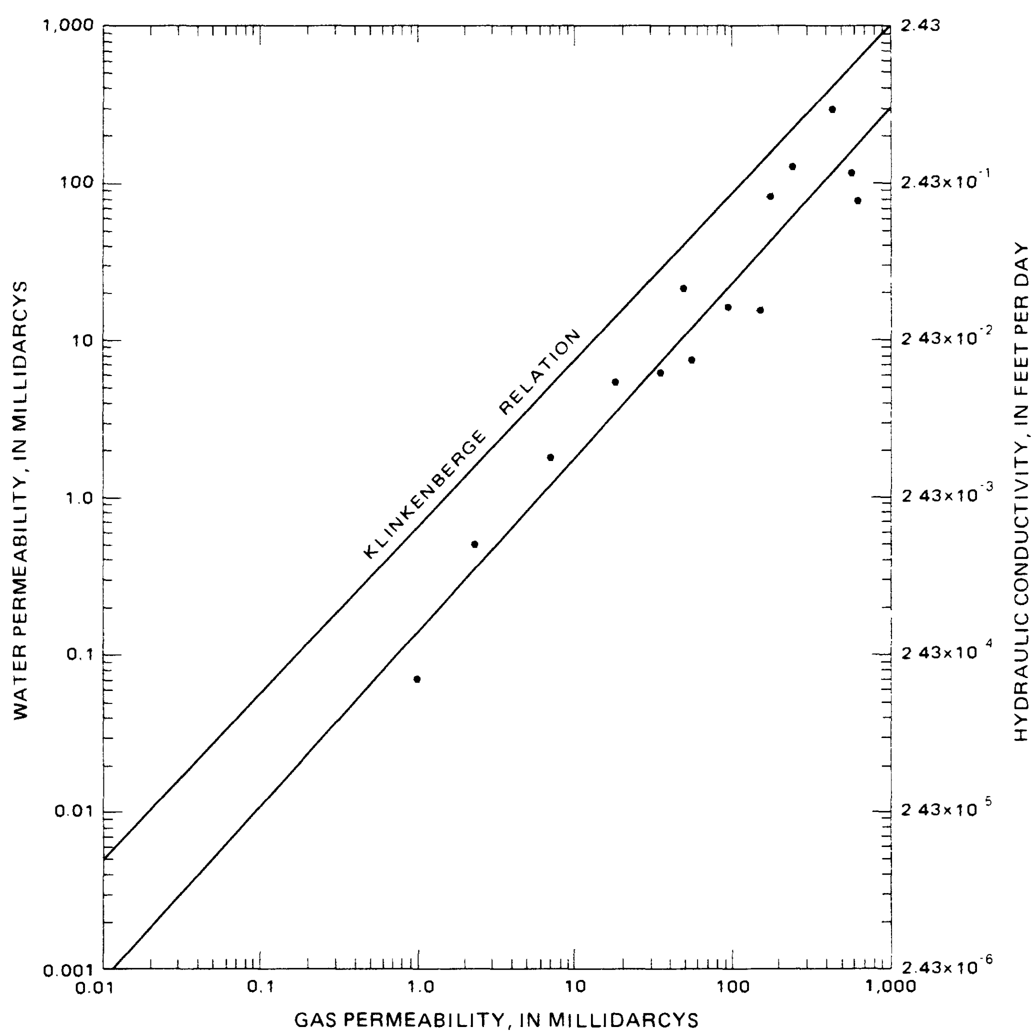


Figure 25.--Relation between gas permeability and hydraulic conductivity in samples from regional aquifers.

Hydraulic Conductivity

The hydraulic conductivity of the Trout Creek aquifer is defined for 33 data points in the eastern part of the area. The areal distribution of data values seems random, and no clear regional trend in hydraulic conductivity is evident. The data are approximately log-normally distributed, have a geometric mean of 5.1×10^{-3} ft/d, a standard deviation of 5.5×10^{-2} ft/d, and range from 4.4×10^{-6} to 2.7×10^{-1} ft/d. The hydraulic conductivity of the basal Williams Fork aquifer is defined for 53 data points. Here again, no clear pattern of regional trend in hydraulic conductivity is evident, although values seem to be larger in Eckman Park and near Trout Creek. The data are approximately log-normally distributed, have a geometric mean of 1.1×10^{-1} ft/d, a standard deviation of 8.3×10^{-1} ft/d, and range from 3.0×10^{-5} to 4.2 ft/d. In the Twentymile aquifer, hydraulic conductivity in the eastern area is defined for 40 data points, which indicate no regional trend in hydraulic conductivity. The data are approximately log-normally distributed, have a geometric mean of 1.4×10^{-2} ft/d, a standard deviation of 1.2×10^{-1} ft/d, and range from 9.7×10^{-6} to 6.9×10^{-1} ft/d.

The geometric mean values for the hydraulic conductivity of the three aquifers indicate that the basal Williams Fork aquifer is about 10 times more permeable than the Twentymile aquifer and is about 20 times more permeable than the Trout Creek aquifer. The difference between the mean hydraulic-conductivity values is statistically significant at the 1 percent level in a Student's t test. The difference in hydraulic conductivity may be due to the effects of secondary permeability produced by fractures in the coal beds in the basal Williams Fork aquifer. Unfractured coal is relatively impermeable. However, results of eight aquifer tests in the Wadge coal indicate that the mean hydraulic conductivity of this coal is 3.5×10^{-1} ft/d--about three times as large as the hydraulic conductivity of the basal Williams Fork aquifer as a whole. Although the data are few, the above results indicate that coal beds in the study area may be relatively permeable.

The effects of secondary permeability in the sandstone aquifers are more difficult to quantify. If fracturing enhances water movement in the sandstone, hydraulic conductivity based on aquifer tests could be larger than hydraulic conductivity based on laboratory analyses of unfractured rock samples. Nine aquifer tests in the Twentymile Sandstone had a mean hydraulic conductivity of 2.1×10^{-2} ft/d. Thirty-one hydraulic conductivity values from laboratory analyses of unfractured rock samples had a mean of 1.2×10^{-2} ft/d. The difference between these two numbers is not stastically significant at the 1 percent level of a Student's t test, indicating that secondary permeability in sandstone may be hydrologically insignificant or highly localized.

Fracture patterns on outcrops of Twentymile Sandstone Member indicate that joint and fracture density is highly variable in the eastern part of the study area. North of Grassy Gap (fig. 26), the sandstone forms massive cliffs that have unfractured intervals of hundreds of feet. Northwest of Twentymile Park (fig. 27), joints and fractures occur at intervals of 10 to 100 ft; to the northeast of Twentymile Park, joints and fractures are present at intervals of 10 ft or less (fig. 28). The effects of secondary permeability at depth in the sandstones likely are small because of lesser density of fracturing in the subsurface and minimal fracture interstice due to overburden load.

Figure 26.--Massive cliffs formed by outcrops of the Twentymile Sandstone Member of the Williams Fork Formation north of Grassy Gap.

Figure 27.--Moderately fractured outcrops of the Twentymile Sandstone Member of the Williams Fork Formation northwest of Twentymile Park.

Figure 28.--Dense joint and fracture pattern enhanced by erosion on the exposed dip slope of the Twentymile Sandstone Member of the Williams Fork Formation northeast of Twentymile Park.

The hydraulic conductivity of the shale and siltstone beds that form confining layers in the study area are computed from 14 lateral hydraulic-conductivity determinations on drill-core samples of unweathered siltstone and 12 vertical hydraulic-conductivity determinations on drill-core samples of unweathered marine shale. The respective mean hydraulic-conductivity values of 8.1×10^{-4} and 4.4×10^{-4} ft/d are not statistically different at the 1 percent level of significance in a Student's t test. Both the lateral and vertical hydraulic conductivity of unfractured siltstone or shale confining layers in the eastern part of the area are assumed to be equal to the mean of the lateral and vertical values (3.6×10^{-4} ft/d). Effects of fracturing are not documented by field data.

One aquifer test in the Lewis Shale indicates a relatively large value of hydraulic conductivity (table 6, sample 5/86-21BCC). Secondary fracturing from weathering or faulty well construction may have caused this anomalously large value.

Transmissivity

The transmissivity distribution in the aquifers in the eastern part of the area was calculated as the product of mean hydraulic conductivity and the aggregate thickness of water-yielding materials in the aquifer. The resulting transmissivity of the Trout Creek aquifer ranges from 0.5 to 0.8 ft^2/d across the area. This small range is the result of the relatively uniform thickness of the aquifer (100 to 150 ft). A median value of 0.65 ft^2/d is consistent with the range and distribution of transmissivity. The 100- to 200-ft aggregate thickness of the basal Williams Fork aquifer produces transmissivity values that range from less than 10 ft^2/d to more than 25 ft^2/d. One area of small transmissivity is located in the southern part of Twentymile Park. Areas of relatively large transmissivity are near Eckman Park, Trout Creek, Grassy Gap, and Hilberry Mountain (fig. 29; pl. 1). The transmissivity of the Twentymile aquifer is irregular because of the large and inconsistent range in thickness (80 to 180 ft). The average transmissivity was 3.5 ft^2/d. In outcrops, the saturated thickness of each aquifer thins rapidly to a point of zero saturation. The rate of thinning and the location of the point of zero saturation are poorly defined by data. Consequently, the rapid decrease in transmissivity at the margin of each aquifer also is poorly defined.

Transmissivity values in the western part of the area generally are larger than transmissivity values in the eastern part of the area (tables 5 and 6). This is not a function of thickness alone because well completions varied in thickness throughout the study area. The three most plausible reasons for the differences are variation in fracturing, diagenesis, and lithology. Lithology likely is the most important of the three. Sediments in the eastern area were deposited in a lower energy, deeper water environment, and consequently contain more marine shale than the western area. The resulting average grain size of the eastern lithology would be smaller, and the resulting permeability also should be smaller. Fracturing and diagenesis are present and cause local variations in permeability, but they do not differ systematically in the two areas and probably are not an important cause of the larger transmissivity in the west.

Porosity

Porosity determinations made on 77 rock samples from outcrops and drill cores indicated regional trends in porosity in some aquifers. Although the data are sparse, the porosity of the Trout Creek aquifer seems to average about 15 percent in a broad band extending from Twentymile Park toward Hayden (fig. 30). Porosity along parts of the northern and southern margins of the aquifer averages about 22 percent. A similar pattern is indicated by the porosity data for the Twentymile aquifer, although the smaller porosity band is narrower than is indicated for the Trout Creek aquifer. Porosity averages about 12 and 23 percent in the two areas indicated in the Twentymile aquifer (fig. 31). Insufficient data are available to define trends in the porosity of the basal Williams Fork aquifer; porosity in the 16 samples ranges from 5.6 to 19 percent, has a mean of 14.1 percent, and a standard deviation of 3.6.

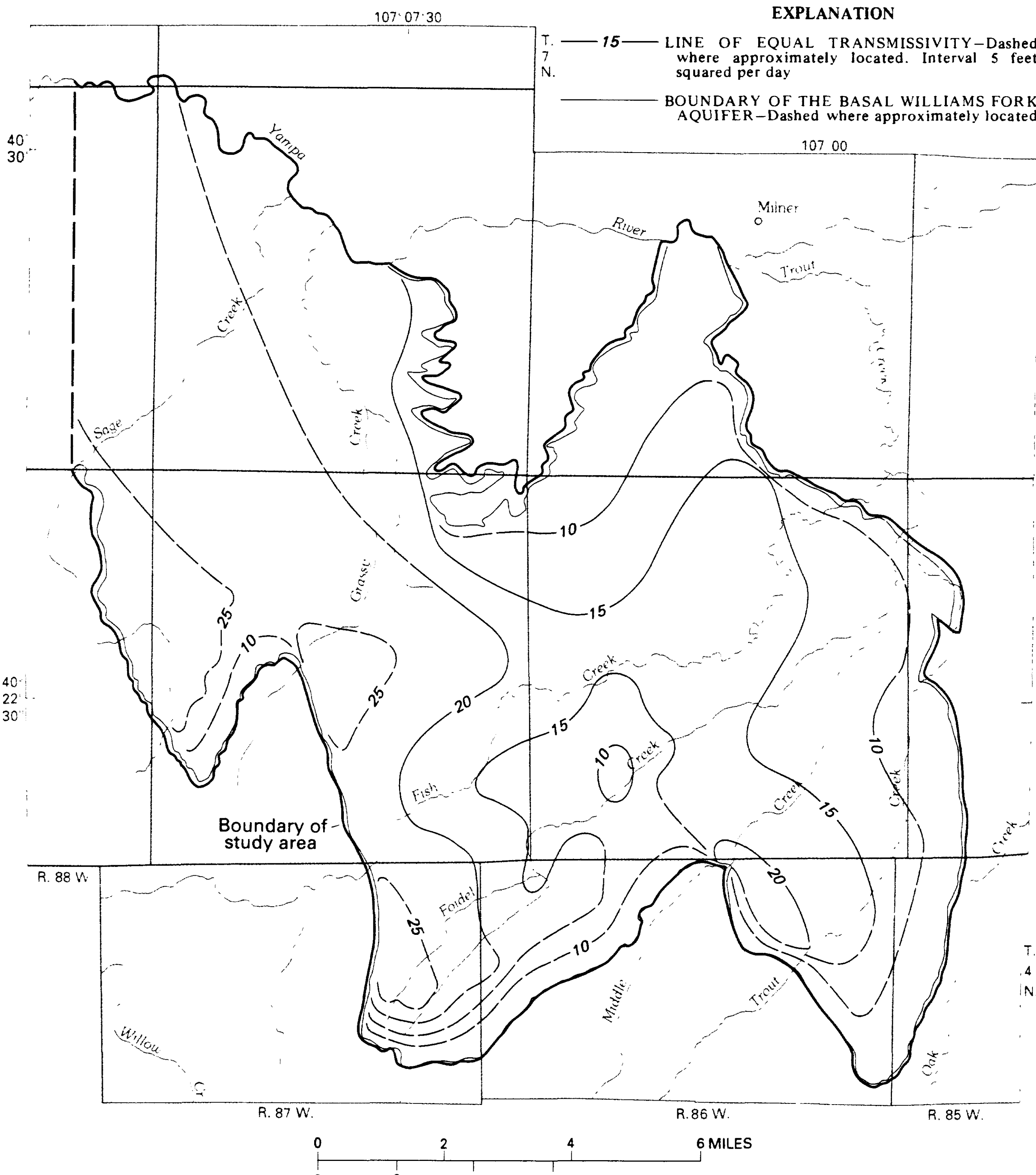
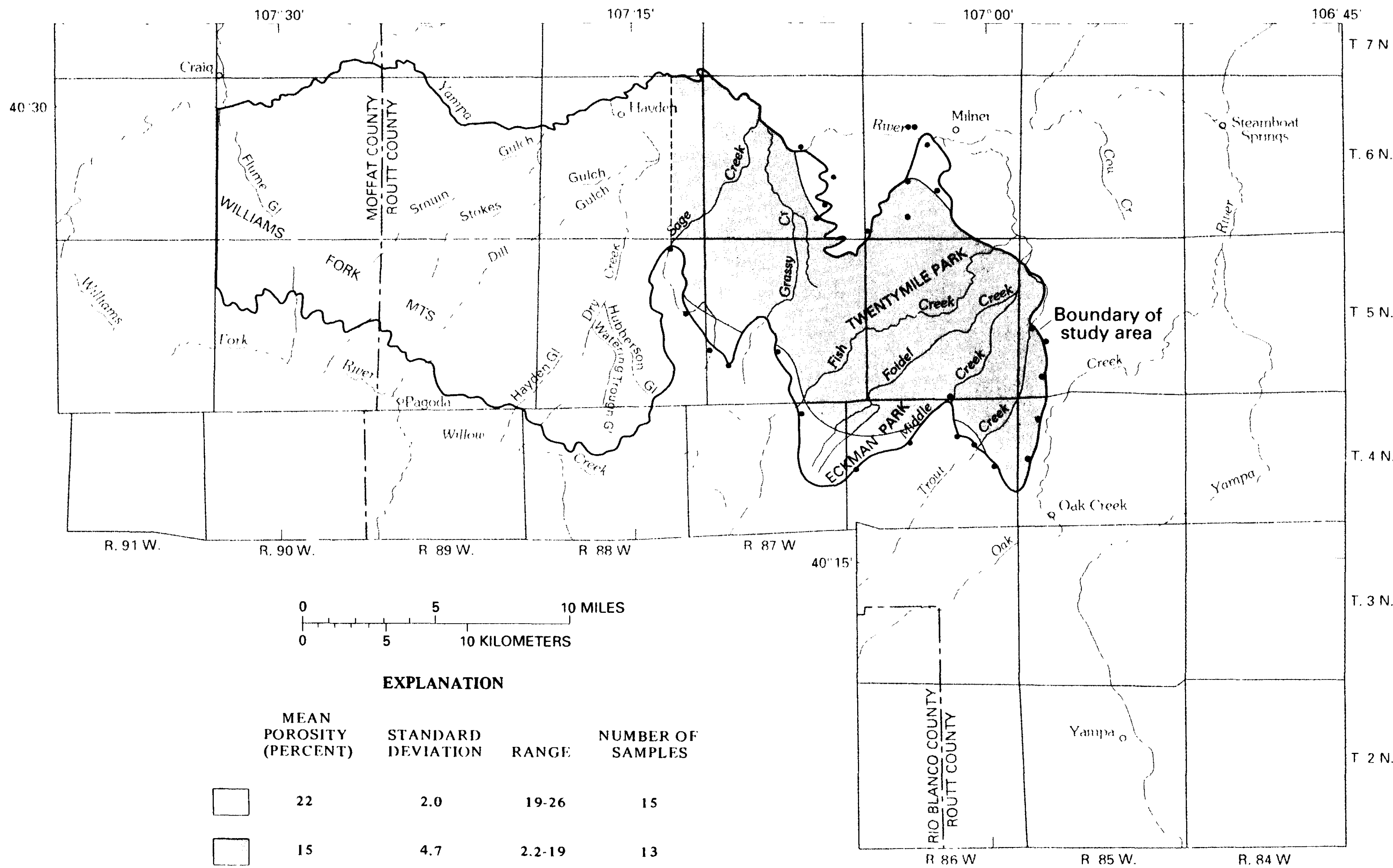


Figure 29.--Transmissivity of the basal Williams Fork aquifer in the eastern part of the study area.



EXPLANATION

	MEAN POROSITY (PERCENT)	STANDARD DEVIATION	RANGE	NUMBER OF SAMPLES
☐	22	2.0	19-26	15
☐	15	4.7	2.2-19	13

• DATA POINT

Figure 30.--Porosity of the Trout Creek aquifer in the eastern part of the study area.

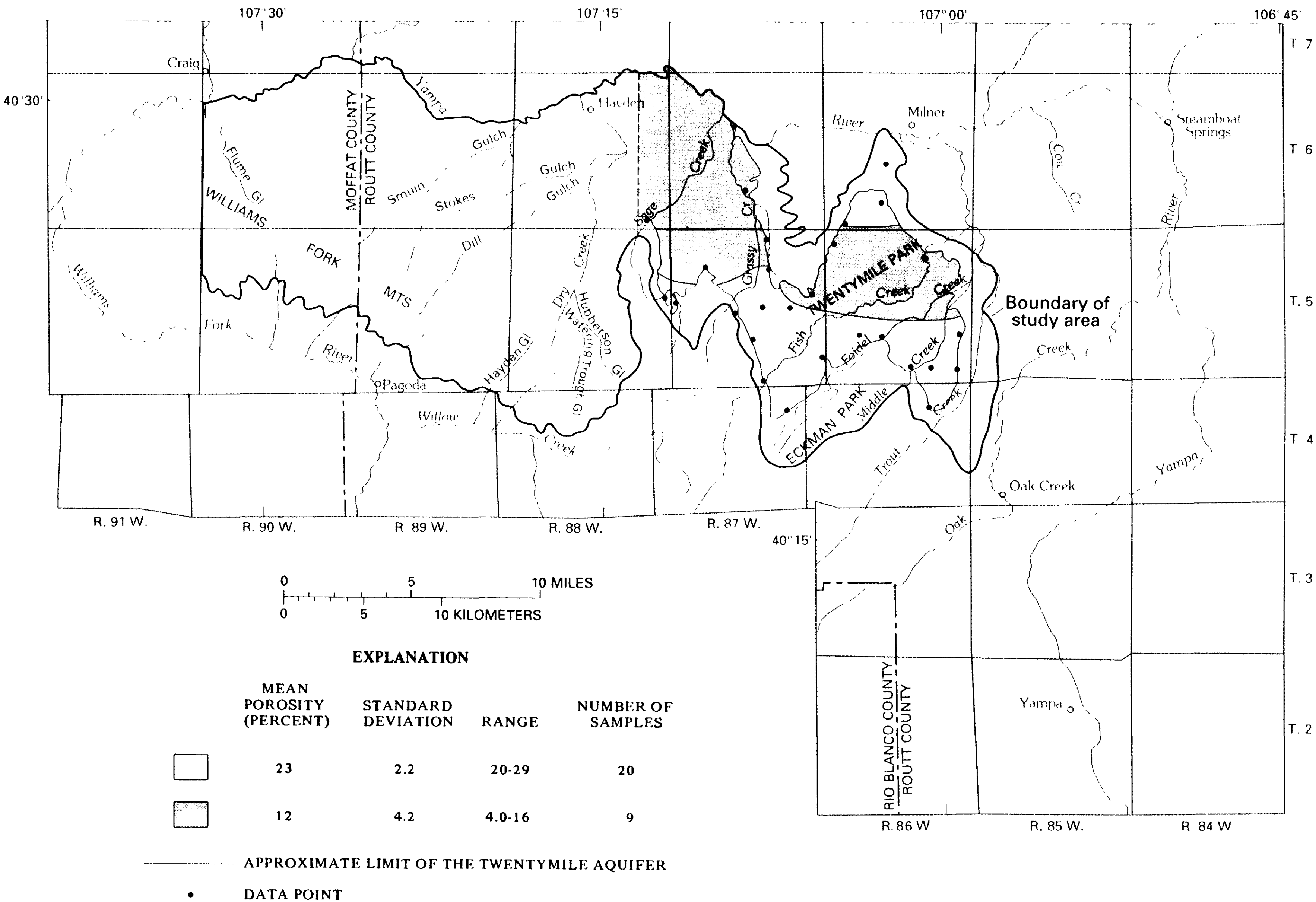


Figure 31.--Porosity of the Twentymile aquifer in the eastern part of the study area.

Specific Storage and Storage Coefficient

In a confined aquifer, the specific storage is related to the porosity and compressibility of the rock and water by the equation:

$$S_s = \gamma\ (\phi\ C_w + C_r)\ , \tag{1}$$

where

S_s = specific storage;

γ = specific weight of water;

ϕ = porosity;

C_w = compressibility of water; and

C_r = compressibility of rock.

Porosity of the sandstone strata in the eastern part of the area commonly ranges from 10 to 25 percent and averages about 20 percent. Compressibility of sandstone similar to that in the study area is about 1.5×10^{-6} in^2/lb (Fatt, 1958). These data, when used with the characteristics of water in the above equation, yield a specific storage of 9×10^{-7} ft^{-1}. This value is the volume of water the confined water-yielding sandstones release from or take into storage, per unit volume of rock, per unit change in head due to the compressive character of the water and rock.

In an unconfined aquifer, the volume of water released from or taken into storage by this process is insignificant when compared to the volume of water released by gravity drainage or filling of pore space in the rock. The storage coefficient of an unconfined aquifer is approximately equal to the specific yield of the water-yielding material and may be several orders of magnitude larger than the confined storage coefficient. No data are available to define the specific yield of the sandstones in the study area. However, sandstone that has a porosity of 20 percent could be expected to have a specific yield of about 1×10^{-1}.

Storage coefficient in a confined aquifer is equal to the product of specific storage and aquifer thickness. Thus, a 100-ft-thick confined aquifer in the Twentymile Sandstone, or Trout Creek Sandstone Members, that has a specific storage of 9×10^{-7} per foot would have a storage coefficient of 9×10^{-5}. Storage coefficient in an unconfined aquifer in either unit would be about 1×10^{-1}.

Three storage-coefficient values obtained from pumping-well aquifer tests in the basal Williams Fork aquifer ranged from 2×10^{-4} to 1×10^{-3}. The accuracy of such tests generally are poor, but results indicate confined conditions exist in this aquifer.

GROUND-WATER MOVEMENT

Ground-water movement occurs as a result of hydraulic-head differences in an aquifer. The head in an aquifer at a well is calculated from water-level-measurement data and normally is expressed in terms of the altitude of the standing water level in the well. Head determinations at many different sites define the altitude and areal distribution of head in the aquifer (a potentiometric surface). Thus, the potentiometric surface indicates the altitude and distribution of the standing water level in wells. Ground water moves from points of higher head (near areas of ground-water recharge) to points of lower head (near areas of ground-water discharge) in a direction that generally is perpendicular to the equipotential lines on a potentiometric-surface map. Potentiometric-surface data and aquifer-characteristics data may be used to calculate the rates and distribution of recharge and discharge (water budget) in the aquifer. Detailed water-budget calculations normally are performed on a digital computer because of their number and complexity.

Ground-Water Recharge

Climate, vegetation, and geology have a direct effect on ground-water recharge. Because potential evaporation exceeds mean annual precipitation in the study area, most infiltration occurs only during snowmelt or intermittent periods of intense rainfall. Part of the water entering the soil is consumed by vegetation and lost to the atmosphere through transpiration. This process (evapotranspiration) is enhanced on south-facing slopes where greater insolation produces maximum evaporation and transpiration. The lower angle of incidence on north-facing slopes produces less evapotranspiration and increases the potential for ground-water recharge. Most recharge in the study area occurs in the spring at the higher altitude margins of the area when snowmelt eventually saturates the ground and enables deep percolation. Some recharge may result from thunderstorms in the summer; however, most of this water is lost to evaportanspiration and little can infiltrate to depth.

The ability of water to percolate to depth and recharge the bedrock aquifer also is controlled by the lithology of the soil and the underlying bedrock formations. Clayey soils or shaley bedrock commonly are of very low permeability and will retard water movement. By contrast, sandy soil or sandstone outcrops or subcrops are relatively permeable and may allow water movement to depth. Aquifer recharge zones in the study area are defined by the outcrop or subcrops of permeable bedrock units within the Williams Fork Formation and the underlying Trout Creek Sandstone Member. In the eastern part of the area, the middle member of the Williams Fork Formation is shale and is not considered to be a recharge zone. The extensive outcrops of the Lewis Shale and outcrops of the Iles Formation shale underlying the Trout Creek Sandstone Member likewise are not considered recharge areas.

The rate of recharge can be estimated from results of previous studies. Watershed modeling techniques were used by Weeks and others (1974) to define a relation between the rates of precipitation and ground-water recharge in the Piceance Basin (a mountainous area 60 mi southwest of Craig). The relation between precipitation and ground-water recharge for the Piceance Basin (fig. 32) initially was defined by a model that uses precipitation, solar insolation, and temperature data in calculating surface runoff and deep percolation (recharge) in a watershed of varied slope, aspect, vegetative cover, and soil type. Subsequent modeling of the ground-water flow system in the Piceance Basin indicated that the relation correctly defined the rate and distribution of recharge needed to properly simulate the geohydrology of the aquifers. Similar surface-water modeling procedures were used by Parker and Norris (1989) in the Foidel Creek watershed. The precipitation-recharge results from Foidel Creek are less extensive than those from the Piceance Basin, but indicate that the relation defined in the Piceance Basin also applies to the Twentymile Park study area (fig. 32).

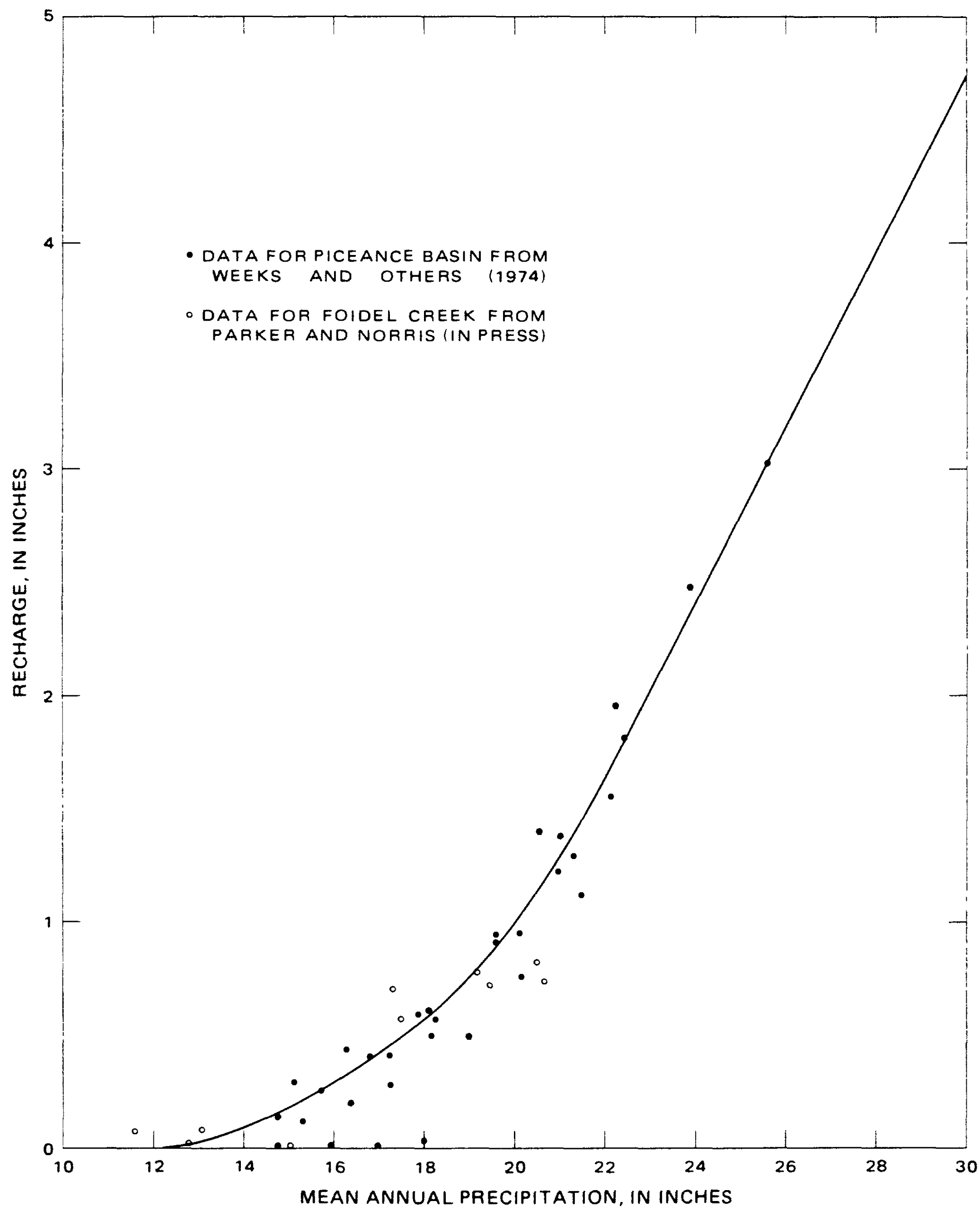


Figure 32.--Relation between mean annual precipitation and ground-water recharge.

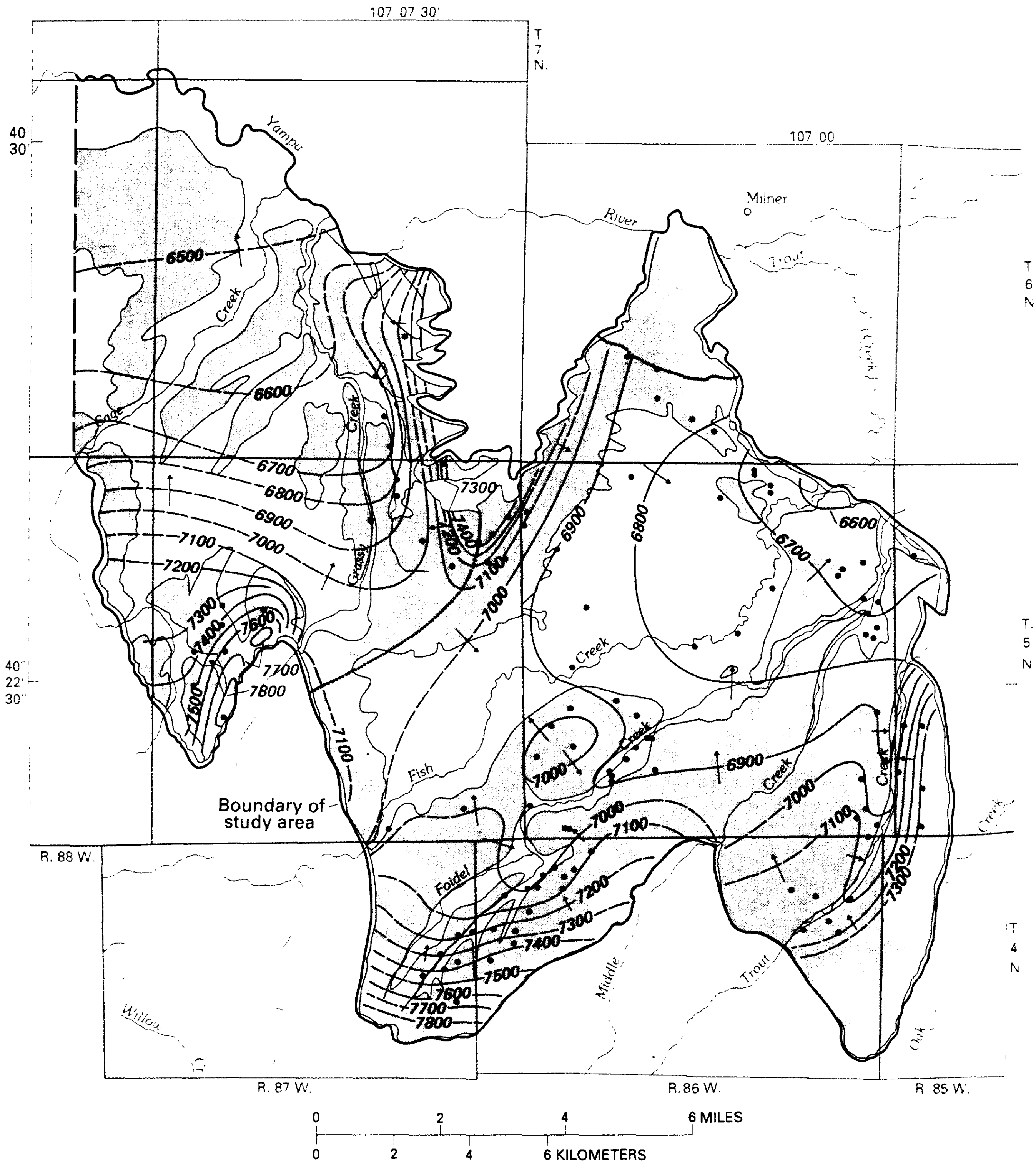
Mean annual recharge to the bedrock aquifers in the eastern part of the study area was estimated by converting the precipitation rates shown on the isohyetal map (fig. 19) to potential recharge rates by using the relation shown in figure 32. The potential recharge rates then were multiplied by the size of the associated segment of the recharge zone to obtain distributed recharge. This technique produced larger estimates of recharge in areas that contain large recharge zones, and effects of altitude and aspect of the recharge zone were incorporated because these two factors are used in producing the isohyetal map. Mean annual recharge to the bedrock aquifers was 2.8 ft^3/s, or about 0.31 in/yr, in the 123 mi^2 eastern study area. Subsequent ground-water modeling in the area indicated that this rate of recharge is compatible with the known hydrology of the bedrock aquifers.

Potentiometric Surface

The potentiometric surfaces in the eastern part of the study are defined by water-level measurements in wells completed in the basal Williams Fork aquifer and the Twentymile aquifer. Too few data are available to define the potentiometric surface in the Trout Creek aquifer. The potentiometric-surface maps (figs. 33 and 34) are based on about 120 water-level measurements selected from an original group of more than 2,500 measurements. Most of the original measurements were made by mining company personnel and were released as part of public documents submitted with mine permit applications. Other measurements were made by U.S. Geological Survey and other Federal or State agency personnel. The maps are constructed to represent predevelopment (or steady state) water-level conditions in the two aquifers. The water-level measurements used on these maps were chosen to ensure that (1) they represent heads that are little affected by drawdown from mines or discharging wells, and (2) the measurements were obtained from wells completed only in one aquifer. These requirements eliminated many measurements from consideration even though heads in most of the area still are near predevelopment levels.

Seasonal changes in depth to water are most pronounced near the aquifer outcrops where recharge from spring snowmelt may cause 10 to 30 ft of water-level rise in the shallow aquifers. Seasonal changes in water level in the more deeply buried aquifers generally are less than 5 ft/yr but may exceed 50 ft/yr near a few pumping wells.

Heads in the basal Williams Fork aquifer are above land surface in much of Twentymile Park and along low-lying areas in most stream valleys (fig. 33). Wells completed in the basal Williams Fork aquifer in these areas can flow at land surface, although most wells are shut in to prevent loss of water from the aquifer. The potentiometric surface of the basal Williams Fork aquifer (fig. 33) ranges in altitude from more than 7,200 ft near recharge areas along parts of the aquifer margin to less than 6,500 ft near the two discharge areas near Hayden and the downstream reach of Fish Creek. The sinuous shapes of the potentiometric contours near Fish Creek, Foidel Creek, Grassy Creek, and Trout Creek are the result of ground-water recharge from, or discharge to, the streams. In some areas, the effects of recharge or discharge extend through the middle confining layer or through both the Twentymile aquifer and the middle confining layer. Heads in the basal Williams Fork aquifer generally were near steady-state conditions in 1986 except near mined areas or near



EXPLANATION

HEAD IN THE BASAL WILLIAMS FORK AQUIFER ABOVE LAND SURFACE

HEAD IN THE BASAL WILLIAMS FORK AQUIFER BELOW LAND SURFACE

BOUNDARY OF BASAL WILLIAMS FORK AQUIFER—Dashed where approximately located

7000 POTENTIOMETRIC CONTOUR—Shows altitude of standing water levels in wells. Dashed where approximately located. Arrow indicates general direction of ground-water movement. Contour interval 100 feet. Datum is sea level

GROUND-WATER DIVIDE

DATA POINT

Figure 33.--Approximate steady-state potentiometric surface of the basal Williams Fork aquifer in the eastern part of the study area.

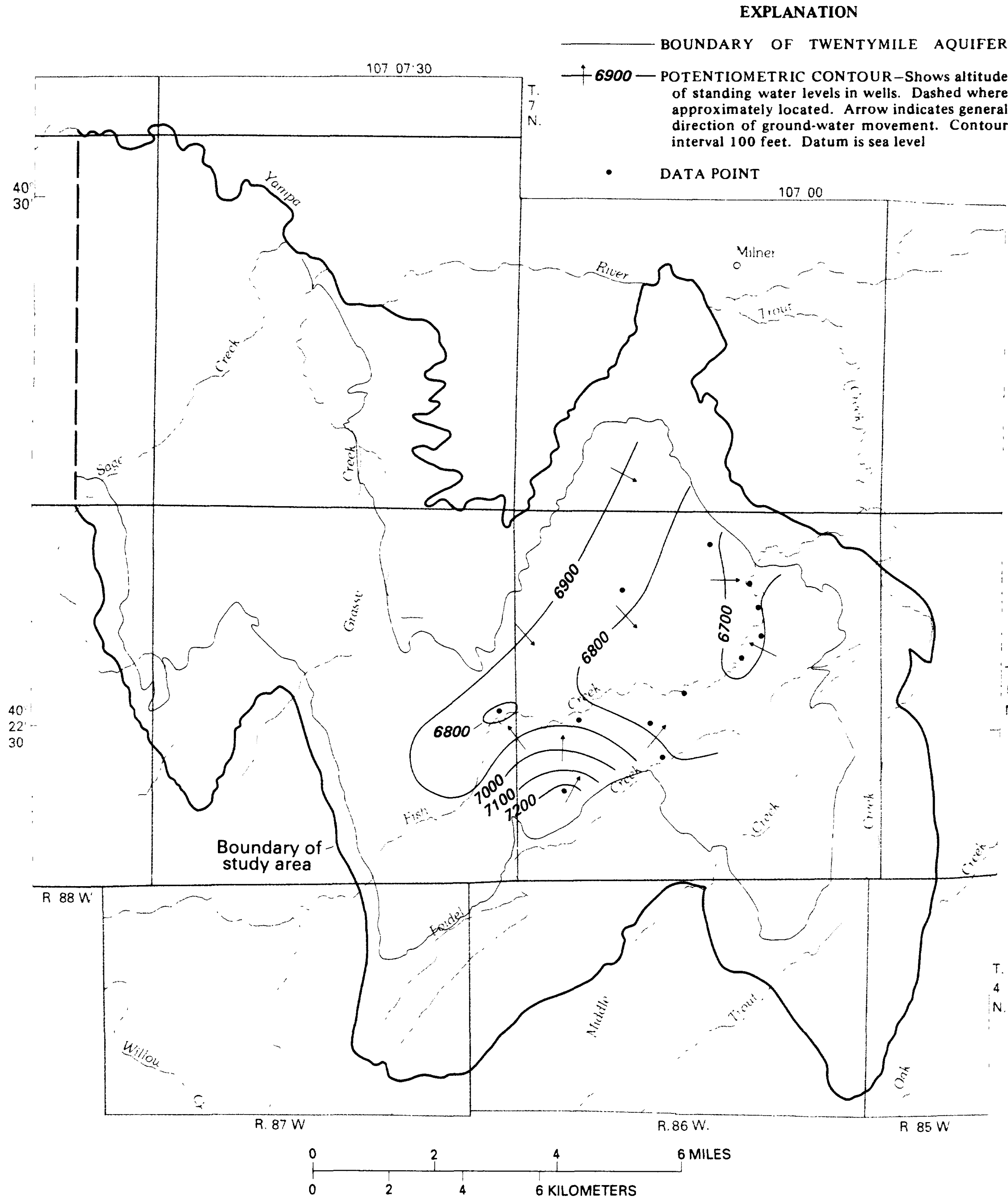


Figure 34.--Approximate steady-state potentiometric surface of the Twentymile aquifer in the eastern part of the study area.

uncapped flowing wells. In some instances, mining has removed part of the aquifer or replaced it with spoils material, either or both of which could disrupt steady-state conditions near the mine.

Ground-water divides in the potentiometric surface of the basal Williams Fork aquifer located near Grassy Gap and north of Twentymile Park form hydrologic boundaries in the flow system because ground water cannot move across the divides. As long as the divides remain in their present position, polluted ground water on one side of the divide can have no effect on ground-water quality on the opposite side of the divide.

The potentiometric surface of the Twentymile aquifer (fig. 34) is less well defined than that of the basal Williams Fork aquifer (fig. 33), for which more data are available. Heads in the Twentymile aquifer near the margin of the aquifer generally are higher than those of the underlying basal Williams Fork aquifer. Near Fish Creek, heads in the Twentymile aquifer generally are lower than those of the underlying basal Williams Fork aquifer. These head relations indicate that the potential exists for water to move from the Twentymile aquifer to the basal Williams Fork aquifer near the margins of the Twentymile aquifer and from the basal Williams Fork aquifer to the Twentymile aquifer near Fish Creek. It is likely that heads in the Twentymile aquifer also are higher near the margins of the aquifer and lower along the valleys of Fish, Foidel, Grassy, and Middle Creeks. Similar recharge-discharge conditions may exist in the Twentymile and basal Williams Fork aquifers. Recharge generally occurs in the outcrop areas of the aquifers and discharge occurs along the valleys of the principal streams draining the area.

Computer-simulation techniques were used to provide additional definition and corroboration of the hydrologic system as conceptualized for the eastern part of the study area. A multilayer model for simulation of quasi-three-dimensional flow (McDonald and Harbaugh, 1984) was constructed to simulate steady-state ground-water flow through the three aquifers. Model parameters, such as hydraulic conductivity and thickness of the aquifer and confining layers, were defined on the basis of the preceding data. Precipitation recharge through outcrops was defined by the previous estimates of distributed recharge. The effects of perennial streamflow and evapotranspiration also were simulated. A 2,000-ft interval grid consisting of 51 rows, 30 columns, and 3 layers was used to discretize the system; this grid also defines the scale at which hydrologic data were defined for use in the model. Model construction procedures are discussed in greater detail in the "Supplemental Information" section at the back of this report.

The model was calibrated to ensure its accuracy by comparing model-calculated heads and rates of discharge with measured values. An acceptable level of calibration was achieved after minor refinements were made to the model-input data. The resulting model-calculated potentiometric surface maps (figs. 35 and 36) are in good agreement with the maps based on measurements (figs. 33 and 34). The mean differences between calculated and measured values of head at 54 points in the model area was 9 ft in the Twentymile aquifer, 16 ft in the basal Williams Fork aquifer, and 20 ft in the Trout Creek aquifer. The differences were approximately randomly distributed over the model area. The model-calculated maps provide more complete definition of the potentiometric surfaces in the Twentymile and basal Williams Fork

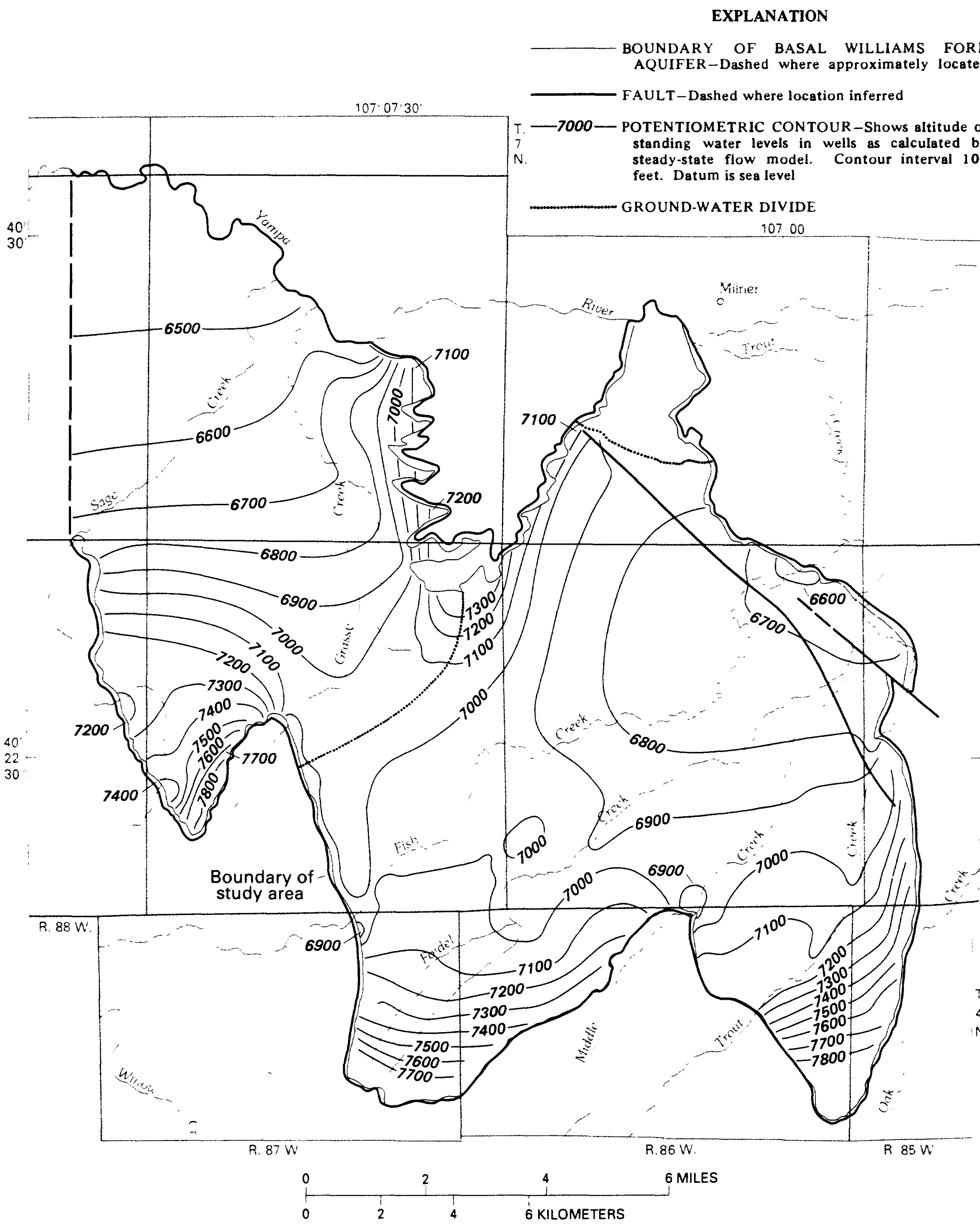


Figure 35.--Model-calculated steady-state potentiometric surface of the basal Williams Fork aquifer in the eastern part of the study area.

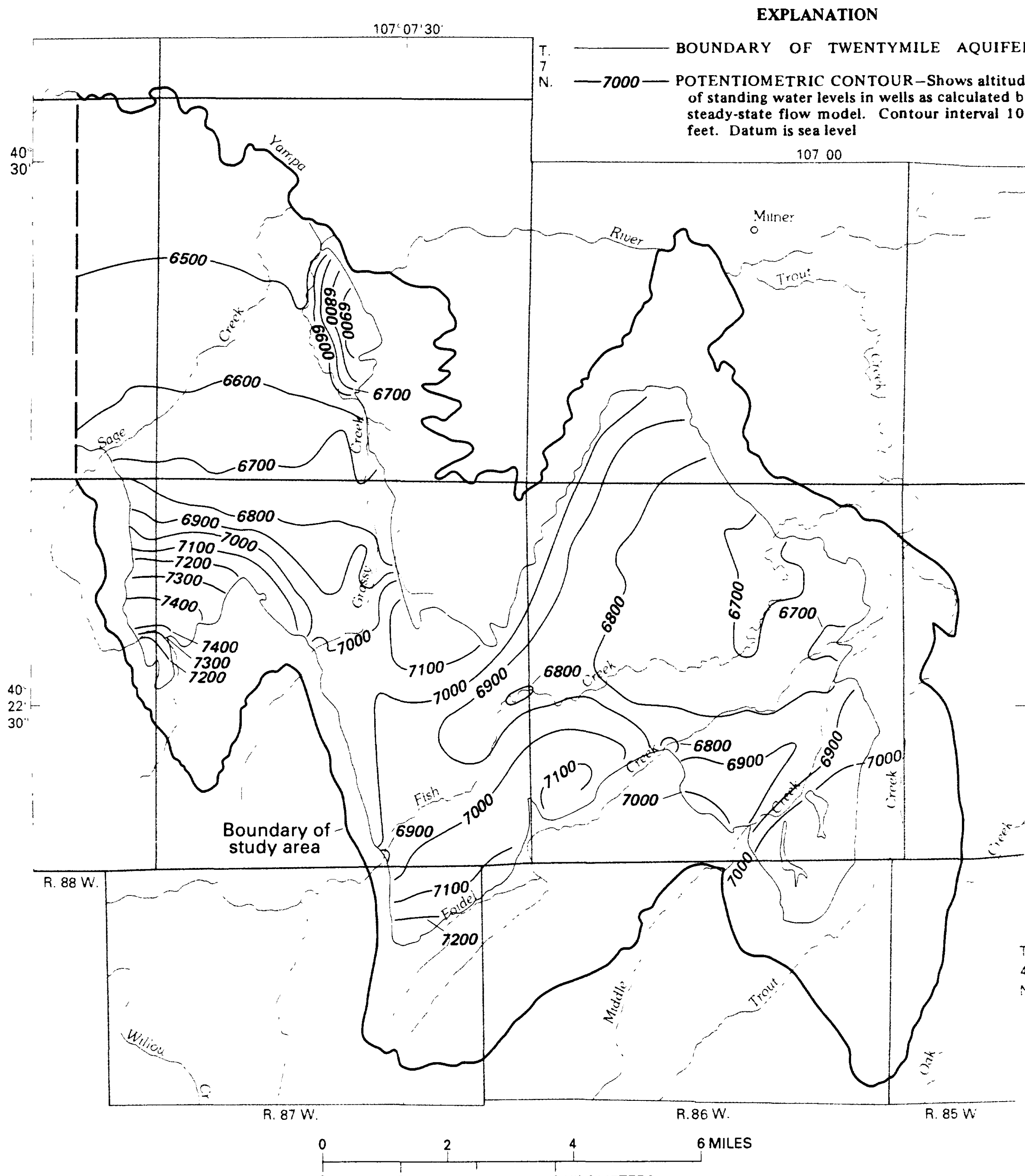


Figure 36.--Model-calculated steady-state potentiometric surface of the Twentymile aquifer in the eastern part of the study area.

aquifers. Model results also indicate that the potentiometric surface in the Trout Creek aquifer essentially is identical to that of the basal Williams Fork aquifer. Model simulations that seem to more closely match the observed heads could be achieved by further changing model input in selected areas, but independent information is not available to justify such changes. As a result, the model was not forced to fit preconceived notions of the flow system.

Most of the water-level measurements made in the western part of the study area were made in uncased drill holes. As a result, the data do not define the potentiometric surface in a single aquifer. Instead, the water-level measurements represent a composite head that occurs in the water-yielding materials penetrated by the individual well. The approximate altitude of these composite heads is indicated in figure 37. The elevation of the heads generally conforms to topography; higher heads occur in the higher altitude areas to the south; lower heads occur in the lower altitude areas to the north. Heads are higher in the deeper aquifers because their recharge areas are at a higher altitude.

Flow Direction

The direction of ground-water flow in the eastern part of the study area is relatively well defined by data and is further corroborated by simulation results. Water moves from recharge areas along the elevated outcrops at the margin of the aquifers toward Twentymile Park and Hayden. A similar pattern of movement is indicated for the basal Williams Fork and Twentymile aquifers. In the higher outcrops of the Twentymile aquifer, the potential exists for interaquifer ground-water movement from the Twentymile aquifer to the basal Williams Fork aquifer. At lower altitudes, the potential is reversed and water may move from the basal Williams Fork aquifer upward to the Twentymile aquifer. In areas where heads are above land surface, water also may discharge to the surface, where it may be lost to evapotranspiration or flow into alluvial aquifers or streams.

The aquifers south and east of the ground-water divide at Grassy Gap form a closed basin because ground-water underflow into and out of the area is insignificant. Ground water in this area is derived from local recharge and moves through the area to discharge at the surface in Twentymile Park. To the northwest of the ground-water divide at Grassy Gap, ground water moves from the outcrop recharge areas to depth along the Hayden syncline. Water may discharge either by vertical leakage into the Yampa River and its alluvium or as underflow into the larger flow system associated with the Sand Wash basin to the north of the Yampa River.

Faulted areas occur near Eckman Park, on the west flank of the Tow Creek anticline, and near the eastern margin of Twentymile Park. Faults near Eckman Park and the Tow Creek anticline are parallel or subparallel to topographic gradients and the general direction of ground-water movement. If sufficient displacement has occurred on these faults, they may form barriers to ground-water movement across the fault plane. Such faults could have little effect on water movement parallel to fault planes. Thus, it is difficult to determine the effects of faulting on ground-water movement in these two faulted areas.

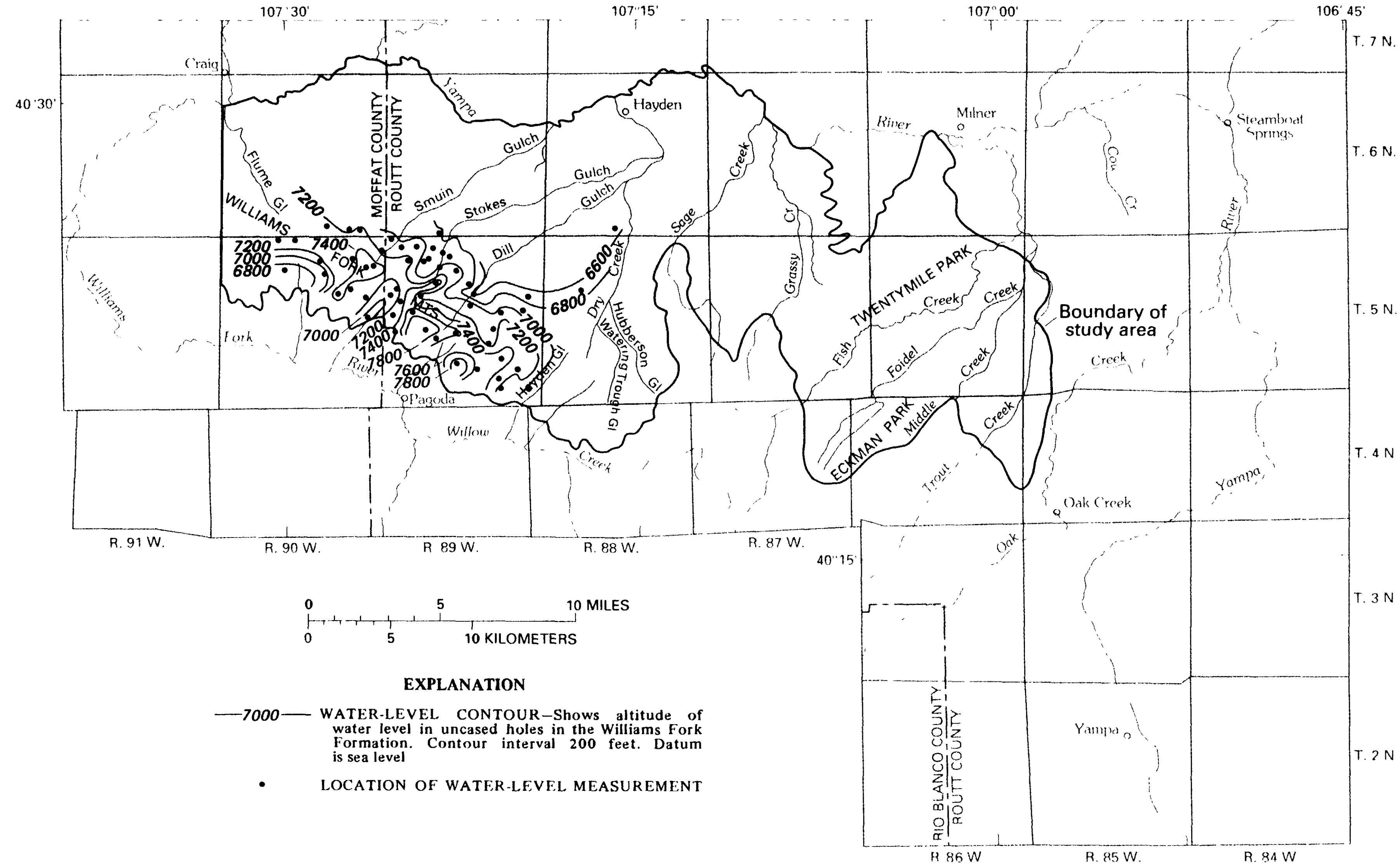


Figure 37.--Water-level altitude in the Williams Fork Formation in the western part of the study area.

Faulting is more extensive along the eastern margin of Twentymile Park. Subparallel fracturing associated with this fault zone has increased hydraulic conductivity parallel to the fault; water moves along the fault more readily than it moves across the fault. This flow pattern causes part of the elongated shape of the 6,700-ft potentiometric contour shown in figure 33.

Differences in hydraulic head in the three aquifers cause water to move vertically across the two confining layers that separate the aquifers. Heads that are above land surface in the uppermost aquifer also may cause upward movement of water across the upper confining layer (Lewis Shale). Although the volume of this interaquifer leakage is very small, it is an important component in the water budget of the area. The rate of vertical ground-water movement also is very small. The time required for a particle of water to move vertically across a confining layer was estimated from Darcy's law using the equation:

$$t = \frac{\Phi L^2}{K \Delta h}, \tag{2}$$

where

t = travel time;
Φ = porosity of the confining layer;
L = thickness of the confining layer;
K = hydraulic conductivity of the confining layer; and
Δh = difference in hydraulic head between the aquifers immediately overlying and underlying the confining layer.

The parameters L, K, and Δh were defined by the corresponding model input data or model-calculated head. Porosity was assumed to equal 13 percent. As indicated in table 8, the mean traveltime for a particle of water to move vertically across the confining layers under steady-state conditions is on the order of thousands to tens of thousands of years. Thus, under steady-state head conditions, the interaquifer movement of ground water requires very long traveltimes and thus is not an important consideration in predicting shorter term movement of contaminants. Traveltimes across the lower confining layer are long, primarily because only small steady-state head differences are present across the confining unit. Under transient conditions, as in response to mine dewatering, larger head differences could develop and the resulting traveltimes could be shortened.

The direction of ground-water flow in the western part of the study area is more difficult to determine because of the lack of potentiometric-surface data for individual aquifers. It is probable that ground water moves in generally northeasterly and southwesterly directions from a ground-water divide that approximately coincides with the topographic divide of the Williams Fork Mountains. Southwestward-flowing ground water moves down the cuesta backslopes toward the Williams Fork. This flow is against structural dip and oblique to the strike of the aquifers. Flow paths likely are short because most water discharges at local springs and seeps. Most of the ground water in the western part of the study area flows to the north or northeast along paths subparallel to the dip of the regional structure. Recharge occurs in the highland outcrops along the Williams Fork Mountains. Down-dip movement carries the water to greater depth in the Williams Fork Formation and under the Lewis Shale. Discharge occurs by upward leakage or by underflow. Water discharged by upward leakage is ultimately lost to evapotranspiration at the land surface or is tributary to streams or alluvial aquifers. Water discharged by underflow moves out of the study area and contributes to the larger regional ground-water flow system in the Sand Wash basin.

Table 8.--*Mean traveltime for steady-state flow of ground water across confining layers in the eastern part of the study area*

[NA, not applicable]

	Traveltime, in years, within surface drainage areas					
	Sage Creek	Grassy Creek	Fish Creek	Foidel Creek	Middle Creek	Trout Creek
Mean upward traveltime for water movement through confining layer separating Twentymile and basal Williams Fork aquifers.	30,300	4,900	11,000	6,300	2,900	NA
Mean downward traveltime for water movement through confining layer separating Twentymile and basal Williams Fork aquifers.	1,400	2,700	16,500	6,600	2,800	2,900
Mean upward traveltime for water movement through confining layer separating basal Williams Fork and Trout Creek aquifers.	98,400	32,700	63,000	28,700	44,300	69,900
Mean downward traveltime for water movement through confining layer separating basal Williams Fork and Trout Creek aquifers.	95,500	36,400	76,200	88,400	78,000	45,100

Lateral ground-water velocities in the eastern part of the area, computed during model simulations, range from less than 1.0 ft/yr to more than 30 ft/yr in the basal Williams Fork aquifer and from less than 0.1 ft/yr to more than 3.0 ft/yr in the Twentymile aquifer. The distribution of lateral velocities shown in figures 38 and 39 indicate that larger velocities generally are located near the margins of the aquifer and smaller velocities are prevalent in the central parts of the aquifers. The combined effects of relatively large hydraulic conductivity and potentiometric gradient result in a larger ground-water velocity along the fault area near the eastern margin of the aquifers. Other faults or interconnected fracture systems could have similar but less pronounced effects on local ground-water velocities. Lack of data prevents individual simulation of any such local features. However, the larger scale effects of local faults and fractures are incorporated in the model through use of spatially distributed hydraulic conductivity, which is based in part on results of aquifer tests in potentially fractured aquifers. Although ground-water velocities cannot be precisely determined because of this lack of data and associated uncertainties in the model parameter values, the magnitude of velocities shown in figures 38 and 39 are significant in that very slow rates of ground-water flow are indicated. A contaminant that enters the bedrock aquifer will not move rapidly to other parts of the area and could remain virtually immobile at some locations.

WATER BUDGET

The ground-water flow model was used to estimate the steady-state water budget for the eastern part of the study area. The simulated water budget (table 9) indicates that total recharge to, or discharge from, the bedrock aquifers is only about 2.8 ft^3/s. This small rate of flow is consistent with the small hydraulic conductivity and small well yields observed in formations that are classified as marginal aquifers in this study. Recharge and discharge for each aquifer by major surface drainage area is listed in table 9. For example, the model calculated that the Twentymile aquifer receives 0.0946 ft^3/s of precipitation recharge from that part of the Sage Creek drainage area that overlies the aquifer; the aquifer loses 0.0777 ft^3/s of discharge to evapotranspiration in the same area. Recharge may come from percolation of water in streams and ponds or from deep infiltration of precipitation. Discharge may be to evapotranspiration or to streamflow and alluvial aquifers. Estimated total recharge to the basal Williams Fork aquifer is about 1.4 ft^3/s, recharge to the Twentymile aquifer is similar, but recharge is only about 0.02 ft^3/s in the Trout Creek aquifer. Recharge and discharge to the Trout Creek aquifer is limited by the small transmissivity and very limited outcrop area of the aquifer. Vertical leakage (the difference between total inflow and total outflow through the lateral boundaries of the aquifer) is the rate of flow through the confining layers that separate each aquifer. The Trout Creek aquifer receives about 75 percent of its recharge as leakage from the overlying basal Williams Fork aquifer and discharges about 90 percent of inflow into the basal Williams Fork aquifer in other areas.

The accuracy of the simulated water budget is affected by the size of the grid interval used in modeling, by the accuracy of the model parameters, and by the extent of the model calibration. The 2,000-ft grid interval used in this model provides a resolution of 930 nodes in the Trout Creek aquifer,

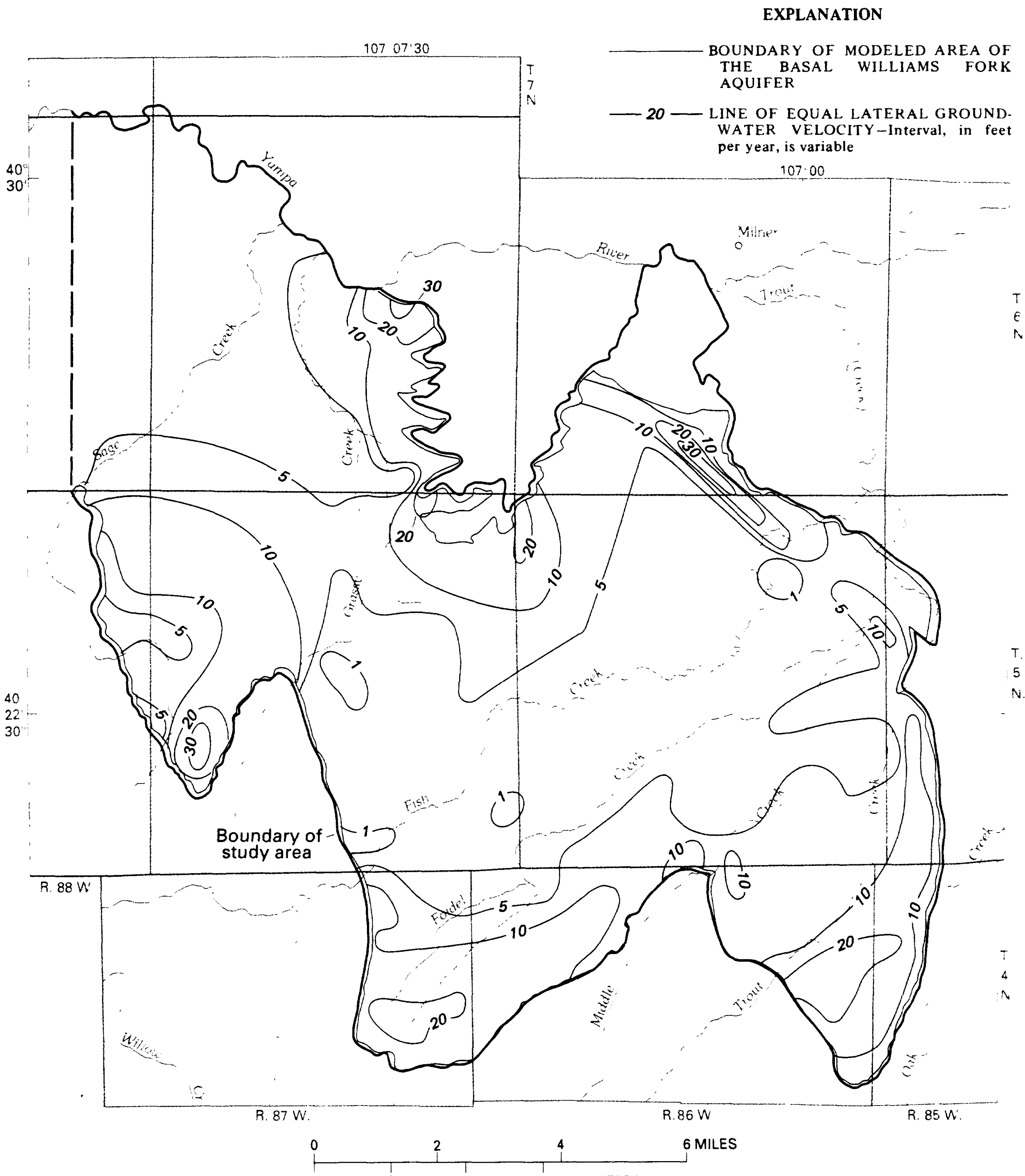


Figure 38.--Magnitude of lateral ground-water velocities in the basal Williams Fork aquifer in the eastern part of the study area.

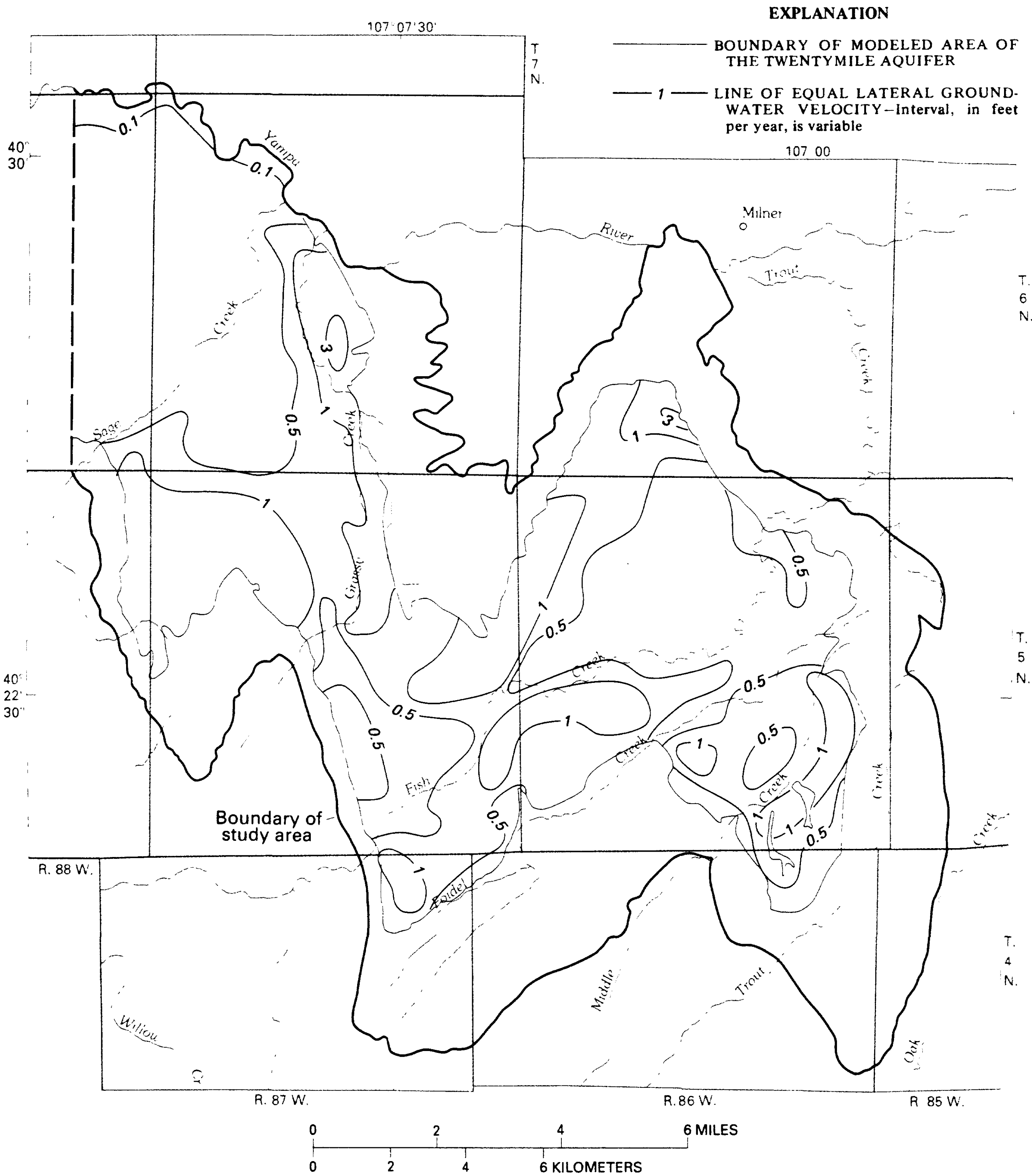


Figure 39.--Magnitude of lateral ground-water velocities in the Twentymile aquifer in the eastern part of the study area.

Table 9.--*Simulated steady-state water budget for aquifers in the eastern part of the study area*

[ft³/s, cubic feet per second]

Component	Layer[1]	Flow rate (ft³/s) in specified drainage area[2]						
		Sage Creek	Grassy Creek	Fish Creek	Foidel Creek	Middle Creek	Trout Creek	Total
				Ground-water recharge				
Precipitation	1	0.0946	0.6209	0.3389	0.1101	0.1366	0.0103	1.3114
	2	0.1869	0.4534	0.1506	0.2934	0.1393	0.2249	1.4485
	3	0.0016	0.0145	0.0040	0.0007	0.0015	0.0012	0.0235
Subtotal		0.2831	1.0888	0.4935	0.4042	0.2774	0.2364	2.7834
Streamflow	1	0	0	0	0	0	0	0
	2	0	0	0	0	0	0.0023	0.0023
	3	0	0	0	0	0	0.0002	0.0002
Subtotal		0	0	0	0	0	0.0025	0.0025
Total recharge		0.2831	1.0888	0.4935	0.4042	0.2774	0.2389	2.7859
				Ground-water discharge				
Streamflow	1	0.1144	0.4020	0.1690	0.0553	0.0804	0	0.8211
	2	0.1938	0.0908	0.0863	0.1640	0.1030	0.2112	0.8491
	3	0.0011	0.0025	0.0005	0.0004	0.0031	0.0012	0.0088
Subtotal		0.3093	0.4953	0.2558	0.2197	0.1865	0.2124	1.6790
Evapotrans-piration	1	0.0777	0.2634	0.2894	0.0072	0	0	0.6377
	2	0	0.2021	0.0278	0.2368	0.0024	0	0.4691
	3	0	0	0	0	0	0	0
Subtotal		0.0777	0.4655	0.3172	0.2440	0.0024	0	1.1068
Total discharge		0.3870	0.9608	0.5730	0.4637	0.1889	0.2124	2.7858
				Vertical leakage				
Downward movement of water through confining unit underlying model layer.	1	0.0352	0.2796	0.1764	0.0620	0.0900	0.0105	0.6537
	2	0.0055	0.0187	0.0118	0.0155	0.0066	0.0108	0.0689
	3	0	0	0	0	0	0	0
Subtotal		0.0407	0.2983	0.1882	0.0775	0.0966	0.0213	0.7226
Upward movement of water through confining unit underlying model layer.	1	0.1359	0.3262	0.2850	0.0264	0.0277	0	0.8012
	2	0.0100	0.0228	0.0204	0.0183	0.0024	0.0098	0.0837
	3	0	0	0	0	0	0	0
Subtotal		0.1459	0.3490	0.3054	0.0447	0.0301	0.0098	0.8849

[1]Layer 1, Twentymile aquifer; layer 2, basal Williams Fork aquifer; layer 3, Trout Creek aquifer.

[2]See figure 22 for location of drainage areas.

920 nodes in the basal Williams Fork aquifer, and 530 nodes in the Twentymile aquifer. This large number of nodes provides sufficiently detailed resolution for the purposes of this investigation. The model parameters range in accuracy from the well-defined data on aquifer extent, thickness, and outcrops, to relatively poorly defined data on hydraulic conductivity and recharge and discharge. These data adequately define the hydrologic conditions in the model area, and a good model calibration was achieved.

The model was calibrated by comparing the model-calculated potentiometric surface maps with measured potentiometric surface maps and by comparing model-calculated discharge with measured changes in streamflow. The close agreement between the calculated and measured potentiometric surfaces for each aquifer indicates that the model is a good simulator of the steady-state flow system in the aquifers. The water budget calculated by the model also should be a good estimate of the actual water budget for the area and likely is of better accuracy than water-budget information based on field measurements.

It generally is difficult or impossible to make direct field measurements of most components of a water budget. In many instances, the component to be measured is spatially or temporally variable or is inaccessible for measurement. Gain or loss in streamflow may occur in response to ground-water discharging into, or recharging from, a stream. The long-term average gain or loss in streamflow is difficult to measure because of the short-term effects of storm runoff, interaction with flow in alluvial aquifers, evapotranspiration from phreatophytes, and diversions. Gain or loss in streamflow was measured along 71 reaches of selected streams in July, August, and September 1986 (table 3). Pertinent gain-loss data are summarized in table 10 for purposes of comparison with surface-water gain-or-loss data calculated by the steady-state model. The difficulty in relating instantaneous measurements of streamflow to long-term average streamflow are apparent. However, the measured and model-calculated values of gain or loss are of comparable magnitude, which indicates that the model-calculated water budget likely is a reasonable estimate of the actual steady-state water budget.

GROUND-WATER GEOCHEMISTRY

The chemical composition of ground water is the result of geochemical processes that include dissolution of soluble minerals from the soil and aquifer matrix, chemical reactions and ion exchange reactions between dissolved constituents, and precipitation of minerals. The large number of dissolved constituents in water, and the complex geochemical processes that may affect the concentrations of these constituents, make identification of most geochemical reactions difficult even when adequate data are available. In the western part of the study area, ground-water-quality data are few and are poorly associated with individual aquifers. In the eastern part of the area, the more numerous chemical analyses associated with specific aquifers enable evaluation of some geochemical processes. The prevalence of these geochemical processes in nature and the similarity of geology, hydrology, climate, and topography, between the eastern and western parts of the study area indicate that geochemical processes identified in the eastern part of the area also likely are occurring in the western part of the area, even though data may be lacking.

Table 10.--*Measured and calculated gain or loss in streamflow in the eastern part of the study area*

Stream	Reach (fig. 23)	Date measured	Flow, in cubic feet per second: Measured gain (+) or loss (-) in streamflow	Flow, in cubic feet per second: Model-calculated long-term average gain (+) or loss (-) in streamflow
Grassy Creek	1-3	07-22-86	+0.12	+0.03
		09-15-86	+0.05	
	3-4	07-22-86	+0.01	+0.01
		09-15-86	-0.06	
	4-5	07-22-86	+0.11	+0.06
		09-15-86	-0.02	
	5-6	07-22-86	-0.02	+0.06
		09-15-86	+0.03	
Foidel Creek	1-3	09-17-86	+0.08	+0.02
Middle Creek	1-2	09-17-86	0.0	+0.08
	2-3	09-17-86	+0.02	+0.05
"S" Creek	2-3	09-16-86	+0.02	+0.01
	4-5	08-13-86	+0.02	+0.01
		09-16-86	-0.02	
	5-6	08-13-86	0.0	+0.01
		09-16-86	+0.01	

About 75 water-quality analyses were available in the study area for U.S. Government observation wells completed in the Williams Fork Formation. About half of these samples were collected during 1980-81; the remaining samples were collected during earlier U.S. Geological Survey studies. In addition, data from about 1,000 chemical analyses of ground water are available from mining companies in the area. Results of most of these analyses have been published in various mine permitting or monitoring documents pertaining to the eastern part of the study area. An additional one-time sampling of domestic wells and springs was conducted by the U.S. Geological Survey in 1977 (Brogden and Giles, 1977). However, these domestic well data are of limited usefulness because of the lack of well-construction data and possible mixing of alluvial and bedrock water in the well.

Dominant Water Types and Distributions

The Twentymile aquifer contains water that primarily is a sodium bicarbonate type. Water in this aquifer commonly has dissolved-solids concentrations that range from 300 to 600 mg/L; the larger concentrations occur in the north-central part of Twentymile Park (fig. 40). Hardness averages about 20 mg/L as calcium carbonate, and the water is classified as soft. Sulfate concentrations generally range from 50 to 140 mg/L.

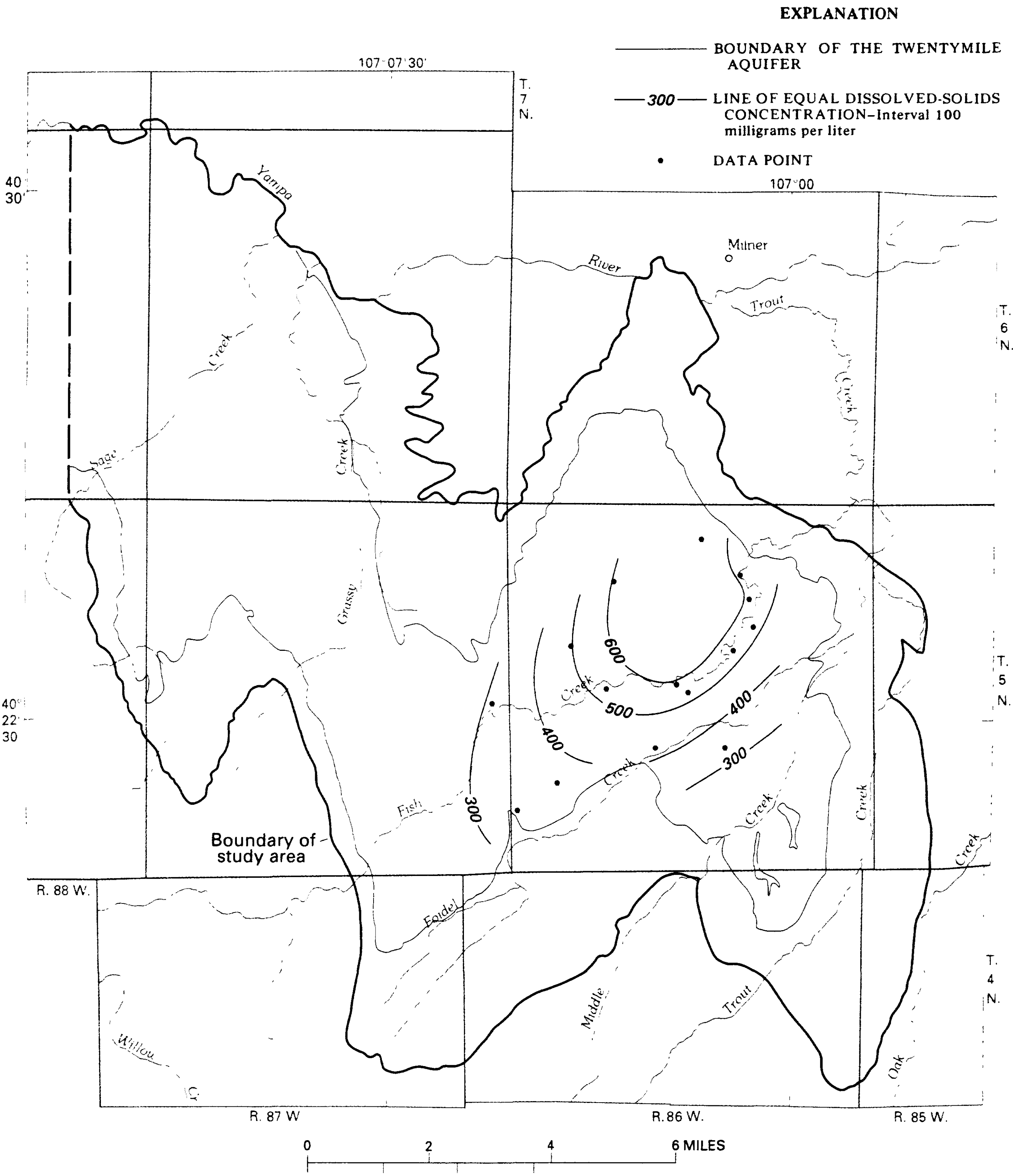


Figure 40.--Dissolved-solids concentrations of water in the Twentymile aquifer in the eastern part of the study area.

Water in the basal Williams Fork aquifer generally is a sodium or calcium bicarbonate type but may be a sulfate type in local areas. Calcium is the predominant cation near recharge areas along the margin of the aquifer. In these areas, calcium plus magnesium concentrations average about 250 mg/L; sodium plus potassium concentrations average about 70 mg/L. As the water moves into deeper parts of the aquifer it becomes a sodium bicarbonate type. Water in the deeper parts of the basal Williams Fork aquifer has calcium plus magnesium concentrations that average about 25 mg/L; sodium plus potassium concentrations average about 280 mg/L. The decrease in calcium plus magnesium concentrations and the concurrent increase in sodium plus potassium concentrations are the result of cation exchange reactions on the clay minerals of marine shales that are interbeded in the aquifer. Cation exchange does not affect water composition substantially until the water has moved about 1 mi into the aquifer (fig. 41). By the time the water has moved about 2 mi into the aquifer, most cation exchange is complete, and the water at greater distances along the flow path retains a relatively uniform sodium-dominant cation composition.

This cation exchange produces a natural softening of the ground water. Near the margins of the basal Williams Fork aquifer, where calcium and magnesium concentrations are large, the water is classified as very hard; the mean hardness is about 960 mg/L as $CaCO_3$. In the central part of the aquifer, where the water has undergone cation exchange, the water is classified as soft to hard; the mean hardness is about 70 mg/L as $CaCO_3$.

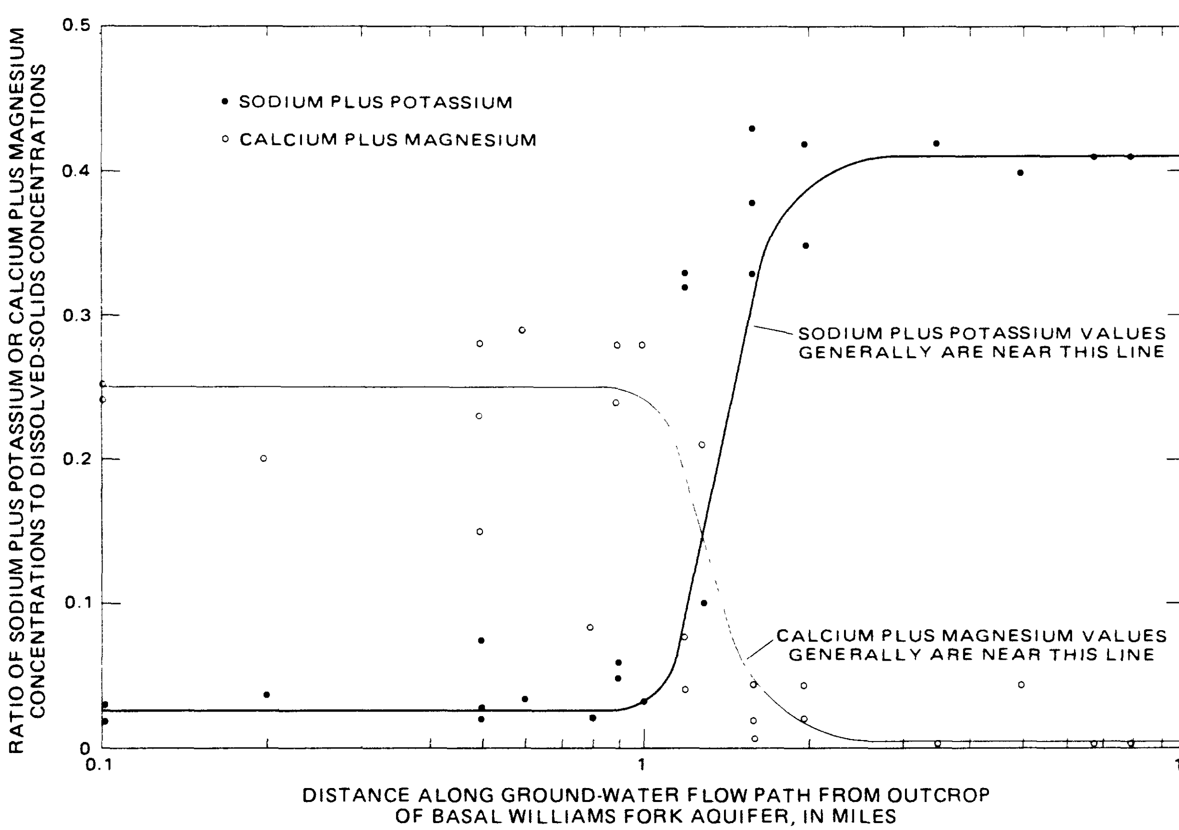


Figure 41.--Change in relative concentrations of cations in the basal Williams Fork aquifer.

In some parts of the basal Williams Fork aquifer, sulfate is the dominant anion in solution. Oxidation of sulfur minerals present in the coal and associated carbonaceous shale and dissolution of gypsum are the likely sources of the dissolved sulfate. Large amounts of sulfate are formed in the oxidizing environment of mine spoils; lesser amounts of sulfate occur in the undisturbed outcrops of the coal-bearing intervals. Ground water containing large natural concentrations of sulfate occurs sporadically along the recharge area near the basin margins. Near areas disturbed by mining, large sulfate concentrations are more prevalent. Sulfate concentrations generally range from 50 to 1,500 mg/L in the mined areas and from 50 to 400 mg/L in the undisturbed areas. The relative concentration of sulfate decreases at greater distance along the ground-water flow path downgradient from mined areas, as shown in figure 42. This decrease likely is caused by a combination of three geochemical processes: (1) Sulfate reduction--the precipitation of sulfate from solution in a reducing environment; (2) dispersion--the mixing and spreading of the water that contains large concentrations of sulfate into the surrounding water that contains small concentrations of sulfate; and (3) limited solute movement--water that contains large concentrations of sulfate has moved only a limited distance away from the mine during the time since mining began.

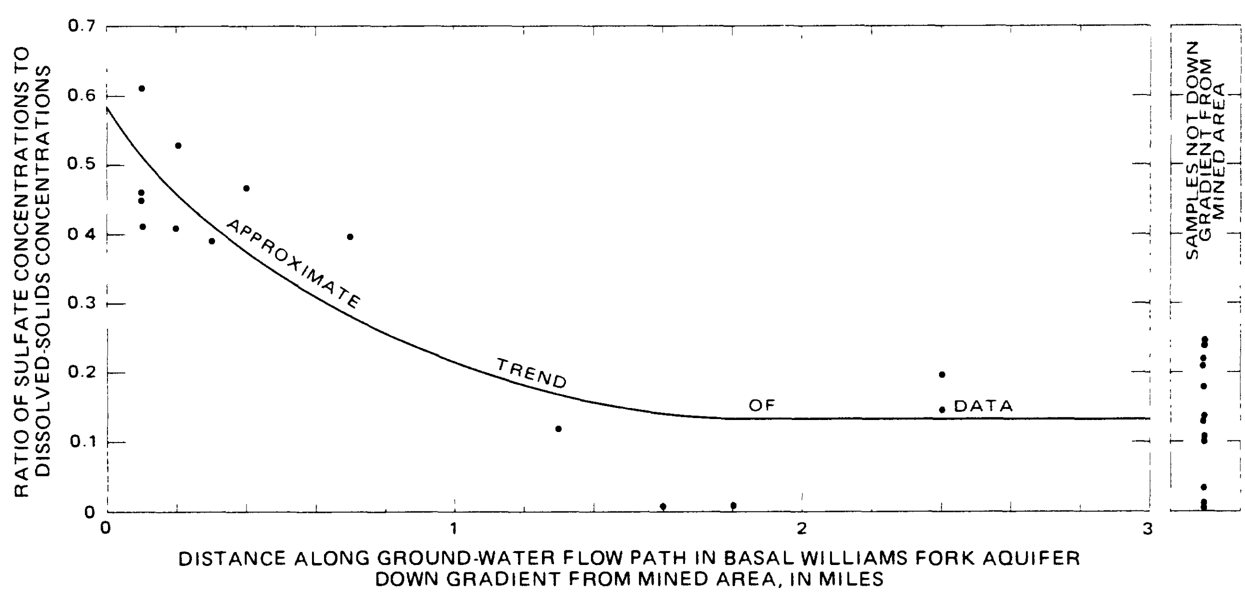


Figure 42.--Change in relative concentration of sulfate in the basal Williams Fork aquifer.

Dissolved-solids concentrations in the basal Williams Fork aquifer generally range from 300 to 1,400 mg/L in those areas where sulfate concentrations are small (fig. 43). Dissolved-solids concentrations larger than 1,000 mg/L are present near the eastern margin of Twentymile Park. In areas where dissolved sulfate concentrations are large (fig. 43), dissolved-solids concentrations may range sporadically from 400 to 3,000 mg/L. The relation between dissolved-solids concentrations and specific conductance also is affected by the anion water type as shown in figure 44. The linear regression line for sulfate water has a correlation coefficient of 0.93 and an equation of the form $Y = 0.805x - 45.8$. The linear regression line for bicarbonate water has a correlation coefficient of 0.95 and an equation of the form $Y = 0.606x + 58.4$.

Geochemical Controls on Cation Concentrations

Calcium, magnesium, and sodium, are the dominant cations in the study area; potassium ions occur in concentrations small enough to be disregarded. Two geochemical processes, carbonate dissolution and ion exchange, have an effect on cation concentrations and distribution.

The first geochemical process, carbonate dissolution, is lithology and surface dependent. Dissolution of calcite and dolomite at low temperatures provide calcium and magnesium cations to the aqueous system. Thin discontinuous limey shales, limestones, and dolomitic limestones present in the fine-grained rocks provide a source of calcium and magnesium.

The solubility of the carbonate minerals is controlled by the pH of the local ground water. Water recharging the aquifer carries oxygen and carbon dioxide gasses into the aquifers. These two gasses tend to decrease the pH of the recharge water and thereby increase the solubility of carbonate minerals. Carbon dioxide forms carbonic acid on dissolution in water:

$$CO_2 + H_2O = H_2CO_3 \ . \qquad (3)$$

Oxidation of sulfide minerals, such as the iron pyrite that is commonly present in the coal or associated carbonaceous shale, forms sulfuric acid:

$$4FeS_2 + 15O_2 + 8H_2O = 2Fe_2O_3 + 8H_2SO_4 \ . \qquad (4)$$

The pH of the water in the recharge areas of the basal Williams Fork aquifer averages about 7.5 and is more acidic than the water in deeper parts of the aquifer where the pH averages about 8.5. The carbonic acid and sulfuric acid produced by reactions 3 and 4 may react with calcite to produce calcium, bicarbonate, and sulfate ions:

$$CaCO_3 + H_2CO_3 \rightarrow Ca^{++} + 2HCO_3^- \ , \qquad (5)$$

$$2CaCO_3 + H_2SO_4 \rightarrow 2Ca^{++} + 2HCO_3^- + SO_4^{--} \ , \qquad (6)$$

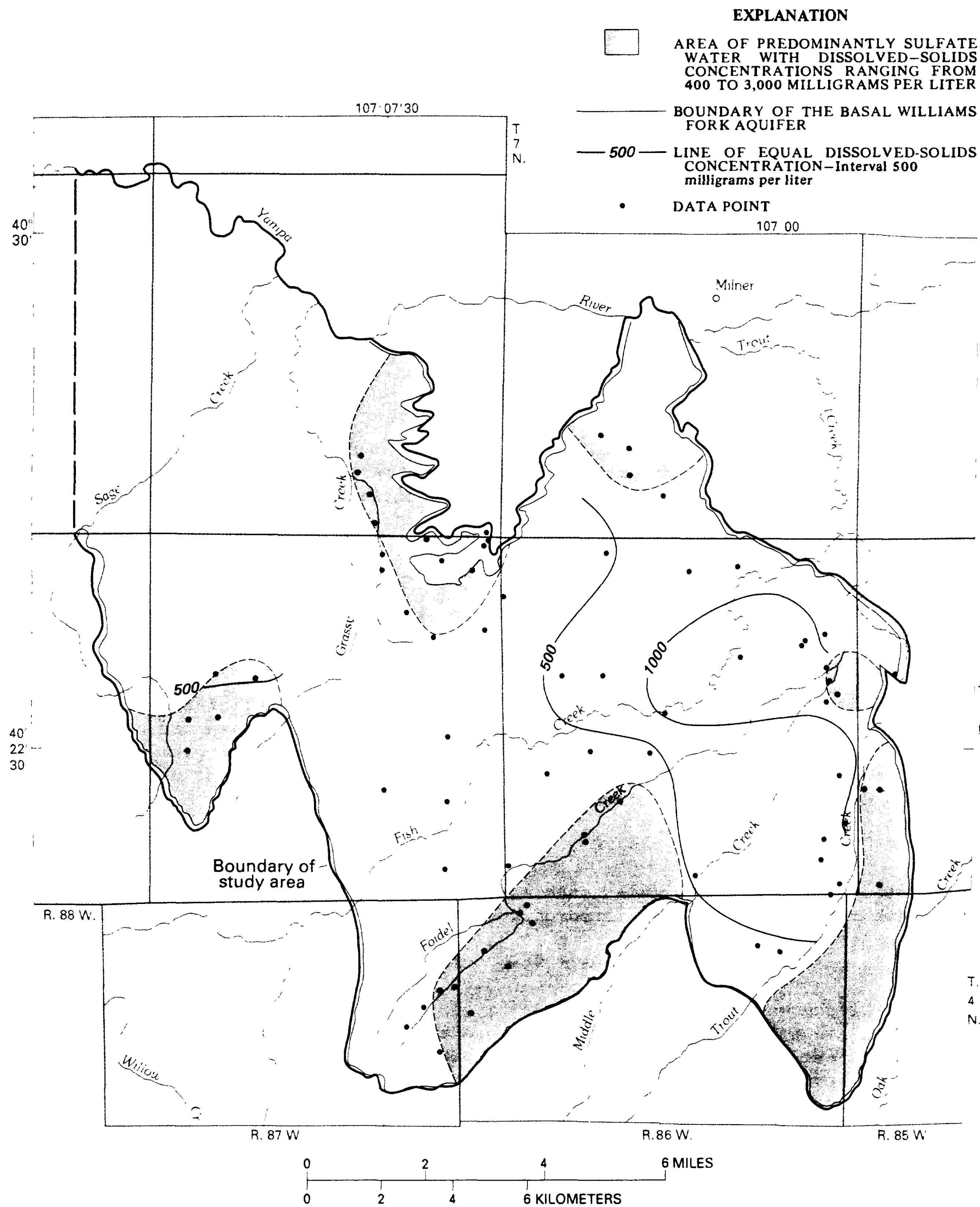


Figure 43.--Dissolved-solids concentrations of water in the basal Williams Fork aquifer in the eastern part of the study area.

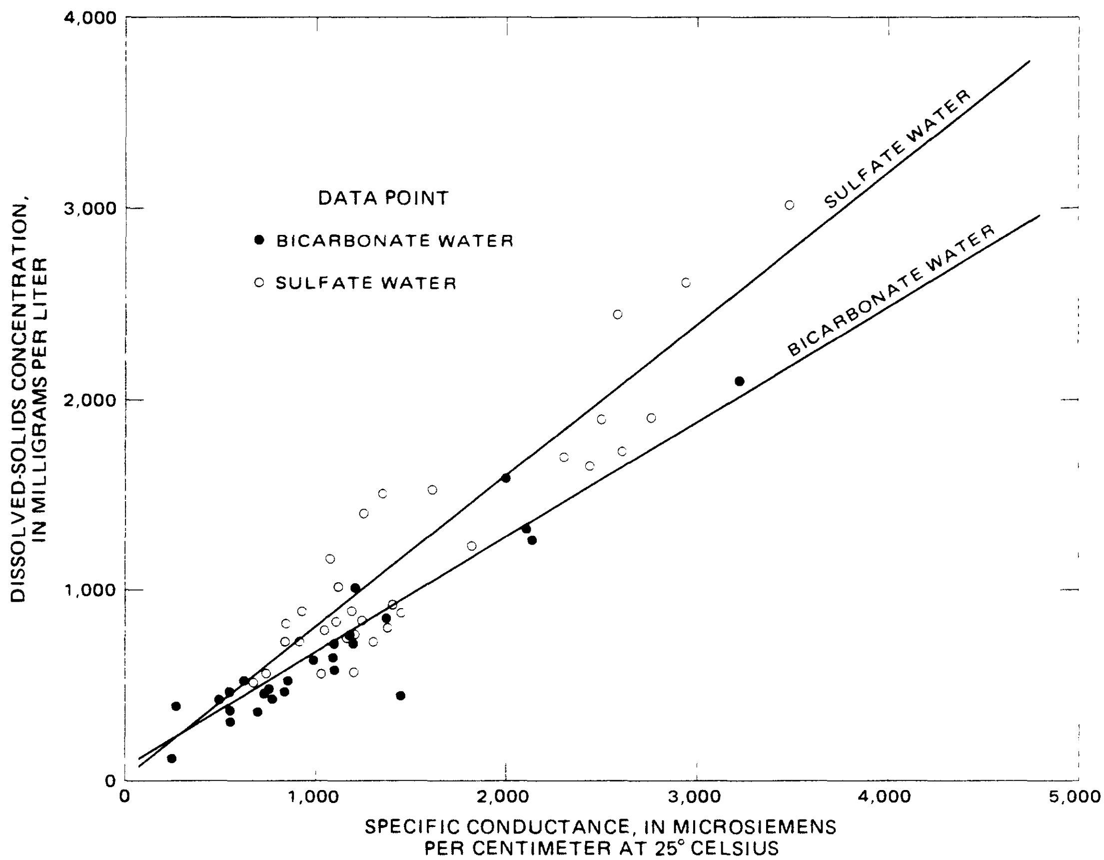


Figure 44.--Relation between specific conductance and dissolved-solids concentrations for sulfate and bicarbonate water in the basal Williams Fork aquifer.

or may react with dolomite to produce calcium, magnesium, bicarbonate, and sulfate ions:

$$CaMg(CO_3)_2 + 2H_2CO_3 \rightarrow Ca^{++} + Mg^{++} + 4HCO_3^- ; \qquad (7)$$

$$CaMg(CO_3)_2 + H_2SO_4 \rightarrow Ca^{++} + Mg^{++} + SO_4^{--} + 2HCO_3^- . \qquad (8)$$

In areas containing marine shales, ion exchange is the predominant geochemical process controlling cation concentration and distribution. The mechanism is an exchange of calcium and magnesium ions in aqueous solution with sodium ions on the clay minerals of the sodium-rich marine shales. This cationic exchange is the principal source of sodium in the ground water. The general equation for monovalent-divalent cation exchange is (Garrels and Christ, 1965):

$$A_2\,X_2 + B^{++} = BX_2 + 2A^+ . \qquad (9)$$

In this example, A represents sodium, and B represents calcium or magnesium. Sodium is continually replaced at the exchange sites of the clay(X_2), until enough sites are filled with divalent cations to establish a chemical equilibrium with ground water. This process accounts for most of the approximately 200 mg/L decrease in calcium plus magnesium concentrations (and corresponding increase in sodium plus potassium concentrations) that occur as water moves from the recharge areas to the central part of the basal Williams Fork aquifer.

Geochemical Controls on Anion Concentrations

The primary anions in the basal Williams Fork aquifer are bicarbonate and sulfate. Carbonate ions also are present in significant concentrations in local areas that have alkaline water of large pH values. Dissolution of carbonate minerals may yield carbonate or bicarbonate ions. At pH greater than 10.5, a shift from bicarbonate ions to carbonate ions may occur:

$$HCO_3^- \rightleftharpoons CO_3^{-2} + H^+ \ . \qquad (10)$$

Dissolved sulfate anions commonly are derived from two sources--dissolution of authigenic gypsum and oxidation of pyrite and marcasite. Reduction rates for sulfur systems often are slow, resulting in nonequilibrium forms of sulfur being present (Hem, 1970); two forms of sulfur, sulfide ions and hydrogen sulfide gas, can be present in the same sample.

Direct dissolution of gypsum may occur as ground water moves slowly through the gypsum-bearing units. However, larger rates of dissolution occur in the weathered zone or in spoils because weathering and crushing create secondary permeability that allows increased ground-water movement through the materials and increased contact of the water with newly exposed soluble minerals such as gypsum.

Oxidation of reduced sulfur, which primarily occurs in pyrite and marcasite in the bedrock organic shales and coals, produces sulfate. As these beds are exposed to oxygenated water, sulfur is oxidized to produce the sulfate ion, as indicated in equations 4 and 6. Pyritic materials are very common in drill samples from the area, and large concentrations of sulfate in wells completed in coalbeds indicate that coal and carbonaceous shale are sources of sulfur.

In a semiarid climate, precipitation may be insufficient to leach all geochemical weathering products out of the soil zone. In areas of fine-grained rocks, production of sulfate by weathering and inflow of sulfate in runoff and precipitation may exceed the rate that the sulfate is removed by runoff and subsurface flow. This can cause large concentrations of sulfate to form near the land surface (Hem, 1970). This process may explain the large sulfate concentrations associated with some wells that are completed in alluvial aquifers and also may account for the alkali deposits present along some poorly drained valley bottoms.

SOLUTE-TRANSPORT SIMULATION

Mathematical models provide a means of simulating processes occuring in hydrologic systems. Models of ground-water flow, for example, can provide information about the water budget, potentiometric surface, and direction of ground-water movement. Solute-transport models commonly incorporate a flow model and thus provide the information typical of a flow model in addition to information about the rate of ground-water movement and the concentration of dissolved chemical constituents. The added simulation capability of a solute-transport model makes it a particularly useful tool for evaluating the effects of mining on the head and water quality in an aquifer.

Selection of a Model Computer Code

Computer programs currently (1988) are available for many different types of solute-transport models that have a wide range of simulation capabilities. Bachmat and others (1980), Science Applications Inc. (1981), and Thomas and others (1982) present evaluations of numerous computer codes for use in ground-water management. Kincaid and others (1982-86) expanded and updated these previous works in order to evaluate the suitability of solute-transport codes for application to subsurface waste-disposal issues associated with coal-fired electric generating plants. Kincaid's work indicated that of the hundreds of codes potentially applicable to such issues, only three were considered suitable for final testing and evaluation. These three codes, available in the public sector, can be used for steady- or transient-state simulations of saturated, single-phase, two-dimensional flow of water through an isothermal, nonhomogeneous, anisotropic porous medium using distributed parameters and varied spatial and temporal boundary conditions. The method of characteristics solute-transport model (Konikow and Bredenhoeft, 1978) code was chosen for use in this study because of: (1) Kincaid's favorable rating of the code with respect to other codes; (2) the extensive history of successful application of the code to real-world solute-transport problems; (3) the continuing support and updating of the code provided by the authors; and (4) the acceptance by the U.S. Geological Survey and the U.S. Environmental Protection Agency of the code as a tool suitable for analysis of ground-water solute-transport problems.

Simulation Procedures

The objectives of the solute-transport modeling in this study are similar to those of the flow modeling in that both models are intended to provide basin-wide evaluations of the geohydrology of the aquifers rather than mine-specific or site-specific evaluations. As a result, solute transport modeling was undertaken using the same grid network used in the flow modeling. This allowed direct incorporation of flow-model data into the solute-transport model without redefining a grid or redigitizing distributed-parameter data. Model evaluation of the water-quality changes in the aquifer can be achieved by simulating the movement of a conservative tracer. Dissolved-solids concentrations commonly are used for this purpose and are better defined by field data in the model area than other chemical constituents.

The use of transient- or steady-flow and transient- or steady-transport simulation procedures are determined in part by the scale of the simulations. In the eastern part of the study area, lateral ground-water velocities range from less than 1 ft/yr to more than 30 ft/yr. Traveltimes for water to move the 2,000-ft distance between node centers of the model grid range from about 10 to 2,000 years. These traveltimes indicate that the solute transport model must compute changes during time periods of at least tens to hundreds of years, rather than short-term changes of a few years, if useful model simulations are to be achieved. Steady-flow, transient-transport simulations are appropriate for such conditions. Simulations of this type are based on the long-term, unvarying flow of ground water. The ground-water velocities produced by the steady flow are used to control the rate and direction of movement of water of differing chemical quality. Thus, the model-computed water quality in the aquifer changes through time (transient transport), even though the heads and rates of flow are invariant (steady flow).

If effects of transient ground-water flow are to be disregarded in the solute-transport model, the water-quality changes produced in the undisturbed aquifer during the transient period must be relatively small. In open-pit and underground mines, an initial period of transient ground-water flow occurs during the several-year interval when the mine is active and is totally or partly dewatered. During this period, ground-water movement is toward the mine, and any poor quality leachate generated in the mine would be unable to move beyond the pit or workings. Water-quality changes in the undisturbed aquifer during this first period of transient flow likely are negligible.

A second period of transient flow occurs once mining is completed, or when the pit or workings begin to flood. The water levels in the mine or spoils will rise until an equilibrium level is reached with heads in the surrounding undisturbed aquifer. Transient-flow conditions cease once approximate equilibrium conditions are achieved. Poor quality leachate in the mine may begin to move beyond the mine during this transient period if the water level in the mine or spoils exceeds the head in the adjacent undisturbed aquifer. This second period of transient flow is relatively brief. Springs have been observed to develop near the low wall of the open-pit, dip-slope mines within a period of a few months to about 3 years following the completion of nearby mining. Water-quality changes in the undisturbed aquifer resulting from this brief period of transient ground-water flow are likely to be insignificant in comparison to the 10 to 100 years of water-quality changes that will be considered in the solute-transport model. Thus, for most simulations, the effects of transient ground-water flow may be disregarded without introducing serious error.

Multiple single-layer models provide an appropriate means of simulating solute transport. Results from the multilayer flow model indicate that downward components of flow exist across confining layers between the principal aquifers near the margins of the basin; upward components of flow exist across confining layers near the central parts of the basin. Traveltime required to move water from one aquifer to another across the intervening confining layers was shown to be on the order of 1,000 to 200,000 years. If a multilayer solute-transport model of the aquifer system were constructed, it would not indicate movement of poor quality water from one aquifer to another within the simulation time period. Both single-layer and multilayer models will correctly simulate the required lateral movement of contaminant in an aquifer, but a single-layer model is less complex, more computationally efficient, and is easier to build and operate than a multilayer model. The Twentymile aquifer and the basal Williams Fork aquifer are of principal concern in mine impact analyses. A single-layer solute-transport model of each aquifer was built. Additional information about the design of these models is contained in the "Supplemental Information" section at the back of this report.

Model Calibration

Large-scale solute-transport models require large-scale historical changes in ground-water quality for use in model calibration. Movement of poor quality water from spoils in several open-pit mines apparently has caused the degradation of ground-water quality at numerous observation wells completed in the basal Williams Fork aquifer. However, virtually all of these wells are located within 1,000 ft of the spoils, a distance too small to provide useful data for calibration of a 2,000-ft grid-interval model. This limited historical movement precludes transport calibration of the solute-transport model. Thus, steady-flow, transient-transport simulations to be made with the model are based on a calibrated steady-state flow model and an uncalibrated transport model.

Transport calibration primarily enables adjustment of model dispersivity and porosity to values that are compatible with other model parameters so that the model-calculated changes in concentration will agree with observed changes in concentration. Dispersivity primarily affects the amount of dispersion, or spreading out, of a zone of poor quality water caused by nonuniform ground-water velocities in the aquifer. Porosity primarily affects the rate of movement of a degraded zone caused by the average ground-water veolocities in the aquifer. Even in fully calibrated models, some uncertainty exists as to the best value for any particular model parameter. The best values for dispersivity and porosity are more uncertain because of the lack of transport calibration. Sensitivity analysis provides a means of determining the relative importance of a parameter value. If the model results are little affected by a large change in a parameter value, the model is said to be insensitive to that parameter. Conversely, if the model results change markedly in response to a small change in a parameter, the model is sensitive to the parameter.

Sensitivity Analyses

The solute-transport model of the basal Williams Fork aquifer was used to simulate contaminant movement by using a range of values for dispersivity and porosity. Concentration profiles calculated by the model at the end of 100-year simulations of contaminant migration away from areas of degraded water quality in spoil aquifers at the Edna mine are shown in figures 45 and 46. Effects of varying transverse and longitudinal dispersivity are shown to produce minimal changes in the concentration profiles (fig. 45, graphs A and B). Porosity values that range from 5 to 15 percent are shown to produce more substantial changes in the concentration profile (fig. 45, graph C). The results of the sensitivity analyses indicate that the model-calculated dissolved-solids concentrations are relatively insensitive to dispersivity but more sensitive to porosity.

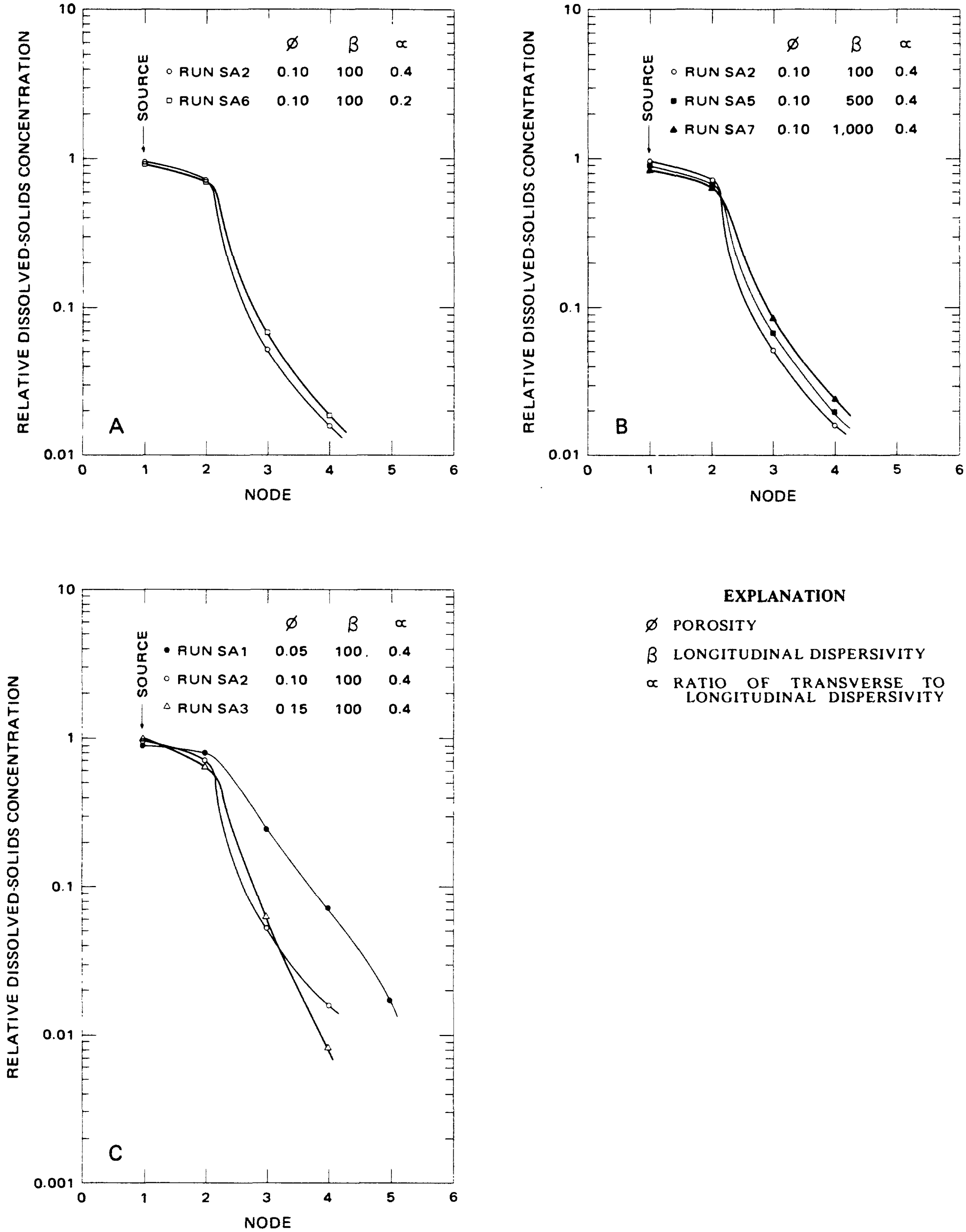


Figure 45.--Sensivity of model-calculated concentration profiles to changes in model parameters.

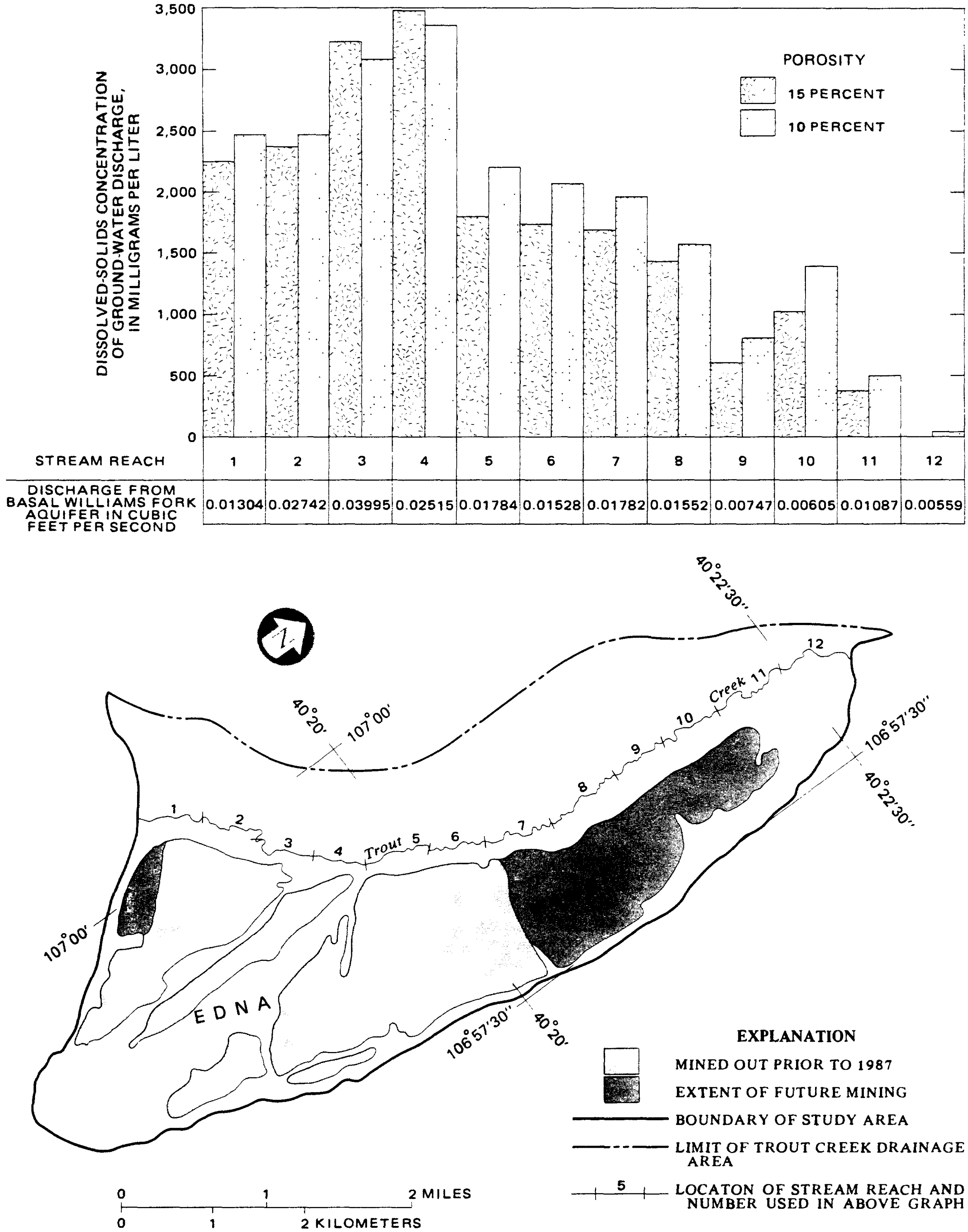


Figure 46.--Changes in simulated concentration of ground-water discharge to Trout Creek.

The dispersivity of the uniform, fine-grained sandstone in the study area likely is less than the dispersivity measured in alluvial aquifers (generally 50 to about 200 ft), which commonly consist of heterogeneous mixtures or interbedded layers of gravel, sand, silt, and clay. This sensitivity analysis (fig. 45) and similar sensitivity analyses of solute-transport models of alluvial aquifers (Robson, 1974) indicate that model results are relatively insensitive to dispersivity values that are much smaller than the dimensions of the model grid. Although the dispersivity of the aquifer in this study likely is smaller than 200 ft, a value of 200 ft was used in the model in order to avoid underestimation of the rate of dispersion in the aquifer. Numerical dispersion produced by the mathematical approximations used in the computer code also is present in model results. The two forms of dispersion likely have only a small effect on model results but would tend to make the model slightly overestimate the rate of contaminant movement rather than underestimate the rate of contaminant movement.

Model simulations that use 10 and 15 percent porosity span the 14-percent mean porosity of the basal Williams Fork aquifer indicated by laboratory analysis of rock samples (table 7). The use of either 10 or 15 percent porosity in the model produces changes in the calculated concentration distribution and also affects the concentration of ground-water discharge to streams. Dissolved-solids concentration of ground water discharging to Trout Creek generally increases in response to the larger ground-water velocities produced by smaller porosity, as shown in figure 46. In the southeastern part of the model area, a 33-percent decrease in porosity (a change in model porosity from 15 to 10 percent) causes an approximately 13-percent increase in the dissolved-solids concentration of ground-water discharge to Trout Creek. If porosity is known within an uncertainty of ±20 percent, then the concentration of ground-water discharge to Trout Creek and possibly other streams in the area will have an uncertainty due to porosity of about ±10 percent.

Previous discussion of the effects of secondary permeability and porosity indicated that fracturing in the aquifer could not be shown to have produced a statistically significant change in hydraulic conductivity of fractured versus unfractured samples. It is unlikely that a significant increase in secondary porosity caused by fracturing could occur without a corresponding and much larger increase in hydraulic conductivity. However, no data were available to make a comparison of the porosity of fractured and unfractured rocks. The sensitivity analyses provide one means of indicating how changes in porosity caused by fracturing could affect the solute-transport simulations. If the porosity of fractured rock is assumed to be about 20 percent larger than that of unfractured rock (17 percent compared with 14 percent porosity) lateral ground-water velocity will decrease by about 17 percent and the dissolved-solids concentration of ground-water discharge to Trout Creek (for example) will be about 10 percent less than that indicated in subsequent simulations.

The model sensitivity to changes in porosity primarily occurs as the result of changes in ground-water velocity. Identical changes in ground-water velocity and model response can be produced by changes in hydraulic conductivity. (However, changes in hydraulic conductivity will cause changes in the water budget.) For example, a 17-percent decrease in lateral ground-water velocity can be produced by a 20-percent increase in porosity or a 17-percent decrease in hydraulic conductivity. Thus, the sensitivity of the model to changes in porosity also provides information on the sensitivity of the model to changes in hydraulic conductivity.

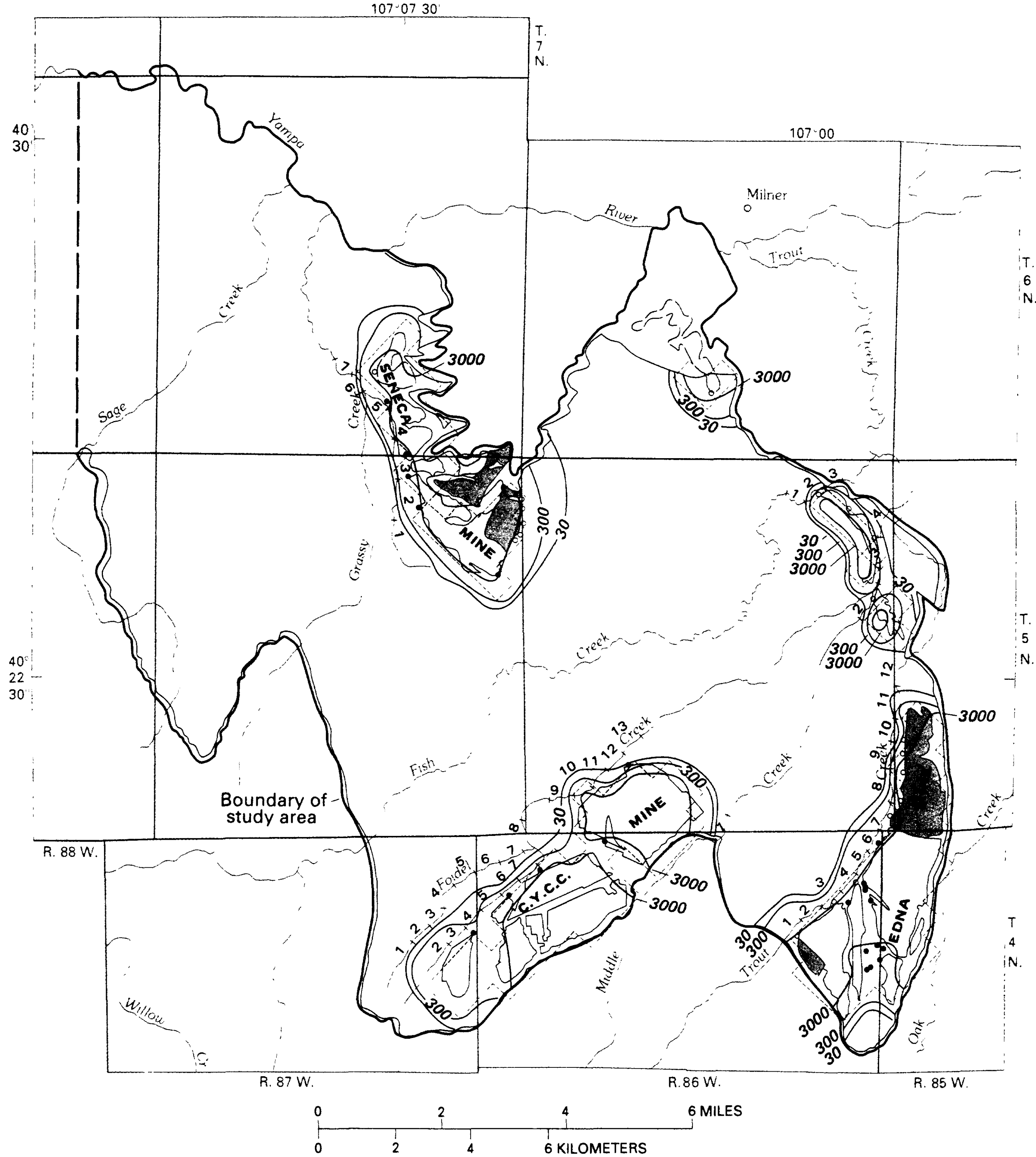
Model Simulations

Three sets of simulations were made using the solute-transport models of the basal Williams Fork and Twentymile aquifers. Set I simulated the effects on the basal Williams Fork aquifer of movement of poor quality water from spoil aquifers at inactive open-pit mines in the lower member of the Williams Fork Formation. Set II simulated the effects on the Twentymile aquifer of migration of poor quality water from spoil aquifers at inactive open-pit mines in the upper member of the Williams Fork Formation. Set III simulated the effects on the basal Williams Fork aquifer of migration of poor quality water from an inactive underground mine in the lower member of the Williams Fork Formation.

Flow of poor quality water from spoil aquifers in mined-out areas of open-pit, dip-slope mines was investigated by use of the first two sets of model simulations. For these simulations, the extent of the spoil aquifers was assumed to include both the present mined-out areas and the areas proposed for future open-pit mining. It also was assumed that heads in the spoil aquifers at the downdip contact with the basal Williams Fork or Twentymile aquifers would be controlled by the altitude of springs that have developed, or likely will develop, along the downdip edge of the spoils. The location (fig. 47) and altitude of these springs was used to determine the head relation between the spoil aquifer and the bedrock aquifers. In some areas, the head in the spoil aquifer was shown to be higher than the head in the adjacent basal Williams Fork or Twentymile aquifer, and poor quality water in the spoil aquifer could move directly into the adjacent bedrock aquifers. In other areas, heads in the spoil aquifers were shown to be lower than the heads in the adjacent bedrock aquifers, and water movement from the spoil to the bedrock would not occur in the local area.

A 30-year simulation period was assumed to begin at the close of open-pit mining in the local area. Mine plans submitted to State reglatory agencies by the local coal companies indicate that all future open-pit mining in the area will be complete prior to 1998. Any transient water-level or water-quality conditions in the aquifers prior to the close of mining are assumed to be negligible, as discussed previously in the "Simulation Procedures" section.

The dissolved-solids concentration of water in the spoil aquifers was assumed to remain constant at 4,500 mg/L. This concentration represents a "worst case"--that is, the largest concentration that Colorado State regulatory agencies assumed could conceivably occur (for purposes of this study) in the spoils during the simulation periods. It further was assumed that the 4,500-mg/L concentration in the spoils represents a 3,500-mg/L increase over the background concentration in the basal Williams Fork and Twentymile aquifers.



EXPLANATION

SOURCE OF POOR QUALITY WATER IN SPOIL AQUIFERS

Open-pit mining completed prior to 1987 in the lower member of the Williams Fork Formation. Boundary dashed where location is approximate

Proposed area of future open-pit mine in the lower member of the Williams Fork Formation

------------ BOUNDARY OF SIMULATED MINE

———— BOUNDARY OF MODELED PART OF BASAL WILLIAMS FORK AQUIFER

—2— STREAM REACH AND NUMBER

—300— LINE OF EQUAL INCREASE IN DISSOLVED-SOLIDS CONCENTRATION IN THE BASAL WILLIAMS FORK AQUIFER—Interval, in milligrams per liter, is variable

• SPRING ASSOCIATED WITH DISCHARGE FROM SPOIL AQUIFER

○ PROBABLE LOCATION OF FUTURE, OR PRESENTLY CONCEALED, DISCHARGE FROM SPOIL AQUIFER

Figure 47.--Simulated changes in dissolved-solids concentrations near open-pit mines after 30 years of solute movement.

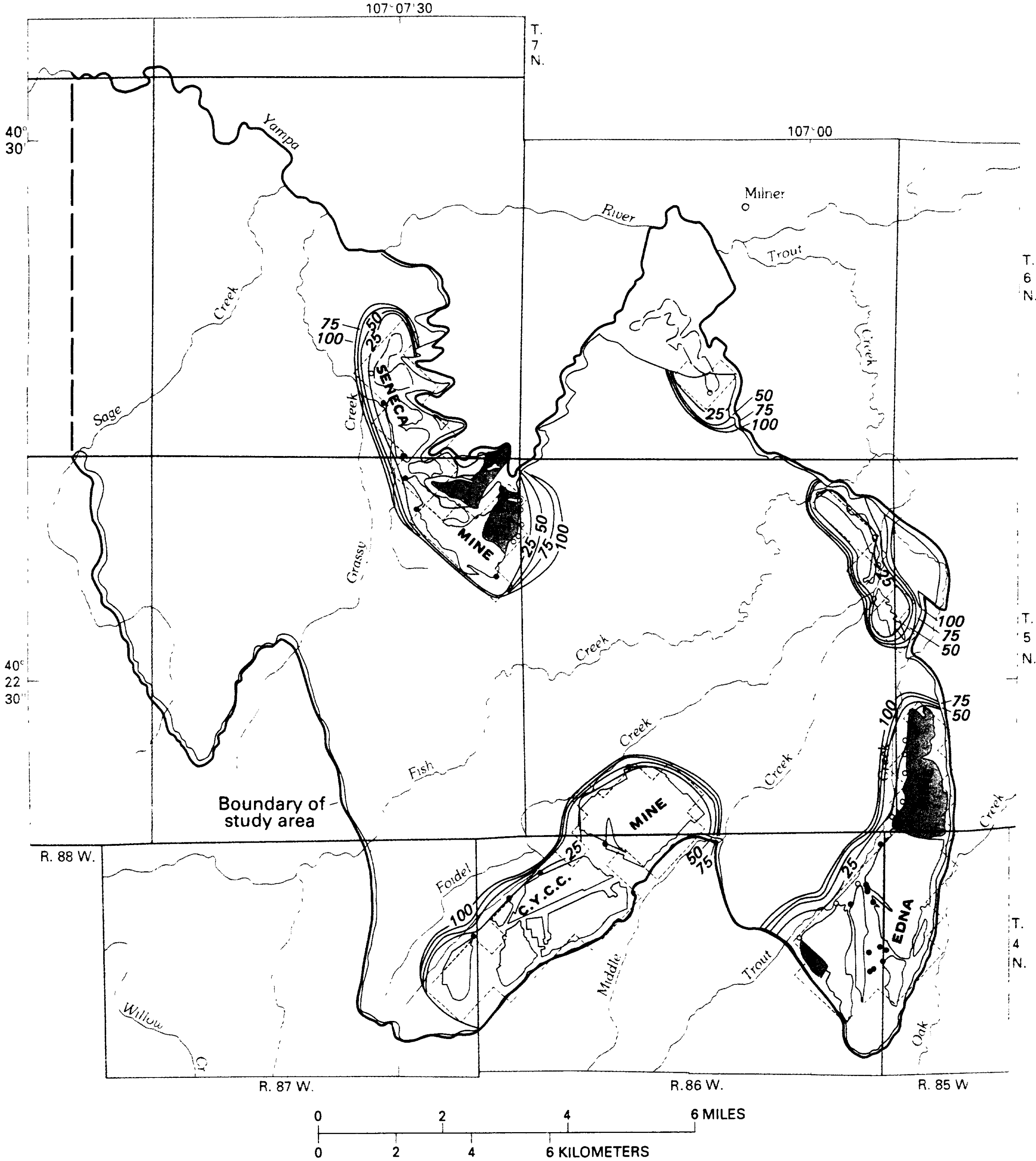
Simulations of the basal Williams Fork aquifer are based on a porosity of 14 percent, longitudinal dispersivity of 200 ft, and a transverse to longitudinal dispersivity ratio of 0.4. Simulations of the Twentymile aquifer are based on a porosity of 20 percent, longitudinal dispersivity of 200 ft, and a transverse to longitudinal dispersivity ratio of 0.4.

Simulation Set I

The increase in dissolved-solids concentration in the basal Williams Fork aquifer caused by 30 years of inflow of poor quality water from spoil aquifers in the lower member of the Williams Fork Formation is shown in figure 47. The spoil water is shown to cause an increase in concentration in the basal Williams Fork aquifer ranging from more than 3,000 mg/L near the spoil aquifer-bedrock aquifer interface, to less than 30-mg/L increase at distances generally less than 0.5 mi from the spoil aquifer. The limited movement of the degraded ground water in the 30-year simulation period primarily is the result of small rates of lateral ground-water movement in the basal Williams Fork aquifer. Simulations of up to 100 years of movement indicate a similar small rate of ground-water movement (fig. 48). In some areas, the movement of the degraded water is restricted further by the proximity of a spoil aquifer to a stream valley. Ground-water discharge to the stream valley may intercept all or part of the degraded water that moves toward the valley, thereby restricting, or terminating, the lateral movement of the degraded water in the aquifer.

Degraded water discharging from the aquifer in stream valleys also can affect the quality of surface flow in the stream. The three largest mines in the model area have the largest effect on the quality of the ground-water discharge. About 0.2 ft^3/s of simulated discharge to Trout Creek undergoes a 1,785-mg/L increase in dissolved-solids concentrations downgradient from the Edna Mine (table 11). The CYCC Mine produces a 1,541-mg/L increase in 0.35 ft^3/s of simulated discharge to Foidel Creek, and the Seneca Mine produces a 2,639-mg/L increase in 0.24 ft^3/s of discharge to Grassy Creek. Changes in streamflow quality produced by the simulated rates of ground-water discharge likely will not be significant because much larger rates of flow from spoil-aquifer springs directly enter the streams, or the stream-valley alluvium, and provide a means for much more rapid and direct change in the chemical quality of the streamflow. In Trout Creek, the 0.2 ft^3/s of simulated ground-water discharge will be greatly diluted by the 10 to 20 ft^3/s of measured base flow in the stream.

Near the southwestern ends of the CYCC and Edna Mines, ground water near the spoil aquifers (figs. 47 and 48) has been diluted by simulated inflow of small dissolved-solids concentration water from parts of the basal Williams Fork aquifer located upgradient of the spoil aquifer. Near the smaller mines in the northeastern part of the area, the concentration changes shown by the model (figs. 47 and 48) do not correspond well to the shape of the spoil aquifer due to the limited resolution of the 2,000-ft grid interval model.



EXPLANATION

SOURCE OF POOR QUALITY WATER IN SPOIL AQUIFERS

Open-pit mining completed prior to 1987 in the lower member of the Williams Fork Formation. Boundary dashed where location is approximate

Proposed area of future open-pit mine in the lower member of the Williams Fork Formation

BOUNDARY OF SIMULATED MINE

BOUNDARY OF MODELED PART OF BASAL WILLIAMS FORK AQUIFER

—100— LINE OF 500 MILLIGRAMS PER LITER DISSOLVED-SOLIDS CONCENTRATION IN THE BASAL WILLIAMS FORK AQUIFER—Time interval 25 years

SPRING ASSOCIATED WITH DISCHARGE FROM SPOIL AQUIFER

PROBABLE LOCATION OF FUTURE, OR PRESENTLY CONCEALED, DISCHARGE FROM SPOIL AQUIFER

Figure 48.--Extent of movement of poor quality water away from open-pit mines during a 100-year simulation period.

Table 11.--*Increase in dissolved-solids concentration of ground-water discharge to streams produced by model simulation set I*

[ft^3/s, cubic feet per second; mg/L, milligrams per liter]

Reach (fig. 47)	Affected ground-water discharge to creek (ft^3/s)	Increase in dissolved-solids concentration in ground-water discharge due to effects of spoil aquifers (mg/L)
	Trout Creek	
1	0.0130	844
2	0.0274	1,892
3	0.0400	3,395
4	0.0252	3,477
5	0.0178	741
6	0.0153	1,034
7	0.0178	770
8	0.0155	1,069
9	0.0075	29
10	0.0061	423
11	0.0109	193
Total discharge	0.1965	
		Discharge-weighted mean concentration = 1,785
	Grassy Creek	
1	0.0149	3,428
2	0.0060	57
3	0.0813	2,432
4	0.0468	3,477
5	0.0133	604
6	0.0667	3,106
7	0.0086	1
Total discharge	0.2376	
		Discharge-weighted mean concentration = 2,639
	Foidel Creek	
1	0.0206	7
2	0.0390	414
3	0.0447	904
4	0.0237	161
5	0.0180	340
6	0.0195	2,105
7	0.0400	2,557

Table 11.--*Increase in dissolved-solids concentration of ground-water discharge to streams produced by model simulation set I*--Continued

Reach (fig. 47)	Affected ground-water discharge to creek (ft^3/s)	Increase in dissolved-solids concentration in ground-water discharge due to effects of spoil aquifers (mg/L)
	Foidel Creek--Continued	
8	0.0543	2,592
9	0.0036	275
10	0.0366	3,425
11	0.0110	434
12	0.0253	2,060
13	0.0102	1
Total discharge	0.3465	
		Discharge-weighted mean concentration = 1,541
	Middle Creek	
1	0.0771	12
2	0.0102	2,280
3	0.0149	56
4	0.0070	29
Total discharge	0.1092	
		Discharge-weighted mean concentration = 231
	Fish Creek	
1	0.0049	4
2	0.0282	25
3	0.0088	54
Total discharge	0.0419	
		Discharge-weighted mean concentration = 29

Simulation Set II

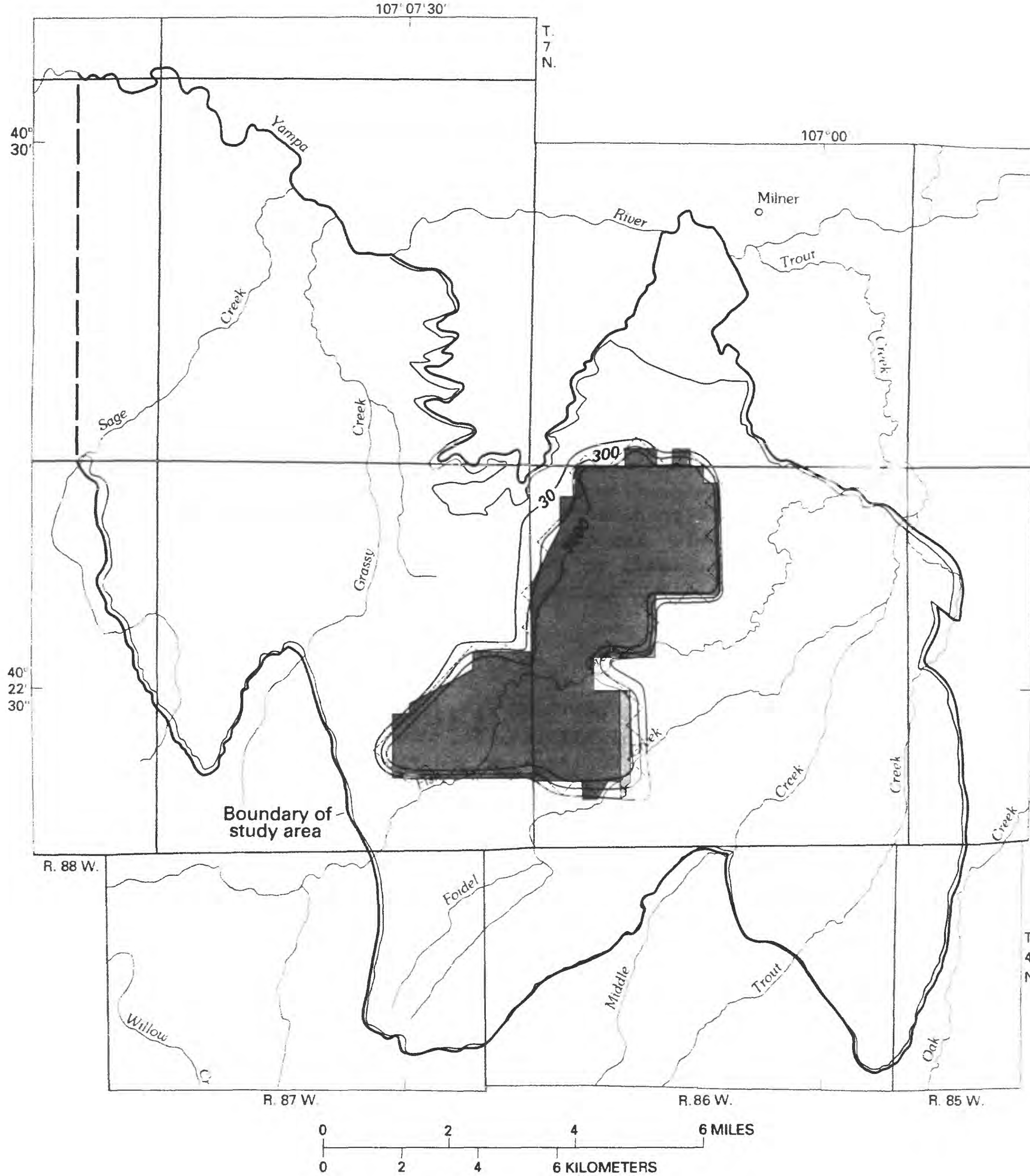
The change in dissolved-solids concentrations in the Twentymile aquifer caused by poor quality water in spoils in the upper member of the Williams Fork Formation was examined in this set of simulations. The only local mines that worked coal seams in this unit were located between Fish Creek and Foidel Creek in the west-central part of Twentymile Park. Heads in the spoil aquifers again were estimated as the basis of the altitudes of springs at the downdip edge of the spoil aquifers. The altitude of these springs and the resulting estimated heads in the spoil aquifers were determined to be critical to the model simulations because heads in the Twentymile aquifer are at or above land surface in most of the area of the spoils. In all but the southeasternmost part of the spoils, the spoil-aquifer heads were 0 to 130 ft below heads in the Twentymile aquifer. This head relation would prevent any significant movement of poor quality water from the spoil aquifer into the bedrock aquifer; model simulations were similar, indicating minimal effect of these spoil aquifers on the Twentymile aquifer.

If the head relation had allowed migration of poor quality water, the effect on the Twentymile aquifer still likely would have been small because Fish Creek valley is the local ground-water discharge area for the aquifer and would have intercepted almost all of the degraded water entering the Twentymile aquifer. Under existing conditions, the spoil aquifers discharge at springs and seeps, or by underflow into the alluvium, and contribute dissolved solids to the streamflow more directly than would be possible by means of flow through the bedrock aquifer. The quantity of direct discharge to Fish Creek is not known but likely is small in comparison to the 1 to 5 ft^3/s of base flow normally present in this reach of Fish Creek.

Simulation Set III

This set of simulations was designed to investigate the changes in ground-water quality in the basal Williams Fork aquifer caused by movement of poor quality water away from an inactive and flooded underground mine located in the central part of Twentymile Park. Mine development plans submitted by the coal company indicate that mining would be completed by 2017. Mine flooding probably would continue for several years after the workings were abandoned. A steady-flow, transient-transport simulation was used to investigate 30 years of movement of poor quality water away from workings flooded to the same level of head as had existed in the aquifer prior to mining. A transient-flow, transient-transport simulation was used to investigate 30 years of movement from workings flooded to the level of the average premining head at the margin of the mined area. The former conditions are more representative of a mined-out area that has hydraulic conductivity similar to that of the original premined materials, such as might occur following the collapse of the workings. The latter conditions are more representative of a mined-out area that has an extremely large hydraulic conductivity due to water flow through uncollapsed mine workings. The principal hydrologic difference between these two conditions is that in the first example head gradients at the boundary of the mined area are identical to the premining gradients, whereas head gradients in the second example generally are larger because of the assumption of a uniform average head throughout the mine.

The water-quality results of the steady-flow, transient-transport simulation (fig. 49) are not markedly different from the results of the transient-flow, transient-transport simulations (fig. 50). This indicates that the rate of lateral movement of degraded water away from the inactive underground mine will not be seriously affected by the collapsed or uncollapsed condition of



EXPLANATION

SOURCE OF POOR QUALITY WATER IN FLOODED UNDERGROUND WORKINGS

Underground mining completed prior to 1987 in the lower member of the Williams Fork Formation

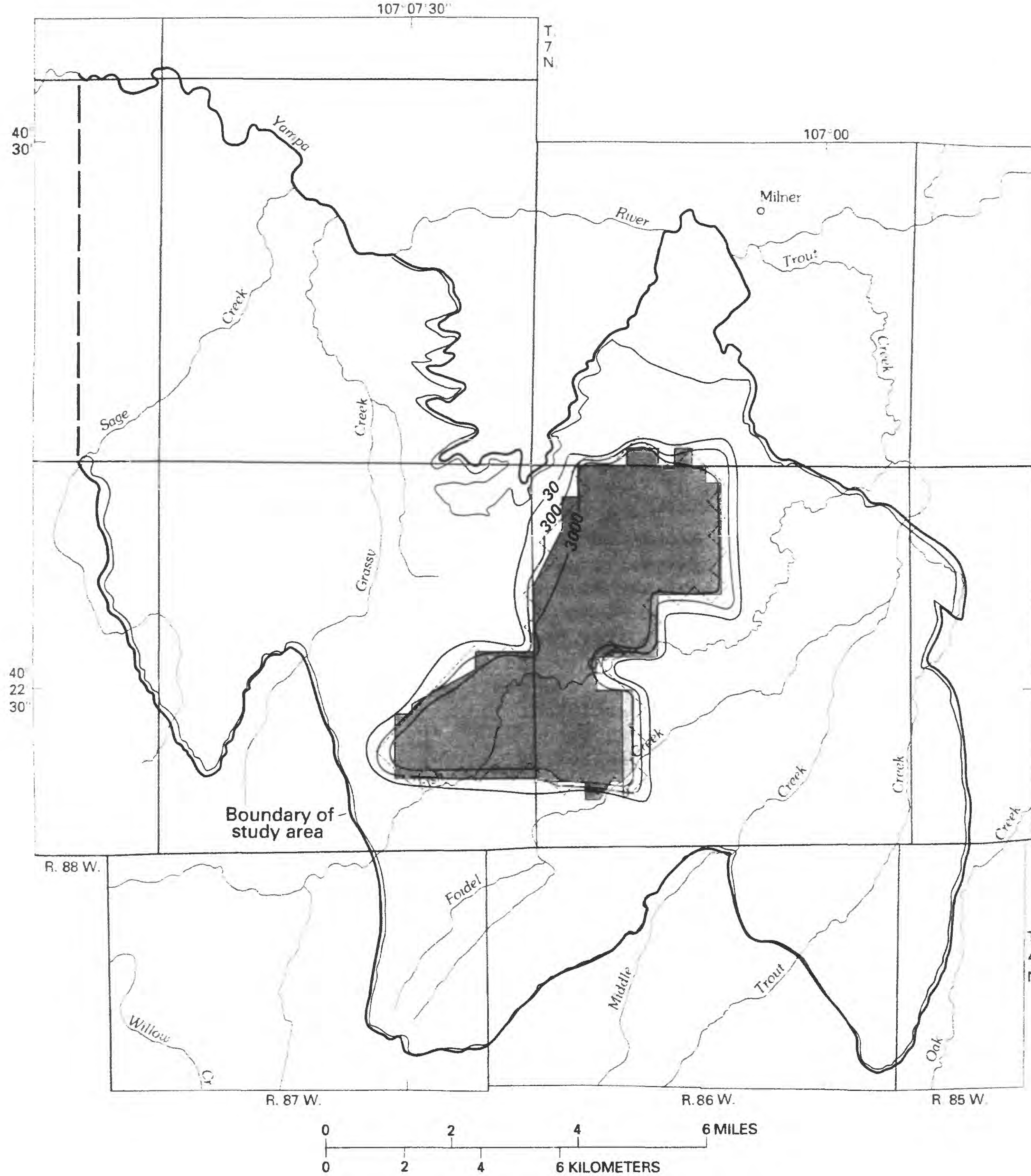
Proposed area of future underground mine in the lower member of the Williams Fork Formation

BOUNDARY OF SIMULATED MINE

BOUNDARY OF MODELED PART OF BASAL WILLIAMS FORK AQUIFER

300 LINE OF EQUAL INCREASE IN DISSOLVED-SOLIDS CONCENTRATION IN THE BASAL WILLIAMS FORK AQUIFER—Interval, in milligrams per liter, is variable

Figure 49.--Thirty-year simulated change in dissolved-solids concentrations near an underground mine using steady-flow, transient-transport conditions.



EXPLANATION

SOURCE OF POOR QUALITY WATER IN FLOODED UNDERGROUND WORKINGS

Underground mining completed prior to 1987 in the lower member of the Williams Fork Formation

Proposed area of future underground mine in the lower member of the Williams Fork Formation

BOUNDARY OF SIMULATED MINE

BOUNDARY OF MODELED PART OF BASAL WILLIAMS FORK AQUIFER

300 LINE OF EQUAL INCREASE IN DISSOLVED-SOLIDS CONCENTRATION IN THE BASAL WILLIAMS FORK AQUIFER—Interval, in milligrams per liter, is variable

Figure 50.--Thirty-year simulated change in dissolved-solids concentrations near an underground mine using transient-flow, transient-transport conditions.

the mine workings. As in previous simulations, the relatively small movement of the degraded water during the simulation period is due to the small ground-water velocities in the aquifer. The model-simulated effects of hydrodynamic mixing (dispersion) cause simulated changes in concentration upgradient and downgradient from the mine.

SUMMARY AND CONCLUSIONS

The bedrock geohydrologic system in the upper part of the Mesaverde Group of northwestern Colorado consists of two regional aquifers separated by three principal confining layers. The confining layers, consisting primarily of marine shale, underlie the Trout Creek Sandstone Member of the Iles Formation (the deepest regional aquifer), separate the Trout Creek Sandstone Member from the younger Twentymile Sandstone Member of the Williams Fork Formation (the second regional aquifer), and overlie the Twentymile Sandstone Member. Numerous aquifers of local extent are present in sandstone beds and coal seams of the middle confining layer in the sandier lithology of the western part of the study area. In the eastern part of the study area, the only local aquifer (the basal Williams Fork aquifer) consists of sandstone and coal within the basal part of the Williams Fork Formation.

The basal Williams Fork aquifer has greater water-yielding potential than either of the two regional aquifers in the eastern area. Sandstones in the Trout Creek and Twentymile aquifers are similar in appearance, composition, grain size, sorting, and thickness (about 100 to 150 ft) but differ in average hydraulic conductivity; the hydraulic conductivity of the Trout Creek aquifer is about one-third that of the Twentymile aquifer. The basal Williams Fork aquifer generally contains more sandstone (about 100 to 200 ft) and has an average hydraulic conductivity about eight times larger than that of the Twentymile aquifer. The resulting mean transmissivity is about 20 ft^2/d for the basal Williams Fork aquifer, 4 ft^2/d for the Twentymile aquifer, and 0.6 ft^2/d for the Trout Creek aquifer. Fractured coal seams may contribute to the larger average hydraulic conductivity of the basal Williams Fork aquifer.

Infiltration of precipitation is the principal source of recharge to the bedrock aquifers in the study area. Precipitation generally increases with altitude because of orographic effects associated with up-valley and cross-valley movement of storms. The upper reach of the Yampa River valley is an exception in that lesser mean annual precipitation occurs at higher altitudes upstream from Steamboat Springs because of rain-shadow effects of cross-valley tracking storms. The mean annual precipitation of 14 to 25 in. in the study area is much less than potential evaporation, which exceeds 40 in/yr. As a result, excess surface water is available to recharge the aquifers only during periods of snowmelt or intense rainfall. Of the approximately 150 ft^3/s of mean annual precipitation that falls on the eastern part of the study area, only about 2 percent recharges the bedrock aquifers.

Geologic structure and the resulting topography of the formations have an important bearing on the ground-water recharge, discharge, and flow system in the aquifers. Structure in the study area has markedly dissimilar eastern and western tectonic forms. In the eastern part of the area, complex deformation associated with the Laramide orogeny has produced a series of four plunging synclinal and anticlinal features that resulted in structural basins southeast of Hayden and in Twentymile Park. Structurally high areas occur at outcrops of the formations in the mountainous areas surrounding Twentymile Park and on the elevated flanks of the Sage Creek, Fish Creek, and Tow Creek anticlines. In the western part of the area, the predominant structure is the southern limb of the Sand Wash basin, which has been only slightly deformed and dips northward. Structurally high areas occur along the crest of the Williams Fork Mountains at the southern margin of the western area.

Exposed outcrops of the aquifer units allow infiltration of water from precipitation and snowmelt. This water may become part of a local ground-water flow system and discharge at local stream valleys crossing the outcrop, or the water may become part of a larger regional ground-water flow system and move to depth in the aquifer. Modeling indicates that recharge to the three aquifers in the eastern part of the study area totals only about 2.8 ft^3/s. Rates of discharge are similar under the steady-flow conditions in the area and occur by upward leakage through leaky confining layers, by lateral flow to stream valleys that cross low-lying outcrops, or by evapotranspiration.

In the eastern part of the study area, ground water generally moves from recharge areas along the elevated margins of the aquifers toward discharge areas in the central low-lying parts of Twentymile Park and the valleys of Grassy, Fish, Foidel, Middle, and Trout Creeks. Lateral ground-water velocities generally range from 0.5 to 30 ft/yr. Head gradients between the shallow and deeper aquifers enable downward movement of water in the recharge areas and upward movement of water in Twentymile Park and near Grassy Creek and the Yampa River. Calculated traveltimes for a particle of water to move vertically through the slightly leaky confining layers separating the aquifers average about 8,000 years. Heads in all the aquifers are above land surface in much of the low-lying area in Twentymile Park.

In the western part of the study area, ground water generally moves in a northeasterly direction from the recharge areas along the upper parts of the Williams Fork Mountains toward discharge areas, or outflow areas, along the study area boundary at the Yampa River. This larger flow system contains smaller flow systems associated with local recharge in upland areas and discharges in nearby outcrops of water-yielding units in stream valleys. Downward head gradients in the recharge areas and upward head gradients in the discharge areas likely occur as they do in the eastern part of the area.

Most streamflow is the result of snowmelt and precipitation runoff and is little affected by ground-water recharge or discharge in the study area. Subparallel streams that drain cuesta dip slopes formed by the Williams Fork Formation or Lewis Shale in the western part of the area generally are ephemeral; snowmelt runoff occurs from March to July. Discontinuous perennial reaches are produced by ground-water discharge at seeps and springs. Larger streams in the eastern part of the area commonly cross structural trends, have perennial flow, and may have drainage areas extending well beyond the study area. Gain-loss measurements in Fish Creek and its unnamed tributaries draining Twentymile Park indicate small gains in streamflow at the points where the streams cross the mountain-front outcrop of the aquifer units. Minimal gain in streamflow occurs downstream from these outcrops even though heads in the aquifers may be above land surface. Surface-water quality is strongly affected by the geology of the drainage area. Older, crystalline-rock drainage areas upstream from the study area generally yield calcium bicarbonate streamflow of excellent quality (100 to 400 mg/L of dissolved solids). Sedimentary rocks of mixed continental and marine origins, such as the Williams Fork Formation, commonly yield streamflow of either calcium magnesium bicarbonate or calcium magnesium sulfate composition; dissolved-solids concentrations range from 300 to 800 mg/L. Marine terrain yields streamflow of magnesium sodium sulfate composition that has dissolved-solids concentrations of about 1,000 to about 8,000 mg/L.

The chemical composition of ground water in the study area is the result of geochemical processes that include dissolution, cation exchange, and precipitation. These processes may differ depending on the aquifer sampled and the location of the sample point in the ground-water flow path in the aquifer. Carbonate dissolution near the margins of the basal Williams Fork aquifer produces the calcium bicarbonate water that predominates within about 1 mi of the outcrop. As the water moves farther into the aquifer, cation exchange naturally softens the water and produces a sodium bicarbonate water type, and dissolved-solids concentrations range from 300 to 1,400 mg/L. Oxidation of pyritic minerals associated with coal and dissolution of gypsum contribute dissolved sulfate to ground water downgradient from spoils and coal outcrops. Sulfate concentrations decrease at greater distance along the ground-water flow path, possibly in response to sulfate reduction.

Solute-transport models that simulate dissolved-solids concentrations in the basal Williams Fork aquifer and in the Twentymile aquifer were constructed. These models were used to evaluate the potential effect on the aquifers of movement of poor quality water away from spoil aquifers and flooded underground mines. Simulation results indicate that ground-water velocities in these aquifers are commonly so small that degraded water does not move a significant distance from its source within the 30- to 100-year modeling timeframe. Thus, mining effects on bedrock water quality are small even when worst-case concentrations are simulated in the spoil aquifers.

The short distance between ground-water discharge areas at streams and the spoil aquifers at open-pit mines may decrease or halt further movement of degraded ground water. Ground-water discharge areas at streams commonly receive inflow from the bedrock aquifer underlying both sides of the stream valley. If degraded water moves toward the discharge area from a spoil aquifer on one side of the valley, the convergent ground-water flow field may prevent the movement of the degraded water beyond the valley. Spoil aquifers at each of the three large open-pit mines, and several of the smaller mines, in the eastern part of the study area are located on dip slopes above the stream valleys of Trout Creek, Foidel Creek, Fish Creek, and a tributary to Grassy Creek. Each of these stream valleys function as ground-water discharge areas and tend to retard movement of degraded water beyond the valley.

Movement of degraded water away from spoil aquifers primarily will affect the chemical quality of the ground water discharging to the nearby stream valley. However, the most rapid and direct effect on surface-water quality is produced by the direct discharge of degraded water to the streams from spoil seeps and springs. The effect on stream quality attributable to movement of degraded water through the bedrock aquifer will be delayed, because of small rates of ground-water movement, and also will be decreased because the small rates of affected ground-water discharge (generally less than 0.3 ft^3/s) will be diluted by the relatively large rates of streamflow (generally 1 to 20 ft^3/s).

Minimal differences in model simulation results were obtained by changing the head configuration in a simulated underground mine to represent hydrologic conditions associated with open mine voids or collapsed mine voids.

Sensitivity analyses of model dispersivity and porosity indicated that simulation results are insensitive to dispersivity but more sensitive to porosity. Porosity variations of 33 percent produced a 13-percent change in the dissolved-solids concentrations of ground water discharging to a stream downgradient from a spoil aquifer.

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SUPPLEMENTAL INFORMATION

Flow-Model Design

The flow model consists of three active layers with head-dependent vertical leakage between layers. Lateral hydraulic conductivity and transmissivity were spatially distributed as previously indicated. Vertical leakance between active layers was calculated as the vertical hydraulic conductivity of shale (3.6×10^{-4} ft/d) divided by the thickness of the confining layers and was spatially distributed on the basis of shale thickness as previously defined.

Several types of boundary conditions were used in the model. Precipitation recharge was simulated at nodes at the outcrops of each aquifer as indicated in figures 51-53. The rate of recharge was calculated as the product of the potential recharge rate times the area of outcrop in each node. A constant-head boundary condition was simulated at each node representing the subcrop of the entire thickness of an aquifer under a perennial stream. Most constant-head nodes are located near the more steeply dipping formations at the margins of the aquifers (figs. 51-53). Constant-head altitudes were defined by the altitude of the stream at the subcrop. Head-dependent leakage into or out of a stream overlying an aquifer (but not in contact with the full thickness of the aquifer, as in the case of constant-head nodes) was simulated by river nodes. Spatially distributed conductance of the river confining layer was estimated on the basis of the shale thickness in the aquifer near the river and the vertical hydraulic conductivity of shale. Ground-water discharge by evapotranspiration and springs in areas where the overlying confining layer outcrops was simulated by head-dependent discharge ("L" in figs. 51 and 52). Spatially distributed conductance of the confining layer was defined by the shale thickness of the outcrop part of the unit and the vertical hydraulic conductivity of shale. No cross-boundary flow was simulated on the periphery of the aquifers because the aquifers outcrop or because model boundaries coincide with the potentiometric gradient.

Solute-Transport Model Design

The design of the solute-transport model is similar to that of the flow model except for those aspects that deal with vertical connection between layers. Both models use the same grid spacing and the same grid network (figs. 54 and 55), although two additional rows and columns of inactive nodes are required by the solute-transport model code. The models share common values and areal distributions of lateral hydraulic conductivity, precipitation recharge, constant-head nodes, rivers, and discharge to outcrops of the confining layers. In the flow model, vertical leakage between model layers is computed by the model as a function of head difference between model layers. The solute-transport model simulates only a single layer and vertical leakage is specified in the model as a constant rate of recharge or discharge. The rate and spatial distribution of vertical leakage used in the solute-transport model is defined by flow-model results. Because both models simulate steady-flow conditions, changes in vertical leakage with time are not considered. In figure 54, vertical leakage between the underlying Trout Creek aquifer ("U") is differentiated from the vertical leakage to the overlying Twentymile aquifer ("M"); however, leakage is used in the solute-transport model as a net value.

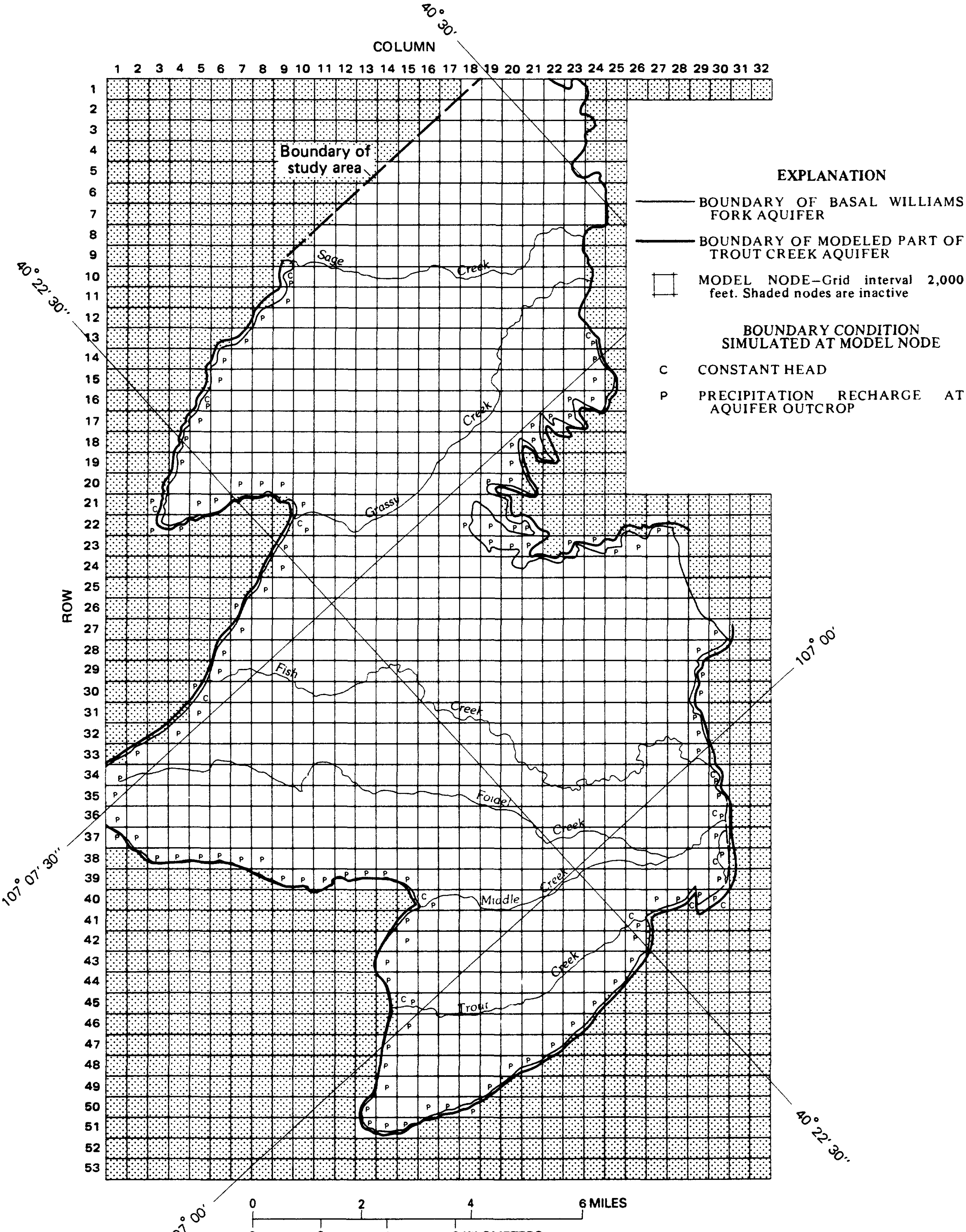

Figure 51.--Flow-model grid and nodal distribution of boundary conditions in the Trout Creek aquifer.

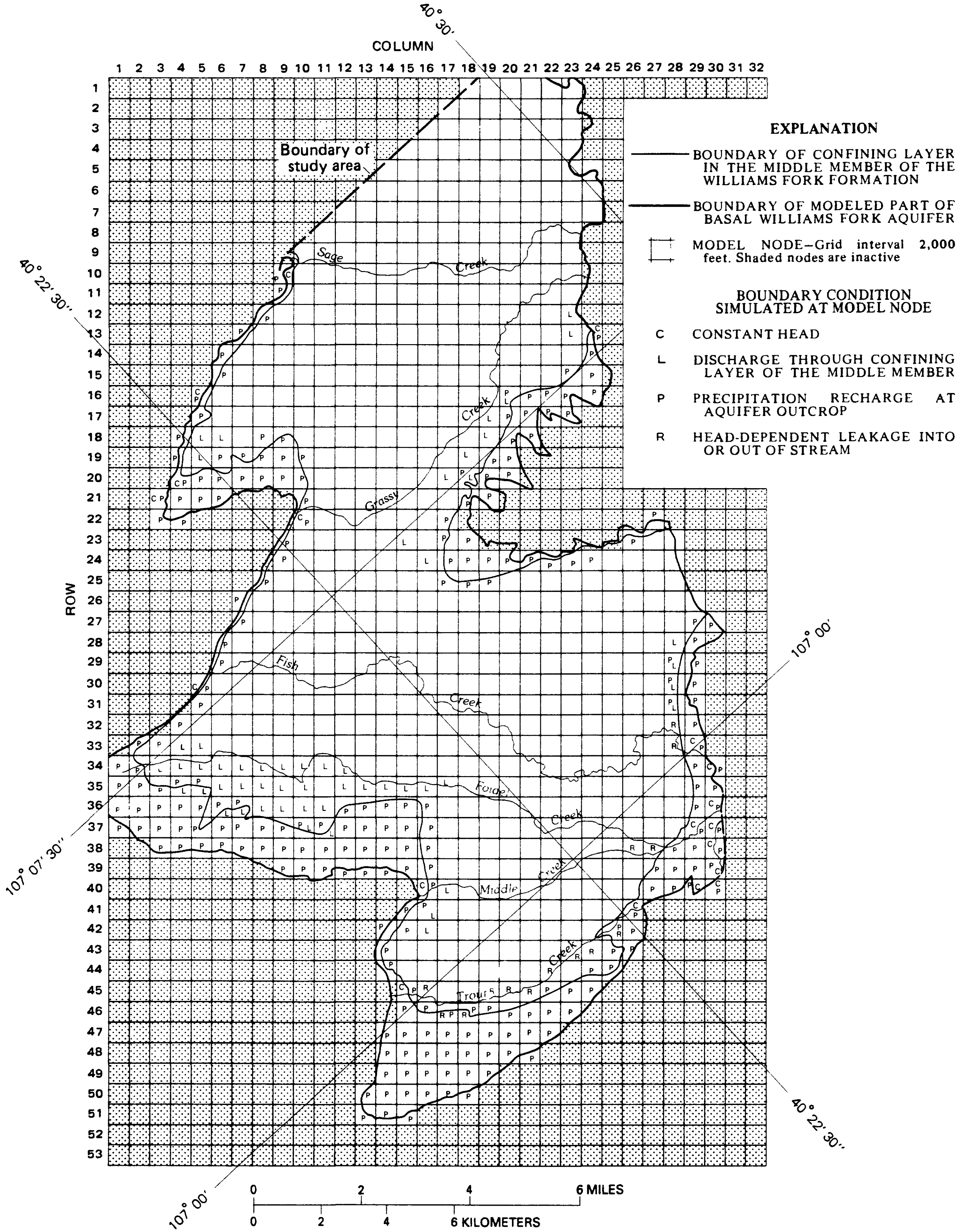

Figure 52.--Flow-model grid and nodal distribution of boundary conditions in the basal Williams Fork aquifer.

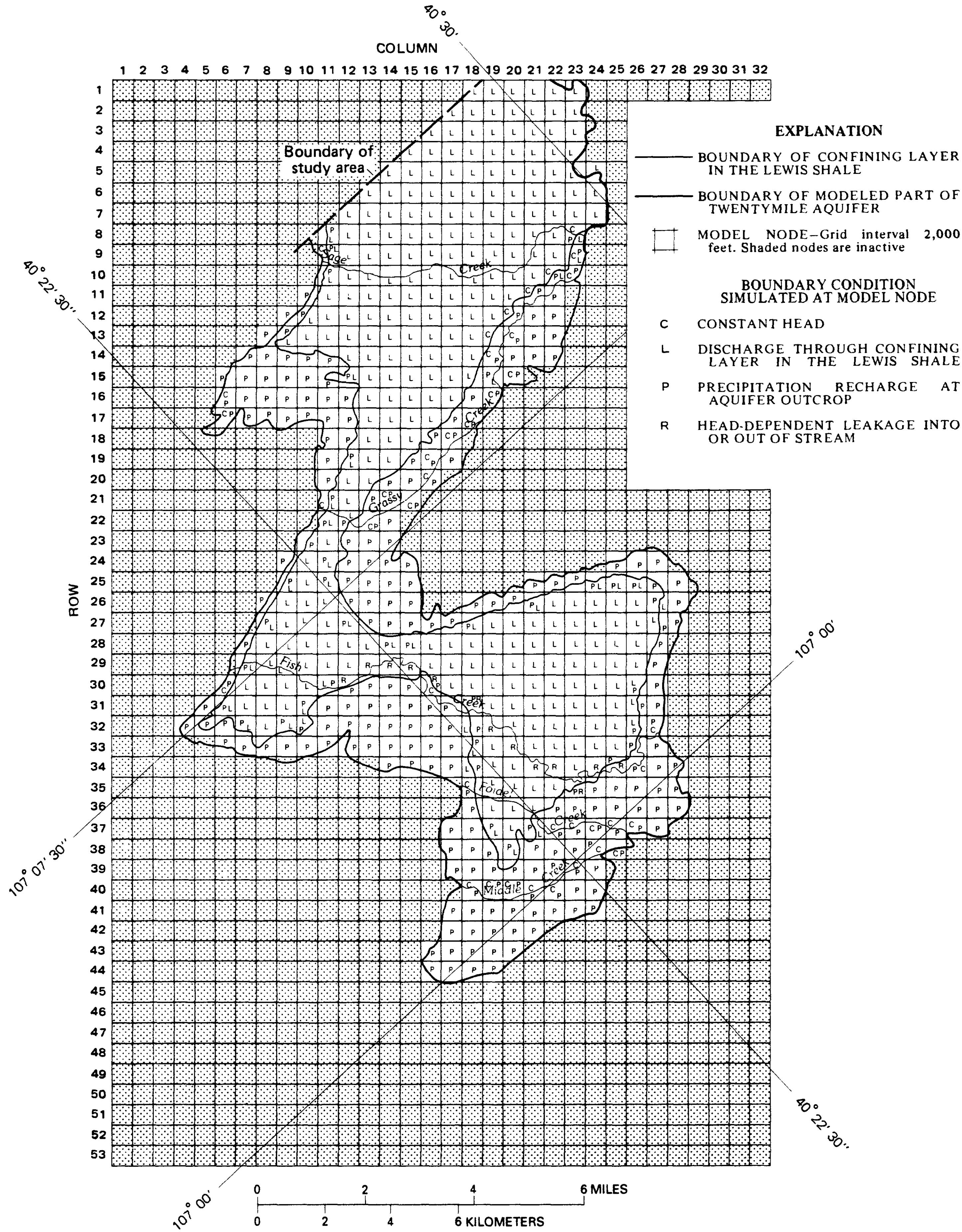

Figure 53.--Flow-model grid and nodal distribution of boundary conditions in the Twentymile aquifer.

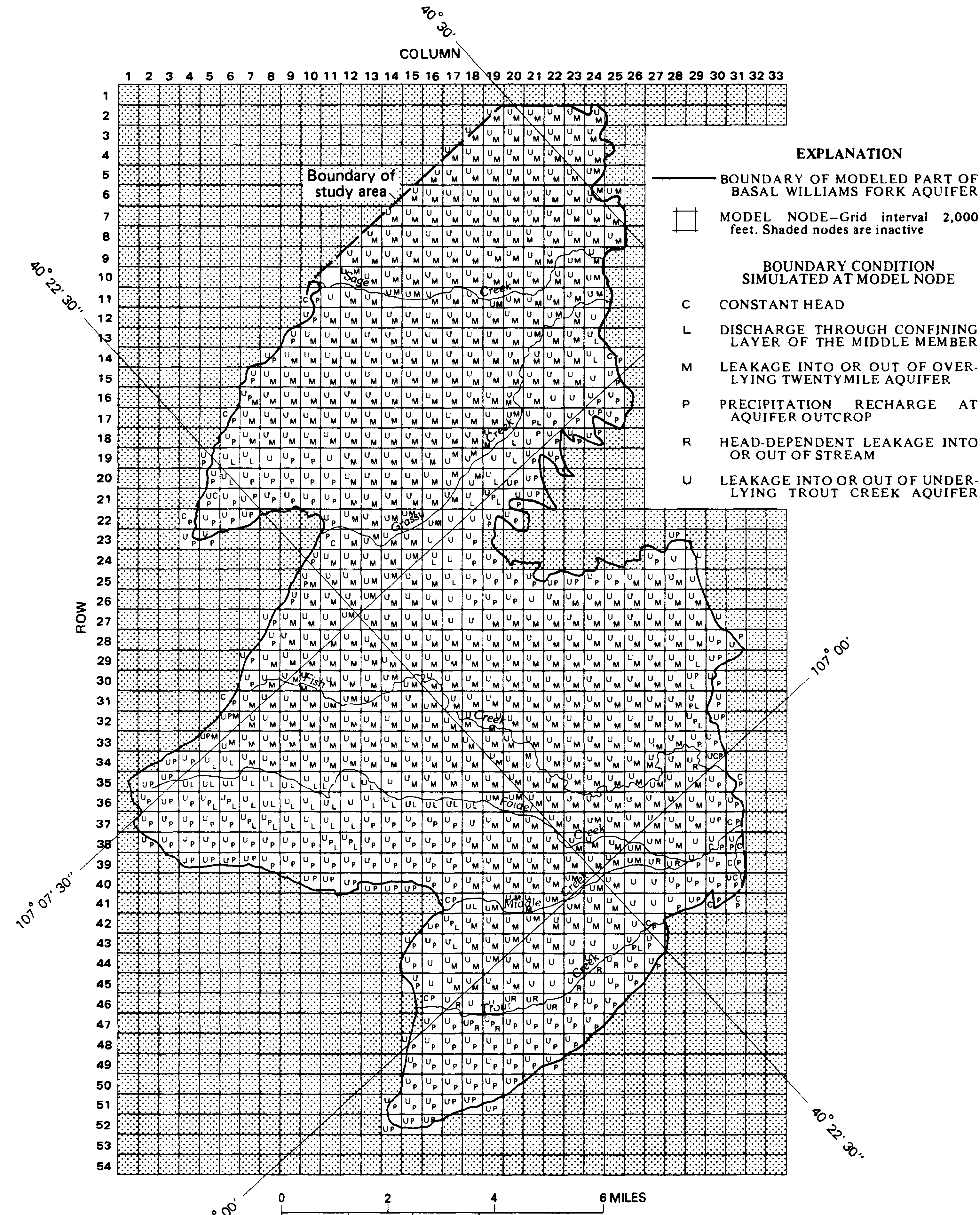

Figure 54.--Solute-transport model grid and nodal distribution of boundary conditions in the basal Williams Fork aquifer.

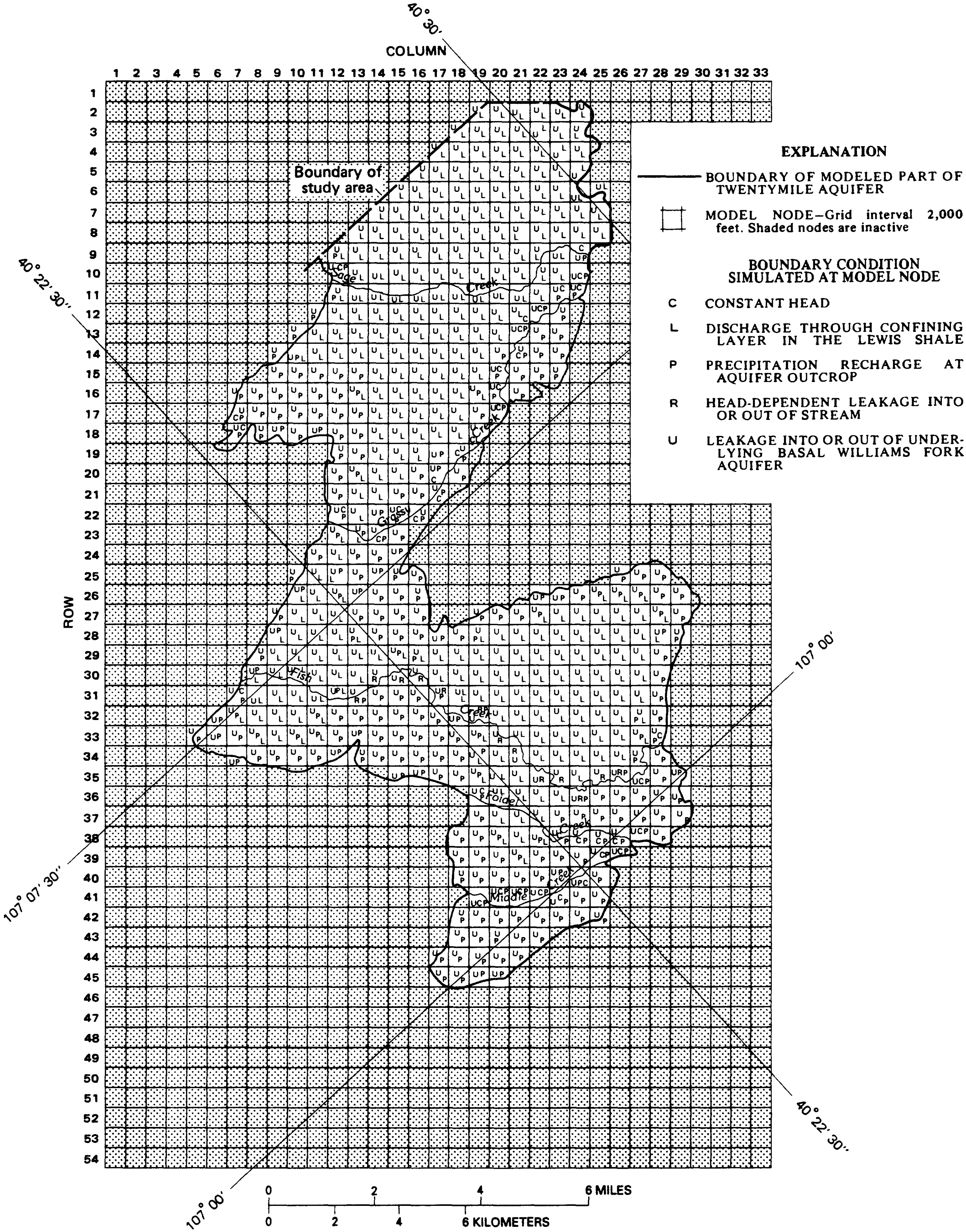

Figure 55.--Solute-transport model grid and nodal distribution of boundary conditions in the Twentymile aquifer.

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