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Geohydrology of Tertiary Rocks in the Upper Colorado River Basin in Colorado, Utah, and Wyoming, Excluding the San Juan Basin

Regional Aquifer-System Analysis

By Kent C. Glover, David L. Naftz, *and* Lawrence J. Martin

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CONVERSION FACTORS, ABBREVIATIONS, AND VERTICAL DATUM

Multiply	By	To obtain
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
cubic foot per day (ft^3/d)	0.02832	cubic meter per day
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
inch (in.)	25.4	millimeter
inch per year (in/yr)	25.4	millimeter per year
mile (mi)	1.609	kilometer
reciprocal foot (ft^{-1})	0.3048	reciprocal meter
square foot per day (ft^2/d)	0.0929	square meter per day
square foot (ft^2)	0.0929	square meter
square mile (mi^2)	2.590	square kilometer

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

$$°F = 9/5(°C) + 32.$$

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".

The following units of measure and their abbreviations are used in this report:

meq/L = milliequivalents per liter
mg/L = milligrams per liter
mmol/L = millimoles per liter

GEOHYDROLOGY OF TERTIARY ROCKS IN THE UPPER COLORADO RIVER BASIN IN COLORADO, UTAH, AND WYOMING, EXCLUDING THE SAN JUAN BASIN

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ABSTRACT

Four hydraulically isolated aquifer systems in Tertiary rocks have been identified in the sedimentary basins of the Upper Colorado River Basin: The Piceance Basin, the Uinta Basin, the Green River Basin, and the Great Divide-Washakie-Sand Wash Basins aquifer systems. The Piceance Basin aquifer system consists of two aquifers separated from each other and from the underlying Cretaceous Mesaverde aquifer by confining units. Aquifer properties generally are related to the degree and interconnection of fractures. The upper Piceance Basin aquifer is contained within the Uinta Formation and upper part of the Parachute Creek Member of the Green River Formation. Hydraulic conductivity ranges from 0.003 to 1.6 feet per day. The Mahogany confining unit, which underlies the upper Piceance Basin aquifer, has an average thickness of 160 feet and a hydraulic conductivity generally one to two orders of magnitude smaller than the hydraulic conductivity of the superjacent and subjacent aquifers. The lower Piceance Basin aquifer is contained within the lower part of the Parachute Creek Member of the Green River Formation; hydraulic conductivity ranges from 0.001 to 1.2 feet per day. The basal confining unit consists of 2,000 to 4,000 feet of older Tertiary rocks and probably has a hydraulic conductivity less than 0.01 foot per day. The Piceance Basin aquifer system receives recharge water in upland areas, transmits part of the water horizontally through the upper Piceance Basin aquifer toward discharge areas along major streams or springs on canyon walls, and transmits the remaining water downward through the Mahogany confining unit into the lower Piceance Basin aquifer. Water in the lower Piceance Basin aquifer moves generally horizontally and, in the vicinity of discharge areas, leaks upward into the upper Piceance Basin aquifer. Recharge totals about 42 cubic feet per second. Discharge occurs primarily as seepage to alluvium along Yellow and Piceance Creeks and springs near Roan and Parachute Creeks. The basal confining unit separates the lower Piceance Basin aquifer from the Mesaverde aquifer.

Two major aquifers have been identified in the Uinta Basin aquifer system: The Duchesne River-Uinta aquifer and the Douglas Creek-Renegade aquifer. The aquifers are separated by a 3,000- to 6,000-foot confining unit. An equally thick confining unit separates the Douglas Creek-Renegade aquifer from the underlying Mesaverde aquifer. Flow-model analysis indicates that hydraulic conductivity of the Duchesne River-Uinta aquifer ranges from 0.5 to 1.0 foot per day and is related to the percentage of sandstone present, but is enhanced by fractures in the lower part of the aquifer and within the central part of the basin. Hydraulic conductivity of the Douglas Creek-Renegade aquifer is related primarily to the percentage of sandstone present and ranges from 0.05 to 0.3 foot per day. Water in aquifers in Tertiary rocks of the Uinta Basin moves generally horizontally from basin margins toward major streams of the basin. Vertical movement of water

has not been detected and, with local exceptions, probably is negligible. Recharge occurs in upland areas peripheral to the Uinta Basin, where precipitation annually exceeds 10 inches. Total recharge is estimated to be about 270 cubic feet per second. Additional local recharge and discharge probably occurs, particularly in the Douglas Creek-Renegade aquifer.

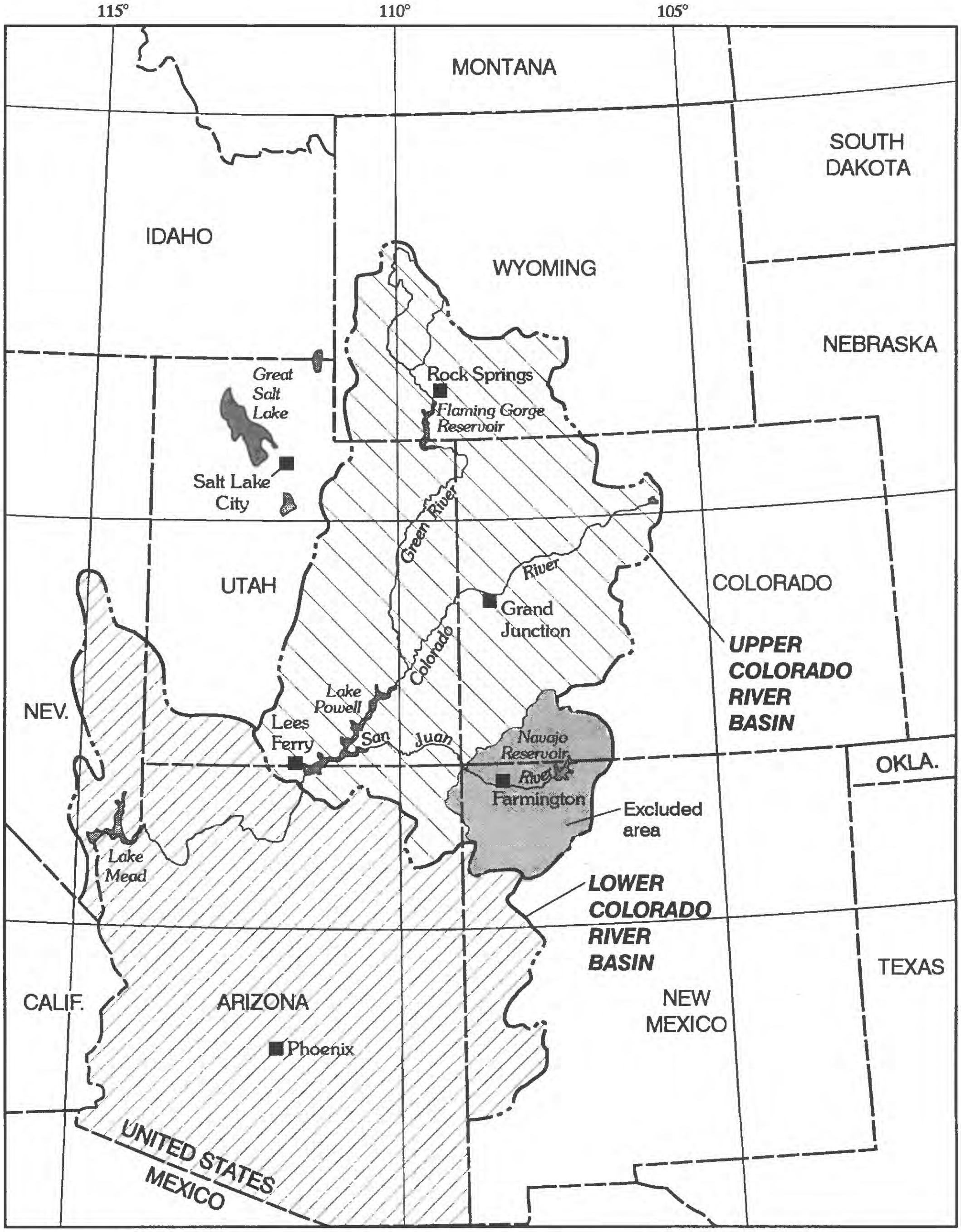
The Green River Basin aquifer system consists of four aquifers and two confining units. Aquifers include the Bridger, Laney, New Fork, and Wasatch-Fort Union aquifers. Confining units are the Wilkins Peak and Tipton units. All units correlate roughly with geologic formations or members of the same names. Hydraulic conductivity was estimated by compiling aquifer-test data and developing a flow model. Hydraulic conductivity is related to the percentage of sandstone present in the Bridger, New Fork, and Wasatch-Fort Union aquifers, and to the degree of fracturing in the Laney aquifer.

Water in the Green River Basin aquifer system moves horizontally from recharge areas along basin margins toward discharge areas, and moves vertically into deeper aquifers. Near the Green and Big Sandy Rivers and Blacks Fork, water leaks upward from deeper aquifers and discharges to streams and associated alluvium. Recharge from precipitation was estimated to be 138 cubic feet per second. Additional recharge of 18 cubic feet per second occurs in the vicinity of Farson and Eden, Wyoming, due to surface water irrigation return flow. Water in the Bridger aquifer moves horizontally toward Blacks Fork. Most water enters the Laney aquifer by upward leakage from the New Fork aquifer and discharges to the Green and Big Sandy Rivers. Water recharges the New Fork aquifer in areas near the Wind River Uplift. Water enters the Wasatch zone of the Wasatch-Fort Union aquifer where the zone is exposed at land surface; the water moves both horizontally and vertically toward discharge areas. Discharge from the Wasatch zone to the Green River and tributaries upstream from Fontenelle Reservoir is approximately 94 cubic feet per second. Smaller quantities of water move vertically into the Fort Union zone of the Wasatch-Fort Union aquifer. Water in the lower part of the Wasatch zone and the Fort Union zone moves toward Flaming Gorge Reservoir and leaks upward through overlying aquifers and confining units at a rate of approximately 13 cubic feet per second.

The Great Divide-Washakie-Sand Wash Basins aquifer system consists of the Wasatch-Fort Union aquifer and an overlying confining unit in the Great Divide, the Washakie, and the Sand Wash Basins. The thickness of the Wasatch zone of the Wasatch-Fort Union aquifer typically ranges from 2,000 to 4,000 feet. Hydraulic conductivity ranges from 0.03 to 9.1 feet per day with large values typical in the Great Divide Basin and along the basins' margins where sandstone is more common. The Fort Union zone underlies the Wasatch zone with thickness typically between 3,000 and 6,000 feet. Hydraulic conductivity in the Fort Union zone is related to the percentage of sandstone present and ranges from 0.001 to 938 feet per day; the median value is 0.02 foot per day. Reliable estimates of the amount of water moving through the system cannot be made with existing data. Recharge areas occur along the basins' margins. Discharge areas are located near the center of the Great Divide Basin, along Bitter and Separation Creeks, and along the Little Snake River.

INTRODUCTION

Historically, streamflow has supplied virtually all the water needs of people in the Upper Colorado River Basin; however, increasing development of energy resources has stimulated interest in the use and management of ground water. Although qualitative basinwide ground-water appraisals have been made by the U.S. Geological Survey, few quantitative appraisals of regional aquifer systems have been undertaken. A comprehensive knowledge of aquifer systems in the Upper Colorado River Basin is needed for the most advantageous development and management of ground-water resources. The U.S. Geological Survey began an extensive study of ground water in the Upper Colorado River Basin (fig. 1) during 1981. The study is part of the Survey's Regional Aquifer-System Analysis (RASA) program. The San Juan Basin is the subject of a separate RASA study and has been excluded from this investigation.



Base modified from U.S. Geological Survey 1:5,000,000 United States base map, 1987

Figure 1. Location of the Upper Colorado River Basin.

PURPOSE AND SCOPE

Characteristics of aquifer systems described in this report about geohydrology of Tertiary rocks include:

1. Geologic characteristics, such as thickness, lateral extent, lithology, and sedimentary character of aquifers and confining units;
2. Hydraulic characteristics, such as hydraulic conductivity and storage coefficient;
3. Characteristics of ground-water recharge, movement, and discharge such as sources and locations of recharge and discharge, volume of water recharging and discharging, and directions of movement (as inferred from potentiometric-surface maps and other data); and
4. Water-quality characteristics, such as the distribution of dissolved solids and ionic ratios, and geochemical trends.

PHYSICAL SETTING

The Upper Colorado River Basin includes the area drained by the Colorado River and its tributaries upstream from Lees Ferry, Ariz., and also includes the Great Divide Basin, a closed basin in Wyoming. The Upper Colorado River Basin comprises an area of about 113,500 mi^2 in parts of Colorado, Utah, Wyoming, New Mexico, and Arizona (fig. 1). The basin is characterized by high rugged mountains, broad basins, and high plateaus that have been deeply entrenched and dissected by streams. Altitudes range from about 3,100 ft above sea level near Lees Ferry, Ariz., to more than 14,000 ft in the mountains of Colorado. The upper part of the San Juan River Basin, although part of the Upper Colorado River system, is not discussed in this report. The upper part of the San Juan River Basin has an area of about 14,600 mi^2. Thus, the study area for this investigation is about 99,000 mi^2.

The climate of the Upper Colorado River Basin is affected more by the movement of the air masses over the mountains than by latitude. The high mountains, particularly on the northern and eastern edges, are comparatively wet and cool; whereas low plateaus and the interiors of basins are dry and subject to large differences in temperature. Average annual precipitation ranges from about 5 inches in some interior valleys to about 50 inches in some mountainous areas (Freethey and Cordy, 1991). Long periods in which average daily temperatures are below freezing are common in mountainous areas.

GEOLOGY

GEOLOGIC SETTING

Consolidated sedimentary rocks of Paleozoic, Mesozoic, and Cenozoic age are present throughout most of the Upper Colorado River Basin. Major rock types include sandstone, limestone, dolomite, and shale, all of which may be aquifers. Sedimentary, igneous, and metamorphic rocks of Precambrian age are exposed in uplifts along the eastern and northern edges of the basin. Unconsolidated alluvium and glacial deposits of Quaternary age occur along streams and in small scattered areas of the river basin. Volcanic rock types, including flows, flow breccias, ash-flow tuffs, and water-laid tuffs, are located in scattered areas of southwestern Colorado and south-central Utah.

Tectonic activity has divided the Upper Colorado River Basin into numerous structural basins, uplifts, and platforms (fig. 2), many of which are bordered or transected by faults. Fracturing has developed in many areas in response to tectonic activity or removal of overburden. Formations typically are deeply buried at basin centers but are exposed along basin margins and uplifts. The tectonic activity in the region has produced frequent lateral changes in depositional environments, resulting in lateral and vertical changes in lithofacies within synchronously deposited sedimentary sequences. A brief summary of tectonic history and resulting sedimentation during the Late Cretaceous Epoch and Tertiary Period will be given in the following paragraphs. This background information is helpful in understanding the classification of aquifers and confining units in rocks of Tertiary age of the Upper Colorado River Basin.

A series of mountain-building movements during Cretaceous time produced highlands west of the Upper Colorado River Basin and subsidence east of the basin. Following each pulse of movement, materials eroded from the western highlands were washed into the eastern seas, resulting in a lateral sequence of eroded highlands, piedmont and coastline fluvial deposits, beach deposits and marine sandstone, marine siltstone and shale, and deep-water limestone. The positions of the various elements of this sequence were controlled by changes in mountain building in the west and rates of subsidence in the area of deposition (Grose, 1972).

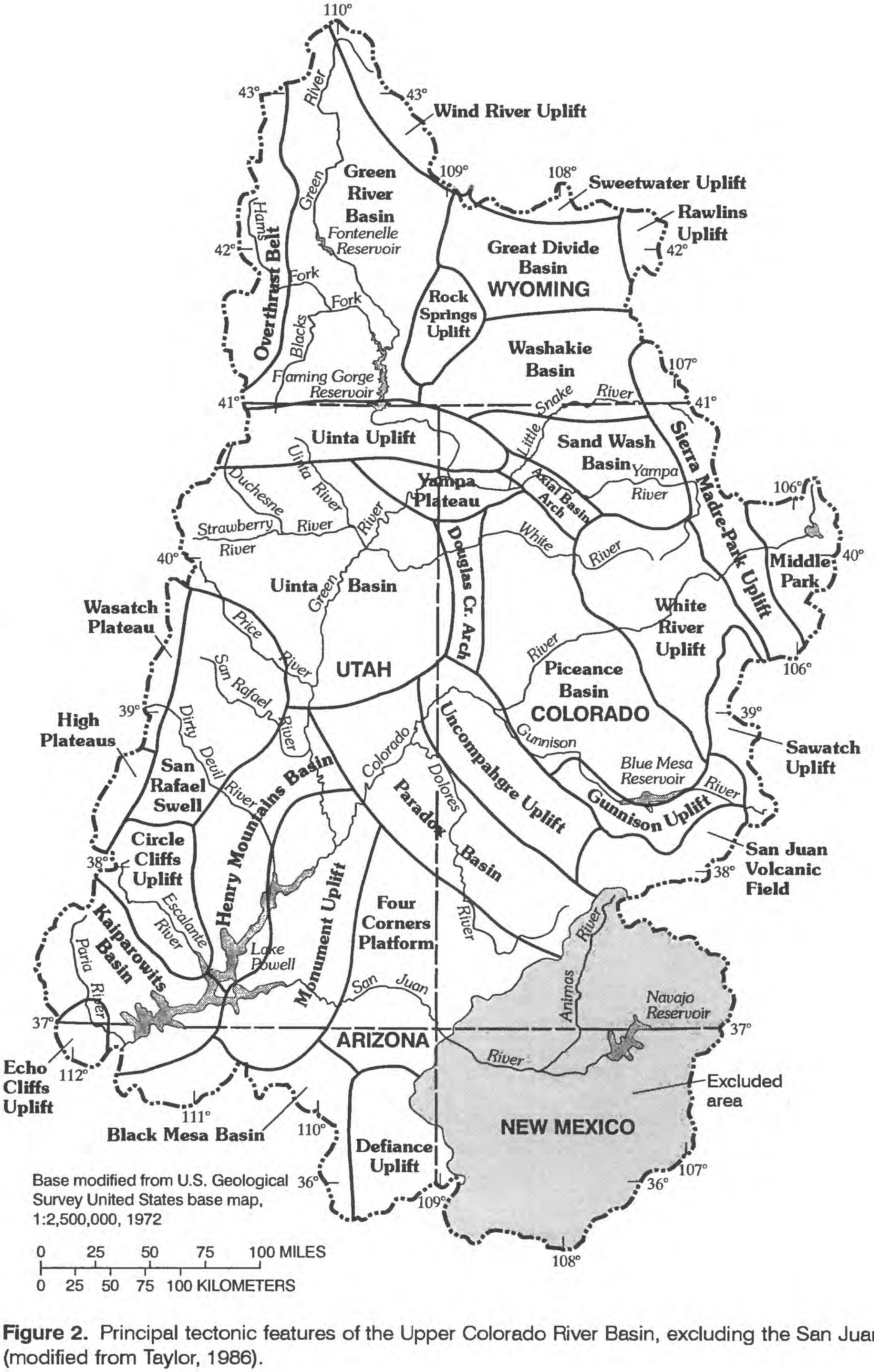


Figure 2. Principal tectonic features of the Upper Colorado River Basin, excluding the San Juan Basin (modified from Taylor, 1986).

A major mountain-building period, apparently unrelated to early tectonic activity, began during Late Cretaceous time, reached its peak during Paleocene to Oligocene time, and continued into the Quaternary (Hansen, 1986). This mountain-building period, the Laramide orogeny, produced the major structural features of the region that have existed to the present. Orogeny did not take place at the same time throughout the region. Generally, mountain building began in the western part of the region and moved progressively eastward.

During Paleocene time, uplift of the Laramide orogeny intensified; sandstone and shale were deposited in the adjacent basins of the study area. The emergence of mountains created internal drainage and ponding in central Utah. Rocks of Paleocene age typically are lake deposits in central Utah, grading eastward into low-energy fluvial deposits. A major drainage system carried sediments from the west and south to a freshwater lake in the Great Divide Basin. Sediments characteristic of high-energy fluvial environments were deposited in the northern Green River Basin, the present north end of the Rock Springs Uplift, and areas adjacent to the Sierra Madre and Rawlins Uplifts. The Piceance Basin probably was the site of high-energy fluvial deposition; however, subsequent erosion removed most of these sediments. Extensive flood plains, swamps, and local ponds developed in the southern Green River Basin, Washakie Basin, and western Sand Wash Basin. By late Paleocene time, the flood plains and swamps extended into the Piceance Basin. Sandstone is the principal rock type of high-energy fluvial environments, and shale and claystone predominate in low-energy flood-plain deposits. Vegetation in swamps was buried and subsequently transformed into coal.

During latest Paleocene time and continuing into early Eocene time, renewed uplifting of mountains was accompanied by subsidence of adjacent basins. The lake in central Utah expanded throughout early and middle Eocene time until it covered virtually all of the Uinta and Piceance Basins. Downwarping of the Green River and Washakie Basins resulted in a change from mixed fluvial, paludal, and lacustrine conditions in early Eocene time to lacustrine conditions in middle Eocene time. The large lake that developed in southwestern Wyoming at times may have been connected to the lake of the Uinta and Piceance Basins. These lakes changed in salinity in response to changing climatic conditions. Large quantities of salts precipitated from the water and were included in the limestone and shale lake deposits. Streams flowing from the adjacent mountains to the lakes deposited fluvial sediments. The relative percentage of sandstone and conglomerate in the fluvial sediments decreases from mountains toward the lakes. Changes in the size of the lakes resulted in complex intertonguing of lacustrine and fluvial sediments. By late Eocene time, the rate of sedimentation exceeded the rate of downwarping and the lakes decreased in size. Deposition, predominately fluvial, continued through the rest of Eocene time and into Oligocene time.

Fluvial sediments of Oligocene, Miocene, and Pliocene age were deposited throughout the region; however, regional uplift and subsequent erosion have removed most of these rocks. Isolated areas of sandstone and conglomerate remain near mountain uplifts, and remnants of siltstone and claystone are found within the basins. Volcanic flows remain in the high plateaus of Utah. Mountain glaciation during Pleistocene time resulted in glacial deposits in uplift areas and terrace deposits along streams. Erosion has dissected the glacial and terrace deposits, leaving some locally important aquifers, but no aquifers of regional importance.

Erosion, which occurred during the end of Tertiary time and continued into Quaternary time, removed large volumes of Tertiary rocks. Most Tertiary rocks that remain are located in the northern half of the Upper Colorado River Basin (fig. 3). Thickness of Tertiary rocks increases from basin margins to about 15,000 ft near the center of some basins. The thickness of Tertiary rocks exceeds 5,000 ft in a substantial part of the northern Upper Colorado River Basin.

GEOLOGIC FRAMEWORK

LATE CRETACEOUS GEOLOGY

The hydrologic system in Tertiary rocks of the Upper Colorado River Basin is affected by geologic and hydrologic conditions within the underlying Upper Cretaceous rocks. A complete description of the regional geohydrology of Cretaceous rocks is given by Freethey and Cordy (1991). However, a summary description of the uppermost regional aquifer within the Cretaceous sequence is helpful in understanding the hydrologic system of Tertiary rocks.

The principal Upper Cretaceous rocks of importance in a description of Tertiary geohydrology include the Mesaverde Group and equivalent rocks, as well as overlying Cretaceous rocks. These rocks have been classified by Freethey and Cordy (1991) as the Mesaverde aquifer and as the North Horn-Mesaverde aquifer by Naftz (1996). In addition to the Mesaverde Group, important formations within the Mesaverde aquifer

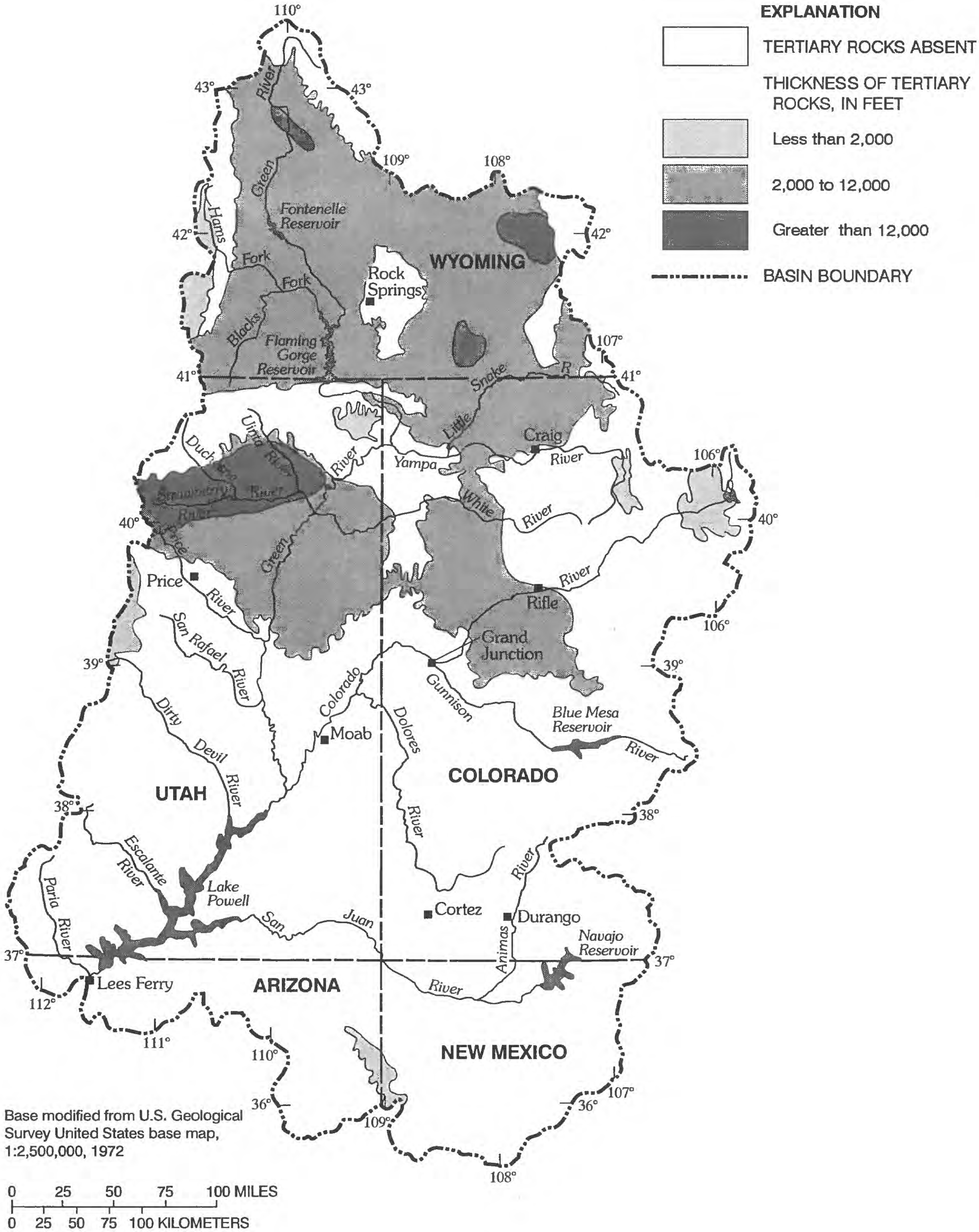


Figure 3. Areal extent and thickness of sedimentary rocks of Tertiary age in the Upper Colorado River Basin.

include the Lance Formation, Fox Hills Sandstone, Lewis Shale, and Adaville Formation (plate 1). Rocks that are both Late Cretaceous and early Tertiary age (North Horn Formation) also are included as part of the Mesaverde aquifer. The Mesaverde aquifer underlies most of the northern half of the Upper Colorado River Basin (fig. 4).

In general, the Mesaverde Group and related formations are thickest where deeply buried in basins and thinnest toward basin margins (fig. 4). In southwestern Wyoming, the average thickness is 4,000 ft, but the Mesaverde is greater than 8,000 ft thick where buried in the Washakie and Great Divide Basins. In northwestern Colorado and northeastern Utah, average thickness is 3,000 ft, but is greater than 7,000 ft in the Piceance Basin. The contact with overlying Tertiary rocks generally coincides with a change in lithology from shallow marine sandstone and shale to fluvial sediments.

Rocks of the Mesaverde Group and related formations represent depositional environments associated with transgressions and regressions of the Late Cretaceous sea. The sediments were deposited in fluvial, deltaic, lagoonal, swampy, and shallow marine environments. As a result, the formations can be characterized by complex lateral and vertical gradations in lithology and frequent intertonguing. In general, the lithology consists of conglomerate, sandstone, siltstone, mudstone, claystone, carbonaceous shale, and coal. In the basins of southwestern Wyoming, the Mesaverde Group and partly equivalent Adaville Formation consist of interbedded sandstone, shale, coal, and carbonaceous shale. In northwestern Colorado and northeastern Utah, the Mesaverde Group is predominately sandstone with interbedded shale and coal.

The Cretaceous formations that overlie the Mesaverde Group in the southeastern part of the Green River Basin differ from the underlying rocks in that they lack carbonaceous shale and coal. The Lewis Shale is primarily marine shale, whereas the Fox Hills Sandstone is a marine to brackish-water sandstone. Interbedded sandstone and shale make up the Lance Formation. In Utah, the Currant Creek and North Horn Formations in the Uinta Basin and the Canaan Peak Formation in the Kaiparowits Basin (fig. 2, not shown on plate 1) are somewhat similar in lithology (conglomerate, sandstone, and shale, but no carbonaceous shale). The Currant Creek Formation is tightly cemented. The Evanston Formation, located in the Overthrust Belt of southwestern Wyoming (not shown on plate 1), is a thick conglomerate, carbonaceous sandstone, siltstone, mudstone, and claystone.

FORT UNION AND WASATCH FORMATIONS AND RELATED FORMATIONS

The geologic framework of the Paleocene and lower Eocene rocks is described collectively because the formations are lithologically similar. In general, the rocks consist of fluvial sediments, and their hydraulic properties depend on the number, thickness, and continuity of sandstone layers and lenses. The most areally extensive formations within the Paleocene and lower Eocene sequence of rock are the Fort Union and Wasatch Formations. These formations are located in the Piceance, Uinta, Green River, Great Divide, Washakie, and Sand Wash Basins, and the high plateaus of Utah. The rocks crop out over large parts of their areal extent.

Lithologically similar rocks make up the Fort Union, Hoback, and Middle Park Formations of Paleocene age; the Coalmont Formation, Flagstaff Limestone known locally in the subsurface of the western Uinta Basin as the Flagstaff Member of the Green River Formation, the main body of the Wasatch Formation of Paleocene and Eocene age; and the Pass Peak Formation of Eocene age. The Fort Union, Hoback, and Middle Park Formations are characterized by interstratified sandstone, mudstone, shale, and coal beds. The Wasatch, Coalmont, and Pass Peak Formations consist of fluvial, mostly piedmont, deposits. The fluvial deposits are coarse grained near the mountains from which the sediments were derived and become more fine grained with increasing distance from the mountains. Coarse-grained fluvial deposits include lenticular sandstone and conglomerate. The Flagstaff Limestone or stratigraphically equivalent Flagstaff Member, where it is a part of the Green River Formation, are lake deposits, but fractures and solution channels have enhanced hydraulic conductivity where the formations are near land surface.

The sequence of Paleocene and lower Eocene rocks is overlain by the Green River Formation. The contact generally coincides with a change in lithology from predominately fluvial sandstone and interbedded fine-grained material to predominately fine-grained lakebeds. In most areas of the Upper Colorado River Basin, the contact corresponds with the base of the oldest member of the Green River Formation. As a result of changes in the size of the Eocene lakes, there is substantial intertonguing of the Wasatch and Green River Formations. Geologic units such as the New Fork Tongue and the Renegade Tongue of the Wasatch Formation, and the Douglas Creek Member of the Green River Formation contain substantial amounts of sandstone and are important basin-wide aquifers. A lithologic description of intertonguing geologic units is

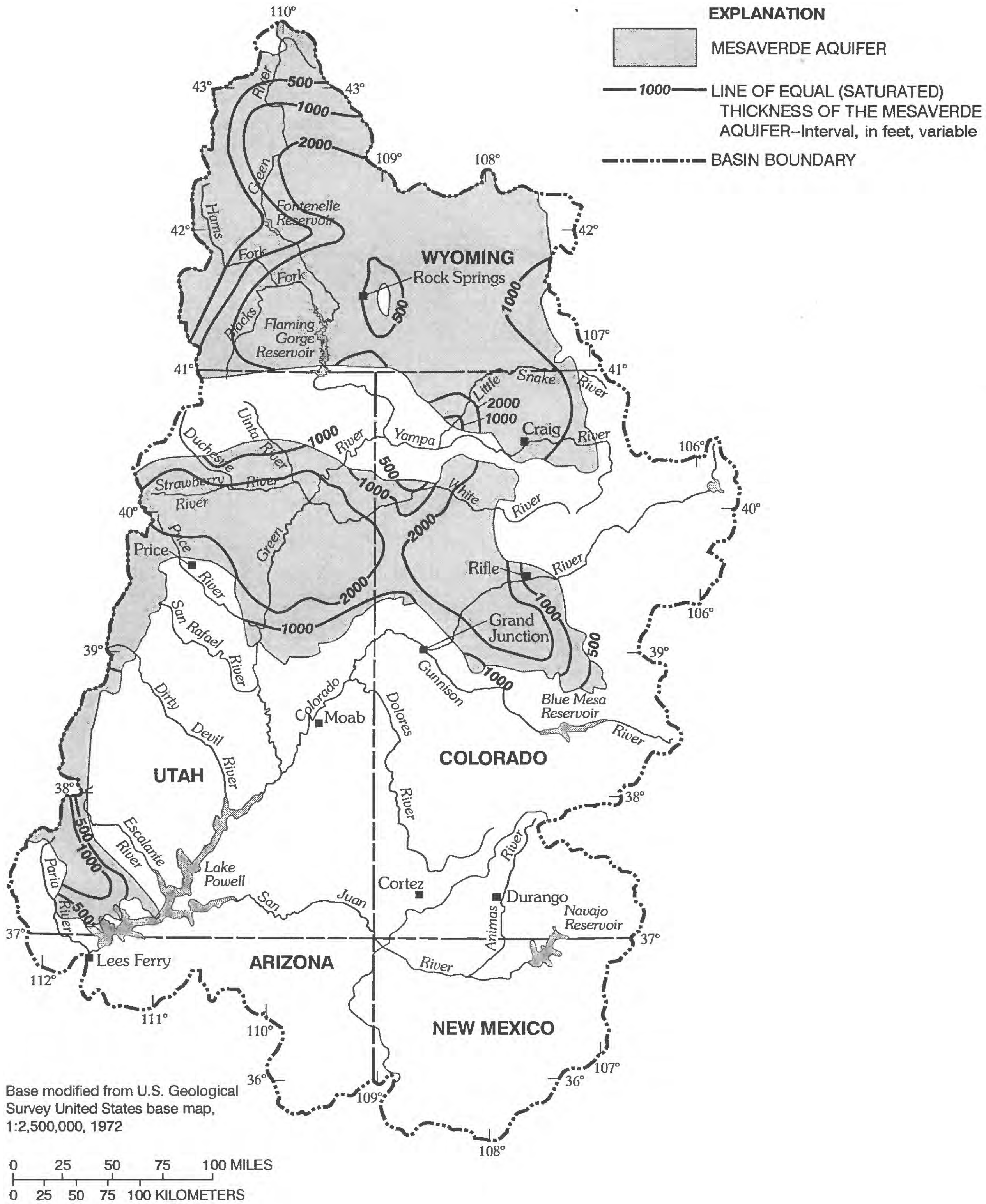


Figure 4. Areal extent and thickness of the Mesaverde aquifer (from Freethey and Cordy, 1991).

given in the following section of this report entitled "Green River Formation and Tongues of the Wasatch Formation."

Rocks in the Fort Union Formation, Wasatch Formation, and related formations have been classified into a number of lithofacies (Pipiringos, 1961; Masursky, 1962; Colson, 1969; Curry, 1969 and 1973; and Dribus and Nanna, 1982). Boundaries between facies generally are gradational and, therefore, somewhat arbitrary. Mapping of lithofacies extent and thickness at a regional scale is complicated by the large number of Paleocene and lower Eocene formations that have been identified. In many places, geologic data are insufficient to accurately describe subsurface relations between lithofacies. Maps of extent and thickness of aquifers and confining units within the Paleocene and lower Eocene sequence of rock are presented in the section of this report entitled "Basin Aquifer Systems." The following lithologic description is helpful in understanding the designation of aquifers and confining units given in that section of the report. Lithofacies maps are not presented in this report. However, maps of estimated hydraulic conductivity, presented later in this report, typically reflect variations in lithology.

The Flagstaff Limestone, known locally in the subsurface of the western Uinta Basin as the Flagstaff Member of the Green River Formation, is present in the Uinta Basin and throughout central Utah (Weiss and others, 1990). The sequence typically is 300 to 800 ft thick, but thins eastward in the Uinta Basin. The Flagstaff loses its identity east of the Green River. Basically a freshwater limestone, the sequence locally includes layers of sandstone, shale, gypsum, volcanic ash, and oil shale. Along western margins of the Uinta Basin, the Flagstaff consists of thick-bedded limestone with interbedded shale and sandstone. The units grade eastward into shaley facies with interbedded limestone and sandstone. The Flagstaff is overlain in most areas by the Wasatch Formation. Along the western margin of the Uinta Basin, the Wasatch Formation is absent and the Flagstaff is overlain by a shale facies of the Green River Formation.

The Wasatch Formation within the Uinta Basin consists primarily of shale. In the southwestern part of the basin, interbedded calcareous and silty shale predominate, although there is some fine-grained sandstone. Fluvial redbeds of fine-grained sandstone and mudstone are common in the south-central part of the Uinta Basin. Sandstone and conglomerate predominate adjacent to the Uinta Uplift; however, the basinward extension of this facies is small. Elsewhere, silty shale and mudstone predominate.

The Fort Union Formation and Wasatch Formation are present within the Piceance Basin. The Fort Union Formation is limited to a thin layer of coarse-grained material. North of the White River, the Fort Union thickens and becomes fine-grained. The Wasatch Formation is as much as 5,000 ft thick and consists primarily of shale with interbedded fluvial sandstone. Sandstone in the Wasatch Formation is lenticular and, with local exception, makes up a minor part of the total lithology. South of the Colorado River, Donnell (1969) classified the Wasatch Formation into three members. The middle member includes an arkosic, ledge-forming sandstone.

The Fort Union Formation accumulated to great thickness within the basins of southwestern Wyoming; however, the quantity of sandstone in the formation varies both areally and vertically. Thickness exceeding 3,000 ft is common, with the greatest thickness being in the Great Divide and northern Green River Basins. Sandstone is the predominate rock type within the depositional trough of the northern Green River Basin. Aquifer tests indicate the formation has low hydraulic conductivity where deeply buried, possibly due to compaction and cementation of the sandstone. Mixed depositional environments in the southern part of the Green River Basin resulted in fluvial sandstone imbedded within a matrix of siltstone, claystone, and shale. Within the Great Divide, Washakie, and Sand Wash Basins, the predominant sequence of rocks is interbedded sandstone of variable thickness with siltstone, shale, lignite, and coal. A massive wedge of sandstone and conglomerate is located near the base of the Fort Union Formation. The basal sandstone thins westward from outcrops along the eastern side of the Great Divide, Washakie, and Sand Wash Basins and cannot be distinguished along the Rock Springs Uplift. In the east-central Great Divide Basin, the Fort Union Formation consists of thick lacustrine shale. A broad east-west trending sequence of thick sandstone, siltstone, shale, and coal beds is present in the north-central Great Divide Basin. The sequence persists westward across the north end of the Rock Springs Uplift and into the northeastern Green River Basin.

The Wasatch Formation overlies the Fort Union Formation in the basins of southwestern Wyoming. The Wasatch consists of siltstone and sandy shale with varying thickness of channel sandstone. The percentage of sandstone generally decreases with distance from uplifts. Thick arkosic sandstone predominates in the northern Green River and eastern Sand Wash Basins. Thick sandstone beds also predominate in the Green River Basin adjacent to the Rock Springs Uplift. The Wasatch Formation is conglomeratic adjacent to the Overthrust Belt and Uinta Uplift. Elsewhere within the Green River Basin, the Wasatch Formation consists of interbedded shale, siltstone, and sandstone; the original sediments typically deposited under mixed fluvial conditions. Within the Great Divide Basin, the Wasatch Formation was deposited under paludal conditions; thick coal beds are interbedded with thick sandstone. The Wasatch is at land surface over most of the western part of the Great Divide Basin but underlies and interfingers laterally with the Battle Spring Formation in the eastern part of the Great Divide Basin. The Battle Spring Formation is predominately arkosic sandstone. Within the Washakie Basin, the Wasatch Formation typically consists of sandy mudstone with sandstone lenses occurring in clusters. Carbonaceous rock and coal is present in the upper part of the formation.

GREEN RIVER FORMATION AND TONGUES OF THE WASATCH FORMATION

The Green River Formation is principally a fine-grained lacustrine deposit with small hydraulic conductivity, except where the rock has been altered by fractures or solution channels. The formation intertongues with the underlying and laterally continuous Wasatch Formation. In general, the overlying Bridger, Uinta, Wagon Bed, and Washakie Formations mark a change from the primarily lake deposits of the Green River Formation to primarily fluvial deposits. In most areas the contact is gradational and intertonguing occurs. Donnell (1961), Bradley (1964), and Cashion (1967) provide detailed descriptions of the geology of the Green River Formation and related Eocene rocks. A summary is provided in this report.

In the Piceance Basin, the Green River Formation has been divided into four members. The basal Douglas Creek Member is a fine-grained sandstone with limestone and some interbedded shale. Although this lithologic description suggests that the member is an aquifer in the Piceance Basin, spring and well yields are small, indicating that hydraulic conductivity also may be small. The overlying Garden Gulch Member is a gray fissile shale with interbedded marlstone, some fine-grained sandstone, and limestone. The Anvil Points Member is a lateral equivalent of the Douglas Creek and Garden Gulch Members. The Anvil Points Member, which is a lakeshore facies that crops out along the eastern margins of the basin, is not extensive in a basinward direction. The Anvil Points Member yields negligible quantities of water to wells and springs. The Parachute Creek Member is a kerogenous, dolomitic marlstone ranging from 500 to 1,700 ft thick. The member has been divided into four zones: A basal unit of oil shale with some zeolite mineralization that is relatively unfractured, a leached zone of oil shale that is fractured and more permeable than overlying or underlying zones, the Mahogany zone of oil shale and saline facies that is fractured where it is near land surface, and an upper zone of fractured marlstone that contains little oil shale.

The Green River Formation within the western and southern Uinta Basin has been divided in ascending order into the Douglas Creek, Garden Gulch, and Parachute Creek Members, but no members have been designated in the western and northern Uinta Basin. The Renegade Tongue of the Wasatch Formation intertongues with the Douglas Creek Member of the Green River Formation. The informally designated black-shale facies (Picard, 1955) forms the basal unit of the Green River Formation in the western and north-central Uinta Basin. The facies is a gray to black dolomitic and calcareous shale that is thinly laminated. Extensively distributed in the subsurface, the black-shale facies thins abruptly to the south and east, where it is indistinct in outcrops. The Douglas Creek Member is a series of predominantly marginal lacustrine deposits of fine-grained sandstone, siltstone, claystone, and limestone. Discontinuous channel sandstone is common. Thickness ranges from 200 ft to 1,300 ft. The Renegade Tongue is a fluvial deposit of massive, irregularly bedded sandstone and siltstone with a thickness of 1,000 ft or less. The Douglas Creek Member and Renegade Tongue have been mapped along the southern and eastern margins of the Uinta Basin and grade laterally into predominately shale with some sandstone and limestone in the center of the basin. The Garden Gulch Member is present only in the eastern part of the Uinta Basin, where it consists of marlstone with siltstone and oil shale. The Parachute Creek Member thickens from southeast to northwest and is about 2,000 ft thick in the center of the basin. In the northern and western parts of the basin, the member merges with undifferentiated Green River Formation. Lithology of the Parachute Creek Member depends on position relative to depositional shorelines. The Parachute Creek Member grades basinward from mostly sandstone and siltstone with little oil shale to predominately marlstone and oil shale.

The diagrammatic section in figure 5, modified from Bradley (1964), shows the relation between members of the Green River Formation and tongues of the Wasatch Formation in southwestern Wyoming. The basal unit of this intertonguing sequence of rock is the Luman Member of the Green River Formation. The member is a low-grade oil shale, siltstone, and mudstone located throughout the Washakie Basin, parts of the Great Divide Basin, and in a narrow east-west band of the southern Green River Basin. The Luman Member is less than 400 ft thick. The Niland Tongue of the Wasatch Formation is a fluvial siltstone and mudstone

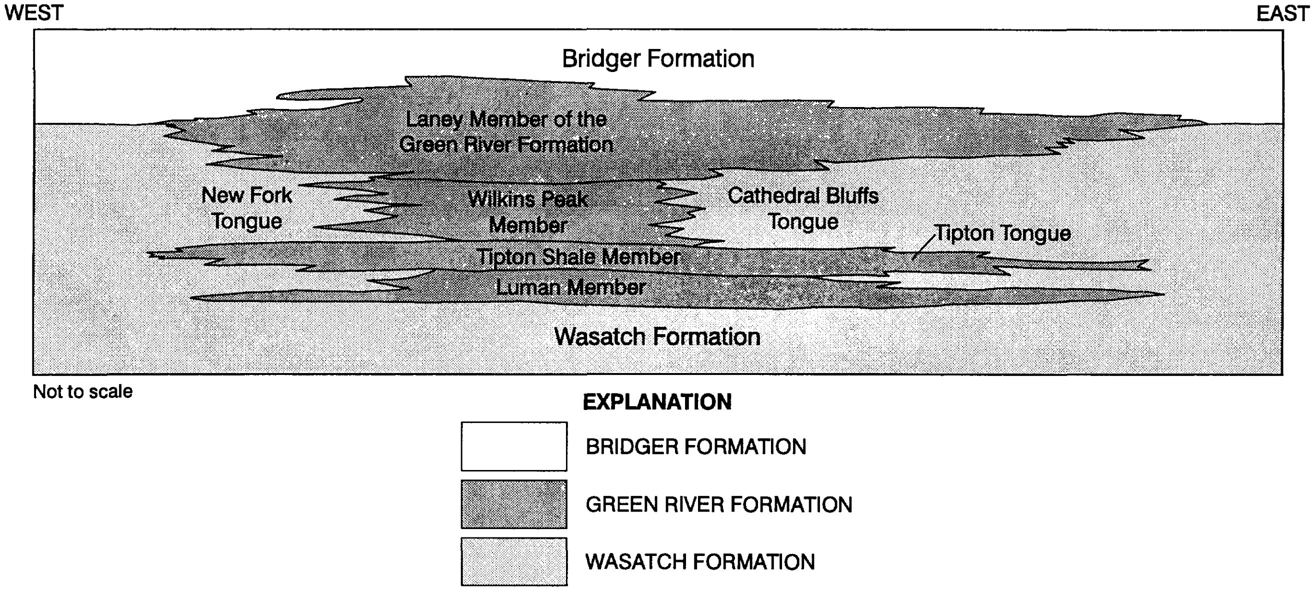


Figure 5. Diagrammatic section showing the intertonguing of rocks in the Green River Formation and Wasatch Formation in southwestern Wyoming (modified from Bradley, 1964).

located approximately in the same areas as the Luman Member. The Tipton Shale Member of the Green River Formation, located in the Green River Basin, consists of less than 200 ft of low-grade oil shale, marlstone, and mudstone. The stratigraphically equivalent Tipton Tongue of the Green River Formation, located in the Washakie Basin, consists of soft fissile shale and flakey marlstone with thin beds of limy sandstone. The Wilkins Peak Member of the Green River Formation is located in the Green River Basin and western part of the Washakie Basin. The Wilkins Peak Member consists of dolomite and thick beds of the saline mineral, trona, which is deposited in the eastern part of the Green River Basin. The New Fork and Cathedral Bluffs Tongues of the Wasatch are approximate lateral equivalents with the Wilkins Peak Member. The New Fork Tongue, located in the northern Green River Basin, is a sandy mudstone that contains numerous lenses and irregular beds of fine- to coarse-grained sandstone. The Cathedral Bluffs Tongue is predominately a claystone and shale, containing coarse arkosic sandstone in the Great Divide Basin and fine-grained sandstone in the Washakie and Sand Wash Basins. The Laney Member of the Green River Formation contains interbedded marlstone, limestone, shale, tuff, and sandstone where present in the Green River Basin. Fractures and solution channels have greatly increased hydraulic conductivity of the Laney Member near the Big Sandy River. In the Washakie Basin, the Laney Member is a chalky to muddy marlstone and shale with only local areas of extensive fracturing.

UINTA AND BRIDGER FORMATIONS AND RELATED FORMATIONS

The Uinta and Bridger Formations and related formations overlie the Green River Formation and mark a change from primarily lacustrine sediments to primarily fluvial sediments. Formations important to the regional geohydrology include the Uinta Formation in the Piceance Basin, the Uinta and Duchesne River Formations in the Uinta Basin, the Bridger Formation in the Green River and Great Divide Basins, and the Washakie Formation in the Washakie Basin. These formations crop out over much of their areal extent but are overlain locally by Miocene and Oligocene rocks (plate 1). The lithology of the Uinta and Bridger Formations and related formations is characterized by a mixture of sandstone, siltstone, shale, and some marlstone. The percentage of sandstone generally increases toward the top of the formations.

Lithology of the Uinta Formation in the Piceance Basin varies greatly but is similar to the Uinta Formation in the Uinta Basin. The formation consists of poorly sorted, coarse- to fine-grained sandstone, siltstone, and some marlstone. The porosity of the rock matrix is negligible. The lower part of the formation is extensively fractured. The percentage of coarse-grained material increases toward the top of the Uinta Formation.

With an average thickness of about 2,500 ft, the lithology of the Uinta Formation in the Uinta Basin varies both areally and vertically. Along the northwestern margin of the basin, the formation is a massive boulder conglomerate. The conglomerate grades abruptly into sandstone and mixed sandstone and shale with distance from the Uinta and Wasatch Uplifts. The Uinta Formation has been divided into three units in the eastern part of the basin (Cashion, 1967). A basal unit of 400 to 1,100 ft in thickness is a resistant sandstone with interbedded shale that thins westward and basinward. West of the Green River, the basal unit grades into lakebed deposits. A middle unit, 300 to 500 ft thick, is a lenticular and interbedded medium-grained to coarse-grained sandstone and claystone. The upper unit also consists of sandstone and claystone and is 600 to 800 ft thick. An 1,100-ft-thick lakebed deposit at the base of the formation is oriented along an east-west trough, about 30 mi wide, in the center of the basin. The limy to dolomitic shale contains substantial quantities of saline minerals. The lakebed deposit grades upward into a mixture of fine-grained calcareous sandstone and limestone which, in turn, grades upward into a fluvial facies composed of red shale, siltstone, and medium- to coarse-grained lenticular sandstone. The lithology of the upper Uinta Formation generally is consistent throughout the basin.

The lithology of the Duchesne River Formation is similar to fluvial facies of the Uinta Formation but with a greater percentage of sandstone and conglomerate. Grain size generally decreases with distance from the Uinta Uplift. Conglomerate and coarse-grained sandstone are common in the northern part of the Uinta Basin, but poorly cemented sandstone and shale predominate in the center of the basin.

The Bridger Formation in the Green River and Great Divide Basins and the Washakie Formation in the Washakie Basin generally are sandy tuffaceous mudstone. The rocks often form a badland topography in outcrops. Interbedded tuffaceous sandstone and volcanic ash are common. As much as 15 to 20 percent of the total rock material in some areas is ash.

GEOHYDROLOGIC UNITS

No regionally continuous aquifers or confining units extend throughout the Tertiary rocks in the structural basins that compose the Upper Colorado River Basin. Breaks in the continuity of stratigraphically equivalent units due to uplifts, together with other local differences in the lithology and hydraulic properties of these units, preclude their being geohydrologically lumped together. Within individual structural basins, however, or among several physically linked basins, aquifers and confining units are continuous and basin hydrologic units have been defined. Ground water moves exclusively along flow paths within these basins. The Tertiary rocks of the Upper Colorado River Basin are considered to comprise four separate aquifer systems. These systems, the Piceance Basin, the Uinta Basin, the Green River Basin, and the Great Divide-Washakie-Sand Wash Basins aquifer systems will be discussed following the "Data Assembly and Analysis" section of the report.

Late Cretaceous sandstones form the Mesaverde aquifer (Freethey and Cordy, 1991). Although the lithologic character of the Mesaverde aquifer is not as uniform as other Mesozoic aquifers, sandstone lenses within the sequence of rocks are sufficiently interconnected to act as a regionally extensive aquifer. The focus of this report is on rocks of Tertiary age in the Upper Colorado River Basin. Therefore, this report discusses those aspects of the Mesaverde aquifer that affect vertical ground-water movement between aquifers in Tertiary rocks and the Cretaceous Mesaverde aquifer.

Previous studies of the Piceance Basin were fostered by the potential for development of oil-shale as an energy source. An early study characterized the geohydrology of the basin (Coffin and others, 1971). Additional studies included simulation of the effects of oil-shale development on basin hydrology (Weeks and others, 1974), and simulation of the hydrogeochemistry and solute transport expected during oil-shale development (Robson and Saulnier, 1981). Later studies included modeling of the Piceance Basin (Taylor, 1982) and simulating mine drainage in the basin (Taylor, 1986).

Previous geohydrologic reports in the Uinta Basin primarily describe water in the rocks of the Duchesne River Formation and the Uinta Formation. In the present report, these formations have been combined in the Duchesne River-Uinta aquifer. Earlier reports included reconnaissance studies of water supply in the southern and northern parts of the basin (Price and Miller, 1975; Hood, 1976; and Hood and Fields, 1978). The lateral boundaries of these various studies and that of a hydrologic model of the Duchesne River-Uinta aquifer (Glover, 1996) have not coincided, but the aquifers in the Tertiary rocks have been well characterized by these reports.

Previous geohydrologic reports in the Green River Basin include a reconnaissance report by Welder (1968), a compilation of ground-water data (Zimmerman and Collier, 1985), and a report on the occurrence and use of ground water in the Green River Basin (Ahern and others, 1981). In addition, Naftz (1996) investigated the water-quality characteristics of the aquifers in Tertiary rocks in the Green River Basin as part of the regional aquifer-system analysis of the

Upper Colorado River Basin. Reports on specific areas include a U.S. Bureau of Reclamation study on the Farson-Eden irrigation area (Barker and Sapik, 1965), a description of ground-water resources in the vicinity of Lyman, Wyo. (Robinove and Cummings, 1963), and study of an in situ oil-shale retort area near Rock Springs, Wyo. (Glover, 1986). Previous geohydrologic reports in the Great Divide and Washakie Basins and adjacent areas include an early reconnaissance level study of the ground water by Welder and McGreevey (1966). Later reports described the plan of study for the regional aquifer-system analysis (Taylor and others, 1983). A still later report described the broad Tertiary geohydrologic framework of the Upper Colorado River Basin (Taylor and others, 1986).

DATA ASSEMBLY AND ANALYSIS

The data used to prepare this report were obtained from many sources. The methods used to analyze these data to characterize the geohydrology of Tertiary rocks in the Upper Colorado River Basin are summarized in this section.

HYDRAULIC CONDUCTIVITY

Hydraulic conductivity of aquifers and confining units in Tertiary rocks was estimated primarily from aquifer and specific-capacity tests conducted in water wells, drill-stem tests conducted in oil-and-gas wells, and digital ground-water models developed at the scale of basin aquifer systems. Analysis of apparent ages of ground water provided hydraulic-conductivity estimates in the northern Green River Basin.

Freethey and Cordy (1991) described determination of hydraulic conductivity from field data for Mesozoic aquifers of the Upper Colorado River Basin utilizing some of the same methods used here in aquifers in Tertiary rocks: aquifer tests, specific-capacity tests, and drill-stem tests. These methods are used to analyze part of the penetrated aquifer that surrounds a well or array of wells and not the entire extent of the aquifer. Each method has advantages and disadvantages regarding reliability of results and cost.

Hydraulic conductivity values calculated from the results of aquifer tests that last several weeks are most reliable. However, even hydraulic conductivity values derived from such long-duration tests may not represent the part of an aquifer that has the largest water-yielding capabilities. Few aquifer tests have been conducted in Tertiary rocks in the Upper Colorado River Basin. One test per basin aquifer is typical; no tests have been conducted in some basin aquifers. The time and expense of designing and conducting an aquifer test seldom is justified unless a substantial water-related problem exists. Interest in oil-shale development has resulted in several multiple-well aquifer tests of the upper and lower Piceance Basin aquifers. Aquifer-test results were reviewed as part of the Upper Colorado River Basin RASA study, but no independent interpretation was attempted. No new aquifer tests were conducted as part of the RASA study of Tertiary rocks. Reports describing the analysis of aquifer tests in the Piceance Basin aquifer system include Coffin and others (1968), Weeks and others (1974), Dale and Weeks (1978), Hood (1976), and Holmes (1980).

Specific-capacity tests commonly are used to estimate hydraulic conductivity for shallow aquifers not deeply buried by younger rocks. A comparison of hydraulic conductivity values derived from specific capacities with the hydraulic conductivity values determined from aquifer tests and laboratory tests shows that hydraulic conductivity values derived from specific-capacity tests usually are within about one order of magnitude of values derived by the other methods. Everitt Zimmerman (U.S. Geological Survey, written commun., 1984) estimated hydraulic conductivity from specific-capacity data in southwestern Wyoming using the method of Theis and others (1963).

Drill-stem tests are performed by the petroleum industry on deep formations that are possible sources of oil or gas. Drill-stem test results can be used to calculate freshwater head and hydraulic-conductivity values. Hydraulic conductivity calculated using drill-stem test results generally are smaller than values determined from laboratory tests. Possible reasons for these small values may be related to the depth of the formations and the short time of test duration. The deep formations may be compressed; pore space considerably decreased due to pressure from the weight of overlying rocks. Drill-stem tests are characteristically 1 to 2 hours long, which is not enough time to incorporate the effect of widely spaced fractures in the aquifer. Drill-stem tests of selected rocks in the Upper Colorado River Basin were analyzed by Teller and Chafin (1986); selection was based on certain quality criteria indicating that the test data were representative of formation conditions. Intrinsic permeability was estimated by the Horner graphical method (Horner, 1952; Bredehoeft, 1965). Hydraulic conductivity was calculated from intrinsic permeability using measured fluid temperature to estimate fluid viscosity.

More than 150 estimates of hydraulic conductivity are available from drill-stem tests of Tertiary rocks in the Upper Colorado River Basin. Virtually all tests were conducted in wells drilled into deeply buried parts of the Green River, Wasatch, and Fort Union Formations. Most tests were conducted in small areas associated with the production of oil and gas. Selection of drill-stem test intervals tends to be biased toward deeply buried sandstone lenses bounded by shale or other very low permeability rock.

Numerical models of ground-water flow were used to estimate basin distribution of hydraulic conductivity. Basin-scale ground-water models have been used within the Upper Colorado River Basin to simulate long-term or steady-state hydrologic conditions. Models of basin flow in Tertiary rocks have been developed for the Uinta Basin aquifer system (Holmes and Kimball, 1987; Glover, 1996), the Piceance Basin aquifer system (Weeks and others, 1974; Robson and Saulnier, 1981; Taylor, 1982, 1986), and the Green River Basin aquifer system (Martin, 1996).

As described in the referenced reports, the basin models of Tertiary aquifer systems in the Upper Colorado River Basin adequately simulate known hydrologic conditions and can be used to estimate basinwide distributions of hydraulic conductivity. Where possible, model-estimated distributions of hydraulic conductivity have been qualified by statistical measures of precision. Values of hydraulic conductivity estimated on the basis of modeling represent averages over distances greater than or equal to the node spacing in the models, the distance over which the ground-water equations are approximated. In basin models, the node spacing typically is between 1 and 10 miles. Estimates obtained by modeling may not correspond to values obtained from analysis of aquifer tests and drill-stem tests that represent much more localized conditions in the aquifer.

Corrected carbon-14 ages of selected water samples were used by Chafin and Kimball (1992) to estimate ground-water velocities along projected flow paths in the Wasatch zone of the Wasatch-Fort Union aquifer within the Green River Basin. The methods used to correct carbon-14 ages in ground water are beyond the scope of this paper, but are described by Wigley and others (1978).

Age of the ground water was used to estimate average ground-water velocities and hydraulic conductivities. The formula used to calculate flow rates along flow paths perpendicular to the potentiometric contours is:

$$\text{Average linear velocity } (\bar{V}) = \frac{\text{flow path distance (in feet)}}{\text{age of ground water (in days)}}.$$

Hydraulic conductivities were calculated according to the formula

$$K = \frac{\bar{V}\theta}{I},$$

where K = hydraulic conductivity, in feet per day;
$\bar{V}$ = average linear velocity, in feet per day;
θ = effective porosity, dimensionless; and
I = hydraulic gradient, dimensionless.

HYDRAULIC HEAD

Hydraulic-head data are available from water-level measurements in wells and from pressure measurements during drill-stem tests. Static water-level data obtained from water wells are most common in aquifers that are at or near land surface. Pressure measurements obtained during drill-stem tests are the primary source of data for deeply buried aquifers. Accuracy must be considered when using either type of data; however, water-level measurements in wells usually are more accurate than water levels determined from pressure measurements in drill-stem tests. However obtained, hydraulic-head data typically are much more accurate than other geohydrologic data. Because the measurements are relatively easy to make, hydraulic-head data also typically are the most plentiful geohydrologic data available.

The distribution of hydraulic head in an aquifer system is indicated by potentiometric-surface maps for each aquifer. A potentiometric surface is defined as a surface connecting points to which water would rise in tightly cased wells open to the same aquifer. If vertical head gradients are substantial within an aquifer, several potentiometric-surface maps for a series of geologic strata may be needed to define the spatial distribution of head. The thickness selected for each hydrogeologic unit would be based on the need to minimize vertical hydraulic-head differences in the unit, thereby improving the accuracy of the potentiometric-surface maps. In contrast to the need for mapping along a relatively thin unit is the need for sufficient hydraulic-head data to describe horizontal distributions of hydraulic head.

Although vertical head gradients within some aquifers in Tertiary rocks of the Upper Colorado River Basin can be substantial, adequate hydraulic-head data are not available on a basinwide basis to compile multiple potentiometric-surface maps for units within individual aquifers. A single potentiometric-surface map for each aquifer is considered to be adequate to determine general directions of ground-water flow areally, as well as to identify general areas of upward or downward leakage. Where possible, potentiometric-surface maps were compiled using water-level measurements in wells that penetrate the entire saturated thickness of an aquifer. This approach was most effective in the Piceance Basin aquifer system. The accuracy of potentiometric-surface maps generally is indicated by the contour interval. A rule of thumb is that accuracy is roughly one-half the contour interval.

The potentiometric surfaces of the aquifers in Tertiary rocks discussed in this report show virtually no pumping depressions, anomalies, or other evidence of broad-scale stress to the ground-water systems. Therefore, the various basins were analyzed as steady-state systems.

GROUND-WATER RECHARGE AND DISCHARGE

Direct measurement of ground-water recharge and discharge rarely is possible. Of the two, discharge can be measured more easily. A variety of indirect methods for estimating recharge and discharge have been devised. Accuracy of the methods varies widely and, in many situations, accuracy cannot be quantified. The general approach used to estimate recharge to and discharge from aquifer systems in Tertiary rocks under pre-pumping steady-state included:

1. Making initial estimates of long-term recharge based on empirical methods or results of watershed-modeling studies;
2. Independently making initial estimates of steady-state discharge based on gain-and-loss studies of streamflow and spring-discharge measurements;
3. Interpreting ground-water-quality data in order to confirm and refine boundaries of steady-state recharge areas; and
4. Revising estimates of recharge and discharge in the course of developing steady-state models of ground-water flow.

In general, ground-water-flow models were developed only for basin aquifer systems for which estimates of ground-water discharge were believed to be reasonably accurate.

Initial estimates of ground-water recharge were made by using a method developed by Eakin and others (1951) and modified by Hood and Waddell (1968). The method is based on an empirically derived assumption that recharge can be estimated as a percentage of average annual precipitation. The actual percentage used as the estimate of recharge is derived in a somewhat subjective manner to account for changes in surface geology, physiography, and seasonal patterns in precipitation. In areas receiving less than 10 inches of annual precipitation, recharge is assumed to be negligible. When using the method of Hood and Waddell (1968), recharge from streams, diversion canals, or surface irrigation is not estimated separately. The percentage values used in estimating recharge from precipitation are adjusted to account for these sources of water. The method has been applied with apparent success in the Uinta Basin by Price and Miller (1975) and Hood and Fields (1978), and in the Green River Basin by one of the authors of this report.

The empirical method of Eakin and others (1951) was developed originally for estimating recharge and discharge in basins in east-central Nevada. Percentage values used in estimating recharge were balanced by trial-and-error against separately derived estimates of discharge in 13 valleys. Recharge in these valleys occurred principally through carbonate, igneous, and metamorphic rocks and large alluvial fans. Hood and Waddell (1968) recognized the need to adjust recharge-percentage values to account for variations in topography and geology. Hood and Waddell (1968) estimated recharge in the Skull Valley, southwest of Salt Lake City, by three separate methods and obtained estimates within 65 percent of recharge estimated by the method of Eakin and others (1951). Discharge estimates were similar to recharge estimates, both in magnitude and the degree of uncertainty.

The previously referenced investigations indicate that recharge estimates based on empirically derived relations between recharge and precipitation should be considered as first-order approximations subject to refinement and calibration. Refinement and calibration are particularly important when applying the method in basins that differ significantly from the basins originally studied by Eakin and others (1951). Significant differences include those in lithology or physiography, presence of numerous diversion canals, and differences in seasonal precipitation patterns. Ground-water flow models have been used in the Uinta and Green River Basin aquifer systems to refine and calibrate empirically derived estimates of recharge. Using model development as a calibration exercise has the advantage of providing a check on the compatibility of recharge, discharge, hydraulic-conductivity, and hydraulic-head distributions.

Estimating of ground-water recharge as a percentage of precipitation lacks a convincing physical basis. An alternative method, used in the Piceance Basin, was to estimate recharge by watershed modeling (Weeks and others, 1974). Watershed modeling is an attempt to simulate precipitation-runoff relations based on physical principles. Model parameters include the distribution of soils and soil properties; vegetation type; land-surface altitude, slope, and aspect; and stream-channel characteristics. Watershed models simulate rainfall, snowpack accumulation and melting, runoff, soil infiltration, and deep percolation (ground-water flow that is not discharged locally to streams). The rate of water leaving the watershed as deep percolation rarely is known when developing a watershed model. Therefore, deep-percolation rates are treated as model-calibration parameters. A complete description of watershed modeling in the Piceance Basin is outside the scope of this report; the model is described by Weeks and others (1974).

Rates of deep percolation obtained by watershed modeling have been used as initial estimates of ground-water recharge in the Piceance Basin aquifer system. Watershed modeling was conducted only in the Piceance Creek drainage. In order to estimate recharge throughout the basin aquifer system, Taylor (1982) noted empirical relations between altitude and recharge estimated by watershed modeling. These empirical relations were used to estimate the distribution of recharge in areas not modeled by Weeks and others (1974).

Water-quality data also were used to substantiate and delineate recharge areas and to help define basin flow paths. Ion-exchange reactions are believed to be common in aquifers in the study area. Chafin and Kimball (1992) reported ion exchange of calcium and magnesium for sodium on clay as a dominant reaction in the Wasatch aquifer (Wasatch zone of the Wasatch-Fort Union aquifer) in the Green River Basin aquifer system.

Because of the large partial pressure of carbon dioxide commonly associated with water in recharge areas, carbonate dissolution is likely. Dissolution of carbonates increases the concentrations of calcium and magnesium compared to their concentrations in precipitation. As recharge water moves downgradient, calcium and magnesium ions exchange with sodium ions on clay materials in the aquifer. Henderson (1985) used the progression from positive to negative log $(([CA]+[Mg])/[Na]^2)$ values to identify areas of recharge and directions of flow in two aquifer systems in Montana and Wyoming. A similar approach was used in the study of the Upper Colorado River Basin. Log-molar ratios of divalent to monovalent cations were calculated as

$$\log\left(([CA]+[Mg])/[Na]^2\right),$$

where Ca = the calcium concentration, in millimoles per liter;

Mg = the magnesium concentration, in millimoles per liter; and

Na = the sodium concentration, in millimoles per liter.

Maps showing the distribution of log-molar ratios of calcium plus magnesium concentrations divided by squared-sodium concentrations were constructed; these maps were compared to recharge and potentiometric-surface maps to confirm and further delineate areas of recharge and ground-water movement in selected aquifers. The largest (positive) log $(([CA]+[Mg])/[Na]^2)$ values were used to identify recharge areas; small (negative) log $(([CA]+[Mg])/[Na]^2)$ values were used to identify nonrecharge (downgradient) areas. Log $(([CA]+[Mg])/[Na]^2)$ values are of limited usefulness in aquifers with substantial sources of sodium other than ion exchange (for example, aquifers containing sodium salts). Because of the large quantities of sodium salts in the Green River Formation, log $(([CA]+[Mg])/[Na]^2)$ ratio maps were not used to identify recharge areas in the Laney aquifer (Green River Basin) and the upper and lower Piceance Basin aquifers (Piceance Basin).

Ground water presently is discharged from aquifers by evapotranspiration, springs, wells, and diffuse seepage along streams. Depending on the method used, estimation of evapotranspiration rates requires knowledge of vegetation type and density, potential rates of water use by plants, rates of precipitation, temperature, soil moisture, and depth to water. Although this knowledge may be available locally, reliable estimation of basin evapotranspiration rates is impractical. Therefore, estimates of ground-water discharge were made when effects of evapotranspiration are minimal. Discharge measurements for springs and flowing wells have been tabulated in several reports (Welder and McGreevy, 1966; Welder, 1968; Hood and others, 1976). Small springs and seeps are numerous in the Upper Colorado River Basin, and measurement of all springs is not practical. However, it is likely that discharge by unmeasured small springs is small compared to total discharge.

Measurement of ground-water discharge by diffuse seepage along a stream is possible, provided that all other sources of water in the stream can be measured and the quantity of water stored in the channel does not change. These stipulations are never completely met but may be approximately true after several months of baseflow conditions when surface-water diversions and return flow of irrigation water are not occurring. Under these conditions, a series of stream discharge measurements, accounting for all tributary inflows and diversions, could be used to calculate a water budget. The difference between water entering the stream and water leaving the stream is assumed to be ground-water recharge or discharge.

Ground-water discharge by diffuse seepage along streams was estimated for aquifer systems in Tertiary rocks in the Upper Colorado River Basin using a water-budget approach and January mean monthly discharge estimates at streamflow gages. No estimates were possible for stream reaches where ground-water discharge was a small percentage of streamflow or where discharge of important tributaries or diversions were not measured. January was selected for analysis because evapotranspiration is minimal, diversion of irrigation water typically stops in October or November, and snowmelt is minimal. Ice forms in many streams during January and may reduce the accuracy of estimated ground-water discharge if the volume of water stored in the channel as ice changes substantially.

The later sections of this report present what is considered to be a fairly complete and reasonably accurate description of ground-water discharge for three basin aquifer systems in Tertiary rocks in the Upper Colorado River Basin. Ground-water models of these aquifer systems have been developed in order to refine the estimated distribution of recharge and hydraulic conductivity. In areas where discharge was unknown or unreliably estimated, the models also were used to estimate the distribution of discharge. Models have been developed for the Piceance Basin (Taylor, 1982, 1986), Uinta Basin (Holmes and Kimball, 1987) and Green River Basin (Martin, 1996) aquifer systems. No model has been developed for the Great Divide-Washakie-Sand Wash Basins aquifer system. Model estimates of recharge and discharge are consistent with measured hydraulic-head data and other information. Detailed descriptions of model development are given in the referenced reports. Where possible, model-estimated distributions of recharge and discharge have been qualified by statistical measures of accuracy.

GROUND-WATER MOVEMENT

Numerical models of ground-water flow are used to estimate rates of ground-water movement from recharge areas to discharge areas and between aquifers. Models are used because of their capability to include complex distributions of hydraulic conductivity, recharge, and discharge. In Tertiary rocks of the Upper Colorado River Basin, quantitative descriptions of ground-water movement have been made in the Piceance, Uinta, and Green River Basin aquifer systems based on digital models. In the Great Divide-Washakie-Sand Wash Basins aquifer system, where no reliable model has been developed, quantitative description of ground-water movement is not possible. However, potentiometric-surface maps can be used to describe general directions of ground-water movement from recharge areas to discharge areas.

QUALITY OF WATER

The water-quality data used in this study are from the National Water Information System (NWIS) files of the U.S. Geological Survey, the files of Petroleum Data Services in Norman, Okla., and from published reports. Data in the NWIS files generally are from water samples collected from wells developed for water supplies. Water-quality analyses from Petroleum Data Services generally are from water samples collected by petroleum companies during drill-stem tests of oil wells. Many of the data were not used in this report because information about the sampling interval was not available or because contamination by drilling fluids was suspected.

The chemical quality of water in the aquifers within the study area was characterized by the use of dissolved-solids-concentration maps. The dissolved-solids data also were separated into major-element components by the method developed by Stiff (1951). Stiff diagrams portray cation and anion concentrations of ground water on three horizontal axes extending on either side of a vertical zero axis. Stiff diagrams for water samples from selected wells in the study area were plotted on maps in conjunction with the dissolved-solids concentration.

BASIN AQUIFER SYSTEMS

Four basin aquifer systems have been identified in the Tertiary rocks of the study area (fig. 6). The aquifer systems are designated by the names of their structural basins: Piceance Basin, Uinta Basin, Green River Basin, and Great Divide-Washakie-Sand Wash Basins. Geohydrologic units for each aquifer system are described in following sections of the report.

PICEANCE BASIN AQUIFER SYSTEM

Tertiary rocks of the Piceance Basin are divided into two basin aquifers and two basin confining units. The aquifers are designated as upper and lower Piceance Basin aquifers to agree with past studies (Weeks and others, 1974; Robson and Saulnier, 1981; Taylor, 1982).

The upper Piceance Basin aquifer includes permeable sandstone and fractured siltstone of the Uinta Formation and fractured marlstone and solution cavities of the upper part of the Parachute Creek Member of the Green River Formation. Average thickness of the aquifer is about 700 ft (fig. 7).

The lower Piceance Basin aquifer includes the fractured marlstone and a leached zone of the lower part of the Parachute Creek Member. The altitude of the top of the aquifer decreases from 7,500 ft above sea level along basin margins to less than 5,000 ft in the north-central part of the basin (fig. 8). The average thickness of the aquifer is about 900 ft (fig. 9).

The Mahogany confining unit separates the upper and lower aquifers and correlates with the Mahogany zone of the Parachute Creek Member. The average thickness of the confining unit is about 160 ft (fig. 10).

A basal confining unit, located stratigraphically below the lower Piceance Basin aquifer, includes the Garden Gulch Member of the Green River Formation and all underlying Tertiary rocks (fig. 11). In the north-central Piceance Basin, shale beds in the Parachute Creek Member with interbeds and nodules of saline minerals (halite and nahcolite) also are included as part of the basal confining unit.

With minor exceptions, the areal extent of the Piceance Basin aquifer system is bounded on the north by the White River and on the south by the Colorado River (fig. 11). The basal confining unit of the aquifer system is present throughout most of the structural basin; however, erosion has removed some of the upper and lower Piceance Basin aquifers in various parts of the basin. Specifically, erosion along the Douglas Creek Arch (fig. 2) also has removed all but a narrow band of the aquifer system. The aquifer system is bounded on the east by the White River Uplift. Reflecting the areal distribution of the upper and lower Piceance Basin aquifers, the geohydrologic description of the aquifer system is limited to the area between the Colorado and White Rivers.

UINTA BASIN AQUIFER SYSTEM

Two basin aquifers, the Duchesne River-Uinta and the Douglas Creek-Renegade aquifers, are present in Tertiary rocks of the Uinta Basin. Sandstone and fractured shale of the Duchesne River and Uinta Formations compose the Duchesne River-Uinta aquifer. The two formations are considered a single aquifer because of similar lithologies and hydraulic-conductivities (Hood, 1976). The Duchesne River Formation directly overlies the Uinta Formation. Thickness of the Duchesne River-Uinta aquifer ranges from 1,000 to 8,000 ft in most of the northern Uinta Basin (fig. 12). Erosion in the southern half of the Uinta Basin and along the Douglas Creek Arch (fig. 2) has limited the Duchesne River-Uinta aquifer to the northern Uinta Basin. Uplifts bound the aquifer on the north and west. Sandstone and limestone beds in the Douglas Creek Member of the Green River Formation and intertonguing sandstone of the Renegade Tongue of the Wasatch Formation compose the Douglas Creek-Renegade aquifer. Average thickness of the aquifer is about 500 ft. The aquifer crops out along the southern margin of the Uinta Basin (fig. 13). The Douglas Creek-Renegade aquifer is present principally in the southeastern part of the basin. Hydraulic conductivity of the aquifer decreases to the north and west, where the aquifer becomes more like a confining unit.

An upper confining unit, the Parachute Creek confining unit, separates the Duchesne River-Uinta aquifer from the Douglas Creek-Renegade aquifer, and a lower confining unit, the Wasatch-Green River confining unit, separates the Douglas Creek-Renegade aquifer from the Mesaverde aquifer, sometimes called the North Horn-Mesaverde aquifer (Naftz, 1996). The Parachute Creek confining unit is composed of the relatively unfractured shale, limestone, and marlstone of the Green River Formation. The Parachute Creek confining unit is 3,000 to 6,000 ft thick and is present throughout most of the Uinta Basin (fig. 14). The Parachute Creek confining unit is absent along the Uinta Uplift where the Duchesne River-Uinta aquifer directly overlies Mesozoic and Paleozoic rocks. The Wasatch-

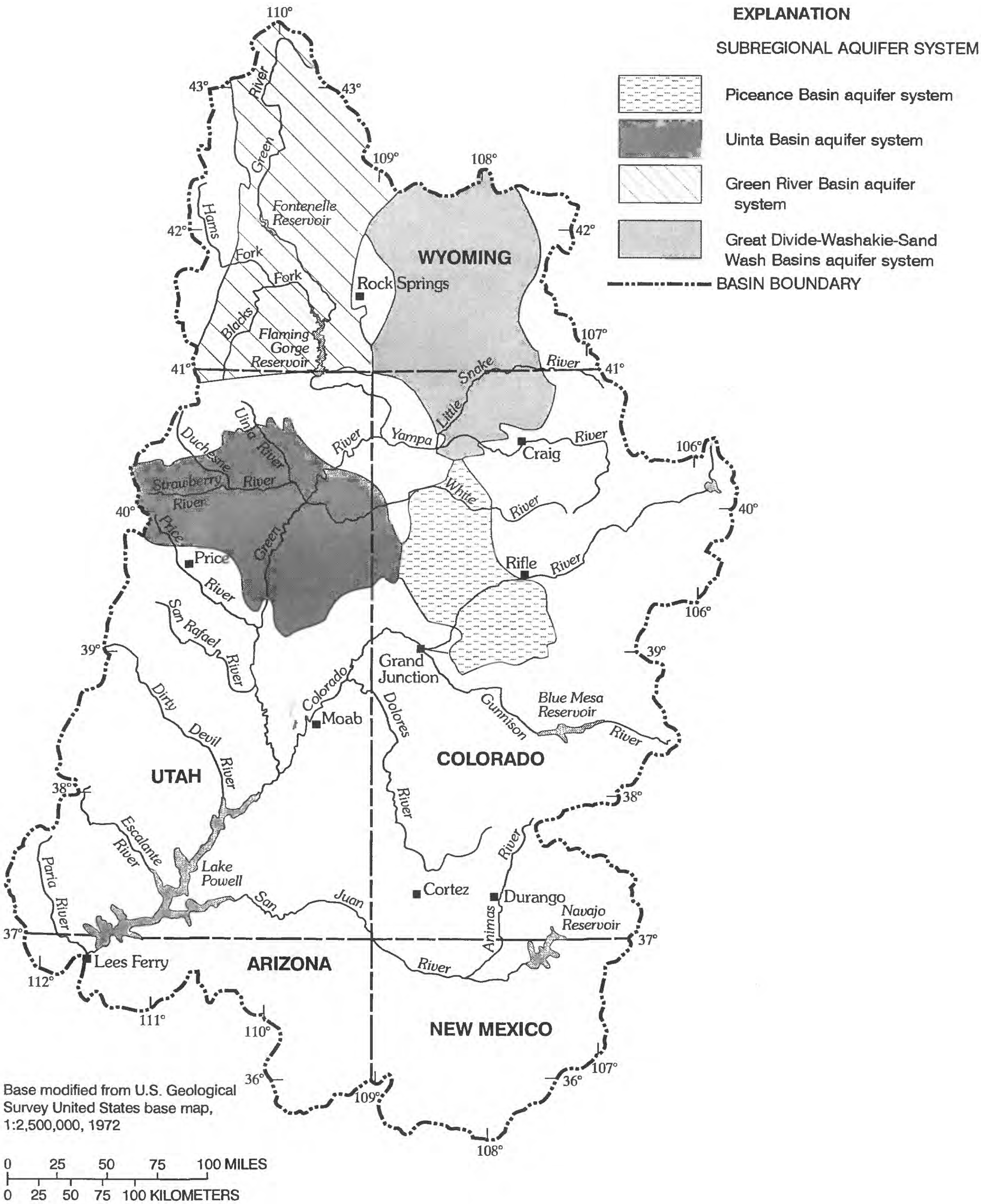
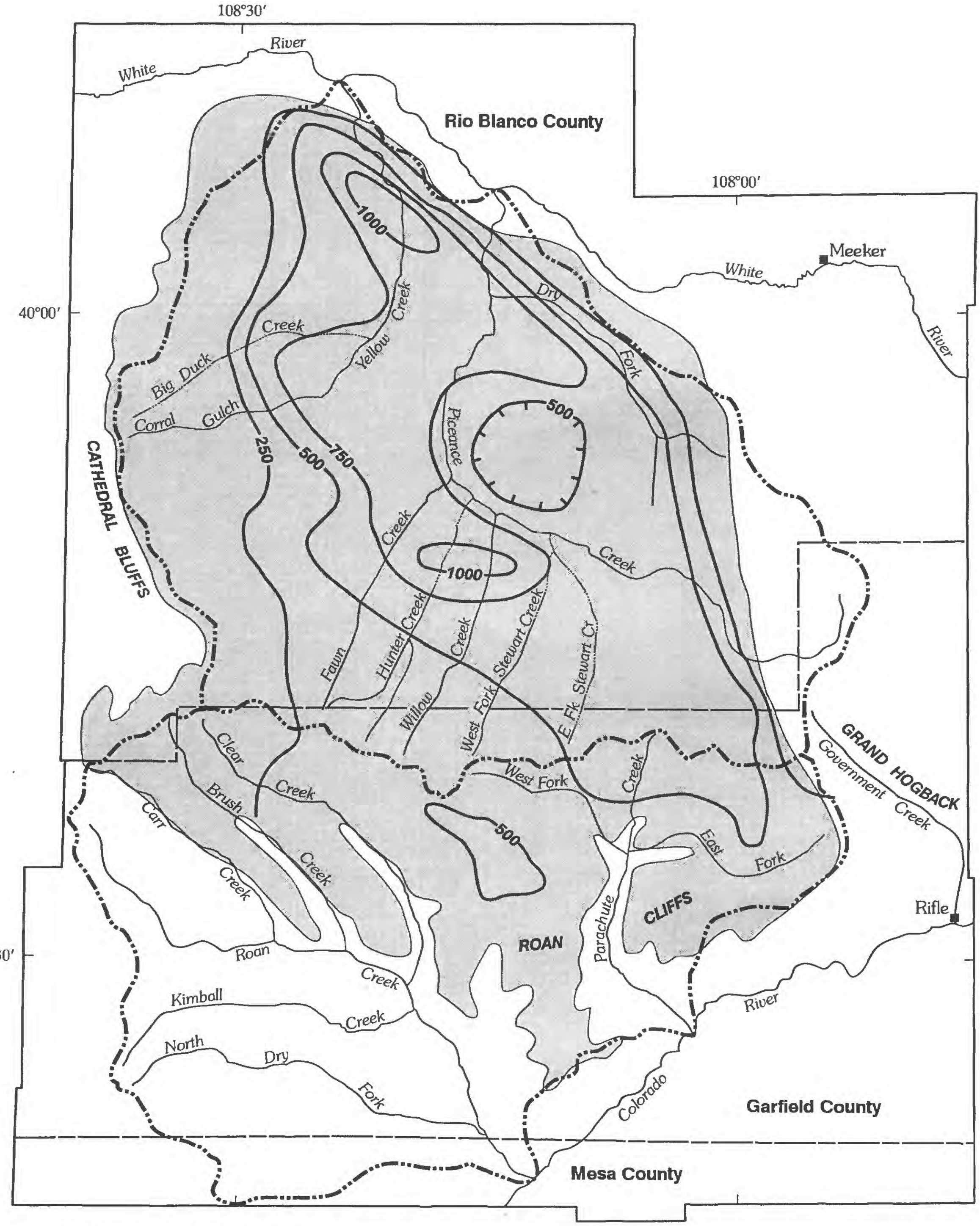


Figure 6. Location of basin aquifer systems in rocks of Tertiary age of the Upper Colorado River Basin.



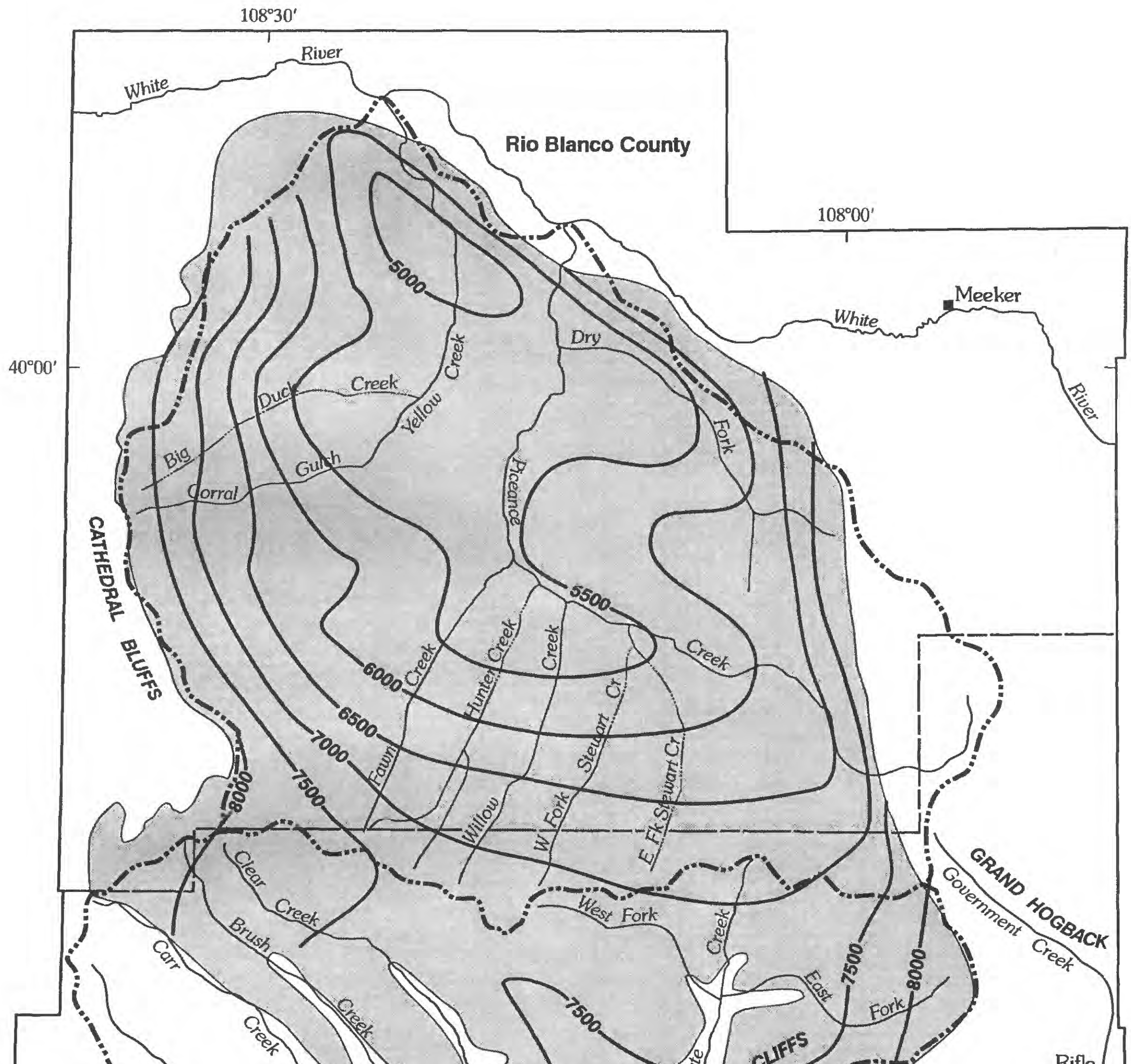
EXPLANATION

AREAL EXTENT OF UPPER PICEANCE BASIN AQUIFER

—500— LINE OF EQUAL THICKNESS OF THE UPPER PICEANCE BASIN AQUIFER--
Interval 250 feet. Hachures indicate closed area of lesser thickness

—··—··— BASIN BOUNDARY

Figure 7. Areal extent and thickness of the upper Piceance Basin aquifer, Piceance Basin aquifer system (modified from Mullens, 1976, and Robson and Saulnier, 1981).



Base modified from U.S. Geological Survey 1:500,000 Colorado State base map, 1969

0 5 10 15 MILES

0 5 10 15 KILOMETERS

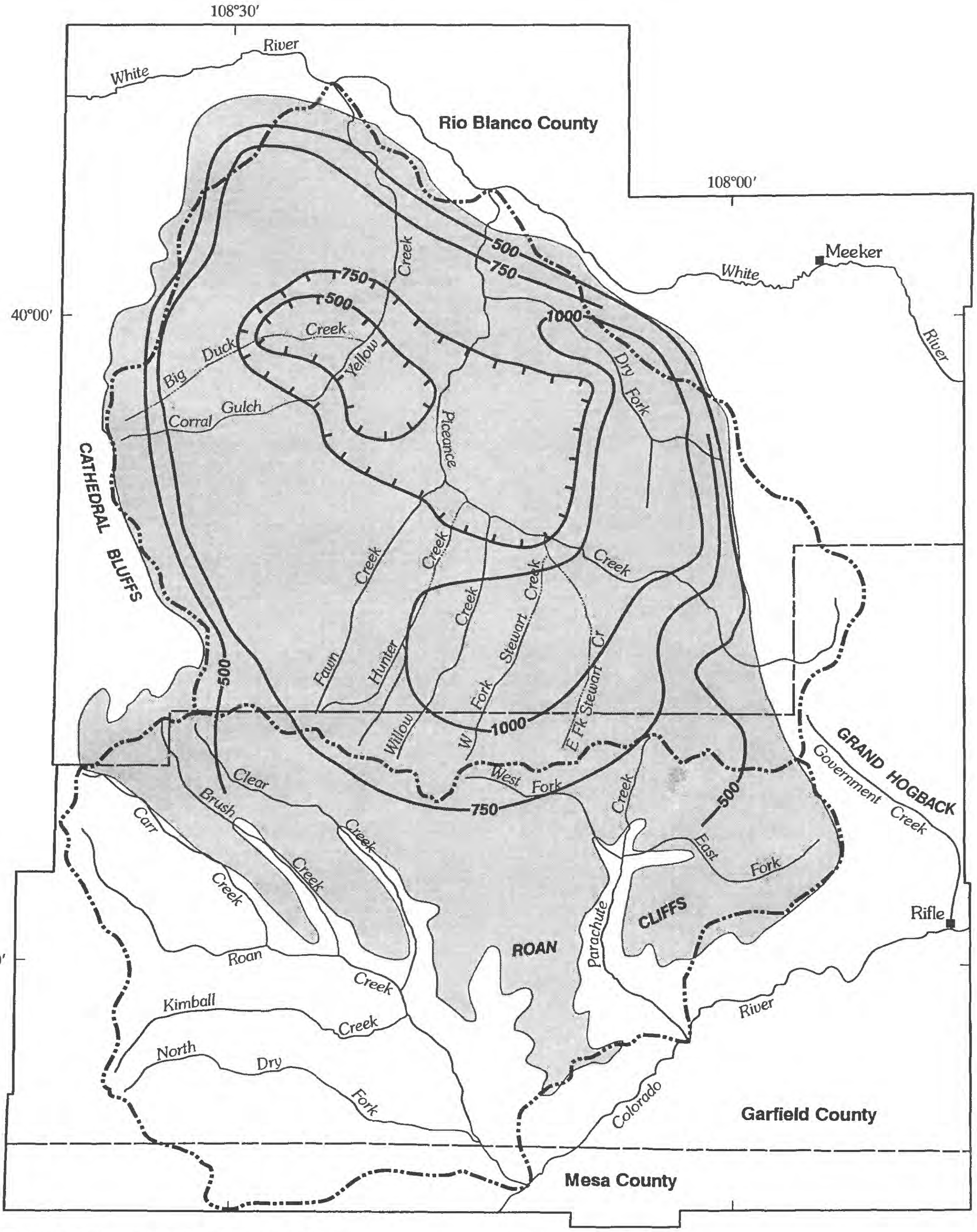
EXPLANATION

AREAL EXTENT OF LOWER PICEANCE BASIN AQUIFER

—7500— SUBSURFACE CONTOUR--Shows altitude of the top of the lower Piceance Basin aquifer. Interval 500 feet. Datum is seal level

BASIN BOUNDARY

Figure 8. Areal extent and altitude of the top of the lower Piceance Basin aquifer, Piceance Basin aquifer system (modified from Robson and Saulnier, 1981).



Base modified from U.S. Geological Survey 1:500,000 Colorado State base map, 1969

0 5 10 15 MILES

0 5 10 15 KILOMETERS

EXPLANATION

AREAL EXTENT OF LOWER PICEANCE BASIN AQUIFER

—500— LINE OF EQUAL THICKNESS OF THE LOWER PICEANCE BASIN AQUIFER-- Interval 250 feet. Hachures indicate closed areas of lesser thickness

—··—··— BASIN BOUNDARY

Figure 9. Areal extent and thickness of the lower Piceance Basin aquifer, Piceance Basin aquifer system.

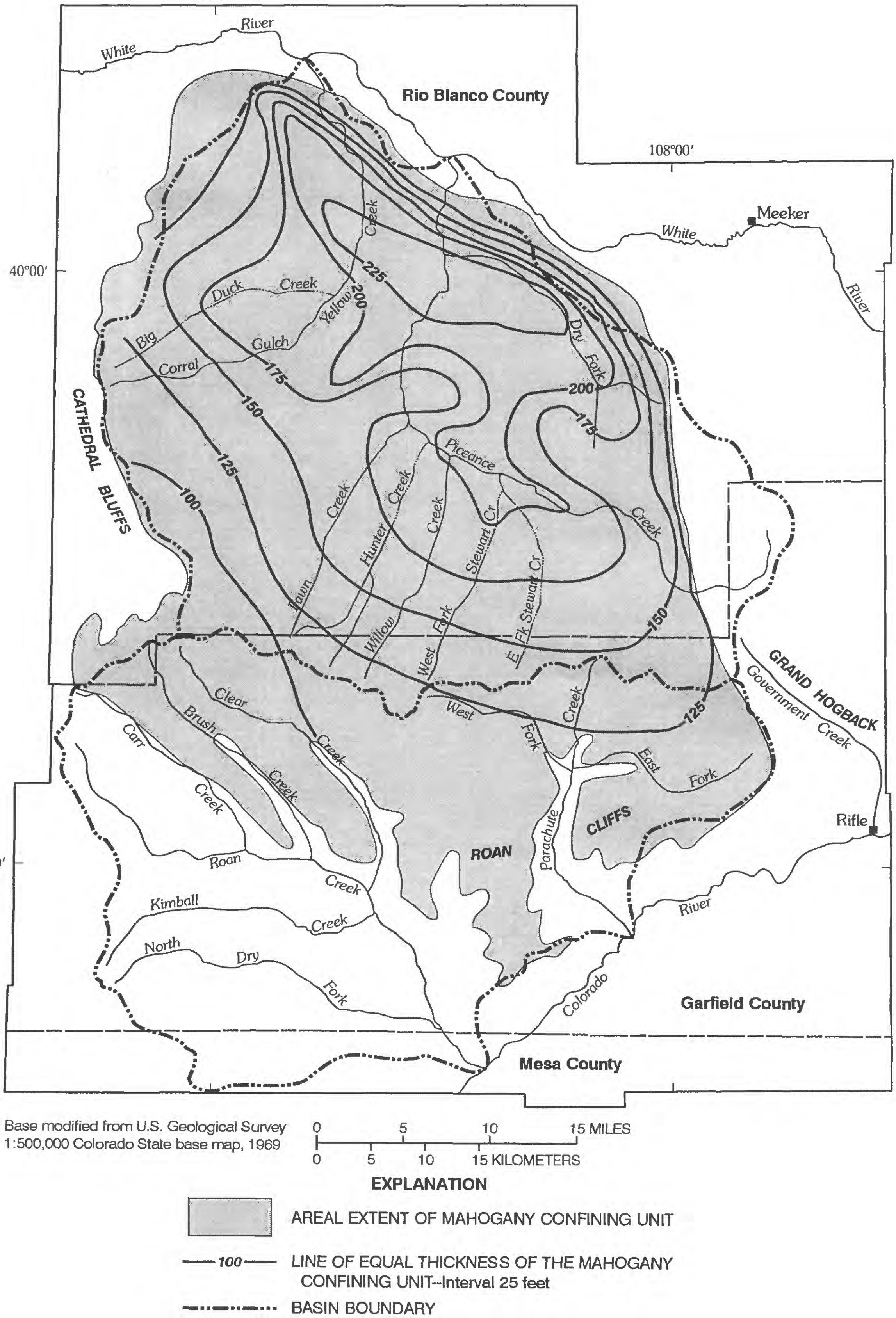
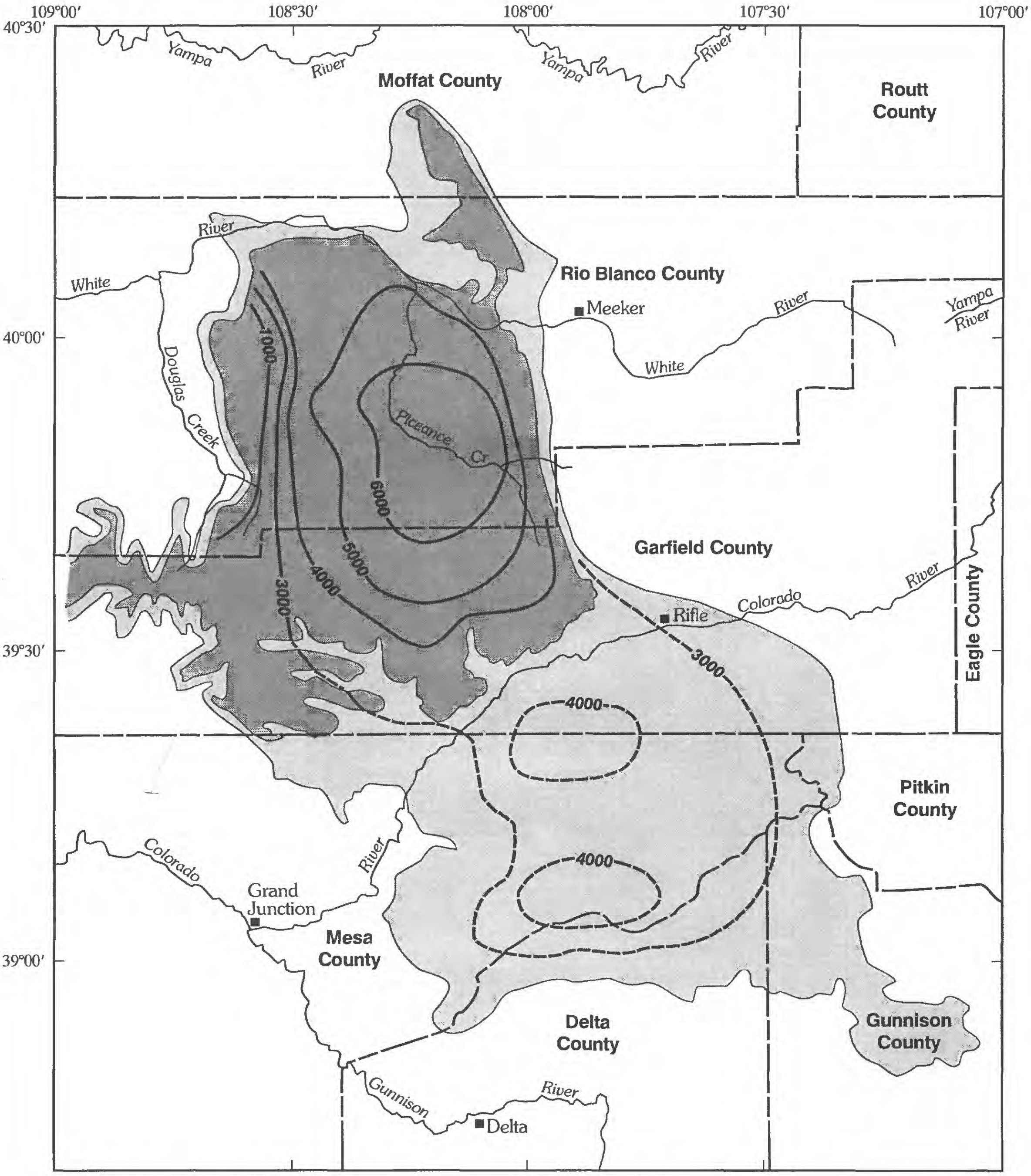


Figure 10. Areal extent and thickness of the Mahogany confining unit, Piceance Basin aquifer system (modified from Donnell and Blair, 1970).



Base modified from U.S. Geological Survey Colorado State base map, 1:500,000, 1969

0 10 20 MILES

0 10 20 KILOMETERS

Geology by K.C. Glover

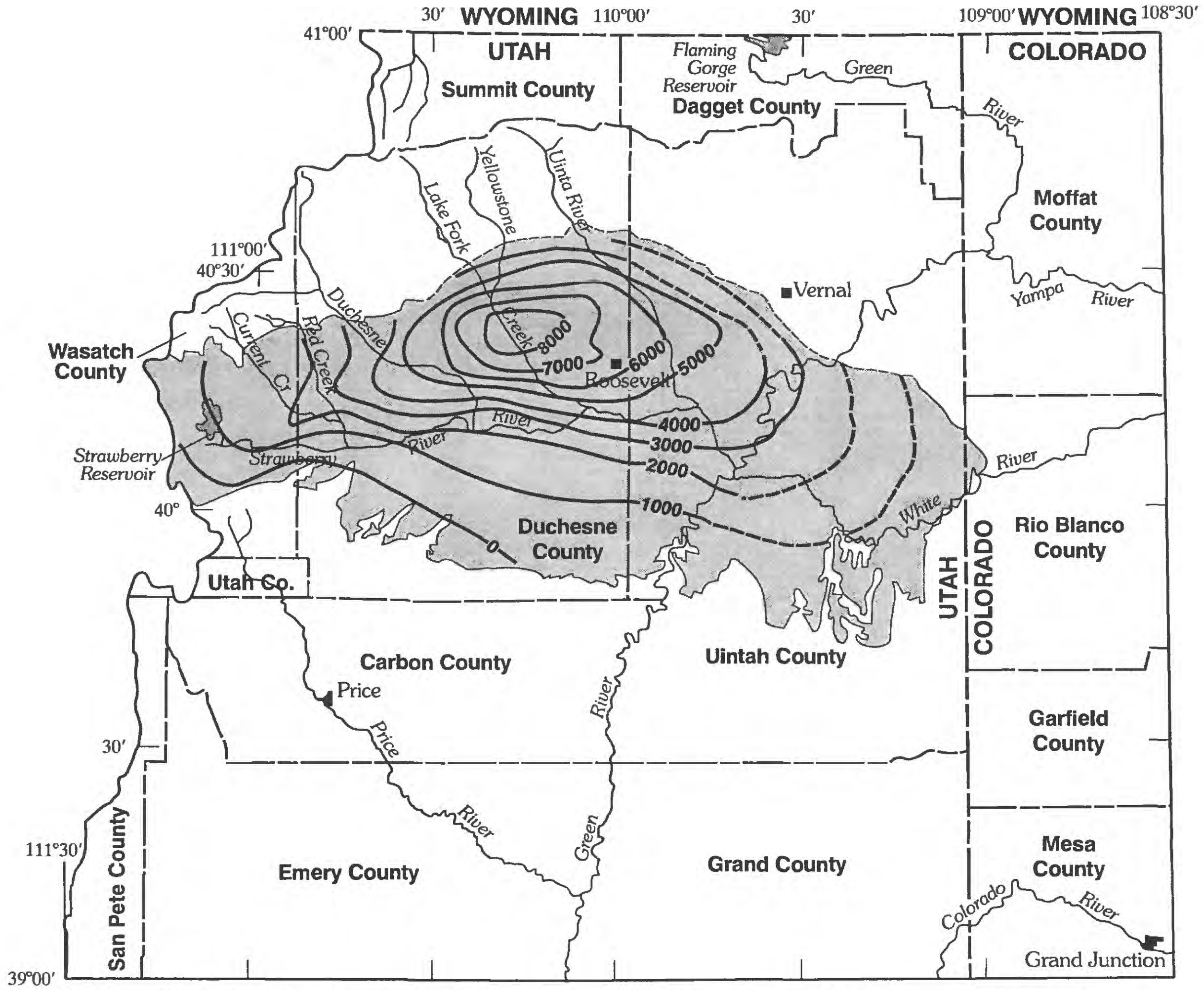
EXPLANATION

BASAL CONFINING UNIT EXPOSED AT LAND SURFACE

BASAL CONFINING UNIT BURIED

—4000— — LINE OF EQUAL THICKNESS OF THE BASAL CONFINING UNIT--
Dashed where approximately located. Interval, in feet, variable

Figure 11. Areal extent and thickness of the basal confining unit, Piceance Basin aquifer system.



Base from U.S. Geological Survey 1:500,000 State base maps: Utah, 1976; and Colorado, 1980

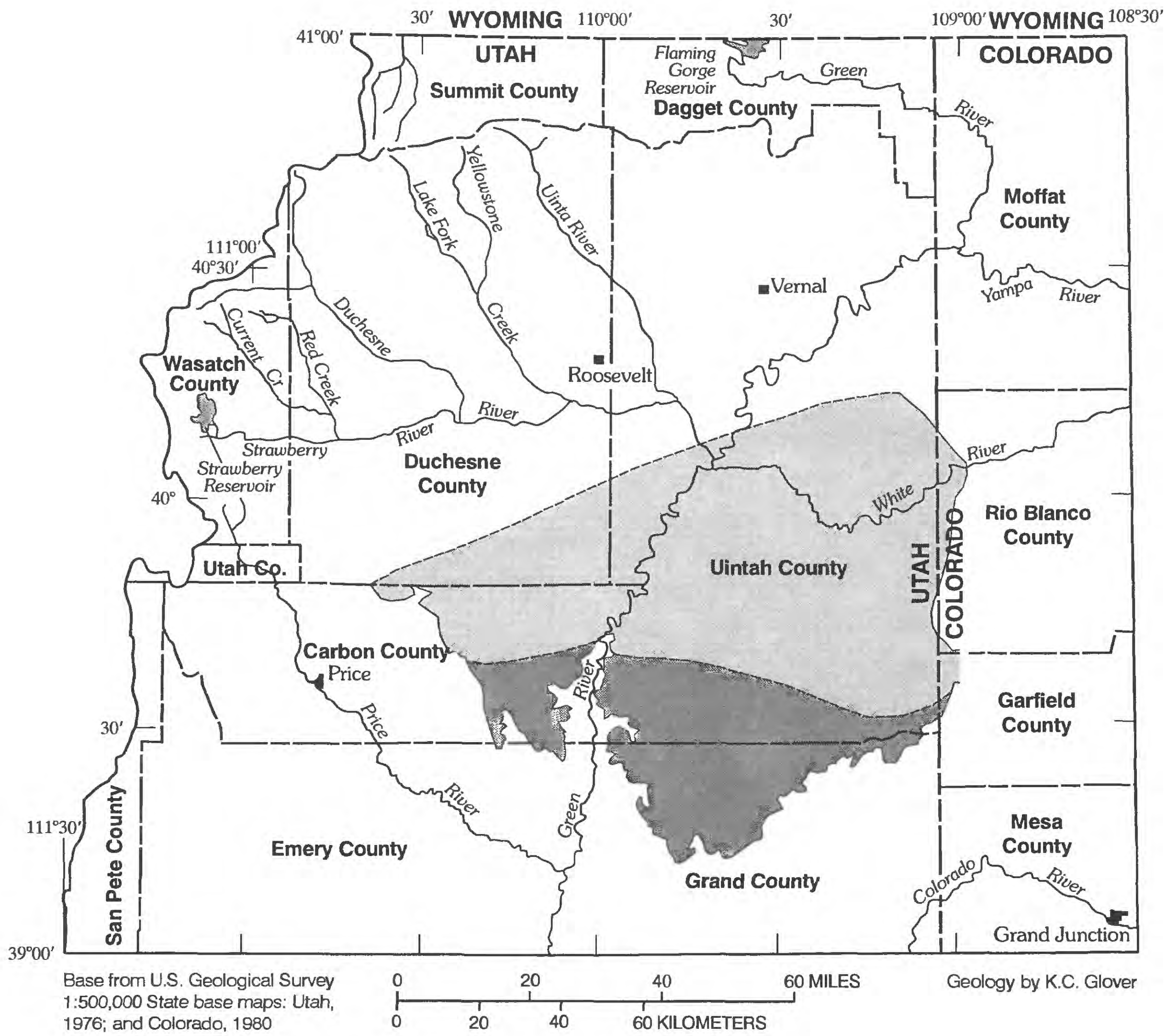
Hydrology by K.C. Glover

EXPLANATION

AREAL EXTENT OF DUCHESNE RIVER-UINTA AQUIFER-- Boundary dashed where approximately located

—1000— LINE OF EQUAL THICKNESS OF THE DUCHESNE RIVER-UINTA AQUIFER-- Dashed where approximately located. Interval 1,000 feet

Figure 12. Areal extent and thickness of the Duchesne River-Uinta aquifer, Uinta Basin aquifer system.

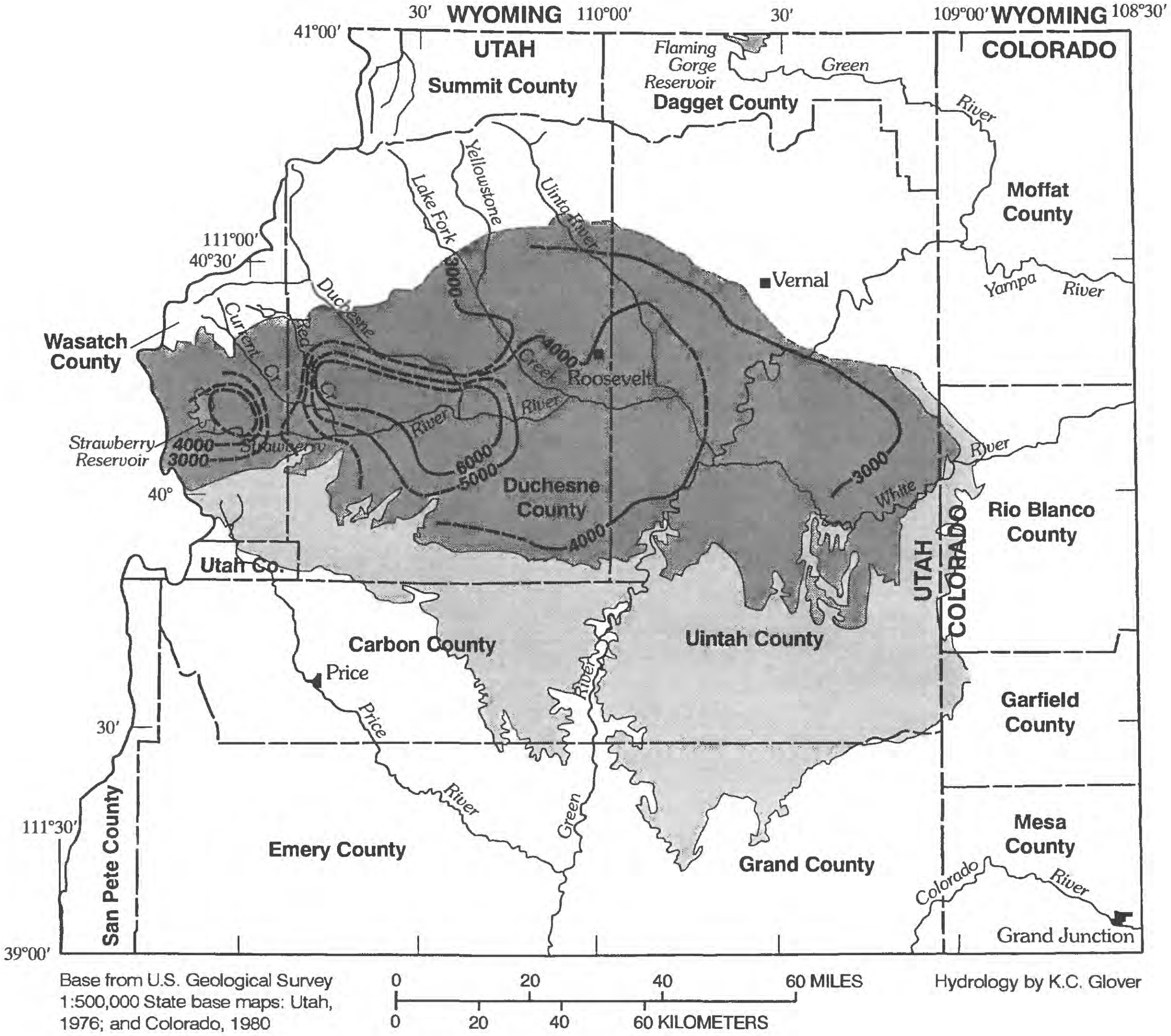


EXPLANATION

DOUGLAS CREEK-RENEGADE AQUIFER BURIED-- Boundary dashed where approximately located

DOUGLAS CREEK-RENEGADE AQUIFER EXPOSED AT LAND SURFACE ALONG MAJOR DRAINAGES-- Aquifer generally is buried in upland areas and exposed only along streams. Boundary dashed where approximately located

Figure 13. Areal extent of the Douglas Creek-Renegade aquifer, Uinta Basin aquifer system.



EXPLANATION

PARACHUTE CREEK CONFINING UNIT BURIED--Boundary dashed where approximately located

PARACHUTE CREEK CONFINING UNIT EXPOSED AT LAND SURFACE

—4000— LINE OF EQUAL THICKNESS OF THE PARACHUTE CREEK CONFINING UNIT--Dashed where approximately located. Interval 1,000 feet

Figure 14. Areal extent and thickness of the Parachute Creek confining unit, Uinta Basin aquifer system.

Green River confining unit is composed of the Wasatch Formation and Flagstaff Limestone, known locally as the Flagstaff Member of the Green River Formation in the western Uinta Basin. The Wasatch Formation is predominately shale and is present throughout most of the Uinta Basin. The Flagstaff Member is relatively unfractured in the subsurface and is restricted to the western Uinta Basin. Thickness of the Wasatch-Green River confining unit is 2,000 to 6,000 ft where overlain by the Douglas Creek-Renegade and Duchesne River-Uinta aquifers (fig. 15).

GREEN RIVER BASIN AQUIFER SYSTEM

Tertiary rocks of the Green River Basin are divided into four basin aquifers and two basin confining units. The Bridger, Laney, New Fork, and Wasatch-Fort Union aquifers are separated geographically or stratigraphically by confining units. The Wasatch-Fort Union aquifer has been separated into the Wasatch zone and the Fort Union zone. Designation of the two zones as upper and lower parts of a combined aquifer was done in view of the thickness of the formations and the known hydrologic differences between them.

Some Tertiary formations of minor areal extent are stratigraphically equivalent to a major geologic unit and considered part of a major aquifer or zone: The Pass Peak Formation is included with the Wasatch zone of the Wasatch-Fort Union aquifer, and the Hoback Formation is included with the Fort Union zone. Similarly, many of the minor tongues and members of the Green River and Wasatch Formations are not discussed individually but are included in major aquifers or confining units (plate 1). Investigation of the following local Tertiary aquifers was beyond the scope of this study: Quaternary alluvial deposits along major streams, Quaternary glacial deposits, Miocene and Oligocene rocks, the Browns Park Formation, the Bishop Conglomerate, and other Oligocene and(or) Eocene rocks.

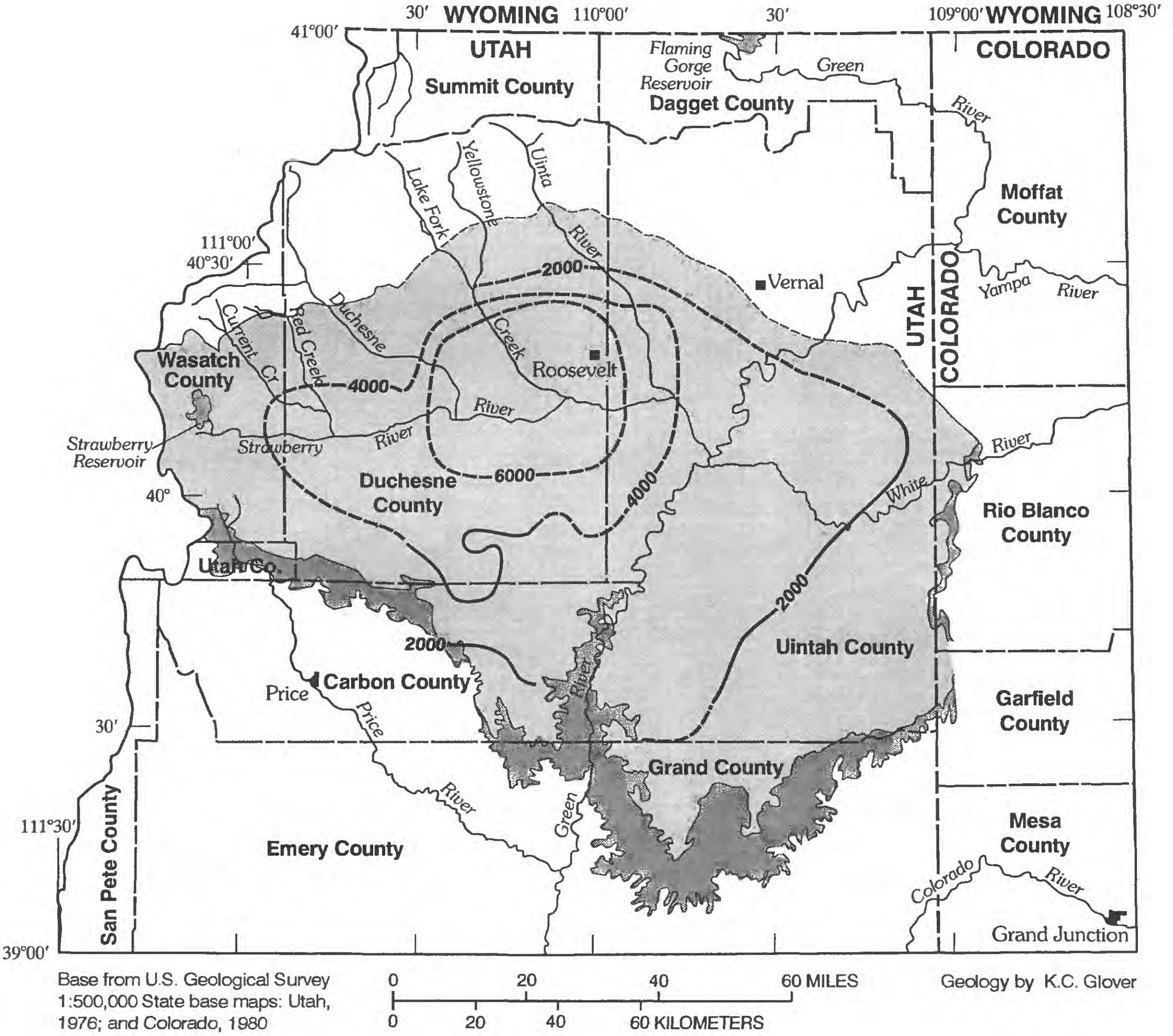
Fractured sandstone, tuff, and shale of the Bridger Formation compose the Bridger aquifer. The aquifer crops out in the southern part of the Green River Basin, where it attains a thickness in excess of 1,500 ft (fig. 16). Erosional remnants of the Bridger aquifer are present in the northeastern part of the basin, but are not important to the basin flow system and are not mapped in figure 16. The saturated thickness of the Bridger aquifer is generally less than 1,000 ft.

The Laney Member of the Green River Formation is designated an aquifer, although for purposes of this study, it could be classified easily as either an aquifer or a confining unit. The extremely small hydraulic conductivity of the rock matrix throughout most of its areal extent suggests the Laney Member be classified as a confining unit. However, fractures and solution channels north of Blacks Fork provide sufficiently high well yields and hydraulic conductivity to justify classification as an aquifer. The large percentage of conglomerate and other permeable rocks along the Uinta Uplift (fig. 2) also indicates that the Laney Member is an aquifer. The alternative designation of the Laney Member as a confining unit is not acceptable because the largest yields in the Green River Basin consistently obtained from wells completed in the Laney Member. Division of the Laney Member into two or more geohydrologic units is not appropriate at basin scales because boundaries between areas of large and small hydraulic conductivity are gradational and not well understood.

The designation of the Laney Member as an aquifer places the Bridger aquifer in direct contact with the Laney aquifer throughout most of the southern Green River Basin. However, the hydraulic conductivity of the Laney aquifer generally is small in areas where it is in direct contact with the overlying Bridger aquifer. Thickness of the Laney aquifer exceeds 1,000 ft in the south-central part of the Green River Basin (fig. 17), but generally is 200 to 600 ft.

The Wilkins Peak confining unit separates the Bridger and Laney aquifers from underlying aquifers. The unit consists of the relatively unfractured Wilkins Peak Member of the Green River Formation. Areas of bedded trona deposits, mapped by Bradley and Eugster (1969), effectively restrict vertical movement of water throughout a large area of the Wilkins Peak confining unit. Thickness of the Wilkins Peak confining unit exceeds 1,000 ft in the south-central part of the Green River Basin, but typically is about 200 to 600 ft (fig. 18).

The New Fork aquifer is comprised of arkosic sandstone within the New Fork Tongue of the Wasatch Formation. The aquifer is located in the northern part of the Green River Basin between the Wilkins Peak confining unit and underlying Tipton confining unit (fig. 19). The aquifer typically is 300 to 350 ft thick (fig. 20). The New Fork aquifer thins to the south and is absent near Green River, Wyo. Where the Tipton confining unit is absent in the northern Green River Basin, the New Fork aquifer cannot be distinguished from the Wasatch zone of the Wasatch-Fort Union aquifer.



EXPLANATION

WASATCH-GREEN RIVER CONFINING UNIT BURIED--Boundary dashed where approximately located

WASATCH-GREEN RIVER CONFINING UNIT EXPOSED AT LAND SURFACE--Boundary dashed where approximately located

—2000— LINE OF EQUAL THICKNESS OF THE LOWER CONFINING UNIT--Dashed where approximately located. Interval 2,000 feet

Figure 15. Areal extent and thickness of the Wasatch-Green River confining unit, Uinta Basin aquifer system.

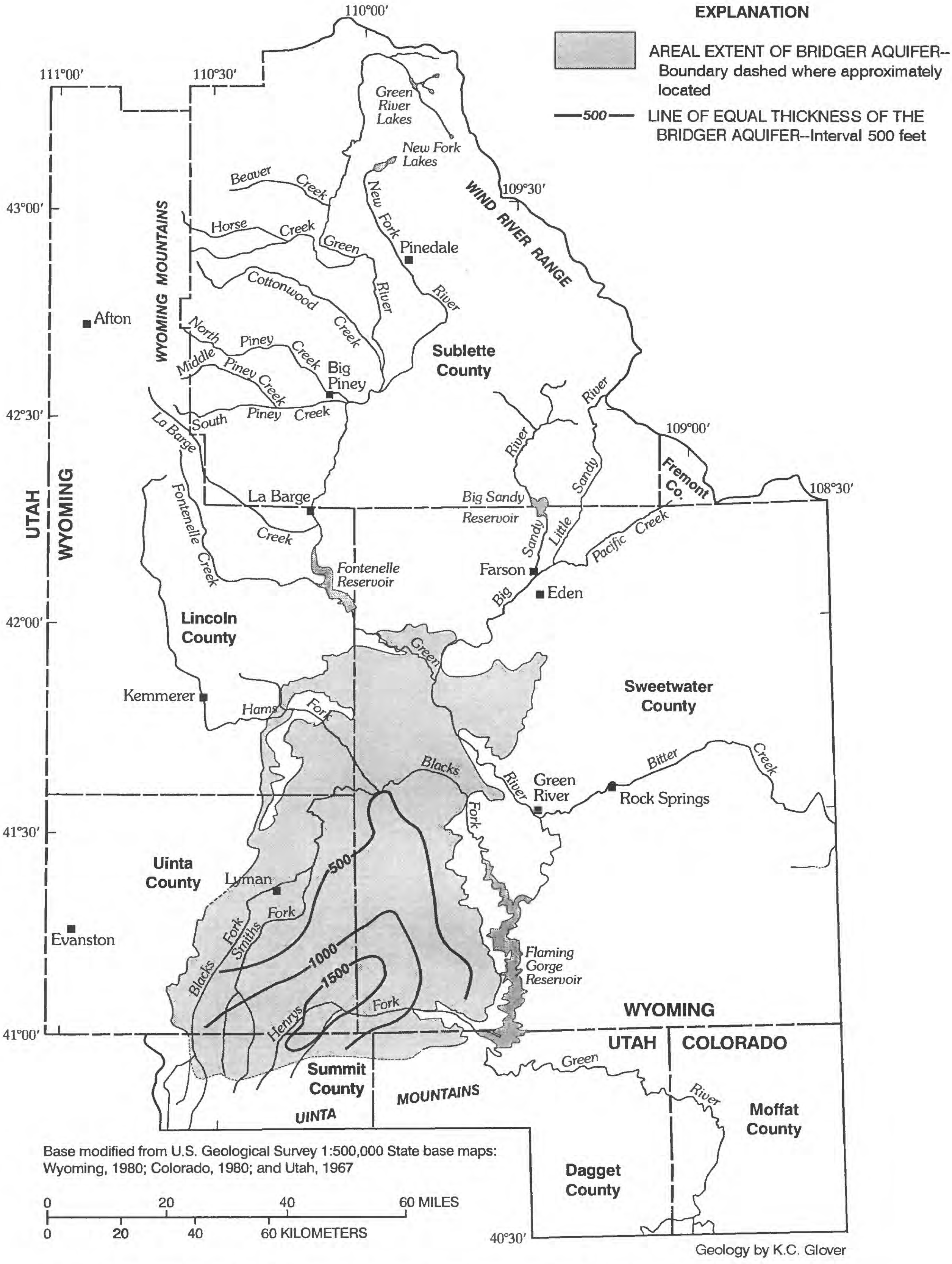


Figure 16. Areal extent and thickness of the Bridger aquifer, Green River Basin aquifer system (from Martin, 1996, p. 17).

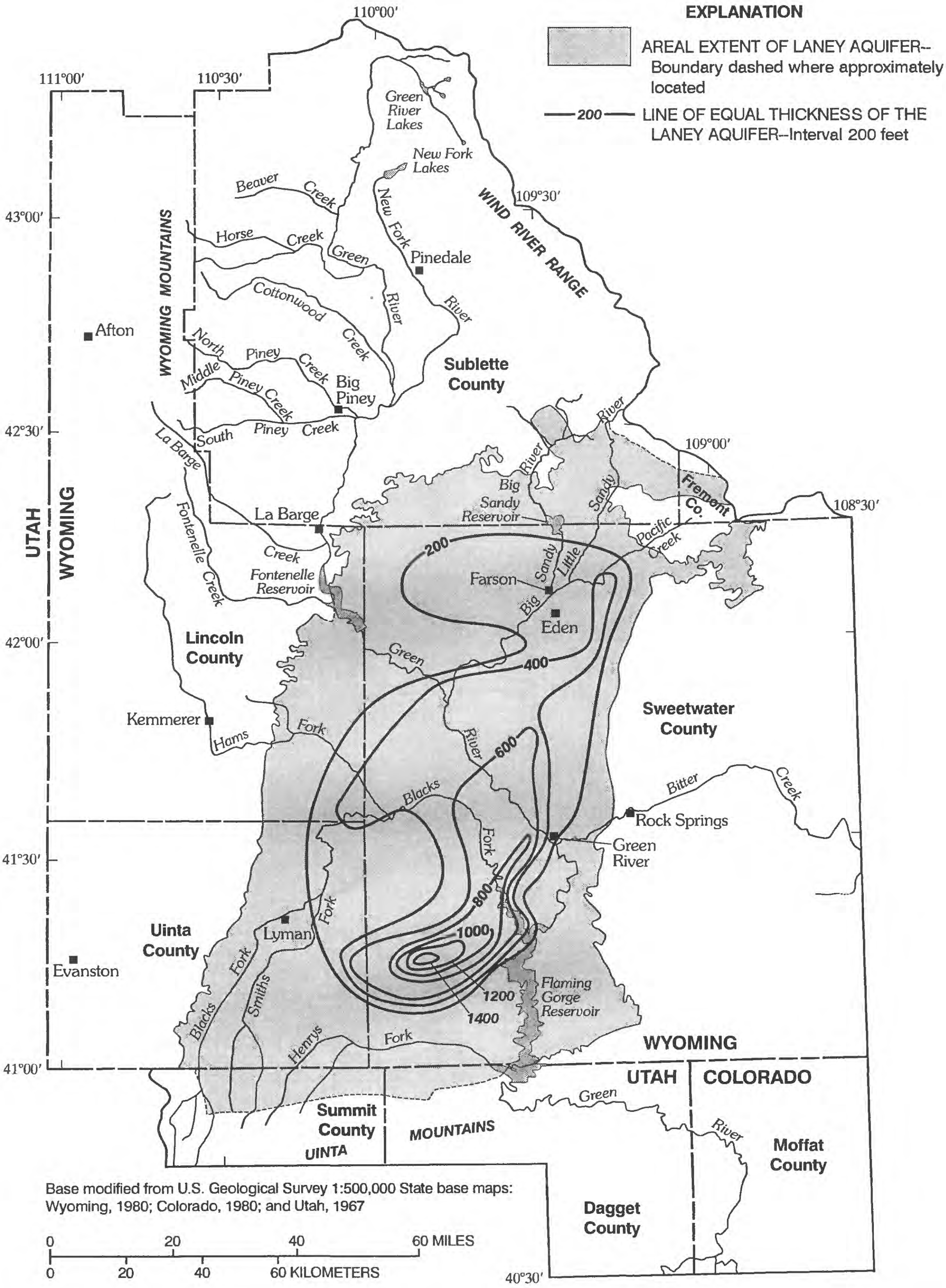


Figure 17. Areal extent and thickness of the Laney aquifer, Green River Basin aquifer system (from Martin, 1996, p. 13).

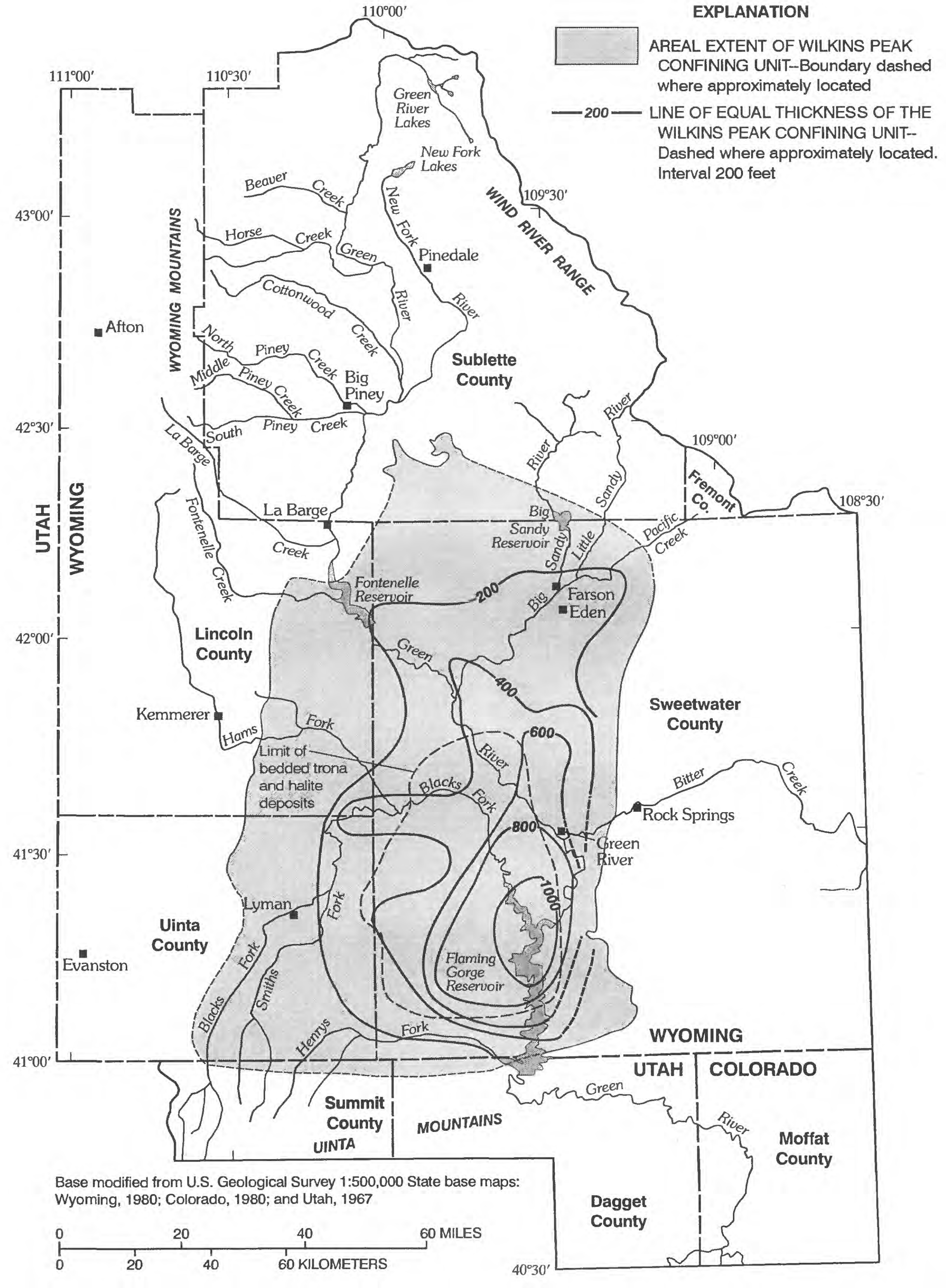


Figure 18. Areal extent and thickness of the Wilkins Peak confining unit, Green River Basin aquifer system (from Martin, 1996, p. 16).

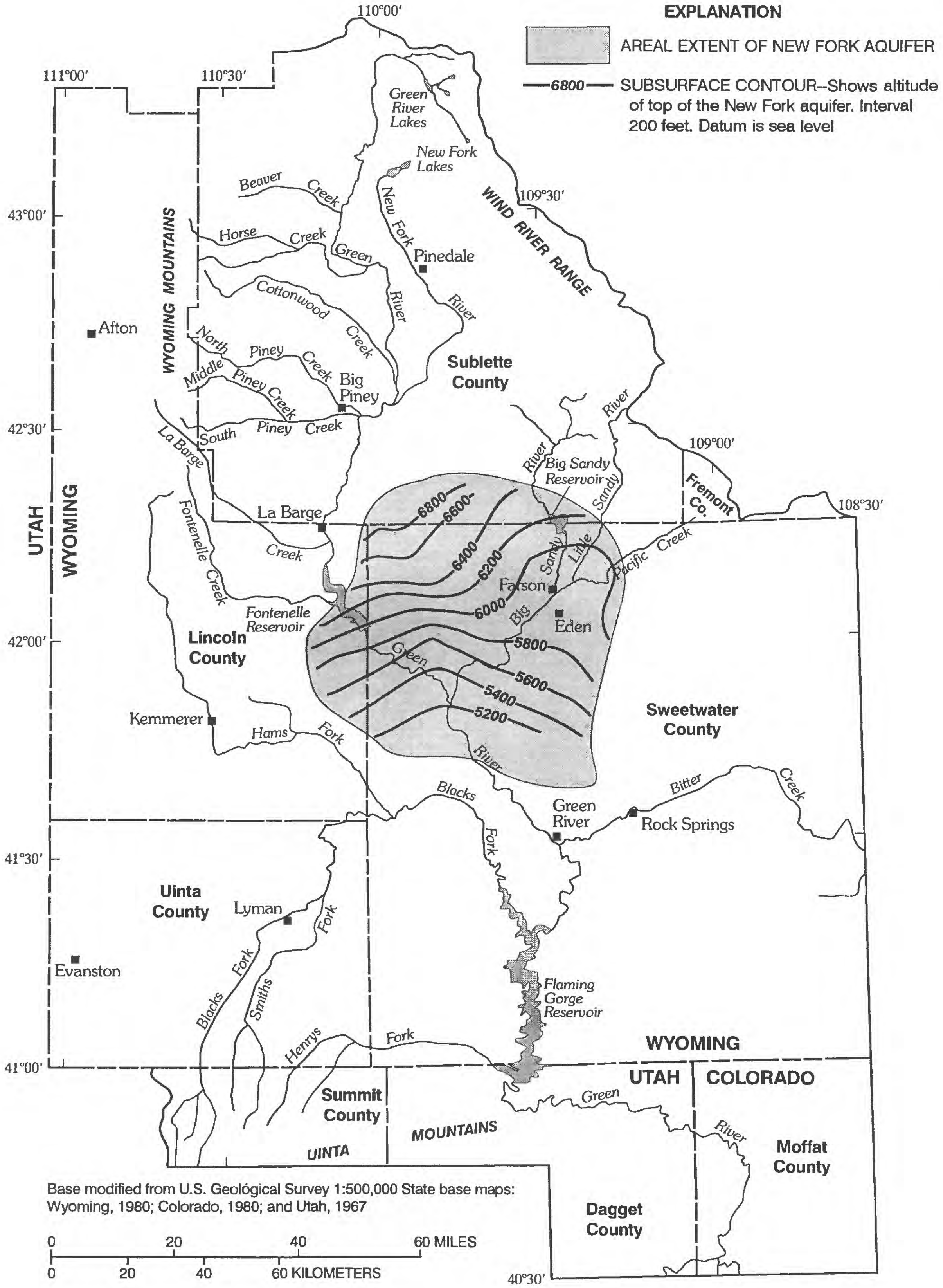


Figure 19. Altitude of the top of the New Fork aquifer, Green River Basin aquifer system (modified from Dana and Smith, 1973, p. 202).

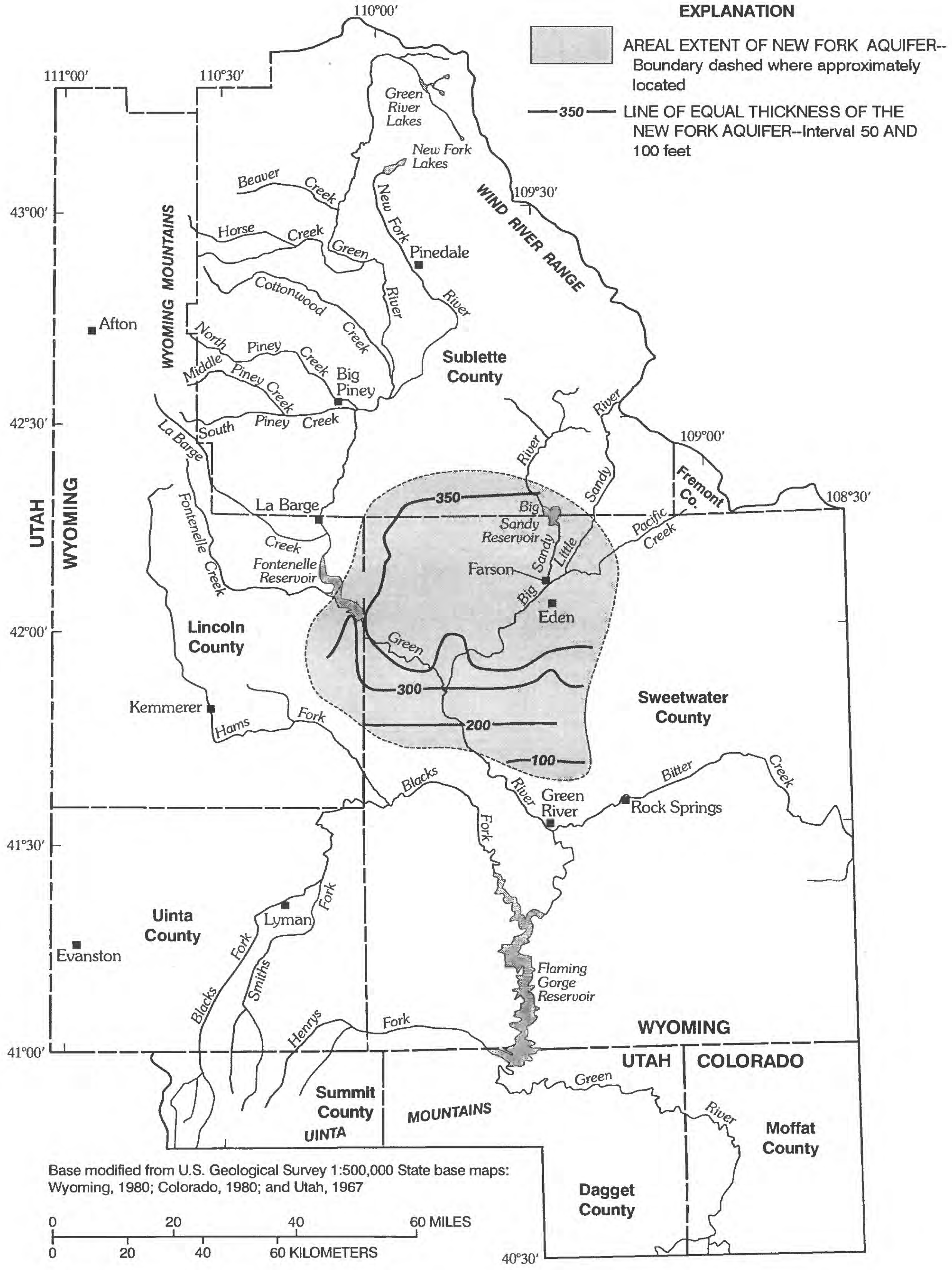


Figure 20. Areal extent and thickness of the New Fork aquifer, Green River Basin aquifer system (from Martin, 1996, p. 17).

Shale and marlstone of the Tipton Shale Member of the Green River Formation compose the relatively thin Tipton confining unit. In parts of the southern Green River Basin where the New Fork aquifer is absent, the Tipton Shale confining unit is directly below the Wilkins Peak confining unit. The Luman Member of the Green River Formation and Niland Tongue of the Wasatch Formation are included as part of the Tipton confining unit in the extreme southern part of the basin. Thickness of the Tipton Shale confining unit generally ranges from 30 to 150 ft, but the unit thickens where the Luman Member and Niland Tongue are present (fig. 21).

The Wasatch, Fort Union, and Hoback Formations present a problem in aquifer classification. Well yields and hydraulic-conductivity estimates from aquifer tests generally indicate that designation of these formations as the Wasatch-Fort Union aquifer is appropriate. However, results of a preliminary ground-water flow model of the Green River Basin do not explicitly substantiate this designation (Martin, 1996). The flow-model results indicate that the Wasatch zone functions better as an aquifer than does the Fort Union zone; however, neither zone has a large simulated hydraulic conductivity (table 1). One possible explanation for the differences between measured values and model results is related to the scale of model analysis. Sandstone lenses and other permeable rocks that greatly affect results of aquifer tests and that result in large hydraulic-conductivity estimates, may not be interconnected at basin scale. Therefore, values of hydraulic conductivity that control basin flow might be smaller than field conductivities from tests of individual sandstone beds. However, insufficient data exist to evaluate differences between local and basin effective values of hydraulic conductivity.

The Wasatch zone of the Wasatch-Fort Union aquifer includes the main body of the Wasatch Formation; a thick sequence of sandy shale and siltstone with varying quantities of coarser-grained channel sandstone. There are extensive areas in the northern half of the Green River Basin where thick permeable sandstone is at or near land surface. In the southern half of the basin, the top of the Wasatch zone is buried at altitudes between 4,000 and 6,000 ft (fig. 22). In the southern half of the Green River Basin, the quantity of sandstone in the Wasatch zone varies areally and vertically but is sufficient for large well yields along basin margins where 1,000 ft or more of saturated rock is penetrated. The Wasatch zone is present throughout the Green River Basin. Thickness typically ranges between 2,000 and 7,000 ft (fig. 23).

The Fort Union zone of the Wasatch-Fort Union aquifer includes the Fort Union and Hoback Formations. The Fort Union zone is similar lithologically to the Wasatch zone, consisting of fluvial sandstone and shale. Although deposition occurred at different times and sources of material were different, distinguishing between the two zones in outcrops, as well as in the subsurface, generally is difficult. Paleontologic and mineralogic studies, combined with a measured decrease in carbonaceous material and an increase in sandstone at the base of the Wasatch Formation, normally are needed to differentiate between the two zones. The Fort Union zone is present throughout the Green River Basin. The zone generally is buried with the altitude of the top of the unit between 0 and 6,000 ft (fig. 24). Thickness ranges between 1,000 and 4,000 ft (fig. 25).

The Wasatch and Fort Union zones extend into the Great Divide Basin north of the Rock Springs Uplift and into the Washakie Basin south of the Rock Springs Uplift (fig. 2). However, ground water does not move through Tertiary rocks between the Green River Basin and either the Great Divide or Washakie Basins. Hydraulic-head and water-quality data, presented later in this report indicate that the physiographic boundary areas at the north and south ends of the Rock Springs Uplift are ground-water recharge areas. Water entering the Wasatch and Fort Union zones in those boundary areas flows westward into the Green River Basin and eastward into the Great Divide and Washakie Basins. Therefore these physiographic boundaries also form boundaries for the basin aquifer systems within the Tertiary rocks.

The base of the Green River Basin aquifer system is defined as the contact with Upper Cretaceous strata. With the exception of the southwestern part of the basin, the Fort Union Formation overlies the Lance Formation, Fox Hills Sandstone, and the Lewis Shale. The Lance Formation and Lewis Shale are differentiated from the Fort Union Formation by a decrease in sandstone and carbonaceous material. The Lance Formation, the Fox Hills Sandstone, and the Lewis Shale are combined with the sandstone of the Mesaverde Group and Adaville Formation to form the Mesaverde aquifer (Freethey and Cordy, 1991).

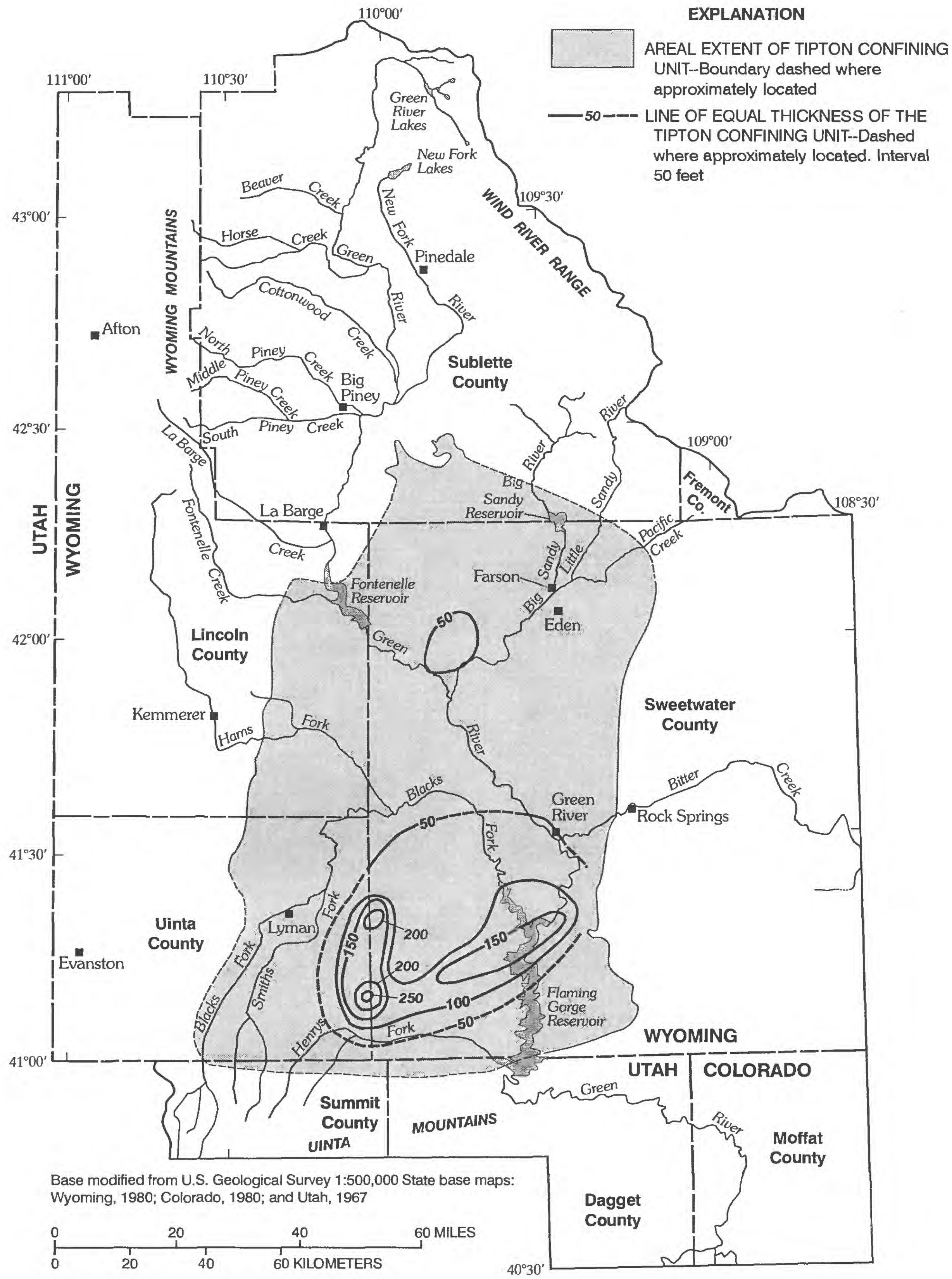


Figure 21. Areal extent and thickness of the Tipton confining unit, Green River Basin aquifer system (from Martin, 1996, p. 18).

Table 1. *Comparison of hydraulic-conductivity estimates from the Green River Basin ground-water flow model to measured values*

[modified from Martin, 1996, p. 39; --, no data]

Geohydrologic unit	Range of hydraulic conductivity, in feet per day		
	Simulated		Measured
	Vertical	Horizontal	Horizontal
Bridger aquifer	0.00001	0.09-0.9	0.03-420
Laney aquifer	.00001-17.3	.04-17.3	2-1,400
Wilkins Peak and Tipton confining units[1]	.00001	.00009	--
New Fork aquifer	.1	6.5	.2-2.0
Wasatch zone of the Wasatch-Fort Union aquifer	.001-4	.04-6.5	.03-2,100
Fort Union zone of the Wasatch-Fort Union aquifer	.00001-0.01	.00001-.3	.02-1,100

[1]The simulated values were used in the part of the confining bed modeled as a layer.

GREAT DIVIDE-WASHAKIE-SAND WASH BASINS AQUIFER SYSTEM

Tertiary rocks of the Great Divide, Washakie, and Sand Wash Basins form the Great Divide-Washakie-Sand Wash Basins aquifer system that is divided into one basin aquifer and one basin confining unit. Hydraulic head and water-quality data presented later in this report indicate that ground water in Tertiary rocks moves between the three basins, forming a single interbasin aquifer system. This interbasin aquifer system directly overlies the Mesaverde aquifer. The aquifer system is bounded on the north by the Sweetwater Uplift, on the east by the Rawlins and Sierra Madre Uplifts, on the south by the Axial Basin Arch and White River and Uinta Uplifts, and on the west by the Rock Springs Uplift (fig. 2).

Tertiary rocks of the Great Divide-Washakie-Sand Wash Basins aquifer system are continuous with Tertiary rocks of other basin aquifer systems in small areas north and south of the Rock Springs Uplift and across the Axial Basin Arch. As indicated previously, the areas near the Rock Springs Uplift are areas of ground-water recharge with ground water flowing into both the Green River Basin aquifer system and Great Divide-Washakie-Sand Wash Basins aquifer system.

Continuity of Tertiary rocks between the Sand Wash and Piceance Basins is provided by the Browns Park Formation. Justification for describing the Tertiary geohydrology of the two basins separately includes:

1. The small area of continuity in the Browns Park Formation,
2. The lack of continuity in other Tertiary formations,
3. Differences in lithology and hydraulic conductivity of the Wasatch and Fort Union Formations in the two basins, and
4. Potentiometric-surface maps indicating that the Little Snake River acts as the major discharge area for the Sand Wash Basin.

The Green River Formation and tongues of the Wasatch Formation combine to form a confining unit for underlying aquifers in Tertiary rocks. The confining unit is present in the Washakie, western Sand Wash, and western Great Divide Basins (fig. 26). The predominant lithology is shale and marlstone in the Green River Formation, and shale and fine-grained sandstone in the tongues of the Wasatch Formation. Well yields and spring discharges generally are less than 30 gal/min. The confining unit typically ranges in thickness from 1,000 to 5,000 ft (fig. 26) in the Sand Wash and Washakie Basins where it is buried by the overlying Washakie Formation. In the Great Divide Basin, the confining-unit thickness generally is less than 2,000 ft.

The Wasatch zone of the Wasatch-Fort Union aquifer is composed of the arkosic sandstone of the Battle Spring Formation and the mixed sandstone, siltstone, and coal of the Wasatch Formation. The Wasatch zone is present at land surface except where overlain by the confining unit (fig. 27) throughout the Great Divide, Washakie, and Sand Wash Basins. Thickness typically ranges from 1,000 to 4,000 ft (fig. 28).

The Fort Union zone of the Wasatch-Fort Union aquifer is composed of the interbedded sequence of sandstone, siltstone, lignite, and coal of the Fort Union Formation. No confining unit separates the Wasatch and Fort Union zones; one aquifer has been designated to simplify the discussion that follows. Nearly all of the Fort Union zone is buried throughout the Great Divide-Washakie- Sand Wash Basins aquifer system with altitudes of the top of the zone between 0 and 7,000 ft (fig. 29). Thickness typically ranges from 1,000 to 7,000 ft (fig. 30).

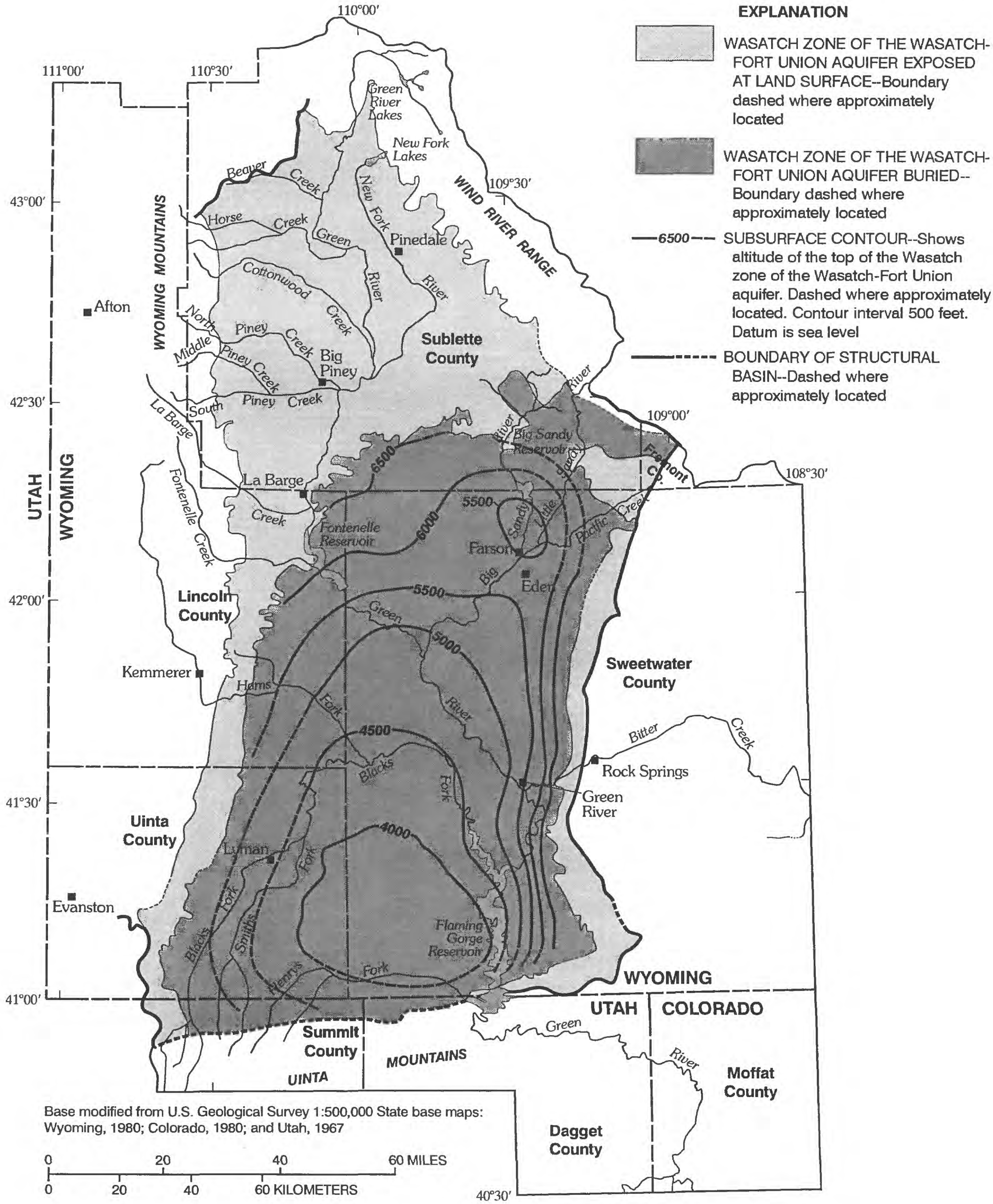


Figure 22. Altitude of the top of the Wasatch zone of the Wasatch-Fort Union aquifer, Green River Basin aquifer system (modified from Welder, 1968).

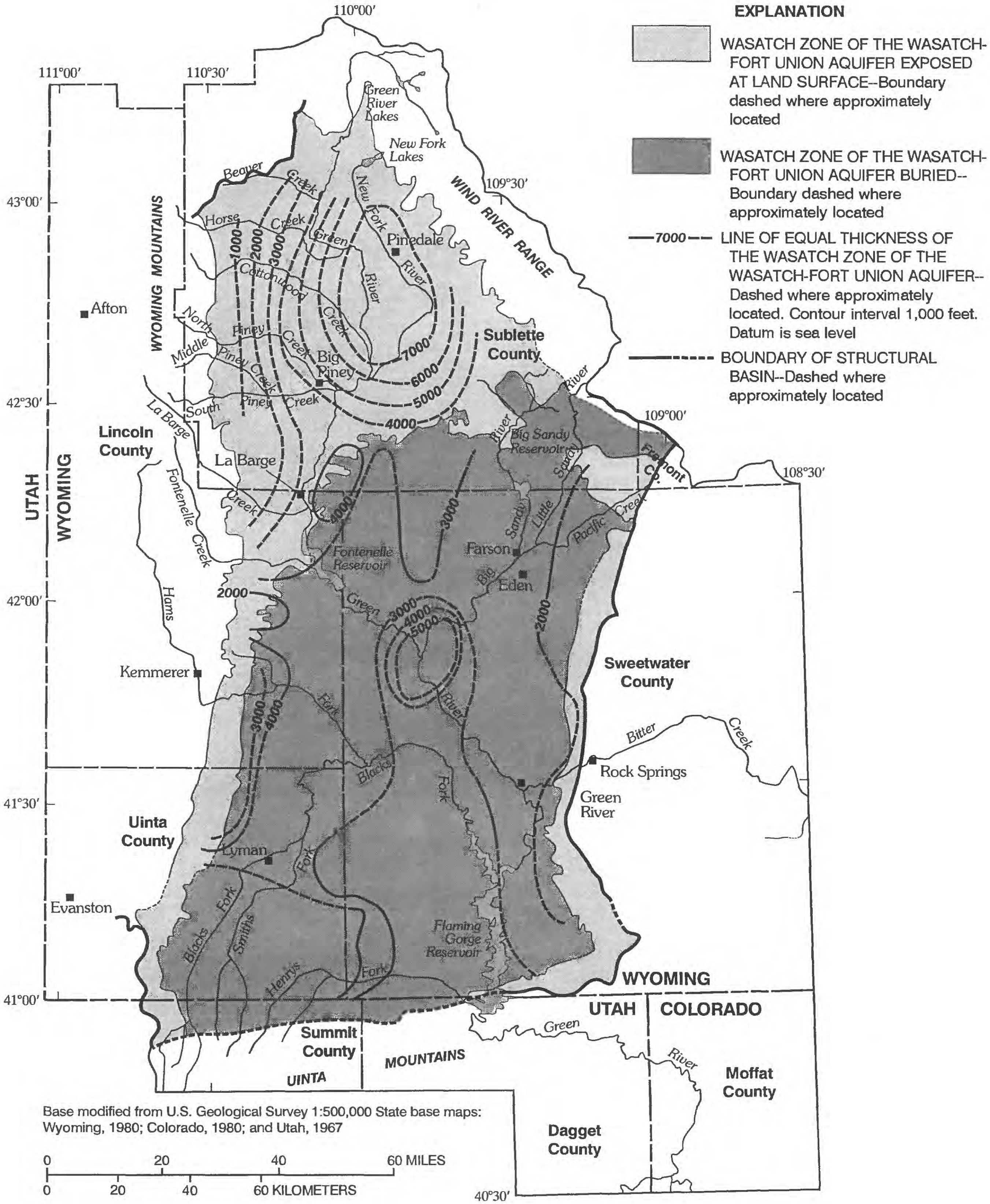


Figure 23. Areal extent and thickness of the Wasatch zone of the Wasatch-Fort Union aquifer, Green River Basin aquifer system (from Martin, 1996, p. 19).

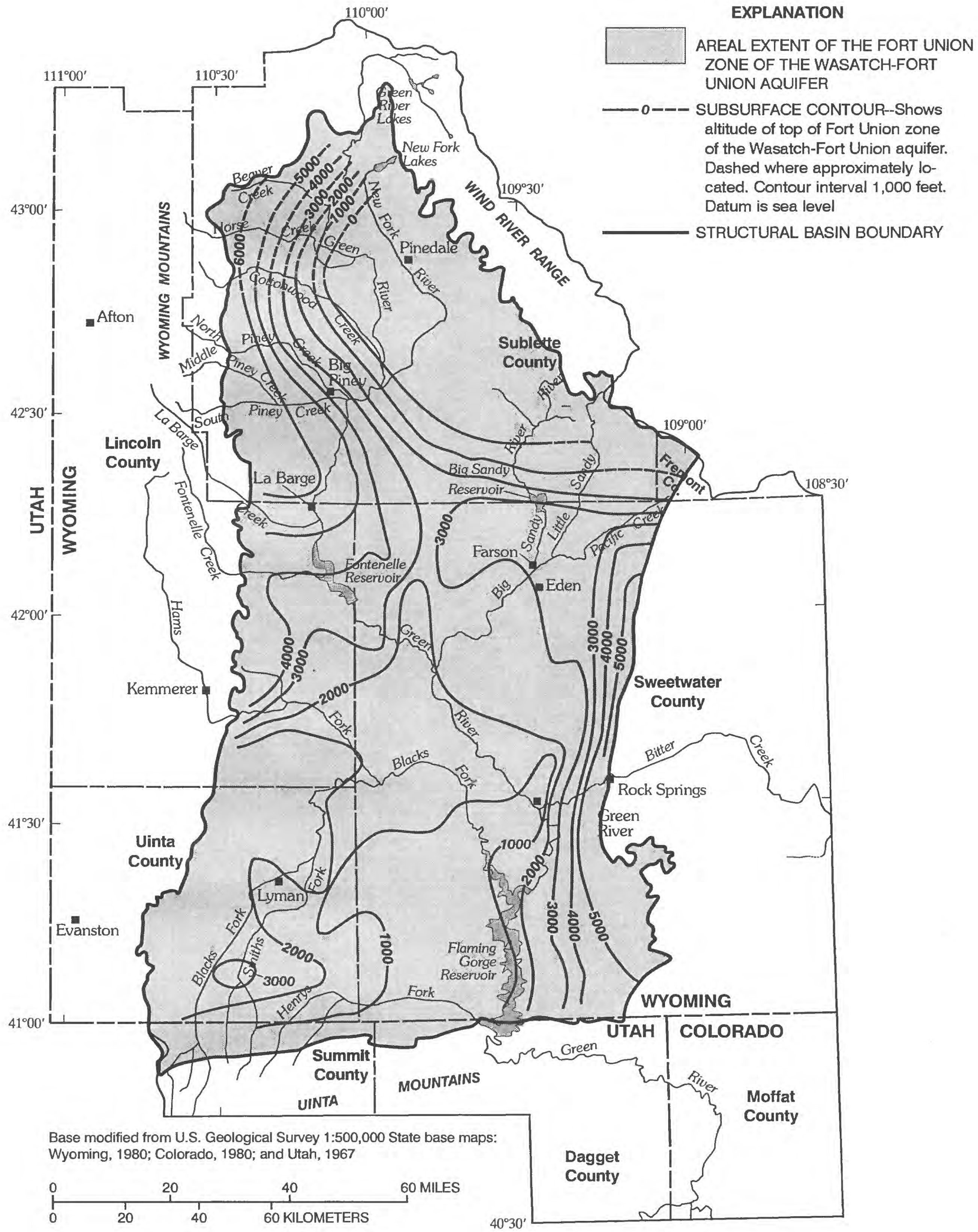


Figure 24. Altitude of the top of the Fort Union zone of the Wasatch-Fort Union aquifer, Green River Basin aquifer system.

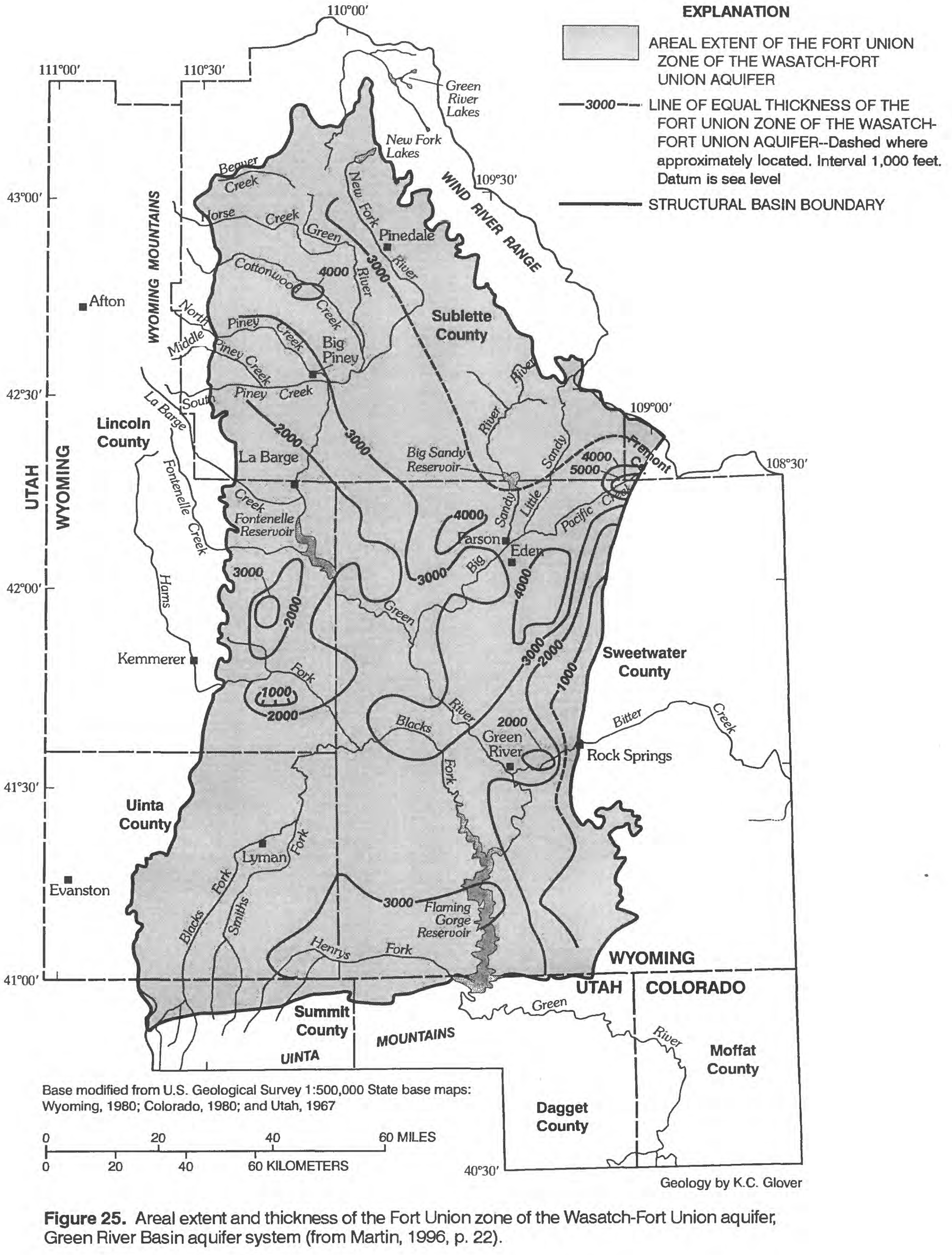


Figure 25. Areal extent and thickness of the Fort Union zone of the Wasatch-Fort Union aquifer, Green River Basin aquifer system (from Martin, 1996, p. 22).

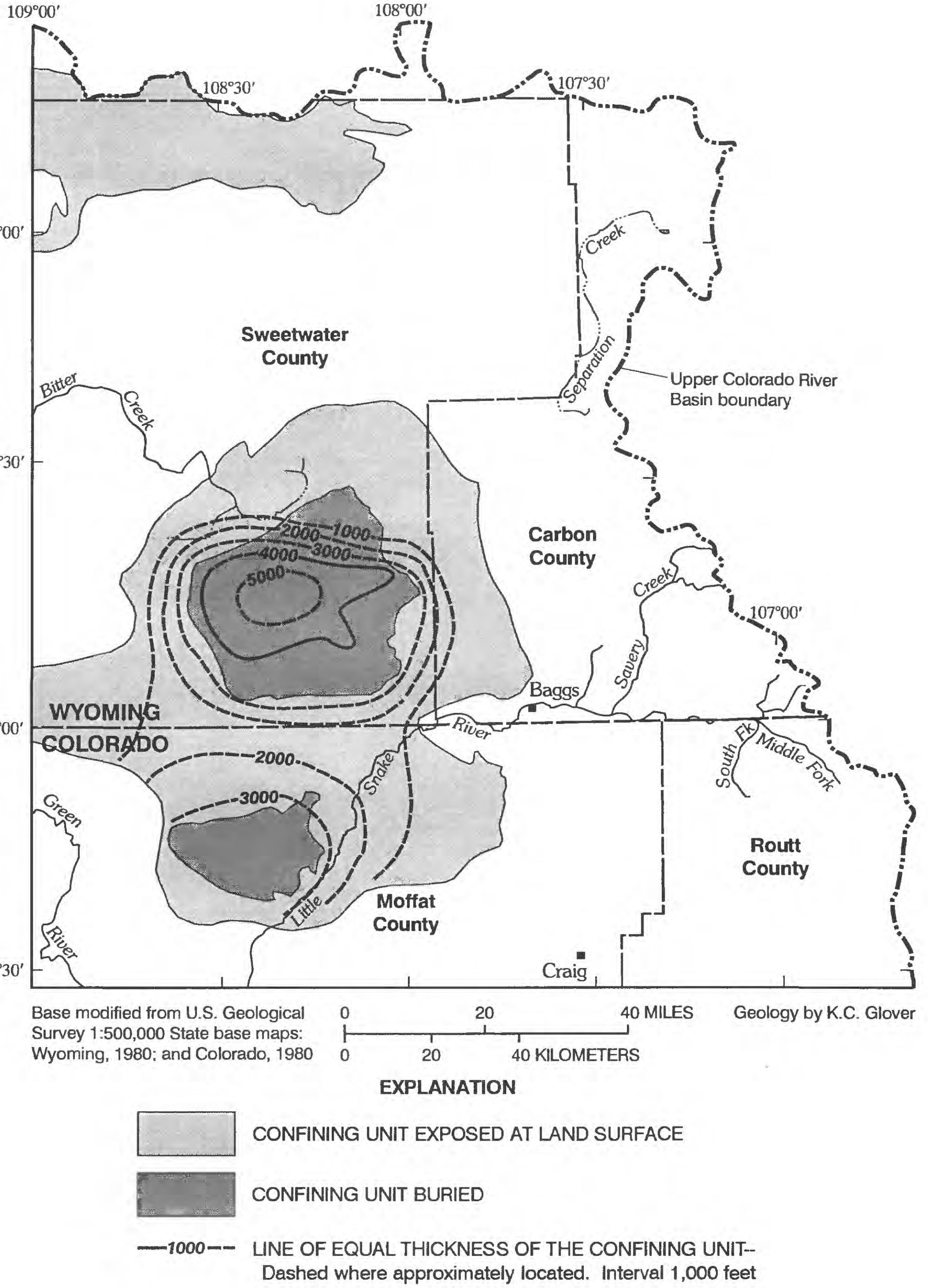
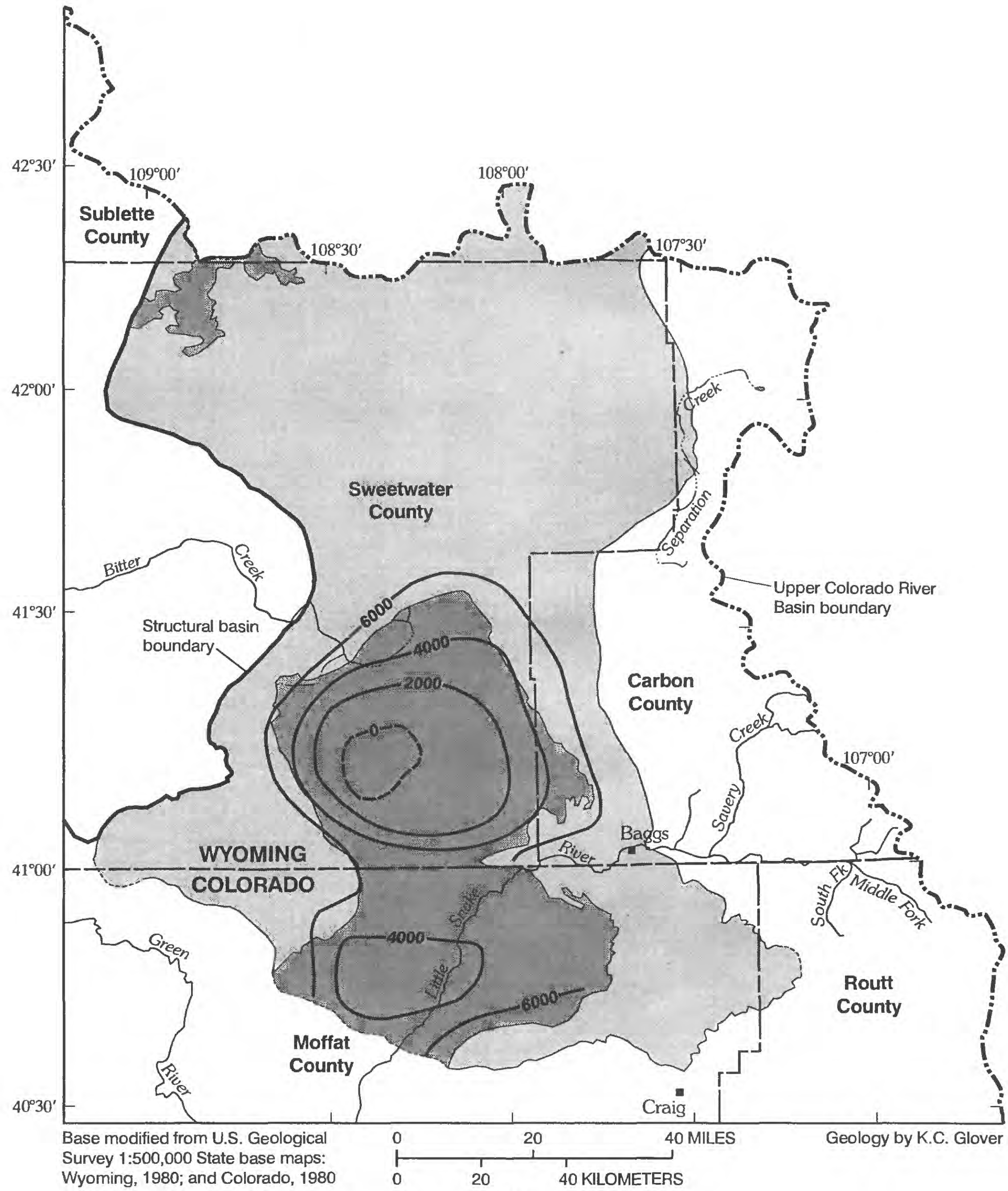


Figure 26. Areal extent and thickness of the confining unit, Great Divide-Washakie-Sand Wash Basins aquifer system (modified from Welder and McGreevey, 1966).



EXPLANATION

WASATCH ZONE OF THE WASATCH-FORT UNION AQUIFER EXPOSED AT LAND SURFACE--Boundary dashed where approximately located

WASATCH ZONE OF THE WASATCH-FORT UNION AQUIFER BURIED--Boundary dashed where approximately located

—*6000*— — SUBSURFACE CONTOUR--Shows altitude of top of Wasatch zone of the Wasatch-Fort Union aquifer. Dashed where approximately located. Contour interval 2,000 feet. Datum is sea level

Figure 27. Altitude of top of the Wasatch zone of the Wasatch-Fort Union aquifer, Great Divide-Washakie-Sand Wash Basins aquifer system (modified from Welder and McGreevey, 1966).

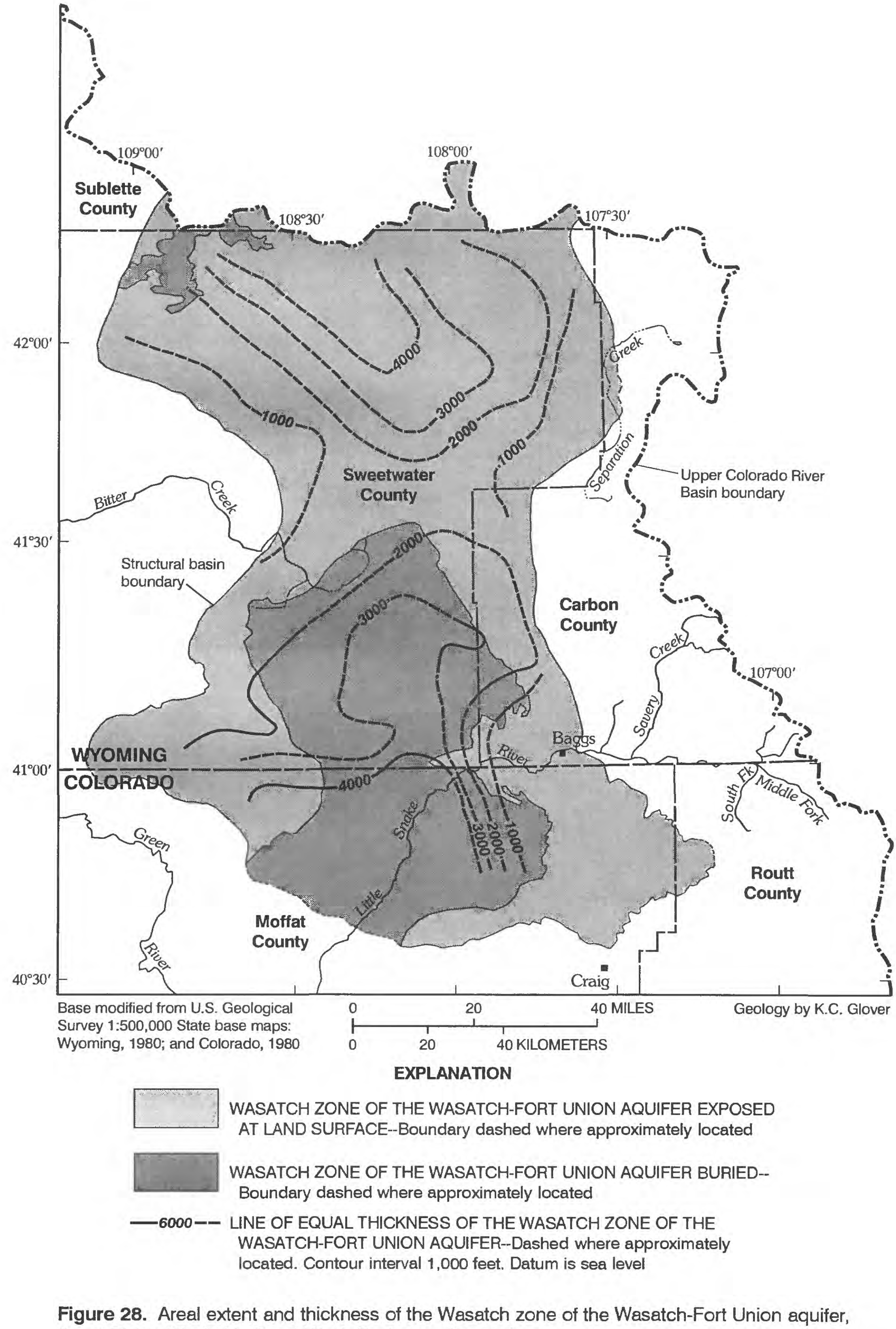
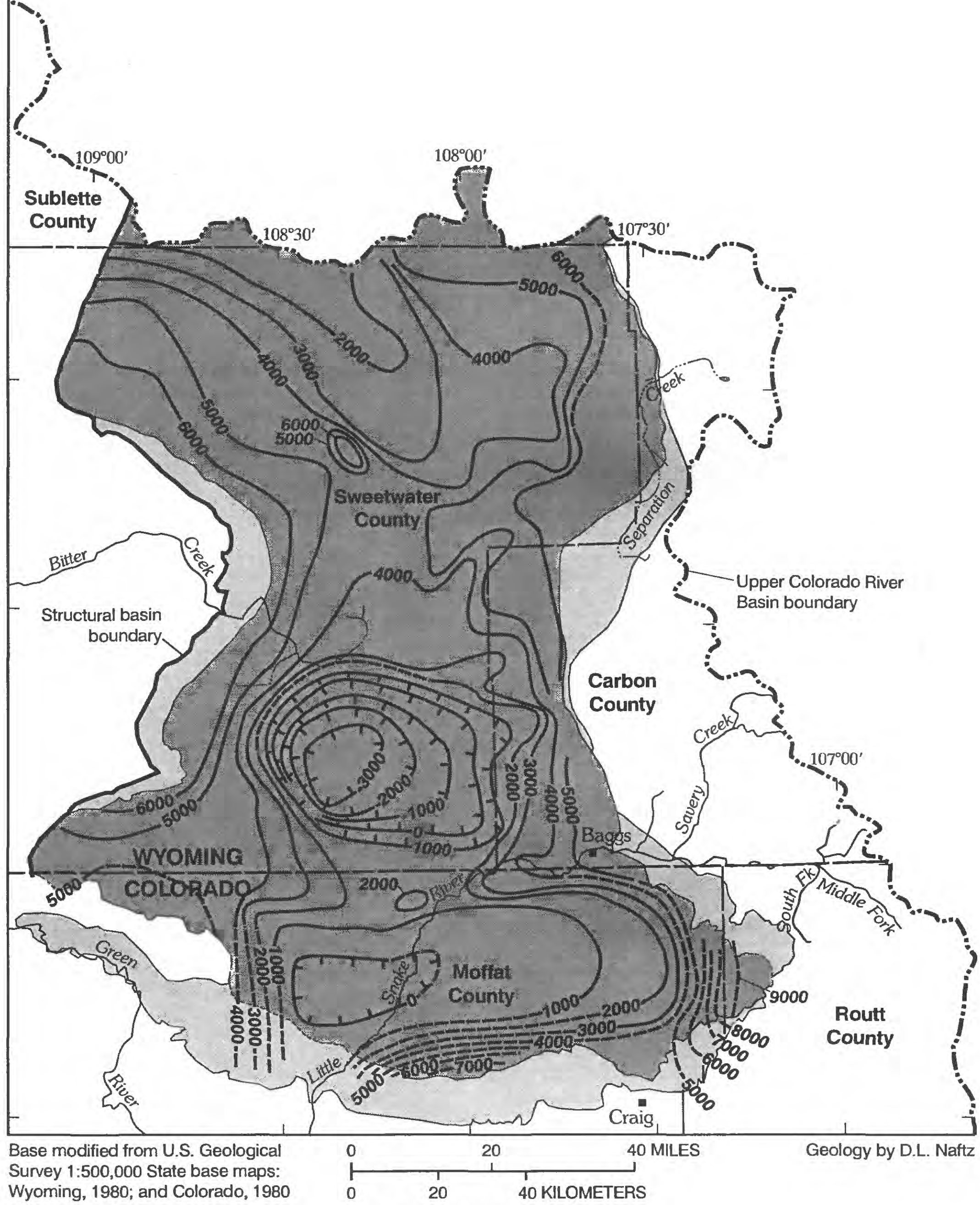


Figure 28. Areal extent and thickness of the Wasatch zone of the Wasatch-Fort Union aquifer, Great Divide-Washakie-Sand Wash Basins aquifer system.



EXPLANATION

FORT UNION ZONE OF THE WASATCH-FORT UNION AQUIFER EXPOSED AT LAND SURFACE--Boundary dashed where approximately located

FORT UNION ZONE OF THE WASATCH-FORT UNION AQUIFER BURIED--Boundary dashed where approximately located

—5000— STRUCTURE CONTOUR--Shows altitude of top of the Fort Union zone of the Wasatch-Fort Union aquifer. Hachures indicate depression. Dashed where approximately located. Contour interval 1,000 feet. Datum is sea level

Figure 29. Altitude of top of the Fort Union zone of the Wasatch-Fort Union aquifer, Great Divide-Washakie-Sand Wash Basins aquifer system.

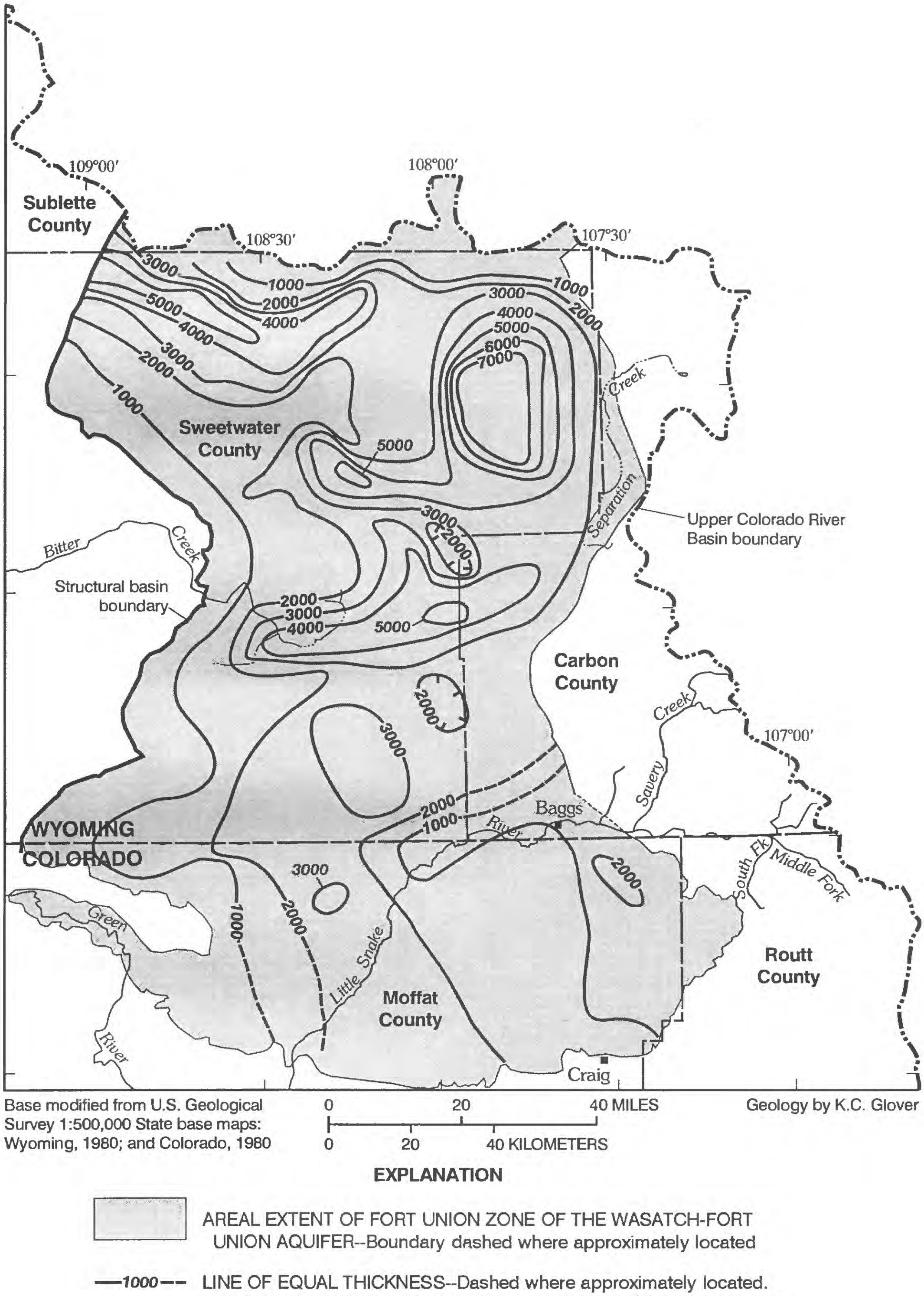


Figure 30. Areal extent and thickness of the Fort Union zone of the Wasatch-Fort Union aquifer, Great Divide-Washakie-Sand Wash Basins aquifer system.

The Fort Union zone overlies a sequence of rocks that includes in descending order the Lance Formation, Fox Hills Sandstone, and Lewis Shale, which are part of the Mesaverde aquifer (Freethey and Cordy, 1991). The Lance Formation and Lewis Shale are distinguished from the Fort Union zone by a decrease in sandstone content; the Fox Hills Sandstone is similar to the Fort Union zone of the Wasatch-Fort Union aquifer.

GEOHYDROLOGY OF THE PICEANCE BASIN AQUIFER SYSTEM

The hydrologic system in Tertiary rocks of the Piceance Basin consists of the upper and lower Piceance Basin aquifers separated from each other and from underlying aquifers in Mesozoic rocks by confining units (plate 1). Confining units are the Mahogany confining unit, which separates the upper and lower Piceance Basin aquifers, and a basal confining unit, which separates the lower Piceance Basin aquifer from the underlying Mesaverde aquifer. The aquifers are truncated laterally by topography.

HYDRAULIC CONDUCTIVITY

Sedimentary rocks of the Piceance Basin aquifer system generally are fine grained and well cemented. As a result, hydraulic conductivity of the rock matrix is very small. Sandstone and siltstone generally occur in lenticular bodies and locally have moderate hydraulic conductivity. With the exception of the upper part of the Uinta Formation, lenses of sandstone or siltstone generally are widely spaced and not interconnected. Therefore, basin values of hydraulic conductivity generally are not related to the percentage of sandstone present. Large well yields and spring discharges (table 2) typically occur only in the upper and lower Piceance Basin aquifers, units with minimal percentage of sandstone.

Hydraulic conductivity of the Piceance Basin aquifer system has been enhanced by the structural deformation of geologic formations and the dissolution of minerals. The term "fracture" is used in this report to include joints and faults, as well as rock breaks of unknown cause. Where sufficiently interconnected, fractures form the primary pathways through which ground water flows within the Tertiary rocks of the Piceance Basin. Hydraulic conductivity, therefore, is related to the number and hydraulic interconnection of fractures and variations in fracture opening along flow paths. Water that has entered the fracture network in recharge areas has dissolved and removed some of the minerals in the rock, further enhancing hydraulic conductivity.

Table 2. *Summary of well-yield, spring-discharge, and hydraulic-conductivity data, Piceance Basin aquifer system*

[gal/min, gallon per minute; ft/d, foot per day; --, not available]

Geohydrologic unit (plate 1)	Well-yield and spring-discharge rates (gal/min)	Hydraulic-conductivity data	
		Aquifer tests (ft/d)	Model analyses (ft/d)
Upper Piceance Basin aquifer	Generally between 1 and 900; occasionally larger.	0.8 to 1.2	0.003 to 1.6
Mahogany confining unit	Generally less than 25	Less than 0.01 (few tests available)	.0003 to 0.1
Lower Piceance Basin aquifer	Generally between 1 and 1,000	0.1 to 1.1	.001 to 1.2
Basal confining unit	Typically less than 10; locally as much as 100	Less than 0.01 (few tests available)	--

Studies of fracture patterns in surface exposures and mines have provided useful insight to estimating hydraulic conductivity (Verbeek and Grout, 1987). Fractures have been classified into several sets. Fractures within each set have similar orientation and appearance, and presumably are of common origin. Although several fracture sets have been recognized, only two sets are dominant and occur throughout the Piceance Basin. The primary set generally trends west-northwest to northwest; the secondary set, north-northeast to northeast, the two sets forming a nearly perpendicular network of intersecting fractures. Dips of both fracture sets are nearly vertical. These two dominant sets of fractures probably have the greatest control on basin hydrologic conditions. The remaining fracture sets commonly are of local extent and probably affect results of individual aquifer tests and local hydrologic conditions.

Within the Parachute Creek Member of the Green River Formation, investigations of fracture patterns have shown that bed thickness, rock type, and depth below land surface affect fracture spacing (Verbeek and Grout, 1987). For any given rock type, fracture spacing increases as bed thickness increases. Based on very limited data, fracture spacing within beds of uniform rock type also appears to increase with increasing depth. In addition, the two dominant fracture sets become more widely spaced with increasing richness of oil shale. Information is not available to describe lateral variations in fracture spacing within a single bed as depth increases.

Bed thickness also affects fracture size (Verbeek and Grout, 1987). Fracture surface area increases with increasing bed thickness. Fracture surface area of the second dominant set (north-northeast to northeast) increases with increasing bed thickness, but also depends on the spacing of the first dominant set (west-northwest to northwest). In general, the first dominant fracture set represents long fractures that cut across bedding planes where rock type is relatively uniform. Where different rock types are interbedded, the first dominant set typically consists of short, strata-bound fractures. The second dominant set of fractures rarely cuts across other fractures or bedding planes but probably is important hydraulically by providing interconnection between other sets of fractures.

Hydrologic research to describe relations between fracture patterns and hydraulic conductivity has been documented recently (Long and others, 1985; Neuman, 1987). However, these investigations have concentrated efforts at scales of an aquifer test or smaller. Proven techniques for application at a basin scale are not available. Nevertheless, several qualitative generalizations regarding relations between fracture patterns and hydraulic conductivity are useful in understanding the basin distribution of hydraulic conductivity in the Piceance Basin. Hydraulic conductivity is strongly influenced by variations in the spacings, surface areas, and interconnections of the two dominant fracture sets. Hydraulic conductivity parallel to bedding planes may be directionally dependent. If they are directionally dependent, the direction of greatest hydraulic conductivity within a bed will not necessarily be parallel to the average orientation of either set of joints. Hydraulic conductivity generally increases with decreasing fracture spacing, or increasing fracture surface area and opening.

The preceding discussion of fracture patterns and the relation to hydraulic conductivity serves as a useful guide to evaluate results of aquifer tests and hydrologic-model development. However, before the results of this evaluation are presented, hydraulic-conductivity estimates from aquifer tests, and hydrologic models are summarized. Detailed descriptions of aquifer tests are provided by Ficke and others (1974); Weeks and Welder (1974); Weeks and others (1974); Dale and Weeks (1978); and Welder and Saulnier (1978); and Loo and others (1979). Descriptions of hydrologic models are provided by Robson and Saulnier (1981); and Taylor (1982, 1986).

Aquifer-test data are available for both upper and lower Piceance Basin aquifers; however, test locations are not well distributed throughout the basin. Much of the data was collected from oil-shale core holes and is concentrated in the vicinity of oil-shale lease tracts. Few reliable tests have been conducted in the drainages of Roan and Parachute Creeks. Most aquifer tests have been conducted without the benefit of multiple observation wells. In fractured rock, where hydraulic conductivity is expected to be anisotropic and highly variable, single-well aquifer tests can produce unreliable results.

Robson and Saulnier (1981) used aquifer-test data as a basis for estimating the basin distribution of horizontal hydraulic conductivity; however, aquifer-test data could not be used to estimate the distribution of vertical hydraulic conductivity. Because most sets of fractures are not extensive throughout the basin, large spatial variations in hydraulic conductivity within an aquifer or confining unit are expected. The aquifer-test data support this conclusion. The nonhomogeneous character of the fractures throughout the fractured rock made determination of effective basin hydraulic-conductivity values based on data from scattered test locations extremely difficult. As a result, initial estimates of hydraulic conductivity were modified during the development of a hydrologic model. Modifications to horizontal hydraulic-conductivity values were slight (S.G. Robson, U.S. Geological Survey, oral commun., 1988). Most modifications were made to the vertical hydraulic conductivity. The model of Robson and Saulnier included five layers; two to represent the upper aquifer, one to represent the Mahogany confining unit, and two to represent the lower Piceance Basin aquifer. The drainage basins of Roan and Parachute Creeks were not included in the model.

Taylor (1982, 1986) developed a hydrologic model that encompassed the entire Piceance Basin aquifer system and utilized a finer mesh finite-difference grid. Based on data not available to Robson and Saulnier (1981), Taylor (1982) more accurately described anisotropy in hydraulic conductivity. The additional data sources, while not extensive, included an analysis of temperature logs from wells tightly cemented across the Mahogany confining unit (Taylor, 1982), and analyses of multiple-well aquifer tests in the upper Piceance Basin aquifer (Taylor, 1982) and lower Piceance Basin aquifer (Loo and others, 1979). Although the horizontal hydraulic-conductivity estimates of Robson and Saulnier (1981) were adjusted, the resulting basin patterns in the two models are fundamentally different. Ratios of horizontal to vertical hydraulic conductivity are significantly different between the models. Because the ratios of Taylor (1982, 1986) are based on data that were not available to Robson and Saulnier, they probably are more reliable.

Generalized basin distributions of hydraulic conductivity (figs. 31-33) indicate that the hydraulic conductivity of the Mahogany confining unit is small compared to the hydraulic conductivity of the two adjacent aquifers. Thickly bedded, rich oil shale of the Mahogany confining unit is characterized by widely spaced fractures. Lithologic conditions in overlying and underlying aquifers have resulted in enhanced fracturing, which has effectively increased hydraulic conductivity. The generalized basin distributions were compiled from more detailed distributions presented by Robson and Saulnier (1981) and Taylor (1982, 1986). The generalized basin distributions can be used to identify trends, but should not be used to predict local hydrologic response to ground-water injection and withdrawal or to predict ground-water contamination. For predictive studies, local, more detailed distributions are needed.

Generalized basin distributions of hydraulic conductivity indicate that hydraulic conductivity generally increases in a northerly direction toward the central part of Rio Blanco County (figs. 31-33). This trend is most apparent in the lower Piceance Basin aquifer and may reflect the enhancement of hydraulic conductivity by dissolution of minerals or increased fracture density or fracture interconnection. Although geologic data are not available to test these assumptions, the trend is supported by aquifer-test data. Within the northern part of the Piceance Basin, hydraulic conductivity appears to be enhanced along a narrow band below Piceance Creek. This pattern is particularly apparent in the Mahogany confining unit. While some aquifer-test data are available to support this pattern, principal justification appears to be that such a pattern was needed to adequately simulate measured hydraulic-head and aquifer-discharge data. Robson and Saulnier (1981, p. 7) note that tectonic activity may have affected the distribution of fractures and concentrated stream channels in areas of enhanced fracturing. If this assumption is true, the coincidence of enhanced hydraulic conductivity with Piceance Creek would have a geologic basis. The ratio of horizontal to vertical hydraulic conductivity appears to increase with depth. Taylor (1982, 1986) reports ratios of 2.0 in the upper Piceance Basin aquifer, 3.3 in the Mahogany confining unit, and from 13.4 to 15.0 in the lower Piceance Basin aquifer. Aquifer-test data to support these ratio estimates are not available. However, fracture-mapping studies suggest a possible geologic basis for the estimates. The presence of alternating beds of rich and lean oil shale in the lower Piceance Basin aquifer probably results in fractures that are stratigraphically bound and generally shorter than fractures in the upper Piceance Basin aquifer. If so, fracture interconnection and hydraulic conductivity in the vertical direction would be less in the lower Piceance Basin aquifer than in the upper Piceance Basin aquifer. At the same time, fracture interconnection and hydraulic conductivity in the horizontal direction may not be impaired significantly by stratigraphically bound fractures.

HYDRAULIC HEAD

Adequate data are available to map two potentiometric surfaces in the Piceance Basin, one for the upper Piceance Basin aquifer (fig. 34) and a second for the lower Piceance Basin aquifer (fig. 35). Hydraulic heads shown in these maps represent water levels in wells that penetrate the entire saturated thickness of each aquifer. Although the use of thinner stratigraphic intervals would have indicated vertical and lateral hydraulic gradients in the upper and lower aquifers more accurately, the basin-wide scarcity of head data would have decreased the overall accuracy of the resulting potentiometric-surface maps. The potentiometric-surface maps compiled in this study are adequate for identifying general northward direction of ground-water flow in the basin, as well as for identifying general areas of upward or downward leakage.

GROUND-WATER RECHARGE AND DISCHARGE

Winter precipitation, stored as snowpack in the higher altitudes of the Piceance Basin, provides most of the recharge to the ground-water system. Areas of substantial natural recharge are shown in figure 36. During the spring, snow gradually melts, and part of the melt water infiltrates soil and rock outcrops and eventually percolates to the saturated zone of the aquifer system. The volume of recharge increases as depth of snowpack increases and is greater at higher land-surface altitudes (Weeks and others, 1974). Little, if any, rainfall infiltrates and percolates to the saturated zone during the summer.

The rate and distribution of recharge have been estimated using precipitation-runoff models (Weeks and others, 1974) and ground-water-flow models (Taylor, 1982, 1986). Initial recharge estimates were obtained by developing a precipitation-runoff model

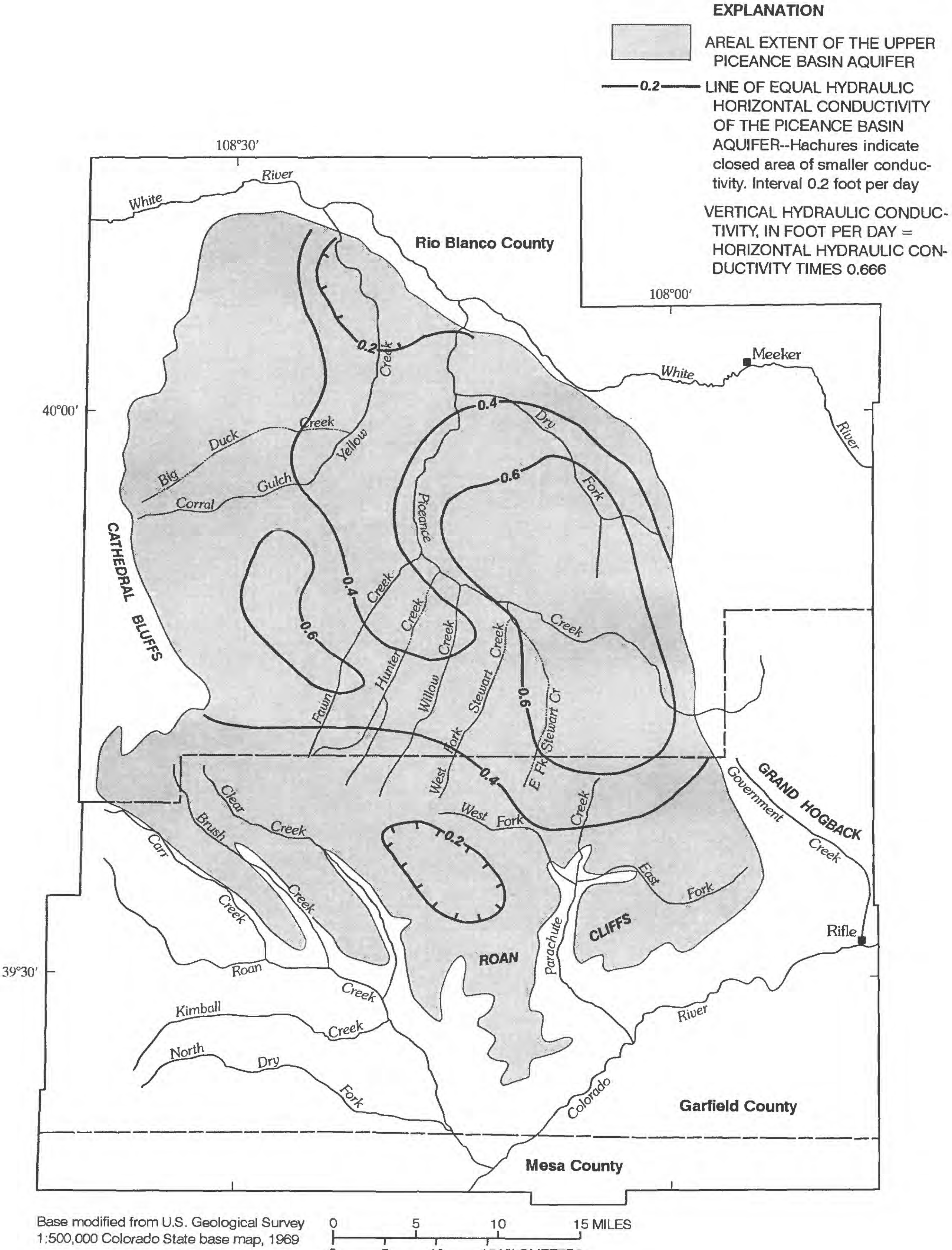


Figure 31. Hydraulic conductivity of the upper Piceance Basin aquifer, Piceance Basin aquifer system (from Taylor, 1982).

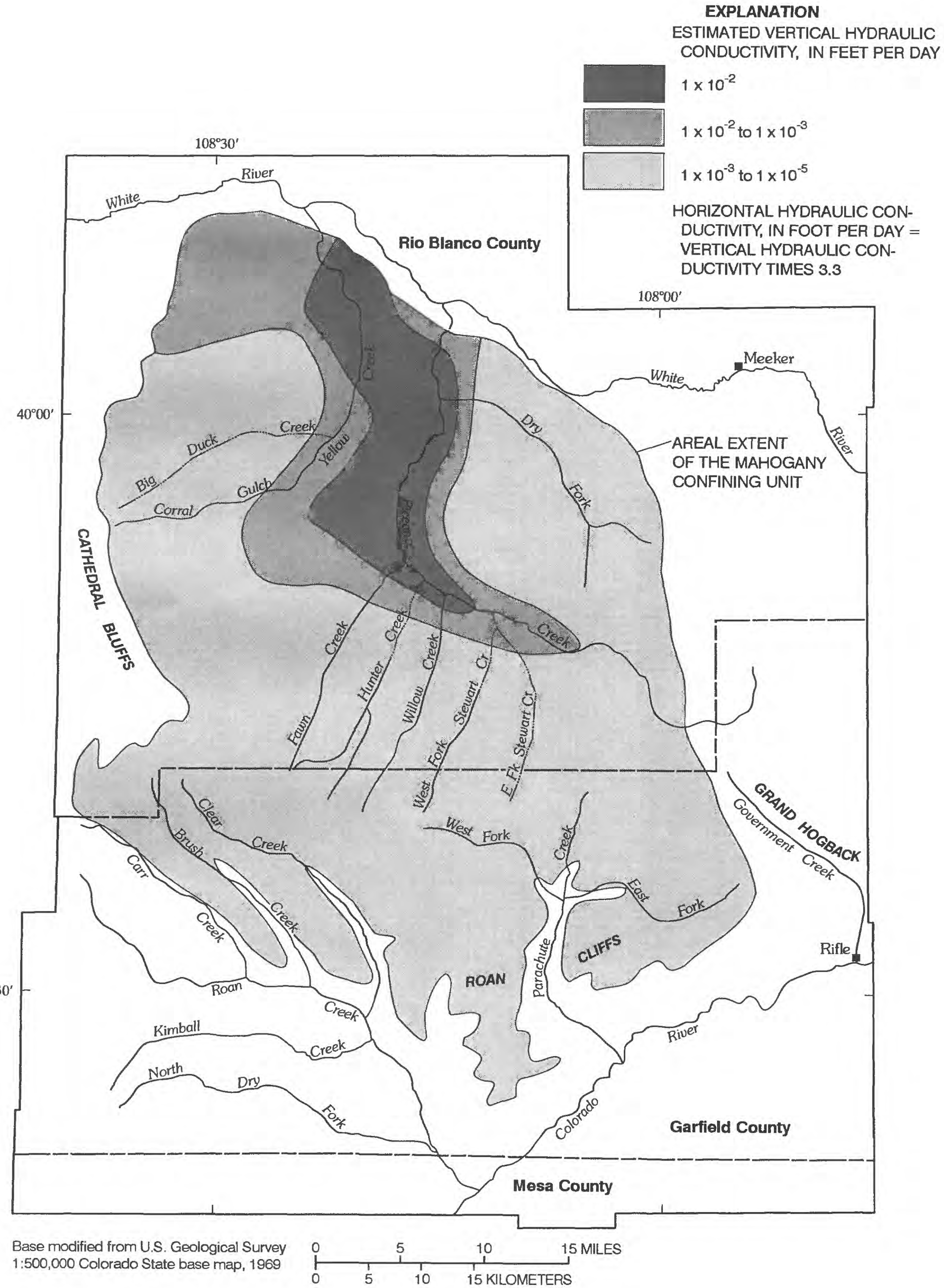


Figure 32. Hydraulic conductivity of the Mahogany confining unit, Piceance Basin aquifer system (from Taylor, 1982).

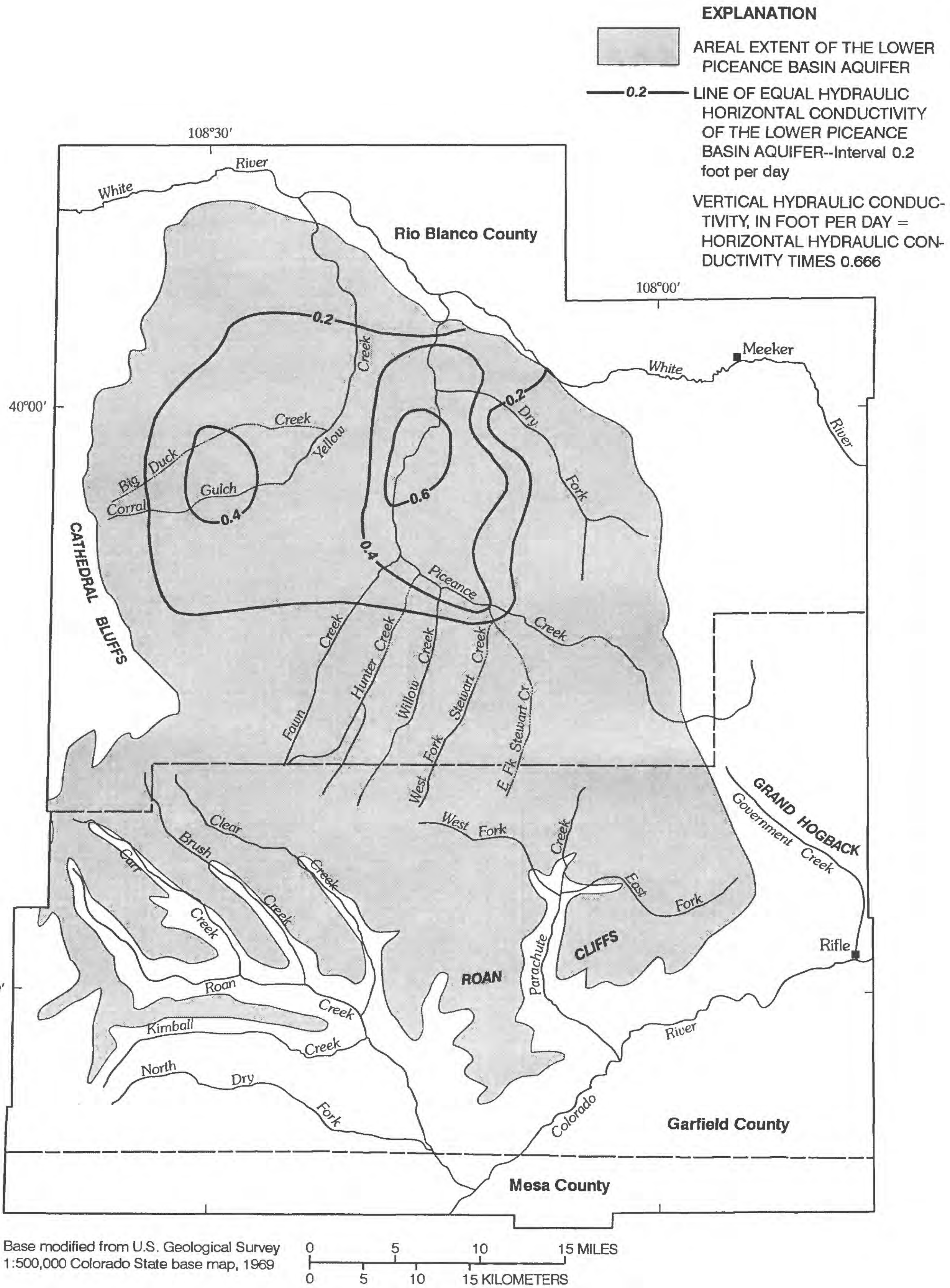


Figure 33. Hydraulic conductivity of the lower Piceance Basin aquifer, Piceance Basin aquifer system (from Taylor, 1982).

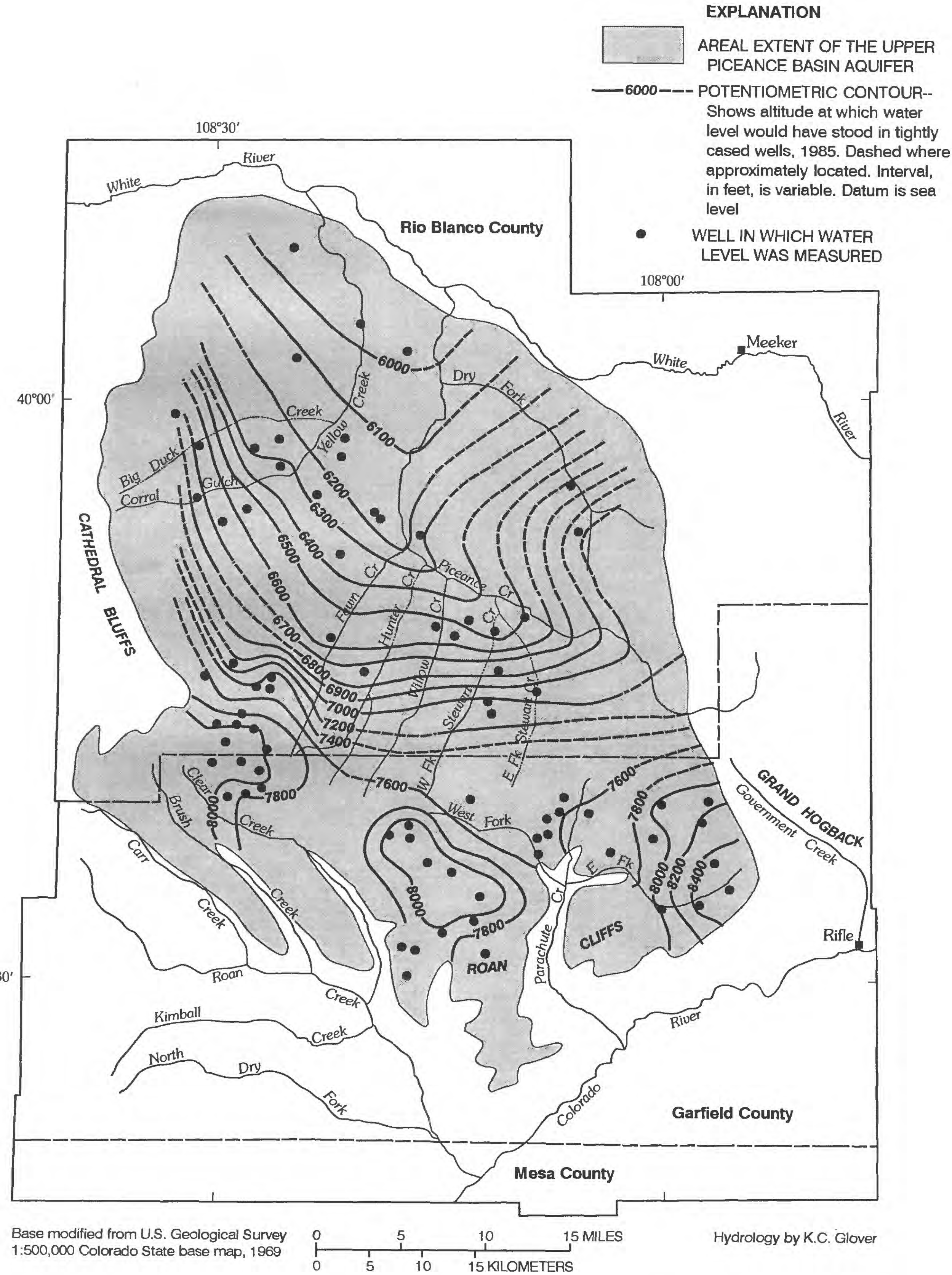


Figure 34. Potentiometric surface (1985) of the upper Piceance Basin aquifer, Piceance Basin aquifer system.

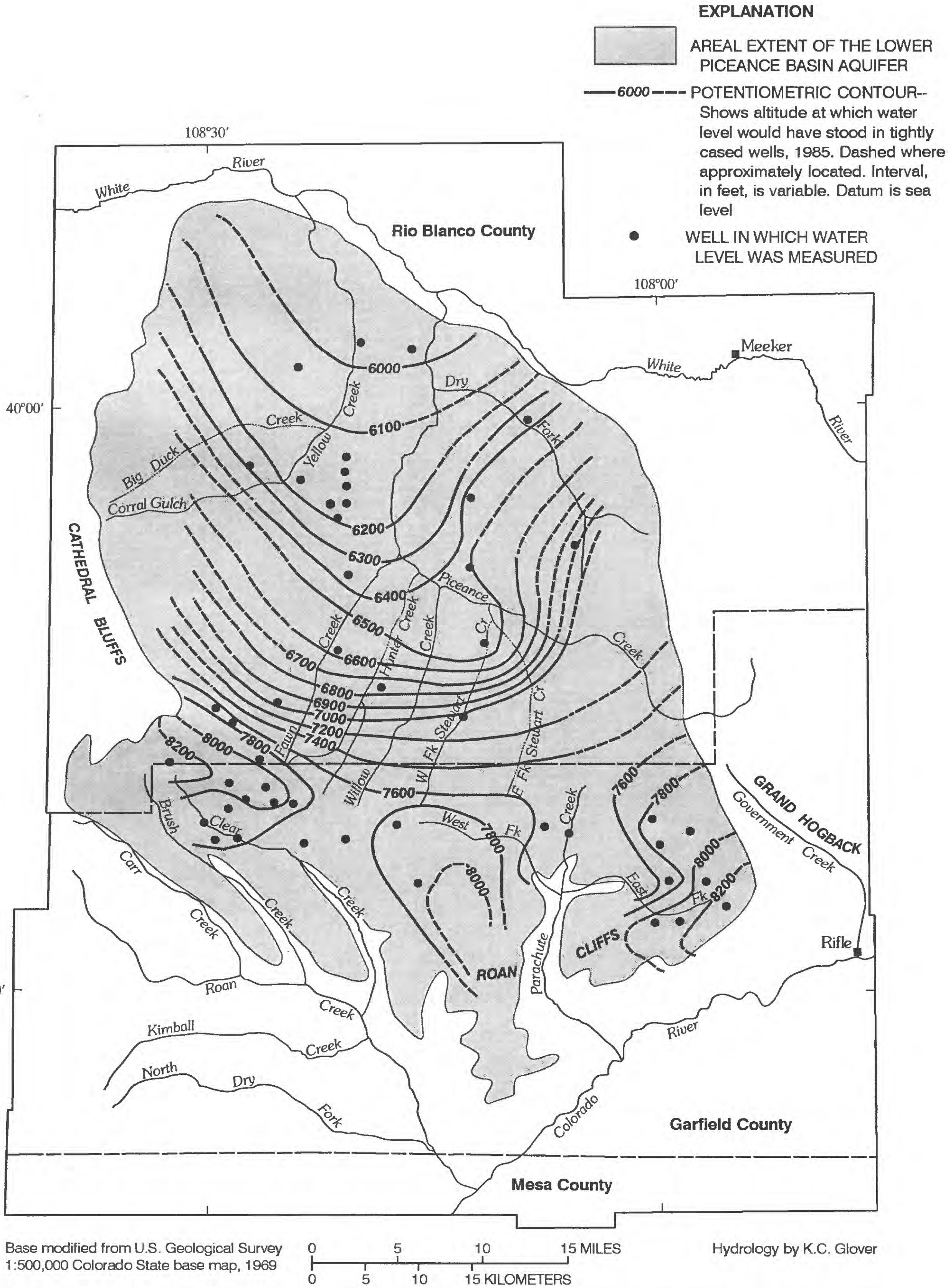


Figure 35. Potentiometric surface (1985) of the lower Piceance Basin aquifer, Piceance Basin aquifer system.

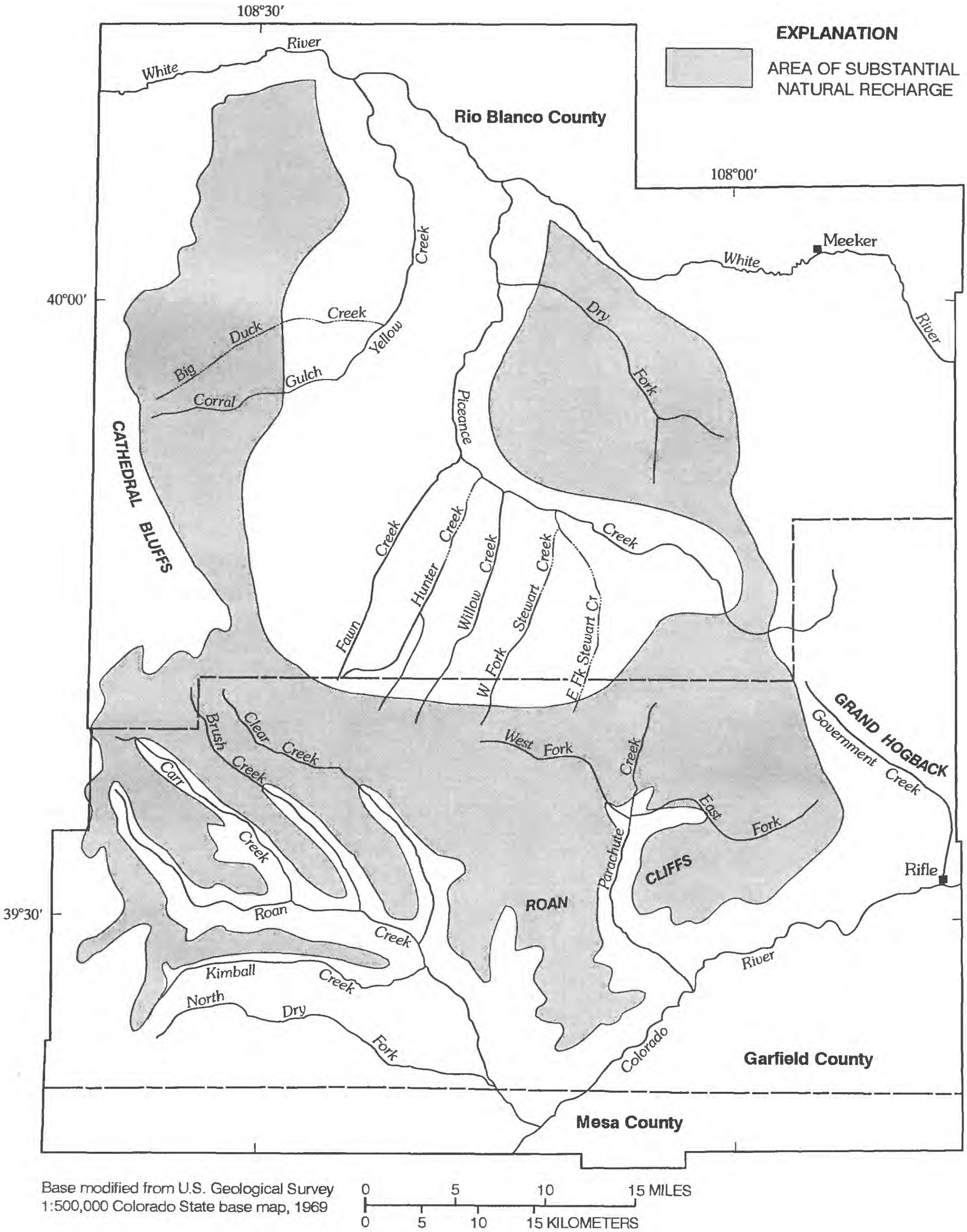


Figure 36. Approximate area of substantial natural recharge, Piceance Basin aquifer system (from Taylor, 1982).

for parts of the basin drained by Piceance Creek (Weeks and others, 1974) and by assuming a linear relation between recharge and land-surface altitude in the remainder of the basin (Taylor, 1982). Adjustments were made, assuming steady-state flow, to ensure that aquifer-system recharge equaled aquifer-system discharge. Estimates of discharge are considered reasonably accurate. Recharge was distributed in quantities approximately proportional to the land-surface altitude and ranged from 0 to 2.3 in/yr. Total recharge to the Piceance Basin aquifer system north of the Colorado River was estimated to be about 42 ft^3/s and included recharge from streams of 0.7 ft^3/s (Taylor, 1986).

A variety of conditions causes water to discharge from the Piceance Basin aquifer system. In parts of the Piceance Creek and Yellow Creek drainages, discharge from the bedrock aquifers occurs through leakage into the valley-fill alluvium that, in turn, yields water to springs. In other places along these drainages, ground water discharges directly to springs near the streams. In contrast, ground water in the Roan Creek and Parachute Creek drainage basins discharges to springs and seeps on canyon walls above the streams.

Streamflow analyses by Taylor (1986) were used to estimate natural discharge from aquifers. The analyses used gain-and-loss studies along Piceance and Yellow Creeks and low-flow statistics for Roan and Parachute Creeks. These estimates were refined during development of a steady-state flow model to ensure consistency with measured hydraulic-head data. Ground-water discharge to springs and stream alluvium in the drainages of Yellow and Piceance Creeks was estimated to be 30.7 ft^3/s, and discharge to springs in the drainages of Roan and Parachute Creeks was estimated to be 11.6 ft^3/s.

GROUND-WATER MOVEMENT

The potentiometric-contour maps of the upper and lower Piceance Basin aquifers (figs. 34 and 35) suggest flow directions and areas of recharge and discharge. These recharge areas include the Grand Hogback, Cathedral Bluffs, and the Roan Cliffs (fig. 36). Part of the recharge water flows downward through the upper aquifer and Mahogany confining unit into the lower aquifer; part flows in a generally horizontal direction toward discharge areas. Water in the lower aquifer flows horizontally toward discharge areas and returns to the upper aquifer in the vicinity of the discharge areas. Stable-isotope and carbon-14 data interpreted by Kimball (1984) confirm that significant mixing between water of the upper and lower aquifers occurs in the Piceance Basin. Discharge from the ground-water system occurs through leakage into alluvium along major streams or through springs and seepage on canyon walls above stream levels.

Areas of substantial vertical leakage between the upper and lower Piceance Basin aquifers have been identified by comparing the potentiometric-surface maps of the aquifers. The east, south, and west margins of the Piceance Basin are areas of possible downward leakage, because the head in the upper aquifer is greater than the head in the lower aquifer. Upward leakage is likely in areas underlying Piceance and Yellow Creeks. Rates of ground-water movement have been estimated during development of a basin ground-water flow model (Taylor, 1986). The resulting ground-water budget for the Piceance Basin aquifer system (table 3) represents present day, relatively undeveloped, steady-state conditions. Model-derived estimates of ground-water discharge total 42.3 ft^3/s, which are similar to estimates determined from streamflow gain-and-loss studies, and therefore are likely to be reliable. Because the system is under steady-state conditions, estimates of recharge also are likely to be reliable.

Table 3. *Estimated ground-water budget, Piceance Basin aquifer system*

[From Taylor, 1986]

Recharge, in cubic feet per second:	
Precipitation	41.6
Losing streams	0.7
Total	42.3
Discharge, in cubic feet per second:	
Yellow Creek and Piceance Creek	30.7
Springs, near Roan Creek and Parachute Creek	11.6
Total	42.3

QUALITY OF WATER

Water in the Piceance Basin aquifer system gains dissolved solids and exhibits changes in major-ion chemistry as it moves along basin flow paths from upland recharge areas to discharge areas. In the upper Piceance Basin aquifer, the dissolved-solids concentration increases from about 500 to 1,000 mg/L (fig. 37). Water type in the upper aquifer is diverse and can range from calcium-carbonate to sodium-carbonate water with large concentrations of sulfate. In the lower Piceance Basin aquifer, the dissolved-solids concentration increases from about 1,000 to 10,000 mg/L along basin flow paths (fig. 38). Water in the lower aquifer is characterized by a large concentration of sodium carbonate. According to Robson and Saulnier (1981), possible reactions governing observed changes in water quality in recharge areas include dissolution of calcite and dolomite. Possible chemical reactions in downgradient areas include dissolution of nahcolite and halite, precipitation of calcite and pyrite, exchange of calcium in the water for sodium in clay minerals, and sulfate reduction.

GEOHYDROLOGY OF THE UINTA BASIN AQUIFER SYSTEM

Two major aquifers have been identified in the Tertiary rocks of the Uinta Basin: the Duchesne River-Uinta and Douglas Creek-Renegade aquifers. The Duchesne River Formation and the Uinta Formation compose the Duchesne River-Uinta aquifer. The Douglas Creek Member of the Green River Formation and the associated Renegade Tongue of the Wasatch Formation compose the Douglas Creek-Renegade aquifer. The stratigraphic relation of these two aquifers and associated confining units is shown on plate 1. Stratigraphic and lithologic descriptions of these aquifers and confining units were given previously in this report. The aquifers are truncated to the east, south, and west by topography and to the north by older rocks of the Uinta Uplift.

HYDRAULIC CONDUCTIVITY

Hydraulic conductivity of the Uinta Basin aquifer system is related to the percentage of sandstone in each geohydrologic unit and to the degree of fracturing. Fractures have enhanced hydraulic conductivity in the Duchesne River-Uinta aquifer. Fractures are particularly important in the lower part of the aquifer and within the central part of the Uinta Basin where sandstone tends to be fine-grained and well-cemented. In contrast, hydraulic conductivity of the Douglas Creek-Renegade aquifer is related primarily to the percentage of sandstone. Sandstone is more common and sandstone beds have greater interconnection in the Douglas Creek-Renegade aquifer than in the Duchesne River-Uinta aquifer. Fractures, although present in the Douglas Creek-Renegade aquifer, are secondary in importance. The Parachute Creek and Wasatch-Green River confining units consist primarily of shale and secondarily of generally isolated beds of fine-grained sandstone. Confining units are virtually unfractured and have small values of hydraulic conductivity. Well-yield and spring-discharge data reflect the estimated ranges of hydraulic conductivity, as shown in table 4.

Hydraulic conductivity of the Duchesne River-Uinta aquifer is greatly enhanced by fractures. Laboratory measurements of unfractured cores (Hood, 1976) indicate that hydraulic conductivity of the matrix is small, ranging from 0.000033 to 3.3 ft/d for the Duchesne River Formation and from 0.16 to 0.32 ft/d for the Uinta Formation. In areas where fracturing was slight, hydraulic conductivity estimated from specific-capacity tests was related to matrix porosity (Hood, 1976, p. 34). Values of hydraulic conductivity for many specific-capacity tests are 100 times, or more, the values expected solely on the basis of matrix porosity. Where the estimated values are large, Hood (1976) concluded that fractures are the principal pathways of water movement to wells. Hydraulic conductivities as large as 600 ft/d have been estimated from specific-capacity tests. Hydraulic conductivities of fractured rock of 1 to 100 ft/d are typical. Hood (1976, pl. 3) presents a map showing the locations of specific-capacity tests in the Duchesne River-Uinta aquifer. The average hydraulic conductivity estimated from specific-capacity tests is 23.2 ft/d. This value is approximately one order of magnitude greater than would be expected from unfractured-core data.

A single aquifer test with observation wells has been reported for the Duchesne River-Uinta aquifer (Hood, 1976). The test was conducted near Roosevelt, Utah. Data from this test were interpreted using image-well theory and assuming the aquifer to be a homogeneous porous medium. By using the Theis method to match various parts of the data record, Hood (1976, p. 25) estimated the hydraulic conductivity to range between 0.3 and 1.3 ft/d. The test location probably is atypical of the entire aquifer because fractures were not believed to contribute very much water to the pumping well. The estimated hydraulic conductivity is an order of magnitude less than the average value obtained from specific-capacity tests.

Estimates of hydraulic conductivity obtained using laboratory and well tests reflect local-scale conditions that may not be applicable when estimating effective basin values. Efforts to contour the distribution of hydraulic conductivity manually (Hood, 1976), and geostatistically by variogram analysis (Glover, 1996), have been unsuccessful. Fracture sets, which may locally control flow, may not be extensive throughout the basin. Fracture data are not available in the Uinta Basin; therefore, evaluation of basin fracture interconnectivity is not possible. However, fracture patterns in the Uinta Basin may be similar qualitatively

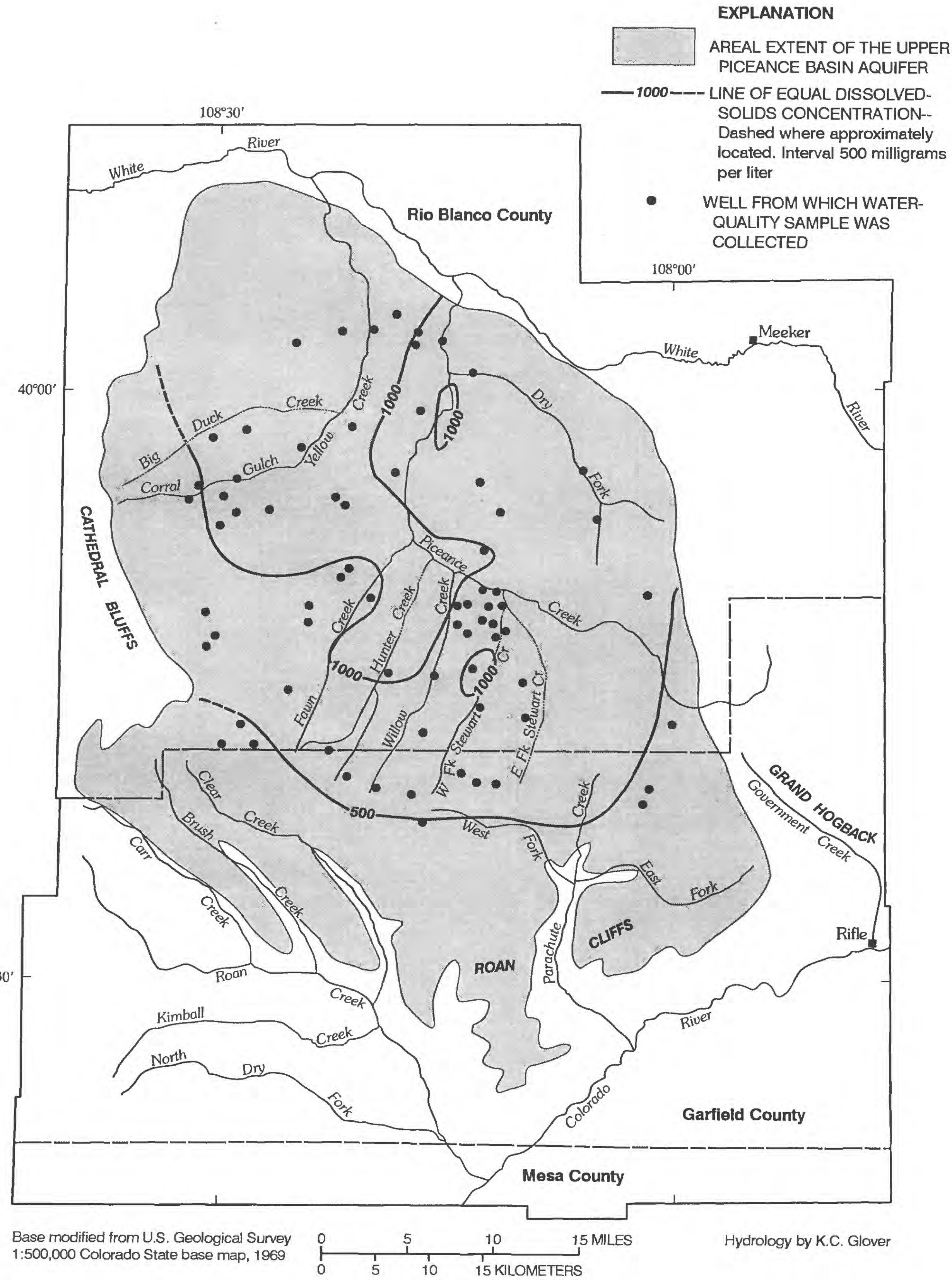
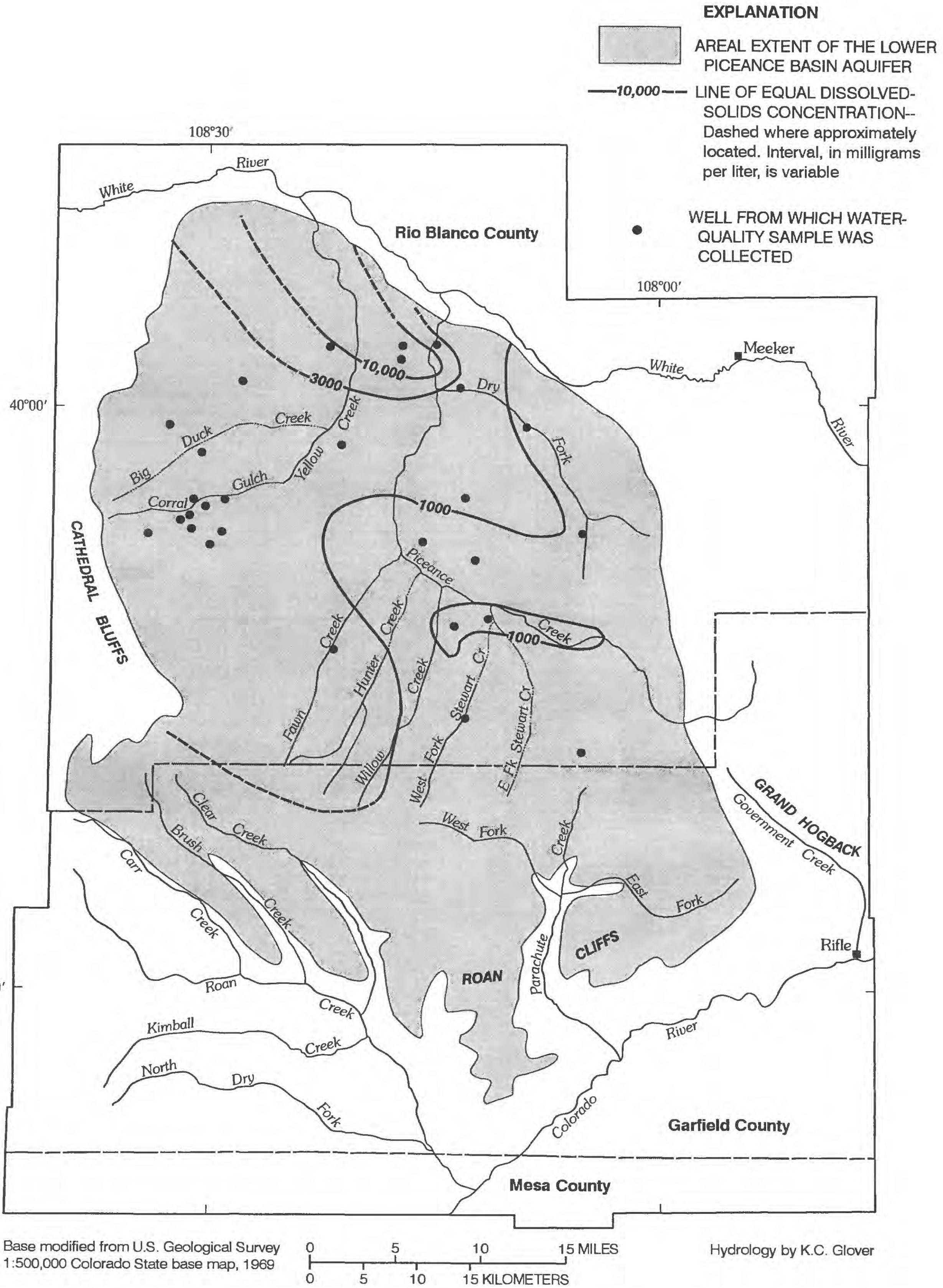


Figure 37. Dissolved-solids concentration in water of the upper Piceance Basin aquifer, Piceance Basin aquifer system.



Base modified from U.S. Geological Survey 1:500,000 Colorado State base map, 1969

Hydrology by K.C. Glover

Figure 38. Dissolved-solids concentration in water of the lower Piceance Basin aquifer, Piceance Basin aquifer system.

Table 4. *Summary of well-yield, spring-discharge, and hydraulic-conductivity data, Uinta basin aquifer system*

[gal/min, gallons per minute; ft/d, foot per day; <, less than; --, not available]

Geohydrologic unit (plate 1)	Well-yield and spring-discharge rates (gal/min)	Hydraulic conductivity	
		Aquifer and specific capacity tests (ft/d)	Model analyses (ft/d)
Duchesne River-Uinta aquifer	Typically 50 to 200; locally less than 5 or more than 500	median: 1.0 range: <0.03 to 600	[1]0.5 to 1.0
Parachute Creek confining unit	Generally less than 10	median: 0.007 range: 0.0002 to 0.11	--
Douglas Creek-Renegade aquifer	Generally less than 100	median: 0.09 range: 0.05 to 0.3	[2]0.05 to 0.1
Wasatch-Green River confining unit	Generally less than 50	median: 0.003 range: 0.0003 to 0.4	--

[1]From Glover, 1996
[2]From Holmes and Kimball, 1987

to those of the Piceance Basin. In the Piceance Basin, only a dominant subset of all fracture sets is present throughout the basin. Basin water movement probably is controlled by the dominant subset, while flow during well tests is enhanced by all fracture sets that are present at the test location.

Hydraulic-conductivity values estimated from specific-capacity tests are approximately an order of magnitude greater than estimated values obtained from model development. As indicated previously, the difference may be the result of differences in fracture interconnection at local (well test) and basin scales. In addition, the larger values obtained from specific-capacity tests may be due to tests generally being conducted in wells that penetrate only the upper, more fractured part of the aquifer.

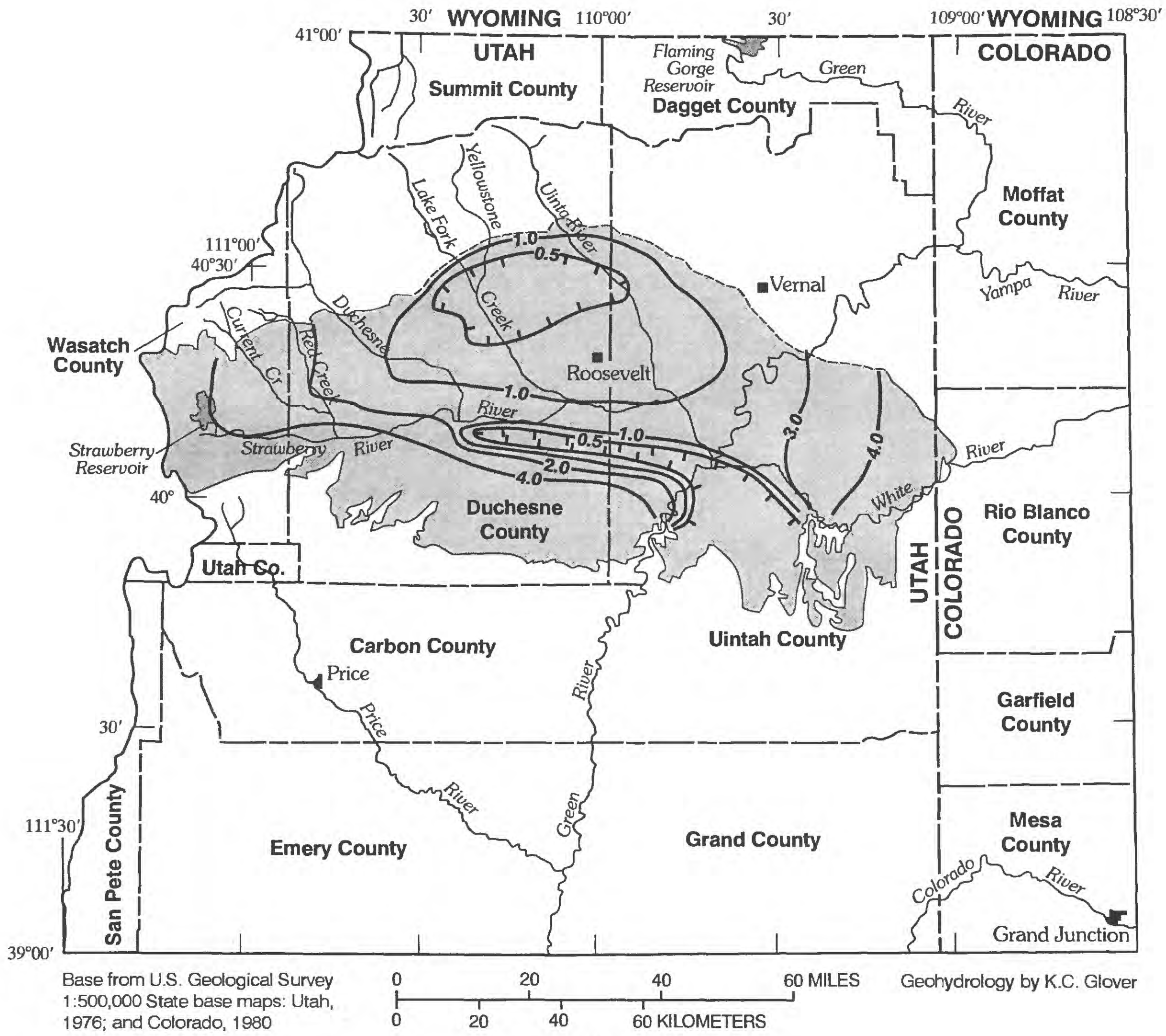
The generalized distribution of hydraulic conductivity (fig. 39) was compiled on the basis of the saturated thickness map for the Duchesne River-Uinta aquifer (fig. 12) and from transmissivity estimates obtained during development of a ground-water flow model for the Duchesne River-Uinta aquifer (Glover, 1996). The resulting hydraulic-conductivity estimates represent effective values for the saturated part of the aquifer. In areas where lithologic or fracture characteristics of the Duchesne River-Uinta aquifer vary with depth, hydraulic conductivity also probably varies.

Hydraulic conductivity of the Duchesne River-Uinta aquifer generally decreases from basin margins toward the depositional center of the basin. A large area with hydraulic conductivity less than 1 ft/d corresponds closely with a thick sequence of shale in the Uinta Formation. Fractures in this sequence apparently are ineffective in enhancing hydraulic conductivity. Although overlain by more permeable sandstone and shale of the Duchesne River Formation, the thick shale sequence effectively reduces the depth-averaged hydraulic conductivity.

Hydraulic-conductivity estimates in the southern part of the Duchesne River-Uinta aquifer are greater than values that would be expected solely on the basis of the percentage of sandstone and conglomerate. Average matrix hydraulic conductivity obtained from laboratory tests (Hood, 1976) is about 0.4 ft/d. Values greater than 2.0 ft/d (fig. 39) may indicate that hydraulic conductivity has been enhanced by fracturing.

A narrow band of small hydraulic conductivity in the south-central part of the Duchesne River-Uinta aquifer corresponds with a fault zone. Solely on the basis of geology, it might be expected that hydraulic conductivity would be enhanced parallel to the strike of the fault zone and hydraulic conductivity would be reduced normal to the strike where permeable strata could be offset. An effort to simulate this anisotropy with the model of the Duchesne River-Uinta aquifer was unsuccessful (Glover, 1996). Hydraulic-head gradient normal to the strike of the fault zone was sufficiently steep to justify reducing hydraulic-conductivity estimates. However, the hydraulic-head gradient along the strike was not sufficiently defined to conclude that hydraulic conductivity was enhanced parallel to the strike.

The area of small hydraulic conductivity along the southeastern margin of the Duchesne River-Uinta aquifer also corresponds with an area of faulting. Hydrologic data in this area are very limited; however, calibration of a ground-water flow model for the Duchesne River-Uinta aquifer was possible only when this area was assigned very small hydraulic conductivity (Glover, 1996). The relatively thin saturated thickness of the aquifer, combined with numerous faults, possibly has resulted in an area where local-scale flow dominates and basin flow is negligible. Additional detailed investigation would be needed to validate or reject this hypothesis.



EXPLANATION

AREAL EXTENT OF DUCHESNE RIVER-UINTA AQUIFER

—4.0— LINE OF EQUAL HYDRAULIC CONDUCTIVITY--Hachures indicate closed area of smaller conductivity. Interval, in feet per day, variable

Figure 39. Hydraulic conductivity of the Duchesne River-Uinta aquifer, Uinta Basin aquifer system.

Limited data are available to describe the hydraulic conductivity of the Douglas Creek-Renegade aquifer and associated confining units. Basin hydraulic-conductivity maps of these geohydrologic units generally would be based on fewer than five aquifer tests, which could present a misleading picture. As an alternative to maps, hydraulic-conductivity estimates are summarized in tabular form (table 4). The hydraulic conductivity of the Douglas Creek-Renegade aquifer is approximately one order of magnitude greater than the hydraulic conductivity of adjacent confining units. Based on a very small number of aquifer tests (Holmes, 1980) and specific-capacity tests (Hood, 1976), the hydraulic conductivity of the Douglas Creek-Renegade aquifer is approximately 0.09 ft/d, but the hydraulic conductivity of adjacent confining units is less than 0.01 ft/d. All three geohydrologic units are substantially less permeable than the Duchesne River-Uinta aquifer.

HYDRAULIC HEAD

The potentiometric-surface map of the Duchesne River-Uinta aquifer (fig. 40) indicates that ground water flows from upland areas adjacent to the Uinta Mountains and along the southwest margins of the basin toward the Strawberry, Duchesne, and Green Rivers.

The potentiometric-surface map of the Duchesne River-Uinta aquifer represents virtually steady-state conditions. Although the total volume of water withdrawn from wells in the aquifer is not large, prolonged pumping has caused local declines in the steady-state potentiometric surface of no greater than 100 ft (Hood and Fields, 1978). These declines are much smaller than the contour interval. Therefore, the potentiometric-surface map does not indicate the effects of pumping and represents steady-state conditions.

Water in the Duchesne River-Uinta aquifer generally is under artesian conditions. Water-table conditions exist in upland areas where the aquifer is dissected by streams. Otherwise, relatively impermeable beds of shale or limestone confine flow within underlying permeable strata. Wells that are open to the aquifer near discharge areas typically flow because the hydraulic head is above land surface.

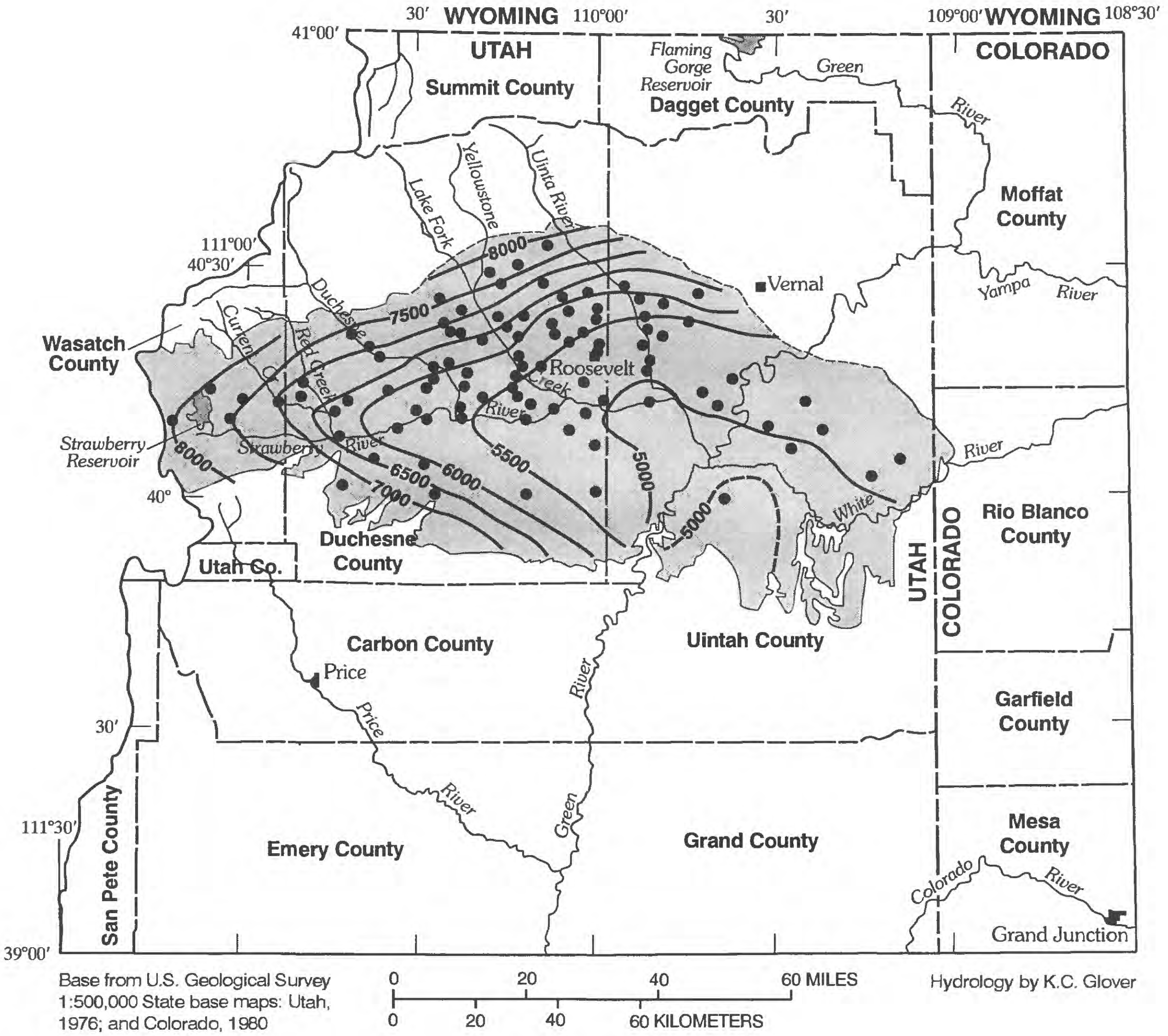
Water-table conditions exist in the Douglas Creek-Renegade aquifer where it crops out along the southern rim of the Uinta Basin, and artesian conditions exist in the central part of the basin. The scarcity of water-level data precluded compilation of a basin potentiometric-surface map. However, limited head data (Holmes and Kimball, 1987) indicate that water in the Douglas Creek-Renegade aquifer generally flows from recharge areas at high altitudes along the southeastern part of the Uinta Basin to discharge areas along the Green and White Rivers where the aquifer is near land surface.

GROUND-WATER RECHARGE AND DISCHARGE

An estimated 273 ft^3/s of water enters the Uinta Basin aquifer system as recharge, primarily in upland areas, and eventually discharges to perennial streams in the center of the Uinta Basin. Of this quantity, 272 ft^3/s enters the system as recharge to the Duchesne River-Uinta aquifer; 1.0 ft^3/s enters as recharge to the underlying Douglas Creek-Renegade aquifer. Additional water enters the system as recharge to confining units; however, the small hydraulic conductivity of confining units limits the recharge rate. For the purposes of basin study, recharge to confining units is considered negligible. Water also recharges and discharges along local flow paths, primarily in upland areas. Local recharge and discharge is not included in quantitative estimates given in this report.

Recharge to the Uinta Basin aquifer system is derived from precipitation and seepage losses from unlined canals and streams (Price and Miller, 1975; Hood and Fields, 1978). Hydraulic-head data indicate that recharge from precipitation occurs principally on the margins of the Uinta Basin, where precipitation is greater than at the center of the basin. Seepage losses have been measured along few of the numerous canals, but Hood and Fields (1978) estimate that as much as 10 percent of diverted streamflow infiltrates and becomes ground water.

Initial estimates of ground-water recharge were based on empirically derived relations that describe recharge as a percentage of average annual precipitation (Price and Miller, 1975; Hood and Fields, 1978). Areas of substantial recharge were subsequently supported by analysis of water-quality data (Naftz, 1996). The distribution of recharge in these areas was refined and calibrated with other hydrologic data during development of ground-water flow models (Holmes and Kimball, 1987; Glover, 1996). Initial estimates were given in detail in Price and Miller (1975), Hood and Fields (1978), and Glover (1996). Only the final estimates are given in this report.



EXPLANATION

AREAL EXTENT OF DUCHESNE RIVER-UINTA AQUIFER

—5000— — POTENTIOMETRIC CONTOUR--Shows altitude at which water level would have stood in tightly cased wells, 1985. Dashed where approximately located. Interval 500 feet. Datum is sea level

● WELL IN WHICH WATER LEVEL WAS MEASURED

Figure 40. Potentiometric surface (1985) of the Duchesne River-Uinta aquifer, Uinta Basin aquifer system (modified from Glover, 1996, p. 7).

Estimated recharge ranges from more than 7 in/yr along the northern and western edges of the basin to less than 0.1 in/yr in the center of the basin (fig. 41). In general, differences between initial and final estimates of recharge were small and probably were less than the error inherent in the method of Eakin and others (1951).

Recharge areas identified by empirical methods and ground-water modeling in the Duchesne River-Uinta aquifer were substantiated by analysis of water-quality data. Results of analyses of water samples from the Duchesne River-Uinta aquifer indicate positive log $(([Ca]+[Mg])/[Na]^2)$ values along the western parts of the Duchesne River-Uinta aquifer (fig. 42), where the potentiometric-surface map indicates recharge. Approximately 10 miles east of Strawberry Reservoir, the log $(([Ca]+[Mg])/[Na]^2)$ values become negative. The absence of positive log $(([Ca]+[Mg])/[Na]^2)$ values in the central parts of the Duchesne River-Uinta aquifer is in agreement with the potentiometric-surface map that identifies the central part of the Duchesne River-Uinta aquifer as a discharge area. The values of log $(([Ca]+[Mg])/[Na]^2)$ adjacent to the Uinta River deviate from the basin trends. This deviation could be indicative of additional processes affecting the exchangeable cation concentrations, other than cation exchange. Site-specific information is needed before further conclusions about water quality in this area can be made.

Discharge from the Uinta Basin aquifer system occurs mostly by seepage to streams, springs, wells, and transpiration by phreatophytes along streams. The principal streams receiving ground-water discharge are the Duchesne, Green, Lake Fork, Strawberry, Uinta, and White Rivers, and Currant and Red Creeks. Springs inventoried by Hood and others (1976) have discharge rates that range from 0.1 to 2,250 gal/min. However, only 12 springs flow at rates greater than 50 gal/min. Most springs in the Uinta Basin are located in upland areas and probably represent discharge points for local flow systems. No extensive inventory of ground-water use by wells has been done. Most water for irrigation, domestic, and industrial use is withdrawn at small rates. Discharge by wells probably has had no substantial effect on the basin distribution of hydraulic head. Ground-water discharge to phreatophytes is large, but difficult to estimate accurately. Price and Miller (1975) and Hood and Fields (1978) estimated transpiration by phreatophytes with a consumptive-use method. The method requires estimation of plant type, plant density, depth to water, and rate of water use. Estimation of these factors is highly subjective.

Components of total ground-water discharge that are important in understanding the basin flow system of the Uinta Basin include diffuse seepage to streams, spring discharge in hydraulically downgradient areas, and transpiration by phreatophytes along stream channels. Spring discharge in recharge areas is assumed to be indicative of local-scale flow. Well discharge is assumed to be negligible because no head declines due to pumping wells are indicated on potentiometric-surface maps. Transpiration by phreatophytes also is assumed to be negligible because discharge estimates were made at low flow during the winter when transpiration is minimal.

Ground-water discharge from the Uinta Basin aquifer system is estimated to be 273 ft^3/s (table 5). Estimates are based on gains or losses in average January streamflow for paired streamflow stations on several perennial streams (Glover, 1996). In these calculations, it was assumed that January streamflow best represents ground-water discharge to streams because the effects of evapotranspiration, ground-water storage in alluvium, surface runoff from ungaged drainages, and irrigation diversions are minimal. However, January streamflow records typically are affected by ice in the channels and may not be as accurate as records for other months. Table 5 includes all major streams of the basin and all streams that noticeably affect the distribution of hydraulic head. However, additional discharge may occur along other streams in the basin. This additional discharge, if present, probably is small. The estimates in table 5 also include effects of spring discharge in the vicinity of stream channels.

The attempt to quantify ground-water discharge to the Green and Uinta Rivers using streamflow records for paired stations on the two rivers was not successful. Differences in streamflow between paired stations were within the accuracy of measurements. Attempts to calculate streamflow gain or loss based on changes in chemical load also were unsuccessful. However, a digital model of the Duchesne River-Uinta aquifer was used successfully to estimate discharge to the Green and Uinta Rivers (Glover, 1996). Model calibration indicated that the Green River upstream from the White River probably is an area of recharge.

GROUND-WATER MOVEMENT

Horizontal movement of water in aquifers in Tertiary rocks of the Uinta Basin generally occurs from basin margins toward the major streams of the basin. Recharge occurs in upland areas peripheral to the Uinta Basin where annual precipitation exceeds 10 inches. Discharge occurs through leakage into alluvium along streams or through springs.

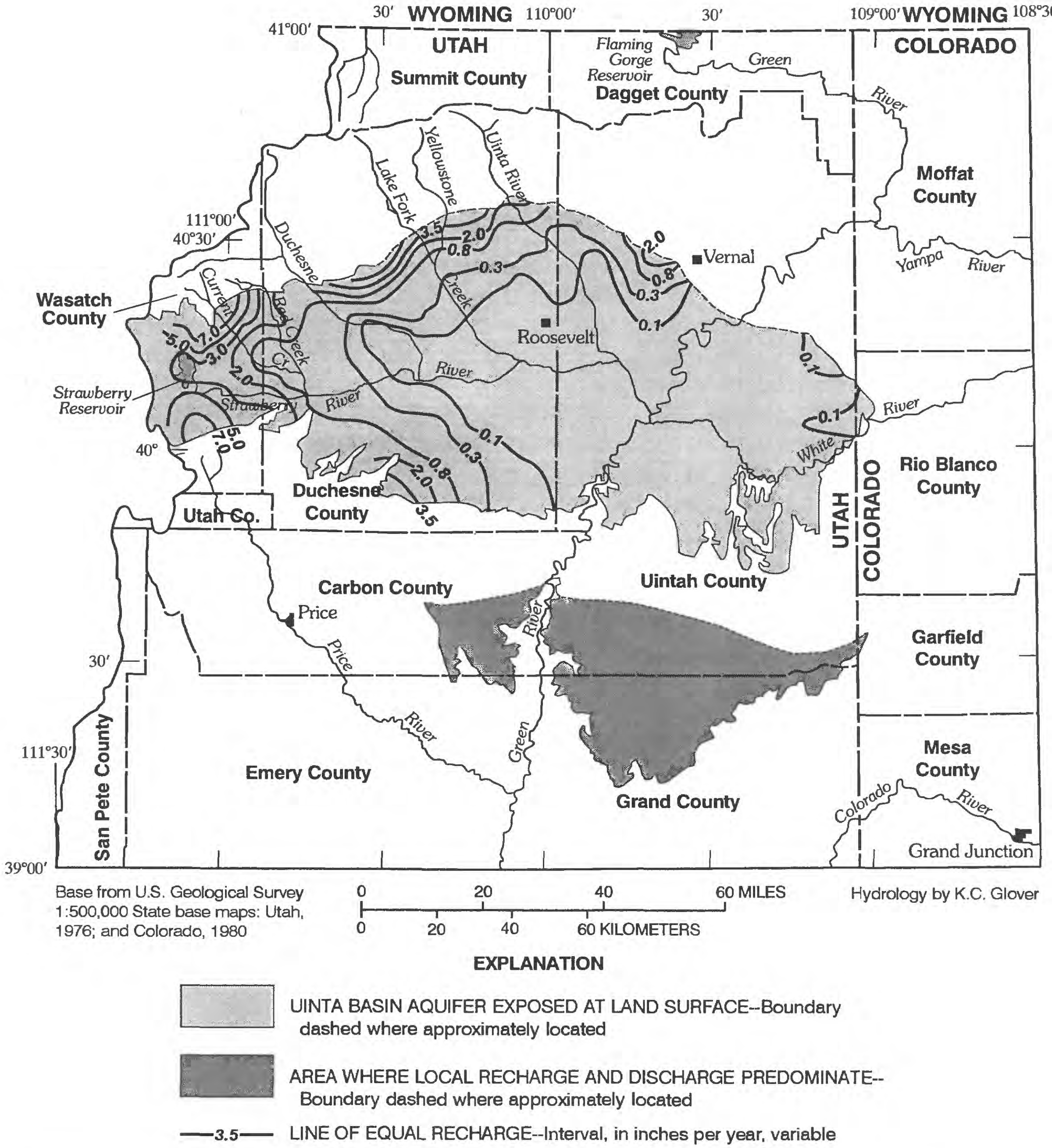


Figure 41. Estimated distribution of recharge to the Uinta Basin aquifer system (modified from Glover, 1996, p. 11; and Holmes and Kimball, 1986).

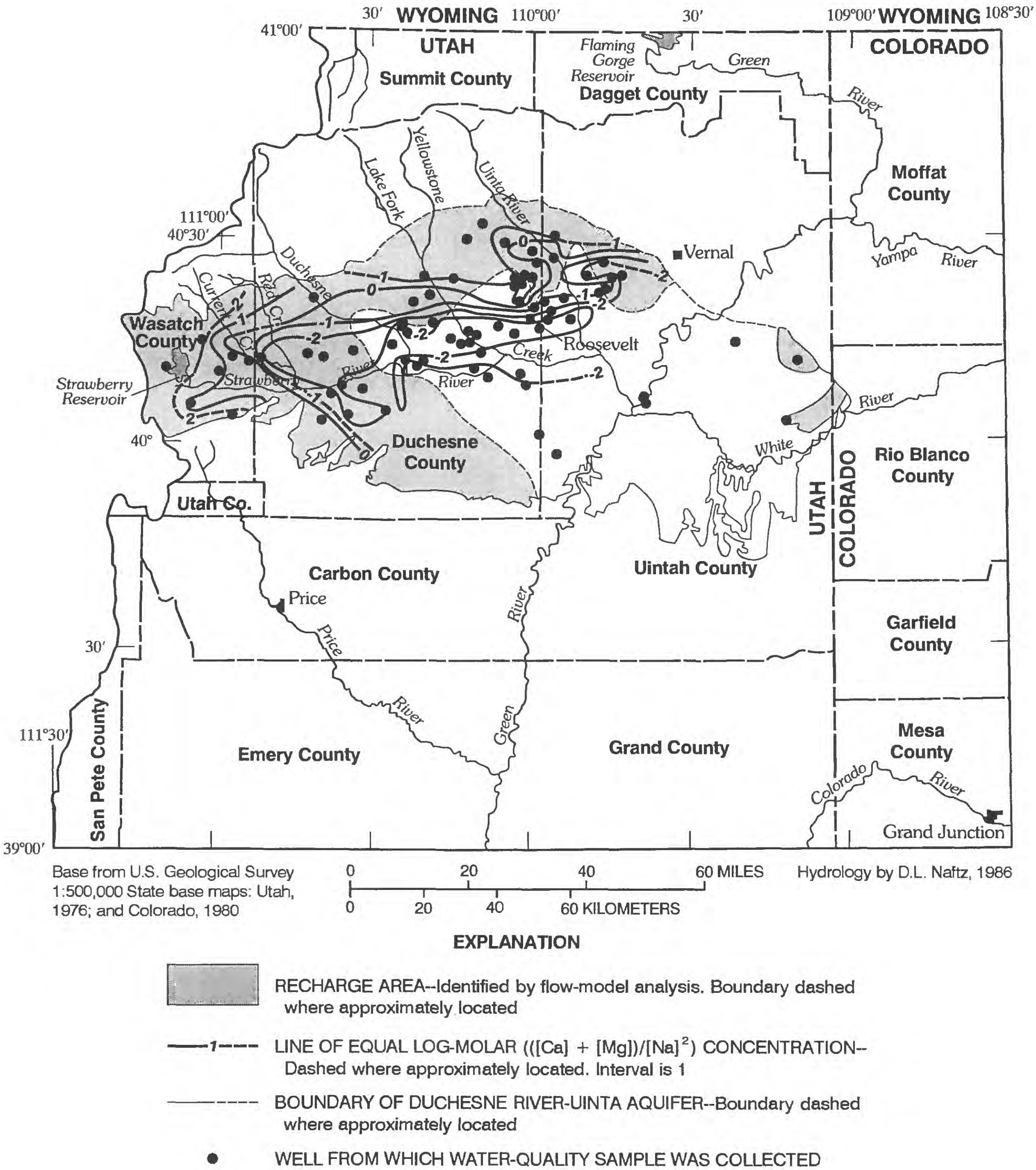


Figure 42. Log-molar ratio of calcium plus magnesium to squared-sodium concentration in water from the Duchesne River-Unita aquifer, Uinta Basin aquifer system (modified from Naftz, 1996, p. 15).

Table 5. *Estimated ground-water recharge from or discharge to selected streams, Uinta Basin aquifer system*

[+, indicates ground-water recharge; -, indicates ground-water discharge]

Stream	Ground-water recharge or discharge (cubic feet per second)
White River	-11
Duchesne River, Uinta River to Lake Fork River	-54
Duchesne River, Lake Fork River to Strawberry River	-76
Duchesne River, Strawberry River to Mesozoic contact including Rock Creek	-30
Lake Creek	-23
Current and Red Creeks	-10
Strawberry River, Starvation Reservoir to Red Creek	-7
Strawberry River, Red Creek to Strawberry Reservoir	-15
Green River, upstream from White River confluence	+9
Green River, downstream from White River confluence[1]	-2
Uinta and White Rocks Rivers[1]	-45

[1]Estimates based on development of ground-water models (Holmes and Kimball, 1987; Glover, 1996)

Rates of vertical leakage between the Duchesne River-Uinta and Douglas Creek-Renegade aquifers are small. Data to test this conclusion are limited, particularly in the Douglas Creek-Renegade aquifer. However, ground-water flow models have been developed for both aquifers on the assumption that no vertical leakage occurs. Both models adequately simulate existing geohydrologic data, indicating that existing data are compatible with an assumption of negligible vertical leakage.

Rates of ground-water movement have been estimated during development of basin flow models (Holmes and Kimball, 1987; Glover, 1996). The resulting ground-water budget (table 6) represents steady-state conditions. Model-derived estimates of discharge equal to 273 ft^3/s (for the aquifer system) are similar to estimates obtained from streamflow loss-and-gain studies and therefore are likely to be reliable. Because the system is under steady-state conditions, estimates of recharge also are likely to be reliable. Rates of movement are much greater in the Duchesne River-Uinta aquifer than in the Douglas Creek-Renegade aquifer (table 6).

Table 6. *Estimated ground-water budget, Uinta Basin aquifer system*

Duchesne River-Uinta Aquifer	
Recharge, in cubic feet per second:	
Precipitation	263
Losing streams	9
Total	272
Discharge, in cubic feet per second:	
Total	272
Douglas Creek-Renegade Aquifer	
Recharge, in cubic feet per second:	
Total	1.4
Discharge, in cubic feet per second:	
Total	1.4

QUALITY OF WATER

Water in the Duchesne River-Uinta aquifer gains dissolved solids and exhibits changes in major-ion chemistry as it moves along basin flow paths from upland recharge areas in the western and northern parts of the aquifer to discharge areas in the central parts of the aquifer. The dissolved-solids concentration increases from about 500 to 3,000 mg/L (fig. 43). Water in the recharge area is a calcium bicarbonate or magnesium bicarbonate type. Water in the downgradient parts of the basin is a sodium bicarbonate or sodium sulfate type.

Water in the Douglas Creek-Renegade aquifer also gains dissolved solids and exhibits changes in major-ion chemistry as it moves north along flow paths from recharge areas in the south (Holmes and Kimball, 1987). The dissolved-solids concentration increases from an average of 785 mg/L in the southern part of the aquifer to an average of 1,450 mg/L in the northern part of the aquifer. Water in the recharge area is diverse in type (Kimball, 1981, p. 9). Water in the northern downgradient areas is sodium bicarbonate or sodium carbonate type.

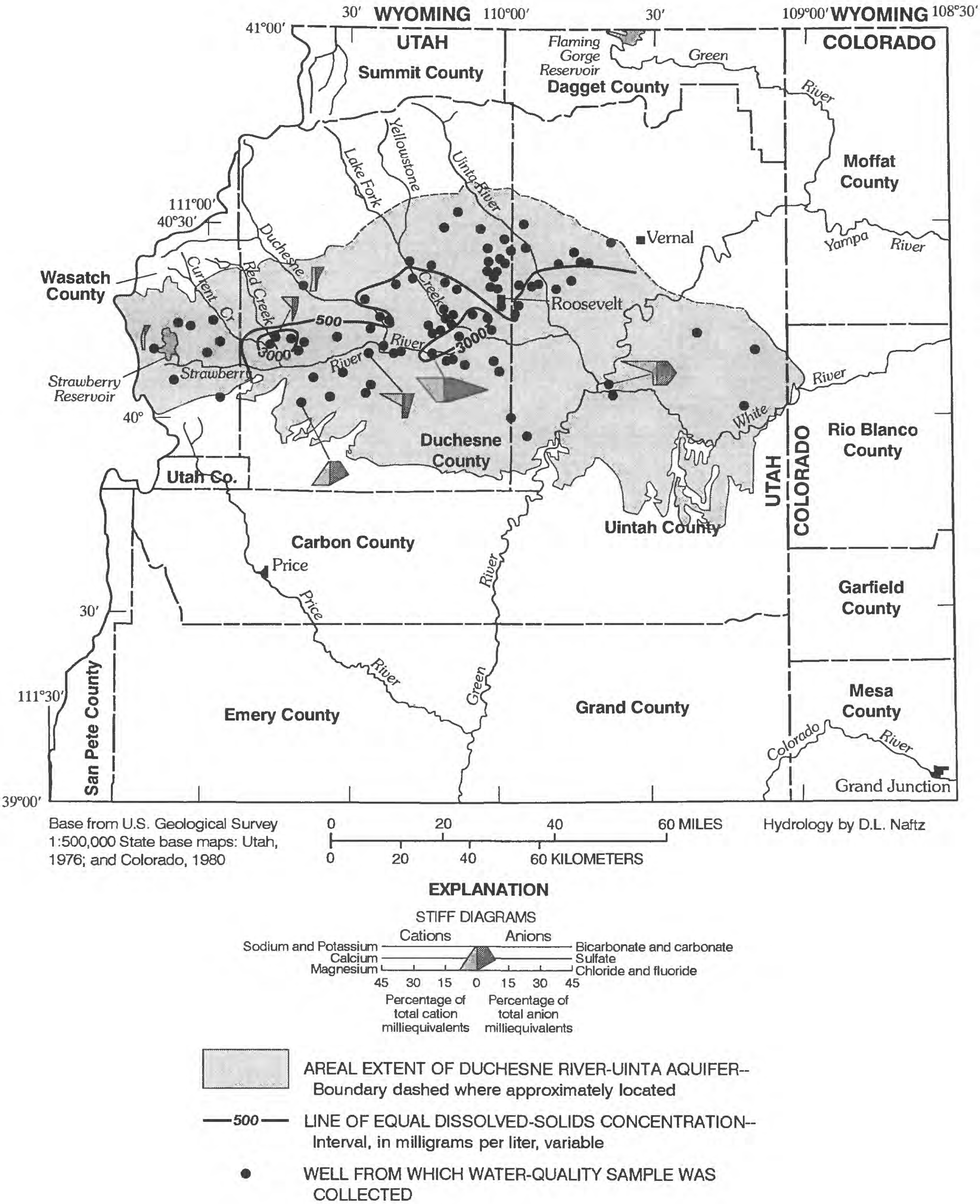


Figure 43. Dissolved-solids concentration and Stiff diagrams for the Duchesne River-Uinta aquifer, Uinta Basin aquifer system (modified from Naftz, 1996, p. 12-13).

GEOHYDROLOGY OF THE GREEN RIVER BASIN AQUIFER SYSTEM

The hydrologic system in Tertiary rocks of the Green River Basin consists of the Bridger, Laney, New Fork, and Wasatch-Fort Union aquifers and the Wilkins Peak and Tipton confining units. Stratigraphic and lithologic descriptions of these aquifers and confining units were given previously in this report. The Bridger, Laney, New Fork, and Wasatch-Fort Union aquifers are separated geographically or stratigraphically by confining units. The Wasatch and Fort Union zones of the Wasatch-Fort Union aquifer are treated as separate parts of a single aquifer because although their hydrologic properties generally are similar, the two units seldom are similar at the same geographic location. No confining unit separates the Wasatch-Fort Union aquifer from the underlying Mesaverde aquifer. The stratigraphic relation of aquifers and confining units is shown on plate 1.

HYDRAULIC CONDUCTIVITY

Estimates of hydraulic conductivity are available for most aquifers and confining units of the Green River Basin. Estimates were made using data from aquifer, specific-capacity, and drill-stem tests; the estimates typically range over several orders of magnitude within each aquifer. Point values of hydraulic conductivity do not correlate spatially and may not represent basin values that control basin ground-water movement. The inability to correlate and contour point estimates may reflect the lenticular nature of sandstone beds in the Wasatch-Fort Union aquifer and the unpredictable pattern of fractures and solution channels within the Laney and Bridger aquifers. Virtually no estimates of vertical hydraulic conductivity are available. Data from aquifer tests are summarized in table 7.

Because point estimates of aquifer properties may not indicate effective basin values, hydraulic conductivity was estimated during calibration of a basin ground-water flow model (Martin, 1996). The model consists of five layers, each subdivided into subareas. Calibration criteria for the model were to acceptably match simulated measured aquifer-system discharge and hydraulic head to measured values. Hydraulic properties were simulated; single values of horizontal and vertical hydraulic conductivity were used to represent subareas that consisted of a number of conterminous finite-difference nodes. Subarea boundaries generally were defined on the basis of lithofacies. Maps showing subarea boundaries and a table listing corresponding estimates of vertical and horizontal hydraulic conductivity are given by Martin (1996, p. 38-39) for each aquifer and confining unit within the Green River Basin aquifer system. A summary of the results of modeling is given in the following paragraphs and in table 1. Estimates obtained by model analysis generally are within the range of estimates obtained by aquifer-test analysis. However, local variations in sandstone content or fracture interconnection can result in hydraulic-conductivity estimates from aquifer-test data that are an order of magnitude different from estimates obtained from the basin ground-water flow model.

Values of hydraulic conductivity from 24 aquifer tests using wells completed in the Bridger aquifer range from 0.03 to 420 ft/d. The median value of these tests is 11 ft/d. Hydraulic conductivity for the Bridger aquifer was not well defined by flow-model analysis. Simulated hydraulic conductivity can be varied by as much as an order of magnitude without significantly affecting the comparison between calculated and measured hydraulic head. Varying hydraulic conductivity within an order of magnitude also has little effect on calculated discharge: values for basin hydraulic conductivity obtained from flow-model analysis range from 0.09 to 0.9 ft/d.

Values of hydraulic conductivity from 8 aquifer tests using wells completed in the Laney aquifer range from 2 to 1,400 ft/d. The median value of these aquifer tests was 13 ft/d. Basin model analysis provided estimates of effective values of hydraulic conductivity in the Laney aquifer. Vertical and horizontal hydraulic conductivity of the Laney aquifer generally are largest in areas adjacent to the Big Sandy River where aquifer-test data also indicate large hydraulic conductivity. Both vertical and horizontal hydraulic conductivity were estimated to be 17.3 ft/d in this area. Fractures and solution channels are believed to be the cause for enhanced vertical and horizontal hydraulic conductivity. The estimated horizontal hydraulic conductivity is small (0.04 ft/d) in areas where the aquifer is buried by the Bridger aquifer. The small hydraulic conductivity of the Laney, where buried, limits vertical movement of water between the Bridger and deeper aquifers. Varying simulated hydraulic conductivity by a factor of 4 or greater significantly changes the calculated distribution of aquifer-system discharge in the flow model.

Table 7. *Hydrologic characteristics and measured hydraulic conductivity, Green River Basin aquifer system (modified from Martin, 1996, p. 9)*

[ft/d, foot per day; --, no information available]

Geohydrologic unit	Hydrologic characteristics	Number of tests	Hydraulic conductivity Range (ft/d)	Hydraulic conductivity Median (ft/d)
Bridger aquifer	Hydraulic conductivity generally is small. Generally yields less than 50 gallons per minute to wells. Locally, however, large yields are possible from fractures and sandstone strata.	24	0.03 to 420	11
Laney aquifer	Hydraulic conductivity is moderate to large where fractures are present and where sandstone dominates. Well yields are greatest near the Big Sandy River. Elsewhere, hydraulic conductivity and well yields are small.	8	2 to 1,400	13
Wilkins Peak confining unit	Very small hydraulic conductivity. Generally yields less than 30 gallons per minute of briney water to wells.	--	--	--
New Fork aquifer	Hydraulic conductivity is moderate. Large well yields are possible where several hundred feet of saturated thickness is penetrated. Source of water for numerous flowing wells near the Big Sandy River.	2	.20 to 2	--
Tipton confining unit	Hydraulic conductivity is small. Hydraulic conductivity probably is large in the east-central part of the Green River Basin, where sandstone units increase in number, and differentiation from the New Fork Tongue of the Wasatch Formation is difficult.	--	[1]0 to 0.67	--
Wasatch zone of the Wasatch-Fort Union aquifer	Hydraulic conductivity varies over several orders of magnitude and reflects wide differences in the lithologic character of the formation. Hydraulic conductivity generally is moderate to large. Vertical variations in hydraulic head occur within the Wasatch zone in some areas.	186	.03 to 2,100	8.7
Fort Union zone of the Wasatch-Fort Union aquifer	Hydraulic conductivity is large in outcrop areas. Where deeply buried or where shale is present, hydraulic conductivity is small.	61	.02 to 1,100	40

[1]Estimates from a flow model developed by Glover (1986, p. 17).

No aquifer tests were available for wells completed in the Wilkins Peak Member of the Green River Formation. Bedded trona deposits within the Wilkins Peak confining unit restrict ground-water flow. As a result, the vertical hydraulic conductivity of the confining unit was estimated by flow-model analysis to be 0.00001 ft/d. This value probably is a maximum estimate and could be decreased several orders of magnitude without significantly altering either calculated hydraulic-head distributions or calculated water budgets.

Hydraulic-conductivity values were available from two aquifer tests using wells completed in the New Fork aquifer. Reported values from these aquifer tests were 0.2 and 2 ft/d. Hydraulic conductivity of the New Fork aquifer was estimated from flow-model analysis to be 6.5 ft/d.

Field measurements of hydraulic conductivity were not available for the Tipton confining unit. The Tipton confining unit is composed of shale and marlstone having small hydraulic conductivity. Vertical hydraulic conductivity of the Tipton confining unit was estimated to be 0.00001 ft/d by flow-model analysis.

Hydraulic-conductivity estimates were available from 186 aquifer tests using wells completed in the Wasatch zone of the Wasatch-Fort Union aquifer. Reported values range from 0.03 to 2,100 ft/d. The median value was 8.7 ft/d. Hydraulic conductivity of the Wasatch zone, as estimated from flow-model analysis, ranges from 0.04 ft/d in the south-central part of the basin where the Wasatch zone is buried by confining units, to 6.5 ft/d in the northern part of the basin where the Wasatch zone is at land surface. The variation in hydraulic conductivity also is related to depositional environment and proximity to the source of sedimentary material. Coarse-grained material tends to be more common near the basin margins where high-energy depositional environments prevailed, and fine-grained material is common in the interior part of the basin where low-energy depositional environments prevailed.

Hydraulic-conductivity values determined from 61 aquifer tests using wells completed in the Fort Union zone of the Wasatch-Fort Union aquifer range from 0.02 to 1,100 ft/d with a median value of 40 ft/d. Hydraulic conductivity of the Fort Union zone is very small (0.00001 ft/d) in the northern part of the Green River Basin where the Fort Union zone is buried by as much as 7,000 ft of overlying sediments. The hydraulic conductivity is greater where the thickness of overlying sediments is less. Maximum values of hydraulic conductivity obtained from flow-model analysis are 0.3 ft/d in the southeast part of the basin. Reasons for the large difference between hydraulic-conductivity estimates obtained from aquifer tests and estimates obtained from flow-model analysis are unknown. Sandstone lenses and other permeable rocks that locally enhance hydraulic conductivity may not be interconnected at basin scale.

Corrected carbon-14 ages of ground water were used by Chafin and Kimball (1992) to estimate the hydraulic conductivity at three sites in the Wasatch zone of the Wasatch-Fort Union aquifer. The carbon-14 ages and estimates for bulk porosity and hydraulic gradient were used to calculate hydraulic-conductivity estimates that ranged from 9 to 30 ft/d at the three sites. These values compare well with simulated flow-model values of 6.5 ft/d reported by Martin (1996) for the Wasatch zone in the northern part of the basin.

HYDRAULIC HEAD

Adequate data are available to map potentiometric surfaces of the Bridger and Laney aquifers and the Wasatch and Fort Union zones of the Wasatch-Fort Union aquifer. No potentiometric surface has been mapped for the New Fork aquifer. However, flowing wells indicate that artesian conditions are common in the New Fork aquifer.

The potentiometric surface of the Bridger aquifer in 1985 (fig. 44) was compiled using water-level measurements of wells completed in the Bridger aquifer. However, wells in the southwestern part of the aquifer typically are completed partly in the overlying alluvium or glacial deposits; these wells were used when better control points were not available. Hydraulic-head data used to map the potentiometric surface of the Bridger aquifer are highly variable over short distances, which probably indicates that local recharge and discharge is occurring. Local flow systems in discontinuous aquifers in Miocene and Oligocene sediments, including the Browns Park Formation and Bishop Conglomerate in the southwestern part of the basin, probably affect water levels in the Bridger aquifer.

The potentiometric surface of the Laney aquifer in 1985 (fig. 45) was compiled primarily using water-level measurements in wells. Because little data were available where the Laney aquifer is buried, the potentiometric surface was compiled only in areas where the aquifer is at or near land surface.

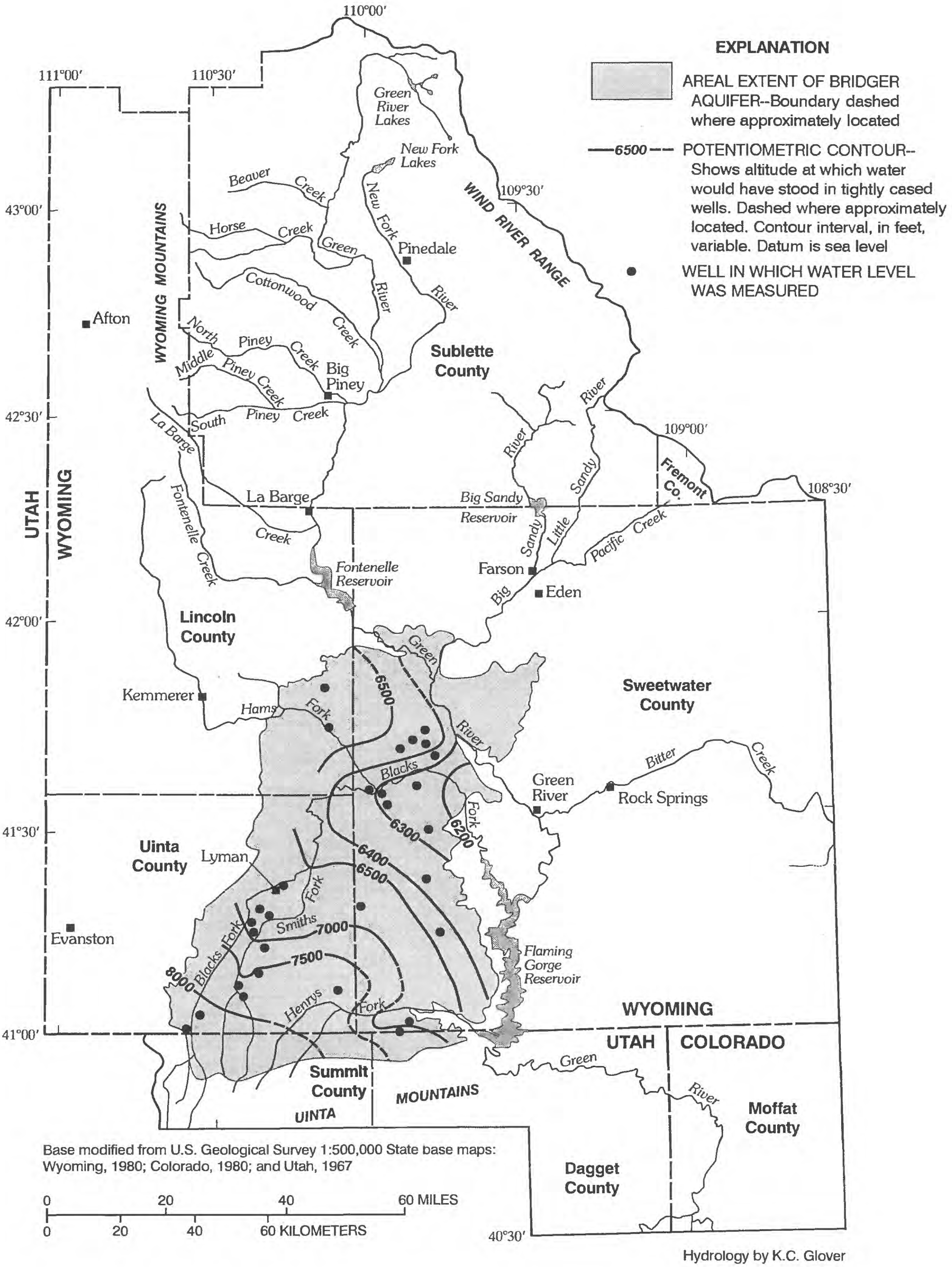


Figure 44. Potentiometric surface (1985) of the Bridger aquifer, Green River Basin aquifer system (from Martin, 1996, p. 12).

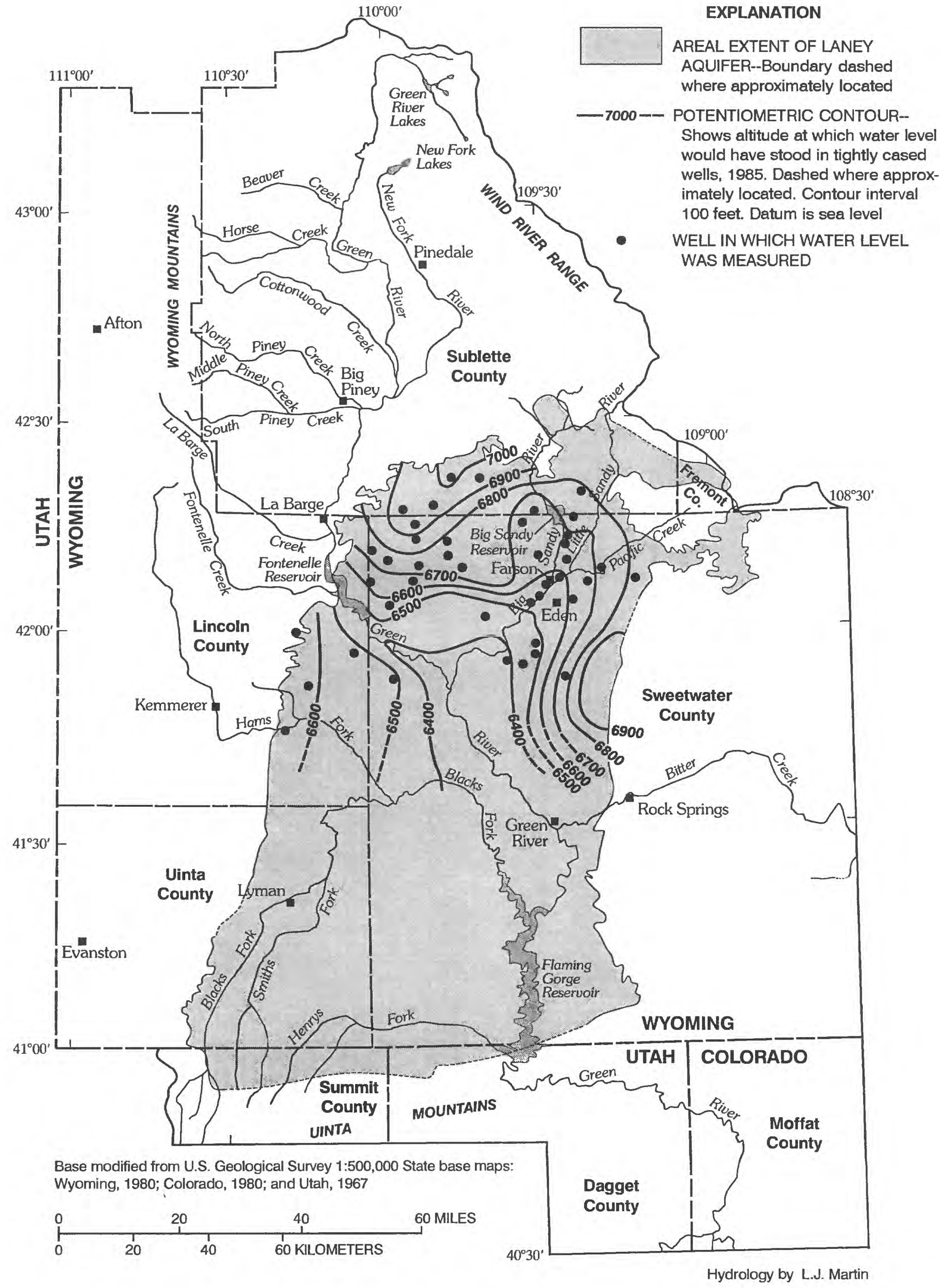


Figure 45. Potentiometric surface (1985) of the Laney aquifer, Green River Basin aquifer system (from Martin, 1996, p. 14).

The potentiometric-surface map of the Wasatch zone of the Wasatch-Fort Union aquifer in 1985 (fig. 46) was constructed primarily using water-level measurements in wells. Water levels in the northern part of the Green River Basin generally were measured in shallow wells (less than 1,000 ft deep) and represent water-table conditions. These data were supplemented by measurements of widely scattered wells where the aquifer is confined below the Tipton confining unit. In areas where the Wasatch zone is thick, there can be substantial differences between the average Wasatch zone hydraulic head and the hydraulic head of the top 1,000 ft of the Wasatch zone.

The 1985 potentiometric-surface map for the Fort Union zone of the Wasatch-Fort Union aquifer (fig. 47) was constructed primarily using pressure measurements from drill-stem tests of wells completed in deeply buried parts of the zone. The accuracy of the potentiometric surface for the Fort Union zone is much poorer than that of the Wasatch zone. Drill-stem tests generally are conducted in Tertiary sediments at depths greater than 2,500 ft below land surface in the Green River Basin. These tests usually are recorded as occurring within the Fort Union Formation. However, such a designation may be unreliable because geologists have not consistently identified the top of the Fort Union using borehole cuttings and geophysical logs.

GROUND-WATER RECHARGE AND DISCHARGE

Although no detailed study of ground-water recharge has been undertaken in the Green River Basin, interpretation of basin water-quality data and basin ground-water flow modeling have identified areas and approximate magnitudes of recharge. Most recharge in the Green River Basin occurs at land-surface altitudes above 7,000 ft. Precipitation at these altitudes generally exceeds 10 in/yr. In general, recharge areas are located along the margins of the basin near the Overthrust Belt, Wind River Uplift, and Uinta Uplift (fig. 2). Surface-water irrigation seepage in the vicinity of Farson, Wyo., also provides recharge to the ground-water system.

The magnitude and distribution of ground-water recharge from precipitation in the Green River Basin was estimated during development of a basin ground-water flow model. Recharge was estimated using linear relations between recharge and average annual precipitation. Two relations, one for the northern part of the basin and another for the southern part, were developed using the assumption of steady-state flow to ensure that aquifer-system recharge equaled aquifer-system discharge. Estimates of discharge are considered reasonably accurate. Recharge from precipitation in the northern part of the basin was estimated to be 117 ft^3/s, and recharge from precipitation in the southern part of the basin was estimated to be 21 ft^3/s. Infiltration recharge to the Green River Basin is estimated to range from 0 to 2.6 in/yr (fig. 48).

In addition to recharge associated with precipitation along basin margins, an area of recharge near Farson, Wyo., is delineated in figure 48. The source of recharge probably is irrigation seepage; annual precipitation in this area is less than 10 in. Attempts to calibrate the basin flow model using measured ground-water discharge along the Big Sandy River proved unsuccessful unless recharge to the Laney aquifer from irrigation also was simulated. Recharge from excess surface-water irrigation in the Farson area is estimated to be 18 ft^3/s.

Concentrations of calcium and magnesium relative to sodium in water from the Bridger aquifer (fig. 49) and the Wasatch zone of the Wasatch-Fort Union aquifer (fig. 50) indicate recharge areas similar to those indicated by hydraulic-gradient data. In the Bridger aquifer, positive log $(([Ca]+[Mg])/[Na]^2)$ values indicate that recharge occurs in the southern part of the Green River Basin adjacent to the Uinta Uplift. The progression of log $(([Ca]+[Mg])/[Na]^2)$ values from positive to negative generally follows the trends indicated by hydraulic-gradient data (figs. 44 and 46) and indicates a flow path from south to north. The localized area of positive log $([Ca]+[Mg])/[Na]^2)$ values in the Bridger aquifer east of Evanston, Wyo., and in the vicinity of Blacks Fork could indicate localized recharge.

Trends in positive values of log $(([Ca]+[Mg])/[Na]^2)$ in water from the Wasatch zone of the Wasatch-Fort Union aquifer (fig. 50) indicate water is recharged in northern, eastern, and western parts of the Green River Basin. Changes in values of log $(([Ca]+[Mg])/[Na]^2)$ occur in this zone of the aquifer along the hydraulic gradient indicating that ground water moves toward the central and southern parts of the basin.

Streams and associated alluvial deposits are the major points of basin ground-water discharge for aquifers in the Green River Basin. Base-flow statistics for streamflow-gaging stations were compared to quantify gains and losses between stations. Streamflow gain or loss calculated for periods of record when effects of evaporation, bank storage, and diversion were minimal, are listed in table 8. With the exception of Blacks Fork, ground-water discharge calculated with a basin flow model compares well with measured streamflow gains or losses. The gain-loss study indicates loss of 9 ft^3/s from the Blacks, Smiths, and Hams Forks. Model results indicate movement out of the aquifer of

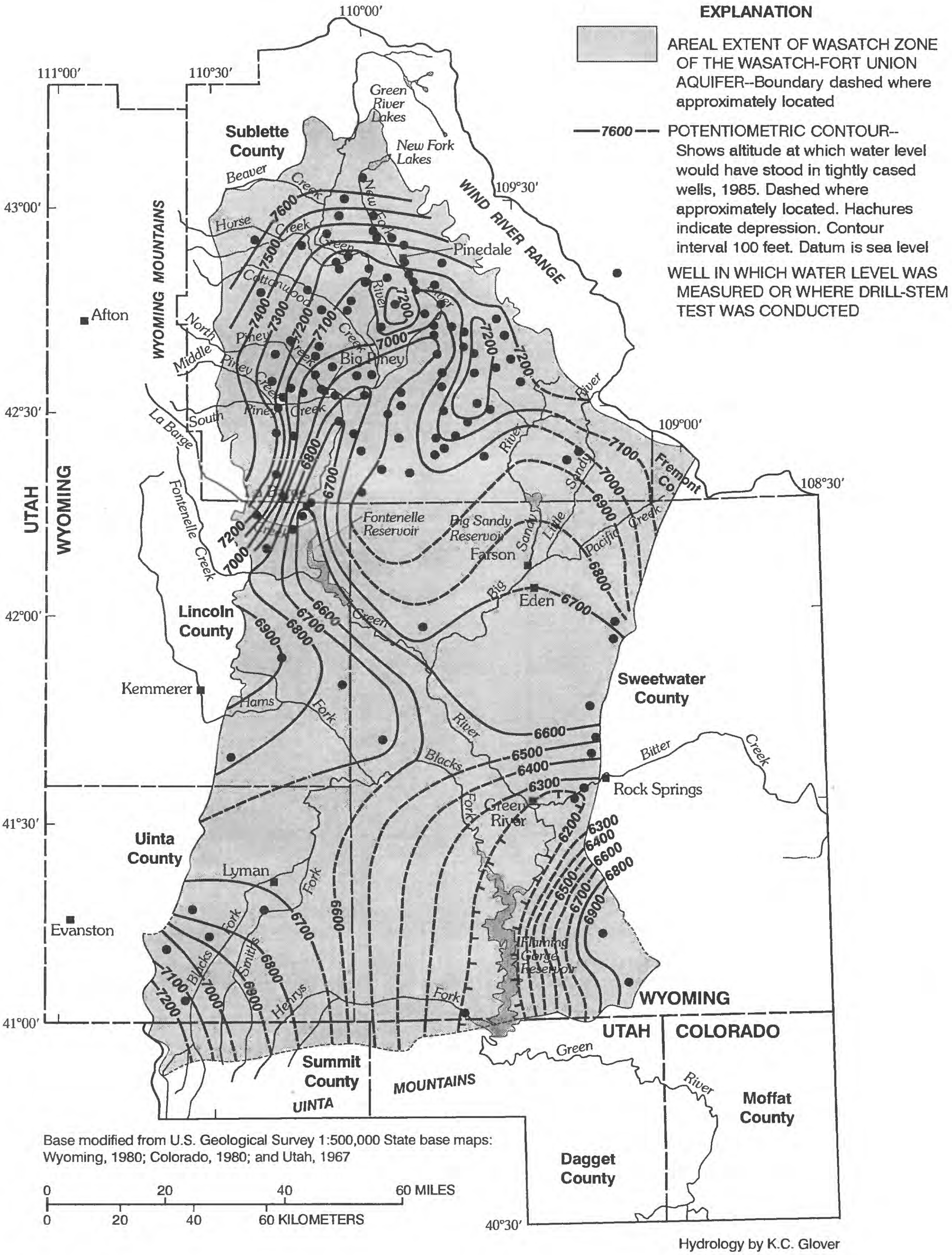


Figure 46. Potentiometric surface (1985) of the Wasatch zone of the Wasatch-Fort Union aquifer, Green River Basin aquifer system (from Martin, 1996, p. 20).

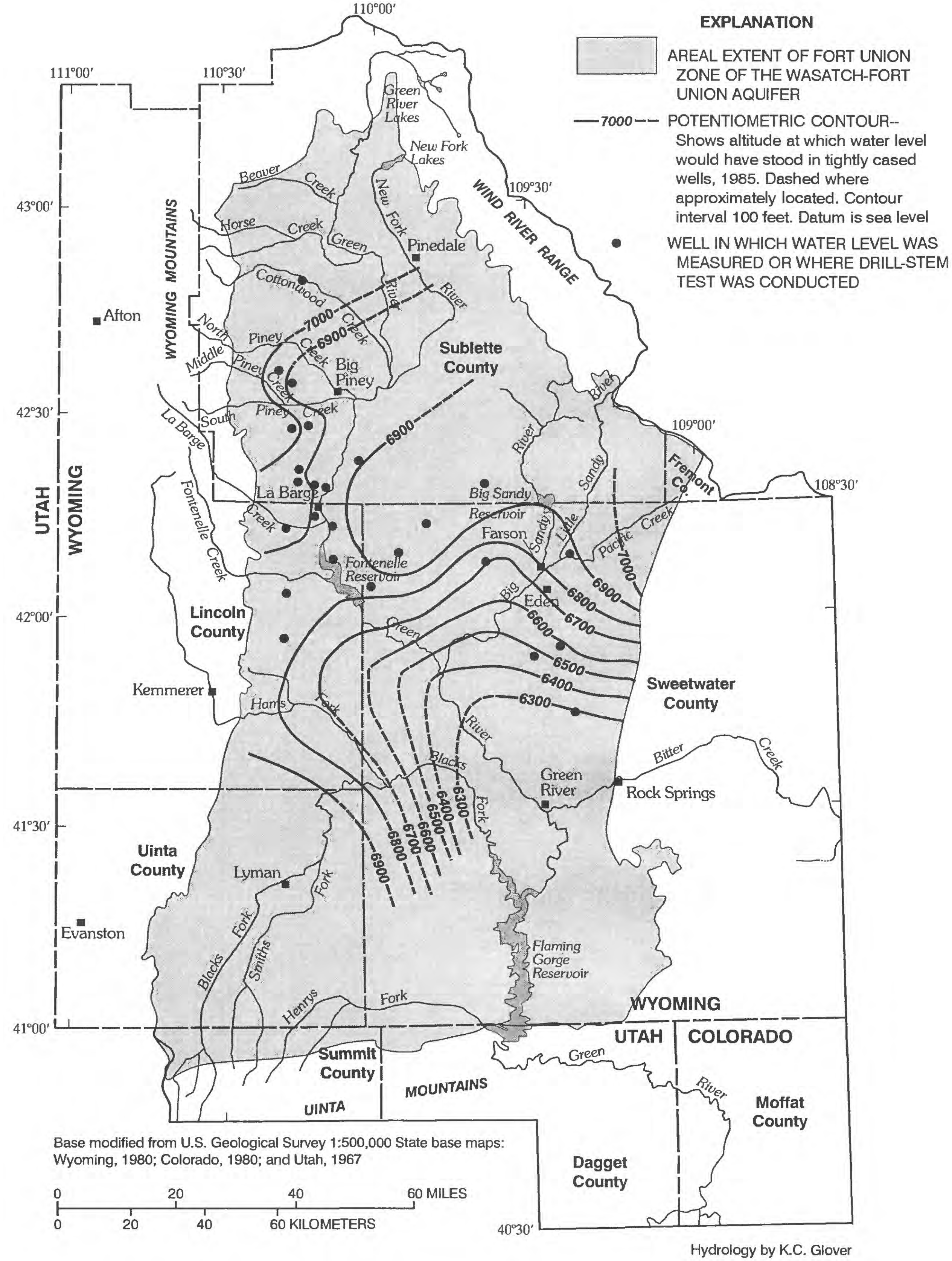


Figure 47. Potentiometric surface (1985) of the Fort Union zone of the Wasatch-Fort Union aquifer, Green River Basin aquifer system (from Martin, 1996, p. 23).

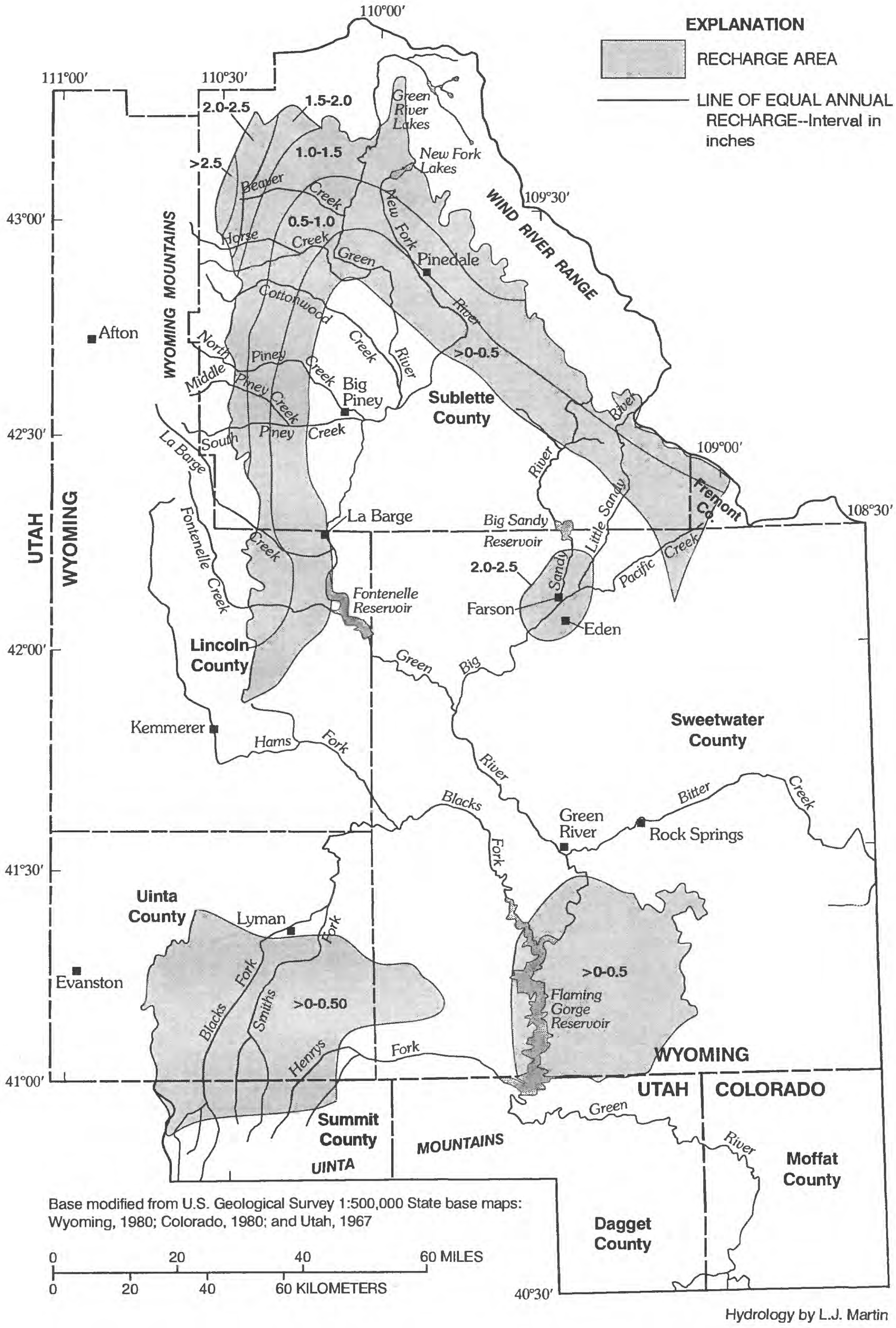


Figure 48. Distribution of estimated recharge to the Green River Basin aquifer system (from Martin, 1996, p. 26).

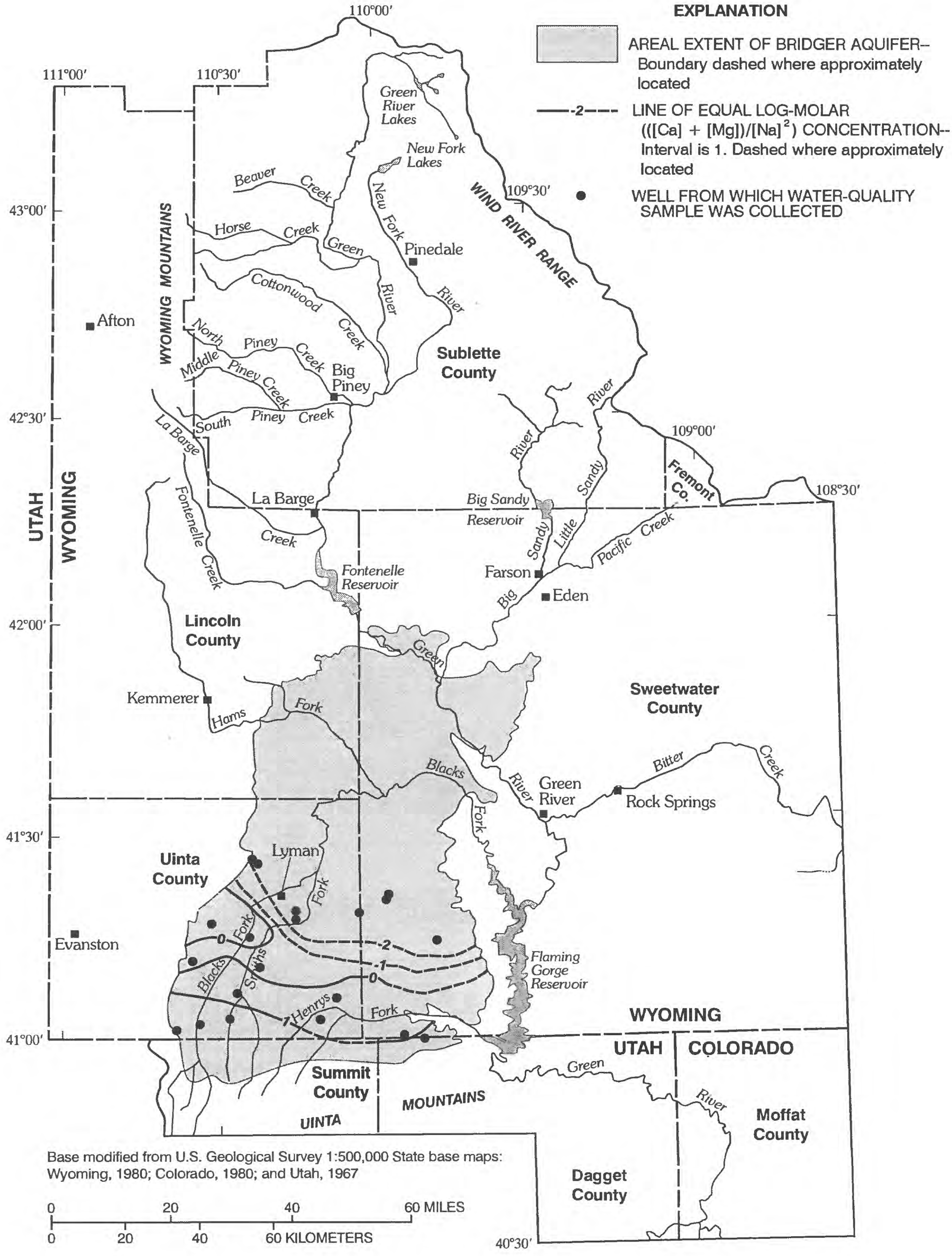


Figure 49. Log-molar ratio of calcium plus magnesium to squared-sodium concentration in water from the Bridger aquifer, Green River Basin aquifer system (modified from Naftz, 1996, p. 15).

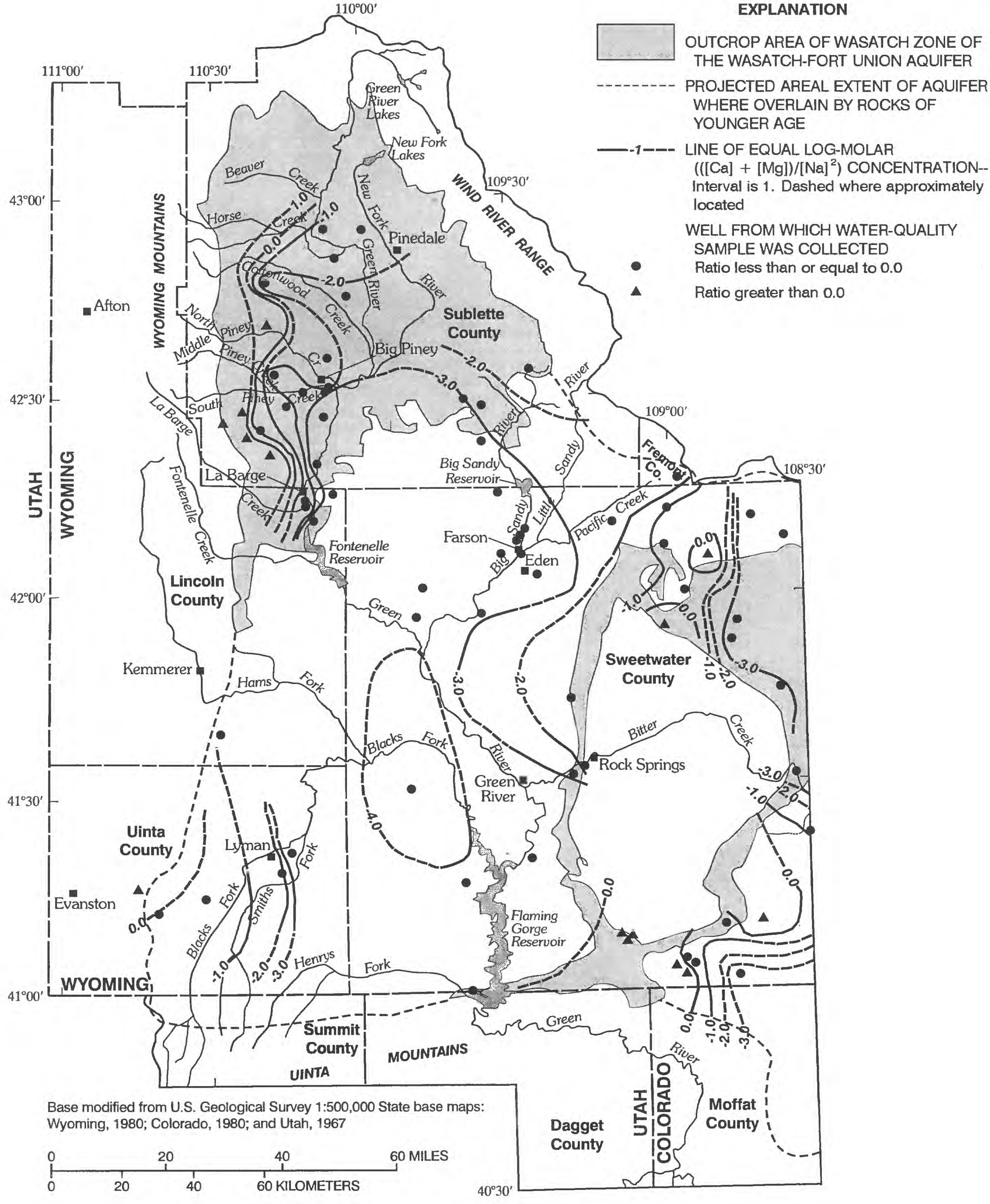


Figure 50. Log-molar ratio of calcium plus magnesium to squared-sodium concentration in water from the Wasatch zone of the Wasatch-Fort Union aquifer, Green River Basin aquifer system (modified from Naftz, 1996, p. 38).

Table 8. *Estimated ground-water recharge from streams and discharge to streams in the Green River Basin*

[+, indicates recharge to aquifers (streamflow loss); - indicates discharge from aquifers (streamflow gain)]

	Estimated ground-water recharge and discharge	
Stream	Streamflow-gaging station data (cubic foot per second)[1]	Flow-model analysis (cubic foot per second)[2]
Green River and tributaries upstream from Fontenelle Reservoir	-94	-98
Green River from Fontenelle Reservoir to Flaming Gorge Reservoir	-23	-23
Flaming Gorge Reservoir	-13	-14
Big Sandy River	-17	-12
Blacks Fork, Hams Fork, and Smiths Fork	+ 9	- 9
Henrys Fork	-16	Not estimated

[1]From Martin (1996, p. 27).
[2]From Martin (1996, p. 40).

about the same magnitude. The reason for the difference between measured and calculated ground-water discharge along Blacks, Hams, and Smiths Forks is unknown. The simulated direction of ground-water movement is consistent with the direction indicated by the potentiometric surface of the Bridger aquifer (fig. 44). The direction indicated by streamflow measurements in the three rivers seems inconsistent; therefore, it is possible that the streamflow measurements are too widely spaced to be useful. It also is possible that losses of water both from the rivers and the aquifer are due to unmeasured evapotranspiration along the stream banks; the anomalous log-molar ratio is caused by mixing of surface water with ground water. The net discharge for the Green River Basin is estimated from measured gains and losses to be 154 ft^3/s.

Springs and seeps are common in the Green River Basin but generally represent discharge points for local flow systems. Springs are particularly common along basal contacts of discontinuous aquifers in Miocene rocks such as those associated with the underlying Bishop Conglomerate of Oligocene age. Other springs and seeps are located in topographically elevated areas where local low permeability zones limit vertical movement.

GROUND-WATER MOVEMENT

Rates and directions of ground-water movement were estimated during calibration of a basin flow model. Ground-water movement along a north-south section across the Green River Basin is summarized in figure 51.

Ground-water movement in the Bridger aquifer is primarily horizontal. Substantial vertical leakage into deeper aquifers occurs only along the Uinta Uplift, where the underlying Laney aquifer is conglomeratic. Ground water in the Bridger aquifer moves from recharge areas along the Uinta Uplift to the north and northwest, and discharges along the Blacks Fork and Smiths Fork. The quantity of water moving in the Bridger aquifer north of Blacks Fork is small compared to other areas.

Rates of ground-water movement in the Laney aquifer are greatest where the Laney is unconfined. Rates of movement in the Laney aquifer, where buried by the Bridger aquifer, are much smaller. Most water enters the Laney aquifer by upward leakage from the New Fork aquifer in the central part of the Green River Basin and by recharge from irrigation return flow near Farson, Wyo. Small quantities of water enter the Laney aquifer adjacent to the Uinta Uplift by downward leakage from the Bridger aquifer. Virtually no water leaks upward into the Laney aquifer in areas underlain by bedded trona deposits (fig. 18) of the Wilkins Peak confining unit in the central part of the basin. Most water in the Laney aquifer discharges along the Big Sandy River and the Green River between Fontenelle and Flaming Gorge Reservoirs.

Water enters the New Fork aquifer from recharge areas adjacent to the Wind River Uplift and by upward leakage from the underlying Wasatch zone of the Wasatch-Fort Union aquifer. Rates of flow are large because of the large hydraulic conductivity of the aquifer. The small dissolved-solids concentration provides additional evidence of the short residence time for water in the New Fork aquifer. Ground water in the New Fork aquifer moves in a southerly direction and discharges by upward leakage into the Laney aquifer. The greatest rates of upward leakage occur along the Green River between Fontenelle Reservoir and the mouth of the Big Sandy River.

Ground-water in the Wasatch and Fort Union zones of the Wasatch-Fort Union aquifer moves along local and basin flow paths. Local flow is common where the Wasatch and Fort Union zones are at land surface along basin margins. Basin flow is common where the aquifers are buried.

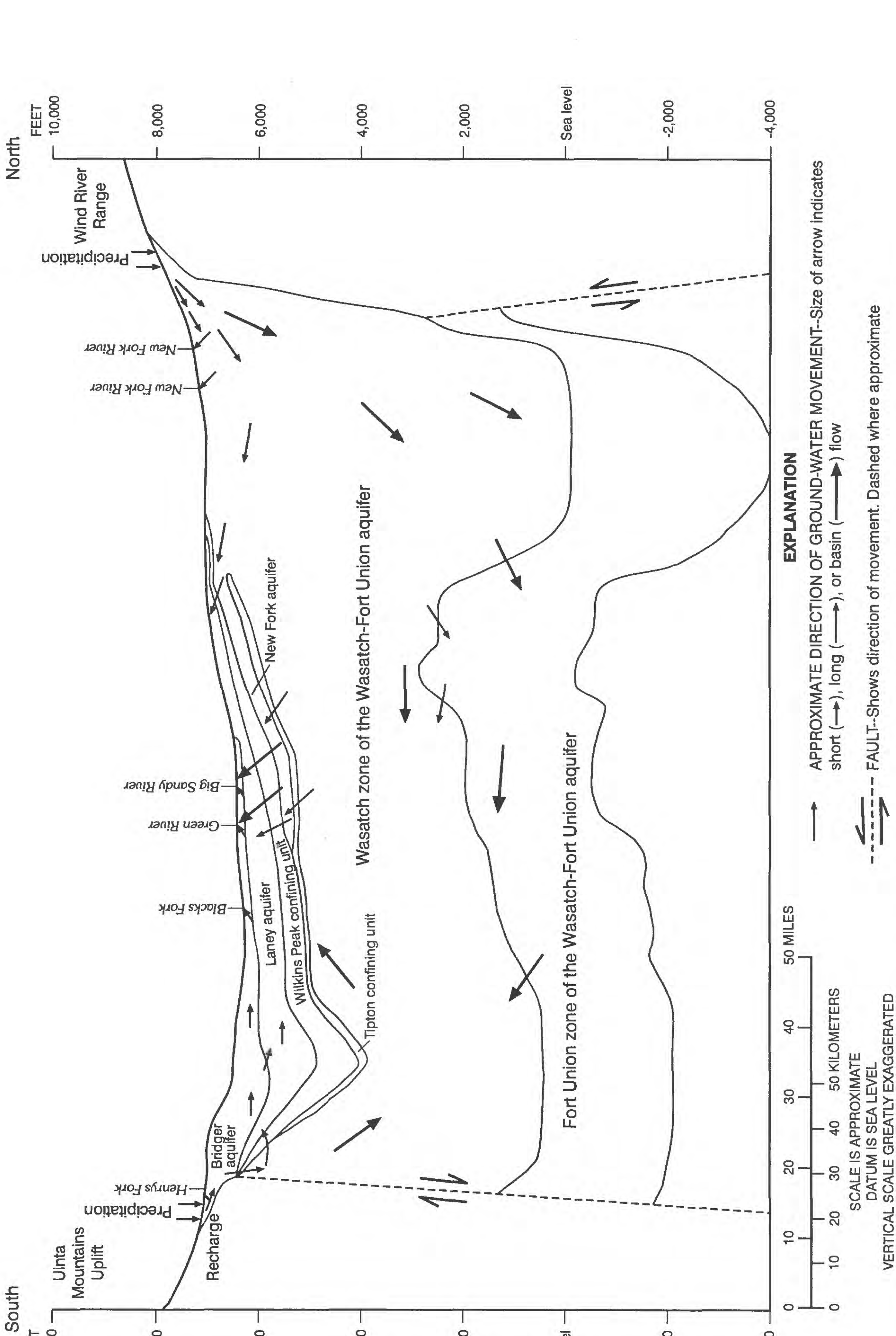


Figure 51. Diagrammatic cross-section summarizing ground-water movement in the Green River Basin aquifer system (from Martin, 1996, p. 29).

Large rates of movement in the Wasatch and Fort Union zones occur from recharge areas adjacent to the Wind River Uplift and Overthrust Belt along relatively short flow paths to discharge areas of the Green and New Fork Rivers. Large rates of movement also exist from recharge areas south of the Rock Springs Uplift to Flaming Gorge Reservoir.

Basin flow occurs in deep parts of the Wasatch and Fort Union zones, and where overlain by the Tipton confining unit. In the northern part of the Green River Basin, substantial volumes of water leak downward from shallow parts of the Wasatch and Fort Union zones and flow horizontally to the south. In the southern part of the basin, water enters the Wasatch and Fort Union zones by recharge and downward leakage along the Uinta Uplift. This water moves horizontally to Flaming Gorge Reservoir. The Flaming Gorge Reservoir area is the principal focus of discharge for the basin flow system. Discharge in the area of the reservoir occurs by upward leakage to overlying geohydrologic units.

QUALITY OF WATER

The dissolved-solids concentrations in water from the Bridger aquifer (fig. 52) increase gradually northward from the southern margin of the Green River Basin to the central part of the basin. This trend in dissolved-solids concentration is locally similar to flow paths indicated by the potentiometric-surface map of the Bridger aquifer (fig 44). Lack of water-quality data for the Bridger aquifer in the northern part of the basin prevents further analysis of trends in dissolved-solids concentration.

Water types of representative samples from the Bridger aquifer are depicted by Stiff diagrams (fig. 52). Calcium bicarbonate and magnesium bicarbonate water predominate in recharge areas of the aquifer adjacent to the Uinta Mountains. Concentrations of sodium plus potassium, bicarbonate plus carbonate, sulfate, and chloride increase along probable ground-water flow paths.

Distributions of dissolved-solids and major ions in water samples from the Laney aquifer are shown in figure 53. The dissolved-solids concentration increases along basin flow paths from the basin margins inward. Sodium bicarbonate and sodium sulfate water types generally predominate in the Laney aquifer.

Distributions of dissolved solids analyzed from water samples of the Wasatch zone of the Wasatch-Fort Union aquifer indicate gradual increases in dissolved-solids concentrations from basin margins inward, along projected basin flow paths (fig. 54). Concentrations of calcium and magnesium decrease along the basin flow paths whereas carbonate-bicarbonate and sodium plus potassium concentrations generally increase along basin flow paths. Water in the central part of the Green River Basin with dissolved-solids concentrations exceeding 3,000 mg/L, usually is dominated by sodium and chloride ions. Sulfate concentrations generally increase along basin flow paths.

Insufficient water-quality data are available for mapping dissolved-solids concentrations in water from the Fort Union zone of the Wasatch-Fort Union aquifer.

GEOHYDROLOGY OF THE GREAT DIVIDE-WASHAKIE-SAND WASH BASINS AQUIFER SYSTEM

The Great Divide-Washakie-Sand Wash Basins aquifer system consists of one multi-basin aquifer and one overlying confining unit. The Wasatch-Fort Union aquifer is composed of the Wasatch and Fort Union zones with no intervening confining unit. However, hydraulic properties of the two units seldom are similar at the same location. Designating two zones simplifies discussion of the geohydrology that follows. Similarly there is no extensive confining unit separating the Wasatch-Fort Union aquifer from the underlying Mesaverde aquifer. The stratigraphic relation of aquifers and confining units is shown on plate 1.

HYDRAULIC CONDUCTIVITY

More than 100 point estimates of hydraulic conductivity were available for the Great Divide-Washakie-Sand Wash Basins aquifer system, but mapping the spatial distribution of hydraulic conductivity in each aquifer zone is not possible. All the estimates were determined using data from aquifer tests (mostly specific-capacity tests) or drill-stem tests. Most aquifer and specific-capacity tests were done in the drainages of Bitter Creek or Separation Creek, and most drill-stem tests were conducted in a narrow band along the Colorado-Wyoming border. Estimates of hydraulic conductivity by interpolation between these three areas cannot be made with any degree of confidence. Even with estimates of hydraulic conductivity in each of the three areas, they do not appear to be spatially correlated. Consequently, the basin distribution of hydraulic conductivity was not mapped.

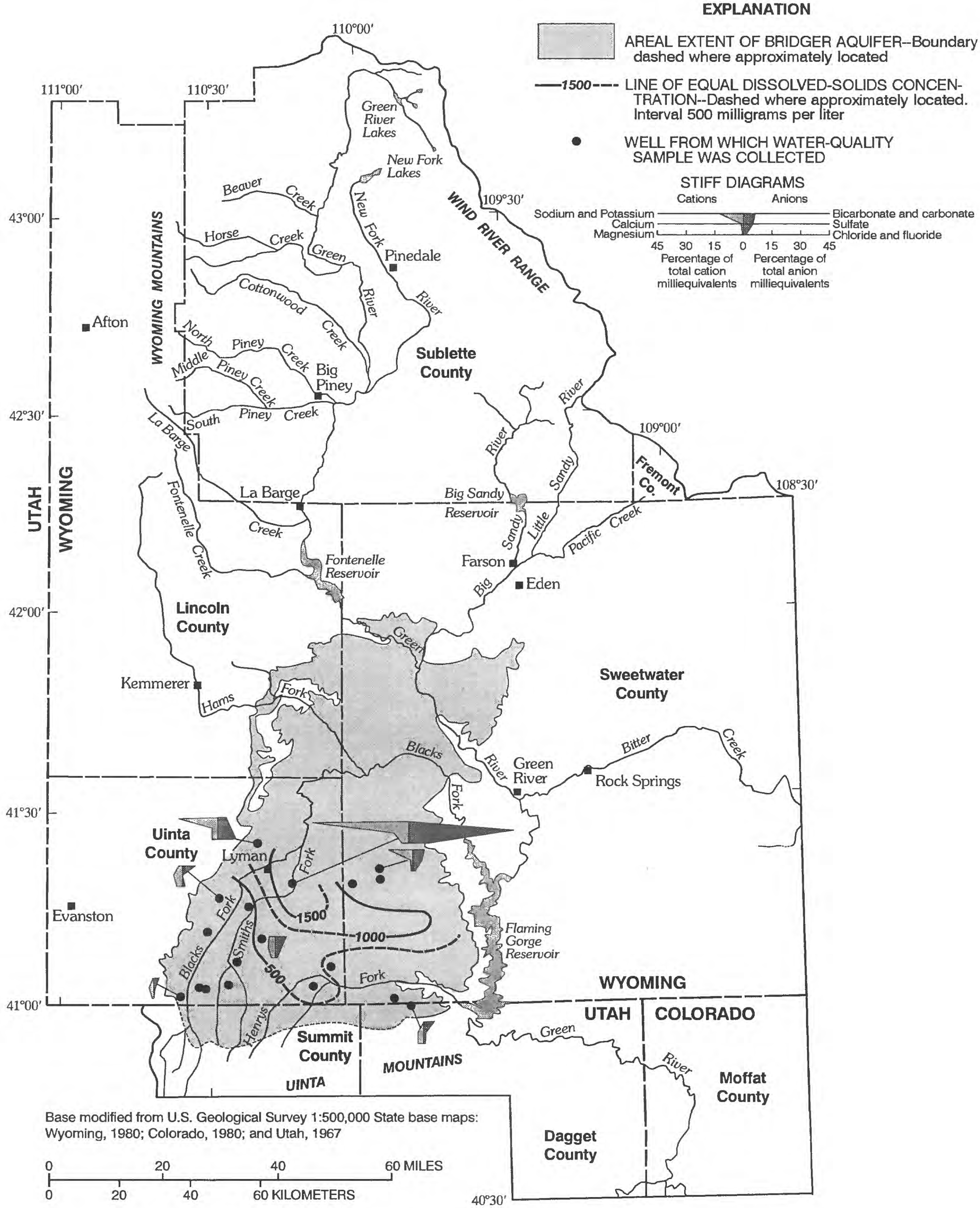


Figure 52. Dissolved-solids concentration and Stiff diagrams for the Bridger aquifer, Green River Basin aquifer system (modified from Naftz, 1996, p. 12-13).

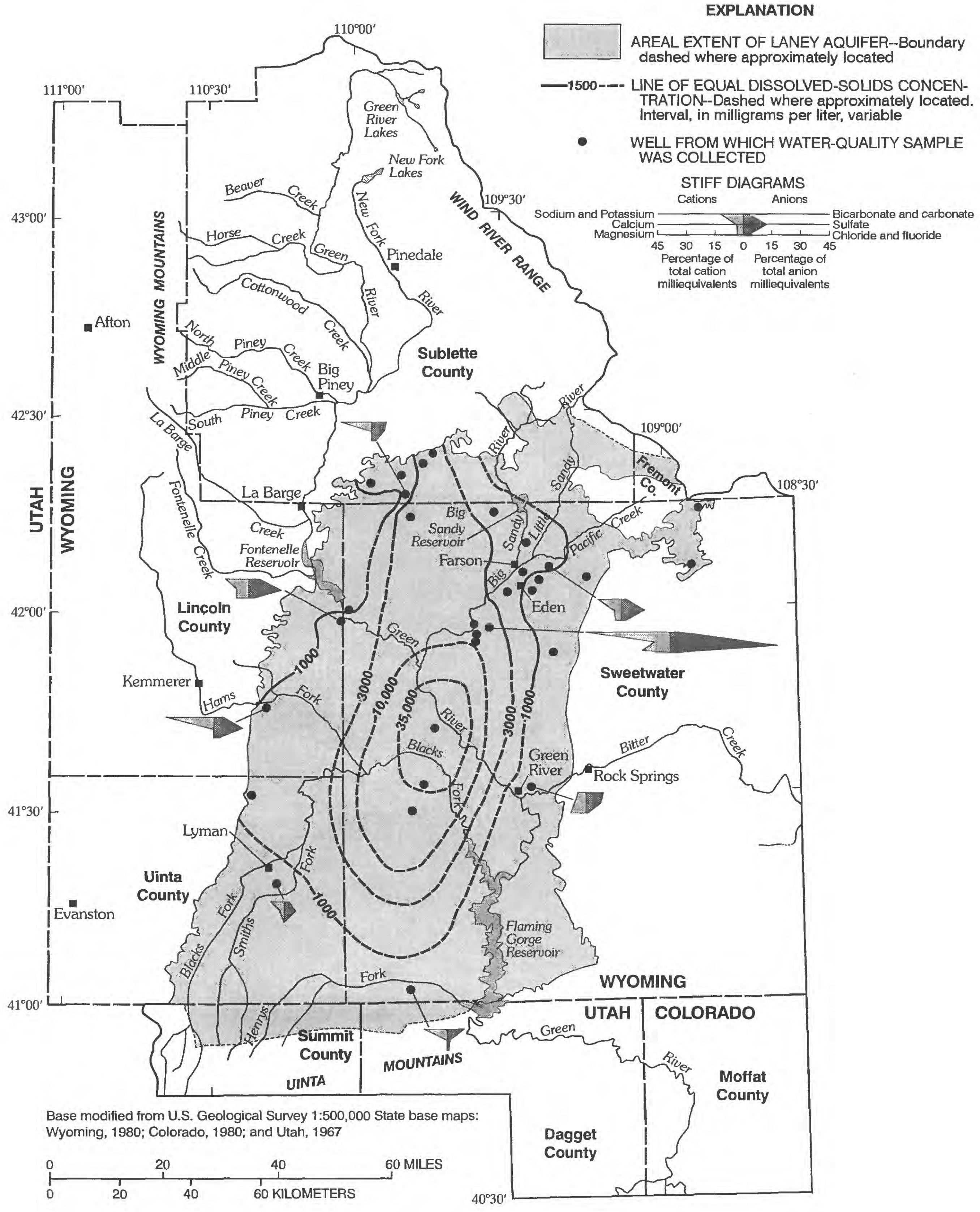


Figure 53. Dissolved-solids concentration and Stiff diagrams for the Laney aquifer, Green River Basin aquifer system (modified from Naftz, 1996, p. 24-25).

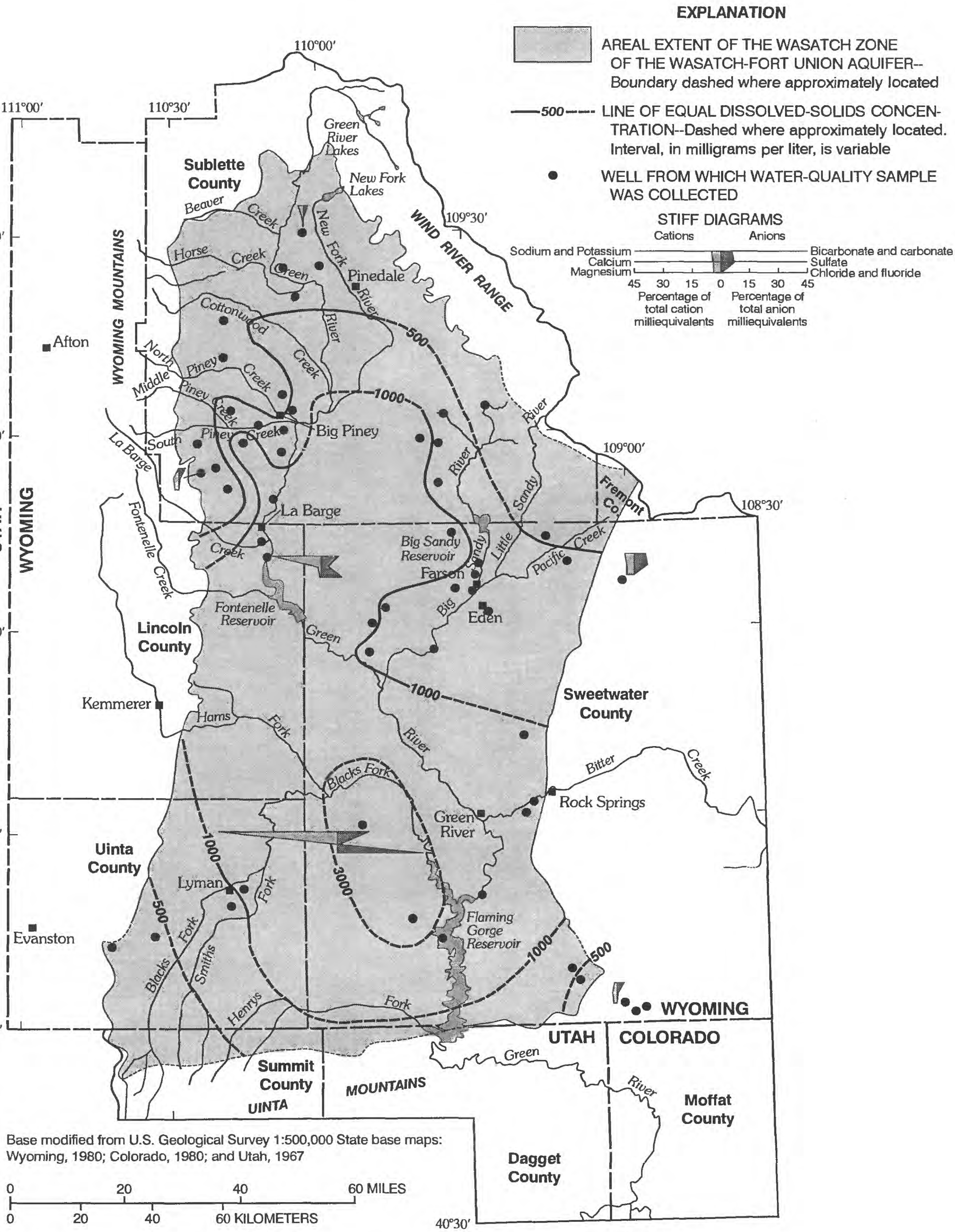


Figure 54. Dissolved-solids concentration and Stiff diagrams for the Wasatch zone of the Wasatch-Fort Union aquifer, Green River Basin aquifer system (modified from Naftz, 1996, p. 36-37).

Although the distribution of hydraulic conductivity in the Great Divide-Washakie-Sand Wash Basins aquifer system cannot be mapped, statistics calculated from point values indicate the possible range in hydraulic conductivity for each geohydrologic unit (table 9). The hydraulic-conductivity estimates summarized in table 9 may be more indicative of the distribution of test locations, however, than differences in hydraulic conductivity between geohydrologic units. Drill-stem tests generally are conducted in deeply buried rock that is suspected to be a trap for oil and gas. Within the geologic setting of the Wasatch and Fort Union zones of the Wasatch-Fort Union aquifer, deeply buried rocks are likely to have less sandstone and smaller hydraulic conductivity than rocks exposed at land surface near basin margins. Aquifer tests generally are conducted in shallow wells near basin margins. Therefore, the larger estimates of hydraulic conductivity obtained from aquifer tests are consistent with known geologic characteristics of the aquifer.

HYDRAULIC HEAD

The distribution of hydraulic head is mapped for the Wasatch zone (fig. 55) and for the Fort Union zone (fig. 56) of the Wasatch-Fort Union aquifer. The potentiometric surface of the Wasatch zone in 1985 was compiled using 1985 water-level measurements in wells that generally penetrated less than 300 ft. Hydraulic-head data used to compile the potentiometric surface of the Fort Union zone generally were obtained from water wells in outcrop areas and from drill-stem tests in sandstone from deeply buried parts of the aquifer. Hachured contour areas generally indicate discharge areas at the centers of the Great Divide Basin in the north and the Washakie Basin in the south where the Wasatch zone is exposed at land surface.

GROUND-WATER RECHARGE AND DISCHARGE

Areas of recharge are indicated by the potentiometric-surface map for the Wasatch zone, but data are not available to estimate the magnitude and distribution of recharge. Presumably, recharge occurs in the Great Divide-Washakie-Sand Wash Basins aquifer system, as in other Tertiary basins, at higher altitudes where increased precipitation and snowpack in combination with permeable formations at land surface facilitate infiltration of water. These conditions occur along the northern, western, and southeastern boundaries of the aquifer system.

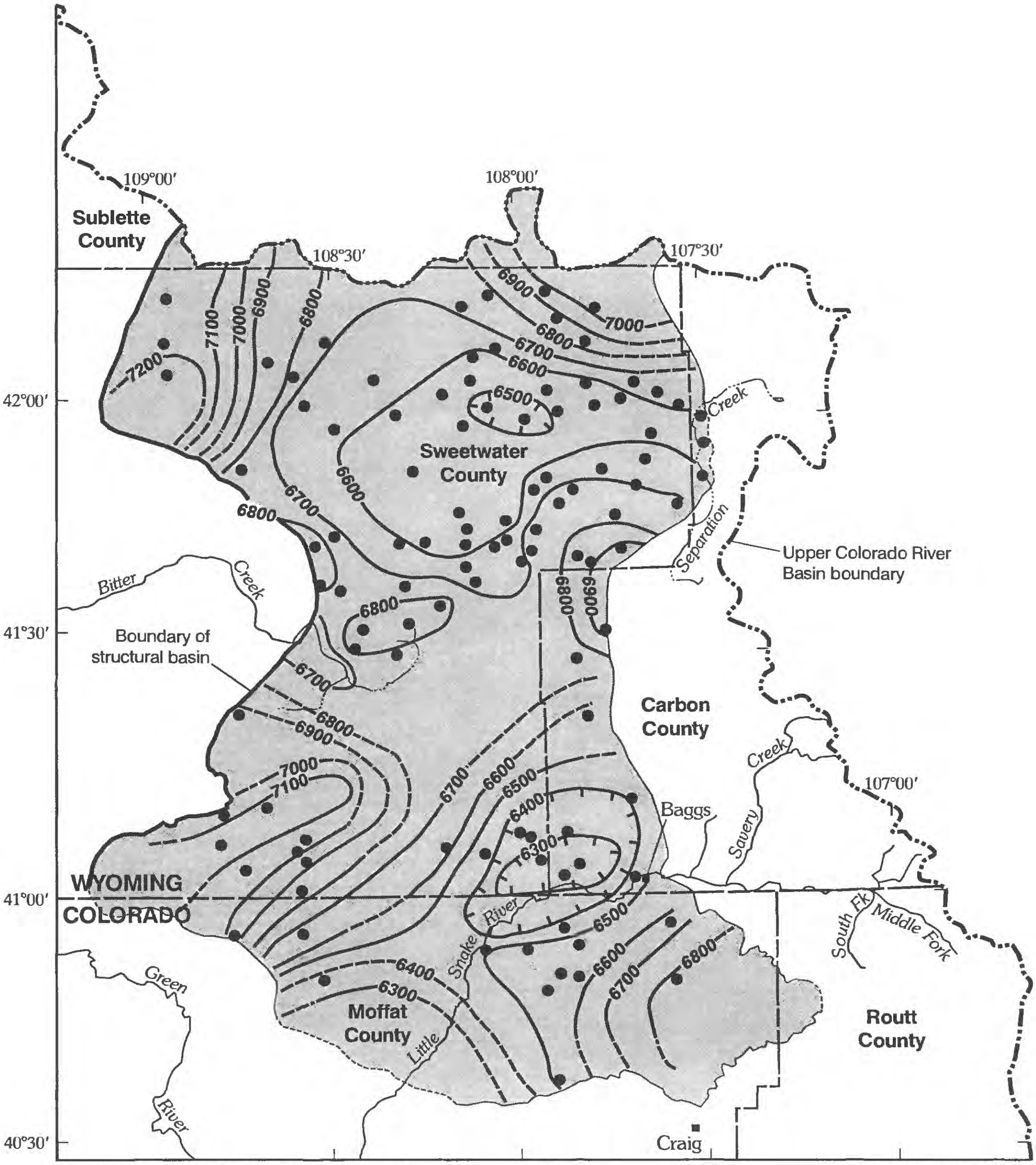
Table 9. *Summary of well-yield, spring-discharge, and hydraulic-conductivity data, Great Divide-Washakie-Sand Wash Basins aquifer system*

[gal/min, gallon per minute; ft/d, foot per day; --, not available]

Geohydrologic unit	Well-yield and spring-discharge rates (gal/min)	Hydraulic conductivity: Drill-stem tests (ft/d)	Hydraulic conductivity: Aquifer and specific-capacity tests (ft/d)
Confining unit	Generally less than 30; locally as large as 100.	range: 0.01 to 0.49	--
Wasatch zone of the Wasatch-Fort Union aquifer	Highly variable; less than 30 to greater than 200.	median: 0.05 range: 0.004 to 1.57	range: 0.03 to 9.1
Fort Union zone of the Wasatch-Fort Union aquifer	Highly variable; less than 30 to greater than 100.	median: 0.02 range: 0.001 to 0.22	range: 0.02 to 938

Concentrations of calcium and magnesium relative to sodium in water from the Wasatch zone were compiled and contoured (fig. 57). Positive log $(([Ca]+[Mg])/[Na]^2)$ values may indicate recharge in the northern and western part of the Great Divide Basin and in the western part of the Washakie Basin. Recharge areas are not mapped on figure 57 because data required to estimate recharge are not available. Negative log $(([Ca]+[Mg])/[Na]^2)$ values indicate downgradient areas and occur toward the center of the Great Divide and Washakie Basins (fig. 57). The positive log $(([Ca]+[Mg])/[Na]^2)$ values in the southern part of the Washakie Basin probably are indicative of localized recharge.

Discharge from the basin-fill aquifer, one of the locally important aquifers, occurs at numerous seeps and springs throughout the Great Divide-Washakie-Sand Wash Basins aquifer system and occurs along several streams including Bitter Creek, Separation Creek, and the Little Snake River. The largest concentration of seeps and springs is located near the center of the Great Divide Basin where the potentiometric surface of the Wasatch zone indicates potential discharge. Discharge from these springs and seeps rarely exceeds 2 gal/min and typically is so small as to preclude direct measurement.



Base modified from U.S. Geological Survey 1:500,000 State base maps: Wyoming, 1980; and Colorado, 1980

0 20 40 MILES

0 20 40 KILOMETERS

Hydrology by K.C. Glover

EXPLANATION

AREAL EXTENT OF WASATCH ZONE OF THE WASATCH-FORT UNION AQUIFER--Boundary dashed where approximately located

—*6300*— — POTENTIOMETRIC CONTOUR--Shows altitude at which water level would have stood in tightly cased wells, 1985. Dashed where approximately located. Hachures indicate depression. Contour interval 100 feet. Datum is sea level

● WELL IN WHICH WATER LEVEL WAS MEASURED

Figure 55. Potentiometric surface (1985) of the Wasatch zone of the Wasatch-Fort Union aquifer, Great Divide-Washakie-Sand Wash Basins aquifer system.

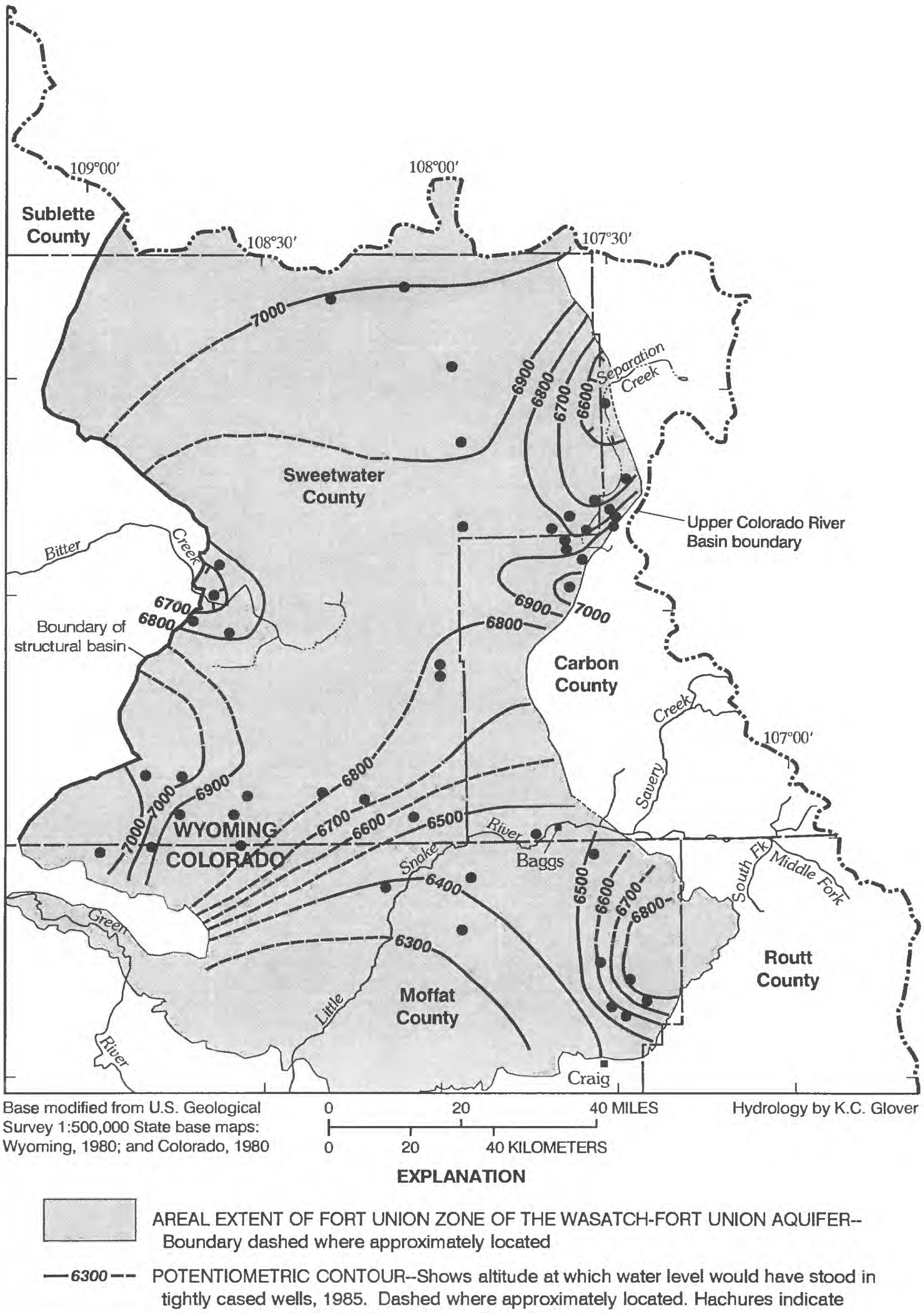
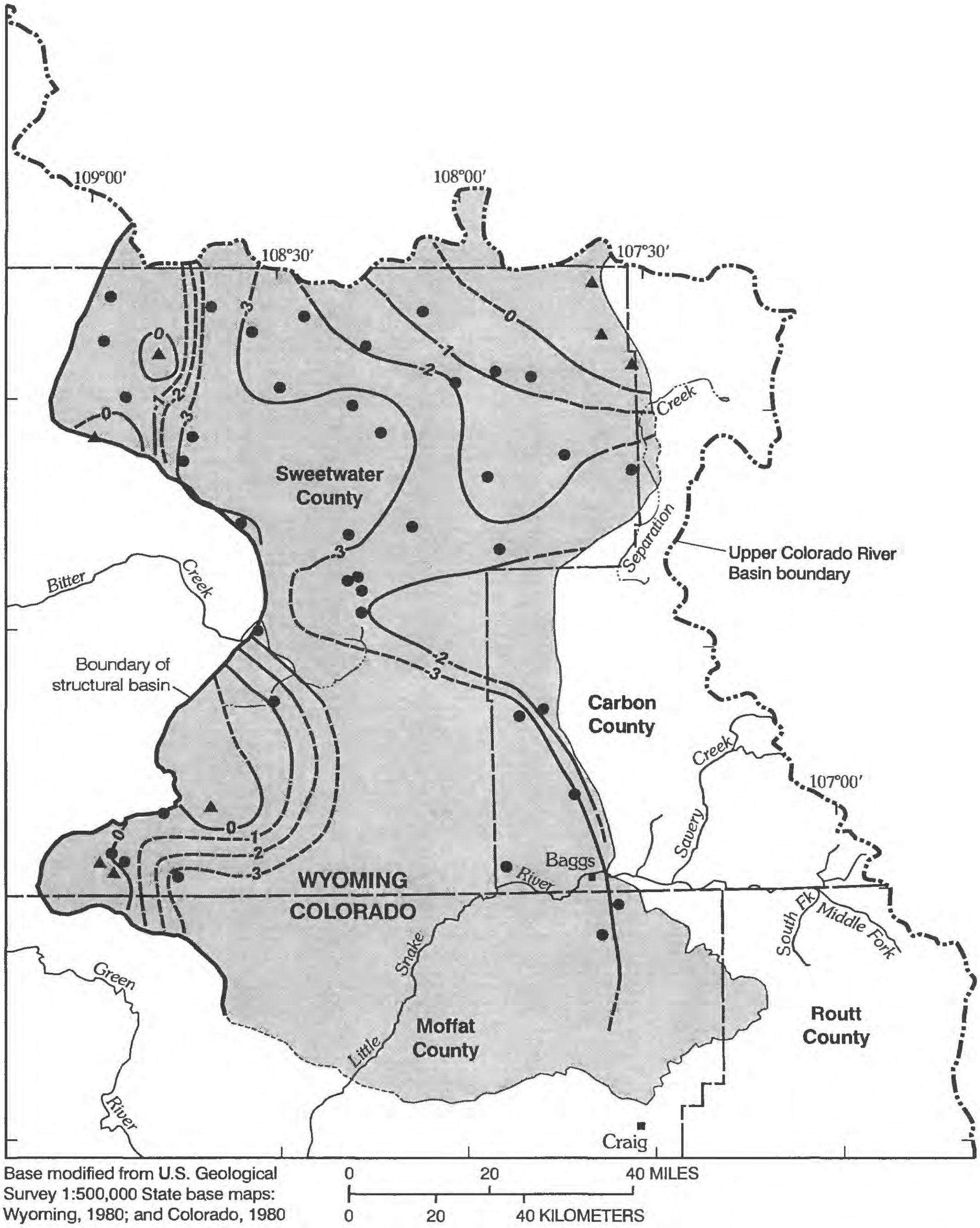


Figure 56. Potentiometric surface (1985) of the Fort Union zone of the Wasatch-Fort Union aquifer, Great Divide-Washakie-Sand Wash Basins aquifer system.



EXPLANATION

AREAL EXTENT OF WASATCH ZONE OF THE WASATCH-FORT UNION AQUIFER--Boundary dashed where approximately located

—-3—- LINE OF EQUAL LOG-MOLAR (([Ca] + [Mg])/[Na]^2) CONCENTRATION--Interval is 1. Dashed where approximately located

WELL FROM WHICH WATER-QUALITY SAMPLE WAS COLLECTED

● Ratio less than or equal to 0.0

▲ Ratio greater than 0.0

Figure 57. Log-molar ratio of calcium plus magnesium to squared-sodium concentration in water from the Wasatch zone of the Wasatch-Fort Union aquifer, Great Divide-Washakie-Sand Wash Basins aquifer system (modified from Naftz, 1996, p. 38).

Ground-water discharge in the Great Divide Basin, combined with surface runoff during snowmelt, forms numerous small lakes and marshes during the spring and early summer. The lakes and marshes are discharge points for closed drainage basins. Water that collects in these lakes and marshes is lost during the summer by evapotranspiration. Estimates of ground-water discharge could be made by conducting detailed lake studies that account for precipitation and runoff, lake evaporation, transpiration by phreatophytes, changes in water stored in lakes, and possible movement of ground water between lakes. However, such studies have not been made, and the rate of ground-water discharge in the Great Divide Basin is not known.

Efforts to estimate ground-water discharge to Bitter Creek, Separation Creek, and the Little Snake River by conducting streamflow gain-and-loss studies were unsuccessful. Apparently, the rates of ground-water discharge along these streams are less than the accuracy possible in stream-discharge measurements. Therefore, ground-water discharge along these streams remains unknown.

GROUND-WATER MOVEMENT

Highlands adjacent to mountains and structural escarpments are recharge areas; lowlands near streams and in the center of the Great Divide Basin are discharge areas. Discharge areas in the Great Divide Basin correspond to a probable decrease in hydraulic conductivity as water flows from the arkosic sandstone of the Battle Spring Formation toward the less-permeable mixture of sandstone and shale in the Wasatch Formation. The flow paths between recharge and discharge areas in the Wasatch zone of the Wasatch-Fort Union aquifer generally are 30 mi or less. Flow paths in the Fort Union zone are much longer.

QUALITY OF WATER

Because of the lack of water-quality data for the Great Divide-Washakie-Sand Wash Basins aquifer system, only water in the Wasatch zone of the Wasatch-Fort Union aquifer has been sampled and analyzed; about 60 percent of the sites sampled contain dissolved-solids concentrations less than 1,000 mg/L. These fresh-water areas generally are located on the southeast and southwest basin margins and the northern margin from the vicinity of Separation Creek on the east to just north of Bitter Creek on the west. The dissolved-solids concentration in the Wasatch zone (fig. 58) gradually increases along basin flow paths from basin margins inward. Stiff diagrams for water from the Wasatch zone indicate decreases in calcium and magnesium and increases in carbonate plus bicarbonate and sodium plus potassium concentrations along basin flow paths.

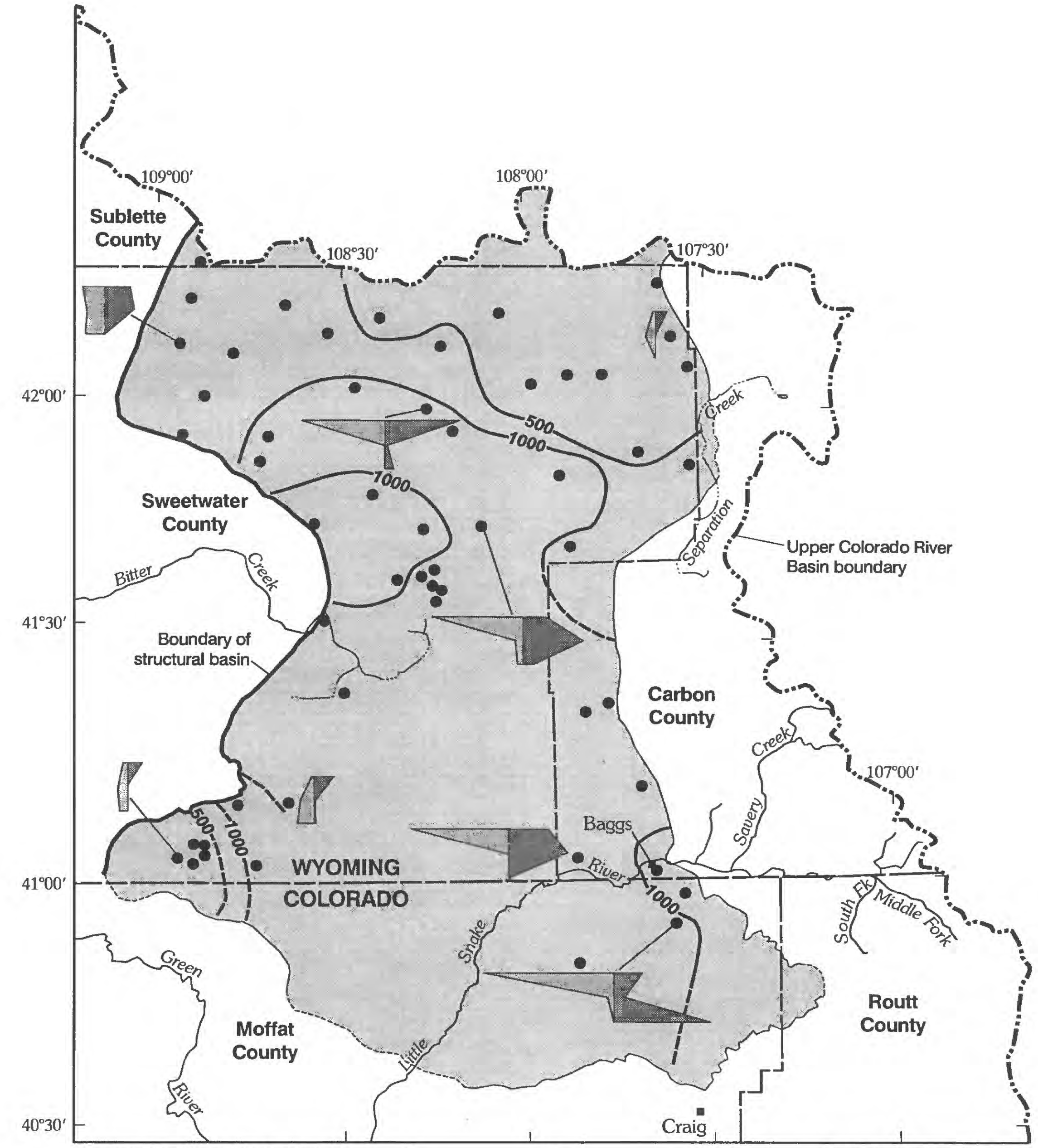
GEOHYDROLOGY OF THE MESAVERDE AQUIFER

The basin aquifer systems in Tertiary rocks described in this report are directly underlain by the Cretaceous Mesaverde aquifer. A summary description of the Mesaverde aquifer emphasizes aspects of the aquifer that affect vertical ground-water movement between aquifers in Tertiary rocks and the Mesaverde. The stratigraphy and lithology of the aquifer are described in the section of this report, Late Cretaceous Geology (p. 6). The aquifer is separated from aquifers in Tertiary rocks by confining units in the Piceance Basin aquifer system and the Uinta Basin aquifer system (pl. 1). No extensive confining unit separates the Mesaverde aquifer from aquifers in Tertiary rocks in the Green River Basin aquifer system and in the Great Divide-Washakie-Sand Wash Basins aquifer system. The following description of the Mesaverde aquifer is summarized primarily by Freethey and Cordy (1991).

Freethey and Cordy emphasized parts of the study area where geohydrologic data were readily available, mostly along the margins of structural basins. Data generally were not available on a regional basis to justify development of hydrologic models or to consider multiple-phase fluid movement of water, oil, and gas. The interpretations presented by Freethey and Cordy (1991) are general and qualitative for parts of the Mesaverde aquifer where few data are available or where the aquifer contains oil and gas.

Large reservoirs of natural gas are common within the stratigraphic units that constitute the Mesaverde aquifer. Located in deeply buried parts of the Green River, Great Divide, Piceance, and Washakie Basins, the reservoirs appear to be unrelated to stratigraphic or structural features, tend to have hydraulic heads that are anomalously large, and have very small hydraulic conductivity. Fluid recovered during drilling and testing of these reservoirs tends to be predominately gas with little water.

The unusual characteristics and large size of gas reservoirs has an important effect on the regional geohydrology of the Mesaverde aquifer. Spencer and Krupa (1985) compiled a bibliography of reports describing characteristics of these low-permeability reservoirs.

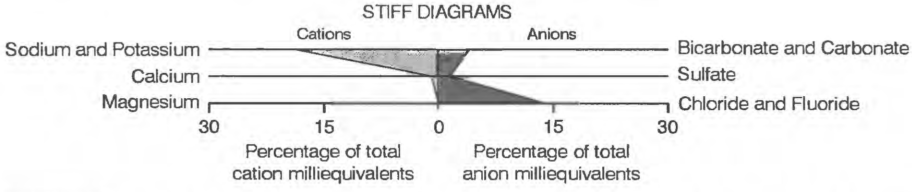


Base modified from U.S. Geological Survey 1:500,000 State base maps: Wyoming, 1980; and Colorado, 1980

0 20 40 MILES
0 20 40 KILOMETERS

Geochemistry by D.L. Naftz

EXPLANATION



AREAL EXTENT OF WASATCH ZONE OF THE WASATCH-FORT UNION AQUIFER--Boundary dashed where approximately located

—3— — — LINE OF EQUAL DISSOLVED-SOLIDS CONCENTRATION FOR THE WASATCH ZONE OF THE WASATCH-FORT UNION AQUIFER--Dashed where approximately located. Interval 500 milligrams per liter

● WELL FROM WHICH WATER-QUALITY SAMPLE WAS COLLECTED

Figure 58. Dissolved-solids concentration and Stiff diagrams for the Wasatch zone of the Wasatch-Fort Union aquifer, Great Divide-Washakie-Sand Wash Basins aquifer system (modified from Naftz, 1996, p. 36-37).

Law and Dickinson (1985) proposed a conceptual model to describe origin and occurrence of abnormally pressured gas reservoirs.

HYDRAULIC CONDUCTIVITY

In general, the regional distribution of hydraulic conductivity in the Mesaverde aquifer is highly variable and does not coincide with the regional distribution of lithofacies. In areas near the Rock Springs and Sierra Madre Uplifts, estimated hydraulic conductivity is as large as 1 ft/d (fig. 59). In areas where the Mesaverde aquifer is deeply buried, estimated hydraulic conductivity is less than 0.01 ft/d. The map of hydraulic conductivity shown in figure 59, based on an assumption that the aquifer is saturated with water, was constructed by contouring point estimates of hydraulic conductivity obtained from drill-stem tests.

HYDRAULIC HEAD

The potentiometric-surface map of the Mesaverde aquifer (fig. 60) was derived from hydraulic-head measurements made during drill-stem tests and from water-level measurements in wells. Drill-stem tests were conducted primarily where the aquifer is deeply buried. Water-level measurements generally were made in wells located in outcrop areas. The map shows the distribution of hydraulic head both in areas where the formation fluid is water and where it is primarily gas.

The potentiometric-surface map for the Mesaverde aquifer (fig. 60) shows several areas where the hydraulic head seems anomalously high. The most anomalous areas are in the Great Divide and Washakie Basins where heads are greater than 8,500 ft. Other areas of anomalously high head are in the northern Piceance Basin, the northern Green River Basin, and possibly a small area along the Colorado-Wyoming border in which heads are in excess of 8,000 ft. These head anomalies occur in association with low-permeability gas reservoirs.

Because the high hydraulic heads have caused problems when completing wells, petroleum geologists have investigated these anomalies extensively (Spencer and Krupa, 1985). Petroleum geologists have considered heads to be anomalously large whenever the vertical pressure gradient is substantially greater than gradients that are typical of hydrostatic conditions. That is, heads greater than those typically measured in shallow wells located in recharge areas would be considered anomalous.

For example, hydraulic heads greater than 8,500 ft in the Great Divide and Washakie Basins would be anomalous because heads in overlying and underlying aquifers are less than 8,500 ft and the land-surface altitude of recharge areas in the vicinity of the Rawlins and Sweetwater Uplifts is less than 8,500 ft. The head anomaly is probably the result of either a transient flow system or the addition of fluid from sources within the aquifer or adjacent aquifers. Similar, but less apparent conditions may cause head anomalies in the northern Piceance Basin, northern Green River Basin, and along the Colorado-Wyoming border.

Law and Dickinson (1985) have proposed a model that explains the origin and persistence of abnormally high heads in the Mesaverde aquifer. The model is based, in part, on (1) the observation that abnormally large heads occur in gas reservoirs with small hydraulic conductivity, (2) formation temperatures generally are greater than 180°F, and (3) the formation contains large quantities of coal or carbonaceous shale. At these temperatures, thermogenic gas probably accumulates at rates that exceed gas loss. Free water in the rock is forced out of the gas-generation zone into overlying and updip rocks where heads are normal. The resulting lack of free water within the gas-generation zone limits dissolution of minerals and enhancement of hydraulic conductivity. Thus, while diagenesis continues, a rock with very small hydraulic conductivity develops, gas accumulates, and pressure head increases.

GROUND-WATER RECHARGE AND DISCHARGE

Locations of potential ground-water recharge and discharge are indicated by the potentiometric-surface map of the Mesaverde aquifer (fig. 60). Recharge occurs along the eastern margins of the Piceance, Sand Wash, and Washakie Basins; the western margin of the Uinta Basin; the Overthrust Belt; and in topographically elevated areas of the Rock Springs Uplift (tectonic features are shown in fig. 2). Other areas of large head in the Mesaverde aquifer coincide with gas reservoirs or possibly receive water by vertical leakage from adjacent aquifers. Discharge from the Mesaverde aquifer occurs as springs in the High Plateaus of Utah and the Rock Springs Uplift. Discharge to streams occurs along Bitter Creek (Rock Springs Uplift), Muddy Creek (east side of the Washakie Basin), the Colorado River east of Grand Junction, Colo., the Price and Green Rivers (Uinta Basin), and the White River near Meeker, Colo. Other hydraulically downgradient areas, such as along the Little Snake River, probably leak water into overlying

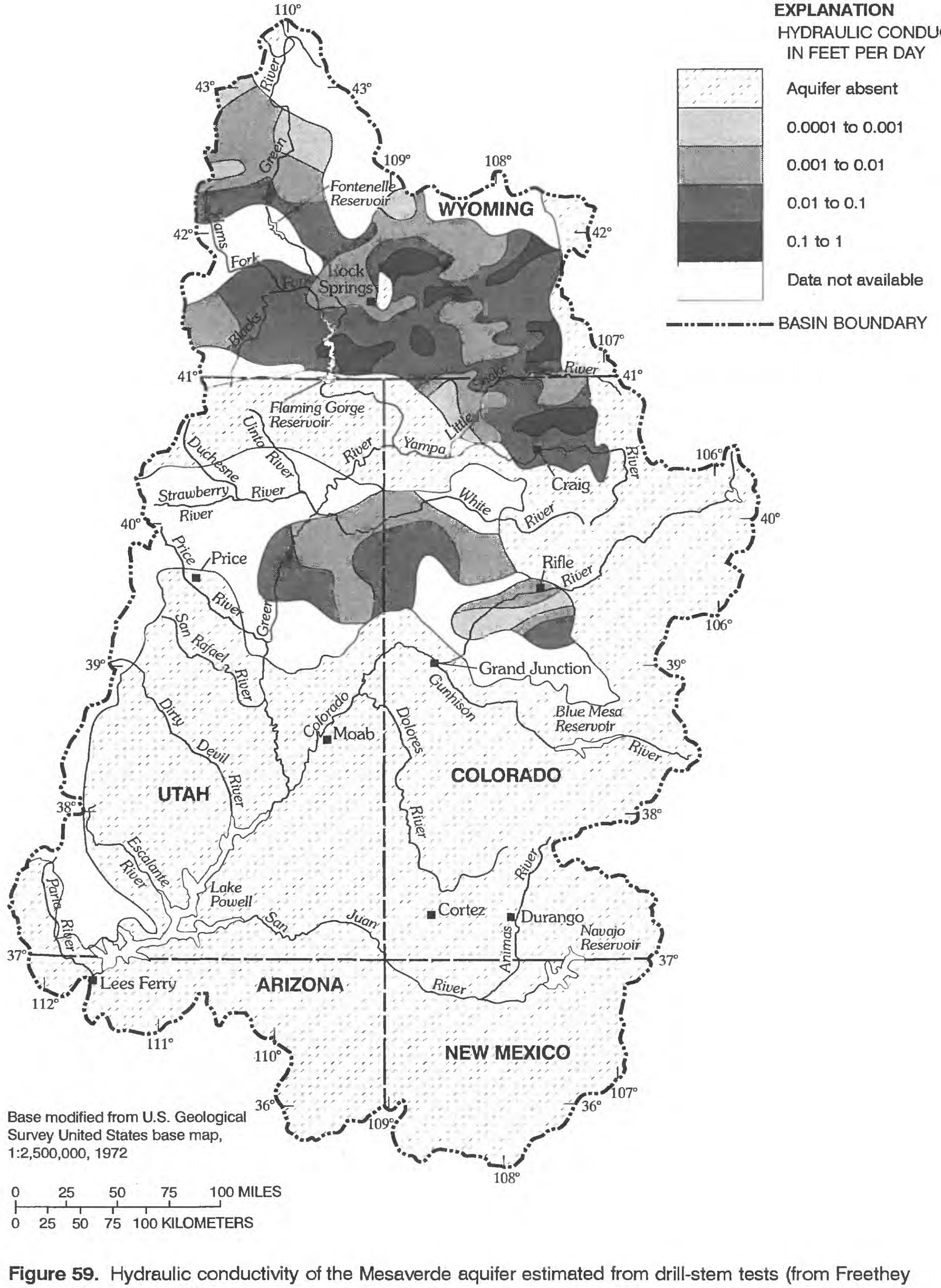


Figure 59. Hydraulic conductivity of the Mesaverde aquifer estimated from drill-stem tests (from Freethey and Cordy, 1991, p. 67).

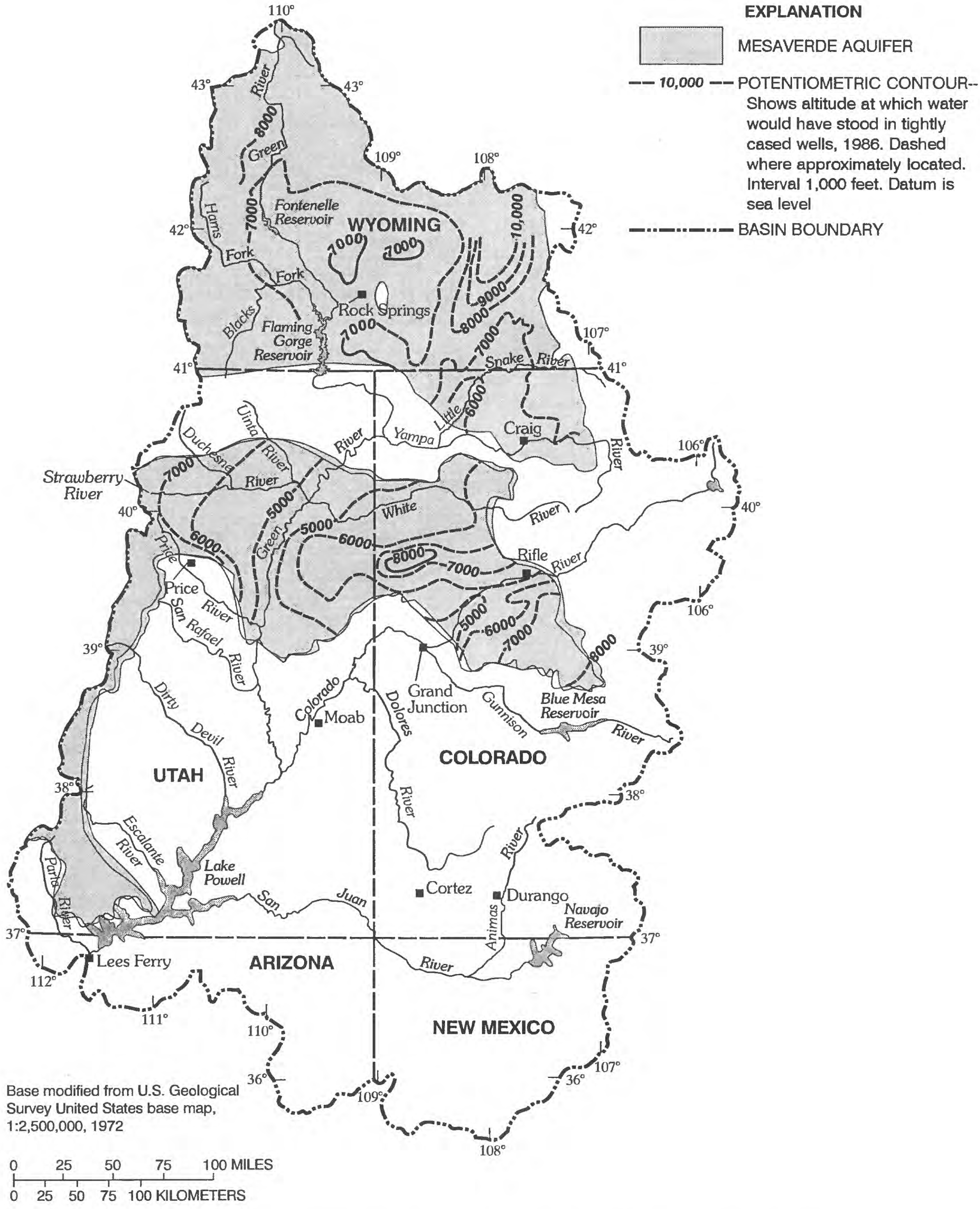


Figure 60. Potentiometric surface (1986) of the Mesaverde aquifer (from Freethey and Cordy, 1991, pl. 5).

aquifers. As described by Freethey and Cordy (1991), recharge was estimated for all aquifers in Mesozoic rocks based on the method of Eakin and others (1951). Data are insufficient to estimate the amount of discharge along most streams.

GROUND-WATER MOVEMENT

Ground-water movement in the Mesaverde aquifer is complicated by variations in hydraulic conductivity and the presence of gas reservoirs. As a result, recharge and discharge areas are localized at topographic high and low areas, creating several discontinuous flow systems rather than a single basin flow system. Movement in deeply buried parts of the aquifer probably is negligible due to the presence of gas reservoirs having small hydraulic conductivity.

Possible areas of vertical leakage between the Mesaverde aquifer and overlying aquifers in Tertiary rocks are identified by comparing potentiometric-surface maps as described in by Freethey and Cordy, 1991. Vertical leakage into the Mesaverde aquifer potentially occurs along the southern end of the Rock Springs Uplift and along the eastern side of the Washakie Basin. Vertical leakage out of the Mesaverde aquifer is possible northwest of the Rock Springs Uplift. Freethey and Cordy (1991) also describes other areas where hydraulic gradients indicate possible leakage out of the Mesaverde aquifer, but these areas coincide with gas reservoirs having anomalously large heads. The lack of water recovery during drill-stem tests, together with the small hydraulic conductivity in these reservoirs, indicates vertical movement of water is limited.

The rate of vertical leakage in the Mesaverde aquifer probably is negligible when compared to the rates of movement within aquifers in Tertiary rocks. Assuming a value of 0.001 ft/d for vertical hydraulic conductivity between the Mesaverde and overlying aquifers, and using available estimates of hydraulic gradient and vertical distance between aquifers, estimated total leakage from the Mesaverde aquifer into aquifers in Tertiary rocks is 0.4 ft/s. Additional leakage of 2.1 ft/s was calculated as possible, assuming that the fluid in the vicinity of head anomalies is water. Water budgets for aquifers in Tertiary rocks described earlier in this report indicate that vertical-leakage rates generally are two to three orders of magnitude greater than the vertical-leakage rates presented by Freethey and Cordy (1991).

SUMMARY

Four basin aquifer systems in Tertiary rocks have been identified in the sedimentary basins of the Upper Colorado River Basin—the Piceance Basin aquifer system, the Uinta Basin aquifer system, the Green River Basin aquifer system, and the Great Divide-Washakie-Sand Wash Basins aquifer system. Each aquifer system is geographically and hydrologically isolated, and is areally coincident with the correspondingly named structural basin or basins. Aquifers and confining units are defined separately for each aquifer system. No region-wide aquifers of Tertiary age have been identified; however, each basin aquifer system is hydraulically connected to the underlying Mesaverde aquifer in rocks of Late Cretaceous age. The quantity of water moving between the Mesaverde aquifer and aquifers in Tertiary rocks is negligible.

PICEANCE BASIN AQUIFER SYSTEM

The hydrologic system in Tertiary rocks of the Piceance Basin consists of upper and lower Piceance Basin aquifers separated from each other and the underlying Mesaverde aquifer by confining units. The Mahogany confining unit separates the upper and lower Piceance Basin aquifers, and a basal confining unit separates the lower Piceance Basin aquifer from the underlying Mesaverde aquifer. The aquifers are truncated laterally by topography and limited to that part of the structural basin between the Colorado and White Rivers. Sedimentary rocks of the Piceance Basin aquifer system generally are fine-grained, well-cemented, and fractured. Aquifer hydraulic properties generally are related to the degree and interconnection of fractures.

The upper Piceance Basin aquifer is contained within the Uinta Formation and upper part of the Parachute Creek Member of the Green River Formation. The aquifer has an average thickness of about 700 ft. Hydraulic conductivity of the aquifer has been enhanced by structural deformation and fracturing and ranges from 0.003 to 1.6 ft/d as estimated by aquifer tests and flow-model analysis. The potentiometric surface for the aquifer indicates the general directions of ground-water movement, but water-level data are not available for many areas where the aquifer is present. The dissolved-solids concentration of water in the upper Piceance Basin aquifer increases from less than 500 mg/L to more than 1,000 mg/L down the flowpath along the direction of the hydraulic gradient.

The Mahogany confining unit, which correlates with the Mahogany zone of the Parachute Creek Member of the Green River Formation, has an average thickness of about 160 ft. Hydraulic conductivity of the confining unit is related to fractures that are generally less common and less interconnected than fractures in adjacent aquifers. As a result, hydraulic conductivity estimated by flow-model analysis generally is one to two orders of magnitude smaller for the Mahogany confining unit than the hydraulic-conductivity estimates for adjacent aquifers.

The lower Piceance Basin aquifer is contained within the lower part of the Parachute Creek Member of the Green River Formation and has an average thickness of 900 ft. Hydraulic conductivity is related to the presence of fractures and solution channels. Aquifer-test and model estimates of hydraulic conductivity for the lower Piceance Basin aquifer are similar to estimates for the upper Piceance Basin aquifer, ranging from 0.001 to 1.2 ft/d. The ratio of horizontal to vertical hydraulic conductivity was estimated to be 13.4 to 15.0. The potentiometric-surface map for the aquifer indicates general directions of ground-water movement from basin margins and upland areas toward streams. Dissolved-solids concentrations range from less than 1,000 mg/L to more than 10,000 mg/L. Water in the lower Piceance Basin aquifer is characterized by large quantities of sodium carbonate.

The basal confining unit of the Piceance Basin aquifer system consists of the lower members of the Green River Formation and underlying Tertiary rocks. The unit generally is 2,000 to 4,000 ft thick, has relatively few fractures, little sandstone, and typically small well yields. Few data are available, but hydraulic conductivity of the basal confining unit probably is less than 0.01 ft/d.

The Piceance Basin aquifer system receives recharge water in upland areas, transmits part of the water horizontally through the upper Piceance Basin aquifer toward discharge areas along major streams or springs on canyon walls, and transmits the remaining water downward across the Mahogany confining unit into the lower Piceance Basin aquifer. Water in the lower aquifer moves generally horizontally and, in the vicinity of discharge areas, leaks upward into the upper aquifer. Winter precipitation, stored as snowpack at higher altitudes of the Piceance Basin, provides most of the recharge to the ground-water system. Total estimated recharge is about 42 ft^3/s. Discharge occurs primarily as seepage to alluvium along Yellow and Piceance Creeks (30.7 ft^3/s) and springs near Roan Creek and Parachute Creek (11.6 ft^3/s). With local exceptions, the flow system is in steady state.

UINTA BASIN AQUIFER SYSTEM

Two major aquifers have been identified in the Tertiary rocks of the Uinta Basin—the Duchesne River-Uinta and Douglas Creek-Renegade aquifers. The Duchesne River Formation and Uinta Formation comprise the Duchesne River-Uinta aquifer. The Douglas Creek Member of the Green River Formation and intertonguing Renegade Tongue of the Wasatch Formation comprise the Douglas Creek-Renegade aquifer. The aquifers are separated by an upper confining unit consisting of the Parachute Creek Member of the Green River Formation. The lower confining unit, which consists primarily of the Wasatch Formation, separates the two aquifers from the underlying Mesaverde aquifer.

The Duchesne River-Uinta aquifer crops out, principally in the northern Uinta Basin where it varies in thickness from 2,000 to 8,000 ft. Hydraulic conductivity of the aquifer generally is related to the amount of sandstone present and is enhanced by fractures. Fracture-enhanced hydraulic conductivity is particularly important in the lower part of the aquifer and within the central part of the Uinta Basin. Hydraulic-conductivity estimates from aquifer tests range over several orders of magnitude. Estimates from flow-model analysis range from 0.5 to 1.0 ft/d and are similar to median value from aquifer tests. The potentiometric-surface map indicates a general direction of water movement from basin margins toward the Duchesne, Green, and White Rivers. Dissolved-solids concentrations of water increase down flow paths from less than 500 to more than 3,000 mg/L.

The upper confining unit separating the Duchesne River-Uinta and Douglas Creek-Renegade aquifers is approximately 3,000 to 6,000 ft thick. Well yields and spring discharges generally are less than 10 gal/min, and hydraulic conductivity, estimated from aquifer tests, ranges from 0.0002 to 0.11 ft/d. This thick, low permeability unit of shale, mudstone, and limestone effectively prevents measurable water from moving between the Duchesne River-Uinta and Douglas Creek-Renegade aquifers.

The Douglas Creek-Renegade aquifer is approximately 500 ft thick and occurs only in the southeastern Uinta Basin. Laterally to the northwest, the percentage of sandstone in the aquifer decreases, hydraulic conductivity becomes correspondingly small, and the unit acts as a confining unit. Hydraulic conductivity of the aquifer is related primarily to the percentage of sandstone present. Hydraulic conductivity, estimated from aquifer tests and flow-model analysis, ranges from

0.05 to 0.3 ft/d. Insufficient head data are available to map the potentiometric surface of the Douglas Creek-Renegade aquifer. Dissolved-solids concentration typically is less than 1,450 mg/L.

The lower confining unit of the Uinta Basin aquifer system typically is 3,000 to 6,000 ft thick and consists of relatively unfractured shale and limestone with minor amounts of sandstone. Well yields generally are less than 50 gal/min and hydraulic conductivity, estimated by aquifer tests, ranges from 0.0003 to 0.04 ft/d.

Horizontal water movement in aquifers in Tertiary rocks of the Uinta Basin occurs generally from basin margins toward major streams of the basin. Vertical water movement between aquifers has not been detected and, with local exceptions, probably is negligible. Recharge occurs in upland areas peripheral to the Uinta Basin where precipitation annually exceeds 10 in. Streams provide a secondary source of recharge water. Total recharge is estimated by empirical methods and flow-model analysis to be 272 ft^3/s to the Duchesne River-Uinta aquifer and 1.4 ft^3/s to the Douglas Creek-Renegade aquifer. Additional recharge probably occurs, particularly in the Douglas Creek-Renegade aquifer. However, discharge of any additional water occurs locally and, therefore, is not included in the estimates given here. Discharge occurs along the Duchesne, Green, and White Rivers, and other major streams of the basin. The Uinta Basin aquifer system, with local exceptions due to pumping from wells, is in steady state.

GREEN RIVER BASIN AQUIFER SYSTEM

The hydrologic system in Tertiary rocks of the Green River Basin consists of the Bridger, Laney, New Fork, and Wasatch-Fort Union aquifers, and the Wilkins Peak and Tipton confining units. All units correlate roughly with geologic formations or members of the same names. The Bridger, Laney, New Fork, and Wasatch-Fort Union aquifers are separated geographically or stratigraphically by confining units. The Wasatch and Fort Union zones of the Wasatch-Fort Union aquifer are geologically similar but separated on the basis of differences in hydrologic properties at any given geographic location.

The Bridger aquifer crops out in the southern part of the Green River Basin and generally is less than 1,000 ft thick. Hydraulic conductivity of the aquifer is related to the quantity of sandstone present and to the degree of fracturing. Hydraulic conductivity estimates from aquifer tests range from 0.03 to 420 ft/d. Model-derived estimates are 0.09 to 0.9 ft/d for horizontal hydraulic conductivity and 0.00001 ft/d for vertical hydraulic conductivity. Data used to compile the potentiometric surface of the Bridger aquifer indicate that local recharge and discharge probably are common and that vertical hydraulic gradients exist within the aquifer. Few water-quality data are available for the aquifer; however, dissolved-solids concentrations vary from less than 500 to more than 1,500 mg/L.

The Laney aquifer is extensively fractured where exposed at land surface in the north-central part of the Green River Basin but consists of relatively unfractured marlstone and shale where buried beneath the Bridger aquifer. Thickness generally ranges from 100 to 600 ft. Hydraulic conductivity is related to the degree of fracturing. Near the Big Sandy River, aquifer-test values as large as 1,400 ft/d have been recorded; well yields are correspondingly large. Where the Laney aquifer is buried and is relatively unfractured, horizontal hydraulic-conductivity values as small as 0.04 ft/d have been estimated by flow-model analysis. Vertical hydraulic conductivity has been estimated by flow-model analysis to range from 0.00001 to 17.3 ft/d. Sufficient head data are available to compile a potentiometric-surface map for the Laney aquifer only in areas where hydraulic conductivity is large—generally within the north-central part of the basin. Dissolved-solids concentrations increase downward along flow paths from less than 1,000 mg/L to more than 35,000 mg/L.

The Wilkins Peak confining unit consists of relatively unfractured marlstone, shale, and salt deposits. The unit is typically 100 to 600 ft thick and separates the Laney aquifer from underlying aquifers. Few data are available to estimate horizontal hydraulic conductivity of the Wilkins Peak confining unit, and it was not simulated. Vertical hydraulic conductivity estimated by flow-model analysis is approximately 0.00001 ft/d.

The New Fork aquifer is predominately sandstone and shale, with a typical thickness of 300 to 400 ft. The aquifer is limited to the north-central part of the basin; laterally to the north, it grades into the Wasatch-Fort Union aquifer. The New Fork aquifer thins to the south. Few aquifer-test data are available, but flow-model analysis has been used to estimate a horizontal hydraulic conductivity of 6.5 ft/d. Vertical hydraulic conductivity is estimated to be 0.1 ft/d. Insufficient data are available to compile a potentiometric-surface map or a map showing dissolved-solids concentrations for the New Fork aquifer.

The Tipton confining unit is a thin unit that vertically separates the New Fork aquifer from the Wasatch-Fort Union aquifer. The Tipton confining unit is absent in the northern Green River Basin; thus, the New Fork and Wasatch-Fort Union aquifers are in direct hydraulic contact. Aquifer-test data for the Tipton confining unit are not available, and horizontal hydraulic conductivity was not simulated. Vertical hydraulic conductivity estimated by flow-model analysis is about 0.00001 ft/d.

The Wasatch zone of the Wasatch-Fort Union aquifer is a thick sequence (typically 2,000 to 7,000 ft) of sandy shale and siltstone with varying quantities of channel sandstone. Sandstone predominates along basin margins and in the northern Green River Basin, and hydraulic-conductivity values estimated from aquifer tests are as large as 2,100 ft/d. Model-derived estimates are as large as 6.5 ft/d. Sandstone is less common in the south-central part of the basin where horizontal hydraulic-conductivity estimates are as small as 0.03 ft/d and vertical hydraulic-conductivity estimates are 0.001 ft/s. Hydraulic-head data are numerous where the aquifer is more permeable near land surface; elsewhere, head data are scarce. Flow paths are short in the northern part of the basin and along basin margins, indicating local recharge and discharge. Flow paths are longer in the south-central part of the basin. In areas where short flow paths are common, dissolved-solids concentrations typically are less than 1,000 mg/L. In the south-central part of the basin, values increase to more than 3,000 mg/L.

The Fort Union zone of the Wasatch-Fort Union aquifer is similar lithologically to the overlying Wasatch zone. Thickness typically ranges from 2,000 to 4,000 ft. Hydraulic conductivities estimated from aquifer tests range over five orders of magnitude with a median value of 40 ft/d. Model-derived estimates are small (0.00001 ft/d) in the northern part of the Green River Basin, increasing to a value of 0.3 ft/d for the southeastern part of the basin. Data from drill-stem tests were used to compile a potentiometric-surface map for the Fort Union zone and indicate general directions of water movement from basin margins toward the center of the basin. Insufficient water-quality data were available to map dissolved-solids concentrations.

Water in the Green River Basin aquifer system moves from recharge areas along basin margins, horizontally toward discharge areas and vertically into the deeper aquifers. In the vicinity of the Green River, Big Sandy River, and Blacks Fork, water leaks upward from deeper aquifers and discharges to streams and associated alluvium. Recharge was estimated by using a linear relation between recharge and average annual precipitation during flow-model development to equal 117 ft^3/s in the northern Green River Basin and 21 ft^3/s in the southern part of the basin. Additional recharge of 18 ft^3/s occurs in the vicinity of Farson, Wyo., due to excess surface-water irrigation. Water movement in the Bridger aquifer is primarily horizontal toward Blacks Fork. Most water enters the Laney aquifer by upward leakage from the New Fork aquifer in the central part of the basin. Water enters the New Fork aquifer as recharge adjacent to the Wind River Uplift, moves in a southerly direction, and leaks upward to the Laney aquifer where it subsequently discharges to the Green and Big Sandy Rivers.

Water enters the Wasatch zone of the Wasatch-Fort Union aquifer where the unit crops out, and moves horizontally and vertically toward discharge areas. Short, generally horizontal, flow paths are common in the Wasatch zone. Estimated discharges to the Green River and tributaries upstream from Fontenelle Reservoir are 94 ft^3/s, based on streamflow data. Smaller quantities of water move vertically into the Fort Union zone. Water in the Fort Union zone and lower parts of the Wasatch zone moves along longer flow paths toward Flaming Gorge Reservoir. In the vicinity of the reservoir, water leaks upward through overlying aquifers and confining units and discharges at an estimated rate of 13 ft^3/s, also based on streamflow data.

GREAT DIVIDE-WASHAKIE-SAND WASH BASINS AQUIFER SYSTEM

The Great Divide-Washakie-Sand Wash Basins aquifer system consists of the Wasatch-Fort Union aquifer composed of the Wasatch zone and Fort Union zone—and an overlying confining unit. The Wasatch zone correlates with the Battle Springs Formation and the main body of the Wasatch Formation. The Fort Union zone correlates with the Fort Union Formation. The Green River Formation and tongues of the Wasatch Formation combine to form the confining unit.

The confining unit is present in the western Great Divide, Washakie, and western Sand Wash Basins, with thickness ranging from 3,000 to 5,000 ft. Well yields and spring discharges generally are less than 30 gal/min. Hydraulic-conductivity values estimated from drill-stem tests range from 0.01 to 0.49 ft/d.

The Wasatch zone crops out where no confining unit exists and typically varies in thickness from 2,000 to 4,000 ft. Hydraulic conductivity is related to the quantity of sandstone present in the zone and varies in aquifer-test estimates from 0.03 to 9.1 ft/d. Larger values typically occur in the Great Divide Basin and along basin margins where sandstone is more common. Hydraulic-head and water-quality data indicate that dissolved-solids concentrations in hydraulically upgradient areas generally are less than 500 mg/L, and that concentrations increase down flow paths along the hydraulic gradient.

The Fort Union zone underlies the Wasatch zone and typically is between 3,000 and 6,000 ft thick. Hydraulic conductivity is related to the quantity of sandstone present; estimated values from drill-stem and aquifer tests range from 0.001 to 938 ft/d. The median value for drill-stem tests is 0.02 ft/d.

Recharge to the Great Divide-Washakie-Sand Wash Basin aquifer system occurs along basin margins and discharge occurs in the center of the Great Divide Basin, along Bitter Creek, Separation Creek, and the Little Snake River. A reliable estimate of the amount of water moving throughout the system cannot be made with existing data. Changes in water quality along inferred flow paths have been used to identify probable recharge areas in the northern and western parts of the Great Divide Basin and in the western part of the Washakie Basin.

MESAVERDE AQUIFER

The basin aquifer systems in Tertiary rocks described in this report are directly underlain by the Cretaceous Measverde aquifer. Geohydrologic data are available for the Mesaverde aquifer along the margins of the structural basins, but not available on a regional basis to justify development of hydrologic models or to consider multiple-phase fluid movement of water, oil, and gas. Large reservoirs of natural gas are common within the stratigraphic units that constitute the Mesaverde aquifer. Generally associated with these reservoirs are anomalously small values of hydraulic conductivity and large hydraulic heads. Because of the importance of the oil reservoirs and their effect on geohydrology, research on these anomalies by petroleum geologists and others is ongoing. Regional distribution of hydraulic conductivity ranges from about 1 ft/d in the Rock Springs and Sierra Madre Uplifts to less than 0.01 ft/d where the aquifer is deeply buried and where anomalous hydraulic heads exist.

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