GEOHYDROLOGY OF MESOZOIC ROCKS IN THE UPPER COLORADO RIVER BASIN IN ARIZONA, COLORADO, NEW MEXICO, UTAH, AND WYOMING, EXCLUDING THE SAN JUAN BASIN

REGIONAL AQUIFER-SYSTEM ANALYSIS

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1411-C

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By GEOFFREY W. FREETHEY *and* GAIL E. CORDY

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FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

The Regional Aquifer-System Analysis (RASA) Program was started in 1978 following a congressional mandate to develop quantitative appraisals of the major ground-water systems of the United States. The RASA Program represents a systematic effort to study a number of the Nation's most important aquifer systems, which in aggregate underlie much of the country and which represent an important component of the Nation's total water supply. In general, the boundaries of these studies are identified by the hydrologic extent of each system and accordingly transcend the political subdivisions to which investigations have often arbitrarily been limited in the past. The broad objective for each study is to assemble geologic, hydrologic, and geochemical information, to analyze and develop an understanding of the system, and to develop predictive capabilities that will contribute to the effective management of the system. The use of computer simulation is an important element of the RASA studies, both to develop an understanding of the natural, undisturbed hydrologic system and the changes brought about in it by human activities, and to provide a means of predicting the regional effects of future pumping or other stresses.

The final interpretive results of the RASA Program are presented in a series of U.S. Geological Survey Professional Papers that describe the geology, hydrology, and geochemistry of each regional aquifer system. Each study within the RASA Program is assigned a single Professional Paper number, and where the volume of interpretive material warrants, separate topical chapters that consider the principal elements of the investigation may be published. The series of RASA interpretive reports begins with Professional Paper 1400 and thereafter will continue in numerical sequence as the interpretive products of subsequent studies become available.

Dallas L. Peck
Director

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GLOSSARY

[Definitions are from Freeze and Cherry (1979), Gary and others (1972), Lohman and others (1972), and Pettijohn (1957). Definitions are stated as they apply to this report. All places mentioned in this report are shown on plate 6]

Anticline.—A fold that is convex upward and whose core contains the stratigraphically older rocks.

Aquifer.—A formation, a group of formations, or a part of a formation that contains sufficient saturated permeable material to yield significant quantities of water to wells and springs.

Aquifer test.—A controlled field experiment made to determine the hydraulic properties of water-bearing and associated rocks.

Base level.—The theoretical lowest level toward which erosion of the Earth's surface constantly progresses but seldom, if ever, reaches. Local base level refers to temporary base level in a particular area.

Bentonite.—A plastic, porous rock consisting largely of colloidal silica and composed essentially of clay minerals in the form of extremely minute crystals.

Capillary fringe.—The lower part of the unsaturated zone, just above the water table, that contains water under less than atmospheric pressure. It is continuous with water below the water table but is held above it by surface tension.

Channel-fill deposit.—An alluvial deposit in a stream channel.

Confining layer.—A body of material distinctly less permeable than the aquifer adjacent to it.

Continental deposit.—A sedimentary deposit laid down on land or in a body of water not directly connected with an ocean.

Crossbedding.—An internal arrangement of the layers in a stratified rock, characterized by minor beds inclined more or less regularly in straight, sloping lines or concave forms at various angles to the original depositional surface or the principal bedding plane.

Discharge.—The removal of water from the saturated zone.

Drill-stem test.—A procedure used to determine productivity of an oil or gas well by measuring reservoir pressures and flow capacities while the drill pipe is still in the hole and the well is still full of drilling mud. Aquifer properties are determined using a method described by Bredehoeft (1965).

Dynamic viscosity.—The property that allows fluids to resist relative motion and shear deformation during flow. More commonly called viscosity.

Eolian deposit.—A sedimentary deposit in which grains were transported and laid down by wind.

Ephemeral.—A stream that flows for a very short time in direct response to precipitation and, therefore, whose channel is at all times above the water table.

Epicontinental sea.—Shallow parts of the sea that cover the continental shelf near the edge of a landmass.

Evapotranspiration.—Loss of water from a land area through transpiration of plants and evaporation from the soil.

Facies.—Part of a rock body as differentiated from other parts by appearance or composition.

Fault.—A fracture or fracture zone along which there has been displacement of the sides relative to one another parallel to the fracture.

Flocculate.—The process by which suspended particles are loosely aggregated into clusters.

Fluid potential.—The mechanical energy per unit mass of fluid at any given point in space and time, with respect to an arbitrary state and datum.

Fluvial deposit.—A sedimentary deposit in which material is transported and laid down by a stream.

Friable.—Said of a rock that crumbles naturally or is easily pulverized.

Gaining stream.—A stream or reach of a stream where ground water is discharging into the stream.

Ground water, confined.—Water in an aquifer that is under pressure significantly greater than atmospheric.

Ground water, unconfined.—Water in an aquifer that has a water table.

Hydraulic conductivity.—A measure of the ease with which a fluid will pass through a porous medium, determined by the size, shape, and interconnection of the openings in the material and by the viscosity of the fluid.

Hydraulic gradient.—The rate of change of pressure head per unit of distance of flow at a given point and in a given direction.

Infiltration.—The movement of water from land surface into the zone of aeration.

Interdunal.—Pertaining to the relatively flat surface between sand dunes.

Intermittent stream.—A stream that flows only at certain times during a year.

Interstitial.—Said of a mineral deposit in which the mineral fills the pores of the host rock.

Kinematic viscosity.—The ratio of dynamic viscosity to fluid density.

Kurtosis.—The peakedness or flatness of the graphical representation of a particle-size distribution; thus, a measure of the concentration of sediment particles about the mean diameter.

Lacustrine.—Describing a deposit laid down in a lake.

Lithofacies.—A lateral subdivision of a stratigraphic unit based on a significant change in lithologic character. The change may be abrupt or gradual and (or) physical or chemical. Laterally equivalent lithofacies may be separated by vertical arbitrary-cutoff planes, by intertonguing surfaces, or by gradual changes.

Lithology, lithologic.—The physical character of a rock.

Losing stream.—A stream or reach of a stream that contributes water to the saturated or unsaturated zone.

Marine deposit.—A sedimentary deposit transported and laid down by the action of the water in an ocean.

Millidarcy.—A customary unit of fluid permeability equal to 0.001 darcy. A darcy is equivalent to the passage of 1 cubic centimeter of fluid of 1 centipose viscosity flowing for 1 second under a pressure differential of 1 atmosphere through a porous medium having a cross-sectional area of 1 square centimeter for a distance of 1 centimeter.

Monocline.—A unit of strata that flexes from the horizontal in one direction only.

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.

Normal fault.—A fault in which the overlying block of rock appears to have moved downward relative to the underlying block.

Overburden.—The consolidated and unconsolidated materials that overlie a designated stratigraphic unit.

Peptization.—The process of forming a colloidal solution.

Perennial stream.—A stream that flows continually throughout the year.

Permeability.—A measure of the relative ease with which a porous medium can transmit a liquid under a potential gradient. It is independent of the nature of the liquid and of the force causing movement of the liquid.

Phreatophyte.—A plant that obtains its main water supply from the saturated zone or through the capillary fringe.

Porcelanite.—A hard, dense, siliceous rock having the texture, dull luster, hardness, fracture, or general appearance of unglazed porcelain.

Porosity.—The ratio of the volume of interconnected voids in a rock to the total volume.

Potentiometric surface.—The level to which water rises in a well. This level, generally called hydraulic head, is the sum of the elevation head and the pressure head. Elevation head is a result of the elevation of the point in question above a datum, and pressure head is the height of the column of water that rises above the point in question.

Recharge.—The entry of water into the saturated zone.

Reverse fault.—A fault in which the overlying block of rock appears to have moved upward relative to the underlying block. A thrust fault is a reverse fault with a dip angle of less than 45°.

Riparian vegetation.—Vegetation growing along the banks of a water body.

Rock fabric.—The sum of all the structural and textural features of a rock.

Saturated zone.—Zone of porous medium in which all voids are filled with water.

Skewness.—A measure of the asymmetry of a particle-size distribution.

Sorting.—The degree of uniformity of particle size.

Specific capacity.—The rate of discharge of water from a well divided by the drawdown of the water level in the well, expressed in gallons per minute per foot of drawdown. Aquifer properties are estimated from specific capacity using a method described by Theis and others (1963), and this method is referred to in this report as the "specific-capacity test."

Specific yield.—The ratio of the volume of water in a rock that will drain by gravity to the volume of that rock.

Steady-state conditions.—Refers to a ground-water system in a state of equilibrium. Flow in equals flow out, and storage is constant.

Storage coefficient.—The volume of water that an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in hydraulic head.

Syncline.—A fold that is concave upward, the core containing the stratigraphically younger rocks.

Tectonic.—Pertaining to the origin, historical evolution, and mutual relation of regional structural and deformational features found in the upper part of the Earth's crust.

Terrestrial.—Consisting of or pertaining to land.

Transgressive-regressive cycle.—Cyclic advance and retreat of seawater over a land area.

Transient conditions.—Refers to a ground-water system undergoing some form of external stress that is causing the volume of ground water in storage to change.

Transmissivity.—The rate at which water at the prevailing kinematic viscosity is transmitted through a unit width of the aquifer under a unit hydraulic gradient.

Transpiration.—The process by which water absorbed by plants is discharged into the atmosphere from the plant surface.

Unconformity.—A surface of erosion or nondeposition that separates younger strata from older strata.

Unsaturated zone.—The subsurface zone containing water under pressure less than that of atmospheric, including water held by capillarity.

Water table.—An imaginary surface within an unconfined ground-water reservoir at which the pressure is equal to that of the atmosphere.

METRIC CONVERSION FACTORS

This report uses inch-pound units as the main system for measurement and International System (SI) units for water chemistry, density, grain size, and intrinsic permeability. Units can be converted from one system to another using the following conversion factors. Multiply inch-pound unit by conversion factor to get SI unit. Divide SI unit by conversion factor to get inch-pound unit.

Inch-pound units	*Conversion factor*	*SI units*
acre	0.4047	hectare (h)
acre-foot (acre-ft)	0.001233	cubic hectometer (hm^3)
	1,233.0	cubic meter (m^3)
acre-foot per year (acre-ft/yr)	0.001233	cubic hectometer per year (hm^3/yr)
cubic foot per second per mile [(ft^3/s)/mi]	0.04557	cubic meter per second per kilometer [(m^3/s)/km]
foot (ft)	0.3048	meter (m)
	30.48	centimeter (cm)
foot per day (ft/d)	0.3048	meter per day (m/d)
foot per year (ft/yr)	0.3048	meter per year (m/yr)
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
inch (in)	2.540	centimeter (cm)
	25.40	millimeter (mm)
inch squared (in^2)	6.452	centimeter squared (cm^2)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.590	square kilometer (km^2)

A millidarcy is 0.987×10^{-11} centimeter squared. Temperature in degrees Fahrenheit (°F) can be converted to temperature in degrees Celsius (°C) by using the following equation:

$$°C=5/9(°F-32)$$

The following terms are used in this report to classify water according to the concentration of dissolved solids, in milligrams per liter (mg/L):

Classification	*Concentration of dissolved solids*
Fresh	Less than 1,000
Slightly saline	1,000 to 3,000
Moderately saline	3,000 to 10,000
Very saline	10,000 to 35,000
Briny	More than 35,000

ALTITUDE DATUM

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

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ABSTRACT

Rocks of Mesozoic age in the Upper Colorado River Basin underlie parts of five States west of the Continental Divide—Arizona, Colorado, New Mexico, Utah, and Wyoming. These rocks consist of conglomerate, sandstone, shale, siltstone, claystone, limestone, and evaporites that have been folded, fractured, and faulted by large-scale tectonic activities that created several large structural basins and uplifts. As of 1987, ground-water withdrawals from these rocks throughout the region have been small, except in some localized areas.

The Mesozoic rocks consist of 10 geohydrologic units—5 aquifers separated by 5 confining units. The lowermost (Navajo-Nugget) and uppermost (Mesaverde) aquifers are the thickest; in places, the saturated thickness of each unit is more than 2,000 feet. Each of the three intervening aquifers, the Entrada-Preuss, Morrison, and Dakota aquifers, exceeds 500 feet in saturated thickness in only a few places. Three of the confining units—the Chinle-Moenkopi, Morrison, and Mancos confining units—are more laterally continuous and typically are thicker than the other two—the Carmel-Twin Creek and Curtis-Stump confining units. In places, the Chinle-Moenkopi and Mancos confining units are several thousand feet thick.

Ground-water recharge occurs along the margins of uplifts at higher altitudes, where precipitation is greatest and rocks are exposed. Ground water flows laterally through interconnected pores and fractures in the rock from areas of high to areas of low hydraulic head. Ground water moves vertically between aquifers through confining units in response to differences in hydraulic head in adjacent aquifers. Discharge occurs in the main surface-drainage network from rocks exposed in canyon walls cut by streams. Inflow to and outflow from the Mesozoic ground-water system are estimated to be 1 million acre-feet per year, but uncertainties in the calculation of discharge and recharge rates create a large margin of error. Recoverable ground water in storage of suitable quality for most uses is estimated to be 530 million acre-feet, about 4 percent of the total volume in storage.

Hydrologic properties of the rocks vary laterally because of changing lithofacies within a geologic formation, stratigraphic intertonguing between formations, and erosional pinchouts of formations. The largest value of hydraulic conductivity derived from the results of an aquifer test is 88 feet per day for a 44-foot interval of fractured Navajo Sandstone. As indicated by other aquifer tests, drill-stem tests, laboratory analyses, and specific-capacity tests, hydraulic-conductivity values most commonly range from 0.1 to 10 feet per day in the Navajo-Nugget, Entrada-Preuss, and Morrison aquifers, and from 0.001 to 1.0 foot per day in the Dakota and Mesaverde aquifers. Data indicating the hydrologic properties of the shale, siltstone, and claystone of the confining units are meager, but hydraulic-conductivity values are typically one order of magnitude smaller than for the adjacent aquifers.

Transmissivity values, derived from the product of hydraulic conductivity and saturated thickness, for the Navajo-Nugget and Mesaverde aquifers are more than 2,000 feet squared per day in a few small areas where the thickness of the saturated rock is large. Transmissivity values for the Entrada-Preuss and Morrison aquifers exceed 500 feet squared per day in only a few locations. Values for the Dakota aquifer exceed 100 feet squared per day only locally.

In general, water in the Mesozoic rocks is fresh in the southern half of the study area, where the aquifers are exposed and easily recharged. Water generally is very saline to briny in the northern half, where the aquifers are confined beneath thick overburden and are distant from recharge areas. Sodium chloride water having a dissolved-solids concentration in excess of 35,000 milligrams per liter is present in deep structural basins; calcium bicarbonate water having a dissolved-solids concentration of less than 2,000 milligrams per liter generally is present where aquifers are at shallow depths. Concentrations of iron and manganese generally are large in water from all geohydrologic units, whereas concentrations of other minor constituents are large only locally.

Use of the ground water is limited by deep burial, small transmissivity, and the presence of saline water in many areas. As a result, there has been little development. The southern half of the study area has the largest potential for development of ground-water resources because Mesozoic rocks are generally exposed, saturated thickness is relatively large, and local fracturing near structurally deformed areas increases the potential for recharge and the probability of developing wells having large yields.

INTRODUCTION

The Upper Colorado River Basin contains abundant natural resources. Subsurface resources include coal, oil,

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oil shale, natural gas, uranium, potash, precious metals, and other commercially important minerals. However, the wealth of the region is not only in the subsurface. Each year tens of thousands of people visit the region to enjoy the grandeur of snow-capped mountains, free-flowing streams, and intricately carved canyons. Development of the subsurface resources and nurturing of the tourist industry require water, as do agriculture and livestock grazing, which are also economically important to the region.

In the Colorado River Basin, surface water has been overappropriated, whereas ground water has not been extensively developed. Surface-water supplies have proved inadequate to meet the demands of local and downstream users. Applications to the several regulatory agencies exceed, in gross rate of water claimed, the flow of the Colorado River. Ground water is used to supplement the water supplies for some communities in the region and is the only source of water for other communities. It is used locally for rural domestic purposes and for stock watering. Few industrial developments use ground water.

Although much ground water is available, depth below land surface, transmissive properties of the aquifers, and salinity limit extensive use of ground water compared with surface water. Because of these limitations, planning is needed to ensure that an adequate supply can be obtained without adverse effects on the ground-water system.

Management of the Nation's water supplies can best be accomplished by careful examination of the hydrologic environment. In 1978, the U.S. Geological Survey instituted the Regional Aquifer-System Analysis (RASA) Program to study the important subsurface water resources of the United States. The objectives of the Upper Colorado River Basin RASA (1982) were to (1) classify strata into intervals of aquifers and confining units, (2) quantitatively describe the geometry, hydrology, and geochemistry of these intervals, and (3) analyze the regional ground-water flow systems under steady-state and hypothetical nonsteady-state conditions (Taylor and others, 1983, p. 2).

Responsibilities for accomplishing these objectives for strata (Cambrian to Holocene) in the Upper Colorado River Basin RASA were divided among the Geological Survey offices in Colorado, Utah, and Wyoming. The Paleozoic ground-water systems were studied by the Colorado district, the Mesozoic systems were studied by the Utah district, and the Cenozoic systems were studied by the Wyoming district. This arrangement ensured lateral continuity of study over the project area. The common goal of the three districts was an integrated quantitative assessment of all three ground-water systems.

PURPOSE AND SCOPE

The purpose of this report is to provide a quantitative analysis of the occurrence, movement, and quality of water and the hydrologic characteristics of aquifers and confining units in the Mesozoic rocks of the Upper Colorado River Basin. The analysis is regional in scope and, hence, does not address site-specific problems caused by intricate localized quality, lithologic, or structural discontinuities. The report is intended to answer questions about the lateral flow of ground water from recharge to discharge areas, its vertical movement between aquifer systems, and the general water-yielding properties of aquifers.

Because the investigation was regional in scope, analyses of recharge, ground-water movement, discharge, and storage were based on data and interpretations from the results of previous investigations and existing files from government and private sources. Analyses of hydrologic properties and water quality were based largely on the same sources but were supplemented with a small quantity of newly collected data for areas for which this type of information was lacking.

PHYSICAL SETTING

LOCATION AND EXTENT

The study area is west of the Continental Divide in Arizona, Colorado, New Mexico, Utah, and Wyoming (fig. 1). It includes most of the drainage of the Colorado River upstream from Lee Ferry, an arbitrary point on the Colorado River about 2 miles (mi) downstream from Lees Ferry, Ariz. (All places mentioned in this report are shown on pl. 6.) The Great Divide Basin, an internally drained area in Wyoming, is included in the study area although it is crossed by the Continental Divide. The upper San Juan River basin (including the San Juan structural basin) was excluded from this investigation because it is the focus of a separate RASA study.

The size of the study area is slightly less than 100,000 square miles (mi^2)—about 7,000 mi^2 in Arizona, 34,000 mi^2 in Colorado, 37,000 mi^2 in Utah, and 21,000 mi^2 in Wyoming. A few hundred square miles in the extreme northwestern corner of New Mexico are also included in the study area. The study area extends about 530 mi from 35°46′ N. to 43°27′ N. latitude, and about 350 mi from 105°38′ W. to 112°19′ W. longitude.

PHYSIOGRAPHIC DIVISIONS

The Upper Colorado River Basin is within four physiographic provinces (Fenneman, 1931)—Southern Rocky Mountains, Middle Rocky Mountains, Wyoming Basin, and Colorado Plateaus (fig. 2). The topography in each of these provinces affects the regional ground-water sys-

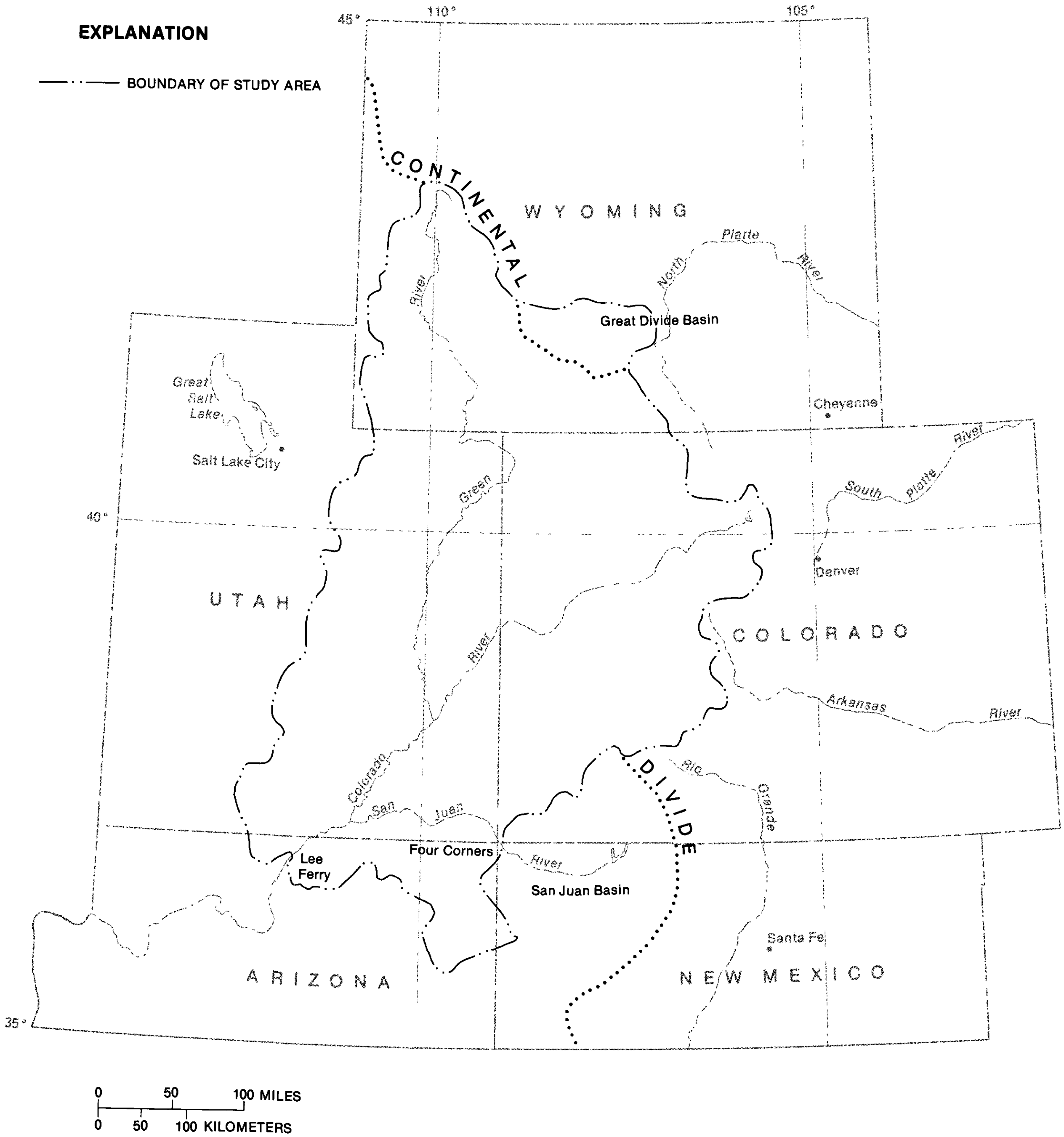


FIGURE 1.—Location of the study area.

tem in different ways. The Southern Rocky Mountains province consists of many mountain ranges with altitudes generally between 8,000 and 14,000 feet (ft).[1] Because of their height, the Southern Rocky Mountains receive a disproportionately large quantity of the annual precipitation in the study area. This precipitation, in the form of rain and snow, is probably the source of ground-water recharge. The Uinta Mountains, in the Middle Rocky Mountains province, trend east-west across the northern part of the basin. The Wyoming Basin province has broad basin floors with altitudes of 6,500 to 7,500 ft that have been interrupted by several structural uplifts and numerous low escarpments; it is drained by the Green

[1]Altitude, as used in this report, refers to distance above or below the National Geodetic Vertical Datum of 1929 (NGVD of 1929).

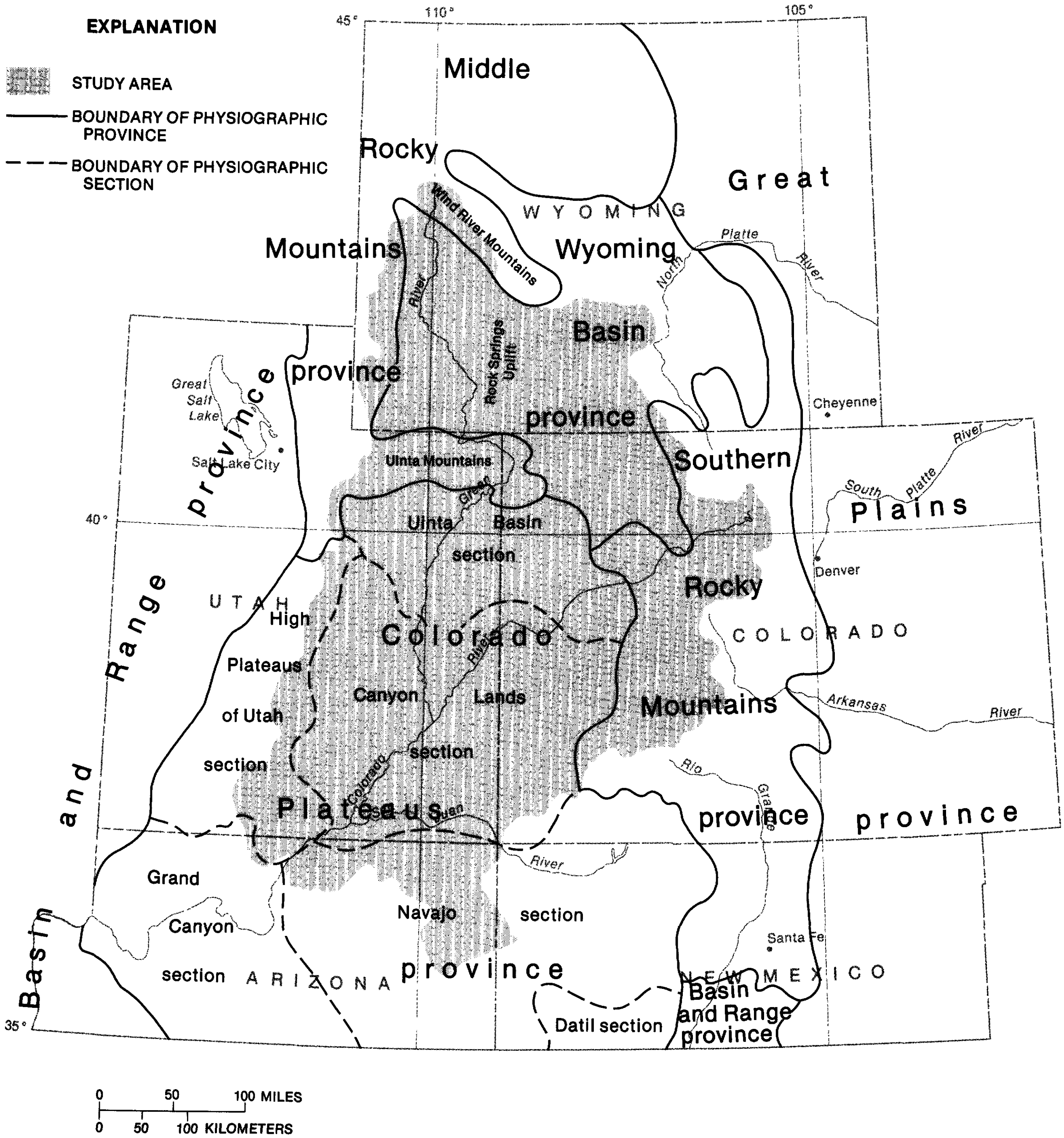


FIGURE 2.—Physiographic divisions of the Upper Colorado River Basin (modified from Fenneman, 1931).

River. Mesozoic aquifers are deeply buried in broad structural basins that compose this province.

The land surface generally is not as high in the Colorado Plateaus province as in the Wyoming Basin province (fig. 3). The Uinta Basin section is similar in geology and in ground-water occurrence to basins in the Wyoming Basin province. Southwest of the Uinta Basin section, the High Plateaus of Utah section forms the western part of the Colorado Plateaus province. The area is extensively faulted and is covered with Tertiary volcanic rocks. Most of the rest of the province (Canyon Lands, Grand Canyon, Navajo, and Datil sections) is

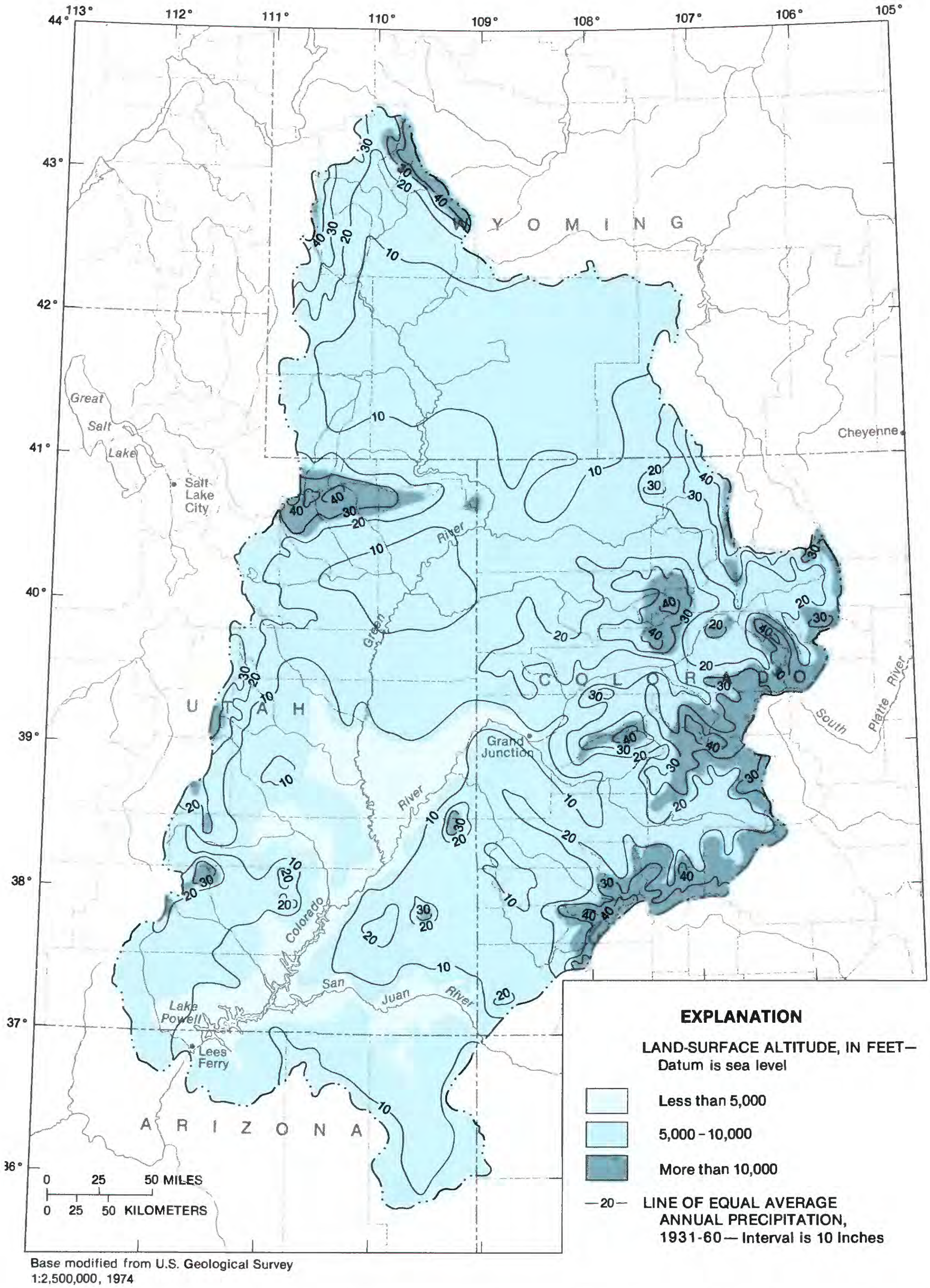


FIGURE 3.—Generalized land-surface altitude and average annual precipitation in the Upper Colorado River Basin (lines of equal precipitation modified from Price and Arnow, 1974, pl. 1B).

dissected by numerous deep canyons that divide the aquifers of Mesozoic rock into subregional systems drained by the streams that incise them.

Landforms that make up each physiographic division affect the areal distribution of recharge and discharge and the occurrence and movement of ground water. They reflect rock type, geologic structure, degree of weathering, and other features, knowledge of which aids understanding of the ground-water flow system. Land-surface altitudes in the study area range from 3,100 ft at Lee Ferry to locally more than 14,000 ft on the Continental Divide. Between these altitudes is a wide range of landforms. About 9 percent of the study area consists of canyon bottoms at altitudes of less than 5,000 ft. These canyon bottoms and other low-lying areas are ground-water discharge areas, indicated by abundant phreatophytes. Sand dunes in parts of the southern deserts of the study area are possible areas of recharge, even though precipitation is small. Hogbacks and reefs indicate areas where erosion-resistant sandstone formations crop out, and these are possible areas of recharge or discharge where traversed by streams. Gentle slopes generally indicate that the exposed formations are easily eroded and have slight permeability. Such formations are composed of shale, siltstone, and claystone, and ground-water movement through them is negligible. Fault zones are areas where surface runoff or precipitation could infiltrate downward into an aquifer. Ground-water flow within an aquifer across or parallel to the line defined by a fault scarp may be significantly different from flow in the surrounding aquifer. Mountains, mesas, and plateaus, because of their altitude, generally indicate recharge areas, but the lithologic character must also be favorable before recharge can occur.

CLIMATE, VEGETATION, AND LAND USE

Average annual precipitation ranges from about 5 to 50 inches (in). It is less than 10 inches in 37 percent of the study area, 10 to 20 inches in 41 percent of the area, 20 to 30 inches in 14 percent of the area, 30 to 40 inches in 7 percent of the area, and 40 to 50 inches in 1 percent of the area. Generally, the higher the altitude, the greater the precipitation (fig. 3). Almost 80 percent of the area receives less than 20 in of precipitation yearly; therefore, the climate of the Upper Colorado River Basin is described as arid to semiarid.

Local topographic control on the movement of airmasses causes temperature and precipitation to vary widely during the year. During fall, winter, and early spring, wet Pacific airmasses bring sustained precipitation to the area. Temperatures are cool, evapotranspiration is minimal, and recharge to aquifers is more likely to take place during this period. During late spring and summer, rainstorms deposit moisture from the Gulf of Mexico on the southern part of the area. Because of the intensity and short duration of the storms, and the higher ambient temperatures during this part of the year, water from these storms quickly runs off or evaporates. Aquifer recharge at this time probably takes place only at higher altitudes, where temperatures remain relatively cool, or where surficial deposits are very permeable.

Vegetative cover is an indicator of the quantity of precipitation, temperature, altitude, and soil cover. It changes from lowland desert shrubs to grassland to mountain forest and then to an alpine biotic community as altitude and precipitation increase and temperature decreases. Other smaller biotic communities grow where hydrologic conditions are favorable. The most prevalent is the dense growth of phreatophytes commonly found along perennial and intermittent stream courses. Phreatophytes affect shallow aquifer systems either by removing water directly from the saturated zone or by consuming infiltrating surface water that would normally recharge the aquifer. These plants also deplete streamflow by consuming ground water that normally discharges to streams, or by lowering the water level below the stream bottom. In contrast, areas covered with crops irrigated by surface water—a human-created biotic community—are recharge areas for aquifers not commonly within reach of crop root systems. Given sufficient time, excess applied irrigation water percolates downward and recharges the aquifer, locally raising water levels and enhancing evapotranspiration downgradient.

People have used the land of the Upper Colorado River Basin in many ways. Uses of the land surface for such purposes as grazing, agriculture, timber production, and recreation have affected the land and vegetation, including the watersheds, but have not appreciably affected the natural occurrence, quality, and movement of ground water. Mining and oil and gas development, however, can locally disrupt the ground-water system and may alter the natural equilibrium in that system. Energy-resource development in the Upper Colorado River Basin likely will continue, and planning for the optimum use and protection of the region's ground water needs to keep pace.

DATA ASSEMBLY AND ANALYSIS

Data used during this study are from many sources. Because the hydrologic data are primarily interpretive rather than measured, the relative accuracy of data from each source is not known. Hydrologic and geologic data points used to compile contour maps were too numerous to show at the map scale used for this report. The data were extracted from computer files and plotted by computer at a scale 2.5 times larger than the scale of plates 1–6. Data values were contoured manually. The data were not qualified except to eliminate obvious inconsistencies. Outcrop areas for the geohydrologic units were generalized from geologic maps of the Upper Colorado River Basin States. Because of the map scale (1:500,000), contacts distinguishing thin units are not shown on these maps. Consequently, outcrop areas of several adjacent geohydrologic units may be shown as a single outcrop. Statisical compilations of aquifer-property, water-quality, and grain-size data were also machine-computed. Two reports, one containing aquifer-property and grain-size data (Weigel, 1987a) and the other containing water-level data (Weigel, 1987b), were the major sources of data. The fence diagram (pl. 4) was compiled from American Stratigraphic Company lithologic logs and data from plates 2 and 3.

To determine water storage in the aquifers, it was necessary to estimate the saturated thickness of those aquifers. To do this, thickness maps were used in conjunction with structure-contour maps defining the top of the Dakota Sandstone to generate maps showing the altitude of the top and bottom of each aquifer. The structure-contour maps were digitized, as were the water-level maps for each aquifer. The water-level and structure-contour maps were then compared by a simple matrix computer program on a grid of 15 minutes of latitude and 15 minutes of longitude (about 17 mi by 14 mi) to determine the total saturated thickness of the respective aquifer.

PREVIOUS INVESTIGATIONS

Although hydrologic studies of all or parts of the Upper Colorado River Basin have been reported, few have addressed the use of the ground-water resources or quantitatively analyzed the aquifers. Probably the first analytical and statistical approach to identifying transmissivities in the exposed rocks of the region was by Jobin (1962). Two reports by Iorns and others (1964, 1965) contain ground-water-quality data for the Upper Colorado River Basin and brief descriptions of the effect of ground-water seepage to streams on the quality of surface water. In 1971, a comprehensive study of the land and water resources of the Upper Colorado River Basin, done by a group of scientists and planners from various government agencies, yielded a qualitative assessment of ground-water availability and an estimate of the volume and distribution of ground water in storage (Hedlund, 1971, p. 20). Price and Waddell (1973) updated Iorns and others' (1964, 1965) hydrologic-data compilation in a map report that shows general availability, depth, and general chemical quality of the ground water. Price and Arnow (1974) also compiled a "summary appraisal" of the ground-water resources of the Upper Colorado River Basin based largely on the interagency work of Hedlund (1971). The report of Price and Arnow contains regionalized descriptions of the occurrence, movement, use, and quality of the ground water and includes estimates of recharge and recoverable water in storage. Ground-water and surface-water resources were again summarized by the U.S. Water Resources Council (1978). This publication contains estimates of projected use of water through the year 2000 and outlines individual problem areas within the region.

Numerous subregional and site studies of ground-water occurrence, aquifer properties, and geohydrology in the Upper Colorado River Basin have been conducted. Some of those studies (several of which are cited later in this report) are, in alphabetical order, Ackerman and Rush (1984), Avery (1986), Blanchard (1986a), Blanchard (1986b), Coffin and others (1971), Cooley and others (1969), Freethey and others (1984), Hood (1976), Hood and others (1976), Lines and Glass (1975), Rush and others (1982), Weir, Maxfield, and Hart (1983), Weir, Maxfield, and Zimmerman (1983), Welder (1968), Welder and McGreevy (1966), and Whitfield and others (1983).

Notable site studies are numerous. Cooperative studies between the U.S. Geological Survey and the Utah Department of Natural Resources have resulted in many publications that focus on the ground-water resources of the Upper Colorado River Basin in Utah. Studies of the Navajo Sandstone in southeastern Utah have generated new interest in this aquifer as a future water supply for that region.

Geologic reports provide knowledge of the geohydrologic framework. The "Geologic Atlas of the Rocky Mountain Region" (Mallory, 1972, p. 166–228) contains a summary of the stratigraphy and structure of the rocks that compose the Mesozoic ground-water system. Hundreds of other publications contain more detailed stratigraphic and lithologic descriptions of the aquifers and confining units of this system, but they are too numerous to mention here.

ACKNOWLEDGMENTS

Many individuals, private companies, and government agencies are acknowledged for their contributions to the hydrologic data of the Upper Colorado River Basin. All data recorded during this study—climatologic, geologic, and hydrologic—are enriching the geohydrologic data base. Syntheses and interpretations of these data enable a better understanding of, and thus better planning for, the allocation and protection of the ground-water resources in the Mesozoic rocks.

REGIONAL GEOHYDROLOGY

Ground water in Mesozoic rocks in the Upper Colorado River Basin is present in numerous sedimentary formations. These formations vary in lithologic and hydrologic character, as determined by depositional environment and by secondary physical and chemical alterations. Deep in the structural basins, the aggregate thickness of these formations exceeds 15,000 ft, yet thicknesses are less than 5,000 ft in most of the study area. Few individual formations retain the same lithologic or hydrologic character regionally. A stratigraphic sequence of formations that exhibits a distinct hydrologic character as a unit can be identified and mapped over a much larger area than can a single formation.

GEOLOGIC SETTING

The Upper Colorado River Basin is divided into several structural basins, uplifts, and platforms (fig. 4). The basins are large, ranging from 25 to 125 mi across. The base of Mesozoic rocks in the Uinta Basin is more than 17,000 ft below sea level (Freethey and others, 1988, fig. 4). Uplifts and platforms that separate individual basins are 6,000 to 12,000 ft above sea level. This structural warping and the associated folding, faulting, and fracturing affect the regional water-flow system in the Mesozoic rocks.

Folds (fig. 5) are more prevalent in western Colorado and southern Utah, where Mesozoic rocks are exposed or are near land surface, than in Wyoming and northeastern Utah, where Mesozoic rocks are buried beneath Tertiary rocks. This apparent difference in density of folding may reflect the lack of data necessary to identify fold structures in these deeply buried Mesozoic rocks. Most major folds have been breached by principal streams, exposing Mesozoic rocks to precipitation and to many miles of streams. These exposures are recharge and discharge areas for the aquifers.

Fractures are common in the folded rocks. Fractures transmit water much more readily than do the connected primary pore spaces in the rock itself, making areas in and near folds and other deformational features conducive to the infiltration of precipitation. The ability of fluids to move through fractures in sedimentary formations decreases with depth because of the increased weight of overlying rocks and the "healing" or closing of fractures with depth.

Faults (fig. 6) are also deformational features that affect the occurrence and movement of ground water. The principal types of faults in the study area that may affect the hydrologic function of the rocks are thrust faults, normal faults, and high-angle reverse faults. Knowledge of the faults and the displacement is necessary to ascertain the continuity of aquifers and confining layers.

Thrust faults, or low-angle reverse faults, are common in the Wyoming thrust belt (fig. 4) and near the boundaries of uplifted areas such as the Uinta and Wind River Uplifts. Normal and high-angle reverse faults can be identified in most of the study area and are most common along the western and southwestern boundaries, in the Paradox Basin, in uplifted areas between basins, and near igneous intrusions. Fault zones may affect ground-water movement locally where the displaced (and fractured) rock has hydrologic properties different from those of the adjacent rock. Depending on rock type and degree of fracturing or recementation in the fault zone, the fault zone could function as either a barrier to or a conduit for ground-water movement.

Many lithologic types are represented in the Mesozoic rocks of the Upper Colorado River Basin. The oldest, of Early and Middle Triassic age, are primarily fine-grained mudstone and shale deposited in a continental shelf environment. These rocks also include less extensive, thin limestone deposits and interbedded lenses of continental sandstone and conglomerate. The Upper Triassic rocks are red shale, sandstone, and conglomerate of continental origin (Mallory, 1972, p. 167). Intermittent eolian deposition began in Late Triassic time and continued during Early Jurassic.

Jurassic sedimentation included four main transgressive-regressive marine cycles culminating with continental deposition. A major subsiding trough at the western margin of the study area, the Utah-Idaho trough, received large thicknesses of sediments throughout the Jurassic Period. Jurassic rocks include several distinct layers of sandstone separated by varying thicknesses of limestone, shale, and mudstone. Lateral facies changes occurred with each transgression and regression of the sea.

Lower Cretaceous deposits consist of fluvial conglomeratic sandstone and mudstone originating in the mountains along the western border of the study area. Subsequently, seas covered the area during most of the Cretaceous Period. Numerous transgressions and regressions of these seas left shoreline-sandstone deposits interfingered with thick sequences of marine shale. Increased tectonic activity, marked by intense overthrust faulting and mountain building to the west and by volcanic activity to the east and south, took place from the Late Jurassic well into the Tertiary.

GEOHYDROLOGIC UNITS

The Mesozoic rocks have previously been grouped into three generalized geohydrologic units (pl. 1) to facilitate

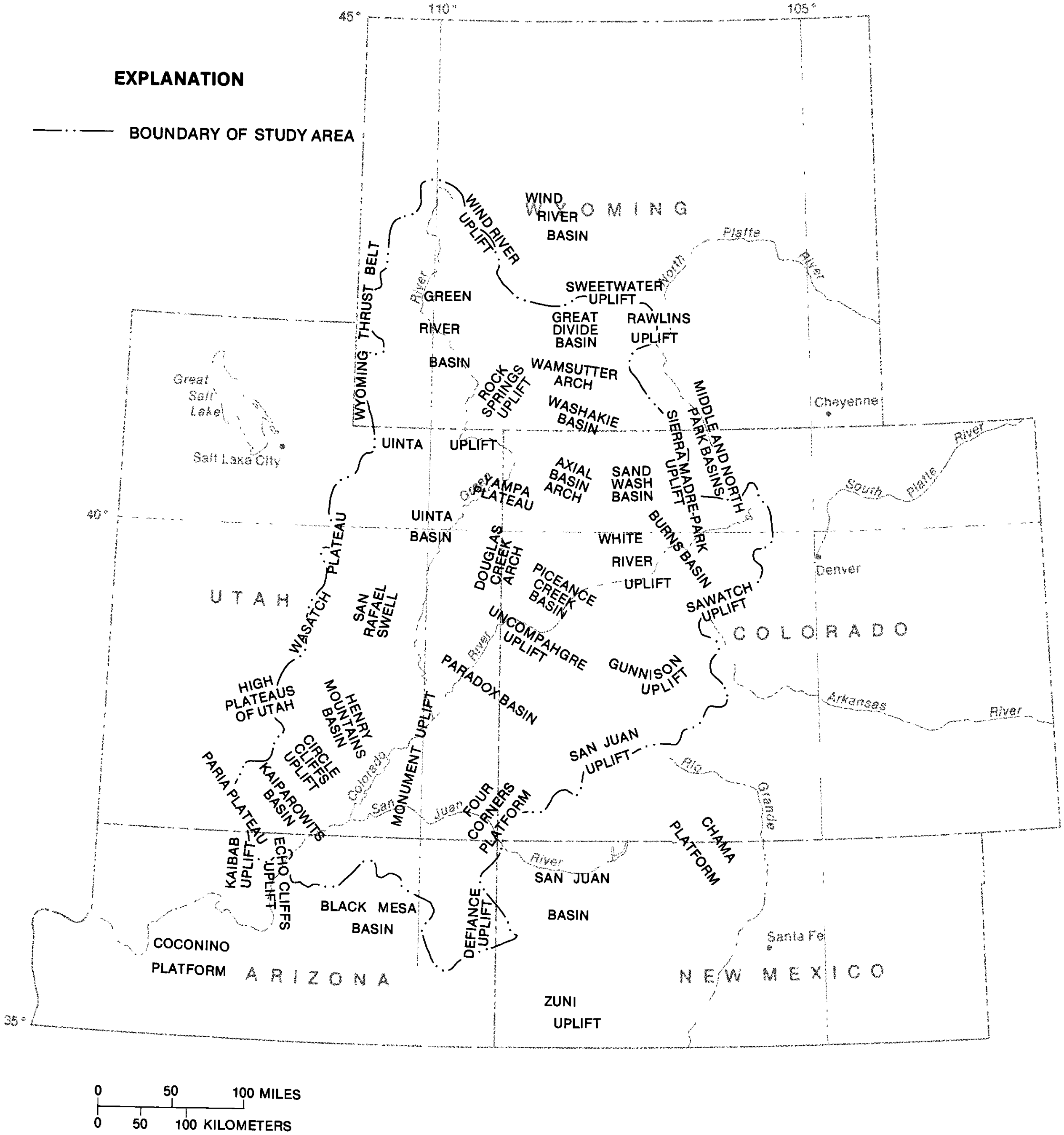


FIGURE 4.—Major structural and topographic features of the Upper Colorado River Basin (from Mallory, 1972, p. 37; and Taylor and others, 1983, p. 10).

analysis and description of this ground-water system (Taylor and others, 1986, sheet 2). Delineation of the three units was based on similarities in lithologic character and on the regional movement of ground water. The lower unit consists of fine-grained Triassic marine and continental deposits; it is considered a confining unit. Upper Triassic and Jurassic rocks form the middle unit; it is considered an aquifer because of the predominance of sandstone. Cretaceous rocks form the upper unit, which consists of a lower aquifer, a middle confining unit, and an upper aquifer.

In this report, the three generalized Mesozoic units are further subdivided into 10 geohydrologic units—5 aquifers and 5 confining units (table 1, pl. 1). The character of

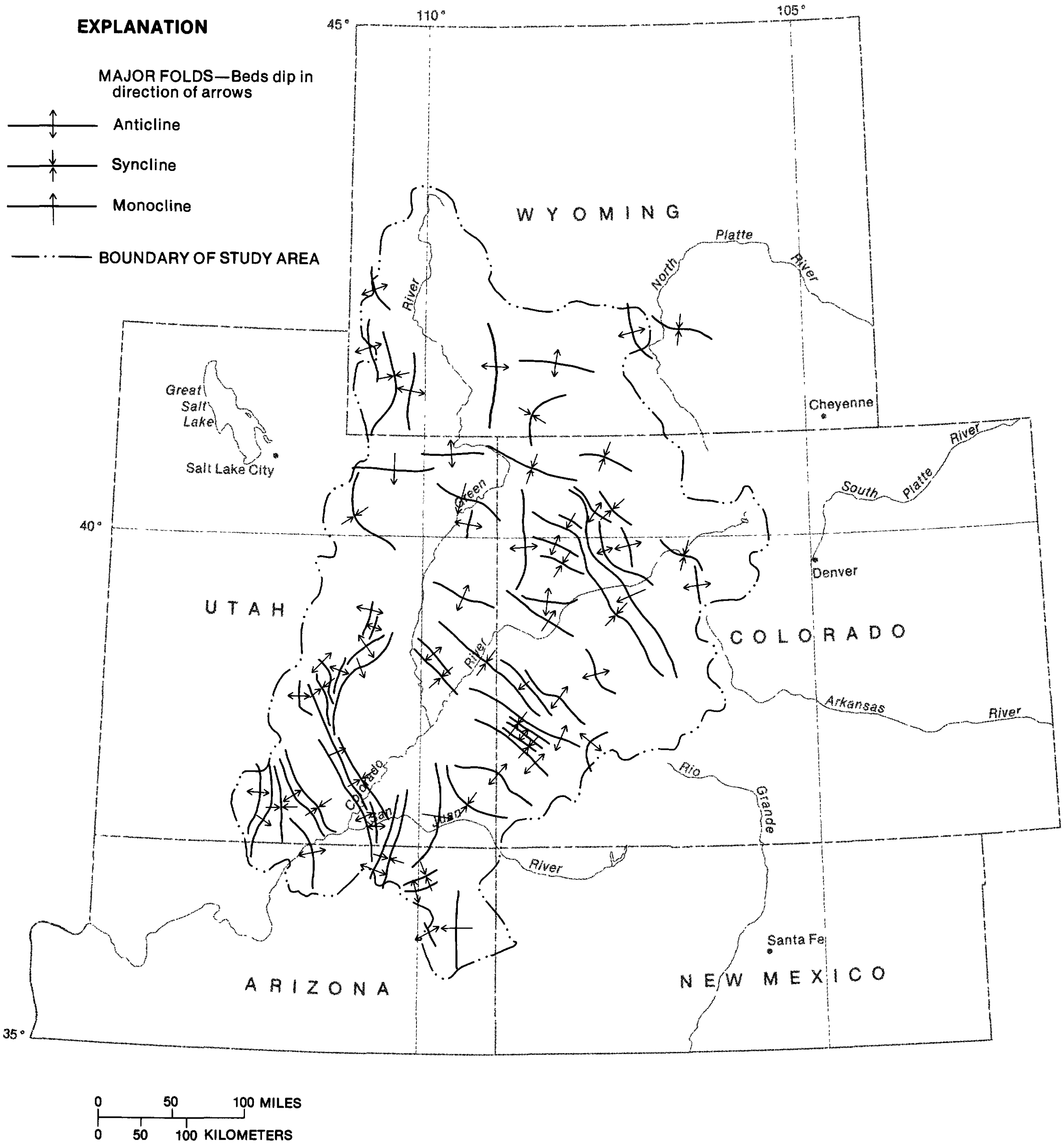


FIGURE 5.—Major folds in Mesozoic rocks of the Upper Colorado River Basin.

these units varies throughout the region, and the designation of a unit as an aquifer or a confining unit depended on its principal function in the regional ground-water flow system. In general, units designated aquifers are composed of sandstone. The number and character of the sandstone beds within one unit vary throughout the region, but the units as a whole are bounded above and below by single layer or multiple layers of rock that have distinctively different hydrologic and geologic properties. Units designated confining units consist principally of shale, siltstone, limestone, and claystone, but they also include interbedded sandstone. Locally, units designated confining units may be aquifers, and parts of units designated aquifers may be confining layers. Because of erosion or nondeposition, certain units are not present in parts of the study area, but all the units are represented

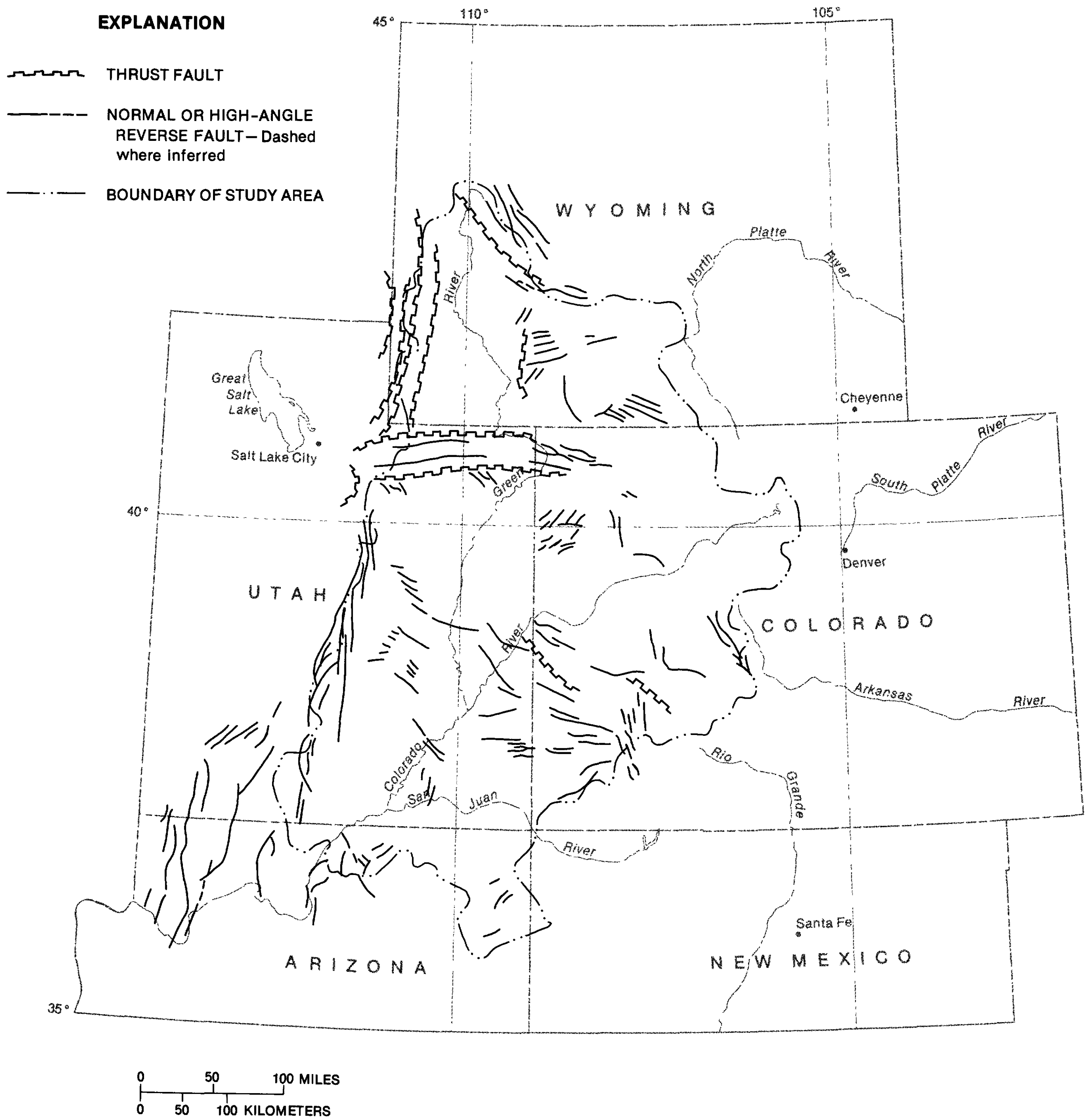


FIGURE 6.—Major faults in Mesozoic rocks of the Upper Colorado River Basin.

in the structural basins that dominate the northern half of the study area. Each of the 10 units, described below, is named for the principal formations that form that unit over most of the study area. The regional nature of these units will be evident to many readers because of their familiarity with some of these names in areas other than the Upper Colorado River Basin.

Understanding the geology of the formations that form the 10 geohydrologic units is paramount to understanding the occurrence and movement of ground water within the Mesozoic rocks. The most important factors are lateral and vertical changes in lithologic character. Negligible changes in an aquifer are indicative of a regional aquifer system rather than many localized aquifers.

TABLE 1.—*Geohydrologic units of Mesozoic age in the Upper Colorado River Basin*

Name	Principal formations represented	Principal lithologic character	Average thickness of unit, in feet (and as percent of total section), at representative sections				
			South-western Wyoming	Northwestern Colorado and northeastern Utah	East-central Utah	South-central Utah and north-central Arizona	Southwestern Colorado
Mesaverde aquifer	Mesaverde Group; Lance, Kaiparowits, Mesaverde, and Adaville Formations; Fox Hills, Wahweap, and Straight Cliffs Sandstones	Sandstone; Shale	4,000 (36)	3,000 (26)	Missing	Missing	Missing
Mancos confining unit	Mancos, Hilliard, Baxter, Aspen, Mowry, Thermopolis, Tropic, Steele, and Cody Shales; Frontier, Niobrara, and Blind Bull Formations	Shale; Sandstone	4,500 (40)	4,000 (37)	2,000 (31)	Missing	Missing
Dakota aquifer	Dakota Sandstone; Cedar Mountain, Bear River, Burro Canyon, and Cloverly Formations; Gannett Group	Sandstone; Conglomerate; Mudstone	200 (2)	200 (2)	300 (5)	Missing	200 (7)
Morrison confining unit	Brushy Basin Member of the Morrison Formation; undifferentiated Morrison Formation	Siltstone; Mudstone; Claystone	Not divided 450 (4)	400 (4)	250 (4)	Missing	300 (11)
Morrison aquifer	Tidwell, Salt Wash, Recapture, Westwater Canyon, and Bluff Sandstone Members of the Morrison Formation; Cow Springs and Junction Creek Sandstones	Sandstone; Conglomeratic sandstone; Siltstone		300 (3)	400 (6)	Missing	450 (16)
Curtis-Stump confining unit	Summerville, Curtis, Stump, Sundance, and Wanakah Formations	Siltstone; Shale; Sandstone	150 (1)	250 (2)	400 (6)	Missing	150 (5)
Entrada-Preuss aquifer	Entrada, Preuss, and Romana Sandstones; Sundance Formation	Sandstone; Siltstone	80 (1)	500 (5)	550 (9)	600 (15)	150 (5)
Carmel-Twin Creek confining unit	Carmel and Gypsum Springs Formations; Twin Creek Limestone	Limestone; Siltstone; Shale	150 (1)	500 (5)	350 (5)	160 (4)	Missing
Navajo-Nugget aquifer	Nugget, Glen Canyon, Navajo, Wingate, and Page Sandstones; Kayenta Formation	Sandstone	500 (5)	750 (7)	1,100 (17)	2,100 (50)	600 (21)
Chinle-Moenkopi confining unit	Moenkopi, Thaynes, State Bridge, Dinwoody, Chinle, Ankareh, and Dolores Formations; Chugwater Group	Siltstone; Claystone; Limestone	1,100 (10)	1,000 (9)	1,100 (17)	1,300 (31)	1,000 (35)

Confining units that have negligible lithologic variability are indicative of minimal water flow between aquifer units. More subtle discontinuities in rock fabric such as unconformities, bedding planes, faults, and joints also affect ground-water movement. Little is known about the hydrologic effect of erosional surfaces in consolidated rocks. The chemical and physical alterations that took place during weathering probably have decreased permeability and would inhibit water movement. Faults and joints may either inhibit or enhance the movement of ground water. The following sections discuss the stratigraphy, lithology, geologic structure, and saturated thickness of the 10 geohydrologic units. Areal extent and thickness maps for each of the geohydrologic units are shown on plates 2 and 3. The relative extent and thickness of the geohydrologic units are shown in a fence diagram (pl. 4).

CHINLE-MOENKOPI CONFINING UNIT

STRATIGRAPHY

The Chinle-Moenkopi confining unit includes the Moenkopi Formation and its equivalents and the Chinle Formation and its equivalents. The areal extent of the Chinle-Moenkopi confining unit is shown on plate 2*A*.

The thickness of the Chinle-Moenkopi confining unit generally increases from east to west (pl. 2*A*). From its eastern margin in west-central Colorado, the unit thickens across western Colorado and eastern Utah to about 2,000 ft in the western Uinta Basin. In Wyoming, the thickness of the confining unit ranges from about 1,000 ft along the eastern border of the study area to 2,500 ft in the Wyoming thrust belt. In northeastern Arizona, the unit thickens in a south-southeasterly direction from 1,200 to 1,700 ft near the southernmost extent of the study area.

The Moenkopi and Chinle Formations form the major parts of this confining unit. Each formation has been divided into several members in various areas, as shown on plate 1.

Individual members of the Moenkopi Formation are limited in areal extent and grade laterally into strata that cannot be differentiated into members (Stewart and others, 1972b, p. 15). Consequently, members in one area do not correlate with those in another area. The Moenkopi Formation as a whole, however, extends throughout the western half of the study area in Utah, Arizona, and northwestern Colorado (fig. 7). Like the Chinle-Moenkopi confining unit, the Moenkopi Formation is thickest along the western margin of the study area. It is about 800 ft thick in Capitol Reef National Park, thinning eastward to zero in eastern Utah and western Colorado (Stewart and others, 1972b, pl. 5).

Lateral equivalents of the Moenkopi Formation include the upper parts of the State Bridge and Goose Egg Formations in northwestern Colorado and the Goose Egg Formation in south-central Wyoming. Because these formations are Permian and Triassic, only the Triassic upper part of each formation is arbitrarily included in the Chinle-Moenkopi confining unit. However, the thickness map (pl. 2*A*) does not reflect the inclusion of the upper parts of these formations because detailed stratigraphic studies needed to define the thicknesses were beyond the scope of this study.

The Moenkopi grades into the Woodside Shale, Thaynes Formation, and Mahogany Member of the Ankareh Formation in the western Uinta Uplift area (Stewart, 1972b, p. 42). In western Wyoming, Moenkopi equivalents include the Dinwoody Formation, overlain by intertonguing Woodside Shale, Thaynes Limestone, and the lower part of the Ankareh Formation, which thin and grade eastward into the Chugwater Group above the Dinwoody Formation (Kummel, 1955, p. 69). In the northeastern corner of the study area, the Dinwoody Formation grades into the upper part of the Goose Egg Formation beneath the Chugwater Group (MacLachlan, 1972, p. 167).

The base of the Moenkopi Formation and equivalents is marked by an unconformity in most areas which represents the boundary between Paleozoic and Mesozoic rocks. However, no unconformity exists in northwestern Colorado and south-central Wyoming, where the upper parts of the State Bridge and Goose Egg Formations are equivalent to the Moenkopi.

The Chinle Formation extends over most of the Colorado Plateaus province (fig. 7), and related strata extend into adjacent areas. In the southern part of the study area, the thickness of the Chinle is generally more than 1,000 ft, with a maximum thickness of about 1,700 ft. It thins irregularly northward, and the thickness ranges from 200 to 500 ft in much of northeastern Utah and northwestern Colorado (Stewart and others, 1972a, p. 1). In western Colorado, the Chinle Formation pinches out along the flanks of the ancestral Uncompahgre and Front Range highlands, in part owing to post-Triassic erosion (MacLachlan, 1972, p. 167).

In southwestern Colorado, the Chinle Formation is laterally continuous with the Dolores Formation. The Ankareh Formation, which is the lateral equivalent of the Chinle Formation in the western Uinta Uplift and Basin areas and western Wyoming, grades eastward into the upper part of the Chugwater Group (Alcova, Jelm, and Popo Agie Formations) in the Rock Springs Uplift of Wyoming (MacLachlan, 1972, p. 167).

An unconformity at the base of the Chinle Formation and equivalent strata marks the contact with the underlying Moenkopi Formation and equivalents throughout the study area (Stewart and others, 1972b, p. 15; Pipiringos and O'Sullivan, 1978, p. A17). However, in the eastern part of the study area where the Moenkopi is absent, the Chinle and Dolores Formations rest unconformably on rocks of Permian age (Stewart and others, 1972a, p. 14).

According to Pipiringos and O'Sullivan (1978, p. A19), an unconformity also marks the top of the Chinle Formation and equivalents throughout much of Arizona, Colorado, Utah, and Wyoming, although locally this surface has not been recognized as an unconformity. Generally, the Chinle and related strata are unconformably overlain by the Glen Canyon Group in the southern part of the study area, the Glen Canyon Sandstone in northwestern Colorado and the eastern Uinta Uplift, and the Nugget Sandstone in the western Uinta Uplift and southwestern Wyoming. Locally, the Twin Creek Limestone in southwestern Wyoming and the Sundance Formation in northwestern Colorado unconformably overlie the Triassic rocks (Imlay, 1980, p. 70, 73).

Lithologic Character

The Moenkopi Formation and equivalent strata were deposited during Early and Middle(?) Triassic time when a broad, long continental shelf occupied the western interior United States, from Canada south to central Arizona. Therefore, the Moenkopi and related strata represent deposits of both continental and marine origin

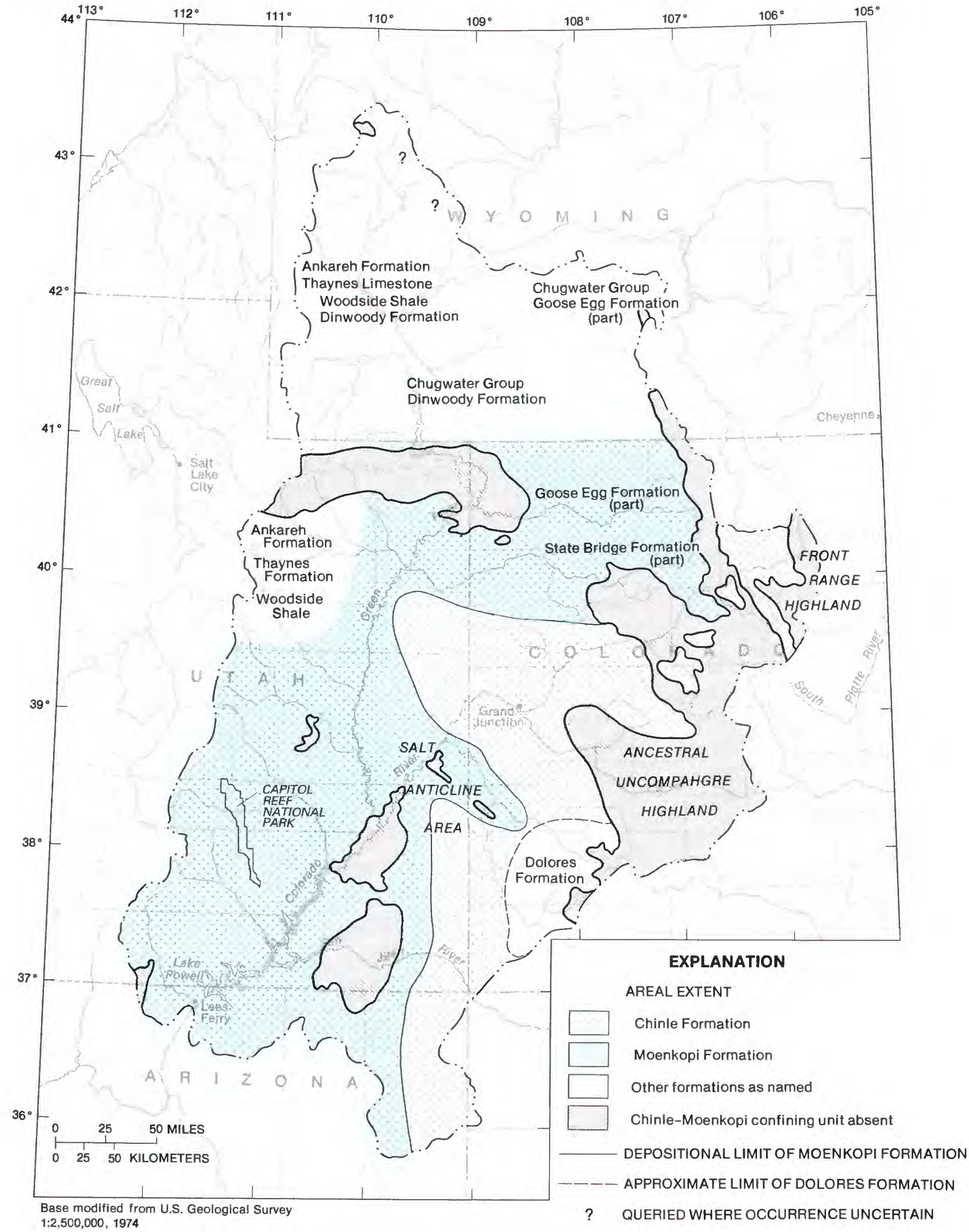


FIGURE 7.—Approximate areal extent of principal formations in the Chinle-Moenkopi confining unit.

(Stewart and others, 1972b, p. 2). Along the eastern margin of the Moenkopi (fig. 7), fluvial sandstone and siltstone grade into marginal marine shale and sandy shale, locally containing interbedded, poorly sorted conglomerate and coarse-grained sandstone. Westward, these marginal marine rocks grade into intertonguing limestone, shale, mudstone, and minor sandstone which represent shallow marine strata transitional to the dark shale and carbonate rocks to the west-northwest of the study area (MacLachlan, 1972, p. 167).

Continental conditions prevailed during the Late Triassic when the Chinle Formation and equivalents were deposited. In contrast to the Moenkopi Formation and related strata, the Upper Triassic red beds are more heterogeneous, poorly sorted, and coarse grained. They include thin limestone of the Alcova Formation, sandstone, siltstone, and shale of the Jelm, Popo Agie, Ankareh, and Dolores Formations, and conglomerate, conglomeratic sandstone, clayey sandstone, and claystone of the Chinle Formation (MacLachlan, 1972, p. 167).

Combined lithofacies of the Moenkopi and Chinle Formations and related strata are shown in figure 8. The main rock types in the Chinle-Moenkopi confining unit are shaly sandstone, sandy shale, and shale. Shale and sandy shale are the dominant lithofacies in most of the area. Shaly sandstone predominates near the Colorado River in Utah and Colorado, in southwestern Colorado, and locally in central Utah. Claystone, sandstone, and conglomerate are major constituents locally, particularly in the Chinle Formation. Because very fine grained strata are characteristic of the Chinle-Moenkopi unit, it is considered a confining unit in the study area.

NAVAJO-NUGGET AQUIFER

STRATIGRAPHY AND SATURATED THICKNESS

The Navajo-Nugget aquifer is made up of the Nugget, Glen Canyon, and Page Sandstones and the Glen Canyon Group, which includes the Wingate Sandstone, the Moenave and Kayenta Formations, and the Navajo Sandstone. The areal distribution and thickness of this aquifer are shown on plate 3*A*.

In southwestern Wyoming and the western Uinta Uplift and Basin, the Nugget Sandstone makes up the aquifer. The thickness of this extensive sandstone is more than 1,000 ft in the Wyoming thrust belt and more than 1,500 ft on the southwestern flank of the Uinta Uplift (pl. 3*A*). South of Wyoming and east of the western Uinta Mountains area, the Nugget Sandstone grades laterally into the equivalent Glen Canyon Sandstone in northeastern Utah and northwestern Colorado (Pipiringos and O'Sullivan, 1978, pl. 1). Southward, the Glen Canyon Sandstone grades into the equivalent Glen Canyon Group (Poole and Stewart, 1964, p. D38).

Formations of the Glen Canyon Group make up the major part of the Navajo-Nugget aquifer in central and southern Utah, northeastern Arizona, and west-central and southwestern Colorado. The Moenave Formation is limited to northeastern Arizona, west of Kayenta (pl. 3*A*), and adjacent parts of Utah (Harshbarger and others, 1957, p. 14). The Wingate Sandstone, Kayenta Formation, and Navajo Sandstone have similar areal extents within the study area, covering central and southern Utah, southwestern Colorado, and northeastern Arizona (Harshbarger and others, 1957, p. 6, 20, 21); however, the Navajo has been removed by Jurassic erosion in most of west-central Colorado (Pipiringos and O'Sullivan, 1978, pl. 1).

The Glen Canyon Group forms a westward-thickening, predominantly sandstone wedge that ranges from zero in northwestern Colorado to about 2,250 ft thick in the southwestern part of the study area (pl. 3*A*). Within the study area, thicknesses of individual formations are as follows: Wingate, 0 to 400 ft (Jobin, 1962, p. 33); Moenave, 125 to 366 ft (Harshbarger and others, 1957, p. 14); Kayenta, 0 to 200 ft; and Navajo, 0 to 1,500 ft (Jobin, 1962, p. 43).

The Page Sandstone is included in the Navajo-Nugget aquifer because of its lithologic similarity and proximity to the underlying Navajo Sandstone. The eastern limit of the Page Sandstone extends from the eastern side of Flaming Gorge Reservoir in Wyoming through Vernal to Green River, Utah, and south through the area just east of Page, Ariz. However, a more recent study by O'Sullivan (1981c) noted an occurrence of the Page Sandstone south of Moab near Kane Springs, Utah, indicating that this unit extends farther east in Utah than previously noted. The western limit is relatively unknown, but the Page Sandstone extends beyond the western border of the study area in southern Utah (Peterson and Pipiringos, 1979, p. B12, fig. 10).

South of Page, Ariz., the Page Sandstone has a maximum thickness of 290 ft (Peterson and Pipiringos, 1979, p. B21). From this area, it pinches out to the east. To the northwest, in south-central Utah, the Page intertongues with the Carmel Formation (Peterson and Pipiringos, 1979, B12, B13). Although the contact between the Page Sandstone and the underlying Navajo Sandstone is a regional unconformity (Peterson and Pipiringos, 1979, p. B21), Blanchard (1986a, p. 12) states that the Navajo and Page Sandstones are hydrologically connected.

Generally, unconformities mark the base and top of the Navajo-Nugget aquifer in the study area. The eastern margin of the aquifer, shown on plate 3*A* in western Colorado and Wyoming, is due to Jurassic erosion which

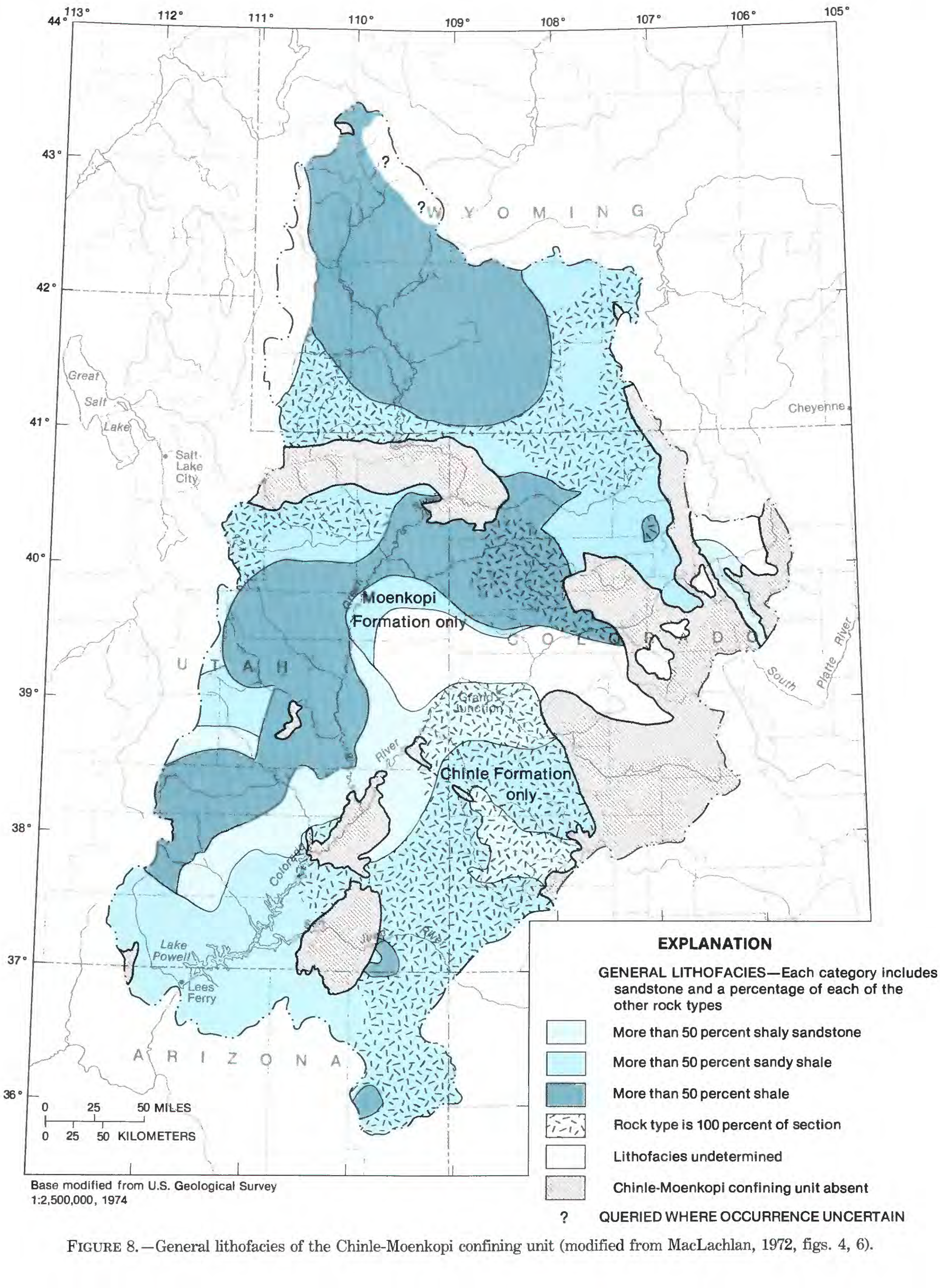


FIGURE 8.—General lithofacies of the Chinle-Moenkopi confining unit (modified from MacLachlan, 1972, figs. 4, 6).

truncated the Glen Canyon Group and the Glen Canyon and Nugget Sandstones (Pipiringos and O'Sullivan, 1978, pl. 1). Similarly, the base of the Navajo-Nugget aquifer is marked by a single major unconformity in most of the area which separates it from the underlying Chinle-Moenkopi confining unit. In northeastern Arizona, this Triassic unconformity lies between the Rock Point Member of the Wingate, an equivalent of the Church Rock Member of the Chinle Formation, and the overlying Lukachukai Member of the Wingate Sandstone (Pipiringos and O'Sullivan, 1978, p. A19). For this study, the silty Rock Point Member is included in the Chinle-Moenkopi confining unit and the sandy Lukachukai Member is included in the Navajo-Nugget aquifer.

The saturated thickness of the Navajo-Nugget aquifer is shown in figure 9. A comparison of the saturated thickness (fig. 9) and the aquifer thickness (pl. 3*A*) indicates that, in most areas, the entire thickness of the aquifer is saturated. Unsaturated zones are present where the aquifer is less than 100 ft thick, as in Colorado and Wyoming (fig. 9). Unsaturated zones are also present on the west side of the Monument Uplift and on the west side of the San Rafael Swell, where less than 100 ft of the more than 1,000 ft of the Navajo-Nugget aquifer are saturated. In these areas, rocks of the Glen Canyon Group are deeply dissected and receive little recharge. The aquifer is only partly saturated along the northwestern flank of the Uinta Uplift, where the rocks dip steeply and receive little recharge because precipitation must percolate through the Tertiary formations that overlie the Navajo-Nugget aquifer.

In most of the western half of the study area, the saturated thickness of the Navajo-Nugget aquifer is 500 to 2,000 ft (fig. 9). The saturated thickness is largest in the southwestern corner of the study area, where the Navajo-Nugget aquifer is thickest.

Although the aquifer contains considerable water in storage, it may not be feasible to withdraw the water through wells. For example, in Wyoming and much of northeastern Utah and northwestern Colorado, the Navajo-Nugget aquifer is overlain by 2,000 ft to more than 12,000 ft of rock, as shown in figure 10. Obtaining water from depths of more than 2,000 ft requires costly drilling, and water from those depths commonly is chemically unsuitable for most uses. Consequently, water in the Navajo-Nugget aquifer is most likely to be developed in the southern half of the study area, where the thickness of the overlying rock is generally less than 2,000 ft and the saturated thickness of the aquifer is more than 500 ft.

Lithologic Character

The Navajo-Nugget aquifer was deposited under arid and terrestrial conditions in the western interior United States. As a result, the rocks are predominantly eolian and fluvial in origin. Sandstone is the major component of the Navajo-Nugget aquifer (fig. 11).

The Moenave Formation is typically fluvial to eolian in origin, consisting of interbedded lenticular sandstone, siltstone, claystone, and minor limestone. The sandstone is very fine to coarse grained, poorly sorted, and crossbedded, and it has a firm calcareous cement (Harshbarger and others, 1957, p. 13–17). The thickness of individual beds within the formation ranges from 1 to 30 ft.

The Kayenta Formation is similar in origin and lithology to the Moenave and consists of fine-grained, crossbedded, lenticular sandstone with interbedded mudstone and interstitial calcareous cement (Harshbarger and others, 1957, p. 18). Although the Kayenta is moderately permeable in most areas, the crosscutting sedimentary structures and mudstone interbeds decrease the permeability in some areas to such a degree that the Kayenta Formation functions as a confining unit (Jobin, 1962, p. 36).

The other formations of the Navajo-Nugget aquifer—the Wingate, Navajo, Page, Nugget, and Glen Canyon Sandstones—are largely of eolian origin with minor fluvial components. They are characteristically massive, very fine to medium-grained, well-sorted, crossbedded sandstone. In outcrops, most of the formations are weakly cemented by calcium carbonate, although silica cement has been noted in the Navajo (Harshbarger and others, 1957, p. 22), Nugget (Picard, 1977, p. 480), and Wingate (Baker and others, 1936, p. 4) Sandstones. In the Navajo Sandstone, carbonate cement is generally a surface phenomenon. J.W. Hood (U.S. Geological Survey, retired, written commun., 1986) states that in cores, the Navajo Sandstone commonly is cemented by secondary silica deposited at grain-to-grain contacts. However, fracture fillings in the Navajo commonly include calcium carbonate, calcium sulfate, and iron oxides. Clay minerals are a cementing agent in the Glen Canyon Sandstone (Poole and Stewart, 1964, p. D38).

The Wingate Sandstone is a crossbedded, well-sorted sandstone of predominantly eolian origin. However, in northeastern Arizona, this formation consists of two members. The lower siltstone and sandstone unit, the Rock Point Member, is equivalent to and continuous with the Church Rock Member of the Chinle Formation in southeastern Utah (O'Sullivan, 1977, p. 141). The upper, fine-grained sandstone, the Lukachukai Member, unconformably overlies the Rock Point Member. Based on lithologic similarities and stratigraphic position, the Rock Point Member is considered part of the Chinle-Moenkopi confining unit, whereas the Lukachukai is included in the Navajo-Nugget aquifer in this study.

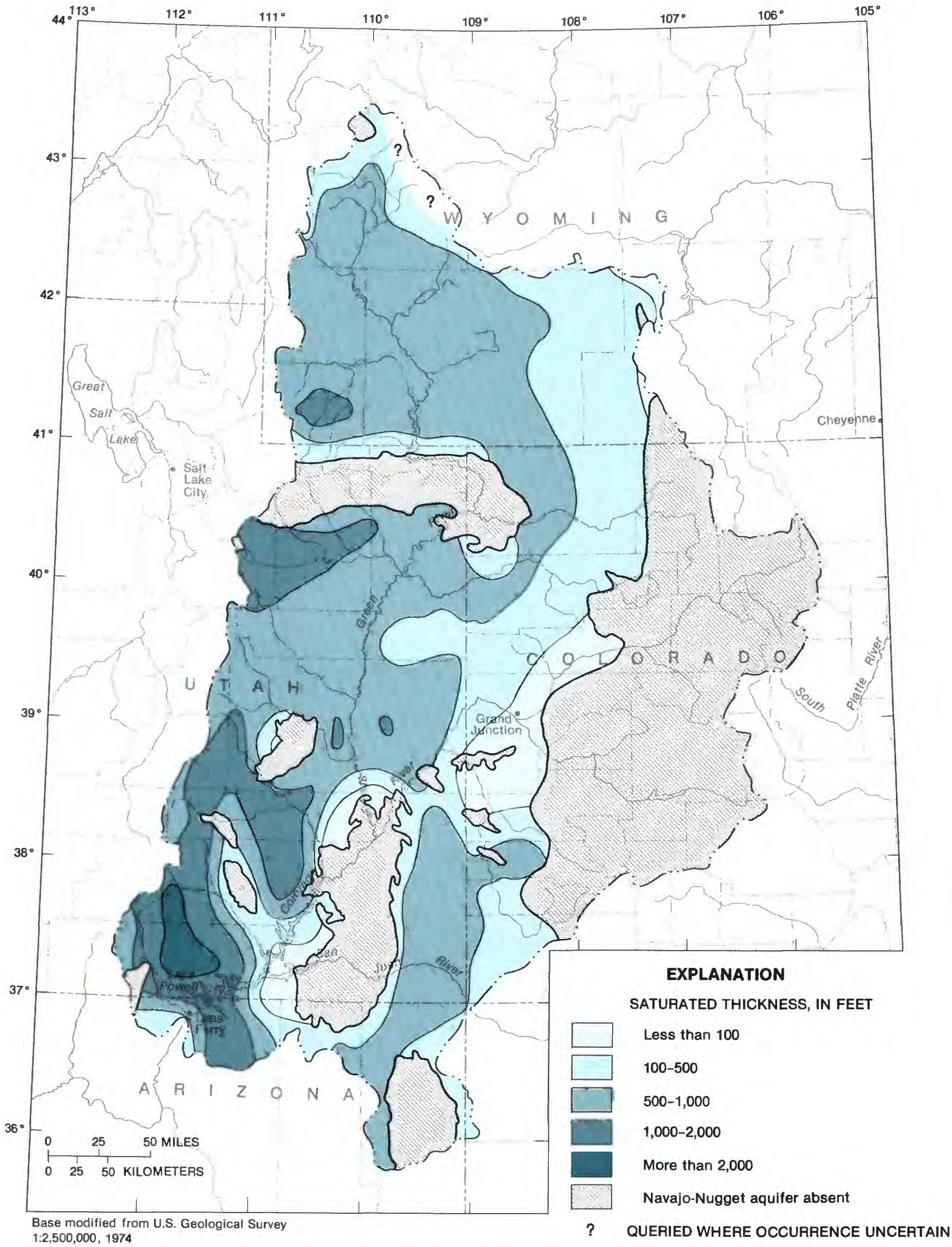


FIGURE 9.—Saturated thickness of the Navajo-Nugget aquifer.

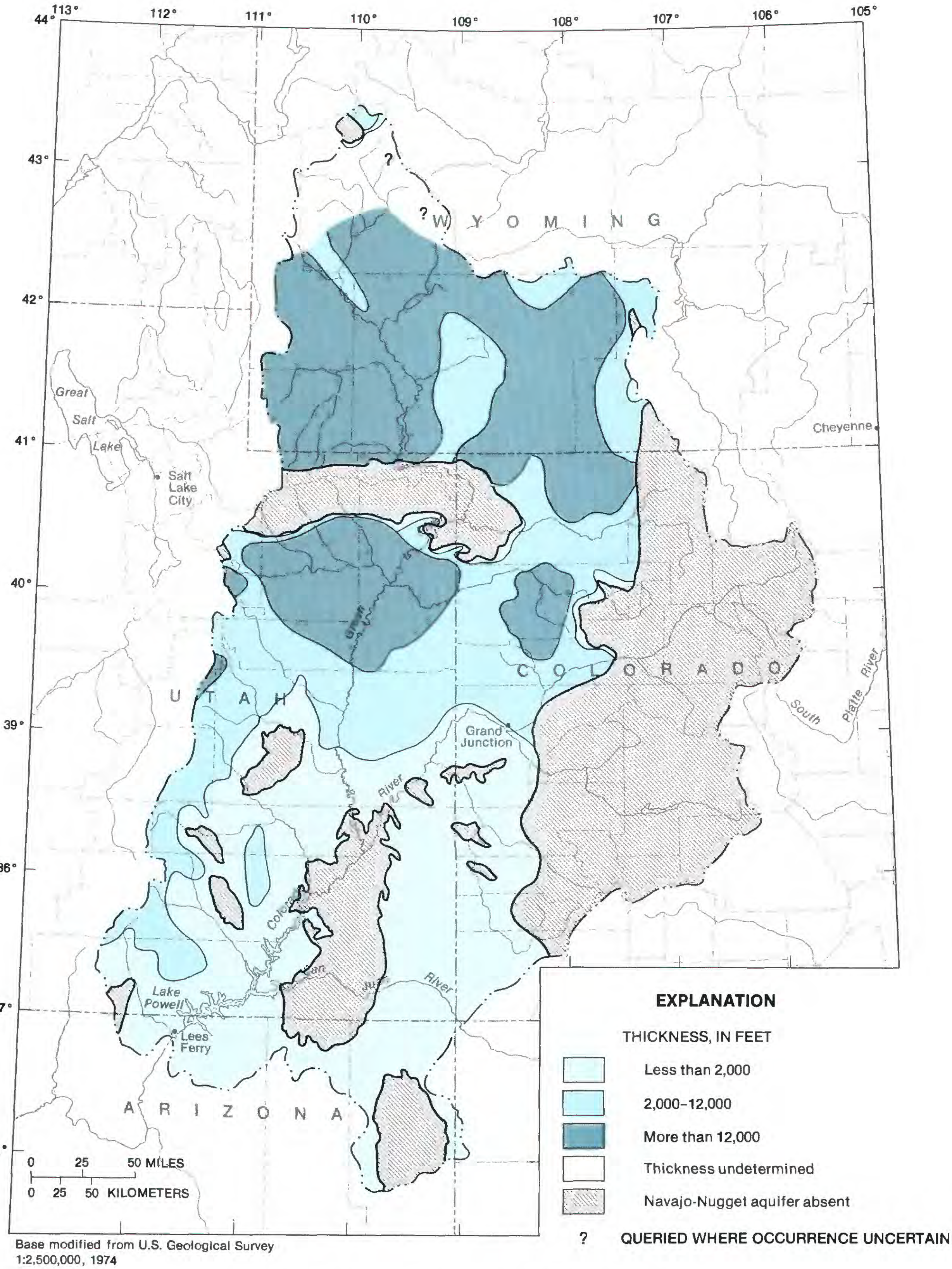


FIGURE 10.—Total thickness of rock overlying the Navajo-Nugget aquifer.

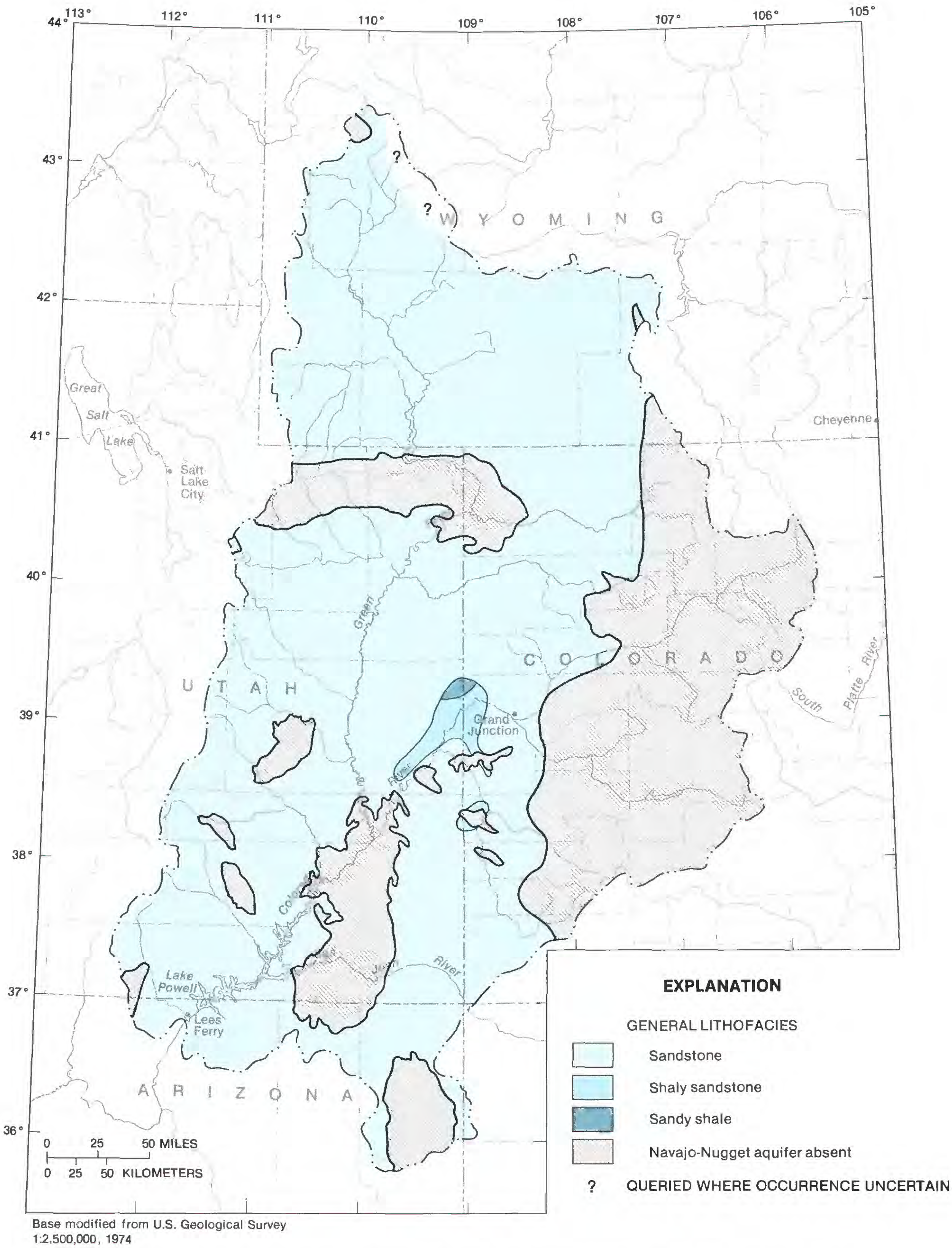


FIGURE 11.—General lithofacies of the Navajo-Nugget aquifer (modified from MacLachlan, 1972, fig. 8).

In a study of the transmissive character of sedimentary rocks of the Colorado Plateau, Jobin (1962, p. 32) states that the Wingate Sandstone has the most uniform permeability of the sandstones. He attributes this characteristic to the relative uniformity of grain size and similarity of interstitial matrix material throughout the formation. Jobin (1962, p. 32) describes the Wingate as "a relatively good transmissive unit" for water over most of the Colorado Plateau based on the thickness and moderately large permeability relative to the other major sandstone formations. However, Jobin analyzed only surface samples in his study, and thus his conclusions may be valid only where the Wingate is at or near the land surface.

The Nugget Sandstone consists of thin-bedded and crossbedded sandstone facies (Picard, 1977, p. 476). The lower, thinly bedded sandstone is very fine grained and exhibits poorer sorting than the crossbedded facies. Consequently, porosity and permeability are smaller in the lower part than in the upper part (Picard, 1977, p. 476). The upper, crossbedded sandstone is thick bedded to massive and exhibits an increase in horizontal and vertical permeability from the base to the top of the sandstone (Uygur and Picard, 1985, p. 27). In the Rawlins Uplift area and north-central Colorado, the Nugget is represented by the Bell Springs Member of the Nugget Sandstone (Pipiringos, 1968, p. D16; Pipiringos, 1972, p. 24), which is similar in lithology to the lower part of the Nugget Sandstone.

The Glen Canyon Sandstone contains varying quantities of flat-bedded strata in the lower 50 to 100 ft (Poole and Stewart, 1964, p. D38). The rest of the formation is thick, crossbedded sandstone similar to the Navajo Sandstone.

The Navajo Sandstone is an important aquifer in the study area. In the northern San Rafael Swell area, the Navajo is very permeable and contains relatively fresh water at a shallow depth (Hood and Patterson, 1984, p. 10). Jobin (1962, p. 42) states that the Navajo Sandstone has the largest transmissivity of the major sandstone strata in the Colorado Plateau because it is thick and well sorted and its permeability is relatively large. Lithologic studies of the Navajo in southern Utah indicate a slight increase in the mean and median grain sizes toward the upper parts of the formation (Uygur, 1980, p. 102). The porosity, hydraulic conductivity, and water content at 100 percent saturation also increase slightly upward; however, cementation decreases slightly upward.

Conspicuous lenses of mudstone, cherty limestone, and dolomite in the Navajo Sandstone account for 2 to 3 percent of the formation in south-central Utah and north-central Arizona (Peterson and Pipiringos, 1979, p. B5). Most lenses are less than 10 ft thick and of limited extent; however, several lenses have been traced 10 to 15 mi in the Circle Cliffs Uplift, northeast of the Kaiparowits Basin. These lenses may function as local impediments to ground-water flow.

A series of regional fractures or joints cuts across the Navajo (Uygur, 1980, p. 15). Where these joints are open, permeability is greatly enhanced. Hood and Patterson (1984, p. 12) note that the permeability of a uniform, planar fracture with an 0.001-in opening is about 132 feet per day (ft/d), 26 times greater than the maximum hydraulic conductivity for unfractured Navajo Sandstone. Conversely, where the joints have been filled by carbonate, iron oxide, or silica, the hydraulic conductivity of the Navajo is smaller (Hood and Patterson, 1984, p. 21).

The Page Sandstone, the youngest formation in the Navajo-Nugget aquifer, had previously been considered part of the Navajo Sandstone. However, recognition of an unconformity at the base of the Page resulted in definition of the Page Sandstone as a separate formation (Peterson and Pipiringos, 1979, p. B20). Lithologically, the Page closely resembles the Navajo Sandstone; however, in the Kaiparowits Basin in the southwestern corner of the study area, a limestone and red-bed tongue of the Carmel Formation splits the Page into two tongues and locally decreases the permeability of the unit.

CARMEL-TWIN CREEK CONFINING UNIT

STRATIGRAPHY

The Carmel-Twin Creek confining unit includes the Carmel Formation in Arizona, Utah, and extreme northwestern Colorado, the Twin Creek Limestone in western Wyoming and adjacent parts of the Uinta Uplift, and the Gypsum Spring Formation in western Wyoming. The Carmel-Twin Creek confining unit is limited to the western part of the study area (pl. 2*B*).

The line of zero thickness on plate 2*B* approximates the ancient shoreline of the Jurassic sea in which the Carmel and equivalent formations were deposited (Wright and Dickey, 1958, p. 174). West of the zero-thickness line, the Carmel-Twin Creek confining unit was deposited in a subsiding trough (Peterson, 1972, p. 178). The confining unit thickens westward to more than 1,400 ft along the western border of the study area in Utah and more than 1,500 ft in western Wyoming (plate 2*B*).

North of the Uinta Uplift, the upper members of the Twin Creek Limestone (Watton Canyon, Leeds Creek, and Giraffe Creek Members shown on pl. 1) grade eastward into equivalent members of the Sundance Formation (Canyon Springs Sandstone, Stockade Beaver Shale, and Hulett Sandstone Members) (Pipiringos and O'Sullivan, 1978, pl. 1; Imlay, 1980, p. 70, 71). For this study, however, the equivalent members of the Sundance Formation have been included in the overlying

Entrada-Preuss aquifer. The Boundary Ridge, Rich, and Sliderock Members of the lower part of the Twin Creek Limestone thin progressively to the northeast, and the Rich and Sliderock Members pinch out in western Wyoming. However, the Boundary Ridge Member grades laterally into the Piper Formation beyond the Wind River Uplift at the northeastern edge of the study area (Pipiringos and O'Sullivan, 1978, pl. 1; Imlay, 1980, p. 70, 71). The basal Gypsum Spring Member of the Twin Creek Limestone grades eastward into the Gypsum Spring Formation (Imlay, 1967, p. 19).

Along the southwestern flank of the Uinta Uplift, the lower five members of the Twin Creek Limestone gradually wedge out eastward from the base up. The two uppermost members of the Twin Creek Limestone (Leeds Creek and Giraffe Creek) closely resemble and are equivalent to the Carmel Formation in northeastern Utah and northwestern Colorado (Imlay, 1980, p. 91).

Southward from the Uinta Uplift, the Carmel Formation is the confining unit in Utah and northeastern Arizona. East of Green River, Utah (pl. 2*B*), the Carmel Formation grades into the Dewey Bridge Member of the Entrada Sandstone along the approximate trend of the line of zero thickness (pl. 2*B*) (Wright and others, 1962, p. 2062). Although the Dewey Bridge Member is similar in lithology to the Carmel Formation in the gradational zone between the two, the Dewey Bridge Member becomes sandy eastward and, therefore, has been included in the overlying Entrada-Preuss aquifer.

The contact of the Carmel-Twin Creek confining unit with the overlying Entrada-Preuss aquifer is conformable, ranging from locally sharp to gradational and interfingering. In general, this contact seems to be gradational in most of the study area. In contrast, the Carmel-Twin Creek confining unit unconformably overlies the Navajo-Nugget aquifer; however, this contact is conformable and gradational where the Carmel either intertongues with or overlies the Page Sandstone, or both, in the southwestern part of the area in Utah and Arizona (Peterson and Pipiringos, 1979, p. B19).

Lithologic Character

The Carmel-Twin Creek confining unit and equivalents were deposited in and on the margins of fluctuating Jurassic seas. Along the western margin of the area, deep-water limestone and shale were deposited in an elongate north-trending trough. Along the margins of the trough in eastern Utah and western Wyoming, red beds, gypsum, and anhydrite were deposited in a marginal marine environment (Hintze, 1982, p. 64).

With the exception of the Gypsum Spring Member, the Twin Creek Limestone is composed of sandy to shaly limestone with interbedded siltstone and minor sandstone. The Gypsum Spring Member is composed of siltstone and claystone, with interbedded brecciated limestone, chert-bearing limestone, and thick masses of gypsum locally (Imlay, 1967, p. 17). The equivalent Gypsum Spring Formation is composed of massive gypsum and anhydrite overlain by an alternating sequence of shale, dolomite, limestone, and thin gypsum beds. The thickness of the gypsum and anhydrite beds is variable, ranging from thin lenses to massive beds 50 to 125 ft thick in western Wyoming (Love, 1945).

The Carmel Formation is diverse in lithology and probably represents lagoonal and estuarine deposits in the east which grade westward into deep-water marine deposits. East of the San Rafael Swell, gray to red siltstone and shale make up about 75 percent of the stratigraphic section, with subordinate limestone, sandstone, and gypsum beds (O'Sullivan, 1981a, p. 90). To the west of the San Rafael Swell, evaporites, including halite, contribute to the deterioration of the chemical quality of both ground and surface waters (Hood and Patterson, 1984, p. 9). Along Comb Ridge (pl. 2*B*) in southeastern Utah, the Carmel grades from crossbedded sandstone in the north to red siltstone and shale near the border with Arizona (O'Sullivan, 1980). In northeastern Arizona and the Kaiparowits Basin of Utah, the Carmel is predominantly siltstone and shale, with lesser amounts of fine-grained, crossbedded sandstone.

ENTRADA-PREUSS AQUIFER

Stratigraphy and Saturated Thickness

The Entrada-Preuss aquifer consists of the Entrada, Preuss, and Romana Sandstones, the Canyon Springs Sandstone, Stockade Beaver Shale, Hulett Sandstone, and Lak Members of the Sundance Formation, and the Cow Springs Sandstone and Cow Springs Sandstone Member of the Entrada Sandstone. The areal extent and thickness of the Entrada-Preuss aquifer are shown on plate 3*B*. Like the Carmel-Twin Creek confining unit, the Entrada-Preuss aquifer was deposited in and adjacent to a subsiding trough trending along the western margin of the study area. Therefore, the Entrada-Preuss aquifer is thickest in the west, where the trough was deepest, and thins eastward.

The Entrada Sandstone is widespread throughout Utah, Arizona, and western Colorado. It generally increases in thickness from east to west, ranging from about 100 ft in western Colorado to 1,100 ft at the western border of the study area, northwest of the San Rafael Swell (pl. 3*B*). West of the Kaiparowits Basin, the Entrada Sandstone thins and has been truncated beyond the study area by erosion prior to the deposition of Cretaceous rocks (Pipiringos and O'Sullivan, 1978, pl. 1). The formation extends well beyond the eastern border of

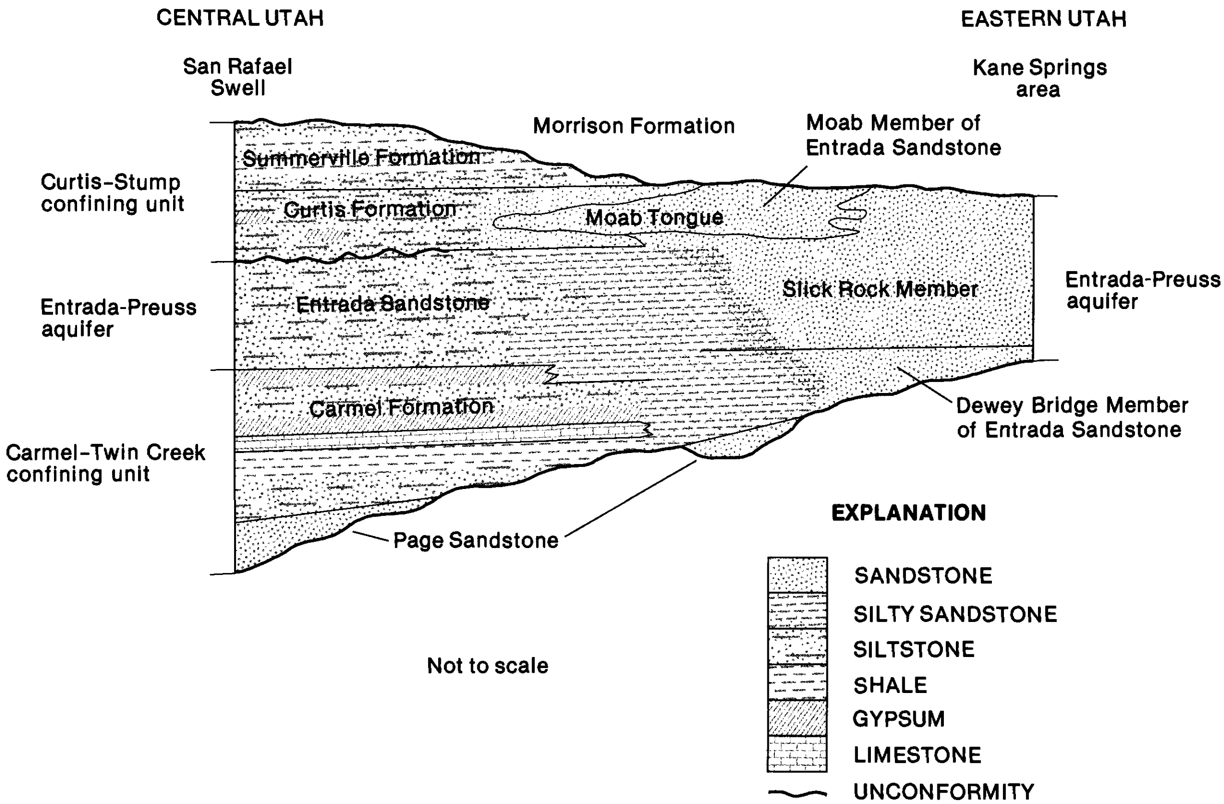


FIGURE 12.—Schematic section showing the stratigraphic relations between the Entrada Sandstone and adjacent formations and geohydrologic units (modified from O'Sullivan, 1981a, fig. 1).

the study area into central Colorado. The Entrada Sandstone in Utah grades into the Preuss Sandstone in the western Uinta Uplift and Wyoming thrust belt and the Lak Member of the Sundance Formation in western Wyoming (Imlay, 1952, p. 1747). In northwestern Colorado and southern Wyoming, the Entrada grades into the Canyon Springs Sandstone Member of the Sundance Formation (Pipiringos, 1972, p. 27).

East of the Green River in east-central Utah, the Entrada Sandstone is divided into three members. In ascending order, they are the Dewey Bridge, Slick Rock, and Moab Members (Wright and others, 1962). The stratigraphic relations among these members and the overlying and underlying formations are complex (fig. 12).

West of the Green River, the Moab Member becomes the Moab Tongue, which interfingers with the Curtis Formation (Curtis-Stump confining unit). The Slick Rock Member grades into the earthy facies of the Entrada, and the Dewey Bridge Member grades into the Carmel Formation (O'Sullivan, 1981a, p. 89). Locally, the Dewey Bridge Member conformably overlies the Page Sandstone; however, in most areas, the base of the Entrada is marked by an erosional surface which truncates the underlying Navajo Sandstone east of the Green River. Westward, the Entrada conformably overlies the Carmel Formation (Wright and others, 1962, p. 2058). The contact of the Entrada with the overlying formations is conformable in most of the area, although locally the contact is unconformable (pl. 1).

The Cow Springs Sandstone Member of the Entrada in northeastern Arizona and the Cow Springs Sandstone (formation) in extreme northeastern Arizona and northwestern New Mexico (not shown on pl. 1) are included in the Entrada-Preuss aquifer unit. On the western side of Black Mesa Basin, the Cow Springs Sandstone Member is a bleached zone at the top of the red Entrada Sandstone (O'Sullivan, 1978). This member grades to the east into the Cow Springs Sandstone, which overlies the Entrada Sandstone. The Cow Springs Sandstone is about 100 ft thick in northeastern Arizona (O'Sullivan, 1978). Locally, in extreme northeastern Arizona and northwestern New Mexico, the Summerville Formation intertongues beneath the Cow Springs Sandstone and isolates it from the Entrada-Preuss aquifer (O'Sullivan, 1978; Condon and Huffman, 1984, p. 100). In these areas, the Cow Springs Sandstone is included with the overlying sandstones of the Morrison Formation in the Morrison aquifer.

The Preuss Sandstone extends from the Wyoming thrust belt south into the western Uinta Uplift. The thickness of the Preuss increases markedly from east to west, ranging from about 200 to 300 ft at the eastern edge of the Wyoming thrust belt to more than 1,100 ft along the southwestern flank of the Uinta Uplift (pl. 3*B*). From western Wyoming, the Preuss Sandstone grades eastward into the Lak Member of the Sundance Formation (Imlay, 1952, p. 1738, fig. 3). The Preuss grades laterally into the Entrada Sandstone in northeastern Utah (Imlay, 1952, p. 1736). The contact of the Preuss Sandstone with the overlying Stump Formation is a sharp and distinct lithologic change. Conversely, the basal contact exhibits a gradual change from red sandstone of the Preuss into the underlying gray, calcareous Twin Creek Limestone.

The inclusion of the Canyon Springs Sandstone, Stockade Beaver Shale, Hulett Sandstone, and Lak Members of the Sundance Formation in the Entrada-Preuss aquifer is based on gradational relations between these members and the Entrada Sandstone. Pipiringos (1972, p. 26) showed that in northwestern Colorado, the Entrada and the equivalent Canyon Springs Sandstone Member of the Sundance Formation extend northward into Wyoming, where they interfinger with the Hulett Sandstone and Lak Members of the Sundance. Farther to the north in Wyoming, the Stockade Beaver Shale Member of the Sundance lies between the Canyon Springs Sandstone and Hulett Sandstone Members and may function as a confining unit within this sequence of sandstones.

In the study area, the total thickness of these four members of the Sundance Formation rarely exceeds 100 ft; however, a section composed of the Canyon Springs Sandstone and Lak Members has been measured at 213 ft thick along the eastern flank of the Park Range in Colorado (Pipiringos and others, 1969, p. N32, N33).

The relations of the lower Sundance members (Canyon Springs Sandstone, Stockade Beaver Shale, Hulett Sandstone, and Lak Members) with the overlying Pine Butte Member of the Sundance are complex owing to gradational changes and intertonguing. The Pine Butte Member conformably overlies and grades downward into the Lak Member of the Sundance Formation in Wyoming and northwestern Colorado. In north-central Colorado, the Pine Butte Member grades down into the Canyon Springs Sandstone Member (Pipiringos, 1972, p. 27). Generally, the Canyon Springs Sandstone Member, the basal unit of the Entrada-Preuss aquifer, lies above rocks that are truncated by a widespread Jurassic unconformity, and the basal contact is sharp.

The Romana Sandstone is limited in lateral extent to the southern Kaiparowits Basin and adjacent parts of Arizona, where it is as much as 150 ft thick (Peterson, 1973). Though stratigraphically above the other formations of the Entrada-Preuss aquifer, the Romana Sandstone is included in this aquifer because of its sandstone lithology. The Romana Sandstone unconformably overlies the Entrada Sandstone and is unconformably overlain by the Tidwell Member of the Morrison Formation (pl. 1).

The saturated thickness of the Entrada-Preuss aquifer is shown in figure 13. In general, water levels indicate that the entire thickness of the Entrada-Preuss aquifer is saturated; however, unsaturated zones are present where the aquifer is less than 100 ft thick, the rocks of the Entrada-Preuss aquifer dip steeply, or it is cut by deep canyons.

The extensive saturated thickness of the Entrada-Preuss aquifer indicates that a large volume of water is stored in the aquifer. However, figure 14 indicates that in more than 50 percent of the study area where the saturated thickness of the aquifer is more than 100 ft, the thickness of the overlying rock exceeds 12,000 ft. The costs to withdraw water from these depths would be large, and the water would be chemically unsuitable for most uses. Moreover, in much of the southern part of the study area where the thickness of the overlying rock is less than 2,000 ft and ground-water resources could be developed, the saturated thickness is commonly less than 100 ft.

Lithologic Character

At the time the Entrada-Preuss aquifer and equivalents were being deposited, fluctuating Jurassic seas surrounded a large island in northern Wyoming and west-central Montana (Imlay, 1952, p. 1735). Connection with marine waters to the north in Canada was restricted by the island, resulting in the development of saline lagoons to the southwest in western Wyoming and central Utah. Marginal marine to continental conditions prevailed to the south and east in Wyoming, Colorado, eastern Utah, and northeastern Arizona. This paleogeography is reflected in the lithology of the Entrada-Preuss aquifer (fig. 15).

The Entrada Sandstone grades from crossbedded eolian sandstone in the east to marginal marine earthy (muddy) sandstone and siltstone westward. In the San Rafael Swell, the earthy facies of the Entrada is a dark-red, fine-grained earthy sandstone. Eastward it becomes a less earthy, irregularly bedded sandstone which grades farther east into a clean, fine- to medium-grained sandstone (Baker and others, 1936, p. 7). East of the Green River, three members of the Entrada are

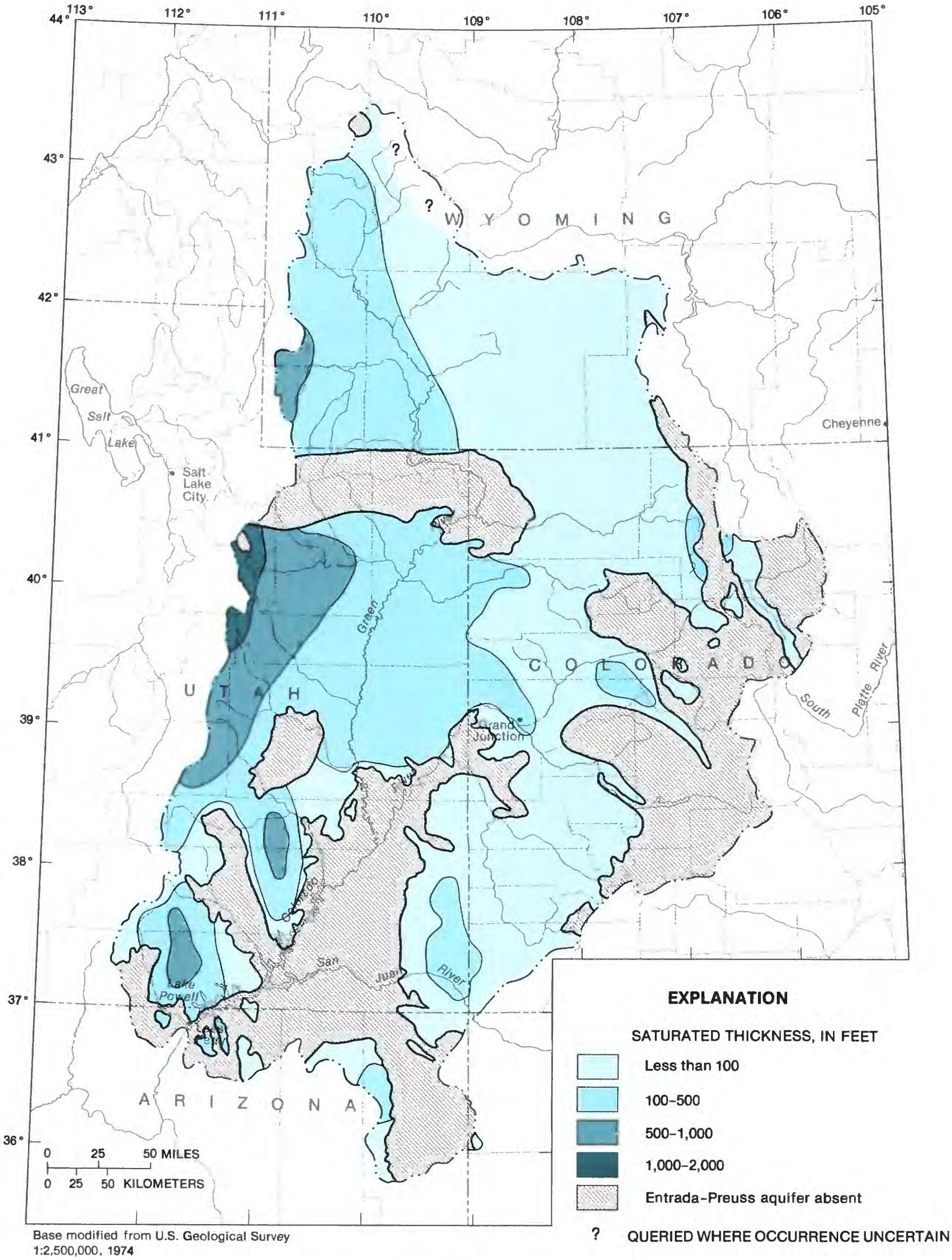


FIGURE 13.—Saturated thickness of the Entrada-Preuss aquifer.

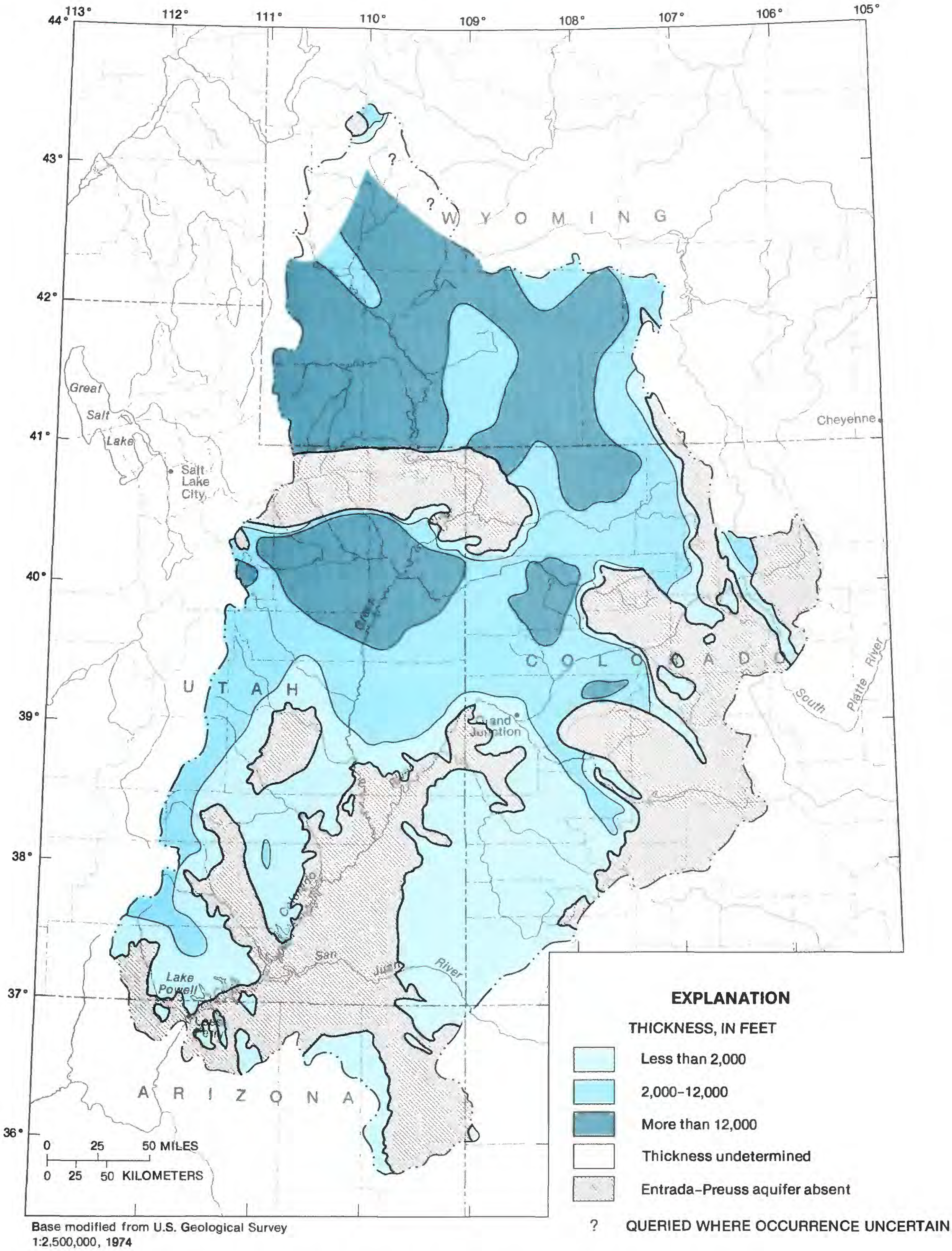


FIGURE 14.—Total thickness of rock overlying the Entrada-Preuss aquifer.

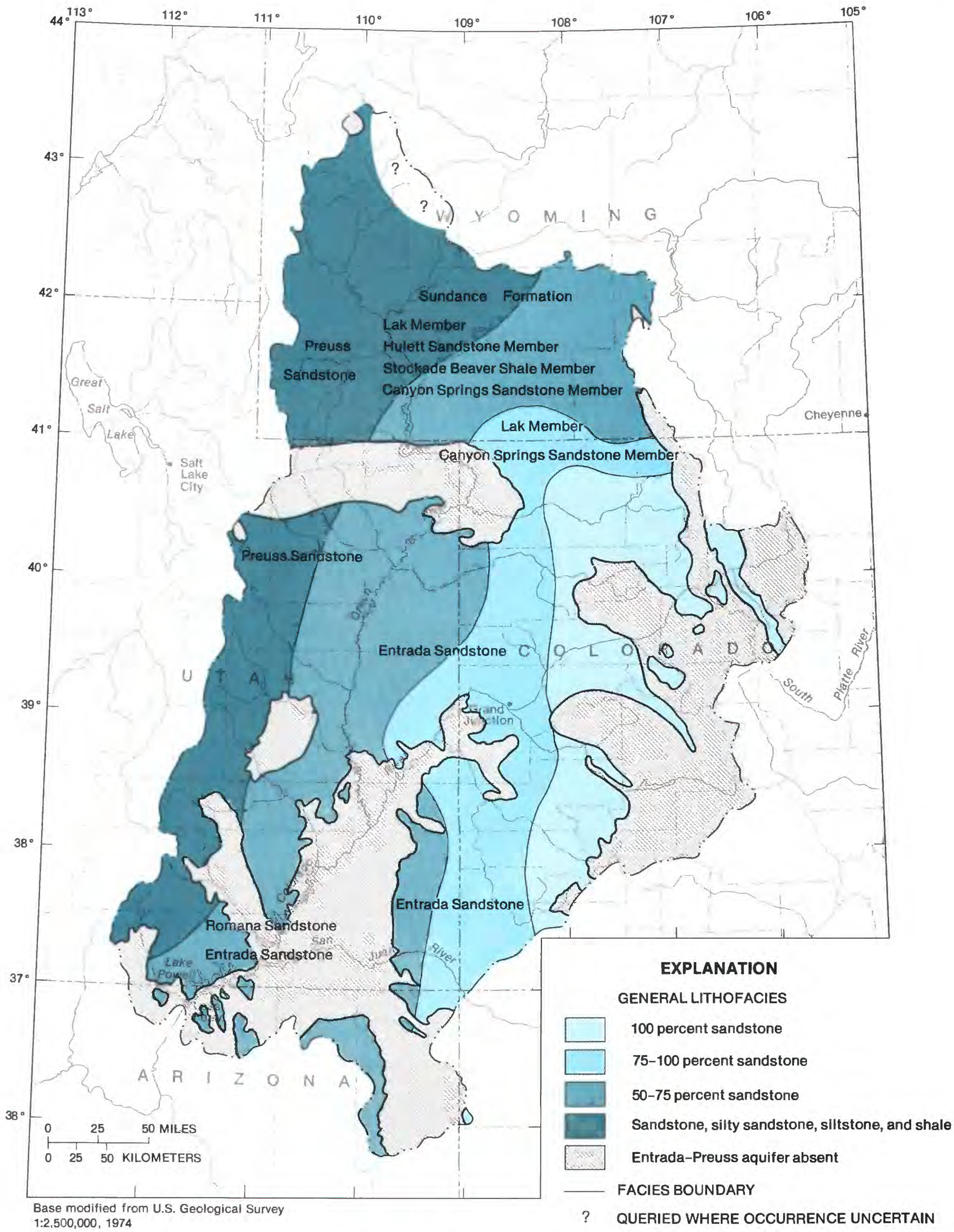


FIGURE 15.—General lithofacies of the Entrada-Preuss aquifer (modified from Peterson, 1972, fig. 6).

recognized. The basal Dewey Bridge Member is predominantly a dark-red, moderately well sorted, very fine grained silty sandstone, which closely resembles the earthy facies of the Entrada farther west (O'Sullivan, 1981a, p. 92).

The Slick Rock Member, which accounts for 50 to 80 percent of the total thickness of the Entrada, consists of light buff to reddish-brown, very fine to fine-grained, crossbedded and massive flat-bedded sandstone of eolian and interdunal origin. Typically, horizontal bedding planes separate the crossbedded and flat-bedded sandstones. At four measured sections near Moab, Utah, 140 individual beds averaging 6 to 7 ft thick were noted in outcrops of the Slick Rock Member (O'Sullivan, 1981a, p. 92). The sandstone is moderately well sorted and generally weakly cemented; however, clay and calcite locally form a firm cement.

The Moab Member forms a single bed of white, crossbedded, eolian sandstone. It is very fine to fine grained and well sorted, and it generally contains less than 1 percent clay (Wright and others, 1962, p. 2067). Similarly, the Cow Springs Sandstone Member of the Entrada Sandstone and the Cow Springs Sandstone are crossbedded to flat-bedded, well-sorted, fine- to medium-grained sandstone (Jobin, 1962, p. 52; O'Sullivan, 1978).

Grain-size-distribution analyses by Wright and others (1962) indicate that the Dewey Bridge Member is approximately equal parts of sand and silt, whereas the Slick Rock and Moab Members are about 90 percent sand, with the rest being silt and clay. Compositional analyses of the three members indicate that clay and calcite form the matrix, which accounts for about 8 percent of the Moab Member, 10 percent of the Slick Rock Member, and 13 percent of the Dewey Bridge Member.

The Preuss Sandstone consists of red, fine-grained sandstone which includes red, sandy siltstone and thin, green siltstone beds. Bedded salt and thin limestone beds are present in the lower part of the Preuss near the Wyoming-Idaho border. The Preuss is well sorted and flat bedded, and it exhibits oscillation ripple marks. This sandstone becomes siltier, softer, and darker red to the south and west (Imlay, 1952, p. 1739).

The Romana Sandstone is mainly yellowish-gray, very fine to fine-grained, moderately sorted sandstone (Peterson, 1973). It is thin to thick bedded, with thick crossbeds. Flat-bedded siltstone is common in the lower part of the formation.

The four members of the Sundance Formation that are included in the Entrada-Preuss aquifer consist of interbedded sandstone, siltstone, and shale. The Canyon Springs Sandstone Member is white to pink, very fine grained, massive to obscurely crossbedded sandstone with interbedded clayey sandstone and thin shale beds. Siltstone and shale are the main components of the Stockade Beaver Shale Member, whereas the Hulett Sandstone Member has equal amounts of sandstone and shale. The clayey sandstone to sandy siltstone of the Lak Member also include interbedded clay, shale, and claystone (Pipiringos and others, 1969, p. N12).

CURTIS-STUMP CONFINING UNIT

STRATIGRAPHY

The Curtis-Stump confining unit is represented by the Wanakah, Curtis, and Summerville Formations, the Curtis and Redwater Members of the Stump Formation, and the Pine Butte, Redwater Shale, and Windy Hill Sandstone Members of the Sundance Formation. The areal extent and thickness of the Curtis-Stump confining unit are shown on plate 2*C*. The formations that form the Curtis-Stump confining unit are not laterally continuous throughout the study area. They occur as five distinct areas separated by zones where the Curtis-Stump and equivalent units were not deposited or have been eroded away.

Recognition of several unconformities in Jurassic rocks (Pipiringos and O'Sullivan, 1978) has led to redefinition of the areal extent and lateral continuity of several formations. In particular, the extent of the Wanakah and Summerville Formations has been modified (R.B. O'Sullivan, U.S. Geological Survey, written commun., 1985).

The Wanakah Formation extends from southeastern Utah into adjacent parts of southwestern Colorado (fig. 16). Based on available data, the Wanakah appears to vary in thickness from less than 100 ft in southeastern Utah and southwestern Colorado to about 300 ft south of the Gunnison River in southwestern Colorado (pl. 2*C*). To the north and northeast, the boundary of the Wanakah is shown by an arbitrary line (R.B. O'Sullivan, U.S. Geological Survey, oral commun., 1985) indicating intertonguing and lateral gradation of the Wanakah into the upper part of the Entrada Sandstone.

The Curtis Formation is recognized in east-central Utah; however, it is the lithologic and time equivalent of the Curtis Member of the Stump Formation (Pipiringos and Imlay, 1979) and the Pine Butte Member of the Sundance Formation (Pipiringos and others, 1969, p. N10). Therefore, it is laterally continuous from central Utah through northern Utah and northwestern Colorado, and into Wyoming (fig. 16). South of the San Rafael Swell, the Curtis Formation thins and is missing in the Henry Mountains Basin. To the east of the San Rafael Swell, the upper part of the Curtis Formation, at least locally, was truncated by Jurassic erosion (R.B. O'Sullivan, U.S. Geological Survey, written commun.,

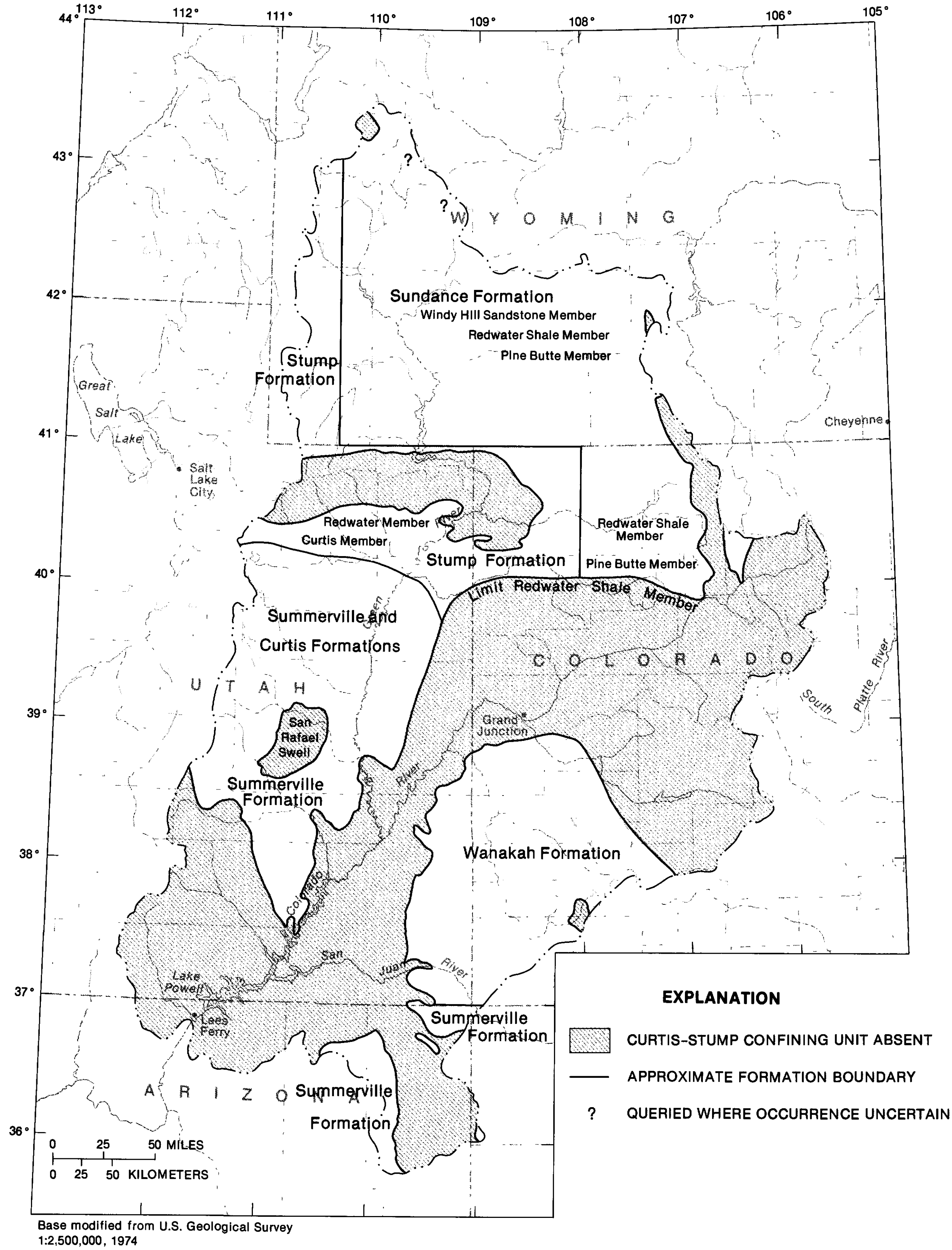


FIGURE 16.—Approximate areal extent of formations within the Curtis-Stump confining unit.

1986), and the lower part grades laterally into the Moab Tongue of the Entrada Sandstone (fig. 12).

The Summerville Formation in the study area is limited to eastern Utah and northeastern Arizona (fig. 16). In Utah, the Summerville Formation is truncated by Jurassic erosional surfaces to the north, south, and east. Around the San Rafael Swell, the Curtis-Stump confining unit is composed of the Summerville and Curtis Formations, which intertongue and increase in thickness westward across the area from about 100 ft in the east to more than 700 ft in the west (pl. 2*C*). South of the San Rafael Swell where the Curtis is absent, the Summerville Formation decreases in thickness southward from 400 ft near the swell to about 200 ft in the south. In extreme northeastern Arizona, the rocks referred to as Summerville Formation are actually equivalent to and continuous with the Wanakah Formation of southeastern Utah and southwestern Colorado (R.B. O'Sullivan, U.S. Geological Survey, oral commun., 1985).

The Curtis-Stump confining unit in south-central Wyoming and northwestern Colorado is represented by the Pine Butte, Redwater Shale, and Windy Hill Sandstone Members of the Sundance Formation (pl. 1). The total thickness of these members is less than 100 ft in northwestern Colorado and adjacent parts of Wyoming and 100 to 200 ft in Wyoming, north of the Uinta Uplift. Westward, the Pine Butte and Redwater Shale Members grade into the Curtis and Redwater Members of the Stump Formation in the Wyoming thrust belt and along the perimeter of the Uinta Uplift of Utah and northwestern Colorado (fig. 16). The Windy Hill Sandstone Member, though widespread, is very thin and locally discontinuous (Segerstrom and Young, 1972, p. 28). It intertongues with the lower Morrison Formation.

The entire Redwater Shale Member as well as the upper part of the Pine Butte Member is truncated by a Jurassic erosional surface along a line roughly paralleling the White River in northwestern Colorado (pl. 2*C*, fig. 16). The lower part of the Pine Butte Member grades southward into the Entrada Sandstone in the same general area.

The Stump Formation, composed of the Curtis and Redwater Members, is less than 100 ft thick along the eastern edge of the Wyoming thrust belt, but it increases in thickness westward to about 300 ft (pl. 2*C*). Along the northern edge of the Uinta Uplift it is more than 300 ft thick, and along the eastern edge it is about 100 ft thick.

Lithologic Character

The strata that make up the Curtis-Stump confining unit are predominantly marine and marginal marine in origin. They were deposited during repeated transgressions and regressions of Jurassic seas encroaching on the area from the northeast and resulting in complex interfingering of continental and marine beds.

Siltstone is the major component of the formations that make up the Curtis-Stump confining unit. Both the Wanakah and Summerville Formations are predominantly siltstone with interbedded sandstone and shale. Jobin (1962, p. 50) states that the abundance of siltstone and shale precludes these formations "from having any significant regional transmissivity" of water.

The Curtis Formation is composed of very fine to fine-grained sandstone with interbedded siltstone, shale, limestone, and thin beds of gypsum. Jobin (1962, p. 50) describes the Curtis as having little transmissive capacity for water in the Colorado Plateau, although sandstone samples from the formation had relatively large permeabilities.

The Stump Formation is also predominantly very fine to fine-grained sandstone with interbedded siltstone, shale, and limestone. Each of the members of the Stump Formation consists of two lithologic units. A lower sandstone unit and an upper shale unit make up the Curtis Member (Pipiringos and Imlay, 1979, p. C3); conversely, a lower shale unit and an upper sandstone unit form the overlying Redwater Member.

The Pine Butte Member of the Sundance Formation ranges from lime-cemented sandstone with interbedded siltstone and shale in western Wyoming to shale with thin sandstone beds in northwestern Colorado. The Redwater Shale Member is mainly clayey siltstone and sandstone, and the Windy Hill Sandstone Member is flat-bedded, calcium-cemented sandstone with locally interbedded clayey siltstone lenses (Pipiringos and others, 1969, p. N15).

MORRISON AQUIFER

Stratigraphy and Saturated Thickness

The Morrison aquifer includes the Cow Springs Sandstone (locally), the Junction Creek Sandstone, and the Tidwell, Bluff Sandstone, Salt Wash, Recapture, and Westwater Canyon Members of the Morrison Formation. The Morrison aquifer is present in the central and southern parts of the study area (pl. 3*C*). In the northern and easternmost parts of the study area, the extent of the Morrison aquifer is unknown; however, the fine-grained lithology of the Morrison Formation in these areas indicates that the presence of an extensive aquifer is unlikely.

The Bluff Sandstone Member of the Morrison Formation and the Cow Springs and Junction Creek Sandstones are limited to the southern part of the study area. The Junction Creek Sandstone is present in southwestern Colorado. The Bluff Sandstone Member extends from southeastern Utah into northeastern Arizona and northwestern New Mexico. The Cow Springs Sandstone in extreme northeastern Arizona and northwestern New Mexico (not shown on pl. 1) is also included in the Morrison aquifer.

The Bluff Sandstone Member and Junction Creek Sandstone overlie a Jurassic unconformity that also marks the base of the Morrison Formation in most of the study area. The two sandstone units are lateral equivalents and are continuous in the subsurface, although the outcrops of the Junction Creek in Colorado are separated from those of the Bluff Sandstone Member in adjacent States (Craig and others, 1955, p. 133). The Bluff Sandstone Member, which is about 340 ft thick at Bluff, Utah, intertongues with and grades laterally into the Tidwell Member of the Morrison Formation in southeastern Utah (O'Sullivan and Pierce, 1983; O'Sullivan, 1984a, p. 15). Southward into Arizona, the Bluff intertongues with the Salt Wash Member (Craig and others, 1955, p. 133). Similarly, the Junction Creek, which is about 275 ft thick, grades northward into what was believed to be the upper part of the Summerville Formation (Haynes and others, 1972) but is now referred to as the Tidwell Member of the Morrison Formation (O'Sullivan, 1984a, p. 9).

Locally, in northeastern Arizona and northwestern New Mexico, the Cow Springs Sandstone is isolated from the Entrada-Preuss aquifer by intertonguing of the Summerville Formation, and thus it is included in the overlying Morrison aquifer. The Cow Springs Sandstone is 70 to 100 ft thick in the area of intertonguing (O'Sullivan, 1978). It is equivalent to the upper Summerville Formation (O'Sullivan, 1978) and is unconformably overlain by the Morrison Formation.

The Tidwell Member of the Morrison Formation, described by O'Sullivan (1984a, p. 15), extends over wide areas of central and eastern Utah and western Colorado, and the average thickness is 50 ft. The Tidwell Member grades upward into the Salt Wash Member of the Morrison Formation. A Jurassic unconformity indicates the base of the Tidwell and the base of the Morrison Formation.

The Salt Wash Member of the Morrison Formation is recognized in western Colorado, northeastern Arizona, and eastern Utah (fig. 17). It is more than 600 ft thick in south-central Utah, northwest of the Colorado River, and thins to zero at its depositional boundary along the western margin of the study area and in northwestern Arizona (Craig and others, 1955, p. 138). Subsurface data from oil exploration wells indicate that the Salt Wash is more than 500 ft thick in the western Uinta Basin and the Wasatch Plateau (pl. 3*C*).

The Salt Wash Member grades to the north and east into the undifferentiated Morrison Formation of northwestern Colorado and southwestern Wyoming, beyond the recognizable limit of the Salt Wash Member shown in figure 17. The Salt Wash intertongues with the Recapture Member in southeastern Utah and grades into the Recapture in northeastern Arizona (Craig and others, 1955, p. 135). Where the basal Tidwell Member of the Morrison is absent, the Salt Wash overlies the Jurassic unconformity that generally marks the base of the Morrison Formation. A distinct lithologic difference distinguishes the Salt Wash from the overlying Brushy Basin Member north of the Four Corners Platform.

The Recapture and Westwater Canyon Members of the Morrison Formation, which overlie the Salt Wash in the Four Corners Platform, have virtually the same areal extent (fig. 17). The Recapture Member ranges in thickness from 600 ft in northeastern Arizona to less than 150 ft along its recognizable limit (fig. 17), thinning to zero along its depositional margin beyond the southern border of the study area (Craig and others, 1955, p. 139).

Exhibiting a similar trend, the Westwater Canyon Member ranges in thickness from about 300 ft along the northern part of the border between Arizona and New Mexico to zero beyond the southern border of the study area. The Westwater Canyon Member grades laterally into the Brushy Basin Member (Craig and others, 1955, p. 136) along its recognizable limit in Utah and Colorado (fig. 17).

The saturated thickness of the Morrison aquifer is shown in figure 18. In northeastern Arizona where the aquifer is as much as 800 ft thick, less than 500 ft of the total thickness is saturated. Similarly, in much of southwestern Colorado and southeastern Utah where the saturated thickness is less than 100 ft, the thickness of the Morrison aquifer is 300 to 500 ft. In general, the aquifer is fully saturated in the central part of the study area. The saturated thickness is unknown in the northern and eastern parts of the study area.

The map of saturated thickness (fig. 18) suggests that a large volume of water is present in the Morrison aquifer. However, the thickness of the rock overlying the Morrison Formation (including the Morrison confining unit) (fig. 19) is more than 2,000 ft in many areas where the saturated thickness of the aquifer is more than 100 ft. In the Uinta and Piceance Creek Basins, the thickness of the overlying rock is more than 12,000 ft (fig. 19). Water from that depth is likely to be unsuitable for most uses and costly to develop. In parts of southeastern Utah and southwestern Colorado where the thickness of the overlying rock is 2,000 ft or less, ground water would be less costly to develop; however, the saturated thickness of the Morrison aquifer in large parts of these areas is less than 100 ft. Conditions favorable for economic withdrawal of water from this aquifer exist in extreme southeastern Utah and southwestern Colorado, where the thickness of the rock overlying the Morrison Forma-

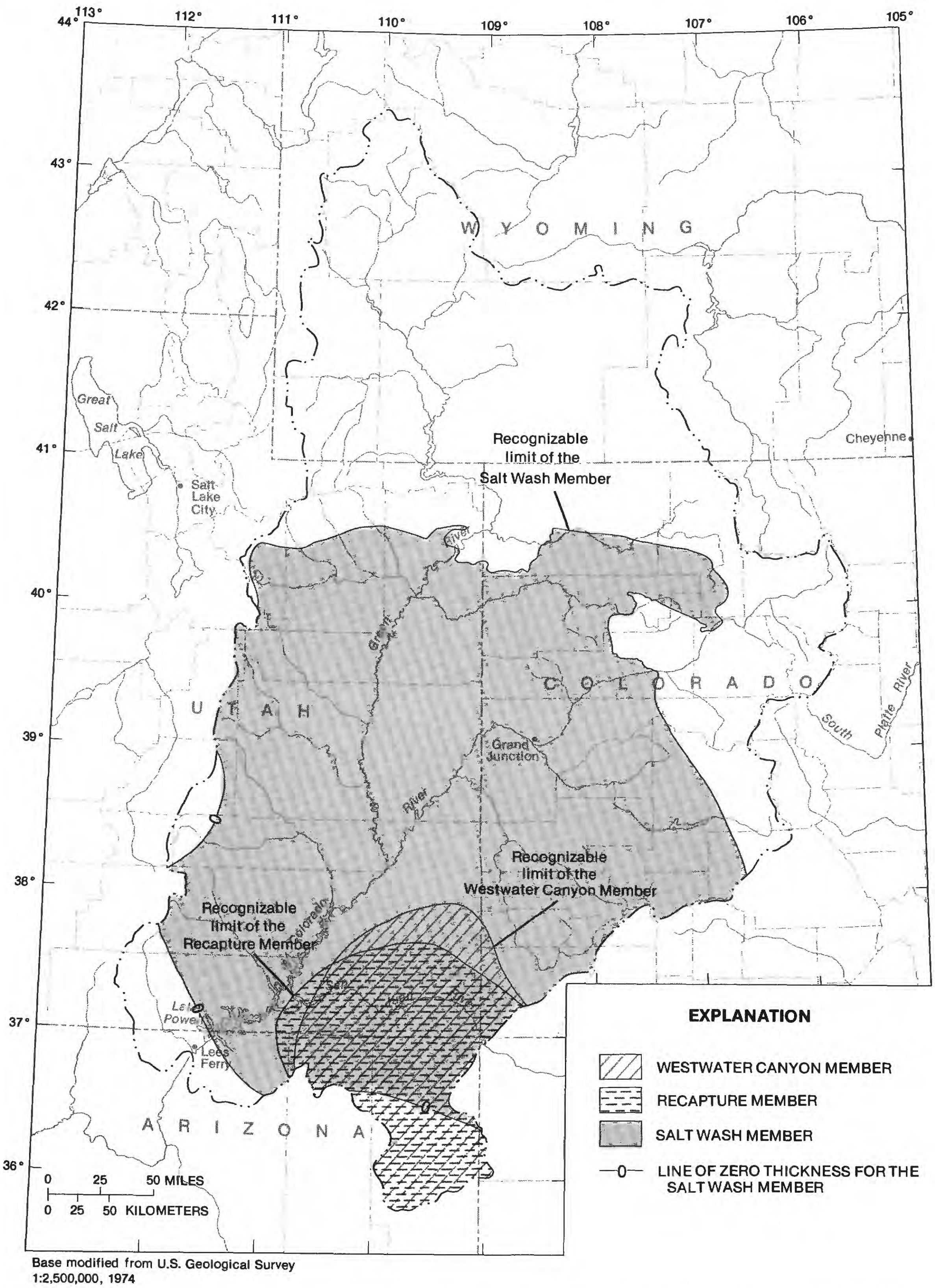


FIGURE 17.—Approximate depositional extent of the Salt Wash, Recapture, and Westwater Canyon Members of the Morrison Formation (modified from Craig and others, 1955, figs. 21, 22, 28).

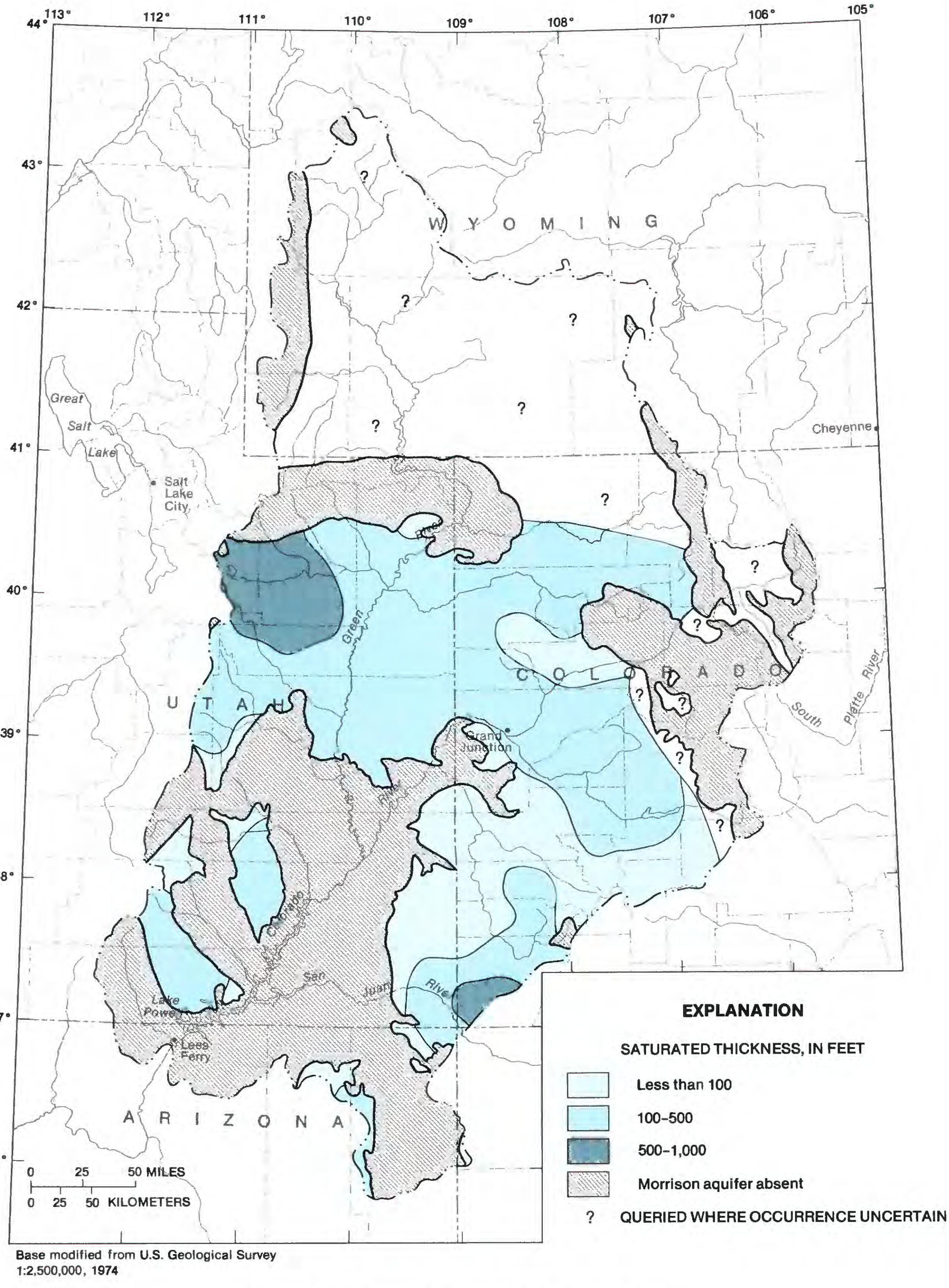


FIGURE 18.—Saturated thickness of the Morrison aquifer.

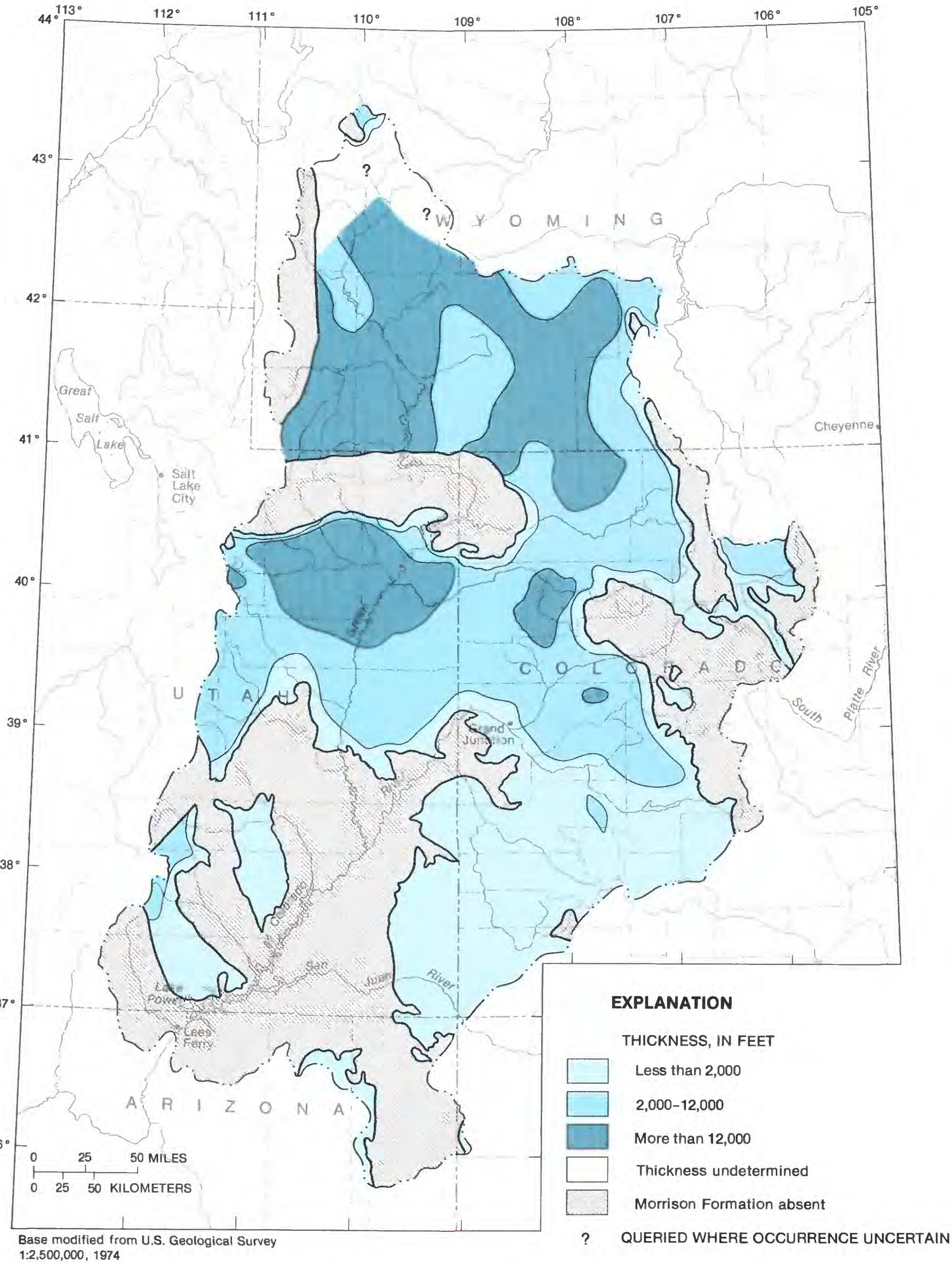


FIGURE 19.—Total thickness of rock overlying the Morrison Formation (Morrison aquifer and confining unit).

tion is less than 2,000 ft, and the saturated thickness of the Morrison aquifer is more than 100 ft.

Lithologic Character

The lithofacies map (fig. 20) shows the general lithologic character of the Morrison aquifer in the area where members of the Morrison Formation can be differentiated (Craig and others, 1955). In the northern and easternmost parts of the study area, the Morrison Formation is undifferentiated.

Siltstone is the dominant lithology of the Tidwell Member of the Morrison Formation, although this unit also contains thin sandstone, chert, and limestone beds as well as limestone nodules. Gypsum is abundant in the Tidwell Member from the San Rafael Swell to east of the Green River (O'Sullivan, 1984a, p. 5).

The Salt Wash, Recapture, and Westwater Canyon Members are composed of interstratified fine- to medium-grained sandstone and claystone. Thin beds of limestone are present in the claystone of the Salt Wash Member along the recognizable limit of the unit (Craig and others, 1955, p. 137). All three members exhibit fluvial sedimentary structures such as scour and fill features, various types of cross laminations, ripple marks, and soft sediment slump structures.

From detailed petrologic studies of the Morrison Formation, Cadigan (1967) describes the following lithologic compositions: Salt Wash—60 percent sandstone, 39 percent siltstone, claystone, and mudstone, and 1 percent limestone and miscellaneous rocks; Recapture—75 percent sandstone and conglomerate, 25 percent siltstone, mudstone, and claystone; and Westwater Canyon—80 percent sandstone and conglomerate and 20 percent siltstone, mudstone, and claystone. He also found that the sandstones of the Morrison Formation are generally fine grained and moderately well sorted.

The Bluff Sandstone Member and the Junction Creek and Cow Springs Sandstones are included in the conglomeratic sandstone and sandstone facies in the south-central part of the study area (fig. 20). These units are chiefly composed of sandstones, subaqueous to eolian in origin, and can be considered relatively homogeneous in lithology and texture compared with the Salt Wash, Recapture, and Westwater Canyon Members of the Morrison Formation.

Statistical analyses of grain-size distributions for sandstones of the Morrison Formation (Cadigan, 1967) show a decrease in grain size and an increase in sorting from southwest to northeast. Grain-size distributions, in addition to thickness (Craig and others, 1955) and lithofacies (Craig and others, 1955; Peterson, 1972) maps, indicate that the major source areas for the Morrison Formation were southwest of south-central Utah, south of Gallup, N. Mex., and in the central Idaho-western Utah-Nevada area. Streams flowed east and northeast, and coarser sediments were deposited near the source areas. As the transporting capacity of the streams decreased, fine-grained sediments were deposited at greater distances from the source areas.

The lithofacies map (fig. 20) reflects the trend of decreasing grain size (increase in clay and mud) with increasing distance from the source area in the south. The Morrison aquifer is largely conglomeratic sandstone and sandstone in the southern part of the study area (fig. 20). Northward, the Morrison aquifer generally increases in shale, claystone, and mudstone, forming the sandstone and mudstone facies. The Morrison aquifer consists solely of the Salt Wash Member in the area of the claystone and lenticular sandstone facies. These rocks are largely claystone with sparse sandstone lenses (Craig and others, 1955, p. 137).

Craig and others (1955, p. 143) found that the ratio of the thickness of stream deposits of sandstone to flood-plain deposits of claystone and sandstone in the Salt Wash and Recapture Members decreases to the north-northeast. Contour maps of permeability and transmissivity by Jobin (1962, p. 60, 61) also show decreasing permeability and transmissivity to the north-northeast in the Salt Wash, Recapture, and Westwater Canyon Members, in part the result of decreasing grain size in that direction.

MORRISON CONFINING UNIT

Stratigraphy

The Morrison confining unit is composed of the Brushy Basin Member of the Morrison Formation in northeastern Arizona, eastern Utah, and western Colorado, and the undifferentiated Morrison Formation in Wyoming and northwestern Colorado. Locally in Wyoming, the Cloverly Formation is included in the confining unit. The areal extent and thickness of the Morrison confining unit are shown on plate 2*D*.

The thickness of the Brushy Basin Member ranges from zero in the southwestern part of the study area, where it was removed by post-Morrison erosion (Craig and others, 1955, p. 156), to more than 400 ft in southwestern Colorado and southeastern Utah (pl. 2*D*). Northward, the thickness of the Brushy Basin is irregular and ranges from about 200 to 400 ft in western Colorado and eastern Utah.

In Wyoming and northwestern Colorado, the Morrison Formation cannot be differentiated into members because the sandstone of the Salt Wash Member cannot be distinguished from the upper part of the Morrison Formation (Craig and others, 1955, p. 156). Craig and others (1955, p. 159) state that the undifferentiated

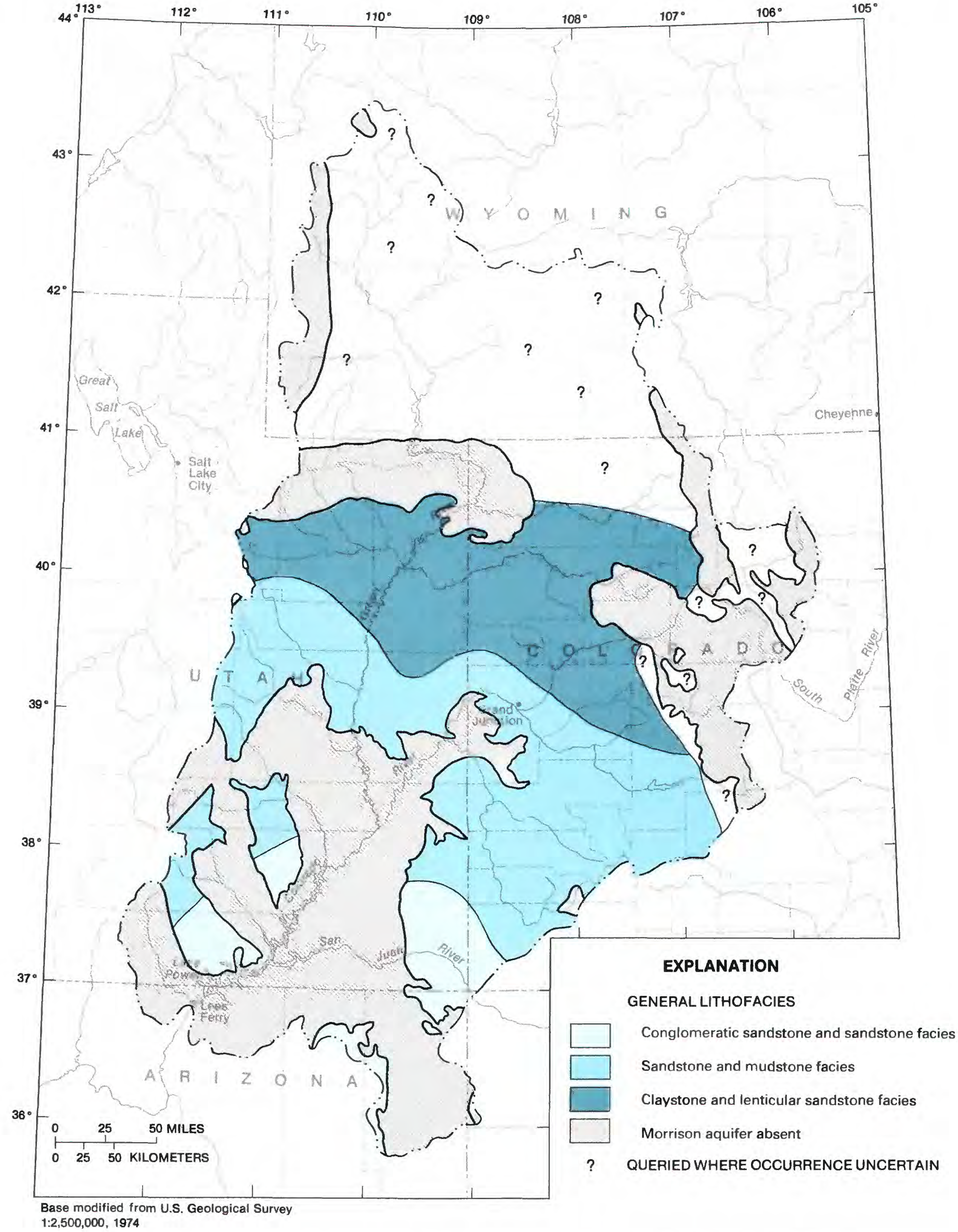


FIGURE 20.—General lithofacies of the Morrison aquifer (modified from Craig and others, 1955).

Morrison is lithologically similar to the Brushy Basin Member. Thus, the undifferentiated Morrison is considered part of the Morrison confining unit in the study area. However, the undifferentiated Morrison Formation may contain aquifers of relatively local extent that are considered part of the Morrison aquifer for this study.

In northwestern Colorado, the undifferentiated Morrison is about 500 ft thick (pl. 2*D*). Thicknesses in Wyoming are 1,000 ft in the northwestern Uinta Uplift, decreasing eastward to 300 ft in the northeastern part of the study area.

The Brushy Basin Member is transitional into the underlying Salt Wash Member where the Recapture and Westwater Canyon Members are absent. In the Four Corners Platform, the lower part of the Brushy Basin grades southward into the Westwater Canyon Member. Farther to the south, the upper part of the Brushy Basin has been truncated by erosion prior to deposition of the Dakota Sandstone. The top of the Morrison confining unit is truncated throughout the study area by a major erosional unconformity which marks the contact with overlying Cretaceous sandstone formations (Pipiringos and O'Sullivan, 1978, p. A26).

In the western Green River Basin, the contact of the undifferentiated Morrison Formation with the overlying Cloverly Formation is selected at the base of a conglomerate bed containing chert clasts (Cloverly), which overlies variegated mudstone of the Morrison. However, in a few places, the conglomerate is absent, and the contact cannot be identified (Furer, 1970, p. 2284). Therefore, the rocks are referred to locally as the Cloverly and Morrison Formations undivided and are included in the Morrison confining unit.

Lithologic Character

Rocks of the Brushy Basin Member are typical of the Morrison Formation throughout the Western United States. About 90 percent of the member consists of poorly sorted, horizontally laminated siltstone, mudstone, and claystone. The remaining 10 percent consists of sandstone and conglomerate (Cadigan, 1967). The principal clay mineral in the Brushy Basin is montmorillonite, a swelling clay from the alteration of volcanic glass shards. Conglomeratic sandstone lenses of fine- to medium-grained, cross-stratified sandstone with stringers of pebbles and granules are common (Craig and Shawe, 1975, p. 163). These sandstone lenses are rarely more than several hundred feet wide, and they pinch out laterally into mudstone. Discontinuous thin-bedded limestone is present locally. Crossbedding and other sedimentary structures in the rocks of the Brushy Basin Member indicate fluvial deposition. Clay and limestone are indicative of deposition in shallow lakes (Craig and others, 1955, p. 160).

The undifferentiated Morrison Formation is similar in lithology to the Brushy Basin, consisting of variegated mudstones with thin, interbedded limestone, sandstone, and conglomerate (Craig and others, 1955, p. 158). As in the Brushy Basin, montmorillonite clay is common. Where the Cloverly and the undifferentiated Morrison Formation are undivided, they are typically variegated mudstone with interbedded siltstone and sandstone. Local aquifers, which are considered part of the Morrison aquifer, may be present in sandstone and conglomerate of the undifferentiated Morrison.

The continuity and relatively great thicknesses of the Brushy Basin Member and the undifferentiated Morrison make these rocks effective confining units. In addition, the abundant swelling clays trap seepage and block its movement through the Morrison confining unit.

DAKOTA AQUIFER

Stratigraphy and Saturated Thickness

The Dakota aquifer includes the Dakota Sandstone, the Burro Canyon, Cedar Mountain, and Cloverly Formations, the Gannett Group, and the Bear River, Smiths, Thomas Fork, and Cokeville Formations. Post-Dakota erosion has removed these rocks from the major uplifted areas, including the Defiance Uplift in Arizona, the Monument and Circle Cliffs Uplifts in Utah, and the White River Uplift in Colorado. The areal extent and thickness of the Dakota aquifer are shown on plate 3*D*.

The thickness of the Dakota aquifer is irregular and variable over short distances, particularly in the eastern and southern extents of the aquifer (pl. 3*D*). The irregular thickness reflects variations in environment and topography when the sediments were deposited, postdepositional compactional differences related to the sandstone-mudstone ratio, and postdepositional erosion (Craig, 1981, p. 198). Because of the variability in thickness, the lines of thickness are necessarily generalized; actual thickness of the unit may vary locally by tens of feet to more than 100 ft from values shown on plate 3*D*.

The thickness of the Dakota aquifer generally increases from 100 ft in the western part of the study area to more than 800 ft in the southwestern Uinta Basin (pl. 3*D*). In southwestern Colorado, the thickness of the Dakota aquifer is 100 to 300 ft. In most of the rest of the area, the thickness averages 100 to 200 ft (pl. 3*D*).

The thickness of the Dakota aquifer is largest in the Wyoming thrust belt where the Gannett Group is 500 to 900 ft thick (pl. 3*D*), and the type sections for the overlying Smiths, Thomas Fork, and Cokeville Formations indicate a total thickness of more than 3,300 ft (Rubey, 1973, p. I8, I14, I17). The aquifer thins to the east and southeast from the thrust belt into the Green River Basin.

In Utah and Colorado, the Dakota Sandstone constitutes the upper part of the Dakota aquifer. In Arizona, the aquifer is composed entirely of the Dakota Sandstone. The thickness of the Dakota Sandstone ranges from zero to 200 ft and averages 100 ft in the study area (Young, 1960, p. 177).

The Burro Canyon Formation and equivalent Cedar Mountain Formation form the lower part of the aquifer in Utah and Colorado. The Burro Canyon Formation is recognized from southeastern Utah to western Colorado, where it pinches out (fig. 21). It grades laterally northwestward into the Cedar Mountain Formation in east-central Utah and northwestern Colorado along a line paralleling the Colorado River in Utah (fig. 21). Commonly, an erosional surface separates the Cedar Mountain and Burro Canyon Formations from the overlying Dakota Sandstone (Craig, 1961, p. 1583); however, Young (1960, p. 176) states that the upper part of the Cedar Mountain grades into the lower part of the Dakota Sandstone in the western Colorado Plateau.

The average thickness of the Burro Canyon is about 130 ft, whereas the thickness of the Cedar Mountain ranges from 130 ft near the arbitrary boundary with the Burro Canyon to more than 500 ft west of Price, Utah (Craig, 1981, p. 196). The Cedar Mountain-Burro Canyon strata account for about 50 to 90 percent of the total thickness of the Dakota aquifer in Utah and Colorado.

The Gannett Group forms the lower part of the Dakota aquifer in the Wyoming thrust belt. It includes, in ascending order, the Ephraim Conglomerate, Peterson Limestone, Bechler Conglomerate, Draney Limestone, and Smoot Formation. The Gannett Group is overlain by the Smiths, Thomas Fork, and Cokeville Formations, which form the upper part of the Dakota aquifer in the thrust belt. Eastward, the Gannett Group grades into the Cloverly Formation in the Green River Basin, and the three overlying formations grade into the Bear River Formation (Rubey, 1973, p. I4). The Bear River Formation grades into the Thermopolis Shale (Mancos confining unit) in the Rock Springs Uplift area, essentially terminating the upper part of the Dakota aquifer.

The base of the Dakota aquifer is marked by unconformities throughout most of the area. In contrast, the contact of the Dakota aquifer with the overlying Mancos confining unit is generally conformable and intertonguing. However, Young (1960, p. 176) notes that this contact in the Colorado Plateau is complicated by several periods of erosion late in the deposition of the Dakota Sandstone, resulting in "a series of overlapping disconformities rising toward the west." In the Wyoming thrust belt, the upper contact of the Dakota aquifer with the Mancos confining unit is gradational and intertonguing.

The saturated thickness of the Dakota aquifer is shown in figure 22. Comparison of the saturated thickness (fig. 22) and the thickness of the aquifer (pl. 3*D*) indicates that the Dakota aquifer is saturated in most of the study area. Generally, this unit is partly saturated or unsaturated along the outcrop margins and where the total thickness of the unit is less than 100 ft. With the exception of the Wyoming thrust belt and the western Uinta Basin, the total saturated thickness of the Dakota aquifer is less than 500 ft.

Comparison of the thickness of the rock overlying the Dakota aquifer (fig. 23) with the saturated thickness of the aquifer (fig. 22) reveals that where the saturated thickness, and thus ground-water storage, is largest, the thickness of the overlying rock is more than 2,000 ft. Below 2,000 ft, water quality is likely to deteriorate and the cost of withdrawal increases. However, in the Kaiparowits Basin, southwestern Colorado, west of the San Rafael Swell, and along the southern and eastern margins of the Uinta Uplift, the saturated thickness ranges from 100 to more than 500 ft, and the thickness of the overlying rock is less than 2,000 ft, making these areas more suitable for development of the ground water.

Lithologic Character

The materials that make up the Dakota aquifer in the Upper Colorado River Basin were deposited in the coastal plain and along the margins of a transgressing epicontinental sea which encroached on the area from the east-southeast (Hintze, 1982, p. 67). Periodic seaward (east) tilting of the land resulted in some scouring of previously formed nonmarine deposits (Young, 1973, p. 12). Hence, numerous unconformities and intertonguing mark the contact between the Dakota Sandstone and the overlying Mancos Shale. As a result of the tectonic instability, the geologic and stratigraphic relations among and within individual rock units are complex and include intertonguing, scour and fill, and local and regional unconformities.

Sandstone and mudstone are the dominant rock types in both the Burro Canyon and Cedar Mountain Formations, although the proportions are different in each. Craig (1981, p. 197) states that "over most of the extent the Burro Canyon consists of 50 percent sandstone, whereas the Cedar Mountain contains more than 30 percent sandstone in only a few places." Minor chert and limestone beds are found in both formations. The sandstone may form single thick beds, but more commonly thin mudstone beds separate the sandstone into several beds. The sandstone ranges from coarse-grained, poorly

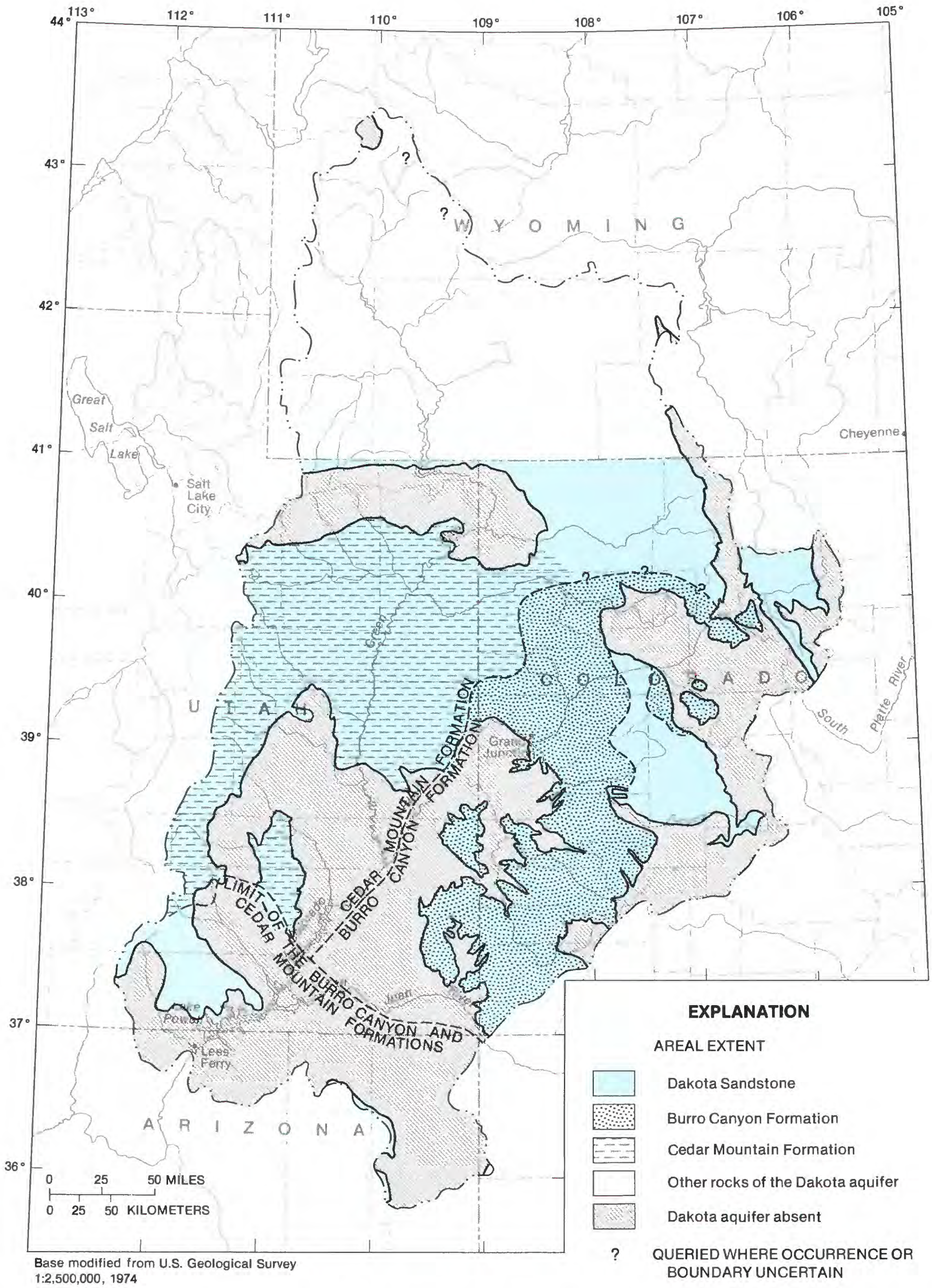


FIGURE 21.—Approximate areal extent of the Cedar Mountain and Burro Canyon Formations, the Dakota Sandstone, and other rocks of the Dakota aquifer (modified from Craig, 1981, fig. 1).

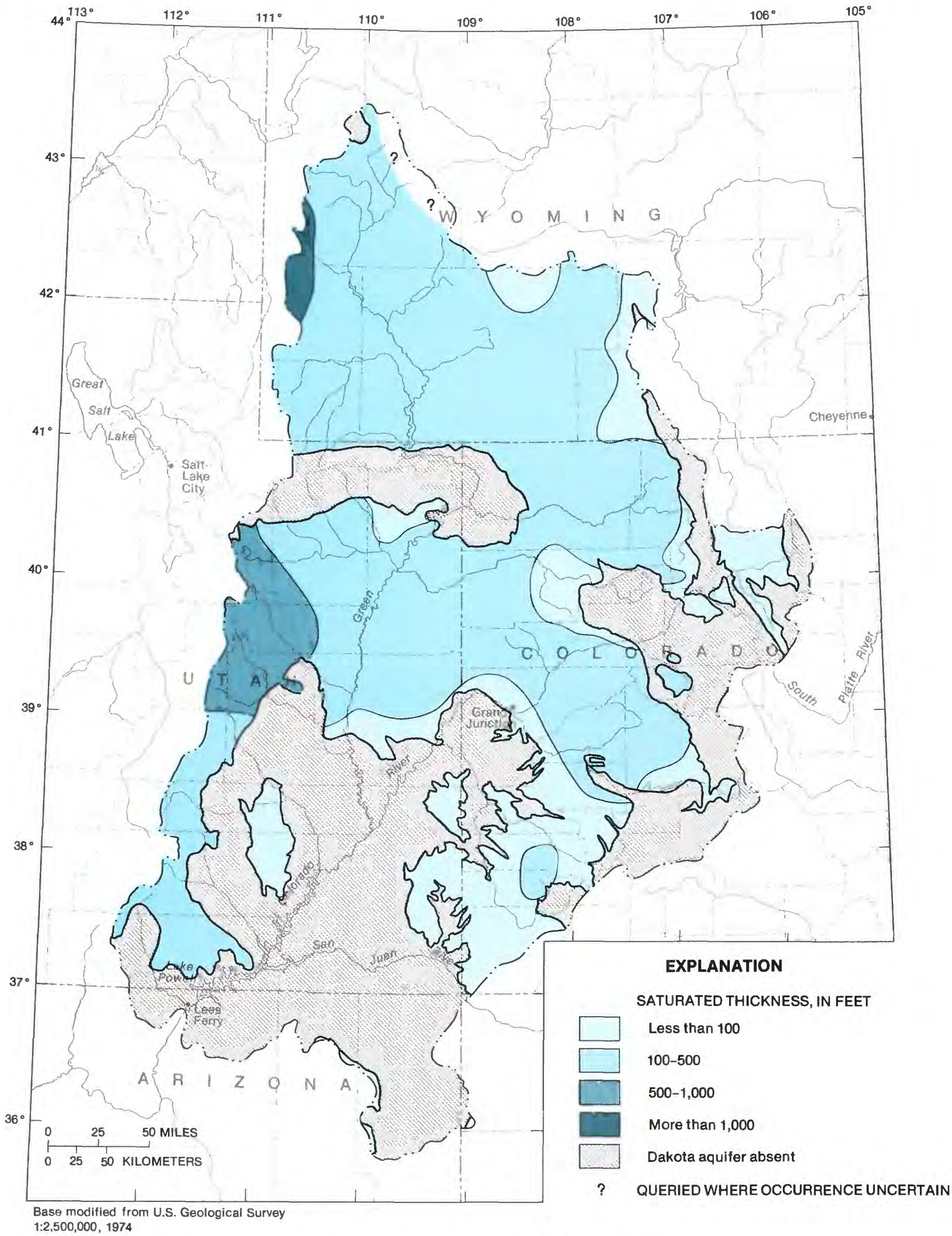


FIGURE 22.—Saturated thickness of the Dakota aquifer.

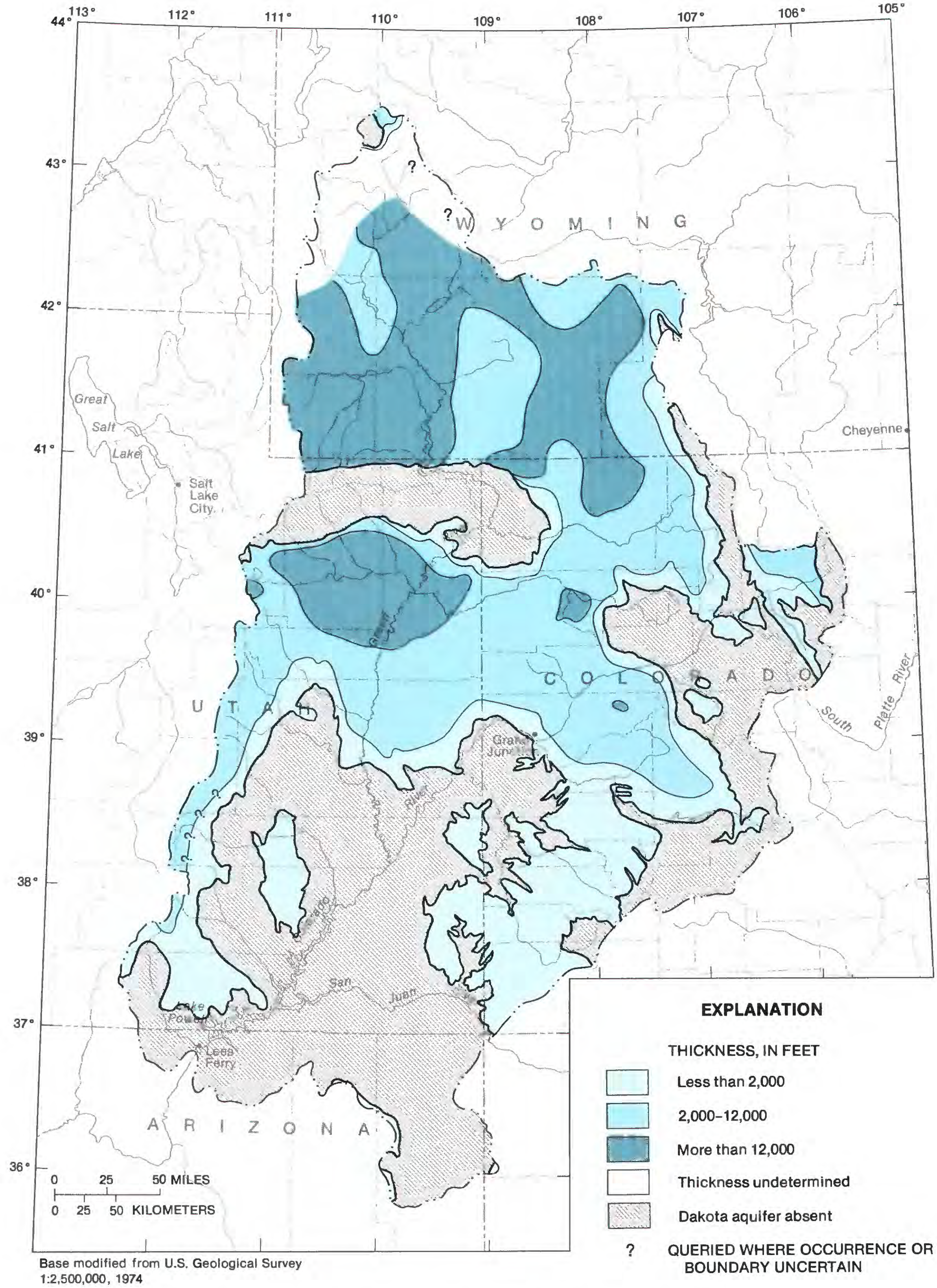


FIGURE 23.—Total thickness of rock overlying the Dakota aquifer.

sorted, cross-stratified sandstone and conglomeratic sandstone such as the basal Buckhorn Conglomerate Member of the Cedar Mountain to finer grained, better sorted, parallel-bedded sandstone (Craig, 1981, p. 197). Calcium carbonate commonly cements the sandstone, and in some places it has been replaced by silica, forming quartzite or chert beds (Young, 1960, p. 172).

Silty to sandy mudstone generally predominates in the interbedded mudstone beds. The major clay mineral in the mudstone of the Cedar Mountain Formation is montmorillonite, a swelling clay, which may increase the confining properties of the mudstone and isolate individual sandstone beds from one another. Clay in the Burro Canyon is predominantly nonswelling illite (Craig, 1981, p. 197).

The Dakota Sandstone contains a large variety of lithologic types, including conglomerate, sandstone, siltstone, mudstone, carbonaceous shale, and coal. However, it can be roughly divided into three units: a widely traceable conglomeratic sandstone at the base, a middle unit of carbonaceous shale and impure coal with sandstone and siltstone lenses, and an upper massive, fine- to medium-grained sandstone (Young, 1973, p. 10). The Dakota generally lacks the calcareous cement of the Cedar Mountain and Burro Canyon Formations (Young, 1960, p. 174). Clay, silica, and iron are the most common cementing agents. Of the six samples of Dakota Sandstone analyzed for carbonate cement during this study (Weigel, 1987, table 6), five ranged from 0.08 to 3.0 percent by weight of calcium carbonate and four had values of less than 0.34 percent. Only one sample had 22.4 percent calcium carbonate.

The percentage of sandstone and conglomerate in the Dakota aquifer in the central part of the study area is shown in figure 24. The sandstone content is generally greater where the Burro Canyon is present, particularly along a northeast-trending zone from Bluff, Utah, to Grand Junction, Colo., which coincides with the thickest part of the Burro Canyon (Craig, 1981, p. 198). In east-central Utah, mudstone of the Cedar Mountain Formation is the thickest part of the aquifer and sandstone is generally less than 40 percent of the aquifer thickness.

The Gannett Group of western Wyoming consists mainly of fluvial and lacustrine sediments deposited in a structural trough under conditions of marked tectonic activity (Eyer, 1969, p. 1368). Rock types vary from conglomerate with interbedded channel-fill deposits of sand, silt, and clay (Ephraim and Bechler Conglomerates) to freshwater limestone and interbedded shale (Peterson and Draney Limestones) and mudstone (Smoot Formation). Rocks of the Gannett Group are well cemented, and samples of the Ephraim and Bechler Conglomerates contain 25 to 31 percent carbonate cement (Furer, 1970, p. 2285).

The Cloverly Formation, equivalent to the Gannett Group, is divided into two parts. An upward-fining sequence of conglomeratic sandstone forms the lower part. The upper part is variegated bentonitic shale with siliceous and calcareous nodular zones (Furer, 1970, p. 2284). Locally, the Cloverly rocks are similar to the underlying undifferentiated Morrison Formation and are included in the Morrison confining unit.

Rocks of the overlying Smiths, Thomas Fork, and Cokeville Formations are generally finer grained than those of the Gannett Group. They include shale and quartzitic sandstone of the Smiths Formation, mudstone and fine- to medium-grained sandstone of the Thomas Fork Formation, and fossiliferous sandstone, limestone, claystone, mudstone, porcelanite, and bentonite of the Cokeville Formation. Like the Gannett Group, these deposits generally are cemented by calcium carbonate (Rubey, 1973).

The Bear River Formation consists of a thick sequence of interbedded black shale, sandstone, mudstone, and limestone. Stokes (1955, p. 81) described the formation in terms of four units: (1) a relatively thin basal sequence of black shale and quartzitic sandstone, (2) interbedded mudstone and sandstone, (3) a thick sequence of sandstone and shale with coal beds and abundant limestone near the base, and, at the top, (4) a thick sequence of variegated mudstone and sandstone beds.

The widely varying lithologies that form the Dakota aquifer, particularly the abundance of mudstones and associated fine-grained rocks, indicate that locally the Dakota aquifer may be a confining unit. In addition, where the formations of the Dakota aquifer are well cemented, interstitial calcium carbonate may hamper ground-water movement.

MANCOS CONFINING UNIT

Stratigraphy

The Mancos confining unit is composed primarily of the Mancos Shale. In Wyoming, the confining unit includes the Quealy and Sage Junction Formations of the Gannett Group, the Thermopolis, Mowry, Baxter, Steele, Cody, Aspen, and Hilliard Shales, and the Frontier, Niobrara, and Blind Bull Formations (pl. 1). The areal extent and thickness of the Mancos confining unit are shown on plate 2*E*.

The Mancos Shale is present in most of the study area in Colorado, Utah, and Arizona, although it has been eroded from major uplifts in the southern half of the area (pl. 2*E*). In the Kaiparowits Basin, the equivalent of the Mancos is the Tropic Shale.

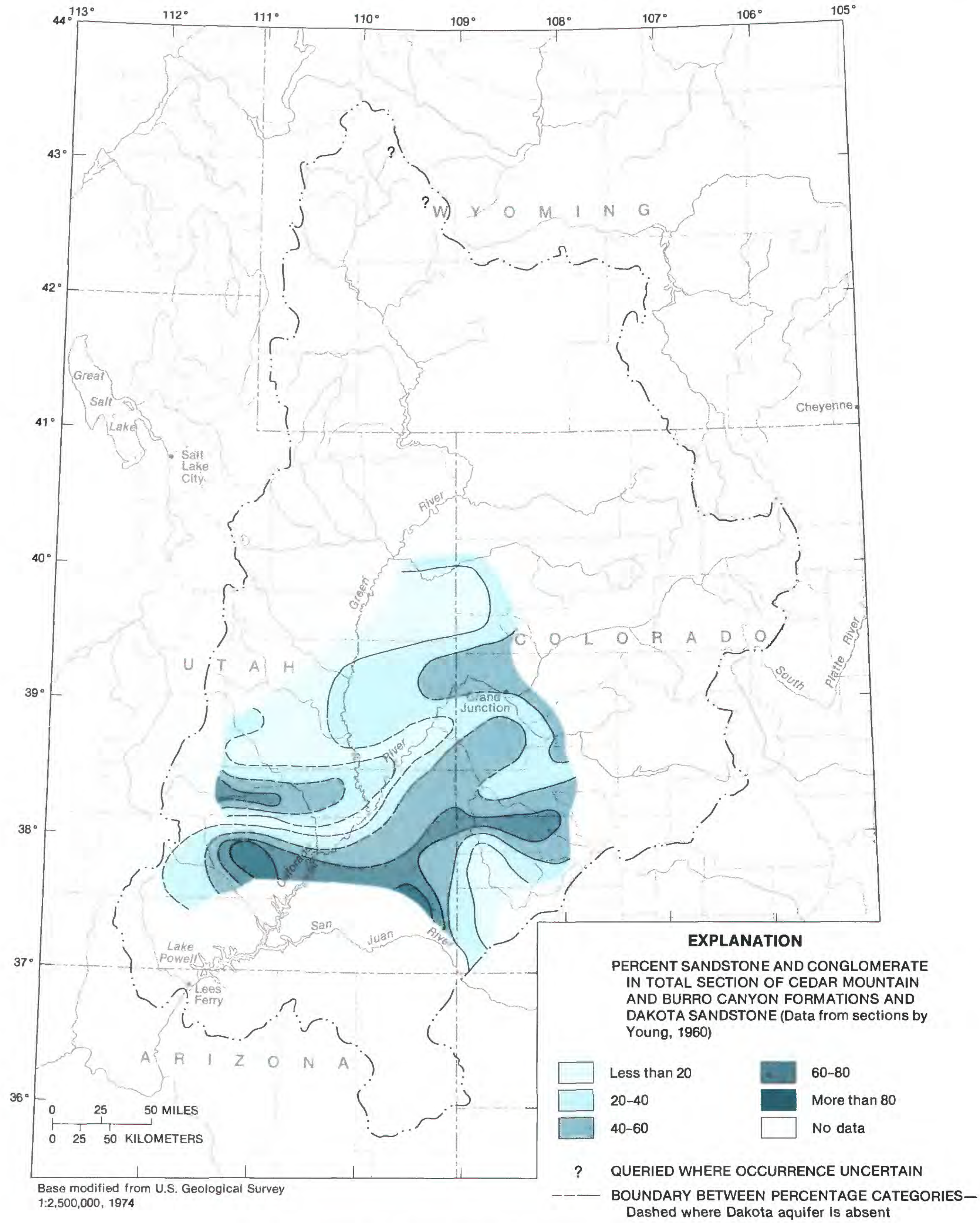


FIGURE 24.—Percentage of sandstone and conglomerate in the Dakota aquifer.

The Mancos confining unit is thickest in the deep basins and in the Wasatch Plateau (pl. 2*E*). It is more than 4,000 ft thick in the center of the Uinta Basin, thinning to less than 1,000 ft near the northern margin, more than 7,000 ft thick in the eastern Piceance Creek Basin, and an average of 5,000 ft thick in the Sand Wash Basin in northwestern Colorado. The Tropic Shale ranges from 600 to more than 1,000 ft thick.

In Wyoming, the Mancos confining unit is as much as two times as thick as it is in Utah and Colorado (pl. 2*E*). In the Green River Basin, the Mancos confining unit ranges from about 5,000 ft thick at the margins to more than 12,000 ft thick in the east-central part of the basin. In the Washakie Basin, east of the Rock Springs Uplift, it ranges from 6,000 to more than 11,000 ft thick. In the central Great Divide Basin, the confining unit is 14,000 ft thick (pl. 2*E*).

From Utah and Colorado northward into Wyoming, the Mancos confining unit grades into a thicker, predominantly shale sequence that has been subdivided into several formations. Included in the Mancos confining unit in western Wyoming is the Thermopolis Shale, which grades into the Bear River Formation of the Dakota aquifer to the west in the Green River Basin (pl. 1). In the Wyoming thrust belt, the Quealy Formation pinches out, but the overlying Sage Junction Formation grades eastward into the Aspen Shale (Rubey, 1973, p. I4), which in turn grades into the Mowry Shale in the Rock Springs Uplift area (McGookey and others, 1972, p. 202). The Mowry Shale in Wyoming is equivalent to the Mowry Member of the Mancos Shale in northeastern Utah and northwestern Colorado (pl. 1).

The Blind Bull Formation of the northeastern Wyoming thrust belt and the western Green River Basin is equivalent to the Frontier Formation and Hilliard Shale to the southeast. The Frontier Formation is continuous eastward across Wyoming and becomes the Frontier Sandstone Member of the Mancos Shale in Utah and Colorado. The overlying Hilliard Shale grades eastward into the Baxter Shale, which in turn grades laterally into the Niobrara Formation and the overlying Steele Shale. Where the Niobrara and Steele are indistinguishable from each other, the rocks constitute the Cody Shale (pl. 1).

The contact between the Mancos confining unit and the underlying Dakota Sandstone or related strata has been variously described as conformable and gradational (Craig and others, 1955, p. 161) to intertonguing and locally unconformable (Young, 1960, p. 176). Several transgressions and regressions of the epicontinental sea during deposition of thick Mancos and related deposits resulted in intertonguing and gradation of the Dakota into the Mancos. Periods of erosion or nondeposition late in the cycle of Dakota deposition resulted in unconformable contacts locally. However, in general, the basal Mancos contact is considered conformable.

Similarly, the upper contact of the Mancos confining unit with the overlying Mesaverde Group and related rocks exhibits gradational and intertonguing characteristics (McGookey and others, 1972, p. 212–214) because of several transgressions and regressions of the Late Cretaceous sea. The intertonguing relationships are important hydrologically where tongues of the Mancos Shale surround sandstone units of the Mesaverde Group. In particular, the Buck Tongue of the Mancos Shale isolates the Castlegate Sandstone from overlying sandstones (fig. 25).

Lithologic Character

The Mancos confining unit is predominantly marine shale, mudstone, and claystone deposited during transgressions of the Cretaceous sea, interbedded with sandstone and shale deposited during regressive cycles of the sea (Hale and Van De Graaff, 1964). The shale, mudstone, and claystone commonly contain thin sandstone lenses, interbedded siltstone, and zones of limestone concretions or limestone beds. These fine-grained rocks have very small permeabilities and inhibit infiltration of precipitation (Hood, 1976, p. 9). In essence, the shale and related deposits form a massive barrier to horizontal and vertical ground-water movement.

Despite the confining characteristics of most of the Mancos confining unit, some of the sandstone strata deposited during regressions of the sea are productive aquifers locally. Hood (1976, p. 9) describes the Frontier Sandstone Member of the Mancos Shale in the Uinta Basin as crossbedded sandstone with a middle shale unit and thin coal beds in the upper part. The permeability is very small to moderate, but freshwater has been obtained from the Frontier (Hood, 1976, p. 9). In the eastern Uinta Basin, the Frontier is overlain by the Mesaverde Group and, thus, is included in the Mesaverde aquifer in that area. Sandstones of the equivalent Frontier Formation in Wyoming are very fine to fine grained.

The Ferron Sandstone Member of the Mancos Shale is a producing aquifer in east-central Utah, where it consists of thin-bedded marine sandstone and shale ranging in thickness from about 80 to 850 ft (Lines and Morrissey, 1983, p. 7). Comparison of the total thickness of the Ferron Sandstone Member and the net sandstone thickness within the Ferron (Walton, 1968, p. 938, 939) indicates that sandstone accounts for 20 to 50 percent of the member. The younger Emery Sandstone Member of the Mancos is dominantly a marine sandstone within the study area and could yield water locally.

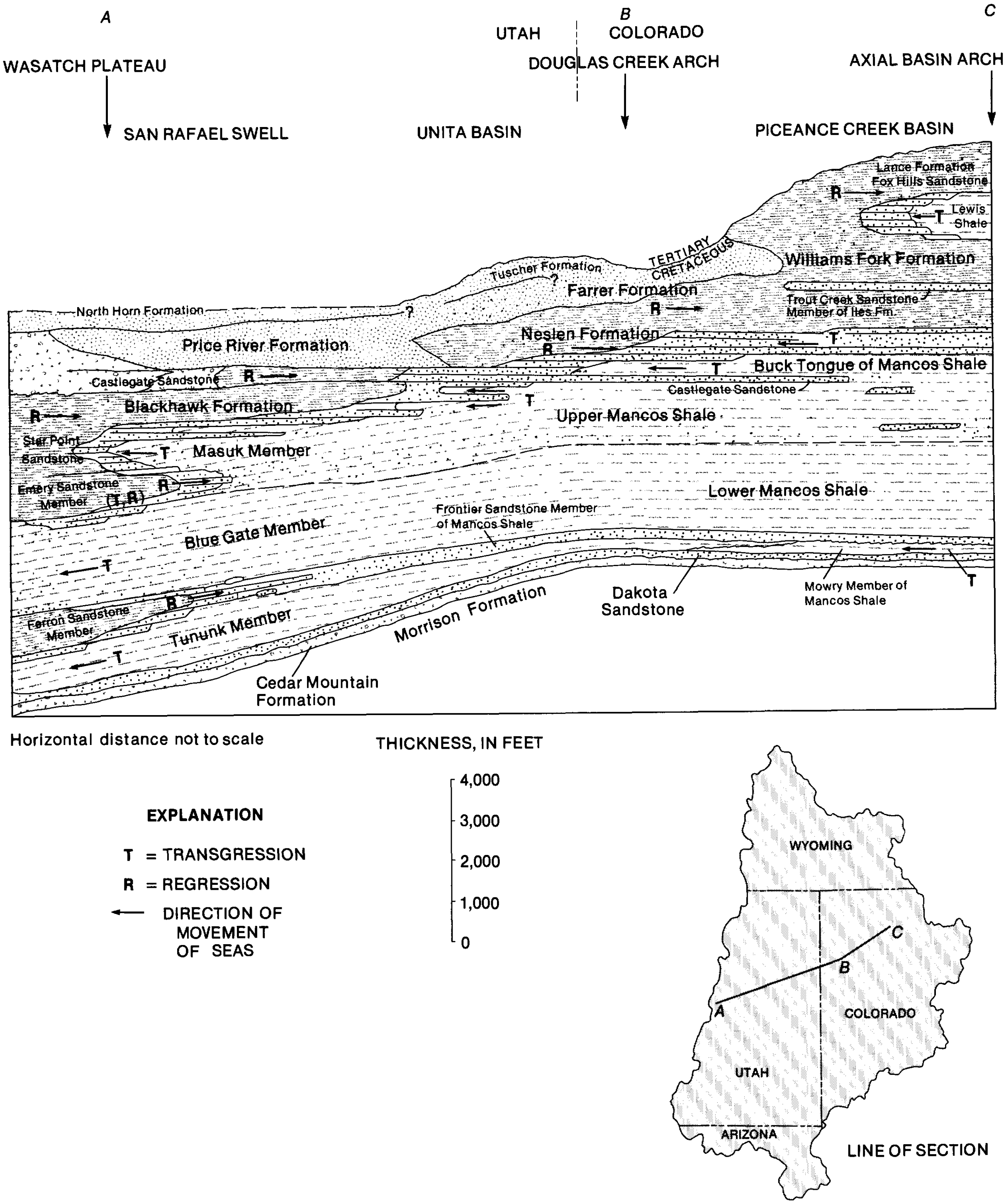


FIGURE 25.—Diagrammatic restored section of Cretaceous rocks, west to east across the central part of the study area (modified from Hale and Van De Graaff, 1964, fig. 9).

MESAVERDE AQUIFER

STRATIGRAPHY AND SATURATED THICKNESS

The Mesaverde aquifer is composed of the Frontier Sandstone Member of the Mancos Shale (locally); the formations of the Mesaverde Group, which are shown on plate 1; the Mesaverde, Adaville, Lance, and Kaiparowits Formations; the Straight Cliffs, Wahweap, and Fox Hills Sandstones; the Lewis Shale; and the lower parts of several formations that are Late Cretaceous to early Tertiary in age, including the Canaan Peak, North Horn, Currant Creek, and Evanston Formations (pl. 1).

Erosion has removed the Mesaverde aquifer from much of the southern part of the study area, although it is exposed in the Henry Mountains, Kaiparowits, and Black Mesa Basins (pl. 3*E*). In the northern part of the study area, the Mesaverde aquifer is continuous across the deep basins but has been eroded from major uplifts such as the Uinta and White River Uplifts.

In general, the Mesaverde aquifer is thickest in the major basins, thinning toward the basin margins (pl. 3*E*). It is 8,000 ft thick in the Washakie and Great Divide Basins of Wyoming and more than 7,000 ft thick in the Piceance Creek Basin in Colorado. In the Uinta Basin near the western border of the study area, the aquifer is more than 4,000 ft thick. In southwestern Wyoming, the thickness of the Mesaverde aquifer is irregular. The aquifer thickens eastward from about 1,000 ft in the eastern part of the Wyoming thrust belt to more than 4,000 ft near the Rock Springs Uplift (pl. 3*E*).

Thicknesses shown on plate 3*E* do not include the Upper Cretaceous to lower Tertiary formations that are part of the Mesaverde aquifer. The thickness of the Upper Cretaceous to lower Tertiary rocks that are hydrologically connected to the Mesaverde is unknown, and they could not be included in the thickness map because of a lack of detailed stratigraphic data.

At the northwestern extent of the Kaiparowits Basin, the Mesaverde equivalents are as much as 4,000 ft thick (Gregory and Moore, 1931, pl. 17). These rocks thin, largely owing to Cenozoic erosion, to about 500 ft along the eastern and southeastern borders of the basin. In Black Mesa Basin at the southern edge of the study area, the Mesaverde aquifer is several hundred feet thick (Cooley and others, 1969, p. A8).

The Mesaverde Group is the major part of the aquifer in much of the study area. As a result of numerous lateral facies changes, formations of the Mesaverde Group have been given different names at different localities. The general location and lateral equivalents of the formations within the Mesaverde Group are shown on plate 1. The Frontier Sandstone Member of the Mancos Shale is included in the Mesaverde aquifer where the Mesaverde Group directly overlies the Frontier Sandstone Member.

In the western Uinta Uplift and Basin and west-central Colorado, the Mesaverde Group is undivided and is referred to as the Mesaverde Formation. Farther south in the Kaiparowits Basin, rocks in part equivalent to the Mesaverde Group are, in ascending order, the Straight Cliffs Sandstone, Wahweap Sandstone, and Kaiparowits Formation. The Straight Cliffs and lower Wahweap Sandstones are actually equivalent to and at one time probably intertongued with the Mancos Shale (Peterson and others, 1980, p. 165), although Cenozoic erosion has physically isolated them from the main Mesaverde outcrops to the north. The upper Wahweap Sandstone, the Kaiparowits Formation, and part of the overlying Canaan Peak Formation are equivalent to and at one time may have graded northeastward into the Mesaverde Group in east-central Utah (Peterson and others, 1980, p. 165). The Mesaverde Group rocks of Black Mesa Basin (Toreva and Wepo Formations and Yale Point Sandstone), which extend into the study area along the southern boundary (pl. 3*E*), are also isolated from outcrops of the Mesaverde to the north.

In the western Wyoming thrust belt, the Upper Cretaceous rocks have largely been removed by erosion and the Mesaverde aquifer is composed entirely of the lower part of the Upper Cretaceous and lower Tertiary Evanston Formation. Eastward in the Green River Basin, equivalents of the Evanston are absent, and the Adaville Formation forms the aquifer. The Adaville is equivalent to the Blair and Rock Springs Formations of the Mesaverde Group to the east (Weimer, 1961, p. 20). From the Rock Springs Uplift to the eastern border of the study area in Wyoming, the Mesaverde aquifer unit includes the Mesaverde Group and the overlying Lewis Shale, Fox Hills Sandstone, and Lance Formation (pl. 1).

The Lewis Shale, together with the intertonguing Fox Hills Sandstone and the overlying Lance Formation, forms the uppermost part of the Mesaverde aquifer in northwestern Colorado and adjacent parts of Wyoming. The Lewis Shale intertongues with the Almond Formation of the underlying Mesaverde Group and with the overlying Fox Hills and Lance Formations (Weimer, 1961, p. 22). To the south in Utah, lateral equivalents of the Lewis, Fox Hills, and Lance sequence include the lower parts of the Upper Cretaceous and lower Tertiary Canaan Peak (Kaiparowits Basin), North Horn (adjacent to the San Rafael Swell), and Currant Creek (western Uinta Basin) Formations.

The basal contact of the Mesaverde aquifer with the underlying Mancos confining unit is generally conformable and commonly intertonguing (fig. 25). Locally, in the northern Uncompahgre Uplift and the western part of the Wyoming thrust belt, an unconformity marks the basal contact of the Mesaverde aquifer. In most of the study area, the upper contact of the Mesaverde aquifer is marked by unconformities that separate the Upper Cretaceous rocks from rocks of Tertiary age. Where the top of the Mesaverde aquifer is in the Currant Creek, North Horn, Canaan Peak, or Evanston Formation, the upper boundary of the aquifer, in the hydrologic sense, is located where these strata become less permeable.

The saturated thickness of the Mesaverde aquifer is shown in figure 26. Comparison of the saturated thickness (fig. 26) and the aquifer thickness (pl. 3*E*) indicates that the Mesaverde aquifer is largely saturated throughout the study area. Locally, as in the Wyoming thrust belt area, the aquifer may be partly saturated. Along the

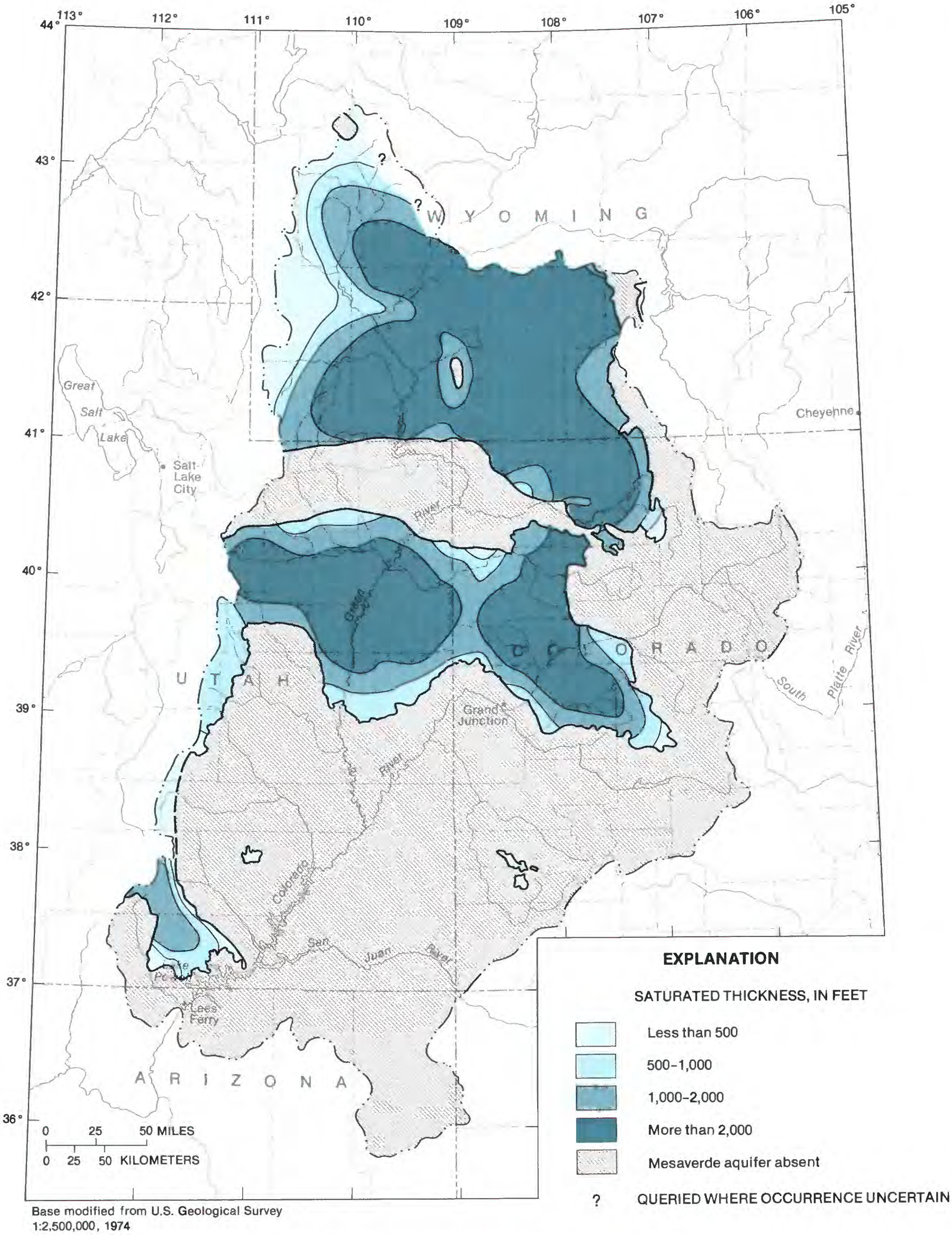


FIGURE 26.—Saturated thickness of the Mesaverde aquifer.

outcrop margins, where the aquifer is commonly less than 100 ft thick, it may be unsaturated.

By comparing the saturated thickness (fig. 26) with the thickness of the overlying rock (fig. 27), it is apparent that in many areas the saturated thickness is more than 2,000 ft and ground-water storage is large; however, in many places the Mesaverde aquifer is overlain by more than 2,000 ft of rock. At depths greater than 2,000 ft, water quality tends to deteriorate and the cost of ground-water development increases. Water in the Mesaverde aquifer is more likely to be developed where the saturated thickness is large and the depth to the aquifer is less than 2,000 ft. The perimeters of the Uinta, Piceance Creek, and Sand Wash Basins and the eastern borders of the Washakie and Great Divide Basins, as well as the Rock Springs Uplift and the Kaiparowits Basin, are possible areas for ground-water development.

Lithologic Character

The rocks of the Mesaverde aquifer represent several 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 of the transgressive-regressive nature of the environments of deposition, the formations that make up the Mesaverde aquifer, and in particular the Mesaverde Group, exhibit complex lateral and vertical gradations and intertonguing (McGookey and others, 1972, p. 212–214).

The lithologic composition of the Mesaverde aquifer is highly variable from formation to formation. The aquifer is composed of conglomerate, sandstone, siltstone, mudstone, claystone, carbonaceous shale, and coal. A lithofacies map of the Mesaverde aquifer was not prepared because detailed stratigraphic compilations were not available. Instead, the general lithology of most of the formations in the Mesaverde aquifer is described below.

The Mesaverde Group in the Rock Springs Uplift area of Wyoming consists of sandstone and shale with interbedded silty sandstone of the Blair Formation, thick coal bed sequences grading into sandstone of the Rock Springs Formation, sandstone with carbonaceous shale of the Ericson Formation, and interbedded sandstone and shale of the Almond Formation. The partly equivalent Adaville, to the west, is composed of interbedded sandstone and siltstone with carbonaceous claystone and coal. In the Rawlins Uplift, the Mesaverde Group is similarly composed of shallow-water marine sandstone and shale to nonmarine sandstone, carbonaceous shale, and coal.

Southward in northwestern Colorado, the Williams Fork and Iles Formations of the Mesaverde Group are predominantly sandstone with interbedded shale and coal beds. Similar lithologies persist westward in the Mesaverde Group and Mesaverde Formation in Utah. In central Utah, the predominant sandstone strata such as the Star Point and Castlegate Sandstones are separated by sequences of interbedded sandstone, shale, and coal of the Blackhawk Formation.

The Straight Cliffs and Wahweap Sandstones and the Kaiparowits Formation of south-central Utah are predominantly fine- to coarse-grained sandstone. The Straight Cliffs includes interbedded shale, mudstone, and major coal beds. The overlying Wahweap contains interbedded sandy shale, and the Kaiparowits includes minor limestone lenses (Gregory and Moore, 1931, p. 100–108). Both the Wahweap and the Kaiparowits lack the coal beds common to the Mesaverde aquifer.

The Toreva and Wepo Formations and the Yale Point Sandstone of Black Mesa Basin consist of fine- to coarse-grained sandstone, and the Wepo also contains interbedded mudstone, siltstone, and coal. These lithologies are typical of the Mesaverde Group rocks in much of the study area.

The strata that overlie the Mesaverde Group and Mesaverde Formation differ from the underlying rocks in that they lack carbonaceous shale and coal beds. The Lewis Shale is chiefly marine shale but contains sandy units. The Fox Hills Sandstone consists of sandstone deposited in a shallow marine environment. Interbedded sandstone and shale make up the overlying Lance Formation.

The lower parts of the Canaan Peak, Currant Creek, and North Horn Formations are somewhat similar in lithology. Conglomerate, sandstone, and shale are common to all three, but they, too, lack the carbonaceous shale and coal of the Mesaverde Group rocks. The Currant Creek Formation, in particular, is extremely well cemented to the extent that fractures commonly break through mineral grains rather than the cement (J.W. Hood, U.S. Geological Survey, retired, written commun., 1986). The lower part of the Evanston Formation includes mudstone, claystone, siltstone, carbonaceous sandstone, and a thick conglomeratic unit (Lines and Glass, 1975, sheet 1).

HYDROLOGIC PROPERTIES OF GEOHYDROLOGIC UNITS AND FLUID CHARACTER

Hydrologic characteristics of the Mesozoic rocks forming the aquifers and confining units of the study area are discussed in the following sections.

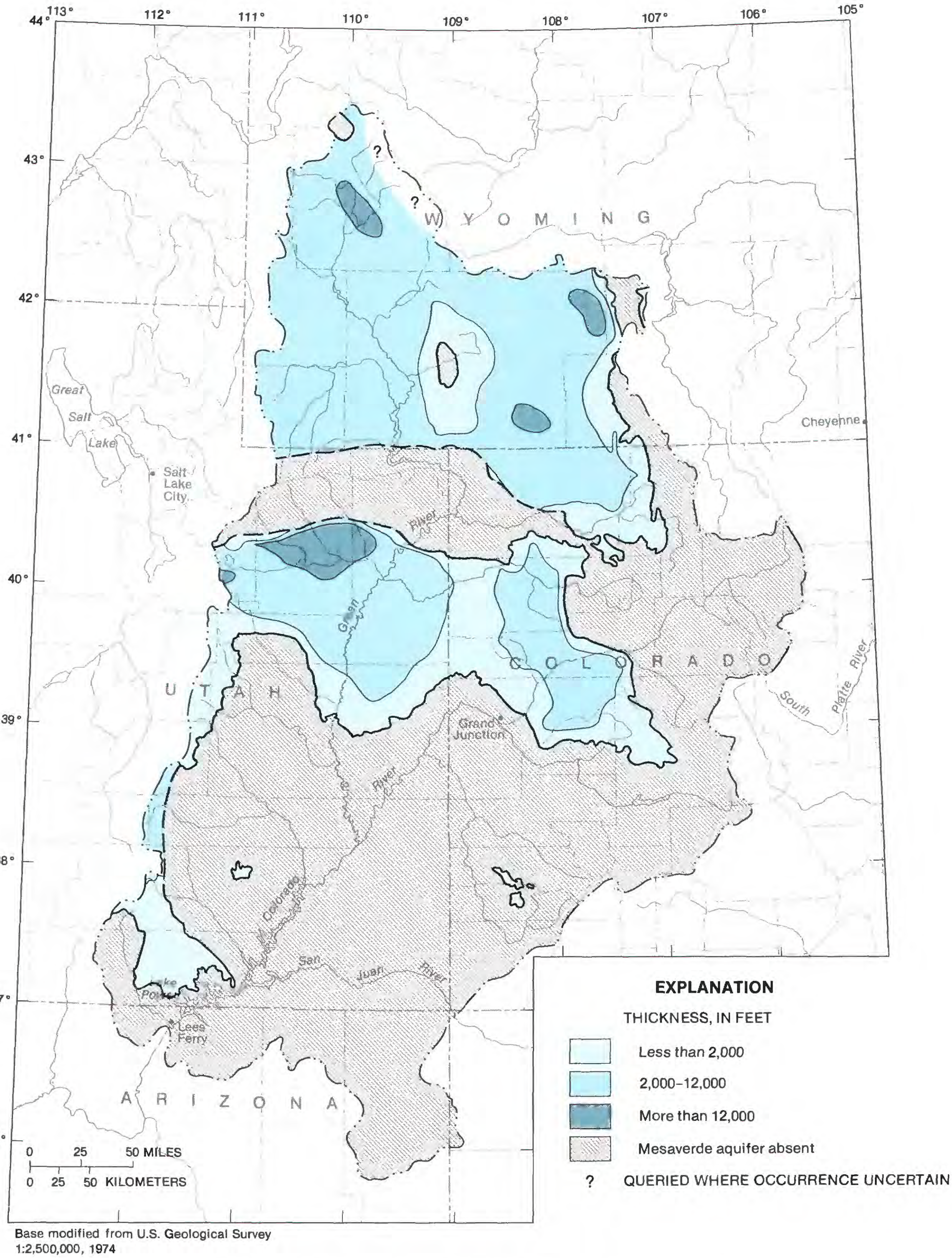


FIGURE 27.—Total thickness of rock overlying the Mesaverde aquifer.

ROCK FABRIC

To describe rock fabric quantitatively, cores of rock from surface outcrops and from wells are customarily analyzed under laboratory conditions to obtain measures that describe the grains, the matrix between the grains, and, most importantly, the void spaces. More than 200 analyses of grain size of Mesozoic rocks from the Upper Colorado River Basin have been reported in published and unpublished reports, and have been compiled by Weigel (1987a). Analyses reported in Jobin (1962) represent the mean values for several samples collected at each site. Most analyses are for sandstones in aquifers and confining units. Most of the sandstone samples from aquifers are categorized as moderately sorted, very fine to fine sand containing less than 10 percent silt or smaller particles (figs. 28*A*, *B*, *C*). The grains are primarily quartz, cemented in varying degrees by silica and carbonate. On average, the Entrada-Preuss, Morrison, and Mesaverde aquifers contain more carbonate minerals than do the Navajo-Nugget and Dakota aquifers (fig. 28*D*).

POROSITY

Porosity, as used in this report, refers to interconnected pore space within a formation. Most workers refer to this as effective porosity. It is measured using a gas to fill the available pore space. Because a gas is able to move through the interconnected pore space with negligible resistance owing to molecular attraction, and because gas does not cause clays to swell, values of effective porosity determined by gas are almost always larger than the porosity values for water. Primary porosity of rock is a result of the original granular arrangement after deposition and consolidation and of subsequent intergranular dissolution or cementation. Secondary porosity results from fracturing of consolidated rocks and dissolution of carbonate rocks. Regionally, fracture openings and solution cavities account for only a small percentage of the void space within the Mesozoic sandstone aquifers, and thus are not meaningful in terms of ground-water storage.

General comparisons of the porosity of the five aquifers measured from 206 minimally weathered rock samples collected near land surface, and from 92 cores taken from wells, are illustrated in figure 29. Samples from the Navajo-Nugget and Entrada-Preuss aquifers most commonly had porosity values ranging from 20 to 30 percent. Most porosity values for the Morrison aquifer ranged between 10 and 20 percent, and most samples from the Dakota and Mesaverde aquifers had porosity values of less than 10 percent. This comparison indicates that the Navajo-Nugget and the Entrada-Preuss aquifers have the largest capacity for storage of water per unit volume of aquifer.

The lateral distribution of porosity measured on 132 samples from the Navajo-Nugget aquifer (fig. 30) indicates that porosity values are largest in the area between the San Rafael Swell and the Circle Cliffs Uplift-Kaiparowits Basin, and along the southern edge of the Uinta Uplift. Values of less than 10 percent were found for samples of the Nugget Sandstone from the Wyoming thrust belt and for one sample of Navajo Sandstone from the southwestern corner of the study area.

The porosity of core samples from a depth of more than 5,000 ft is smaller than that of surface samples and cores from a depth of less than 5,000 ft. The maximum value for a Navajo Sandstone core from more than 5,000 ft was about 16 percent. The largest porosity value for the Mesaverde aquifer, for cores from the Washakie, Uinta, and Piceance Creek Basins at depths of more than 5,000 ft, was about 15 percent.

INTRINSIC PERMEABILITY

Data that could be used to derive intrinsic permeability values for the Mesozoic rocks are not available. Values that approximate the intrinsic permeability were derived from drill-stem tests (Teller and Chafin, 1986) and were converted to hydraulic conductivity for water at 60 °F. Values derived by laboratory analyses that used water at room temperature are also close approximations of the intrinsic permeability. They are discussed in the section on "Hydraulic Conductivity."

In most ground-water systems, the prevailing water temperature is about 60 °F; therefore, hydraulic conductivity determined in a laboratory may be considered a field value. However, in parts of the Upper Colorado River Basin the viscosity of water in aquifers buried in structural basins can be an order of magnitude smaller than water in aquifers near land surface because of great temperature differences. Hydraulic conductivity determined at 60 °F is notably different from hydraulic conductivity measured when the fluid temperatures are as great as 200 °F (fig. 31).

Air permeability in the horizontal direction was determined for 219 samples of Mesozoic rock from the Upper Colorado River Basin. Figure 32 summarizes the air permeability of aquifers and confining units. The smallest mean air permeability values are for samples from the Mancos and Morrison confining units. The mean value of air permeability for the Mesaverde and Dakota aquifers is smaller than that for the other three aquifers, but still it is larger than the values for overlying and underlying confining units. Two of the confining units—the Chinle-Moenkopi and Curtis-Stump confining units—

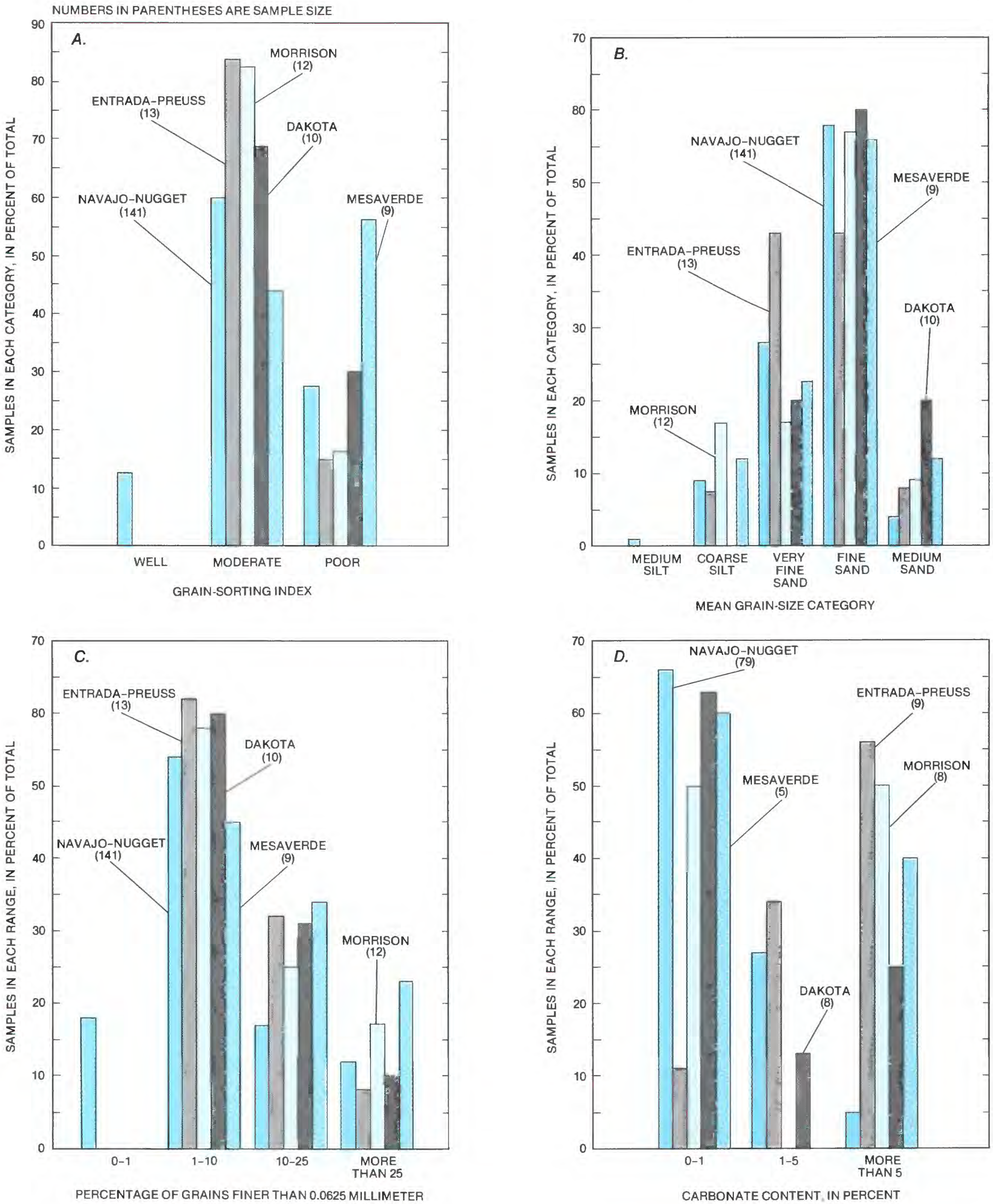


FIGURE 28.—Lithologic characteristics of the Mesozoic rocks forming the aquifers.

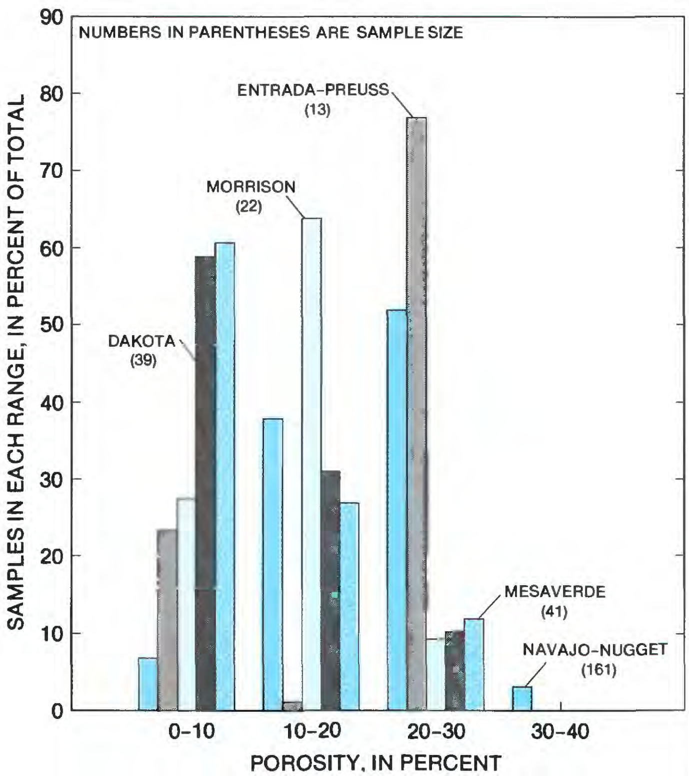


FIGURE 29.—Comparison of porosity distribution in the five aquifers.

have wide ranges of permeability. Permeable sandstone and conglomerate formations occurring locally within these units probably are the cause.

FLUID CHARACTER

The important characteristics of ground water are dynamic viscosity, density, and compressibility. These properties are functions of temperature and pressure, and in most cases they do not vary enough to affect the rate of ground-water movement. However, in the deep structural basins of the Upper Colorado River Basin, water many thousands of feet below land surface is in an environment of much greater pressure and temperature than is water in exposed or less deeply buried formations. The density of freshwater is about 1 gram per milliliter (g/mL), varying slightly with temperature. The density of briny water having 140,000 mg/L of dissolved solids is about 1.1 g/mL, a 10-percent increase.

Water samples from Mesozoic aquifers have dissolved-solids concentrations ranging from 28 to 172,000 mg/L. The difference in density is about 15 percent. At depths where drill-stem tests were conducted, temperatures as high as 295 °F were also recorded, thus changing the viscosity of the water at this depth notably. The viscosity at 212 °F is about 75 percent less than the viscosity at the standard temperature of 60 °F, resulting in a hydraulic conductivity value about 3.5 times larger than that in a 60 °F environment. Movement of ground water under these conditions may be evaluated on the basis of water-level differences only if corrections for density and viscosity are made.

HYDROLOGIC PROPERTIES

HYDRAULIC CONDUCTIVITY

Hydraulic-conductivity values used in defining the aquifers in Mesozoic rocks include values derived from laboratory analyses and analyses of field data.

LABORATORY ANALYSES

During this study 266 hydraulic-conductivity values for Mesozoic rocks, representing aggregate and single rock determinations from laboratory analyses, were examined. The samples analyzed were cores obtained from saturated and unsaturated rock; they represent extremely small parts of the geohydrologic units. Hydraulic-conductivity values determined from these samples do not represent the overall hydrologic effects of many bedding planes, erosional surfaces, and secondary porosity. Measurements were done by several independent testing laboratories. The testing conditions were similar, and values obtained are nearly equal to intrinsic permeability values. Values reported as "water permeabilities" ranged from 0.00001 to 13.7 ft/d for aquifer samples, and from 0.00001 to 10.9 ft/d for confining unit samples. Median values were 0.45 ft/d for aquifers and 0.06 ft/d for confining units. Both "air and water permeabilities" were measured for numerous samples, and a regression relation between air and water permeabilities for Mesozoic sandstones (Weigel, 1987a) was used to increase the number of hydraulic-conductivity values available for interpretation. To avoid confusion, all laboratory hydraulic-conductivity values have been converted to units of feet per day, the unit used for values measured or estimated from field data.

Lateral distributions of hydraulic conductivity based on laboratory-measured permeability to water and on calculated permeability to water derived from measurements of permeability to air for the five aquifers are shown in figures 33–37. Distributions for the confining units could not be shown because the number of hydraulic-conductivity values was too small. Depth of burial is the most notable control on the hydraulic-conductivity values shown in these figures. In general, hydraulic-conductivity values for sandstone cores from wells where the formation is deeply buried are small—generally less than 0.001 ft/d. Values of hydraulic conductivity for cores obtained from near land surface are

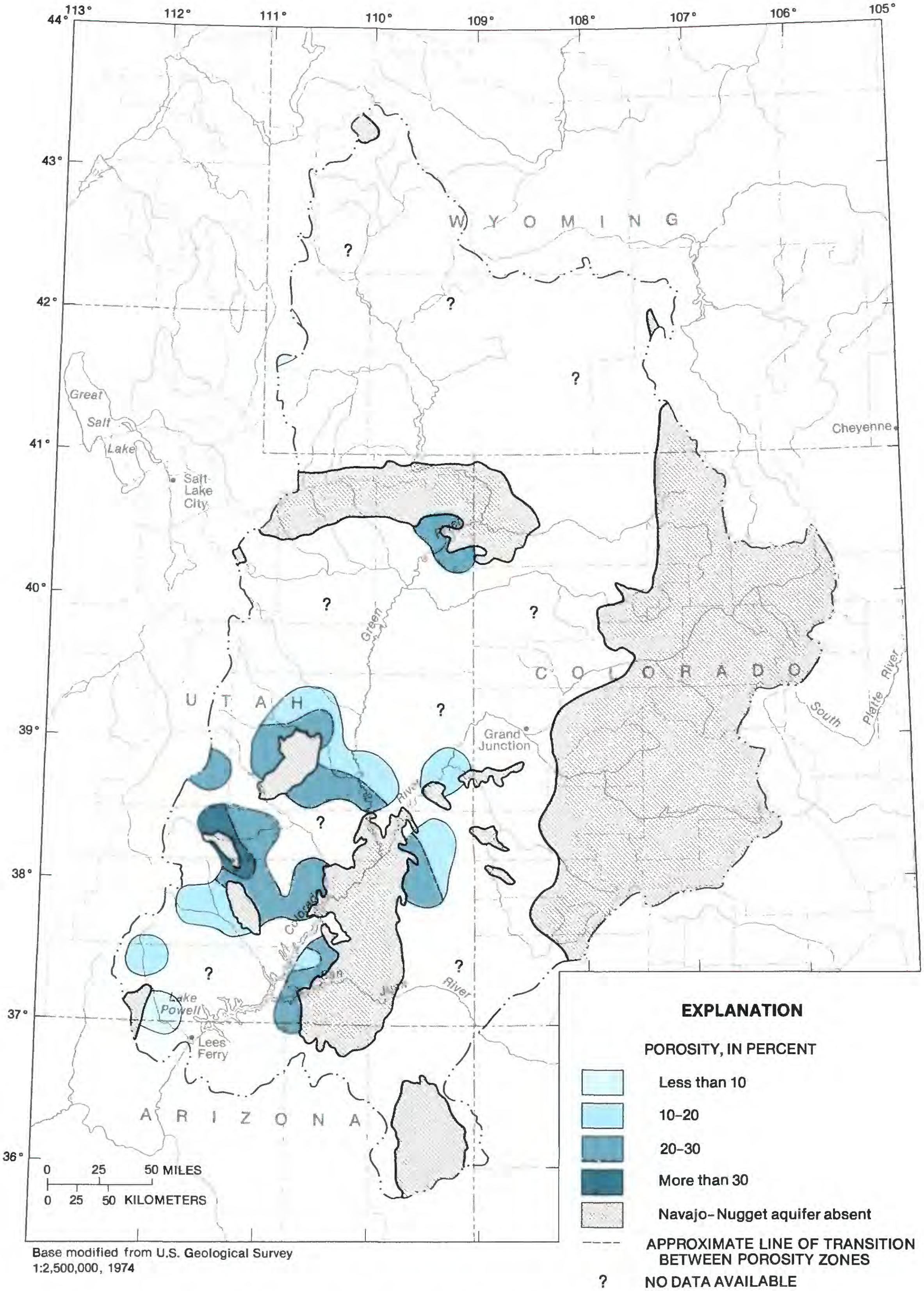


FIGURE 30.—Distribution of porosity values determined from surface samples of the Navajo-Nugget aquifer.

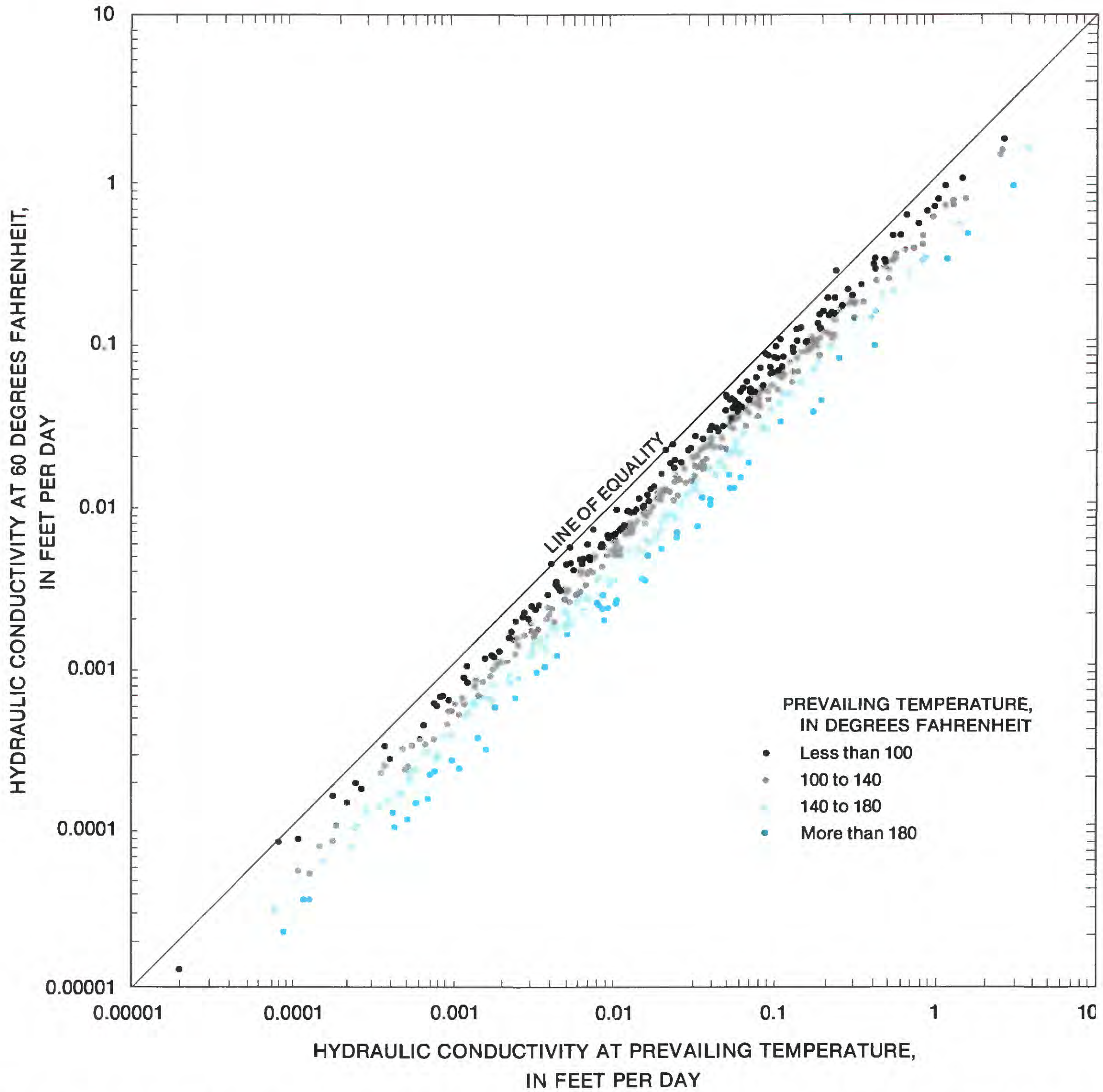


FIGURE 31.—Relation between hydraulic conductivity of Mesozoic aquifers measured at the prevailing temperature and hydraulic conductivity measured at 60 °F.

generally larger—from 0.01 to more than 1.0 ft/d. This difference may also be caused by varying degrees of cementation.

In general, the hydraulic-conductivity distributions do not correlate with lithologic-facies changes. However, certain hydraulic-conductivity values do indicate this relation. Two values of less than 0.01 ft/d for the Navajo-Nugget aquifer northwest of Moab, Utah, coincide with the finer grained facies indicated in figure 11. The small value of hydraulic conductivity for the Entrada-Preuss aquifer at the northern end of the San Rafael Swell represents the "dirty" facies of this aquifer that extends along the western side of the study area (fig. 15).

Hydraulic conductivity is related to porosity. Generally, hydraulic-conductivity values for sandstones measured by laboratory methods increase with increasing porosity (fig. 38). Comparison of the frequency of occurrence (fig. 39) of different ranges of hydraulic conductivity measured in the laboratory indicates that the Morrison aquifer is most likely to have hydraulic-conductivity values larger than 1.0 ft/d. Hydraulic-conductivity values for the Entrada-Preuss and Navajo-Nugget aquifers are likely to be larger than 0.1 ft/d, but fewer values will be larger than 1.0 ft/d. Thirty-five to about fifty percent of the hydraulic-conductivity values for the Dakota and Mesaverde aquifers are likely to be less than 0.1 ft/d, and

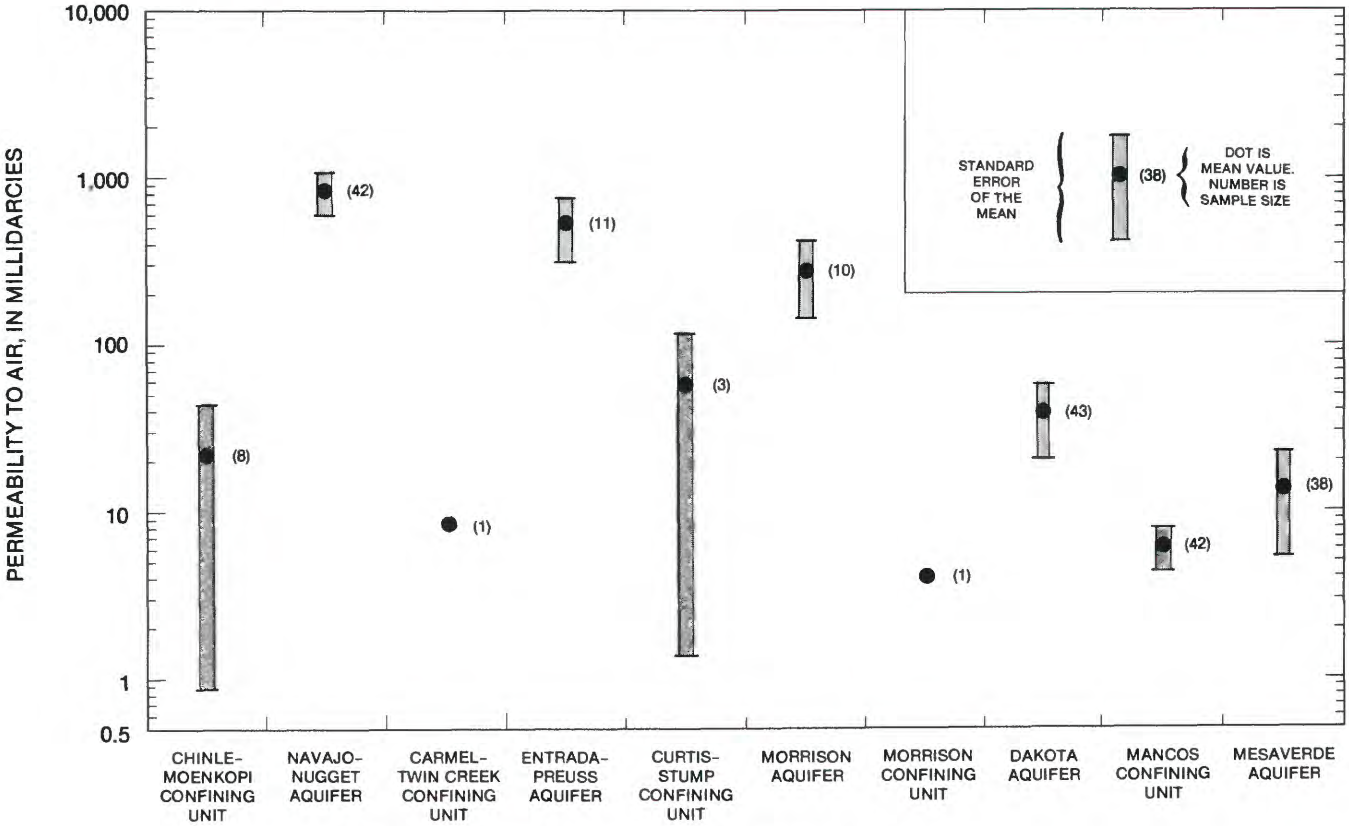


FIGURE 32.—Permeability to air in the Mesozoic geohydrologic units.

a significant number of samples from these two aquifers will have values of less than 0.001 ft/d.

The 95-percent confidence interval of hydraulic-conductivity values for sandstone samples from five aquifers and two confining units are shown in figure 40. The intervals for the Navajo-Nugget and Dakota aquifers and the Ferron Sandstone Member of the Mancos confining unit are relatively small, indicating a regional consistency in the properties that govern hydraulic conductivity. Intervals for the Chinle-Moenkopi confining unit and the Entrada-Preuss, Morrison, and Mesaverde aquifers are, by comparison, large, indicating greater regional differences in these geohydrologic properties. Permeability maps by Jobin (1962) show a similar magnitude of variation in the same geohydrologic units.

The link between hydraulic conductivity and the various measures of grain size and shape is tentative. Experimental values for intrinsic permeability of unconsolidated sand (Masch and Denny, 1966) indicate that hydraulic conductivity increases with increasing median grain size and with sorting of the grains. For consolidated sandstones that form the Mesozoic aquifers in the Upper Colorado River Basin, this relationship is affected by the lithification process. If the mean grain sizes of all analyzed samples are grouped according to the Wentworth Size Classification (Udden, 1914; Wentworth, 1922) and the average grain size for each group is plotted against the corresponding average hydraulic conductivity for each group (fig. 41*A*), an increase in hydraulic conductivity is seen as size class changes from coarse silt to fine sand. However, sandstones having the largest mean grain size do not fit the pattern by having the largest mean hydraulic-conductivity values. This condition is probably related to the degree of cementation in the coarser grained sandstones in the study area and does not apply to sandstones in general. Similar nonlinear results are evident when hydraulic conductivity is compared with the other grain-size parameters, sorting, skewness, and kurtosis; however, the greatest values of hydraulic conductivity occur as expected, when the skewness is nearly symmetrical (figs. 41*B*, *C*, *D*). In general, the Mesozoic rocks in the Upper Colorado River Basin tend to have the largest hydraulic-conductivity values if they are moderately well to well sorted, do not have excess fine- or coarse-grained material, and have a normal to flat rather than a peaked grain-size distribution.

FIELD-DATA ANALYSES

Hydraulic-conductivity values from field data for aquifers in Mesozoic rocks of the Upper Colorado River Basin have been derived using three different methods—

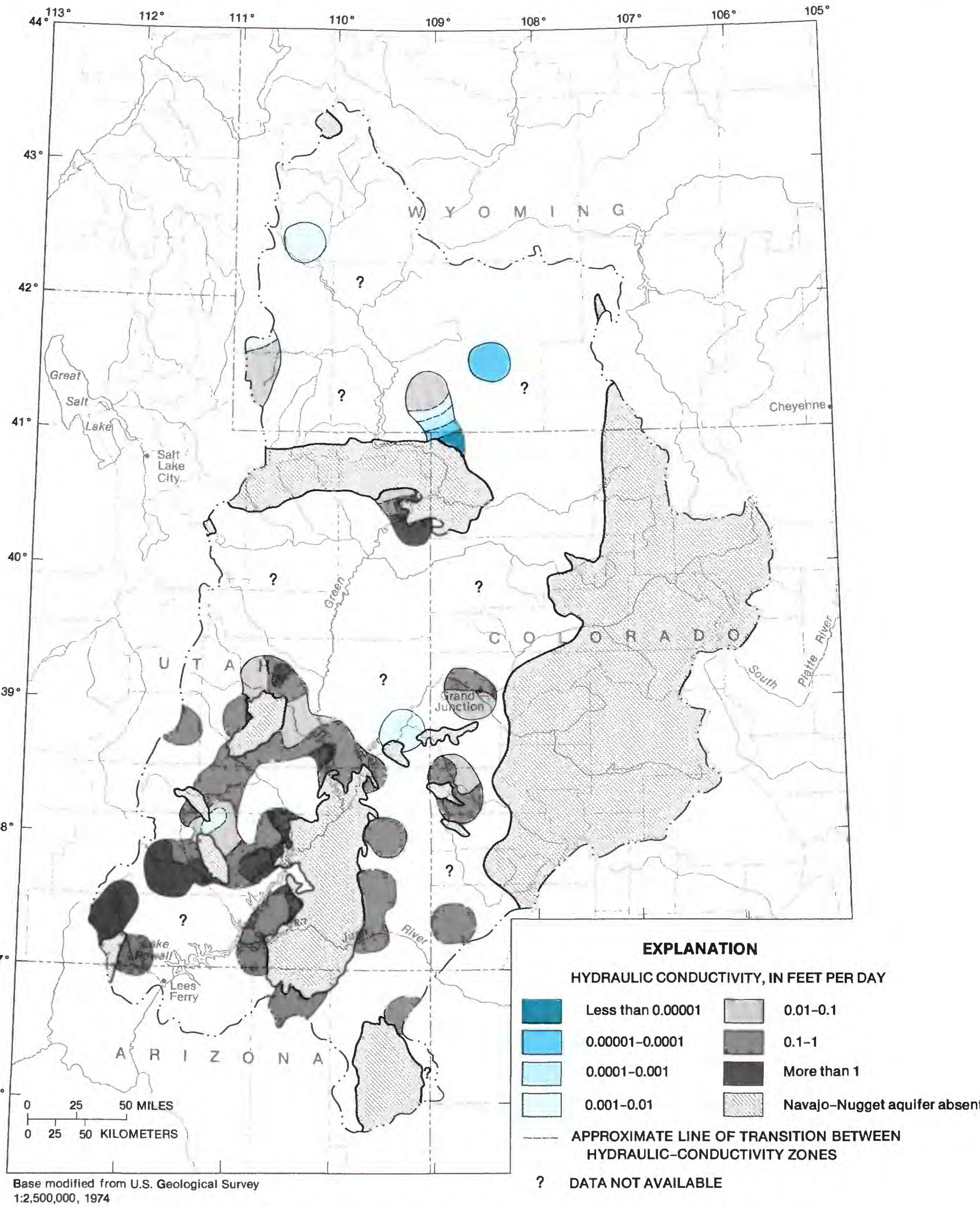


FIGURE 33.—Distribution of laboratory hydraulic-conductivity values in the Navajo-Nugget aquifer.

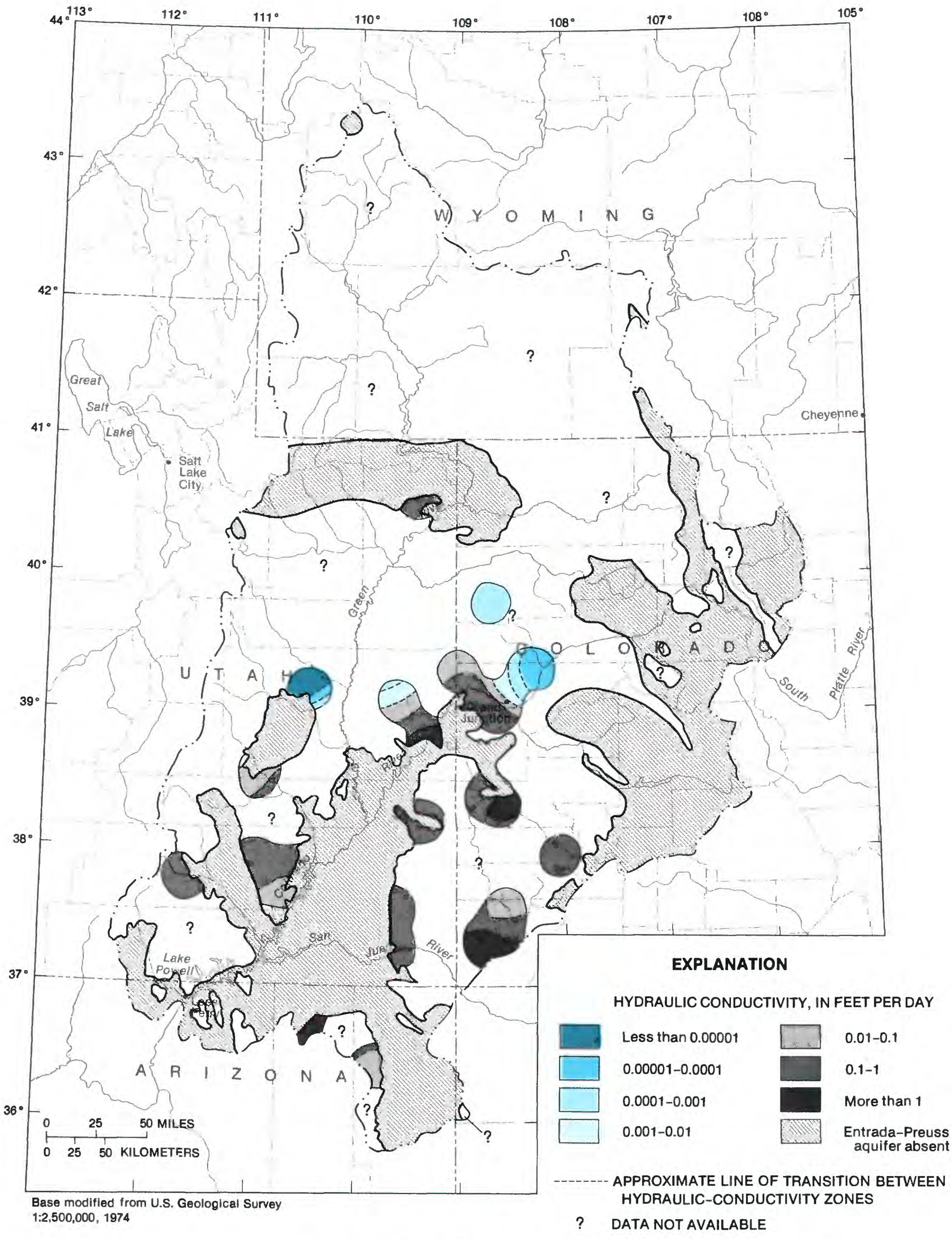


FIGURE 34.—Distribution of laboratory hydraulic-conductivity values in the Entrada-Preuss aquifer.

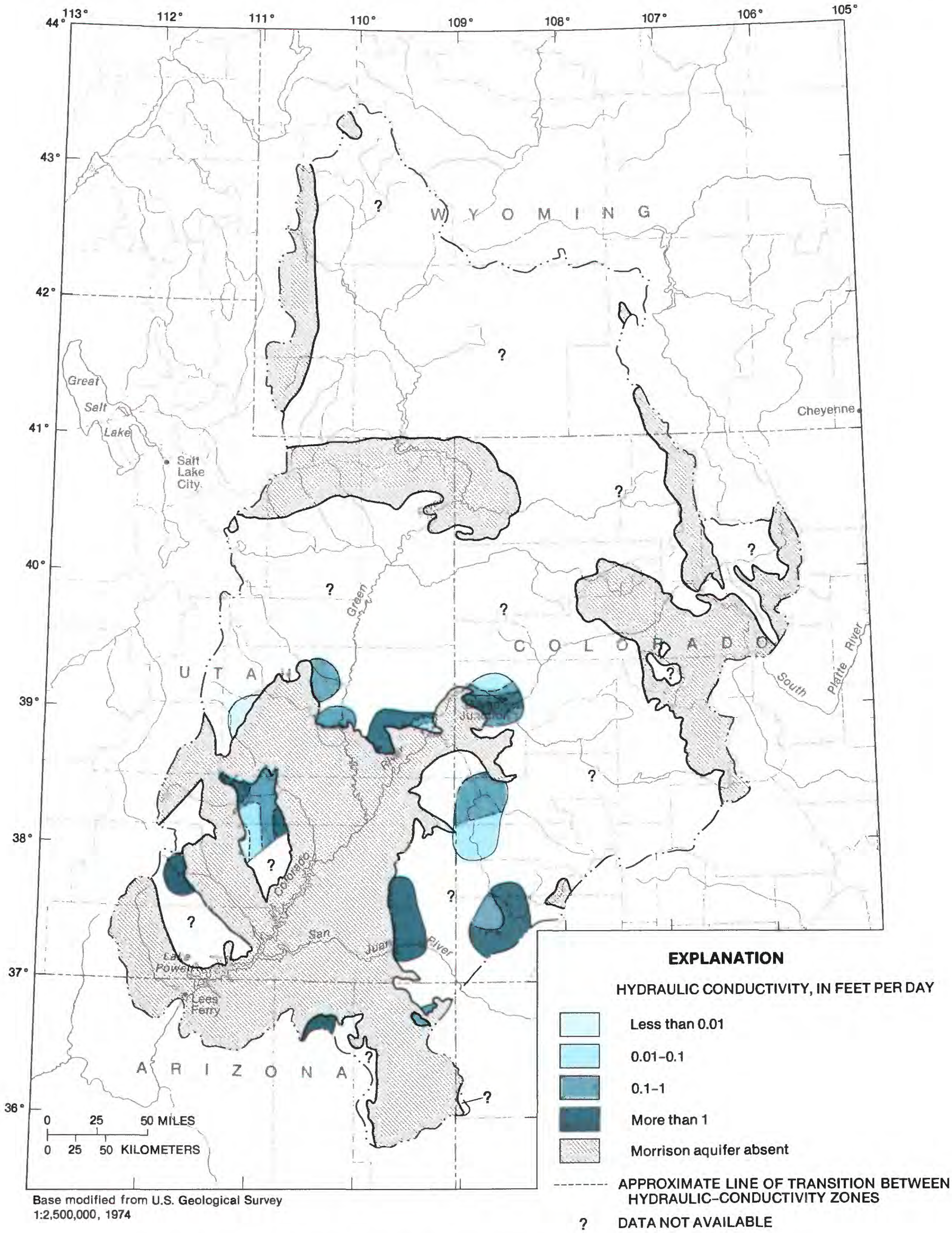


FIGURE 35.—Distribution of laboratory hydraulic-conductivity values in the Morrison aquifer.

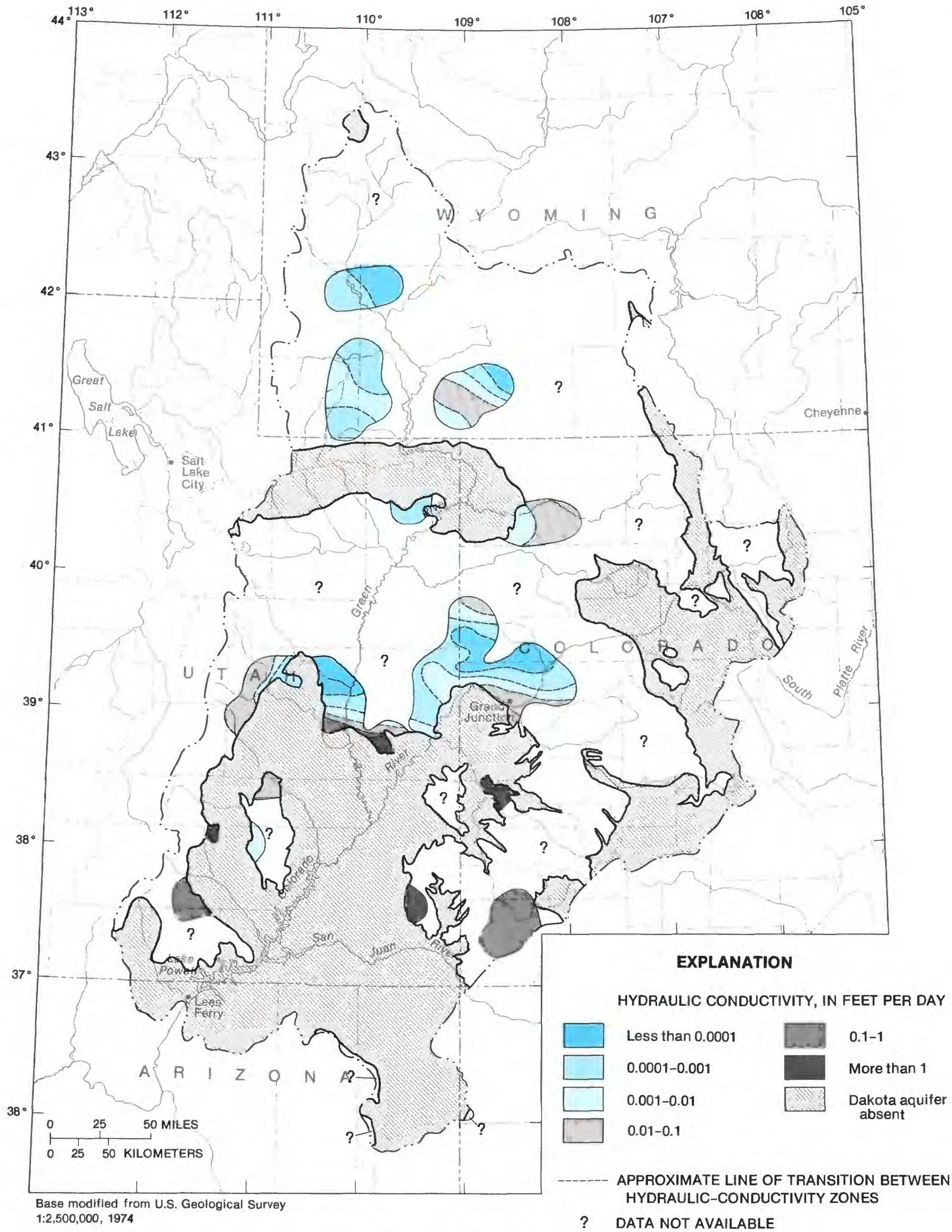


FIGURE 36.—Distribution of laboratory hydraulic-conductivity values in the Dakota aquifer.

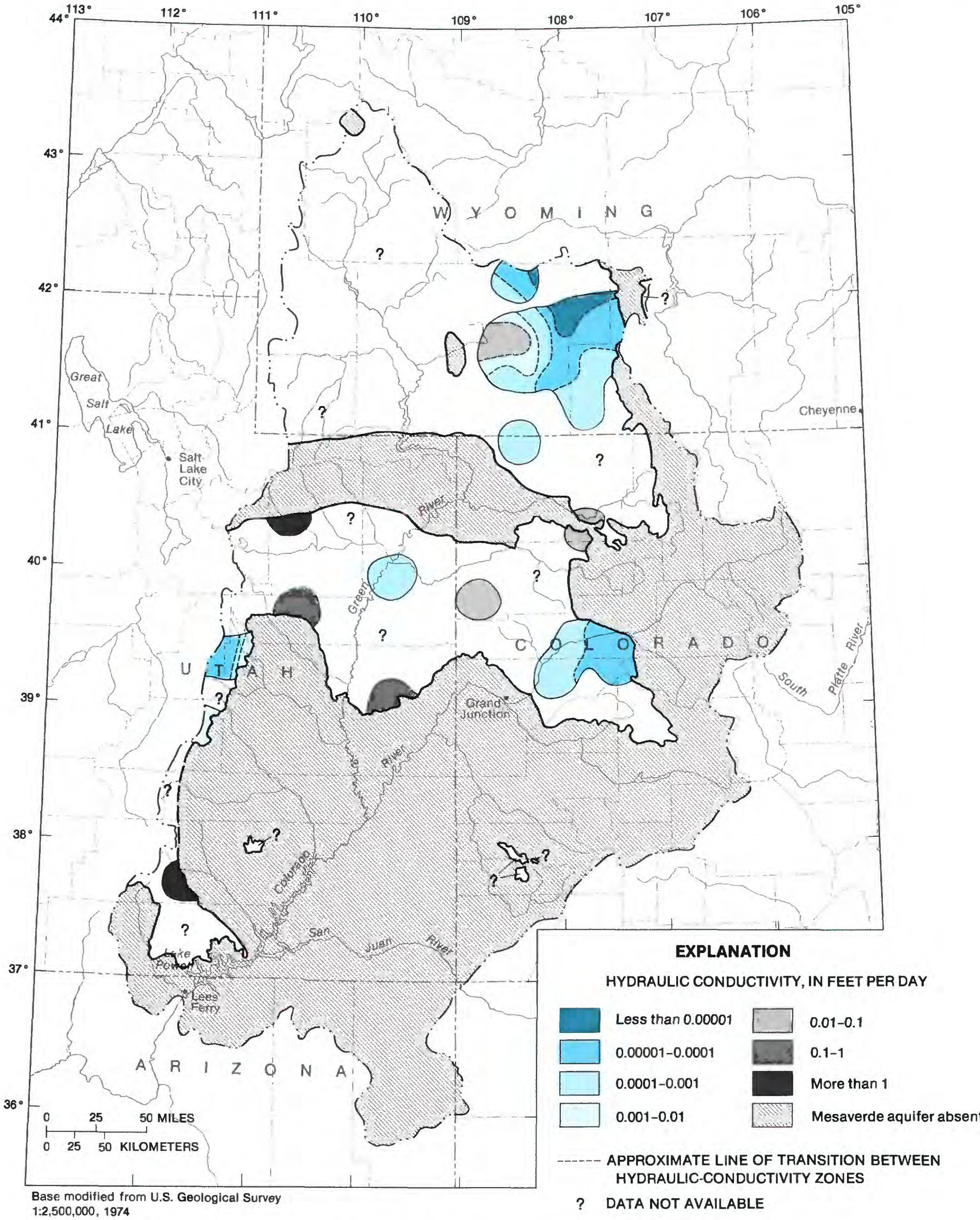


FIGURE 37.—Distribution of laboratory hydraulic-conductivity values in the Mesaverde aquifer.

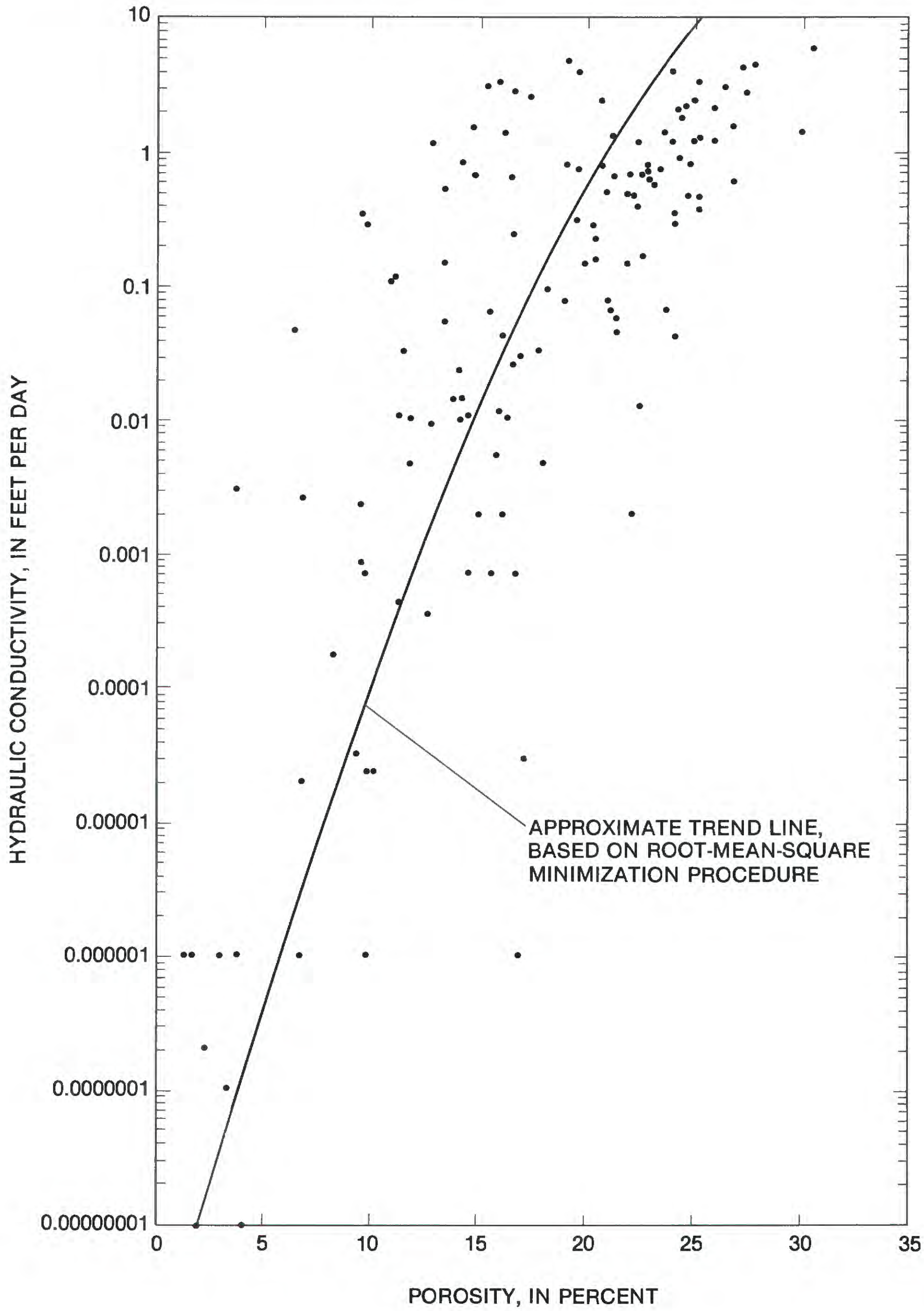


FIGURE 38.—Relation between porosity and hydraulic conductivity in Mesozoic rocks.

aquifer tests, drill-stem tests, and specific-capacity tests. The methods analyze part of an aquifer system surrounding a well, or an array of wells, that penetrates the aquifer. Each method has advantages and disadvantages in terms of cost and reliability of results. Results from tests concerning aquifers and confining units in Mesozoic rocks are presented in Weigel (1987a); site information can be obtained from that publication.

Transmissivity values calculated from the results of aquifer tests that last several weeks are the most reliable data for determining hydraulic-conductivity values. However, because these tests usually represent the part of an aquifer that has the largest water-yielding capabilities (this being only a small part of a regional aquifer), hydraulic-conductivity values derived from these tests may not represent average regional values. The largest

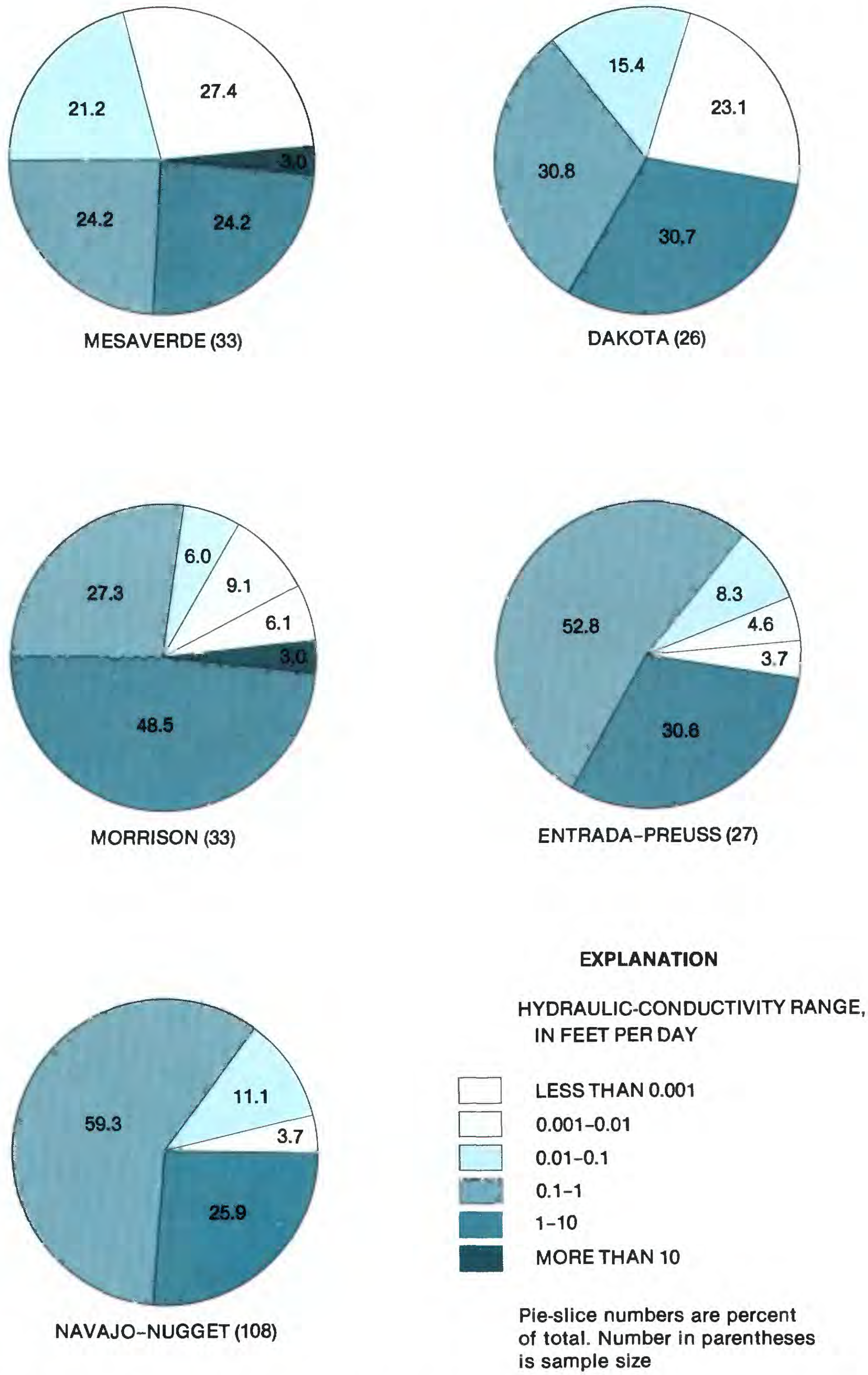


FIGURE 39.—Percentage of laboratory-analyzed samples having hydraulic-conductivity values in the indicated ranges for the five aquifers.

value of hydraulic conductivity derived from an aquifer test is 88 ft/d for a 44-ft interval of fractured Navajo Sandstone. More commonly, values for parts of the Entrada, Glen Canyon, and Navajo Sandstones are between 0.1 and 1.0 ft/d. Transmissivity values from these tests are presented in figure 42.

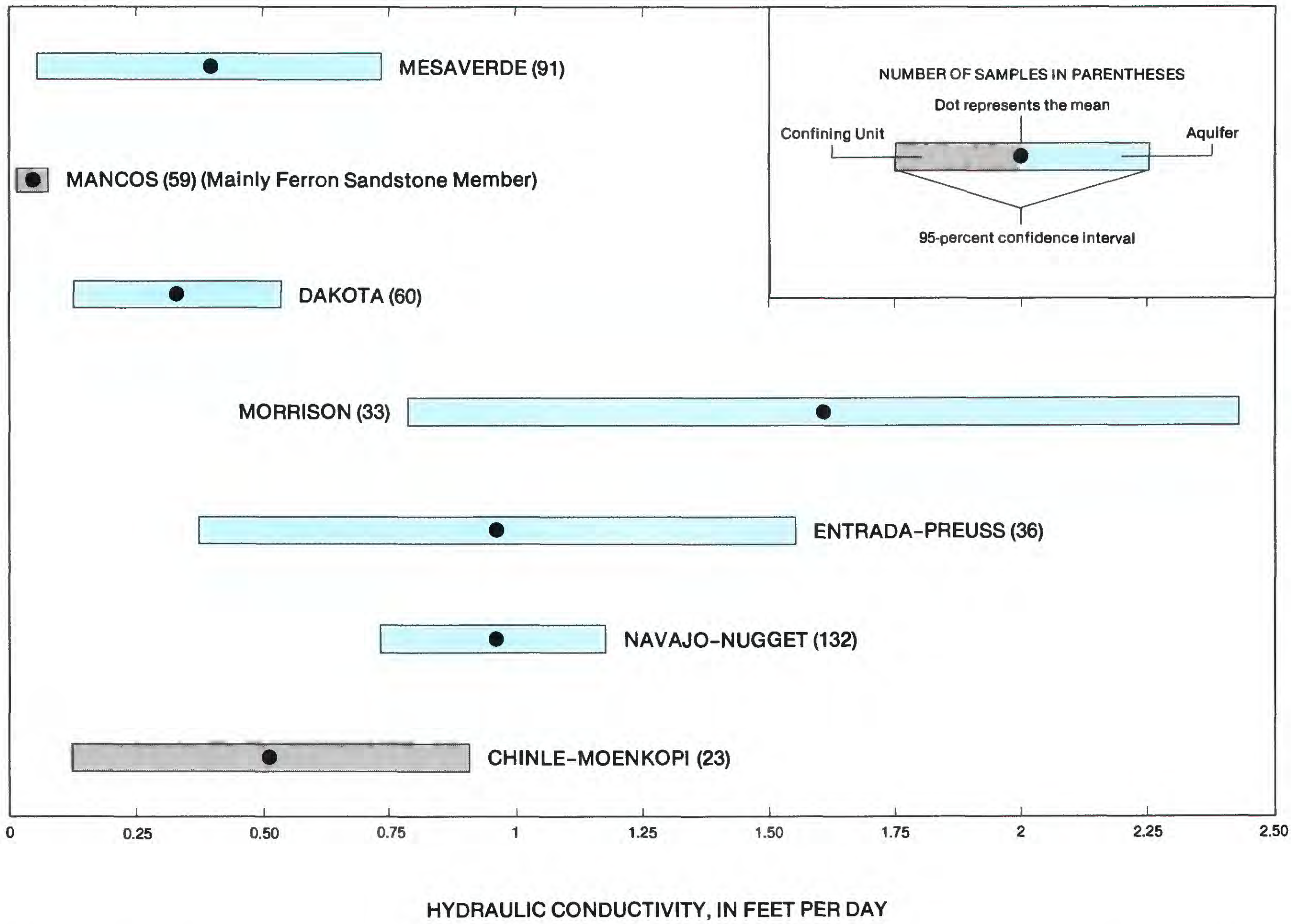


FIGURE 40.—Ninety-five percent confidence interval for hydraulic conductivity determined from laboratory analyses of water-yielding formations found in 7 of the 10 geohydrologic units.

Drill-stem tests are performed by the petroleum industry on rocks that are possible sources of oil or gas. Results of these tests can be used to calculate equivalent freshwater heads and hydraulic-conductivity values. The most reliable of these values are determined from analysis of a graph of pressure buildup in a drill stem caused by formation pressures (Bredehoeft, 1965, p. 35). Other analyses that use only initial and final pressures (Weigel, 1987a) are not as reliable, but they do provide useful information about the relative distribution of hydraulic-conductivity values.

Values of hydraulic conductivity calculated using drill-stem test results are slightly smaller than values determined from laboratory tests. From 620 drill-stem test analyses of Mesozoic rocks, the maximum value of hydraulic conductivity was 3.7 ft/d for the Navajo Sandstone. The median was 0.021 ft/d for aquifers and 0.017 ft/d for confining units.

Possible reasons for these small values involve the physical environment of the formation itself and certain characteristics of the tests. Most drill-stem tests are conducted on deeply buried formations that have been subjected to tremendous pressure from the weight of overlying rocks. The formation may have long since been compressed, and pore space considerably decreased, compared with identical rocks subjected to little or no overburden pressure. Drill-stem tests are characteristically 1 to 2 hours long. This short duration allows only a small part of the formation adjacent to the well bore to be tested. Widely spaced fractures are not likely to be incorporated in test results. Perforations in the casing commonly are inadequate, and the wall of the well bore, if not cased, generally is contaminated with drilling mud, preventing a free flow of fluid from the formation to the borehole walls.

The distributions of hydraulic conductivity estimated from drill-stem tests for the Dakota aquifer are shown in figure 43 and for the Mesaverde aquifer in figure 44. Both maps represent hydraulic-conductivity values for the aquifers where they are deeply buried. The largest

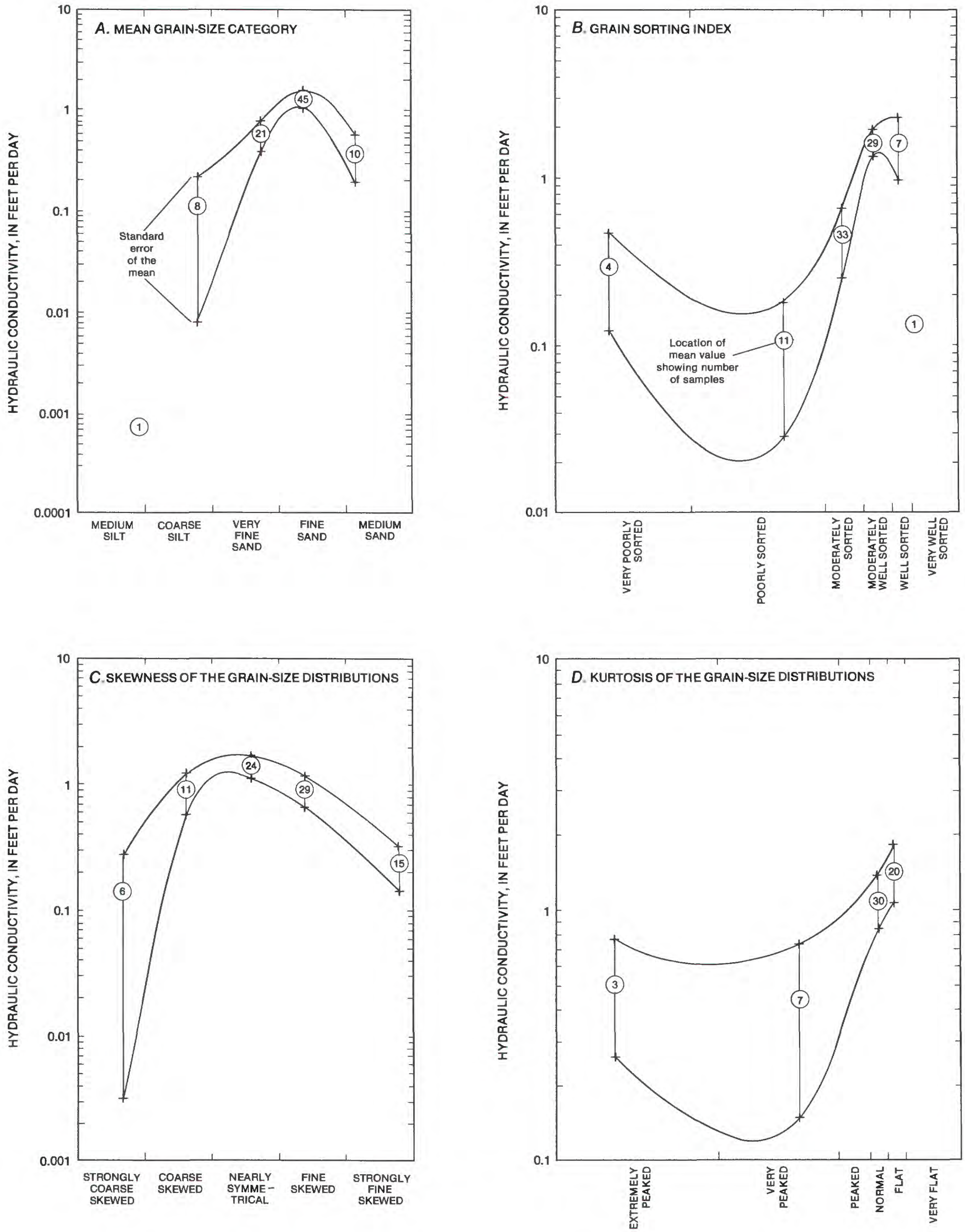


FIGURE 41.—Variation of the mean hydraulic conductivity with mean grain size, sorting, skewness, and kurtosis for Mesozoic sandstones.

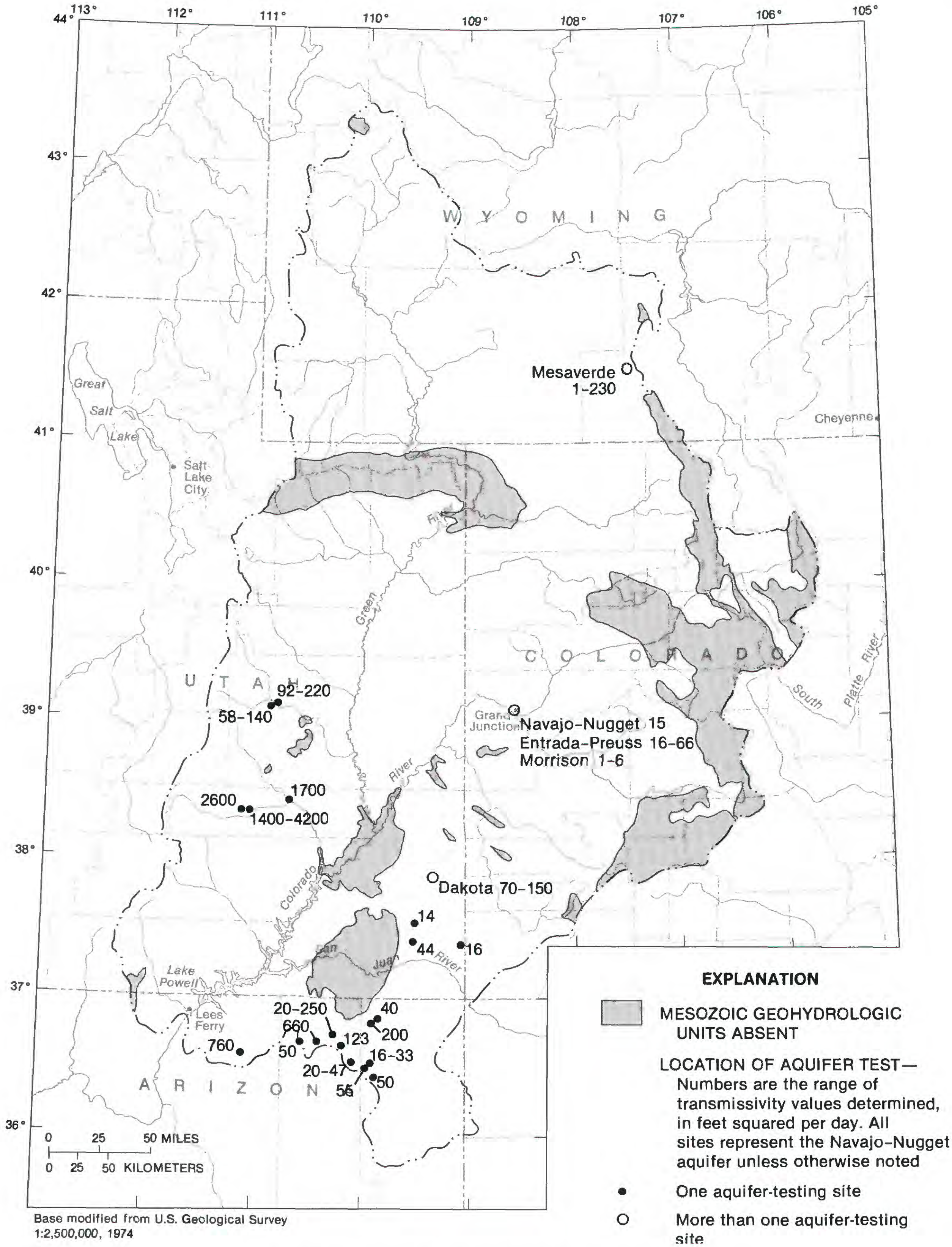


FIGURE 42.—Transmissivity values derived from aquifer tests.

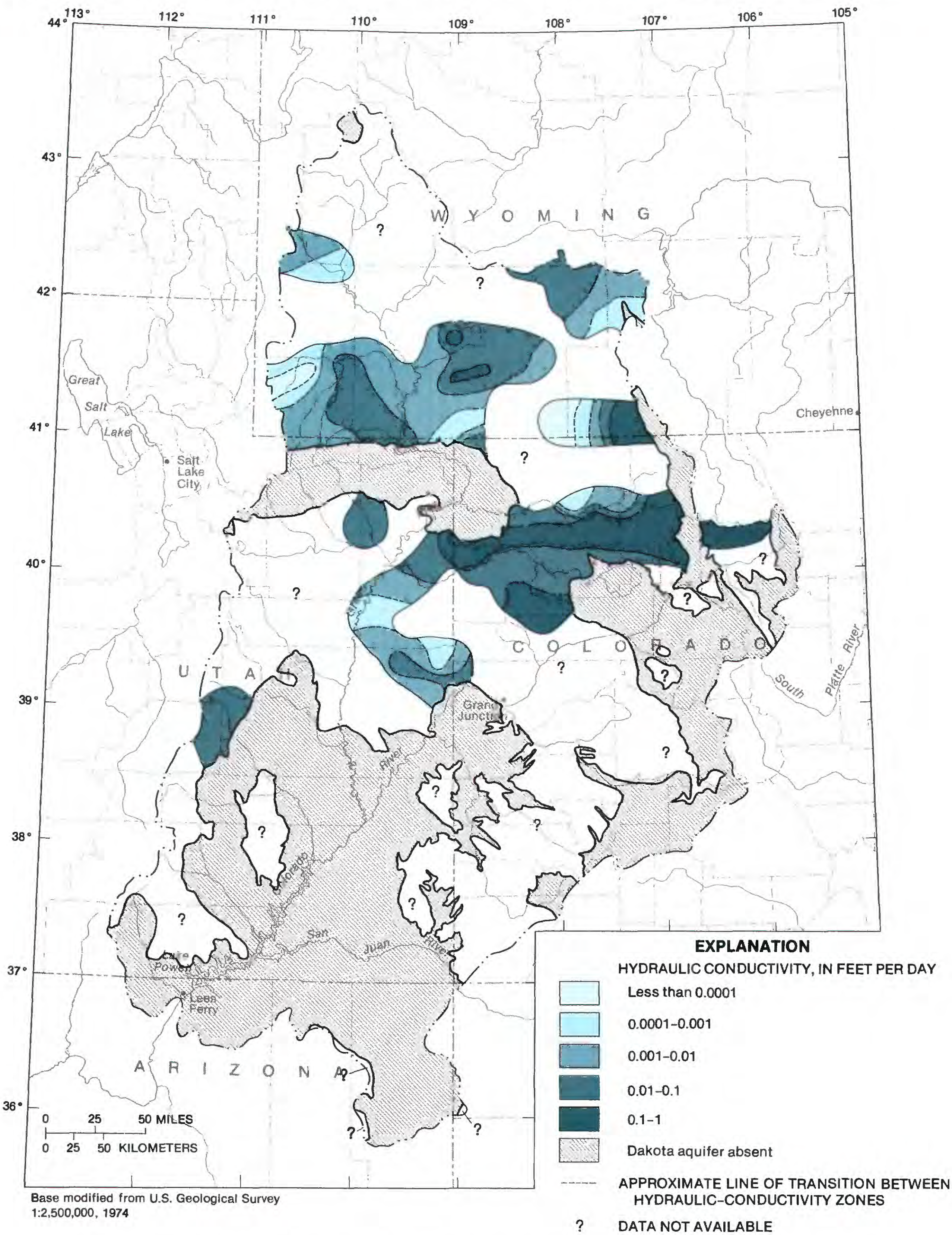


FIGURE 43.—Distribution of hydraulic-conductivity values derived from drill-stem tests in the Dakota aquifer.

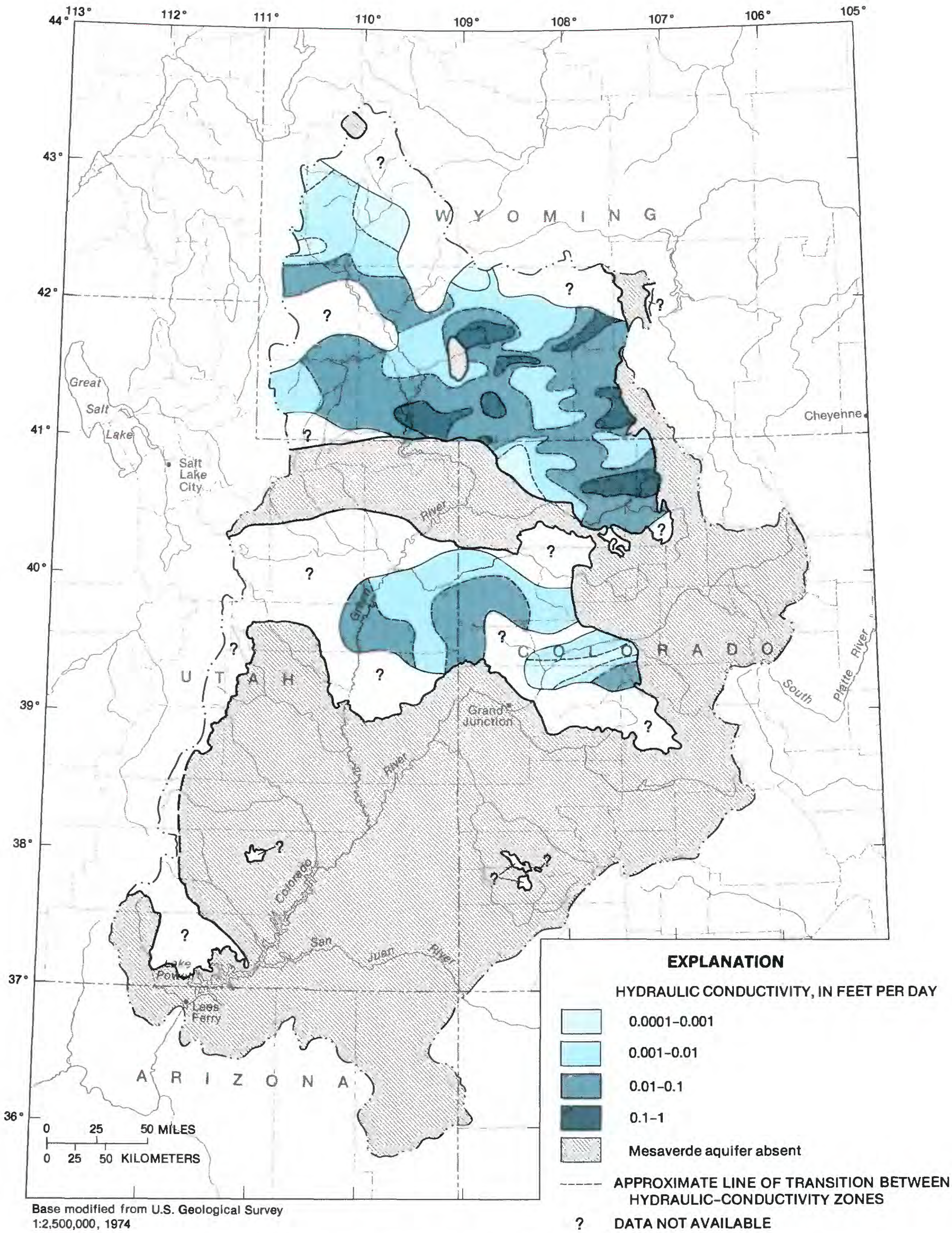


FIGURE 44.—Distribution of hydraulic-conductivity values derived from drill-stem tests in the Mesaverde aquifer.

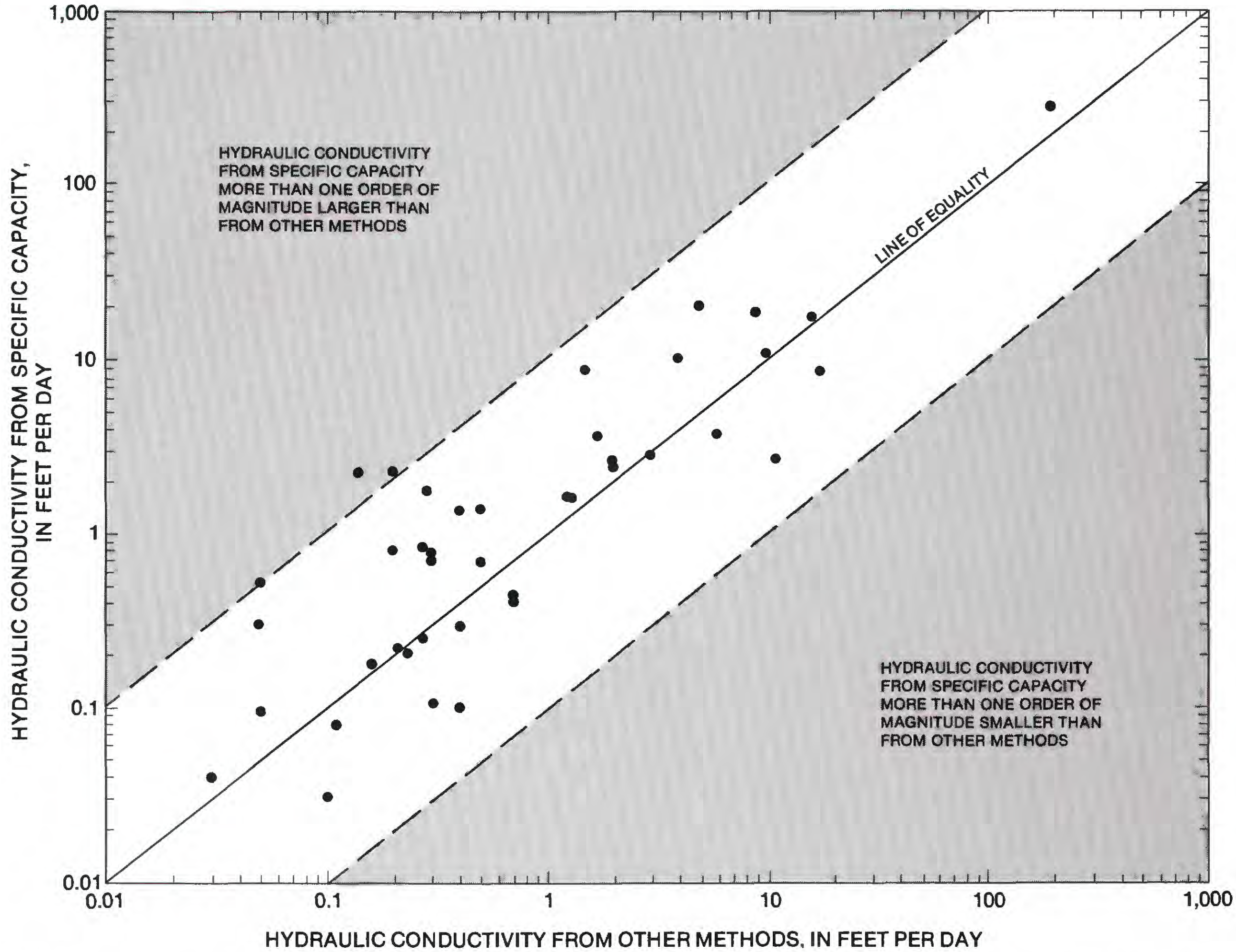


FIGURE 45.—Relation between hydraulic-conductivity values generated from specific-capacity analyses and those obtained using aquifer, drill-stem, and laboratory tests in Mesozoic rocks.

values in the Dakota aquifer are in the less deeply buried rocks in the area that divides the Washakie and Piceance Creek Basins, and in the general area of the Rock Springs Uplift. Values of less than 0.0001 ft/d were determined for deeply buried rocks in the Washakie Basin and in the Wyoming thrust belt. The distribution of hydraulic-conductivity values for the Mesaverde aquifer is less dependent on the thickness of the overburden. The largest values are north and southwest of the Rock Springs Uplift and in localized areas within the Washakie, Great Divide, and Sand Wash Basins. Hydraulic-conductivity values for the Mesaverde aquifer in the southern half of the Green River Basin are, in general, larger than values in the northern half of that basin. In the area between the Piceance Creek and Uinta Basins, hydraulic-conductivity values for the Mesaverde aquifer are slightly larger than values in the deepest part of these basins. Drill-stem test results for the Navajo-Nugget, Entrada-Preuss, and Morrison aquifers were too meager to illustrate.

Estimates of hydraulic conductivity for aquifers that are not deeply buried by younger rocks were obtained from transmissivity values calculated from the specific capacity of water wells using a computer adaptation of the approach described by Theis and others (1963). The method and calculated values for all Mesozoic geohydrologic units in the Upper Colorado River Basin are presented in Weigel (1987a).

Comparison of values derived from specific capacities with values determined from aquifer and laboratory tests shows that values derived from specific capacities are usually within about one order of magnitude of values derived by the other methods (fig. 45). Values derived from specific capacity ranged from 0.001 to 940 ft/d. The median value was 1.31 ft/d for aquifers and 1.26 ft/d for confining units.

Distributions of hydraulic conductivity for the Navajo-Nugget, Dakota, and Mesaverde aquifers derived from specific capacity are shown in figures 46–48. Data were insufficient to allow mapping of the distribution of hydraulic conductivity in the other two aquifers or in any of the confining units. The largest values of hydraulic conductivity for the Navajo-Nugget aquifer are near Moab, Utah, southwest of Grand Junction, Colo., and northwest of Lees Ferry, Ariz., and the smallest values for the unit are in the Piceance Creek Basin and in northeastern Arizona. Zones of largest values for the Dakota aquifer are east of Monticello, Utah, southwest of Montrose, Colo., and near Bluff, Utah, and zones of smallest values are north of Montrose and Gunnison, Colo. Zones of large hydraulic-conductivity values for the Mesaverde aquifer are near the western border of the study area northwest of Price, Utah, in the Rock Springs Uplift, and north of Meeker, Colo. Where the aquifers are thin, the degree of fracturing is a principal factor governing these hydraulic-conductivity distributions.

Comparison of Analyses

The ranges of hydraulic-conductivity values obtained from laboratory analyses and from the three types of field analyses differ (fig. 49) because (1) the test environments for the four analyses are different and (2) the volumes of aquifers or confining units tested are different. Hydraulic-conductivity values derived from laboratory analyses and drill-stem tests are smallest because even the localized effects of fractures are not accounted for in tests for which the samples are small, the test period is short, the test interval is compacted by overlying rock, or the walls of the tested interval are plugged with drilling mud. Hydraulic-conductivity values derived from single- or multiple-well pumping tests are larger because generally the tests are performed in developed water wells at shallow depths for longer time periods. This allows a larger volume of aquifer to be tested, thus allowing the effects of fractures, common at shallow depths near faults and axes of folds, to be incorporated in the results.

TRANSMISSIVITY

In general, aquifers in Mesozoic rocks of the Upper Colorado River Basin are characterized by transmissivity values ranging from about 5 to about 5,000 feet squared per day (ft^2/d). Distributions of transmissivity for each of the aquifers were constructed by mapping the product of saturated thicknesses and hydraulic-conductivity values generalized for each grid block of a 15- by 15-minute (about 17 mi high and 14 mi wide) grid. Hydraulic-conductivity values derived by all methods described previously were used. The distribution for areas for which no data were available was estimated on the basis of lithologic descriptions, thickness of overlying sediments, and values from similar areas for which data were available. As a result, the transmissivity distributions (figs. 50–54) are extremely generalized. They indicate regional variations in transmissivity that parallel variations in saturated thickness, lithologic character, and thickness of overlying sediments. Transmissivity values for aquifers buried in structural basins are small because hydraulic-conductivity values are decreased by the pressure of thick overlying sediments. Transmissivity values for aquifers that are exposed are large where saturated thicknesses are large.

The distributions of transmissivities indicate that the potential for ground-water development in Mesozoic rocks on a regional scale is relatively poor in all five aquifers. The Navajo-Nugget aquifer has the largest transmissivity values because of its large thickness. Locally, because of intense fracturing, transmissivity values in any of the aquifers may exceed the values shown in figures 50–54 by several orders of magnitude, but these localized large-transmissivity areas do not greatly affect regional ground-water movement through the aquifers.

STORAGE COEFFICIENT

Based on the available information, estimated values of storage coefficient, specific yield (unconfined Navajo-Nugget aquifer only), and porosity for the five aquifers in Mesozoic rocks are summarized in table 2. For the Mesaverde and Navajo-Nugget aquifers, storage coefficients may be larger than 10^{-3} in many areas where saturated thickness exceeds 1,000 ft.

As Lohman (1972, p. 8) suggests, specific storage of confined aquifers is approximately equal to 1×10^{-6} per foot of saturated thickness. Therefore, for the areas of confined aquifers the storage coefficient may be estimated approximately by 1×10^{-6} times the saturated thickness from maps in figures 9, 13, 18, 22, and 26. However, in areas of unconfined aquifers the specific yield may be estimated approximately as one-half of the porosity for rocks having large interconnected pore spaces, or less for rocks having small pore spaces.

THE GROUND-WATER FLOW SYSTEM

RECHARGE

Natural recharge to the ground-water system in the Mesozoic rocks of the Upper Colorado River Basin originates from (1) infiltration of precipitation (including snowmelt) through the unsaturated zone to the water table (fig. 55*A*), (2) infiltration of streamflow from stream channels into the zone of saturation (fig. 55*B*),

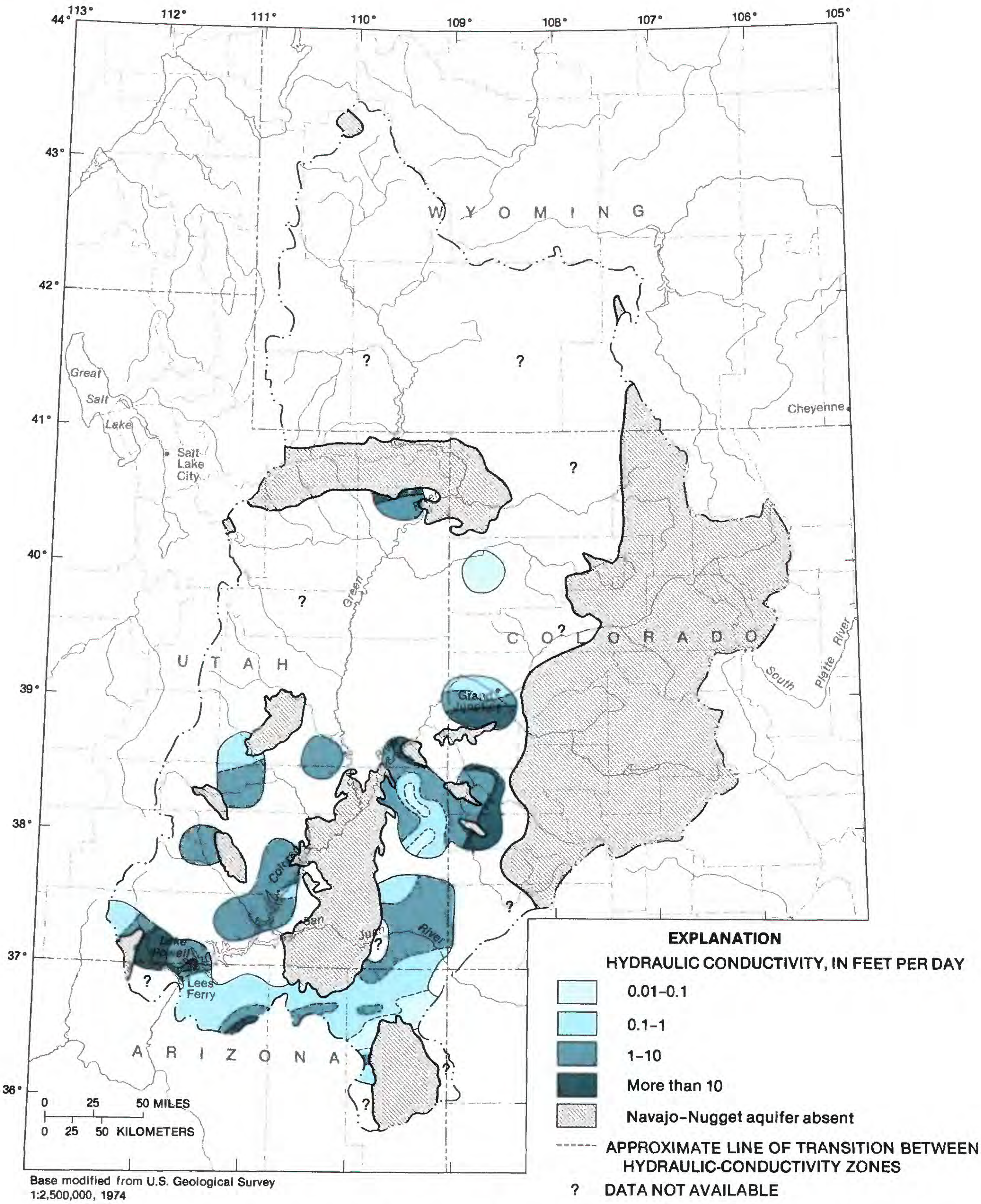


FIGURE 46.—Distribution of hydraulic-conductivity values derived from the specific capacity of wells in the Navajo-Nugget aquifer.

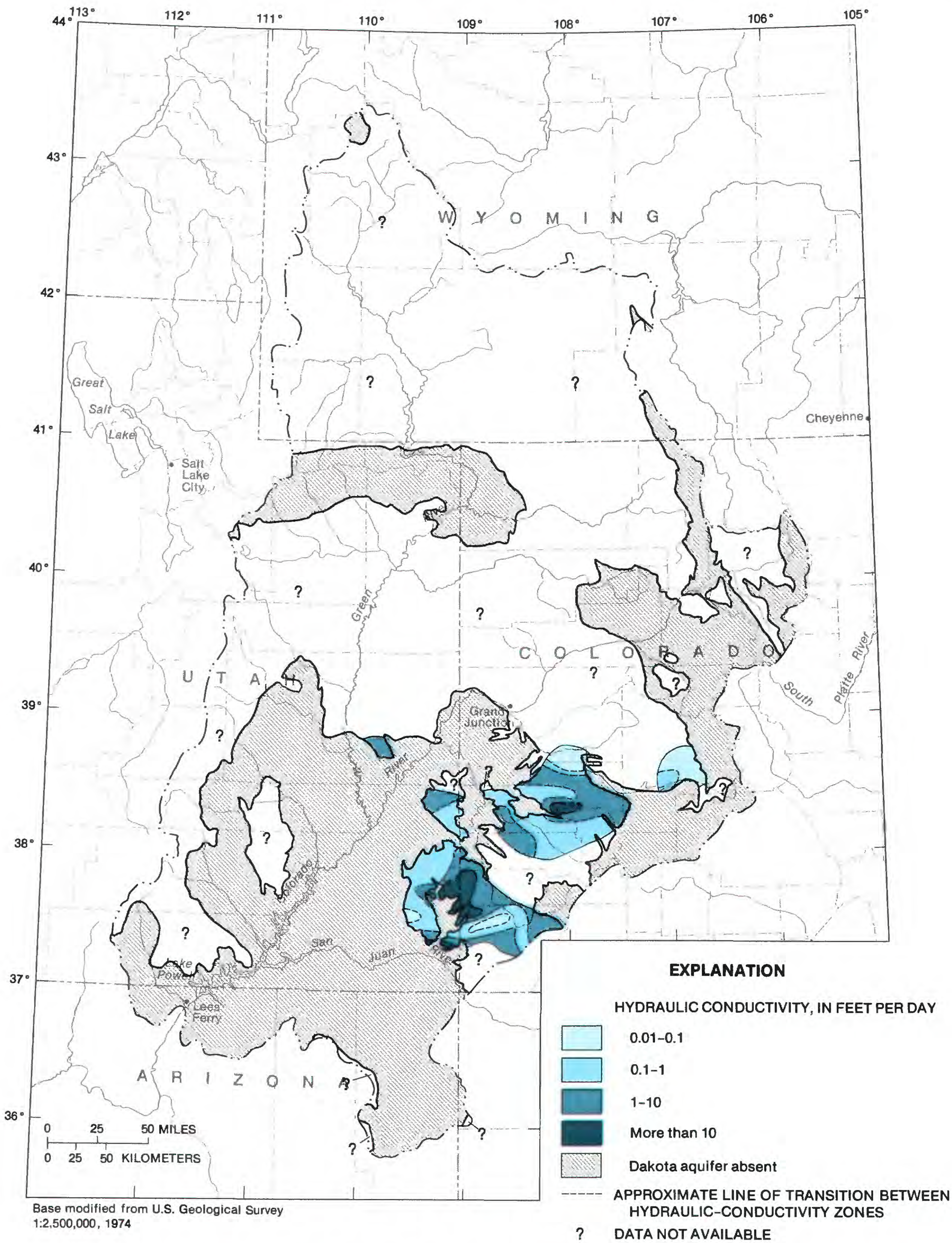


FIGURE 47.—Distribution of hydraulic-conductivity values derived from the specific capacity of wells in the Dakota aquifer.

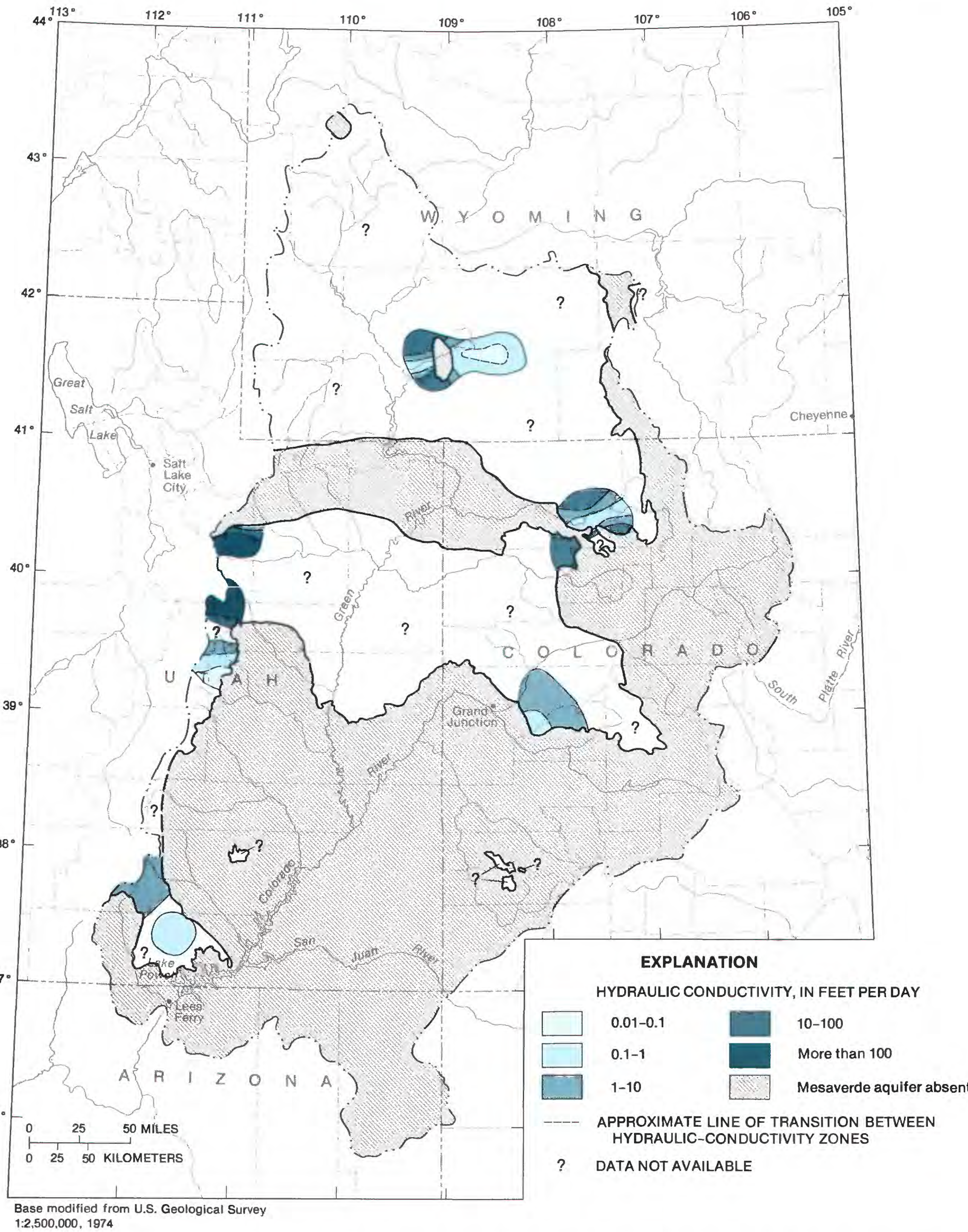


FIGURE 48.—Distribution of hydraulic-conductivity values derived from the specific capacity of wells in the Mesaverde aquifer.

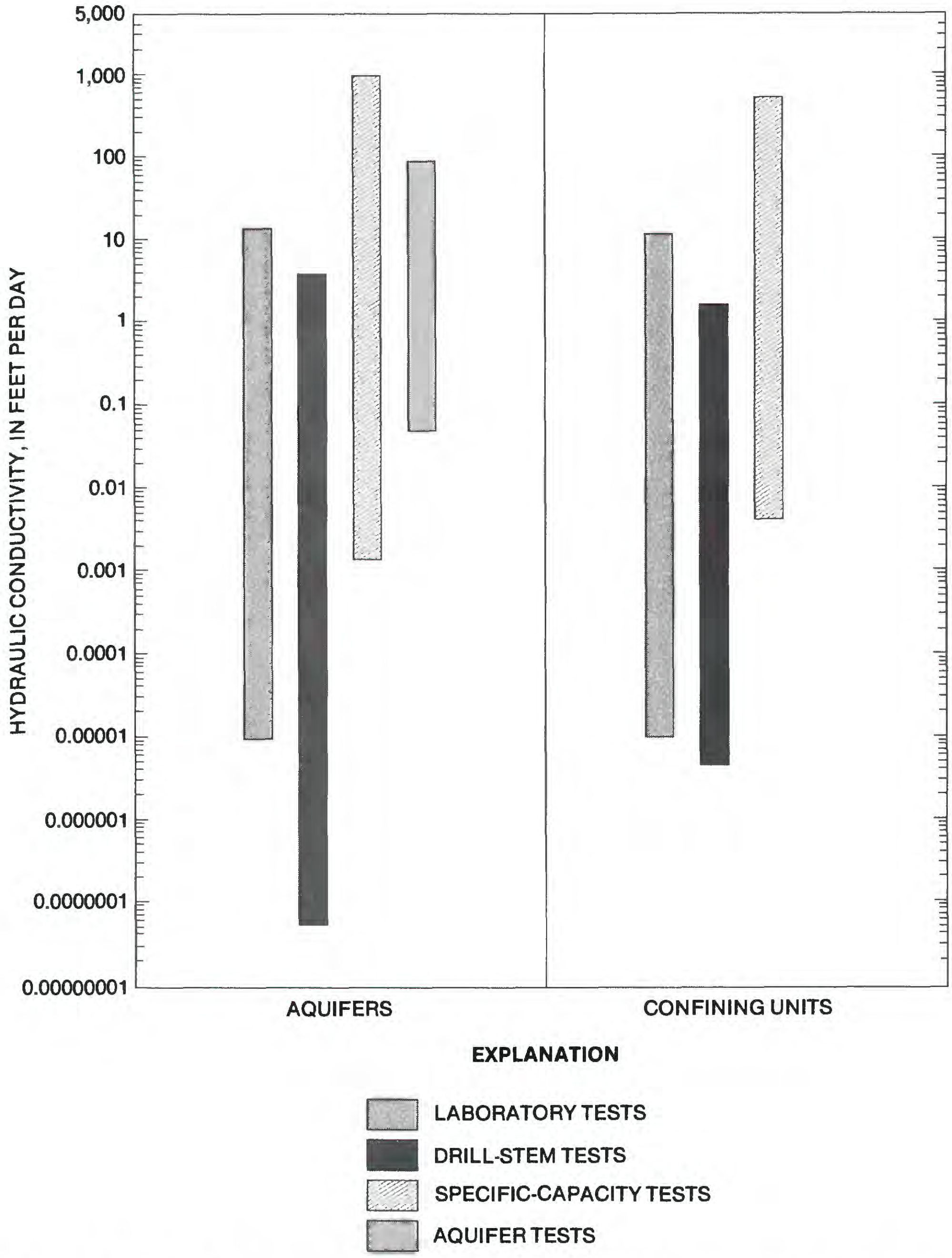


FIGURE 49.—Ranges in hydraulic-conductivity values obtained from laboratory, drill-stem, specific-capacity, and aquifer tests.

(3) horizontal or vertical interformational movement of ground water (fig. 55*C*), and (4) lateral movement of ground water into the study area from the defined boundary, in most cases the surface-water divide (fig. 55*D*). The first two means of recharge are most common.

LOCATION

Most recharge to aquifers in the Mesozoic rocks takes place in or near the areas of largest precipitation. This precipitation reaches the aquifers by infiltrating the unsaturated zone either at the place the precipitation falls or by infiltration where surface runoff is intercepted along ephemeral stream channels. Because flow in ephemeral channels is sporadic in time and quantity, identifying the exact location of recharge to underlying aquifers is impractical, if not impossible.

Recharge to bedrock aquifers occurs during prolonged wet surface conditions; thus, winter precipitation is

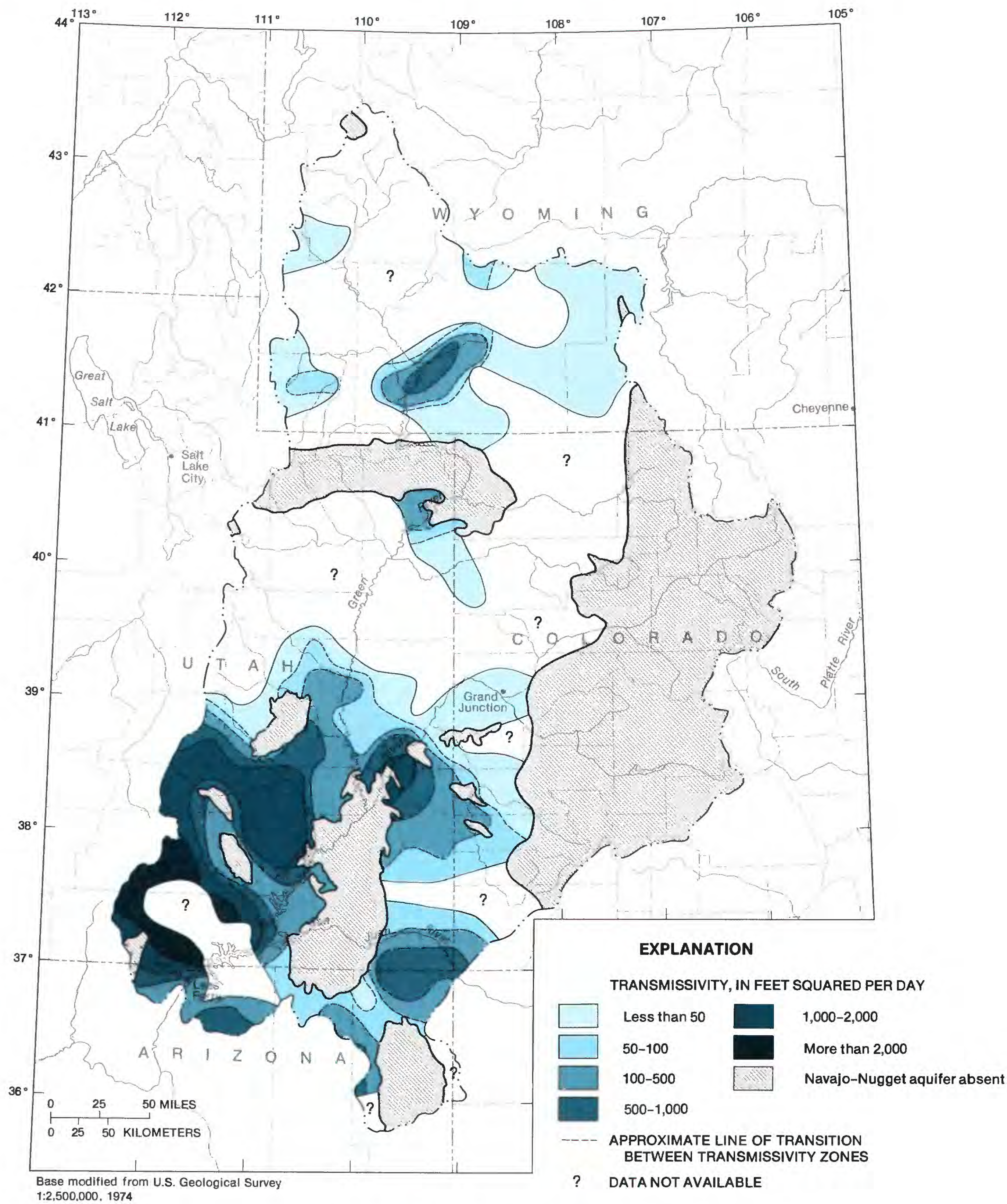


FIGURE 50.—Generalized distribution of transmissivity for the Navajo-Nugget aquifer.

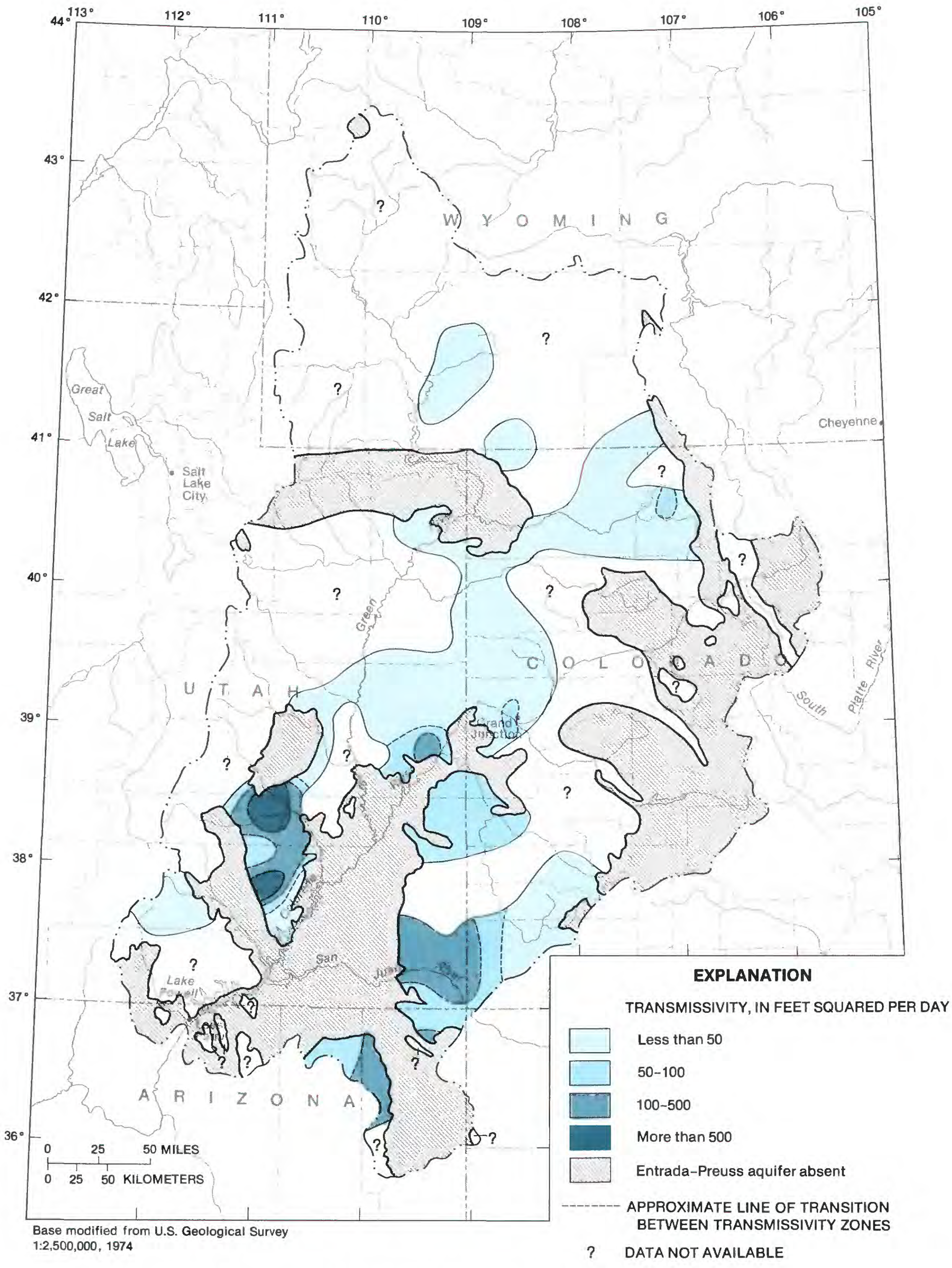


FIGURE 51.—Generalized distribution of transmissivity for the Entrada-Preuss aquifer.

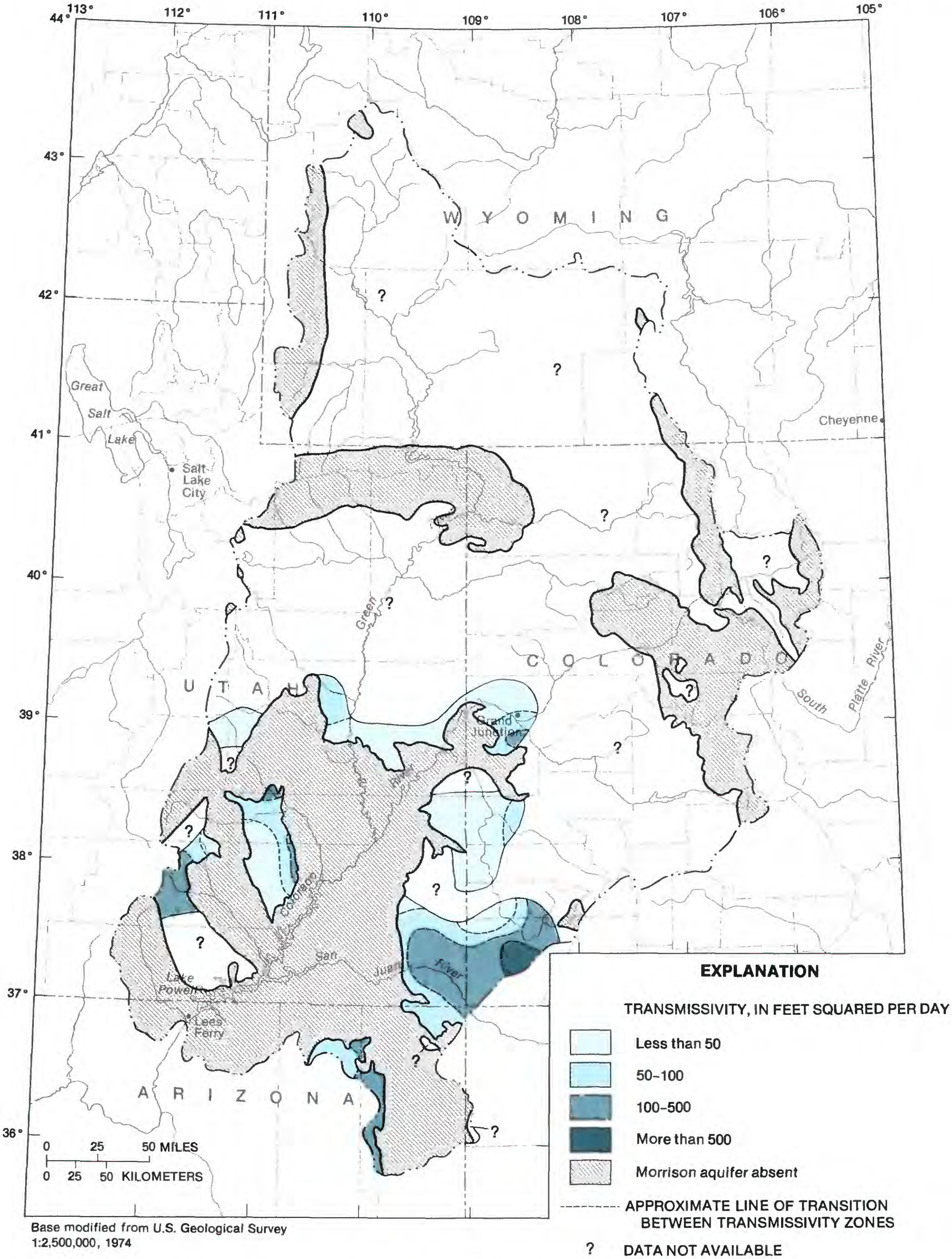


FIGURE 52.—Generalized distribution of transmissivity for the Morrison aquifer.

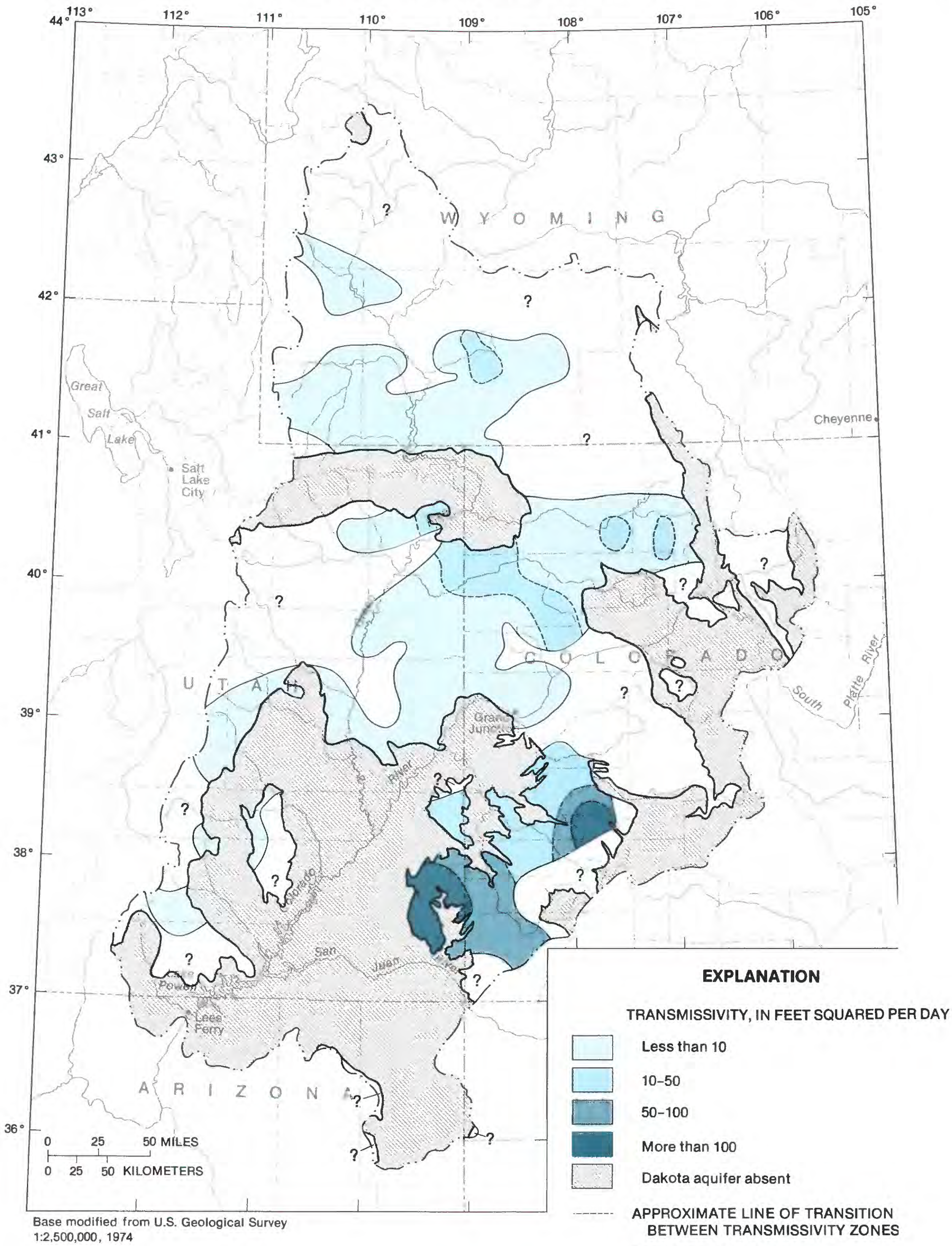


FIGURE 53.—Generalized distribution of transmissivity for the Dakota aquifer.

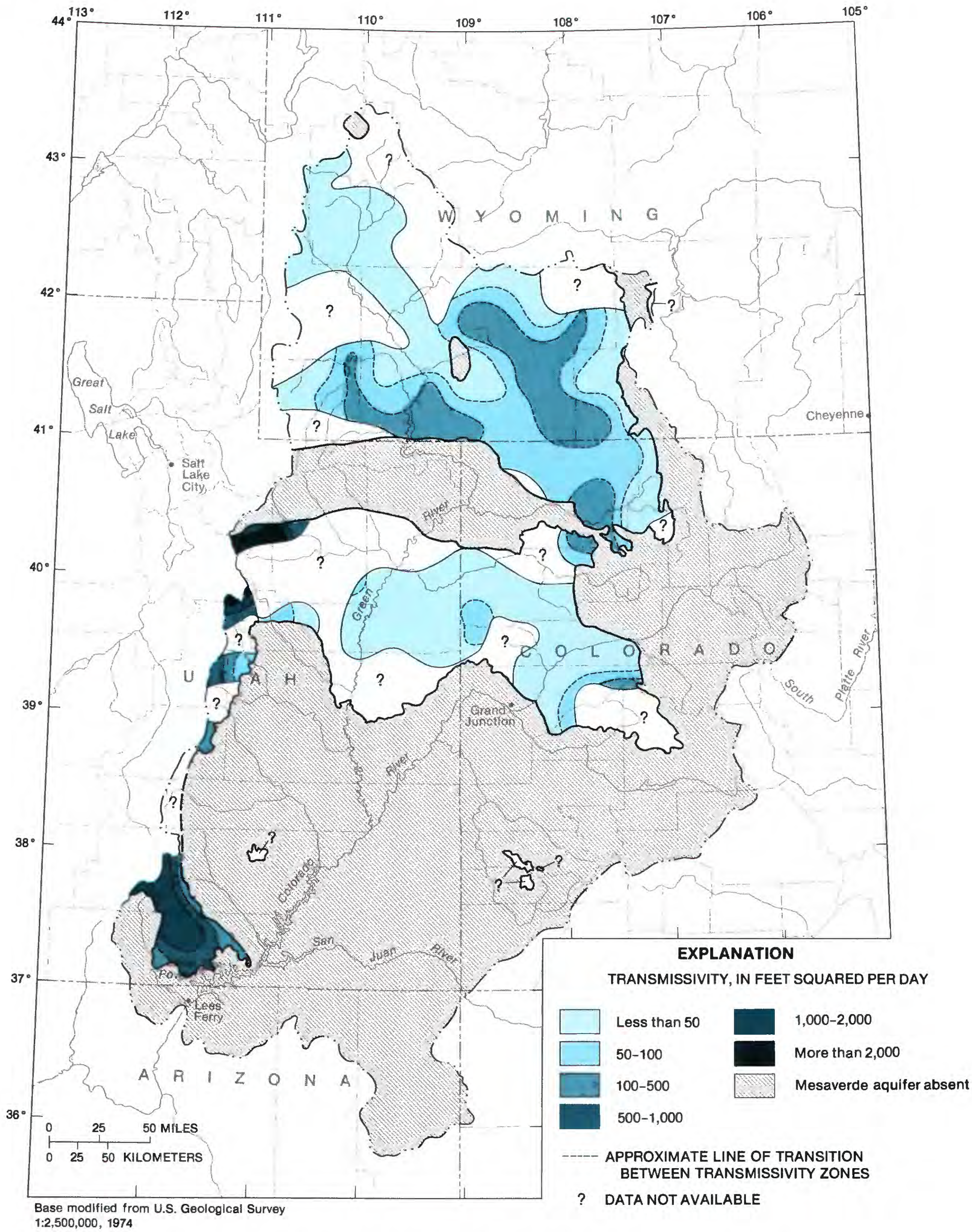


FIGURE 54.—Generalized distribution of transmissivity for the Mesaverde aquifer.

TABLE 2.—*Estimated ranges for the values of the storage coefficient, specific yield, and effective porosity in aquifers in Mesozoic rocks*

[Dashes indicate no data available]

Aquifer	Storage coefficient (confined aquifer)			Specific yield (unconfined aquifer) (percent)	Effective porosity (percent)			Source of estimate
	Range	Geometric mean	Number of available values		Range	Mean	Number of available values	
Mesaverde	0.000002–0.007	0.00004	5	----				Aquifer tests
					2.0–25.0	11	41	Laboratory analyses
Dakota	0.001	.001	2	----				Aquifer tests
					2.0–22.0	10	39	Laboratory analyses
Morrison	0.00003–0.0004	.0001	2	----				Aquifer tests
					4.0–25.0	13	22	Laboratory analyses
Entrada-Preuss	0.000005–0.00008	.00004	5	----				Aquifer tests
					10.0–26.0	20	13	Laboratory analyses
Navajo-Nugget	0.0003–0.008	.0008	21	0.05–0.10				Aquifer tests (Hood and Patterson, 1984)
					2.0–35.0	19	161	Laboratory analyses

responsible for most recharge (Hood and Patterson, 1984, p. 30). Areas of thick snowpack receive the greatest quantity of recharge because late winter and spring snowmelt increase the saturation of the rocks. Increased saturation facilitates recharge more than runoff. Recharge from infiltrating precipitation is most likely to occur where exposed aquifers in Mesozoic rocks receive more than 8 in of normal winter precipitation (fig. 56). (Normal winter precipitation is defined as that falling from October 1 through April 30.) Recharge potential is less where the units are buried and where winter precipitation is small.

According to a map of average annual snowfall (Baldwin and others, 1968, p. 53), the area where average annual snowfall exceeds 100 in corresponds closely to the area where normal winter precipitation is more than 8 in. The use of 8 in of winter precipitation as a smaller threshold for the initiation of recharge is based on results of infiltration and recharge experiments in the Dirty Devil River basin in Utah (Danielson and Hood, 1984). Local conditions elsewhere in the study area could affect the reliability of this threshold value.

As examples, Maxey and Eakin (1951) determined that recharge to Nevada basins originates where average annual precipitation exceeds 8 in, and Price and Arnow (1974) used 12 in of annual precipitation for a smaller threshold value when estimating total annual recharge to the Upper Colorado River Basin.

Excess applied irrigation water is thought to recharge the alluvial deposits in river valleys. Water from the alluvium may recharge the aquifers in Mesozoic rocks in a few locations within the study area where the hydraulic gradient is downward. However, because most irrigation is in the large river valleys of the study area, and because these rivers are mainly discharge areas for the aquifers, recharge to these aquifers by excess applied irrigation water is considered negligible.

Areas of perennial streamflow where exposed aquifers in Mesozoic rocks are being recharged are few. This type of recharge is possible at aquifer outcrops along the flanks of topographically high areas such as the Uinta Uplift, the Uncompahgre Uplift, the High Plateaus of Utah at the western border of the study area, and the Rocky Mountains in Colorado (fig. 57). Many aquifer exposures in the valleys and canyons of the major drainages are sites of ground-water discharge.

Aquifers in Mesozoic rocks may receive some recharge by vertical movement of water from the underlying and overlying rocks. Ground water moves vertically where there is a vertical hydraulic gradient between aquifers. The significance of this exchange is determined by the vertical hydraulic properties of the rocks through which the water moves and the size of the area affected. A small rate of interformational ground-water movement may account for a significant volume if the area affected is large. To reach the Navajo-Nugget aquifer, water in the upper Paleozoic rocks must move through as much as 2,000 ft of siltstone and claystone of the Chinle-Moenkopi confining unit. Because of the regional integrity of this confining unit, vertical ground-water movement from underlying rocks may be limited to areas of faulting, fracturing, and unsealed drill holes.

The degree of hydraulic connection between the Mesaverde aquifer and overlying aquifers in Tertiary rocks varies. In the Uinta Basin, the connection is poor because of the fine-grained nature and large thickness of the Wasatch and Green River Formations at the base of the Tertiary System (Hood, 1976, p. 9). In the Wyoming basins, the Fort Union Formation is at the base of the Tertiary System and is itself an aquifer (Welder and McGreevy, 1966, sheet 3). There, the vertical exchange of water between the Fort Union Formation and the Mesaverde aquifer is more probable, but interbedded shale, siltstone, and claystone within these two aquifers

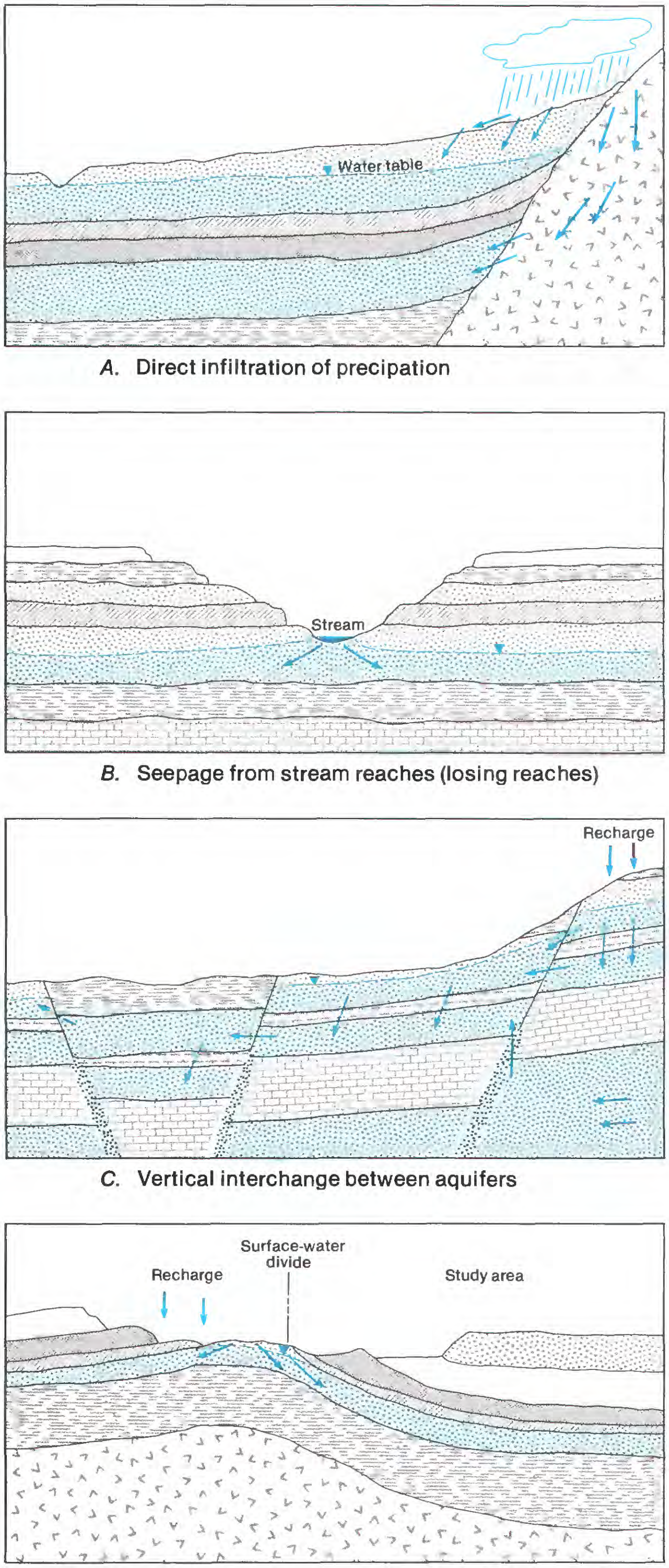


FIGURE 55.—Diagrammatic sections showing means of recharge to aquifers (arrows show direction of water flow).

likely impede exchange of water unless hydraulic gradients are increased substantially by pumping.

Comparing the potentiometric-surface maps of the upper and lower aquifers in Mesozoic rocks with those of adjacent aquifers indicates possible areas of exchange of water (fig. 58). Upward movement of ground water from aquifers in the upper Paleozoic rocks could occur in the Escalante River basin, in the San Juan River basin near Montezuma Creek, Utah, east of the Green River near Green River, Utah, in the southern part of the Paradox Basin in Colorado, at the southern edge of the Uinta Uplift, and at the western side of the Rock Springs Uplift. Downward movement of ground water from aquifers in lower Tertiary rocks is possible near the southern end of the Rock Springs Uplift and the eastern side of the Washakie Basin. Other similar areas may exist but cannot be identified because of the lack of water-level data.

Recharge originating outside the study area takes place where ground-water divides are outside surface-water divides of the Colorado River watershed, as illustrated in figure 55*D*. The locations of these recharge areas are defined by the potentiometric surfaces of the principal aquifers. The potentiometric surface of the Navajo-Nugget aquifer (pl. 5*A*) indicates probable ground-water movement into the study area at the southwestern side of the Paria Plateau and at the northeastern corner of the Black Mesa Basin. The potentiometric surfaces of the Entrada-Preuss aquifer (pl. 5*B*) and the Morrison aquifer (pl. 5*C*) are not as well defined as that of the Navajo-Nugget aquifer but nevertheless show that a small quantity of ground water may move into the study area from the San Juan Basin near the Four Corners Platform area. Water-level data for the Dakota aquifer (pl. 5*D*) are not sufficient to show any interbasin ground-water movement, and water-level contours for the Mesaverde aquifer (pl. 5*E*) indicate possible inflow at the High Plateaus of Utah along the southwestern boundary.

Artificial recharge to aquifers in Mesozoic rocks also occurs locally in the study area. Lake Powell, formed by the Glen Canyon Dam, is hundreds of feet deep, thus creating a local artificial recharge system for the Navajo-Nugget and Entrada-Preuss aquifers (see Thomas, 1985). Because regional movement of ground water indicates that the gorge is a discharge area, the effect of the lake has been to raise the discharge base level and increase the volume of water in storage in the formations surrounding the lake. This increase in base level extends about 186 mi upstream from the dam along the Colorado River and for many miles along tributary canyons. Smaller reservoirs in Utah and Colorado cause similar localized increases in ground-water storage in other aquifers in Mesozoic rocks, but on a much smaller scale.

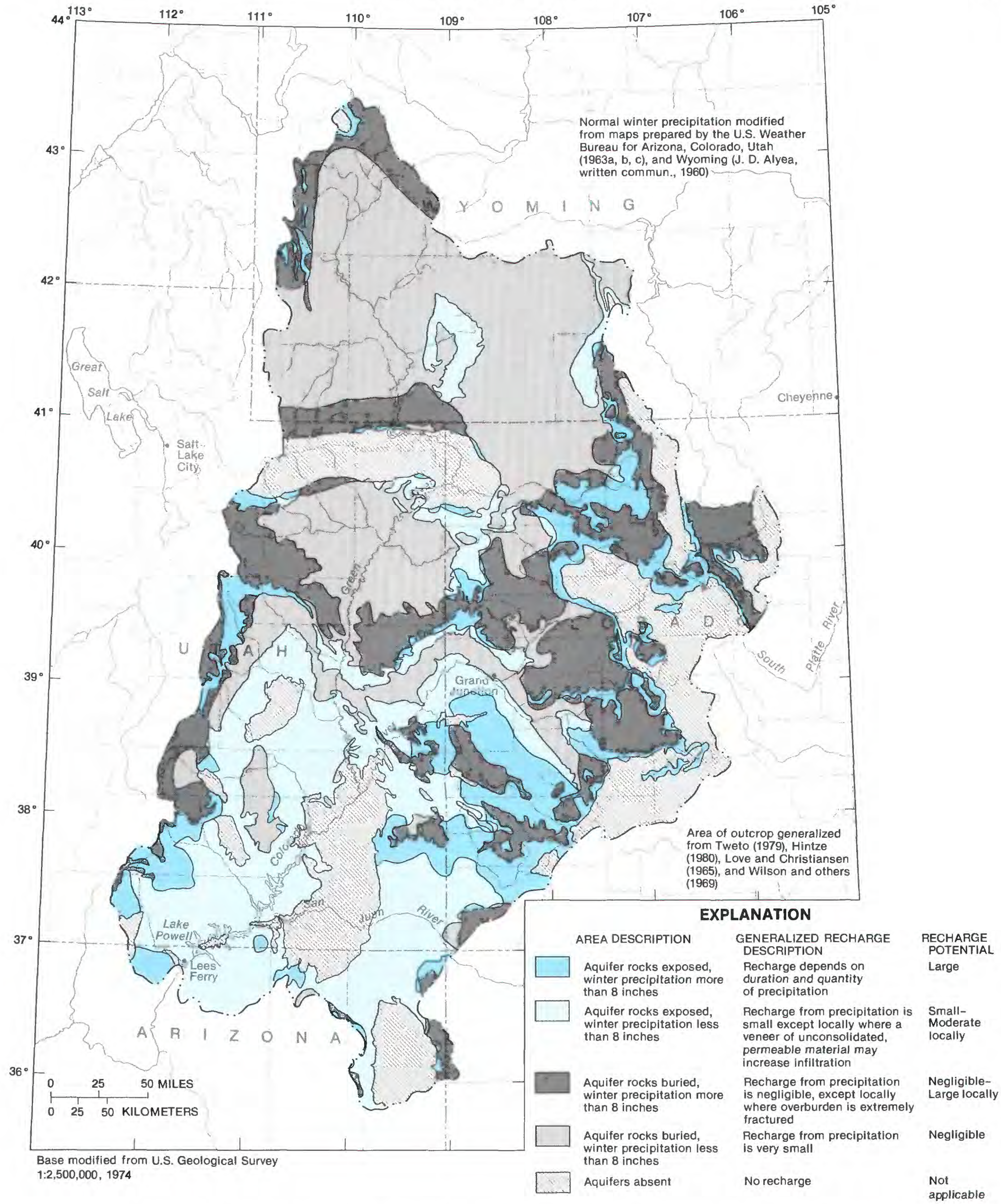


FIGURE 56.—Potential for recharge by direct infiltration of precipitation to the aquifers in Mesozoic rocks.

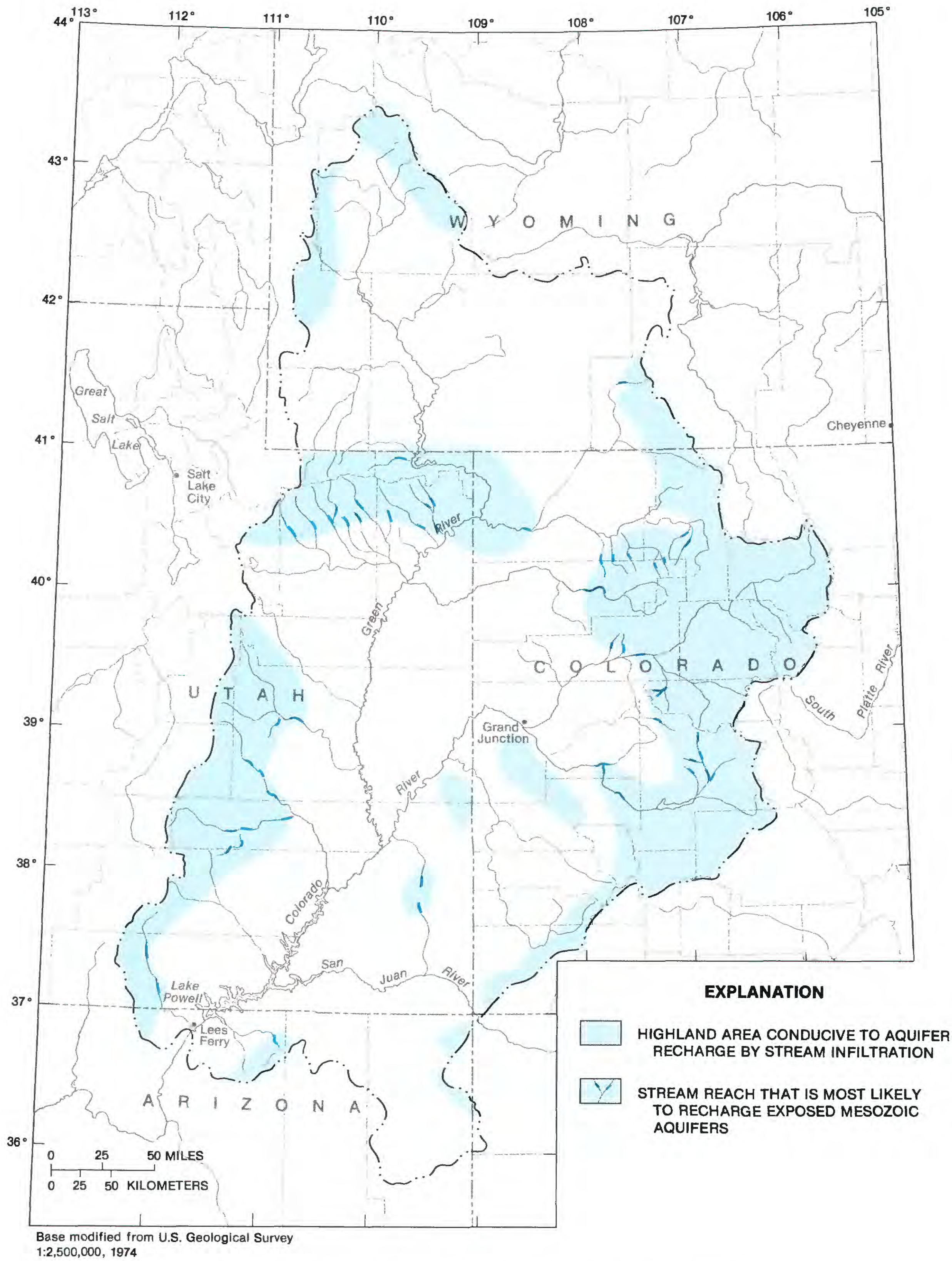


FIGURE 57.—Areas of streamflow recharge to aquifers in Mesozoic rocks.

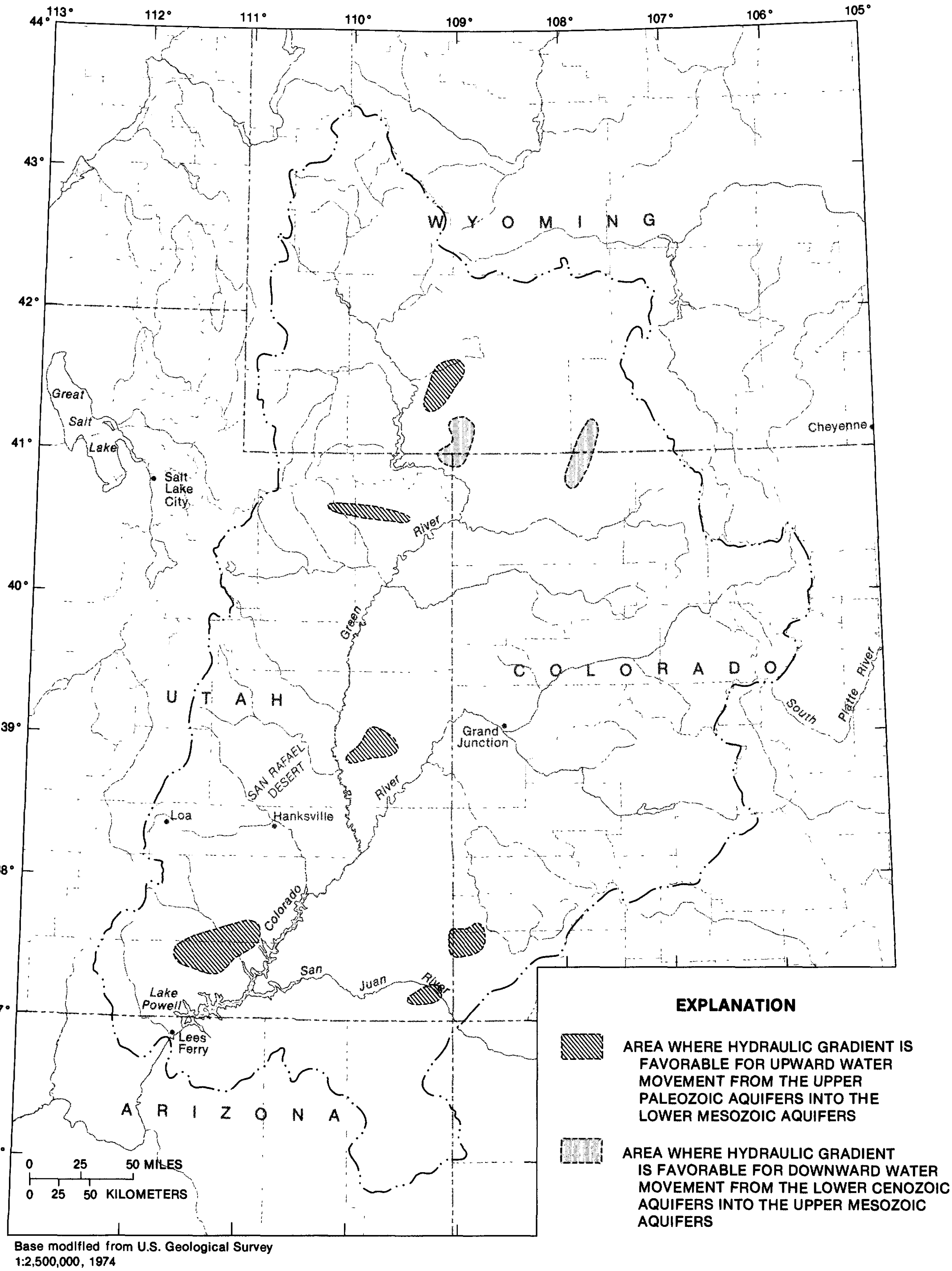


FIGURE 58.—Areas of possible vertical water flow into aquifers in Mesozoic rocks from underlying and overlying rocks.

QUANTITY

Over a long period and under natural conditions, inflow of water to aquifers in the Mesozoic rocks in the study area should equal outflow from the aquifers. The rate cannot be measured on a regional scale, but magnitude can be stated in terms of likely minimum and maximum values using empirical methods developed from the results of previous investigations.

Of the four previously mentioned methods by which aquifers in Mesozoic rocks are recharged in the area, direct infiltration of precipitation into outcrops and infiltration of runoff are the most common. Recharge resulting from infiltration of precipitation and runoff has been estimated by previous investigators for selected subareas and for the entire study area. Almost all estimates of recharge by infiltration of precipitation are based on the Maxey-Eakin method (Maxey and Eakin, 1951) or a modification of that method. The Maxey-Eakin method was derived for alluvial basins of the Basin and Range province in Nevada and incorporates both areal recharge and ephemeral stream-channel recharge. The use of this method for any area other than that of the Basin and Range environment involves many adjustments because of differences in lithology, topography, and potential evaporation.

The most striking difference between the Basin and Range province and the study area is the lithologic character of the aquifers. The sandstone and conglomerate that are aquifers in the study area have hydraulic-conductivity values as much as one to two orders of magnitude smaller than corresponding values for alluvial basin fill in the Basin and Range province. For this reason, the rate of infiltration and recharge to the Upper Colorado River Basin aquifers is probably smaller than that to the basin fill for areas where average annual precipitation, duration, and runoff are similar. Recharge values estimated using the Maxey-Eakin method can be considered the largest possible values.

Using a modified version of the Maxey-Eakin method for areas where average annual precipitation is less than 12 in and all exposed Mesozoic formations are considered rechargeable, the estimated annual recharge to the aquifers in Mesozoic rocks is about 3.3 million acre-ft from precipitation and stream infiltration. Using the same assumptions, a statistical modification of the Maxey-Eakin method based on multiple-linear regression (Watson and others, 1976, p. 346) indicates an annual recharge of 3.1 million acre-ft from precipitation and stream infiltration.

Certain physical conditions increase the potential for recharge. Fracturing of the rock in a recharge area allows more precipitation or runoff to percolate downward and eventually enter the saturated zone. However, the spacing of the joints and fractures and the degree to which the fractures are plugged by chemical precipitates or fine-grained sediments are extremely variable and difficult to quantify. A surface covering of material that is easily infiltrated and prevents rapid evaporation of runoff also increases recharge potential. The covering may be a soil zone, dune sand, or another permeable or poorly consolidated deposit overlying the aquifer. Dune sand covering the San Rafael Desert (fig. 58) north and east of Hanksville, Utah, and Tertiary lava and ash flows on the plateaus north and south of Loa, Utah, are examples of areas that are favorable for increased recharge potential.

Study of infiltration from precipitation and streamflow, and of the recharge resulting from this infiltration (Danielson and Hood, 1984), conducted at Navajo Sandstone outcrops in the southwestern part of the study area provides some insights into ground-water recharge in the Upper Colorado River Basin. In the study by Danielson and Hood, infiltration and recharge at various altitudes and topographic settings were measured during various periods of precipitation using neutron-moisture probes and tensiometers. It was concluded that about 14 percent of the precipitation recharged the aquifer at a site where winter precipitation was about 20 in, but that virtually no recharge occurred at a second site where winter precipitation was slightly less than 8 in, even though infiltration was occurring. Recharge at these sites is considered to be minimal because the sites were established mainly on barren, unfractured, well-cemented sandstone.

The two measurements by Danielson and Hood were assumed to represent the relation between percentage of precipitation recharging the aquifer and altitude in the study area. Because no other data points are available, the simplest relation, a linear one, was assumed (fig. 59). These two measurement points probably result in recharge values smaller than values obtained by the Maxey-Eakin method, because the measurement areas are small and lack fractures and joints. Therefore, recharge values determined using this relationship probably are near the lower limit of recharge estimates. This method probably yields a realistic estimate of average areal recharge in sandstone areas where fractures are widely spaced or where fractures are ineffective because of secondary filling. If the relationship shown in figure 59 is used, the annual quantity of recharge to outcrop areas for aquifers in Mesozoic rocks, and to areas where the aquifers are covered by an extremely permeable veneer of unconsolidated sediments, is about 300,000 acre-ft—25,000 acre-ft to the Navajo-Nugget aquifer, 40,000 acre-ft to the Entrada-Preuss aquifer, 40,000 acre-ft to the Morrison aquifer, 65,000 acre-ft to the Dakota aquifer, and 130,000 acre-ft to the Mesaverde aquifer. Additional infiltration and recharge measurements are needed to determine the validity of this relationship, and

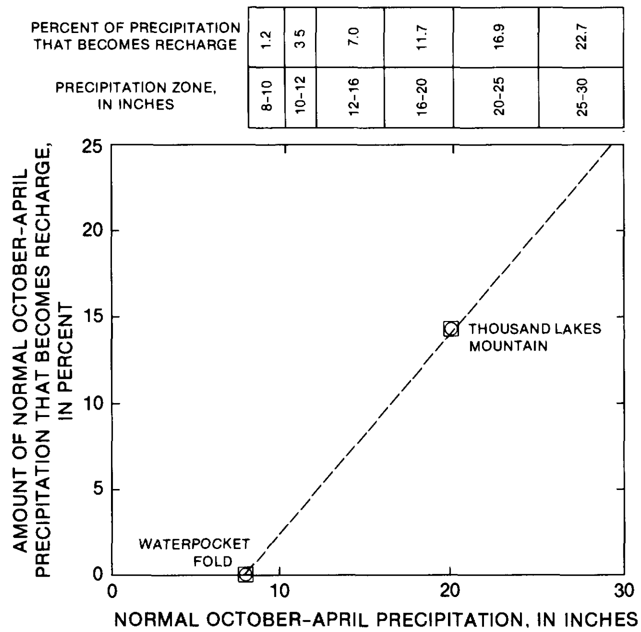


FIGURE 59.—Possible recharge-precipitation relation (from results of infiltration study by Danielson and Hood, 1984).

should incorporate the effect of fractures and soil coverage on the estimates of recharge rates.

Many areas in the Upper Colorado River Basin where permeable plateau-capping igneous rocks occur have the potential to transmit water to the underlying aquifers in Mesozoic rock. By applying recharge rates estimated from the study by Danielson and Hood, an additional annual recharge rate of 280,000 acre-ft was calculated—about 160,000 acre-ft to the Morrison aquifer, 80,000 acre-ft to the Dakota aquifer, and 40,000 acre-ft to the Mesaverde aquifer.

Each of the five aquifers is crossed by perennial streams for short distances where structural warping and erosion have removed overlying layers. If a measurable decrease in streamflow occurs through such a reach, and if other stream losses such as evapotranspiration are accounted for, the remaining loss can be attributed to infiltration into the underlying aquifer.

Danielson and Hood conducted seepage investigations on three streams in the Dirty Devil River basin. Measured losses indicate that recharge from these perennial streams crossing sandstone outcrops ranges from negligible to about 2 cubic feet per second (ft^3/s) per mile of stream channel. Average streamflow losses range from 0.5 to 0.7 ft^3/s per mile.

Avery (1986, table 3) conducted seepage investigations on small streams originating in the Abajo Mountains near Monticello, Utah. Streamflow losses to the Navajo-Nugget and Morrison aquifers range from 0.01 to 0.7 ft^3/s per mile and average about 0.2 ft^3/s per mile.

Based on the cited studies and the length of stream reaches crossing sandstone outcrops shown in figure 57, it is possible to estimate recharge from streamflow. Using a minimum streamflow loss of 0.01 ft^3/s per mile, the minimum recharge where the potentiometric surface of the traversed aquifer is below the streambed would be about 1,500 acre-ft. Using a maximum streamflow loss of 2 ft^3/s per mile, the maximum recharge by streams would be about 290,000 acre-ft. Using a value of 0.4 ft^3/s per mile, a probable value of average stream losses determined from these studies, annual recharge to all aquifers in Mesozoic rocks would be about 58,000 acre-ft—about 20,000 acre-ft to the Navajo-Nugget aquifer, about 10,000 acre-ft to the Entrada-Preuss aquifer, about 3,000 acre-ft to the Morrison aquifer, about 5,000 acre-ft to the Dakota aquifer, and about 20,000 acre-ft to the Mesaverde aquifer.

The quantity of water moving into the aquifers in Mesozoic rocks from overlying Tertiary sedimentary formations and from underlying Paleozoic sedimentary formations is relatively small. Values for vertical flow were estimated from geologic logs, water-level measurements, and hydraulic-conductivity measurements. Representative values for each area where vertical flow into the aquifers in Mesozoic rocks has the greatest potential (table 3) result in a total flow of about 1,100 acre-ft. Because of the uncertainty in estimated values of vertical hydraulic conductivity and of differences in hydraulic head between aquifers, the calculated flow between aquifer systems could be in error by several orders of magnitude.

Lateral flow of ground water across study-area boundaries is also small in terms of a regional ground-water budget. Much of this flow occurs at the southern and southeastern boundaries of the study area in Arizona and southern Utah. Based on simulations of the ground-water flow in the Navajo Sandstone, flow into the study area from the Paria Plateau is about 3,000 acre-ft (Thomas, 1985), and flow from Black Mesa Basin is about 500 acre-ft (Eychaner, 1983, p. 11).

MOVEMENT

Water in aquifers in Mesozoic rocks of the Upper Colorado River Basin moves from areas of high fluid potential to areas of low fluid potential. Differences in fluid potential within the aquifers are most commonly caused by differences in elevation head and pressure head, but they can also result from variations in fluid density, chemistry, and temperature. These variations may be natural occurrences during steady-state conditions or may be induced by natural or manmade changes

TABLE 3.—*Estimated vertical flow into aquifers in Mesozoic rocks based on the ground-water flow (Darcy) equation:* $Q=K_v\ A\ dh/dl$

Area affected: Approximate geographic area shown in figure 58.
Vertical hydraulic conductivity (K_v): Typical vertical hydraulic conductivity for the rock through which ground water moves.
Size of area (A): Estimated from figure 58.
Difference in hydraulic head (dh): Representative head difference between the lowermost aquifer in Tertiary rocks and the uppermost aquifer in Mesozoic rocks, or between the uppermost aquifer in Paleozoic rocks and the lowermost aquifer in Mesozoic rocks.
Distance of vertical flow (dl): Representative vertical distance water must move from one aquifer system to another.
Quantity of flow between aquifer systems (Q): Flow calculated using the flow equation.

Area affected	Vertical hydraulic conductivity (feet per year)[1] K_v	Size of area (acres) A	Difference in hydraulic head (feet) dh	Distance of vertical flow (feet) dl	Quantity of flow between aquifer systems (acre-feet per year) Q	Direction of flow between aquifer systems
Escalante River basin	0.001	750,000	300	1,300	170	Paleozoic to Mesozoic
San Juan River basin	.001	150,000	300	1,100	40	Do.
East of Green River, Utah	.001	300,000	500	1,100	140	Do.
Southern Paradox Basin	.001	300,000	700	700	300	Do.
Southern Uinta Mountain Front	.001	200,000	1,000	900	220	Do.
West of Rock Springs Uplift	.001	300,000	500	1,000	150	Do.
South of Rock Springs Uplift	.001	150,000	300	1,000	50	Cenozoic to Mesozoic
Eastern Washakie Basin	.001	100,000	200	1,000	20	Do.
Total (rounded to nearest 100 acre-feet per year)					1,100	

[1]Values for claystone, siltstone, and shale range from about 0.000001 to 0.01 foot per year (Davis and DeWiest, 1966, p. 349; Morris and Johnson, 1967, p. D36; Freeze and Cherry, 1979, p. 29). The value of 0.001 foot per year represents the most common value for all three types of confining layers.

during transient conditions. The flow of ground water can be interrupted, redirected, depleted, or enhanced by subsurface geologic structure and stratigraphy, land surface features, vegetative cover, and spatial and temporal changes in climatic conditions.

The rate and direction of horizontal ground-water movement are generally defined by gradients of the potentiometric surfaces (pl. 5). Hydraulic gradients in the aquifers vary from about 2 to 100 feet per mile (ft/mi). The direction of lateral ground-water movement typically is from mountains at the borders of the study area toward main rivers that drain the basins, but recharge areas other than those at the borders, and discharge areas other than the rivers, do exist.

The rate and direction of vertical water movement usually depend on vertical differences in hydraulic head between aquifers, and on the thickness and vertical hydraulic conductivity of the confining units. By comparing the potentiometric-surface maps on plate 5 to determine head differences, and by examining thickness of the confining units on plate 2, the relative degree of potential vertical movement within the aquifers and confining units in Mesozoic rocks can be determined.

STRUCTURAL CONTROLS

The large basins and uplifts tend to impede regional movement of water through aquifers in Mesozoic rocks. At uplifted areas where these aquifers have been removed by erosion, regional flow is prevented because these are areas of recharge and discharge. Because recharge and discharge occur locally, water more likely flows through local and intermediate flow systems instead of a regional system. The San Rafael Swell, the Rock Springs Uplift, and the Uncompahgre Uplift are examples.

Large structural basins impede water movement for different reasons than do uplifted areas. Aquifers that are deeply buried in these basins are affected by changes in their hydrologic properties owing to the weight of overlying sediments. Hydraulic-conductivity values determined in a laboratory for four rock samples subjected to five different simulated overburden thicknesses decreased by 7 to 26 percent when simulated overburden thickness increased from 400 to 6,000 ft. Samples containing the largest quantities of fine-grained material (silt size or smaller) had the smallest decreases in hydraulic conductivity, suggesting that the presence of some fine-grained material between the sand grains may increase the skeletal strength of a sandstone.

Porosity also tends to decrease with depth of burial. A plot of porosity values determined from laboratory core analyses versus depth of core source (fig. 60) shows a general decreasing trend in porosity values for sample depths ranging from less than 500 to more than 15,000 ft. The average porosity of samples were

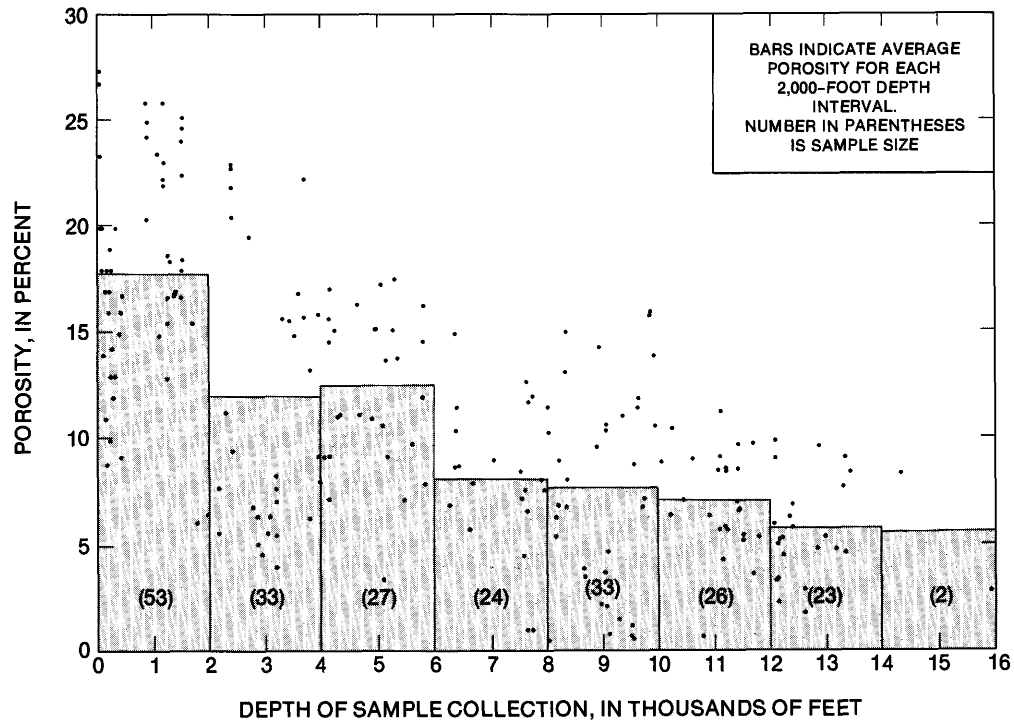


FIGURE 60.—General relation between porosity and depth of sample in Mesozoic rocks.

Depth of collection (feet)	*Arithmetic average of porosity (percent)*
10 to 2,000	17.8
2,000 to 4,000	12.0
4,000 to 6,000	12.5
6,000 to 8,000	8.1
8,000 to 10,000	7.7
10,000 to 12,000	7.0
12,000 to 14,000	5.7
More than 14,000	5.5

Values are widely scattered but generally indicate about a 0.9-percent decrease in porosity per 1,000 ft of depth in the first 6,000 ft, and about a 0.3-percent decrease per 1,000 ft where depths exceed 6,000 ft. Secondary porosity caused by dissolution of grains and cementing matrix may be one reason for this decrease with depth (Shanmugam, 1985). Other reported decreases in porosity with depth, summarized in Freeze and Cherry (1979, p. 153), generally are larger than 1 percent per 1,000 ft.

Faults and fault zones can impede or enhance ground-water movement. Huntoon (1985, p. 177) suggests that the large-displacement thrust faults that border the Green River and Great Divide Basins in Wyoming prevent recharge to deeply buried formations and effectively isolate many aquifers. Aquifers on the southern flank of the Uinta Uplift may be similarly isolated from the systems deep in the Uinta Basin. This condition could be one reason for the apparent movement of ground water from west to east along the southern flank of the Uinta Mountains, indicated by the potentiometric contours for the Navajo-Nugget and Morrison aquifers (pls. 5*A*, 5*C*). However, because the water along these mountains is also moving toward a regional drain, the Green River, the opposite interpretation, that thrust faults have little effect on water movement, could be argued. Normal faults, in general, do not appear to cause separate circulations, either because fault displacement is too small to cause consequential offsetting of formations, or because faults created by tension produce less fault gouge and more fracturing, thus forming permeable conduits that allow water to move horizontally and vertically from one formation to another. Normal faults that parallel the western border of the study area in Utah have no apparent effect on water movement in the Navajo-Nugget aquifer, based on the configuration of the potentiometric surface. The east-west-trending faults that form graben structures south of Monticello, Utah, cause displacement in the Navajo-Nugget, Entrada-Preuss, and Morrison aquifers but have negligible effect on the flow of water from north to south.

Fracturing caused by faulting, folding, and erosional unloading generally enhances the movement of ground water. However, the spacing and interconnection of fractures generally diminish with depth and with increasing distance from the fold or fault causing the fracture. Snow (1968, p. 87) notes that fracture openings in rocks beneath dam sites range from 75 to 400 microns in the upper 30 ft but decrease to 50 to 100 microns at depths of 30 to 400 ft. Nelson and Handin (1977, p. 234) report that fracture permeability in the Navajo Sandstone decreases to less than 2 percent of original when burial depth, simulated under laboratory conditions, increases from 0 to 10,000 ft. Tension fractures caused by anticlines, monoclines, and synclines generally are most pronounced at the convex part of the fold, where tension is largest. This flexure generally is less deeply buried for anticlines and monoclines than for synclines. Because the weight of overlying material tends to close fractures, enhancement of ground-water movement is probably more pronounced near anticlines and monoclines. Because the causes of fracturing are not regionally consistent, ground-water movement enhanced by fractures is local rather than regional.

LITHOLOGIC CONTROLS

The lithologic character and areal extent of the Mesozoic rocks of the Upper Colorado River Basin affect the direction and rate of water movement. Mesozoic rocks extend throughout about 85 percent of the study area (Freethey and others, 1988), but each individual rock unit may cover less area than the size of the study area because of different degrees of erosion or lack of deposition. Water movement between aquifers can be different owing to thin, absent, very coarse grained, or significantly fractured confining units.

In most of western Colorado and parts of eastern Utah, the Carmel-Twin Creek confining unit is absent. In these areas the Navajo-Nugget and the Entrada-Preuss aquifers are hydraulically connected and act as a single aquifer. To the west the Carmel-Twin Creek confining unit increases in thickness, and its low permeability is more restrictive to vertical water movement. Locally, in the drainage of Muddy Creek at the southern end of the San Rafael Swell, the Carmel-Twin Creek confining unit is fractured and contains saline water that may leak downward and mix with fresher water in the Navajo Sandstone (Hood and Danielson, 1981, p. 46).

The Carmel-Twin Creek confining unit in the southern Green River Basin and the Great Divide Basin is coarser grained and more permeable than its counterpart in Utah, but in these basins the overlying Entrada-Preuss aquifer grades to less permeable limestone and shale. The Entrada-Preuss aquifer extends farther east than the Navajo-Nugget aquifer, and thus receives more recharge from the eastern border of the study area, where precipitation is large. The movement of water through the Entrada-Preuss aquifer is well defined along the eastern side of the study area, where it is commonly sandstone, but becomes undefined to the west, where the lithologic character changes to shale and limestone.

The overlying Curtis-Stump confining unit is sandy at the southern and northern ends of the study area and shaly in the midlatitudes of the study area. Thin interbedded shale in the sandy facies retards vertical movement between the Entrada-Preuss and Morrison aquifers. The Morrison aquifer becomes progressively less sandy from southwest to northeast in Utah and Colorado. Although the Morrison aquifer is not defined in Wyoming and northern Colorado, the Morrison Formation becomes more sandy on the northwestern and northeastern sides of the study area in Wyoming and northern Colorado, and aquifers of relatively limited extent may be present.

Water movement in the Morrison Formation is poorly defined everywhere except in the Four Corners Platform area, the southern Paradox Basin, and the Yampa River drainage basin. The Morrison includes a shaly confining unit, the Brushy Basin Member, in Utah and Colorado. This shale is not identified in the northern and eastern parts of the area, but where the Brushy Basin exists, it effectively impedes water movement between the sandstone beds in the lower Morrison aquifer and the overlying Dakota aquifer.

Most of the Dakota aquifer is between 100 and 250 ft thick and includes shale and siltstone layers interbedded with permeable sandstones. Regional direction of water movement is from east to west, mainly in Colorado and Wyoming. The overlying Mancos confining unit includes about 1,000 to 14,000 ft of shale with some isolated sandstones interbedded along the western side of the study area. Because of its large thickness and the fine-grained nature of its shale, vertical movement of water through the Mancos confining unit is probably negligible.

Movement of water in the Mesaverde aquifer is complex because of the stratigraphy of the formations. Recharge and discharge areas are more localized at topographic high and low areas, creating several discontinuous flow systems rather than a single regional system.

FLUID-CHARACTER CONTROLS

Water in the aquifers in Mesozoic rocks varies in chemical character, density, and viscosity because of diverse environments. This variation affects the movement of water through the aquifers and confining units.

The concentration of dissolved solids in water flowing through the five aquifers ranges from 28 to 138,000 milligrams per liter (mg/L) (Freethey and others, 1988, table 2). The transition from fresh to briny water reflects

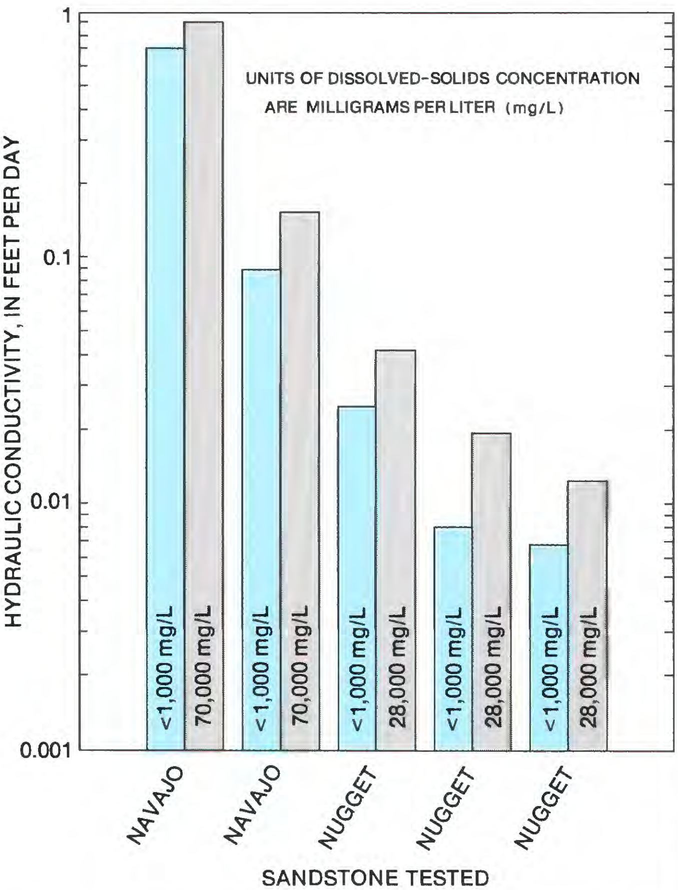


FIGURE 61.—Relation between dissolved-solids concentration and hydraulic conductivity in the Navajo-Nugget aquifer.

a change in water type from mostly calcium bicarbonate to mostly sodium chloride. The water that contains 1,000 to 35,000 mg/L dissolved solids is generally a mixed type containing calcium, magnesium, sodium, sulfate, and bicarbonate ions.

Concentration of dissolved solids in water affects the hydraulic conductivity of the aquifers. Hydraulic-conductivity values for two cores of Navajo Sandstone using water containing about 70,000 mg/L of dissolved solids were 24 and 45 percent larger than values for the same cores tested with freshwater (fig. 61). Hydraulic-conductivity values for three cores of the Nugget Sandstone from about 10,000 ft below land surface measured using water containing about 28,000 mg/L of dissolved solids were 66 to 142 percent larger than values for the same cores tested with freshwater. Although no mineralogic analyses are available to confirm the presence of clays in the tested samples, the observed decrease in hydraulic conductivity with decreasing dissolved solids is believed to be caused by peptization of clays in freshwater and subsequent deposition of microscopic filter cake across smaller pore openings (van Olphen, 1977, p. 127). In saline water, clay minerals flocculate, thus preventing the formation of the filter cake.

Fluid movement in aquifers is dependent on properties of the porous medium and of the fluid in the pores. Fluids contained in aquifers in Mesozoic rocks in the study area are subjected to large chemical, pressure, and temperature differences between points deep in structural basins and outcrop areas of the aquifers. Lateral variations in the dissolved-solids concentration results in lateral variations in fluid density. Vertical changes in temperature result in vertical changes in fluid viscosity. Both variations affect movement of fluids within the aquifer. The effects of fluid viscosity and density variations in the aquifers may cause ground-water flow directions to not be perpendicular to lines of equal potentiometric head in the deep structural basins. The potentiometric contours shown on plate 5 have not been adjusted to compensate for these variations. Thus, actual directions of ground-water flow may be different from those shown on the maps.

DISCHARGE

Discharge from the aquifers in Mesozoic rocks comes from (1) springs and seeps, (2) evapotranspiration, (3) subsurface lateral flow across study area boundaries, and (4) vertical flow to overlying Cenozoic and underlying Paleozoic rocks.

Springs and seeps not discharging to streams are found in canyon bottoms and on canyon walls. Springs that have been located and recorded are major discharge sites, but numerous smaller unrecorded springs and seeps exist throughout the study area. Areas where spring discharge from each aquifer is largest are identified in figures 62 and 63. The largest areal density of spring sites is where the aquifers are incised by deep canyons. The largest density of springs in the lowermost aquifers, the Navajo-Nugget and the Entrada-Preuss, is along the Colorado River canyon at the southern end of the study area. Spring discharge from the Morrison aquifer occurs at the San Juan River, along the flanks of the Henry Mountains, and in canyons cut into the Uncompahgre Uplift. Spring discharge from the Dakota aquifer occurs in canyons of southwestern Colorado, relatively close to its recharge area. Spring discharge from the uppermost aquifer, the Mesaverde, occurs in the High Plateaus of Utah near the western border of the study area, in the Rock Springs Uplift, the Wyoming thrust belt, and along the eastern side of the Washakie Basin.

Ground-water discharge to streams is the base flow of those streams, and such discharge sites can generally be located by comparing the relative position of perennial stream courses with the shape of the potentiometric

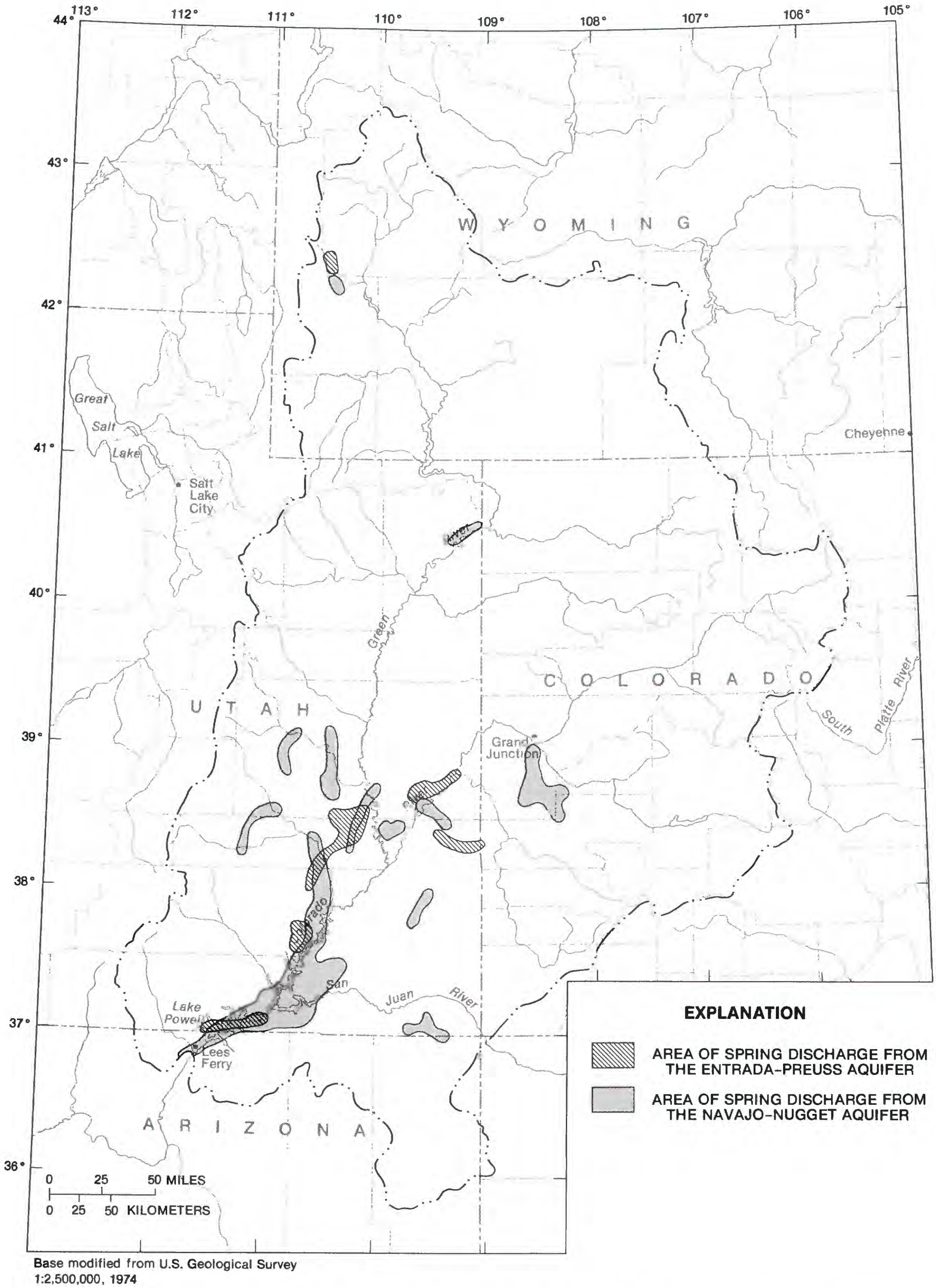


FIGURE 62.—Areas of largest spring discharge from the Navajo-Nugget and Entrada-Preuss aquifers.

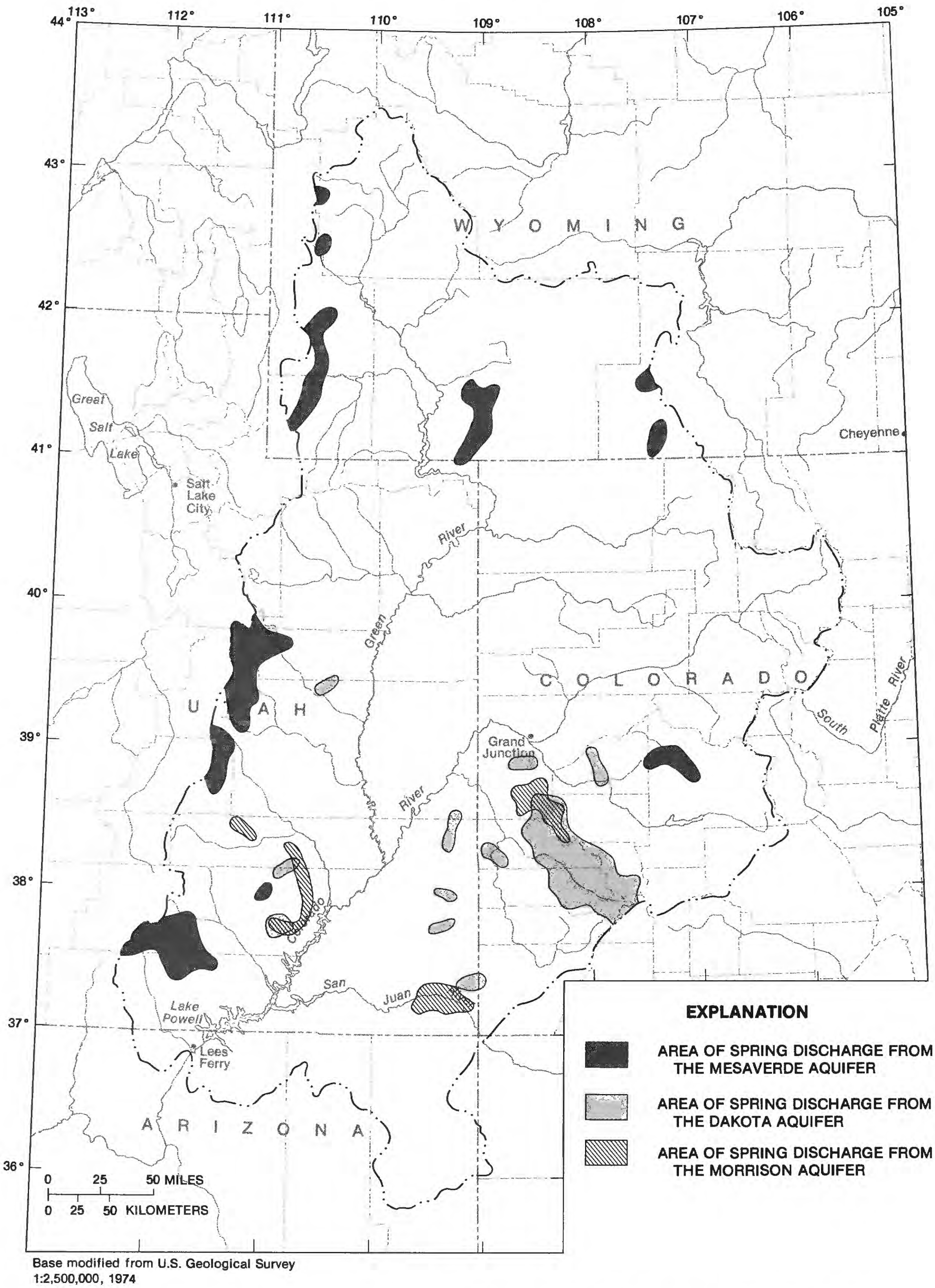


FIGURE 63.—Areas of largest spring discharge from the Morrison, Dakota, and Mesaverde aquifers.

surface of the aquifers. A potentiometric surface that defines a water-level gradient toward a stream indicates that ground water is discharging to that stream.

Based on the water-level configurations of the five aquifers and the areas where they are exposed, discharge to streams from aquifers in Mesozoic rocks takes place in reaches that total about 600 mi (fig. 64). Base flow in the remaining perennial streams in the study area originates from older and younger consolidated aquifers and from the unconsolidated alluvial aquifers.

The Navajo-Nugget aquifer primarily discharges into the main stem of the Colorado River, and secondarily to short reaches of the Green, Dolores, San Juan, Dirty Devil, Escalante, and Paria Rivers, and numerous small creeks. Primary discharge from the Entrada-Preuss aquifer is to small streams flowing from the Uncompahgre Uplift and the San Miguel, Dolores, San Rafael, and San Juan Rivers. Other discharge from this aquifer is to the Dirty Devil and Paria Rivers, and to tributaries flowing from the northwest into the downstream reaches of the Colorado River. The Morrison aquifer discharges to the San Juan, Dolores, San Miguel, and Gunnison Rivers. The Dakota aquifer discharges to the upstream reaches of the Dolores and San Miguel Rivers, and to small streams originating high on the Uncompahgre Uplift. The Mesaverde aquifer probably discharges to Bitter Creek in the Rock Springs Uplift, to the upstream reaches of Muddy Creek on the eastern side of the Washakie Basin, to the Colorado River east of Grand Junction, Colo., to the Price and Green Rivers north of Green River, Utah, and to the White River north of Grand Junction.

Areas in the Upper Colorado River Basin where evapotranspiration occurs are most easily identified by the type and density of vegetation (see Robinson, 1958). Phreatophyte growth is typically dense on the alluvial plains of perennial and some longer intermittent streams where depth to water is less than 15 ft. Salt cedar (*Tamarix gallica*), salt grass (*Distichlis stricta*), and willow (*Salix*) are common in these locations. Greasewood (*Sarcobatus vermiculatus*) and cottonwood (*Populus*) trees grow more sparsely on adjacent stream terraces and near ephemeral-stream channels, where they obtain water at depths as much as 60 ft. Evapotranspiration is also significant in spring areas. Hood and Danielson (1981, p. 44) estimated that more than 10 percent of evapotranspiration in the lower Dirty Devil River basin occurs at springs away from streams, and this may hold true for the remainder of the study area.

To locate areas where evapotranspiration is causing ground-water discharge from the Mesozoic rocks, the direction of ground-water movement and the proximity of the water table to the land surface need to be considered. It is conceivable that phreatophytes growing within certain reaches of perennial stream channels do not extract water from aquifers in Mesozoic rocks, but from the alluvial deposits. Nevertheless, if the configuration of the potentiometric surface of any of the aquifers indicates ground-water movement toward a river, then part of the discharge from that aquifer probably contributes to evapotranspiration (fig. 65*A*). If the opposite is true, then part of the water from the river recharges the aquifer and probably also is intercepted by evapotranspiration (fig. 65*B*). Based on satellite false-color imagery for July, August, and September, the areas of dense vegetation where Mesozoic aquifers are most likely to incur large evapotranspiration losses are along the perennial stream channels in the southern half of the study area (fig. 66). Evapotranspiration losses of smaller magnitude likely take place on terraces and upland areas adjacent to these stream channels. These aquifers are not extensively exposed in the northern half of the study area, thus decreasing evapotranspiration losses there considerably.

Because the study area boundaries are topographically high, they are usually recharge rather than discharge areas. Discharge across the boundary of the study area probably occurs near the exit of the Colorado River at Lee Ferry. Water-level contours for the Dakota aquifer indicate that ground water may also flow north out of the Great Divide Basin. The rate of vertical leakage to younger and older rocks is probably small because of the flow-retarding aspects of bedding planes and the small hydraulic conductivity of younger and older rocks. This vertical discharge may occur in areas where hydraulic heads in the aquifers in Mesozoic rocks are higher than the heads in the aquifers in overlying Cenozoic rocks or the aquifers in the underlying Paleozoic rocks (fig. 67). Extensive vertical fracturing of intervening confining layers would increase probability of this ground-water movement.

QUANTITY

Differing methods were used to determine the quantity of ground-water discharge. Spring discharge was determined by summing the recorded discharge of all springs originating in the Mesozoic rocks and estimating the discharge from unrecorded springs. Discharge from about 750 springs and seeps was recorded. Total discharge from these springs and seeps is about 24,000 acre-ft. These springs probably account for the large, most easily located discharge points, but it is likely that thousands of unrecorded seeps are located at small wetted areas supporting a small population of phreatophyte growth. Aggregate discharge from these unrecorded seeps is not known, but based on an average discharge of 0.1 gallon per minute (gal/min) per seep, the

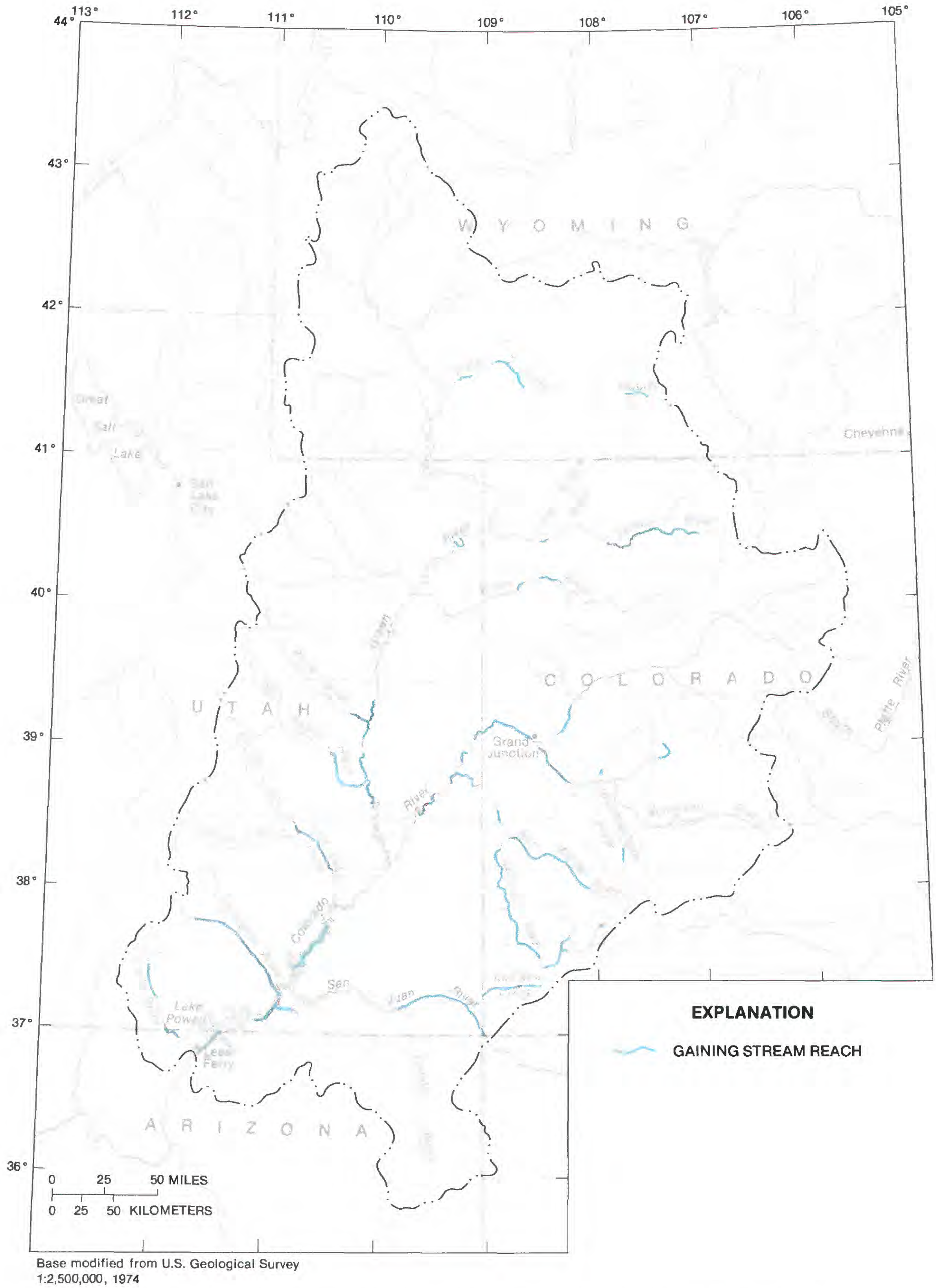
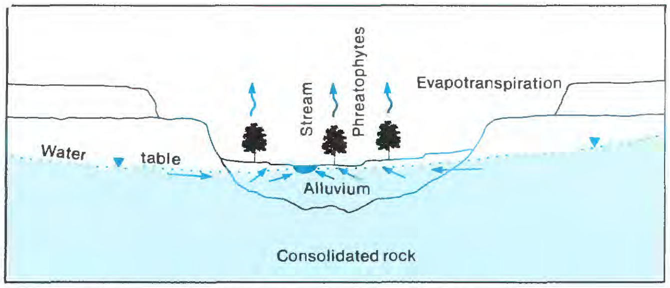
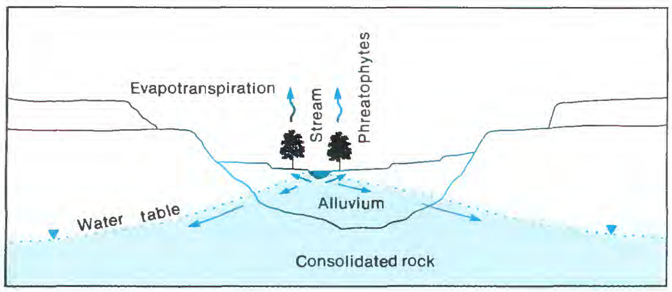


FIGURE 64.—Reaches of perennial streams that receive water from aquifers in Mesozoic rocks.



A. AQUIFER DISCHARGE = STREAM GAIN + EVAPOTRANSPIRATION



B. AQUIFER RECHARGE = STREAM LOSS − EVAPOTRANSPIRATION

FIGURE 65.—Diagrammatic sections showing the effect of evapotranspiration in aquifer-stream relations (arrows show direction of water flow).

annual discharge from these unrecorded seeps is estimated to be about 25,000 acre-ft.

The quantity of ground water discharging to streams has been estimated by previous investigators using low-flow records, seepage investigations, base-flow measurements, and calculations using the ground-water flow equation. Results are summarized in table 4. Total discharge from the aquifers in Mesozoic rocks to streams is estimated to be about 1,000 acre-ft for a 1-mi length of aquifer along the streams.

Evapotranspiration of water directly from the five aquifers in Mesozoic rocks can occur only where the capillary fringe of water in that aquifer reaches a shallow depth below the land surface or where the roots of phreatophytes penetrate near the water table. Flood plains of larger streams commonly support dense growths of riparian vegetation. In stream reaches that receive discharge from underlying aquifers, evapotranspiration decreases the quantity of ground water discharging to the stream.

The quantity of ground water consumed by vegetation was determined from results of previous evapotranspiration investigations in the area covered by phreatophytes where aquifers in Mesozoic rocks crop out along stream channels and near springs. The area of dense phreatophyte growth has been estimated from satellite imagery to be about 51,000 acres. The area of additional losses, where phreatophyte growth is less dense, has been estimated to be about 10 times the area where dense growth occurs. This estimate is based on comparisons of the total area of greasewood population mapped by Hackman (1973) for part of south-central Utah along the Fremont, Muddy, and San Rafael Rivers, and the total area identified as dense phreatophyte growth from satellite imagery along the same reaches of river.

Average annual evapotranspiration rates reported in Robinson (1958, p. 61–75) for vegetation densities of 100 percent vary from 1.3 to 8.3 ft, depending on type of vegetation, depth to water, and length of growing season. Because vegetation density infrequently approaches 100 percent, a typical range of evapotranspiration values applicable for the Upper Colorado River Basin would be smaller. Rush and others (1982, p. 32) used 0.3 ft for greasewood and 3.0 ft for salt cedar, cottonwood, and willow for the central part of the study area. Hood and Danielson (1981, p. 44) estimated aggregate consumptive use of ground water by phreatophytes in the Dirty Devil River basin to be 0.5 ft. A range of values from 0.2 to 3.5 ft was used to estimate evapotranspiration of ground water in the northern Uinta Basin (Hood and Fields, 1978, p. 43). If the minimum and maximum rates from these previous investigations are used to calculate the quantity consumed by dense vegetation, annual ground-water discharge from aquifers in Mesozoic rocks by evapotranspiration would range from 150,000 to 180,000 acre-ft. Water used annually by the less dense greasewood community would range from 100,000 to 250,000 acre-ft. Total annual evapotranspiration by vegetation would range from 250,000 to 430,000 acre-ft.

The quantity of water that may be moving from the aquifers in Mesozoic rocks into overlying Tertiary formations and underlying Paleozoic formations is small. Using estimates for hydraulic gradients, hydraulic conductivity, thickness of confining layers, and area of leakage, the resulting annual discharge is about 5,900 acre-ft (table 5). Because of the uncertainty in estimated values for vertical hydraulic conductivity and for differences in hydraulic head between aquifers, the calculated flow between aquifer systems could be in error by several orders of magnitude. In fact, the extremely large apparent hydraulic head differences in the Washakie and Great Divide Basins indicate that vertical hydraulic conductivity values in these areas are significantly smaller than the estimated values of 0.001 foot per year (ft/yr).

Based on the hydraulic gradient, the cross-sectional area, and the range of hydraulic-conductivity values applicable to the Navajo Sandstone near Lake Powell (Thomas, 1985, p. 23), outflow of ground water to the

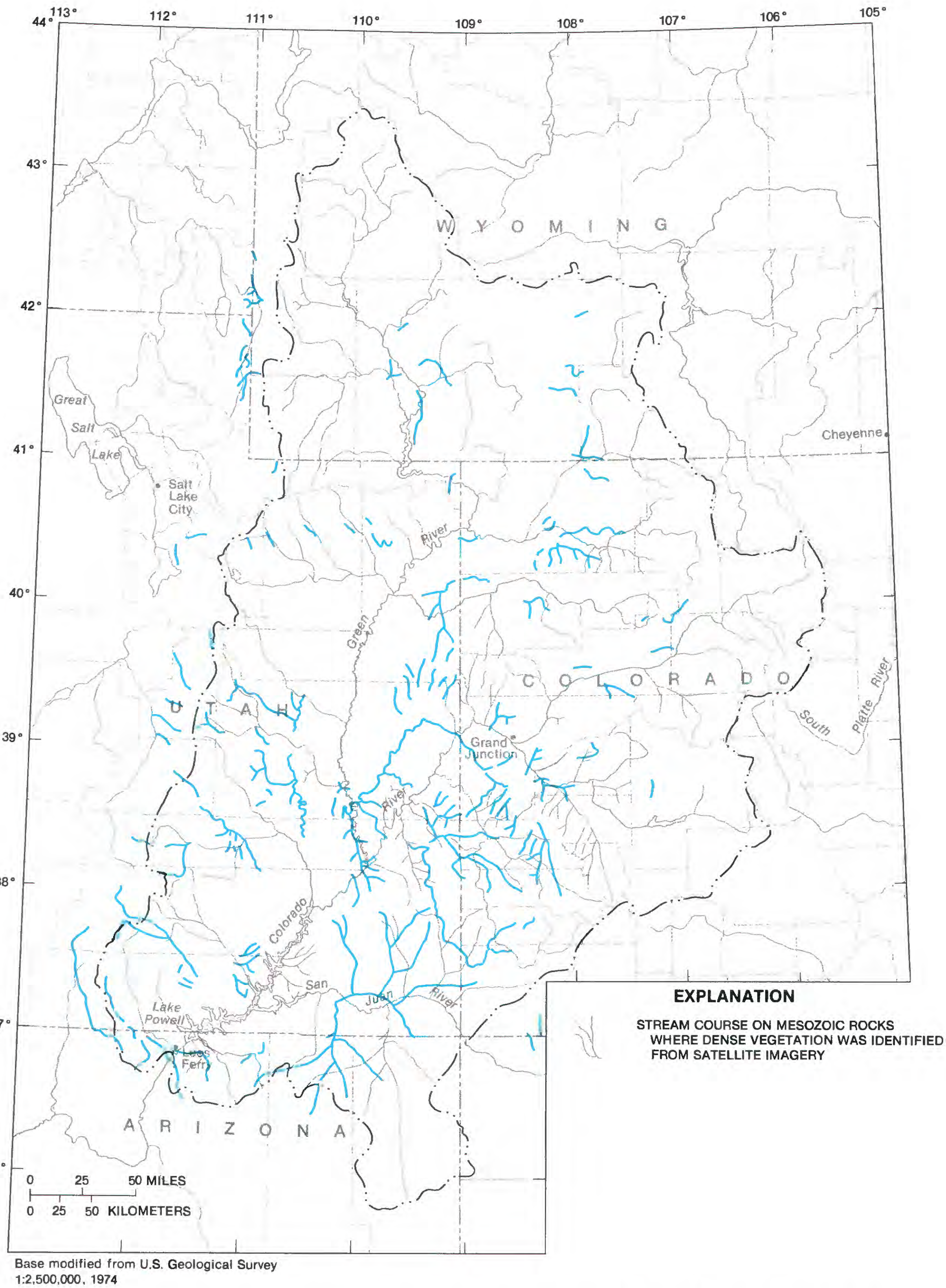
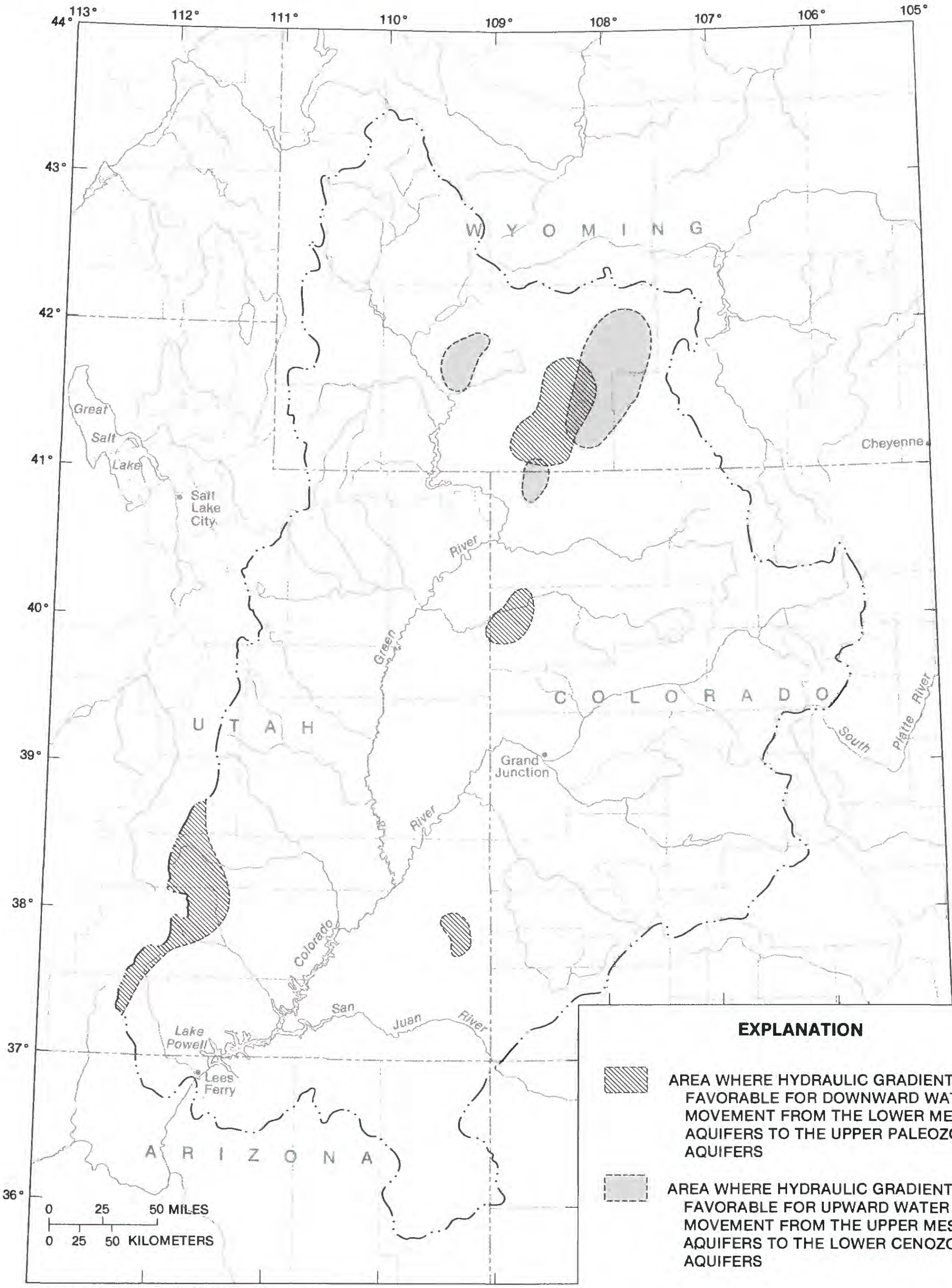


FIGURE 66.—Stream courses on Mesozoic rocks where dense vegetation was identified from satellite imagery.



Base modified from U.S. Geological Survey 1:2,500,000, 1974

FIGURE 67.—Areas of possible vertical water flow from aquifers in Mesozoic rocks into underlying Paleozoic and overlying Cenozoic rocks.

TABLE 4.—*Estimated water discharge from aquifers in Mesozoic rocks to principal streams*

Name of stream: Listed in order of decreasing average annual discharge.
Stream reach: Approximate geographic location.
Length of aquifer: Represents distance of aquifer rather than stream length.
Ground-water discharge: Values rounded to nearest 1,000 acre-feet per year.

Stream	Stream reach	Contributing aquifer	Length of aquifer (miles)	Ground-water discharge (thousands of acre-ft per year)	Source of discharge estimate—Remarks
Colorado River	Glenwood Springs to Cameo, Colo.	Mesaverde	12	17	From streamflow gains calculated for the average discharge in September 1949 through 1958
	Cameo, Colo., to Cisco, Utah	Mesaverde	2	3	Calculated based on same per-mile gain in stream discharge observed between Glenwood Springs and Cameo, Colo.
		Dakota	6	8	
		Morrison	2	3	
		Entrada-Preuss	7	9	
		Navajo-Nugget	2	3	
	Cisco to Hite, Utah	Morrison	8	15–30	Smaller values are from streamflow gains calculated for average discharge for September 1949 through 1958. Larger values from Rush and others (1982, tables 8 and 9)
		Entrada-Preuss	1	2–4	
		Navajo-Nugget	9	17–33	
	Hite, Utah, to Lees Ferry, Ariz.	Entrada-Preuss	20	27–33	Ditto.
		Navajo-Nugget	40	55–66	Rush and others (1982, table 10)
Green River	Greendale to Jensen, Utah	Dakota	2	4	From streamflow gains calculated for average discharge for September 1951 through 1958
		Morrison	4	8	
		Entrada-Preuss	2	4	
		Navajo-Nugget	2	4	
	Jensen to Ouray, Utah	Mesaverde	2	5	Do.
	Ouray to Green River, Utah	Mesaverde	18	17	Do.
	Green River, Utah, to confluence with Colorado River	Dakota	1	1	Values determined from seepage investigations in 1948 (Thomas, 1952), and estimated by Rush and others (1982, table 7)
		Morrison	2	2	
		Entrada-Preuss	6	5	
		Navajo-Nugget	10	8	
San Juan River	Four Corners to Mexican Hat, Utah	Morrison	25	5–23	Smaller values determined from base flow for July 1959 (Whitfield and others, 1983, p. 42). Larger values are from streamflow gains calculated for average discharge in September 1949 through 1958
		Entrada-Preuss	14	3–13	
		Navajo-Nugget	7	1–6	
	Mexican Hat, Utah, to confluence with Colorado River	Navajo-Nugget	5	7–8	Values based on per-mile ground-water inflow rates to the Colorado River between Hite, Utah, and Lees Ferry, Ariz.
	Tributaries to San Juan River	Morrison	8	6	Determined from results of base-flow measurements in 1982 and 1983 (Avery, 1985, p. 43, 57, and 64)
		Entrada-Preuss	2	Total	
		Navajo-Nugget	7		
Gunnison River	Black Canyon of Gunnison to Grand Junction, Colo.	Dakota	3	4	From streamflow gains calculated for average discharge in September 1949 through 1958
		Morrison	12	25	
		Entrada-Preuss	12	25	
Yampa River	Steamboat Springs to Maybell, Colo.	Mesaverde	24	26	Value based on the mean per-mile rate (1,070 acre-feet per year per mile of aquifer) of ground-water inflow to all other streams in Upper Colorado River Basin
White River	Meeker, Colo., to Ouray, Utah	Mesaverde	17	8	From streamflow gains calculated for average discharge in September 1949 through 1958
Dolores River	Rico to Dolores, Colo.	Dakota	2	1–4	Small values based on per-mile rate of aquifer discharge as reported by Weir (Weir and others, 1983, p. 32). Large values from streamflow gains calculated for average discharge in September 1949 through 1958
		Morrison	13	5–24	
	Dolores to Gateway, Colo.	Dakota	2	Negligible	Values bases on per-mile rate of aquifer discharge as reported by Weir (Weir and others, 1983)
		Morrision	17	2	
		Entrada-Preuss	16	2	
		Navajo-Nugget	30	4	
	Gateway, Colo., to confluence with Colorado River	Morrison	4	2	Do.
		Entrada-Preuss	2	1	
		Navajo-Nugget	3	1	

TABLE 4.—*Estimated water discharge from aquifers in Mesozoic rocks to principal streams*—Continued

Stream	Stream reach	Contributing aquifer	Length of aquifer (miles)	Ground-water discharge (thousands of acre-ft per year)	Source of discharge estimate—Remarks
San Miguel River	Placerville to Uravan, Colo.	Morrison	25	19	As reported by Ackerman and Rush (1984, p. 19) from data of Iorns and others (1965, p. 53)
		Entrada-Preuss	55	41	
Uncompahgre River	Ridgeway to Delta, Colo.	Morrison	8	9	Value based on the mean per-mile rate of ground-water inflow to all other streams in Upper Colorado River Basin
San Rafael River	San Rafael Swell to confluence with Green River	Morrison	5	5	Do.
		Entrada-Preuss	15	16	
		Navajo-Nugget	7	7	
Dirty Devil River	Hanksville, Utah, to confluence with Colorado River	Entrada-Preuss	10	7	Calculated using approximate values for the components of Darcy's Law from Hood and Danielson (1981, p. 37)
		Navajo-Nugget	9	7	
Price River	Woodside, Utah, to confluence with Green River	Mesaverde	10	11	Value based on mean per-mile rate of ground-water inflow to all other streams in Upper Colorado River Basin
Escalante River	Escalante, Utah, to confluence with Colorado River	Navajo-Nugget	55	30	From streamflow gains calculated for average discharge in September 1950 through 1955
Muddy Creek	East side of Washakie Basin	Mesaverde	10	6	Calculated using approximate values for the components of Darcy's Law. Water-level gradient = 0.007 ft/ft. Hydraulic conductivity = 1 ft/d. Saturated thickness = 1,000 ft
Bitter Creek	Rock Springs Uplift	Mesaverde	14	5	Do. Water-level gradient = 0.004 ft/ft. Hydraulic conductivity = 1 ft/d. Saturated thickness = 1,000 ft
Total			606	508–614	

southwest from the study area near Lee Ferry is estimated to range from 100 to 600 acre-ft per year. Lateral outflow to the north from the Great Divide Basin is possible, but the quantity would be small. Total lateral flow across the study-area boundary is estimated to be about 1,000 acre-ft per year.

TABLE 5.—*Estimated vertical flow from aquifers in Mesozoic rocks based on the ground-water flow (Darcy) equation:* $Q=K_v\ A\ dh/dl$

Area affected: Approximate geographic area shown in figure 67.
Vertical hydraulic conductivity (K_v): Typical vertical hydraulic conductivity for the rock through which ground water moves.
Size of area (A): Estimated from figure 67.
Difference in hydraulic head (dh): Representative head difference between the lowermost aquifer in Tertiary rocks and the uppermost aquifer in Mesozoic rocks, or between the uppermost aquifer in Paleozoic rocks and the lowermost aquifer in Mesozoic rocks.
Distance of vertical flow (dl): Representative vertical distance water must move from one aquifer to another.
Quantity of flow between aquifers (Q): Flow calculated using the flow equation.

Area affected	Vertical hydraulic conductivity (feet per year)[1] K_v	Size of area (acres) A	Difference in hydraulic head (feet) dh	Distance of vertical flow (feet) dl	Quantity of flow between aquifers (acre-feet per year) Q	Direction of flow between aquifers
Washakie Basin	0.001	1,500,000	2,000	1,200	2,500	Mesozoic to Paleozoic
Piceance Basin	.001	300,000	1,500	700	640	Do.
Monticello, Utah, area	.001	200,000	1,500	1,000	300	Do.
High Plateaus area	.001	1,000,000	1,000	1,600	620	Do.
Northern Washakie and Great Divide Basins	.001	700,000	2,000	1,000	1,400	Mesozoic to Cenozoic
Southern Rock Springs Uplift	.001	150,000	600	1,000	90	Do.
Green River Basin	.001	400,000	800	1,000	320	Do.
Total (rounded to the nearest 100 acre-feet per year)					5,900	

[1]Values for claystone, siltstone, and shale range from about 0.000001 to 0.01 foot per year (Davis and DeWiest, 1966, p. 349; Morris and Johnson, 1967, p. D36; Freeze and Cherry, 1979, p. 29). The value of 0.001 foot per year represents the most common value for all three confining units.

STORAGE

Ground-water storage can be estimated in terms of total, drainable, or recoverable storage. Total storage is all water contained in the connected, open, intergranular spaces, fractures, and solution cavities of the aquifers and confining units, regardless of quality or depth. It is the product of porosity and the volume of saturated material. If porosity is assumed to decrease by about 1 percent for every 1,000 ft of depth, then total storage in the aquifers in the Mesozoic rocks of the Upper Colorado River Basin could range from 10 to 12 billion acre-ft. Drainable storage is the ground water that would drain by gravity. It is the product of the specific yield and the volume of saturated material. Hood and Patterson (1984, p. 20) estimate specific yield of the Navajo-Nugget aquifer to be 50 percent of the average porosity. If this estimate is used for all aquifers in Mesozoic rocks, drainable storage is 5 to 6 billion acre-ft. Storage volumes were calculated by computer, by generalizing porosity, specific-yield, and saturated-thickness values for 15- by 15-minute (about 17- by 14-mi) grid blocks of the study area. Porosity was decreased by 1 percent for every 1,000 ft of depth. This estimate was made on the basis of the assumption that the water is physically drained from pore spaces of the rock. However, there is also drainable water due to elastic storage as hydraulic head in the aquifers is reduced. Because of the complexity of calculation, this elastic-storage release was not included in the calculation of drainable storage; however, the value would be very small, within the range of error of the previous calculation.

In practice, only a small part of this drainable storage is recoverable because of the difficulties in pumping it from the aquifer, and an even smaller part is usable because of the spatial variability in its quality. To obtain a general estimate of the volume of usable, recoverable water within the aquifers in Mesozoic rocks, several qualifying assumptions were applied:

1. The depth of wells used to recover the water would not be more than 2,000 ft.
2. Only one-half of the saturated thickness penetrated can be dewatered.
3. Specific yield of the aquifers is 50 percent of their average porosity.
4. Only ground water having a dissolved-solids concentration of less than 3,000 mg/L would be pumped.

Based on these assumptions, the quantity of recoverable, fresh to slightly saline water in the upper 2,000 ft of aquifers in Mesozoic rocks is about 530 million acre-ft, about 4 percent of total volume of ground water in storage.

QUALITY OF GROUND WATER

Locally, the quality of water in the aquifers and confining units in Mesozoic rocks of the Upper Colorado River Basin is as diverse as the depositional and structural history of the formations that form the aquifer system. Whether the ground-water quality is acceptable depends on its intended use. Water used for agricultural purposes can be of poorer quality than water for domestic use, and water suitable for certain industrial purposes may be extremely bad for crops. A detailed description of the quality of water found in the aquifers and confining units in Mesozoic rocks is contained in a companion report on the geochemistry of the aquifers in the Upper Colorado River Basin. The following discussion is generalized and of limited scope.

DISSOLVED SOLIDS

Concentrations of dissolved solids in water from various Mesozoic formations vary considerably. Concentrations nearly as small as that in rainwater have been measured, as have concentrations almost five times larger than that in average seawater. As noted in Freethey and others (1988), large concentrations of dissolved solids are present in water from deeply buried Mesozoic rocks far from recharge areas. This water is mainly a sodium chloride type. Small concentrations are in water from formations lying near land surface that are in or near recharge areas. This water is mainly a calcium bicarbonate type. Significant local exceptions are near the southern end of the San Rafael Swell, where dissolved-solids concentrations exceed 50,000 mg/L in several samples from the Navajo-Nugget aquifer. To demonstrate the lateral variation of water quality, a map modified from Freethey and others (1988) that represents the generalized quality of water in the Navajo-Nugget and the Entrada-Preuss aquifers is shown in figure 68. The deep structural basins in the northern half of the study area are well defined by the distribution of brine.

MAJOR CONSTITUENTS

The chemical constituents that account for most of the total concentration of dissolved solids are sodium, calcium, magnesium, potassium, bicarbonate, carbonate, sulfate, chloride, and silica. In this report, these ions are termed “major constituents.” Calcium and bicarbonate ions are most common in ground water in and near recharge areas. Sodium and chloride ions are present in large concentrations in water at depth, where movement is slow and recharge areas are distant (Freethey and others, 1988).

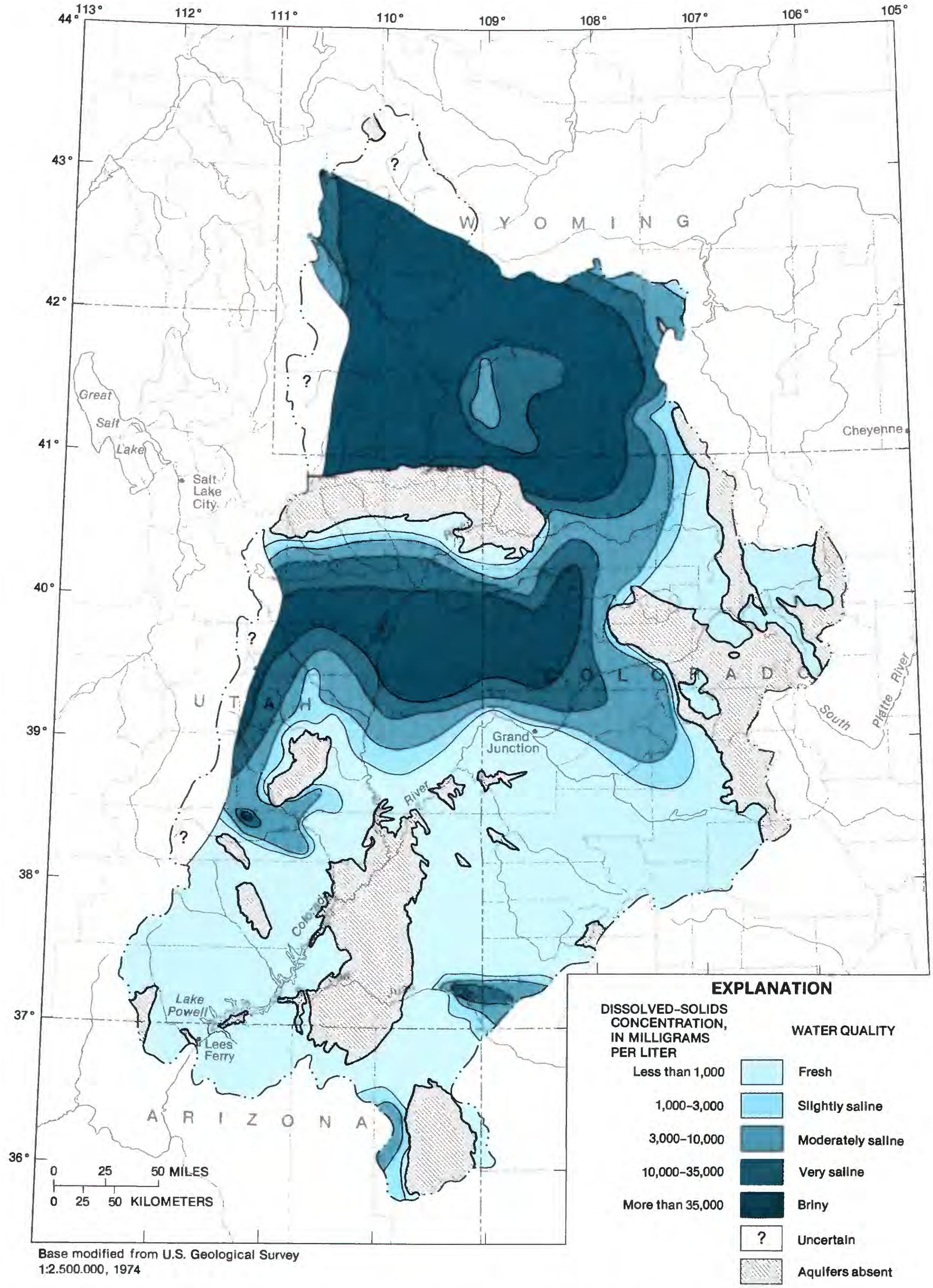


FIGURE 68.—Quality of water (based on the dissolved-solids concentration) in the Navajo-Nugget and Entrada-Preuss aquifers (from Freethey and others, 1984, fig. 19).

Differences in the relative concentration of the major constituents are evident in waters of varying salinity and, to a lesser degree, in waters from aquifers versus water from confining units (fig. 69). The increase in the relative concentration of chloride, sodium, and potassium, as salinity increases, is most evident. The relative concentrations of bicarbonate, carbonate, calcium, magnesium, and silica decrease with the increase in dissolved solids. Water from confining units contains a larger concentration of sulfate than does water from aquifers. Briny water from confining units is a sodium sulfate type, and briny water from aquifers is a sodium chloride type.

Comparison of the mean values of the concentrations of major constituents in water from four aquifers (fig. 70) indicates that the Dakota and Mesaverde aquifers generally have larger percentages of dissolved calcium and bicarbonate than the Navajo-Nugget and Entrada-Preuss aquifers. This probably is the result of the proximity of the Dakota and Mesaverde aquifers to recharge sources. Younger units are less deeply buried and more easily recharged.

The same comparison of confining units (fig. 71) shows virtually no difference in the relative proportions of major chemical constituents. The size of the diagrams indicates that total concentrations of the major constituents are larger in the Curtis-Stump and Morrison confining units than in the Chinle-Moenkopi and Mancos confining units.

MINOR CONSTITUENTS

Chemical constituents that have concentrations of less than 0.1 mg/L (100 micrograms per liter (μg/L)) are termed "minor constituents." These are primarily metals that at larger concentrations can be toxic or cause undesirable staining. The mean, its spread based on the standard error of the mean, and the maximum concentration measured in water from aquifers and confining units that had more than 40 analyses on record are shown in figure 72. The mean values depicted in these graphs represent samples from shallow water wells and deep petroleum test wells. Large chemical concentrations in the brine samples from these deep wells tend to bias the means toward much larger values than would be normal for domestic water samples. Conclusions drawn from a mix of shallow-aquifer samples with deep-brine samples admittedly are questionable. However, the graphs indicate that for the aquifers shown, even with this added bias by the brine samples, mean concentrations of boron, selenium, and lead still do not exceed the maximum contaminant levels for drinking water set by the U.S. Environmental Protection Agency (1976). Mean concentrations of manganese and iron slightly exceed the standards. Samples in which the concentration of selenium exceeds the maximum level set by the Environmental Protection Agency are mostly from aquifers closely associated with uranium ore deposits.

GROUND-WATER BUDGET

The components of a ground-water budget, even for a small localized area, cannot be measured in the Upper Colorado River Basin. Estimating these components requires simplification of a complex chain of interrelated hydrologic processes. The simplified estimates of recharge and discharge into and out of aquifers in Mesozoic rocks in the study area are presented in table 6. About 93 percent of recharge is from infiltration of precipitation that falls on outcrop areas or thinly covered areas of the Mesozoic rocks, or infiltrates through overlying sediment and vertically leaks into the buried Mesozoic rocks. Nearly 6 percent of recharge occurs along losing-stream reaches or comes from other surface-water sources. The remaining 1 percent is vertical leakage from underlying and overlying rocks and lateral boundary flow into the study area.

About 59 percent of the water in aquifers in Mesozoic rocks discharges to streams, and about 36 percent is consumed by evapotranspiration. About 4 percent discharges to springs, and only about 1 percent discharges as vertical leakage or outflow across the study-area boundary.

FACTORS AFFECTING GROUND-WATER DEVELOPMENT IN MESOZOIC ROCKS

Development of the ground-water resources in the Upper Colorado River Basin has to date (1987) been minor. Several reasons for this lack of development are evident. Average precipitation and recharge to the aquifers are small. Thus, the ground-water resources in the Upper Colorado region are also small. The consolidated aquifers, overall, have hydraulic properties that preclude large-scale ground-water development. At present, withdrawal of ground water from deep buried Mesozoic rocks is costly throughout large parts of the study area. The quality of the ground water in large parts of the area is only marginally suitable for certain industrial purposes and is unsuitable for domestic and agricultural purposes. Despite these negative aspects, ground water is available for some uses.

Aquifers and confining layers that compose the aquifers in Mesozoic rocks are present in about 85 percent of the study area. They contain an estimated 530 million acre-ft of recoverable water containing less than 3,000 mg/L dissolved solids. The maps referred to in this

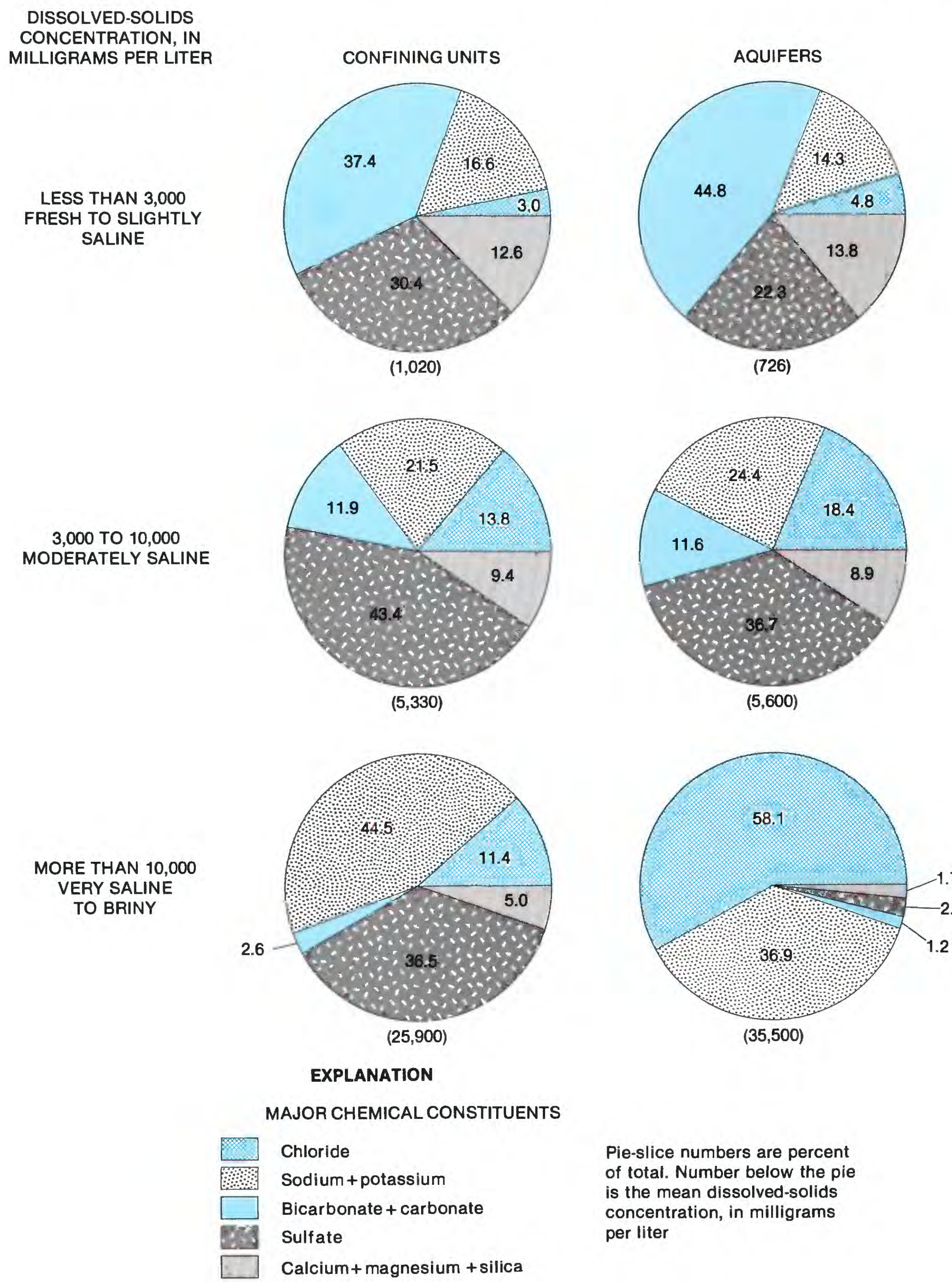


FIGURE 69.—Comparison of average concentration of major chemical constituents in fresh to briny water from confining units and aquifers.

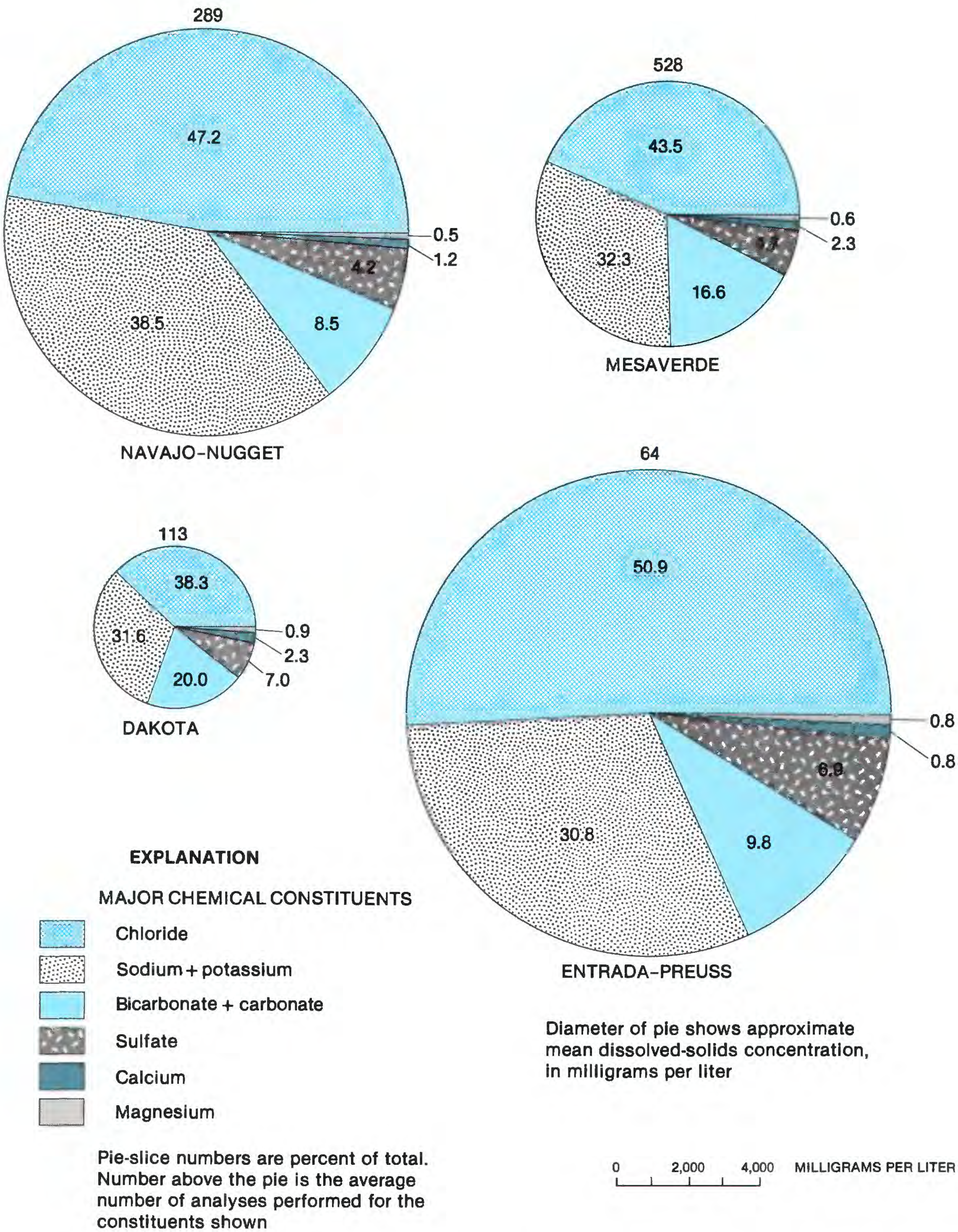


FIGURE 70.—Comparison of major chemical constituents in water from four aquifers in Mesozoic rocks.

section summarize the effects of the different components of the geohydrological environment on the regional aquifer system. Each component affects some aspect of the ground-water resource to different degrees or in different ways.

PERMEABILITY RESULTING FROM FRACTURING

Large-scale structural features such as basins and uplifts may impede regional movement of ground water.

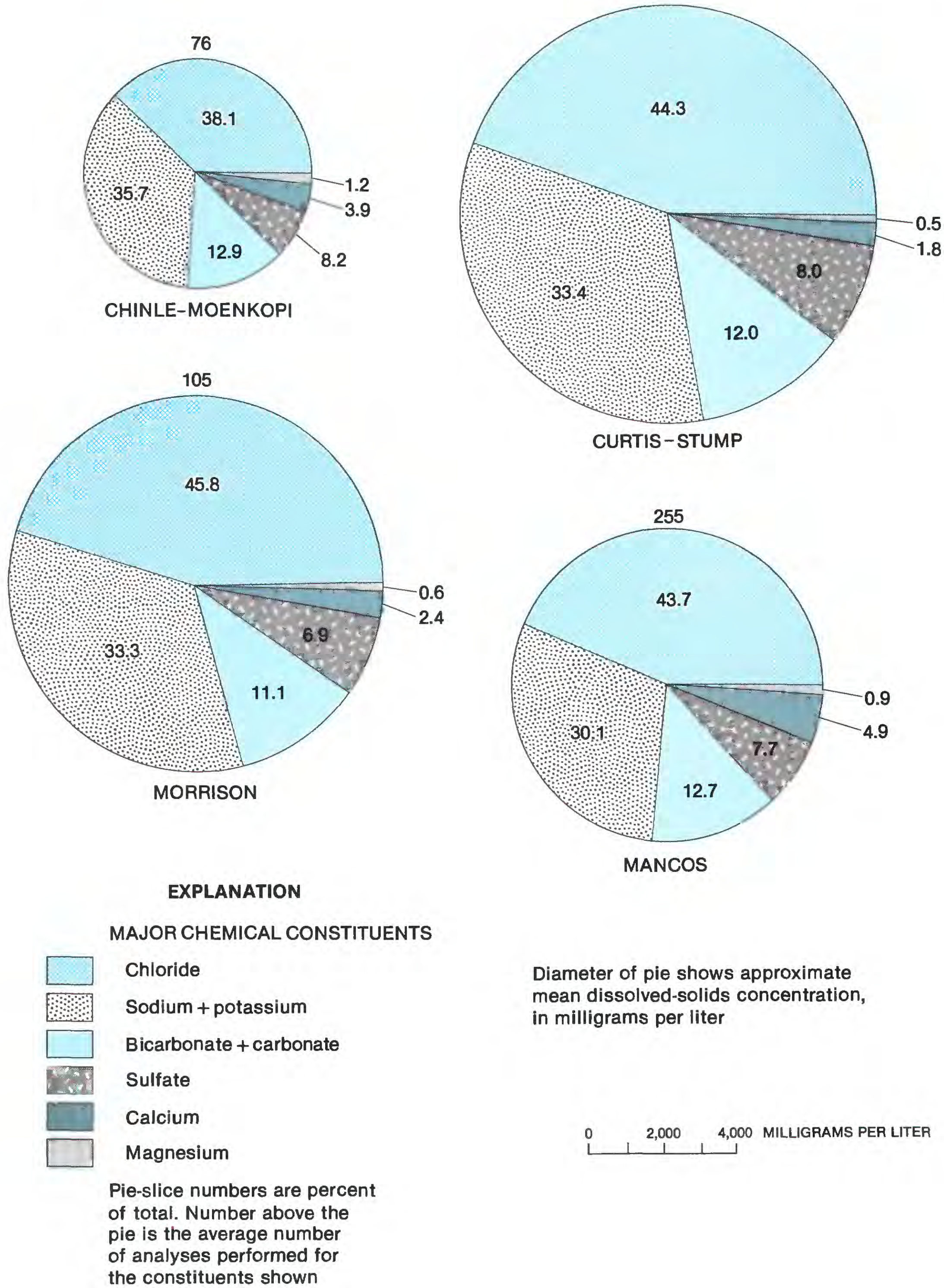


FIGURE 71.—Comparison of major chemical constituents in water from four confining units in Mesozoic rocks.

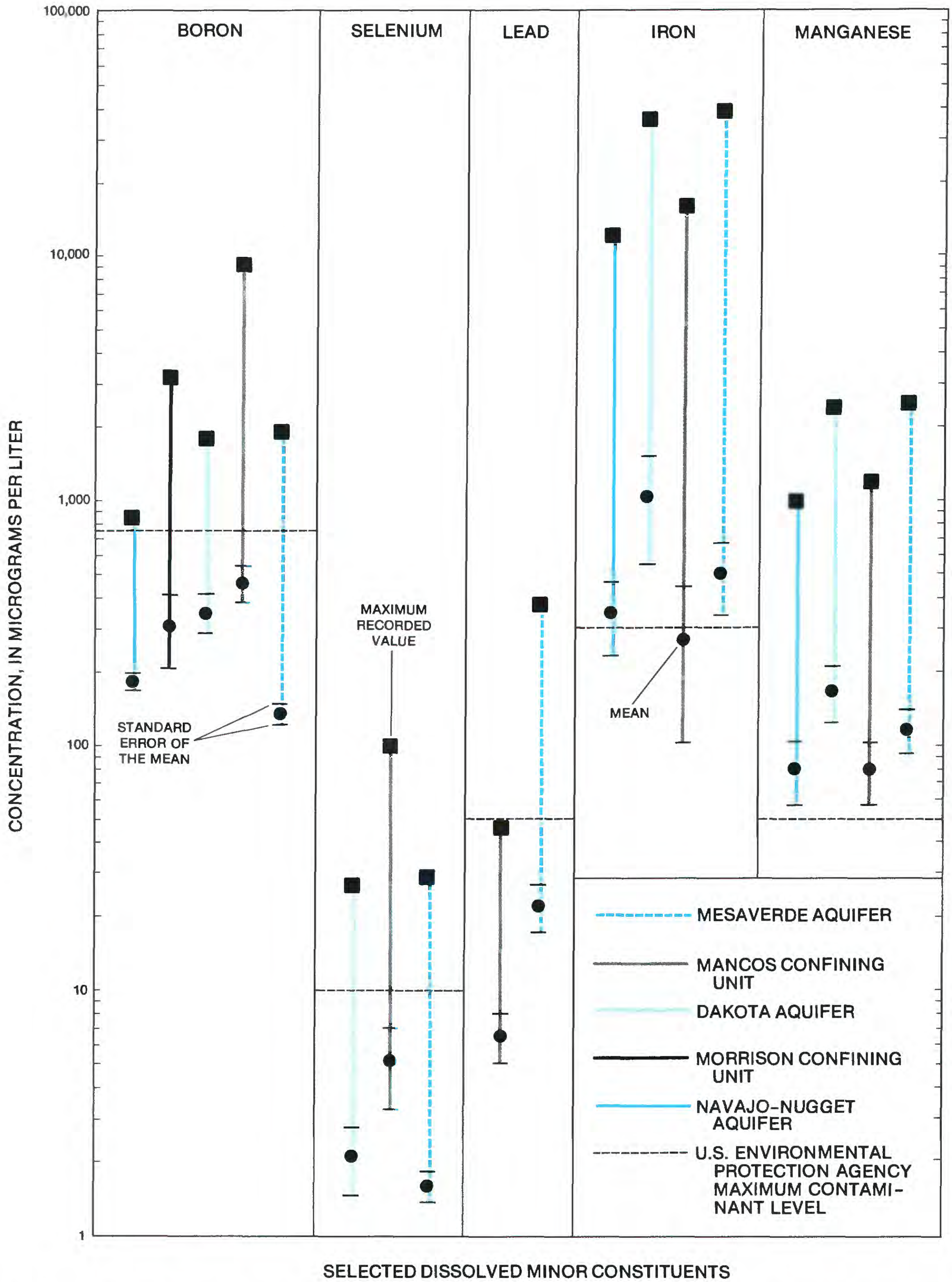


FIGURE 72.—Mean, standard error of the mean, and maximum recorded values of selected minor constituent concentrations in selected aquifers and confining units in Mesozoic rocks.

TABLE 6.—*Water budget for the hydrologic system in the Mesozoic rocks*

[Estimated quantities, in acre-feet per year: e, estimated; r, calculated using most likely rate of stream loss; ?, no estimate made; m, median between the minimum and maximum values; v, calculated using median evapotranspiration rate]

Budget components	Estimated quantities			Method used or source of estimate
	Minimum	Most reasonable value	Maximum	
Inflow:				
Precipitation	580,000	885,000e	3,300,000	Infiltration study by Danielson and Hood (1984); Maxey-Eakin method (Maxey and Eakin, 1951)
Recharge from stream loss	1,500	58,000r	290,000	Stream-seepage investigations by U.S. Geological Survey in south-central and southeastern Utah
Vertical leakage from Paleozoic and Tertiary rocks into Mesozoic rocks	0	1,100	?	Flow calculated using Darcy's Law
Lateral inflow across boundary of study area.	0	3,500	?	Flow calculated using Darcy's Law
Total (rounded to 2 significant digits)		950,000		
Outflow:				
Discharge to streams	508,000	561,000m	614,000	Calculation of stream gains and losses based on streamflow records
Evapotranspiration	250,000	344,000v	434,000	Estimate based on measured areal coverage of phreatophytes from false-color satellite imagery in conjunction with published vegetation maps
Spring discharge	24,000	36,000e	49,000e	Records of U.S. Geological Survey
Vertical leakage from Mesozoic rocks into Paleozoic and Tertiary rocks	0	5,900	?	Flow calculated using Darcy's Law
Lateral outflow across boundary of study area	0	1,000	?	Flow calculated using Darcy's Law
Total (rounded to 2 significant digits)		950,000		

Fault zones form conduits or barriers, altering the rate and direction of local and regional flow. In particular, fracture zones associated with folding and faulting generally create openings and increase permeability. However, the effectiveness of the openings may be decreased by chemical precipitates filling the fractures or by the closing of fractures at great depth as a result of high temperatures and pressures.

Model studies for the San Rafael Swell (Weiss, 1986) and the Lake Powell (Thomas, 1985) and Four Corners areas (Thomas, 1989) demonstrate that average hydraulic-conductivity values that best characterize regional flow in these parts of the study area are small, probably reflecting control by primary pore openings rather than fracture openings. However, fracturing is a dominant factor affecting hydraulic conductivity and ground-water development on a local scale.

Hood and Danielson (1979) demonstrated the importance of fracture permeability on a local basis in a comparison of transmissivity values obtained from laboratory and aquifer tests of the Navajo Sandstone west of Hanksville, Utah, in the southwestern part of the study area. The Navajo Sandstone in this area is folded into an anticline and syncline, and the rocks exhibit closely to widely spaced joints that parallel the trends of the folds. Transmissivities for fractured Navajo Sandstone are 3.3 times greater than those calculated for unfractured rocks tested in the laboratory. Hood and Danielson (1979, p. 30) conclude that secondary permeability, probably the result of fracturing, "has an important effect on well yields and water-level response to pumping."

Geologic studies on a regional basis indicate that sedimentary rocks are more extensively fractured and contain more open fractures along faults, anticlines, monoclines, and flanks of basins and in areas of tightly folded rocks (Cooley, 1986). On a smaller scale, Harris and others (1960, p. 1869) note that the density of fractures on compressional deformational features (such as monoclines, anticlines, and synclines) is largest in areas of maximum curvature (greatest change in dip or strike). They also indicate that the susceptibility of strata to fracturing is dominantly controlled by the thickness and lithology. Thin rock units are more susceptible to fracturing, and the "concentration of fractures is in approximate inverse proportion to the thickness of the individual rock units" (Harris and others, 1960, p. 1856). In addition, the more brittle lithologic types (quartz sandstone, siliceous limestone, dolomite, and shale) show better development and greater density of fractures than more ductile rocks (soft shale, friable quartz sandstone) (Harris and others, 1960, p. 1869).

Fracture openings and permeability are dependent on the aperture of the fractures, their density, and their continuity. Aside from chemical precipitates filling the fractures, the continuity of fractures is affected by the depth at which the fractures occur. Based on a study of the Navajo Sandstone by Nelson and Handin (1977), fracture permeability decreases with depth to about 12,000 ft, at which point it is negligible. In addition, Nelson and Handin state that the first 1,000 ft of overburden have relatively little effect on fracture permeability.

To evaluate the effects of fracturing on overall rock permeability, a map of inferred fracture permeability of Mesozoic rocks was derived (fig. 73). This map is a compilation of three types of data:

1. A potential fracture permeability map by Cooley (1986) based on distribution of structures and lineaments in sedimentary rocks;
2. A compilation of surface-fault frequency determined from 1- × 2-degree (about 68- × 108-mi) geologic maps; and
3. Areas where depth to shallowest aquifer in Mesozoic rock is more than 12,000 ft (thickness of overlying rock is more than 12,000 ft).

Several assumptions provide the basis for the map of inferred fracture permeability. Figure 73 shows the surficial expression of fracturing in the shallowest rocks; however, the relative density of surficial fractures is considered to be representative of the probable density of fracturing at depth. This assumption does not take into account the possibility that some fractures may be filled or "healed" at depths of less than 12,000 ft, thereby decreasing the number of open fractures at depth compared with the number of open surficial fractures. Based on the work of Nelson and Handin (1977), the permeability of open fractures at depths of less than 12,000 ft is assumed to be greater than the permeability of the rock they cut across. Deeper than 12,000 ft, fractures are assumed to be closed; therefore, the fracture permeability is negligible.

Fractured zones tend to yield more water. Thus, the map of inferred fracture permeability indicates broad areas where the potential for greater ground-water yield exists. Six categories of inferred fracture permeability are shown in figure 73. Areas of negligible potential are shown where the shallowest aquifer in Mesozoic rocks is more than 12,000 ft deep. The inferred permeability from fracturing is unknown in those areas where volcanic rocks overlie Mesozoic rocks in Utah and Colorado and where Precambrian igneous and metamorphic rocks have been thrust over younger rocks in Wyoming and Colorado.

Areas identified as category 1A (least fracture permeability) and category 1 are predominantly in the central parts of the structural basins and in the relatively flat lying rocks in parts of the Colorado Plateaus province. Areas identified as category 2 include arches, the perimeters of deep basins, platforms, and uplifts, as well as anticlines between basins (category 1) and uplifts (category 3), and basins with some folding and faulting. Uplifts, monoclines, anticlines, arches, and flanks of basins, each with extensive folding and faulting, are identified as category 3. Category 4 (largest inferred fracture permeability) includes areas where the rocks have been tightly folded and faulted, namely in the Wyoming thrust belt and the southern Park Range of Colorado. Also included in this category are the Paradox Basin and adjacent areas of thick salt deposits where movement or removal of salt by solution has increased local fracturing in the overlying rocks.

HYDROLOGIC CHARACTER

Saturated thickness and hydraulic conductivity are two important hydrologic factors that affect the quantity of water an aquifer can yield. The relative ease with which the storage of an aquifer can be recharged affects that aquifer's capability of yielding water for extended periods. Figure 74 indicates the potential water yield related to saturated thickness and to recharge potential, and figure 75 indicates the potential water yield related to transmissivity. The descriptive words used to indicate potential water yield are relative, applicable only within the study area. The descriptor does not consider the quality of the water or the economic and physical problems associated with well drilling. In general, figures 74 and 75 indicate that the most favorable areas for development of ground water from aquifers in the Mesozoic rocks are the southwestern and southeastern parts of the study area.

WATER-QUALITY CONSTRAINTS

The quality of ground water is often judged by its intended use. Generally, the most stringent standards for quality are for water for domestic uses. Standards are usually, but not always, less strict for recreational, agricultural, and industrial uses. Maps showing the areal distribution of the concentration of dissolved solids (fig. 76) are a good general indicator of how usable the ground-water resources are, based on quality. If quality is the only consideration, areas where all or most water samples contain less than 2,000 mg/L dissolved solids would be most favorable for development. Areas of dissolved-solids concentrations of more than 10,000 mg/L would be least favorable. Because part of the information shown in figure 76 was obtained from samples collected from petroleum test wells, the generalization expressed by that figure may erroneously indicate the water quality in the aquifers in petroleum-producing areas. Areas where data are not available are primarily in the struc-

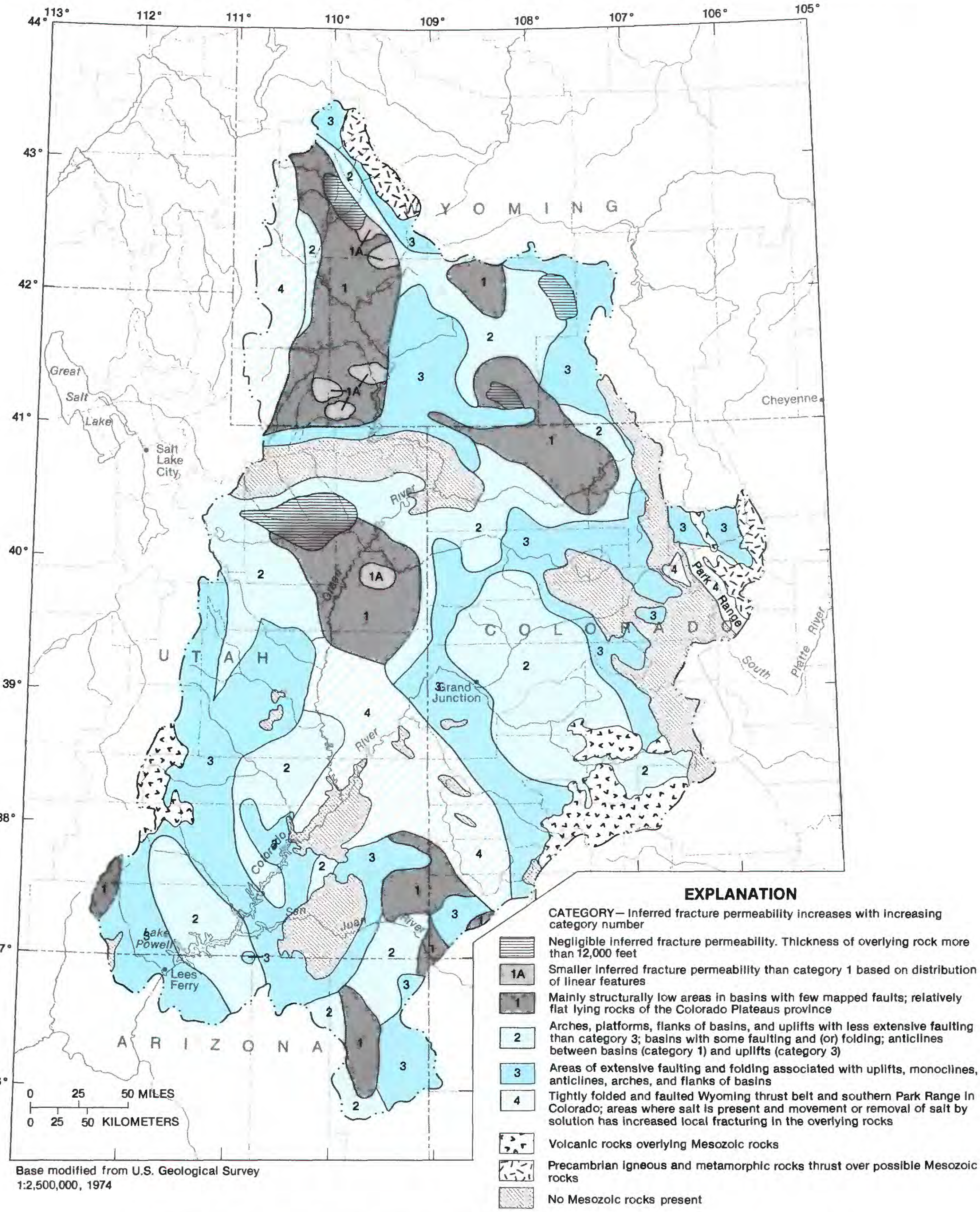


FIGURE 73.—Inferred fracture permeability of Mesozoic rocks (modified from Cooley, 1986, pl. 1).

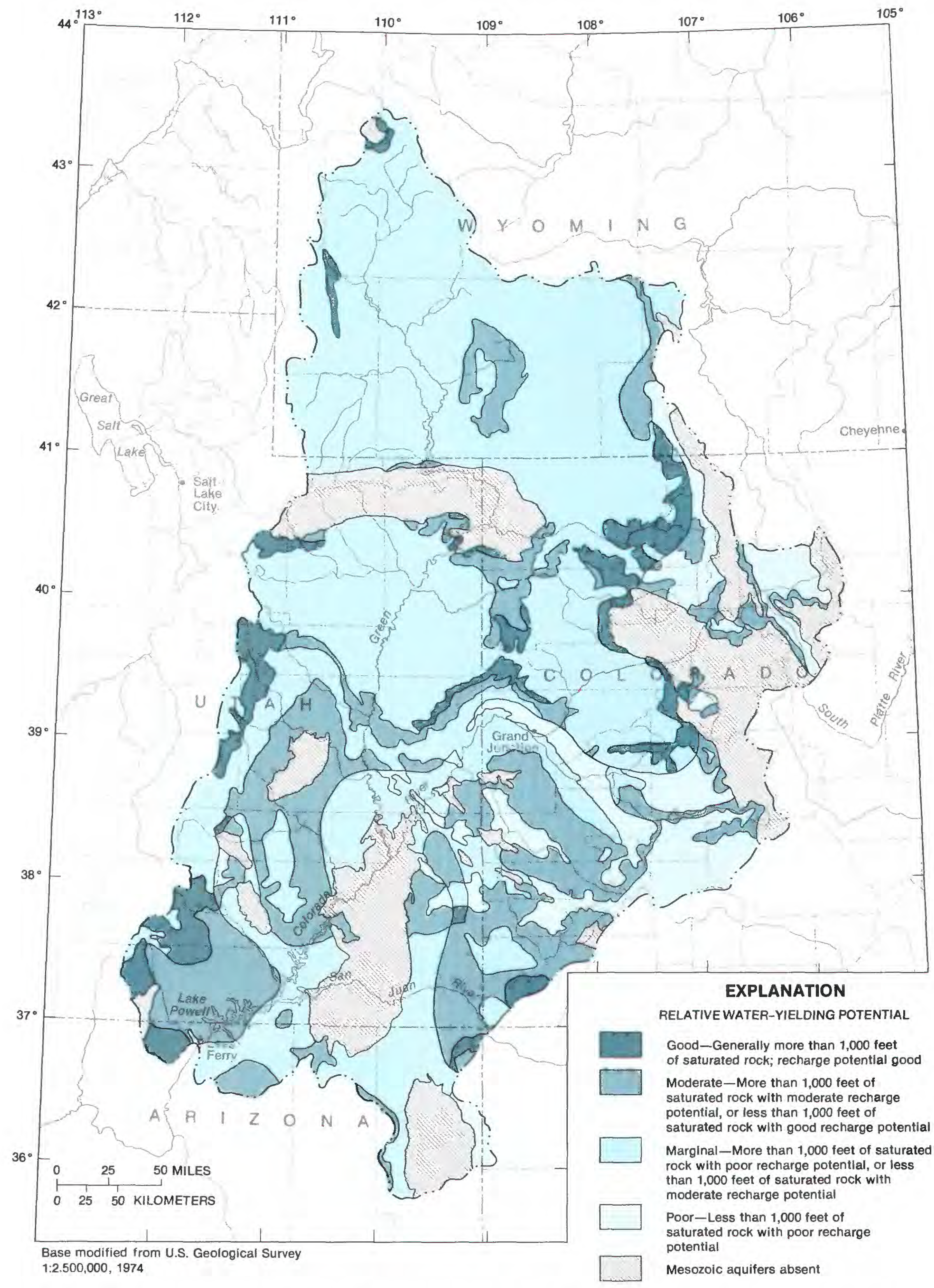


FIGURE 74.—Regional water resources in Mesozoic rocks as defined by saturated thickness and recharge potential.

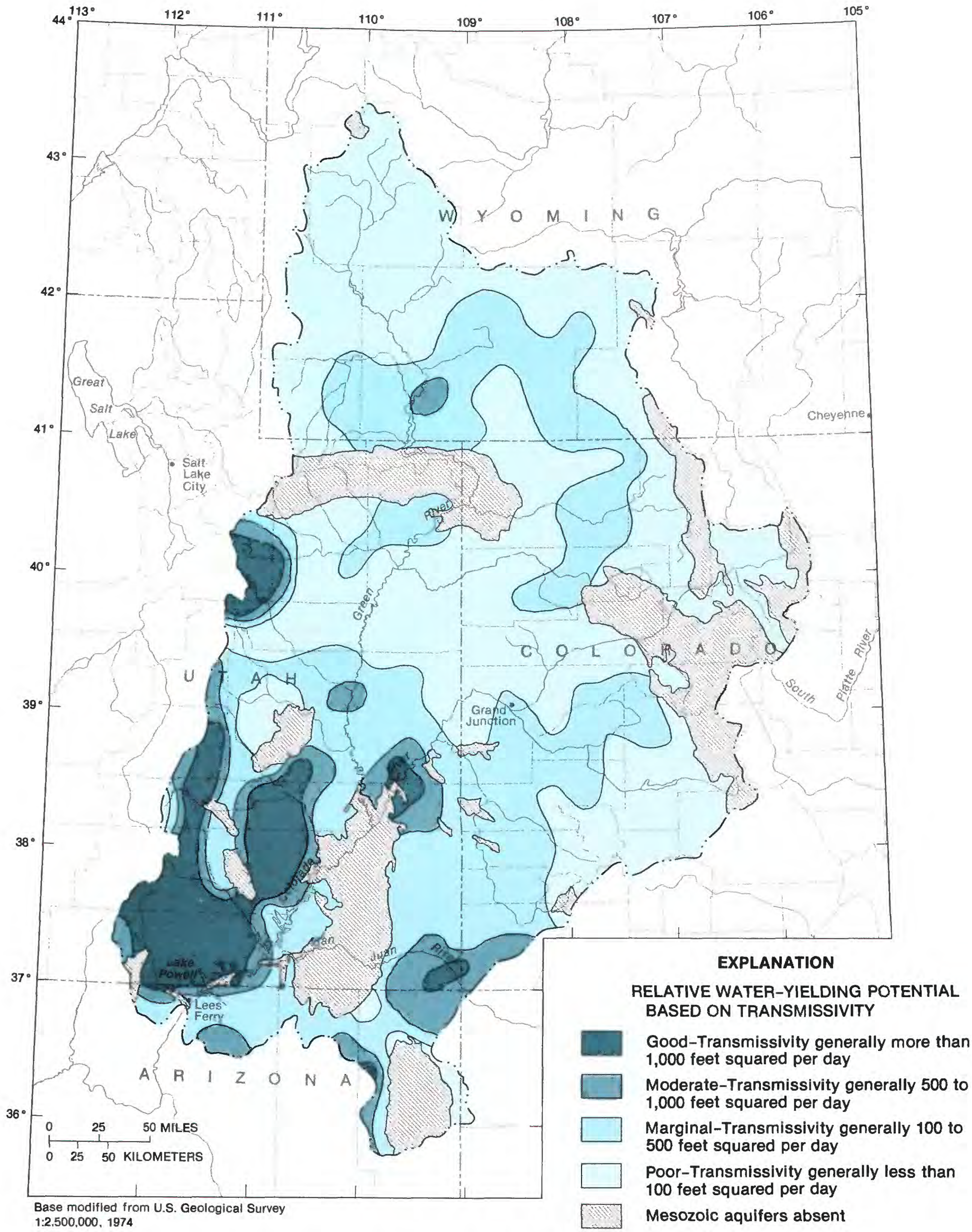


FIGURE 75.—Regional water resources in Mesozoic rocks as defined by estimated aggregate transmissivity.

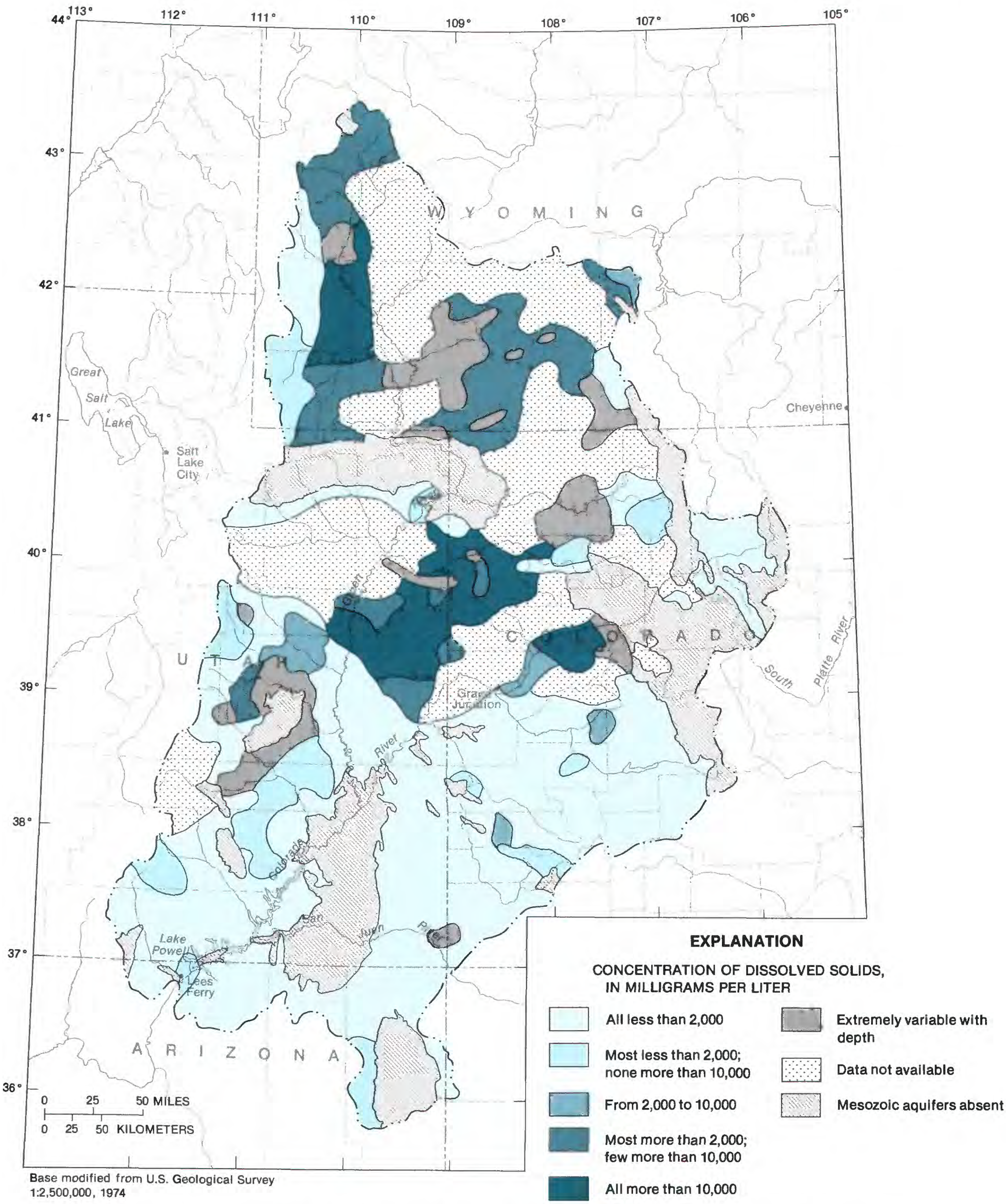


FIGURE 76.—Generalized quality of water in Mesozoic rocks based on the concentration of dissolved solids.

tural basins where Mesozoic rocks are deeply buried and dissolved-solids concentrations are typically large.

NEED FOR FUTURE INVESTIGATIONS

Joints and faults associated with major structural features can function as conduits or barriers to ground-water flow, as discussed previously. However, few studies have actually summarized the occurrence and interrelation of fractures in the Upper Colorado River Basin. Fracturing in confining units can increase leakage between aquifers, possibly affecting water quality, potentiometric surfaces, and other properties of the ground-water flow system. In addition, fractured confining beds can yield usable ground water locally. Further studies are needed to determine where fracturing can affect the hydrologic properties of aquifers and confining units.

Major unconformities mark the boundaries between many of the hydrologic units in this study. Little is known about the effects of these discontinuities, particularly where they separate two lithologically similar aquifers such as the Page and Navajo Sandstones. The hydrologic effects of such discontinuities within the aquifers as bedding planes, thin confining units, and vertical and lateral changes in lithofacies are largely unknown. Research and carefully planned onsite testing are needed to identify the effects of discontinuities on the hydrology of Mesozoic rocks.

Further geologic studies, including subsurface stratigraphic correlation, and lithofacies analyses related to the hydrology of the aquifers would be useful in developing a better understanding of the regional flow system. In addition, mineralogic and petrologic studies could be useful in understanding water quality, both locally and regionally.

Additional investigations of the hydrology of the Mesozoic ground-water system would improve definition of certain aspects of this flow system. The hydrology of the confining units is relatively unknown. Investigations are needed to define horizontal and vertical movement through shales and siltstones. The occurrence and movement of water in confining layers may affect the potential for developing ground water in specific locations in the Upper Colorado River Basin.

Igneous rocks are present in a small percentage of the study area, yet they are extremely important to the hydrology of the Mesozoic aquifer system. Igneous rocks are located, almost without exception, in the principal recharge areas. Investigations concerning unsaturated and saturated flow characteristics of these rocks are needed before the factors that affect the location and quantity of recharge to aquifers in Mesozoic rocks can be understood.

The relation between annual precipitation and aquifer recharge depends on numerous meteorologic, geologic, and hydrologic factors. Empirical formulas approximating this relation for unconsolidated sediments in the Basin and Range province have been developed and used with some success. No relation has been developed for fractured and unfractured consolidated rock in the Colorado Plateaus area. Formulation of this relation is one of the most important keys to evaluating the ground-water resources of the Upper Colorado River Basin.

The quality of the Nation's ground water will be of vital concern in the coming decades as the demand for this resource increases. More baseline information is needed to detect and evaluate changes in quality that may occur. Water in the aquifers in Mesozoic rocks in the Upper Colorado River Basin that has not yet been affected by mining, oil and gas exploration, irrigation practices, or waste disposal, needs to be monitored for salinity, heavy metal concentrations, radionuclides, and nutrient content to establish baseline information on natural concentrations. Aquifers where the state of equilibrium may become disturbed need to be sampled periodically to evaluate changes in quality that might take place. Site investigations to determine the origin and migration of chemical constituents where concentrations exceed established limits would help identify other areas where quality may become a problem.

SUMMARY AND CONCLUSIONS

The hydrologic characteristics of Mesozoic rocks of the Upper Colorado River Basin were analyzed as part of the Regional Aquifer-System Analysis (RASA) Program. In addition to classifying rocks into intervals of aquifers and confining units, the occurrence, movement, and quality of water in these units were quantitatively determined from existing data.

The Mesozoic rocks of the Upper Colorado River Basin were divided into 10 geohydrologic units—5 aquifers and 5 confining units—on the basis of general lithologic and hydrologic character. The confining units are predominantly siltstone, shale, claystone, and limestone. Locally, confining units may contain thick interbedded sandstones that are potential sources of ground water for local use. The Ferron Sandstone and Frontier Sandstone Members are notable examples of local aquifers within the Mancos confining unit. Similarly, the predominantly sandstone aquifers, almost without exception, contain interbedded deposits with confining characteristics. For example, 39 percent of the Salt Wash Member of the Morrison Formation (Morrison aquifer) is composed of siltstone, claystone, and mudstone interbedded with the sandstone.

Areal extent and thickness maps for each of the 10 geohydrologic units are shown on plates 2 and 3. It is apparent from these maps that the Mesozoic rocks are missing from the Uinta Uplift and Monument Uplift in Utah, the White River Uplift, Burns Basin, and Park Uplift in Colorado, and parts of the Sierra Madre Uplift, Rawlins Uplift, and Wind River Uplift in Wyoming. The relative extents and thicknesses of the various Mesozoic geohydrologic units are shown on the accompanying fence diagram (pl. 4). Although the fence diagram lacks geologic structure, it is meant to show the continuity of each geohydrologic unit and the relative thickness.

The thickness of the Mesozoic section and the percentage of that section that consists of aquifer material are shown in figure 77. North of Green River, Utah, and Grand Junction, Colo., where the thickness exceeds 5,000 ft, aquifer material composes 25 to 50 percent of the total thickness. However, 50 to 75 percent of the more than 5,000 ft of thickness consists of confining units that must be penetrated to reach successively deeper aquifers. In contrast, Mesozoic rocks in the area south of Green River and Grand Junction are less than 5,000 ft thick, and in a large part of the area 50 to 75 percent of this thickness is aquifer material, indicating that only 25 to 50 percent of less than 5,000 ft must be penetrated to tap the aquifers. Thus, even though the total thickness of aquifers is smaller in the southern area than in the northern area, the southern area appears to offer more opportunity for obtaining water from the aquifer system in Mesozoic rocks.

General measures of rock fabric indicate that most Mesozoic sandstones consist of very fine to fine sand that is moderately sorted. Particles that are silt size or smaller compose less than 10 percent of the rock volume. Carbonate content generally ranges from less than 1 percent in the Navajo-Nugget aquifer to more than 5 percent in the Entrada-Preuss aquifer.

The most easily determined properties that describe the hydrologic aspects of rock fabric are porosity and gas permeability. The Navajo-Nugget and parts of the Entrada-Preuss aquifers have a larger porosity than the Morrison, Dakota, and Mesaverde aquifers. Gas permeability values for samples from the Navajo-Nugget, Entrada-Preuss, and Morrison aquifers are typically larger than 250 millidarcies (mD)—one to three orders of magnitude larger than those for adjacent confining layers. Values for the Dakota and Mesaverde aquifers are less than 100 mD and 11 mD, respectively, yet these aquifers are much more permeable than the adjacent shale and siltstone confining layers.

Fluid character affects water movement most significantly in aquifers that contain brine and that are buried thousands of feet below land surface. Increased density due to large concentrations of dissolved solids and decreased viscosity due to high temperatures at depth result in hydraulic-conductivity values larger than those at shallower depths for the same rock types in an aquifer.

Ranges of hydraulic-conductivity values, derived from laboratory analyses, aquifer tests, drill-stem tests, and specific-capacity tests, differ because, in general, the test environment and the size of sample tested differ. Values from laboratory analyses represent the smallest tested samples. Values derived from aquifer tests and specific-capacity tests represent zones within an aquifer that are close to land surface where open fractures are most likely. Values from drill-stem tests generally represent zones within a deeply buried aquifer that have the greatest potential for yielding oil or gas and are often affected by invasion of mud into the hole wall.

The most common values of hydraulic conductivity range from 0.1 to 10.0 ft/d in the Navajo-Nugget, Entrada-Preuss, and Morrison aquifers and from 0.001 to 1.0 ft/d in the Dakota and Mesaverde aquifers. Values for confining units are generally one order of magnitude smaller than for the adjacent aquifers. A regional representation is best shown as an aggregate of all values because each group of data depicts an important environment within the regional aquifer system. Transmissivity values that describe ground-water movement through the regional system are small, but locally, where fractures are numerous and open, transmissivity values are undoubtedly much larger.

The Mesozoic rocks of the Upper Colorado River Basin contain about 10 million acre-ft of water. This water moves slowly from areas of recharge to areas of discharge. Principal recharge areas are located where annual winter precipitation is more than about 8 in and where Mesozoic rocks are exposed or are covered by only a thin layer of permeable younger rocks. Secondary recharge areas are located in topographically low areas where surface water directly infiltrates into exposed aquifers in Mesozoic rocks. Recharge originating outside the study area or attributable to vertical movement from an overlying or underlying aquifer is small, but probably important locally. The estimated annual recharge to all aquifers in Mesozoic rocks in the Upper Colorado River Basin is about 1 million acre-ft. Recoverable storage of potable water in the upper 2,000 ft of rock is about 530 million acre-ft.

Annual discharge from these aquifers in Mesozoic rocks is approximately the same as recharge to them. Principal discharge is into the river systems where erosion has breached the aquifer rocks, by direct and indirect evapotranspiration, and to springs and seeps. Discharge through vertical movement to overlying and underlying formations, and by lateral flow across the study-area boundary, is small.

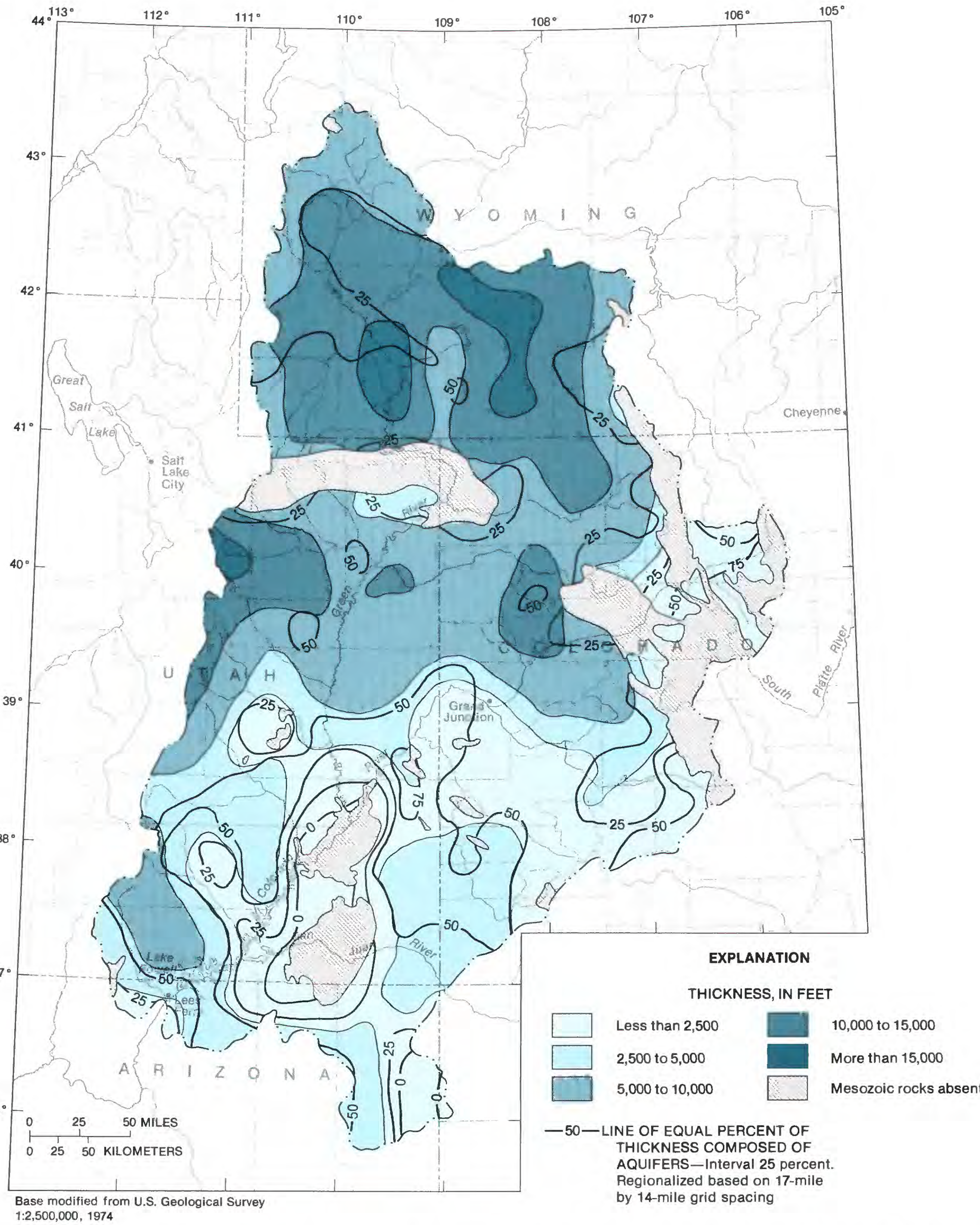


FIGURE 77.—Thickness of Mesozoic rocks and the percentage of that thickness composed of aquifers (generalized on the basis of 17- by 14-mile grid spacing).

Ground-water movement within the Mesozoic rocks takes place horizontally and vertically. Hydraulic gradients in the five aquifers indicate that ground water flows laterally from areas of high altitude to rivers that drain the area. Water levels in the aquifers that are at or near land surface indicate that ground water moves toward the smaller local surface drainages. Water levels in buried aquifers indicate a regional flow toward the main rivers—the Colorado, Green, and San Juan Rivers. Vertical movement of ground water occurs where vertical differences in hydraulic gradients are large. Exchange of ground water between aquifers is most likely where confining units are thin, coarse grained, or absent.

The occurrence and movement of ground water in the Upper Colorado River Basin is also affected by folds, fractures, and faults, which in most cases are associated with the structural basins, uplifts, and platforms that compose the study area. Folds in the study area have little effect on the direction of regional flow; however, eroded folds exposing Mesozoic rocks are recharge and discharge areas for the aquifer system. Open fractures and faults provide conduits which transmit water more readily than the surrounding rock. Conversely, closing of fractures with depth or recementation of faults can cause these features to function as barriers to ground-water flow.

The quality of water in aquifers in Mesozoic rocks is extremely variable, ranging from fresh to briny. Water in or near recharge areas is generally a calcium bicarbonate type. Water far from recharge areas, such as in deep structural basins, is generally a sodium chloride type. Minor constituents, such as trace metals, are present in concentrations smaller than the maximum contaminant level for drinking water established by the U.S. Environmental Protection Agency. Average concentrations of manganese and iron, constituents that cause undesirable staining and taste when the water is used for domestic purposes, slightly exceed Federal standards. These average values for most constituents are biased by exceedingly large concentrations from the brine samples that were included in calculations of the average values.

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