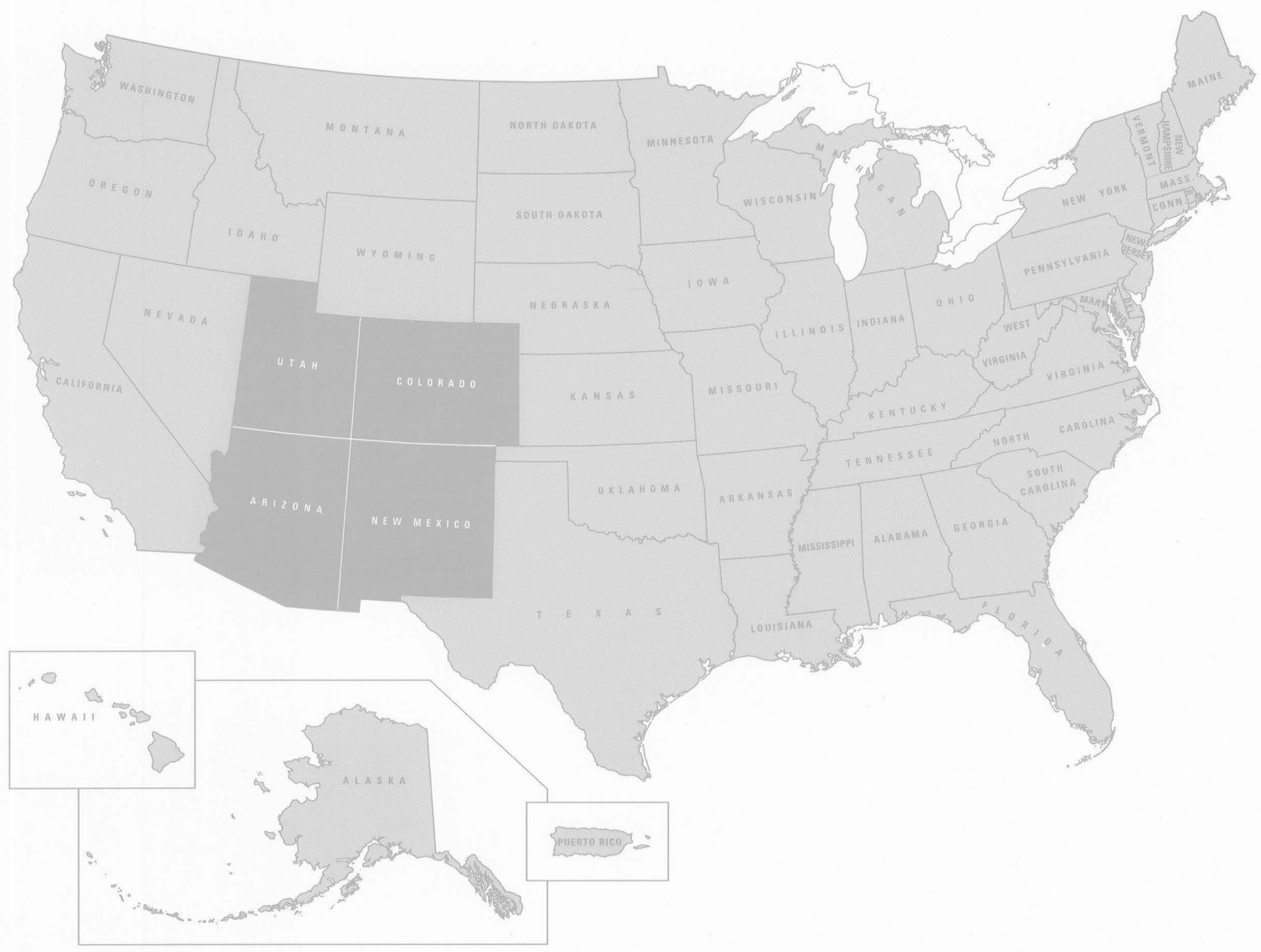
GROUND WATER ATLAS OF THE UNITED STATES

SEGMENT 2

Arizona
Colorado
New Mexico
Utah



HYDROLOGIC INVESTIGATIONS ATLAS 730–C
U.S. Geological Survey

Reston, Virginia
1995

GROUND WATER ATLAS OF THE UNITED STATES
Hydrologic Investigations Atlas 730–C

FOREWORD

The Ground Water Atlas of the United States presents a comprehensive summary of the Nation's ground-water resources, and is a basic reference for the location, geography, geology, and hydrologic characteristics of the major aquifers in the Nation. The information was collected by the U.S. Geological Survey and other agencies during the course of many years of study. Results of the U.S. Geological Survey's Regional Aquifer-System Analysis Program, a systematic study of the Nation's major aquifers, were used as a major, but not exclusive, source of information for compilation of the Atlas.

The Atlas, which is designed in a graphical format that is supported by descriptive discussions, includes 13 chapters, each representing regional areas that collectively cover the 50 States and Puerto Rico. Each chapter of the Atlas presents and describes hydrogeologic and hydrologic conditions for the major aquifers in each regional area. The scale of the Atlas does not allow portrayal of minor features of the geology or hydrology of each aquifer presented, nor does it include discussion of minor aquifers. Those readers that seek detailed, local information for the aquifers will find extensive lists of references at the end of each chapter.

An introductory chapter presents an overview of ground-water conditions Nationwide and discusses the effects of human activities on water resources, including saltwater encroachment and land subsidence.

Gordon P. Eaton

U.S. DEPARTMENT OF THE INTERIOR
BRUCE BABBITT, *Secretary*

U.S. GEOLOGICAL SURVEY
Gordon P. Eaton, *Director*

CONVERSION FACTORS

For readers who prefer to use the International System (SI) units, rather than the inch-pound terms used in this report, the following conversion factors may be used:

Multiply inch-pound units	*By*	*To obtain metric units*
	Length	
inch (in)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
square mile (mi^2)	2.590	square kilometer (km^2)
	Flow	
gallon per minute (gal/min)	0.06309	liter per second (L/s)
million gallons per day (Mgal/d)	0.04381	cubic meter per second (m^3/s)
billion gallons per day (Bgal/d)	3.785	million cubic meters per day (Mm^3/d)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
acre-foot per year	0.00003909	cubic meter per second (m^3/s)
acre-foot	1,233	cubic meter (m^3)
	Transmissivity	
foot squared per day (ft^2/d)	0.0929	meter squared per day (m^2/d)
	Temperature	
degree Fahrenheit (°F)	5/9(°F–32)=°C	degree Celsius (°C)

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.

ATLAS ORGANIZATION

The Ground Water Atlas of the United States is divided into 14 chapters. Chapter A presents introductory material and nationwide summaries; chapters B through M describe all principal aquifers in a multistate segment of the conterminous United States; and chapter N describes all principal aquifers in Alaska, Hawaii, and Puerto Rico.

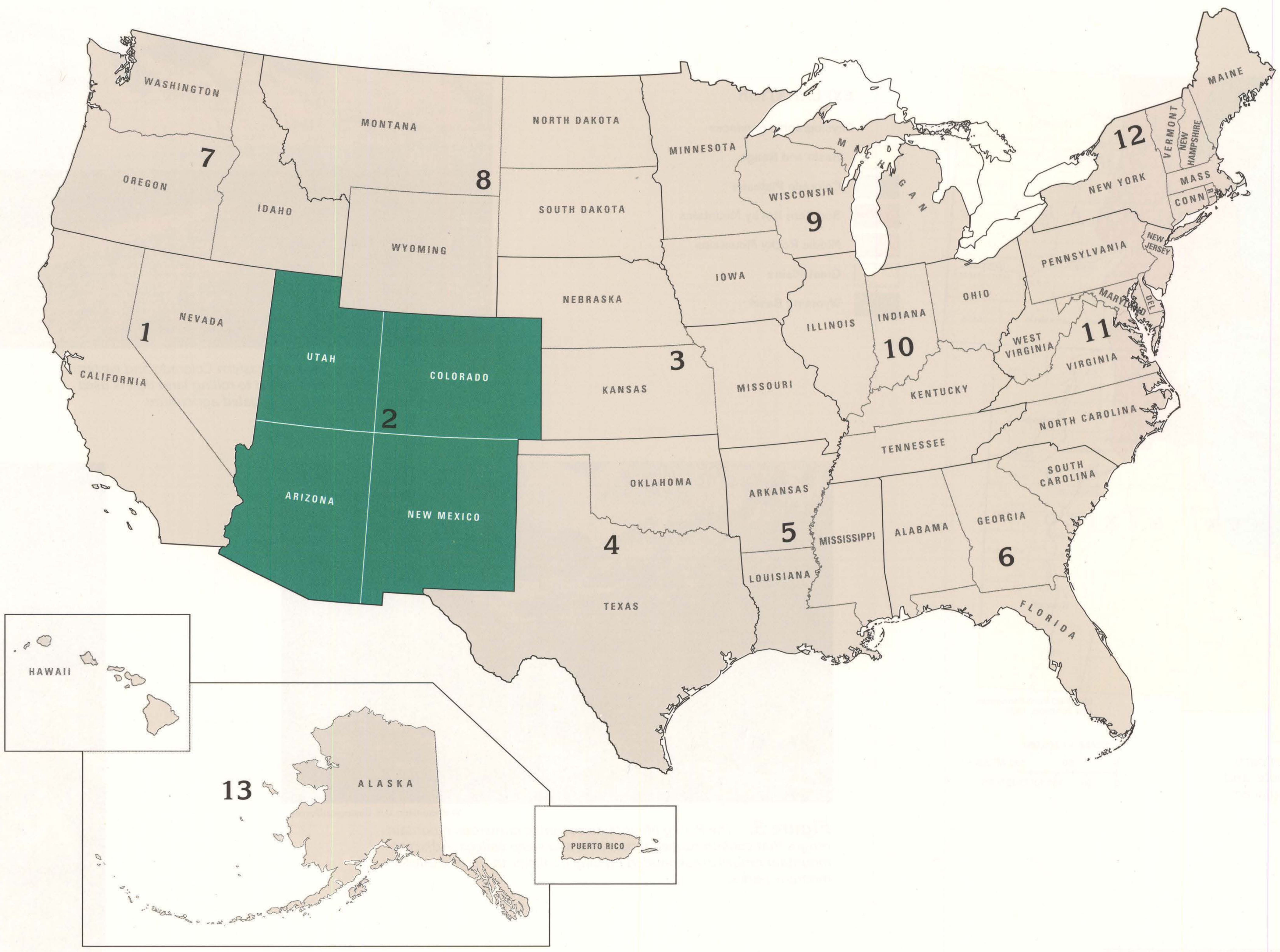
Segment Number	*Chapter content*	*Hydrologic Atlas Chapter*
—	Introductory material and nationwide summaries	730–A
1	California, Nevada	730–B
2	Arizona, Colorado, New Mexico, Utah	730–C
3	Kansas, Missouri, Nebraska	730–D
4	Oklahoma, Texas	730–E
5	Arkansas, Louisiana, Mississippi	730–F
6	Alabama, Florida, Georgia, South Carolina	730–G
7	Idaho, Oregon, Washington	730–H
8	Montana, North Dakota, South Dakota, Wyoming	730–I
9	Iowa, Michigan, Minnesota, Wisconsin	730–J
10	Illinois, Indiana, Kentucky, Ohio, Tennessee	730–K
11	Delaware, Maryland, New Jersey, North Carolina, Pennsylvania, Virginia, West Virginia	730–L
12	Connecticut, Maine, Massachusetts, New Hampshire, New York, Rhode Island, Vermont	730–M
13	Alaska, Hawaii, Puerto Rico	730–N

Ground Water Atlas of the United States

Segment 2

Arizona, Colorado, New Mexico, Utah

By S.G. Robson and E.R. Banta



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Cartographic design and production by Loretta J. Ulibarri, Derald L. Dunagan, and Wendy J. Danchuk

Regional summary

INTRODUCTION

This chapter of the Ground Water Atlas of the United States describes the aquifers in Arizona, Colorado, New Mexico, and Utah. These four States, which comprise Segment 2 of this Atlas, are located in the Southwestern United States and extend from the rolling grasslands of the Great Plains on the east across the Rocky Mountains and Continental Divide to the desert basins of the Southwest. The 425,000-square-mile area ranges in altitude from about 14,400 feet above sea level in the Rocky Mountains of Colorado to about 100 feet near the lower Colorado River in southwestern Arizona. All the ground water in Segment 2 ultimately is derived from infiltration of precipitation, which varies considerably with the altitude and topography of the area.

The Great Plains Physiographic Province of the Central United States extends into eastern Colorado and New Mexico (**fig. 1**), where flat to rolling prairie (**fig. 2**) with scattered hills and bluffs gradually rises westward to 5,000 to 7,000 feet above sea level and abruptly gives way to the frontal ranges of the Rocky Mountains in the Southern Rocky Mountain and Basin and Range Physiographic Provinces. West of the frontal ranges in Colorado and northern New Mexico are additional and higher mountain ranges generally oriented north-south but with many spurs and extensions oriented in other directions. The many ranges of the Rocky Mountains are separated by valleys and high mountain parks (**fig. 3**). Colorado contains about three-fourths of the Nation's land area above 10,000 feet and has 53 mountain peaks higher than 14,000 feet. Most of these high peaks are located near the Continental Divide (**fig. 1**), which extends approximately north-south through central Colorado and western New Mexico. The altitude of the divide decreases in southern New Mexico to less than 4,500 feet in a few areas.

Farther westward, the mountains are less prevalent and are interspersed with broad structural basins. These basins and the broad valleys of the middle Colorado River and its tributaries form the irregular intermontane topography of the Colorado Plateaus Physiographic Province (**fig. 4**). Plateaus and high mesas are formed where the surface has been dissected by rugged canyons carved by the Colorado River and its tributaries (**fig. 5**). The largest of these canyons—the Grand Canyon—extends about 220 miles southwestward from the mouth of the Little Colorado River in Arizona and ranges from 4 to 18 miles in width and from 2,700 to 5,700 feet in depth below the rim.

Small mountain ranges and intervening broad desert valleys of the Basin and Range Physiographic Province are prevalent to the west and south of the Colorado Plateaus in western Utah, southern Arizona, and southern New Mexico (**fig. 1**). These mountain ranges generally protrude 3,000 to 6,000 feet above the surrounding valley floor (**fig. 6**) and commonly extend from 20 to 50 miles in a north or northwesterly direction.

Small parts of the Middle Rocky Mountains and Wyoming Basin Physiographic Provinces extend into northwestern Colorado and northeastern Utah (**fig. 1**). The topography, geology, and hydrology of the two areas are described in Chapter I of this Atlas.

Four of the Nation's major river systems have headwaters in the mountainous areas of Segment 2. The South Platte River of the Missouri River system drains the eastern slope of northern Colorado; the Arkansas River and its tributary, the Canadian River, drain southeastern Colorado and northeastern New Mexico; the Rio Grande and its tributary, the Pecos River, drain south-central Colorado and central New Mexico; and the Colorado River and its tributaries drain Arizona, eastern Utah, northwestern New Mexico, and western Colorado (**fig. 1**). Western Utah is drained by numerous streams that terminate in local desert basins, the Great Salt Lake, or other local lakes and reservoirs. Because the Great Salt Lake lies in the Great Basin, which is the largest closed basin in North America, it has no outlet to the sea. The salinity of the lake water is about 20 percent or about 6 times the salinity of seawater.

Most of Segment 2 is sparsely populated. The average population density of counties is less than 8 persons per square mile in about 65 percent of the four-State area (**fig. 7**). Population densities range from less than 0.5 person per square mile in a few rural counties to more than 4,000 persons per square mile in populous urban areas. The 1990 population of the four States was about 10 million; almost 70 percent of this population was in Arizona and Colorado. Most land in Segment 2 is undeveloped forest grassland, or desert shrubland, much of which is used for livestock grazing. Land used for production of commercial crops primarily is in eastern Colorado and eastern New Mexico.

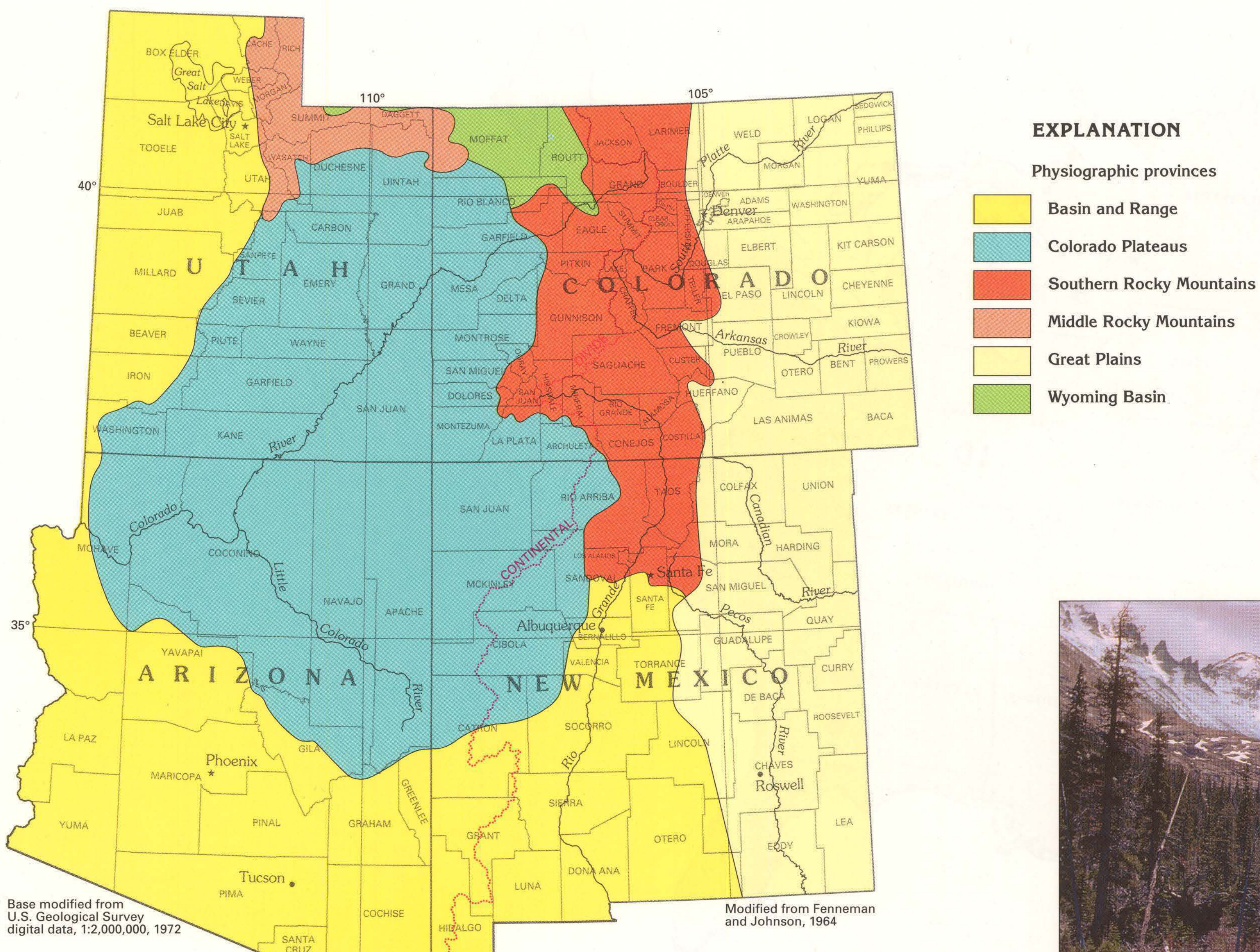

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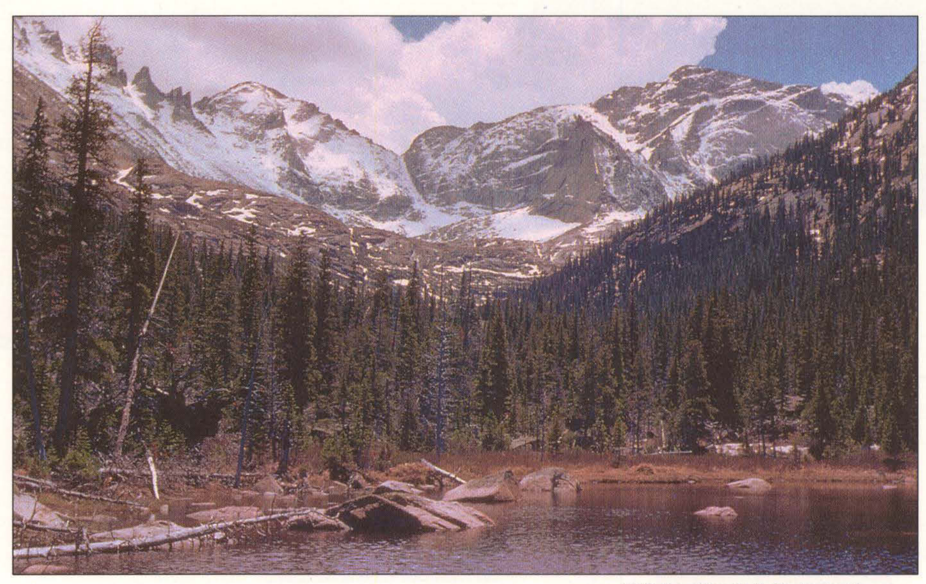
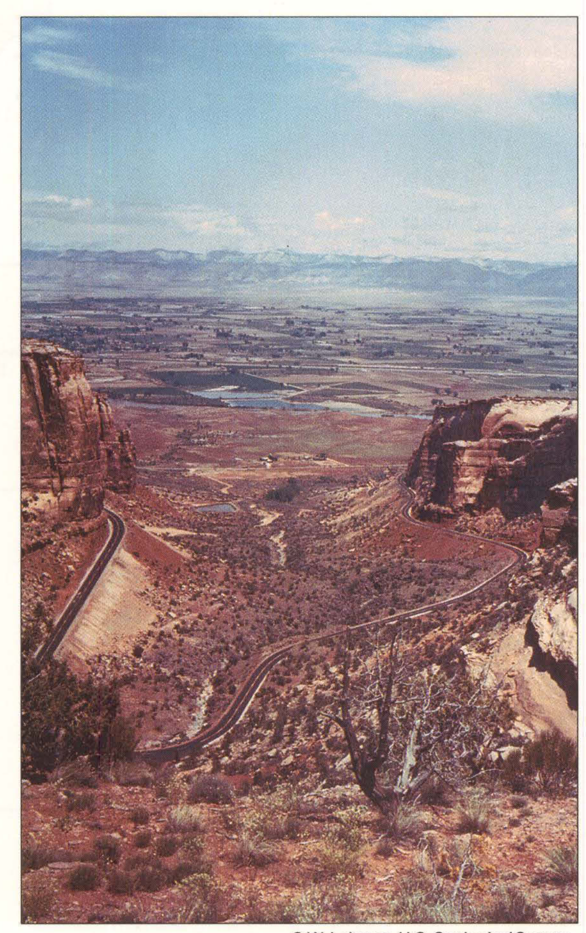
Figure 1. *The four-State area of Segment 2 consists of parts of six physiographic provinces. The geology, terrain, climate, and vegetation in each province can be distinct from that of adjacent provinces.*

Figure 2. *The Great Plains of eastern Colorado and eastern New Mexico generally consist of flat to rolling land that is used for grazing, dryland farming, or irrigated agriculture.*



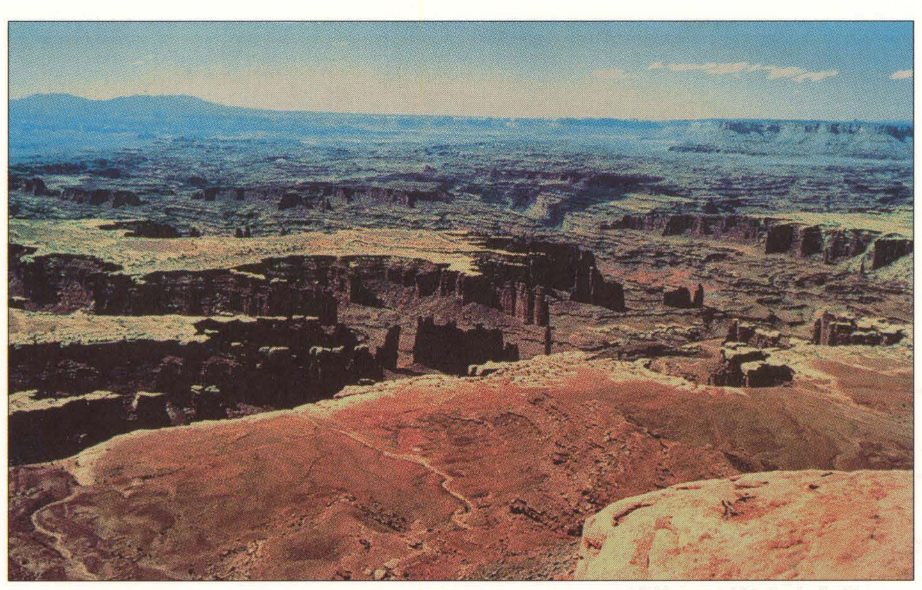
W.B. Hamilton, U.S. Geological Survey

Figure 3. *The Rocky Mountains comprise numerous mountain ranges that contain rugged mountains and steep valleys. Adjacent mountain ranges are separated by larger valleys and broad intermontane parks.*



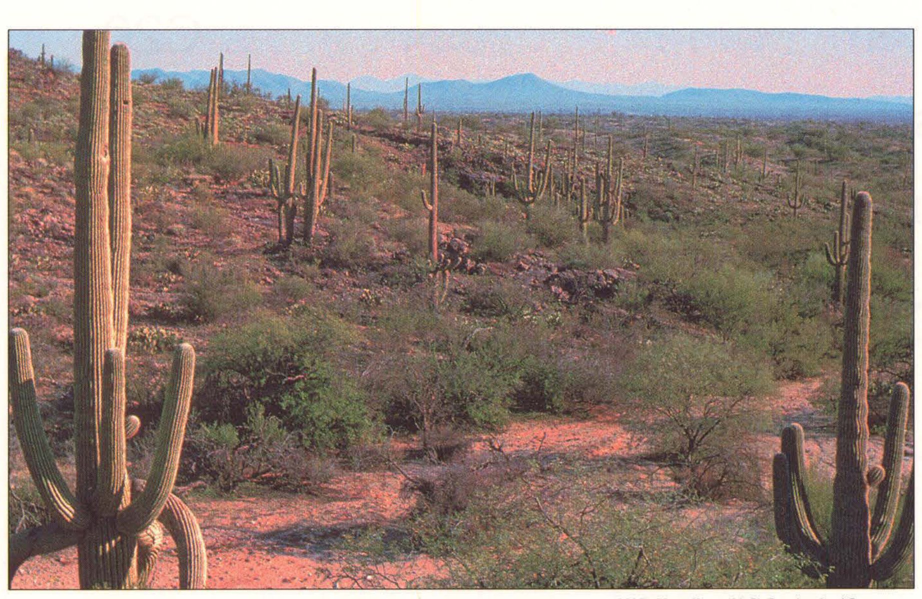
S.W. Lohman, U.S. Geological Survey

Figure 4. *The Colorado River in western Colorado has a broad, gently sloping valley that is bordered by mountains and by canyon walls that become more rugged farther down the Colorado River.*



S.W. Lohman, U.S. Geological Survey

Figure 5. *Rugged canyons such as these in southern Utah extend for about 200 miles along the Colorado River and its tributaries upstream from the Grand Canyon.*



W.B. Hamilton, U.S. Geological Survey

Figure 6. *The desert valleys of southern Arizona generally are bordered by small mountain ranges. The valleys slope gently to dry stream channels or dry lakebeds.*

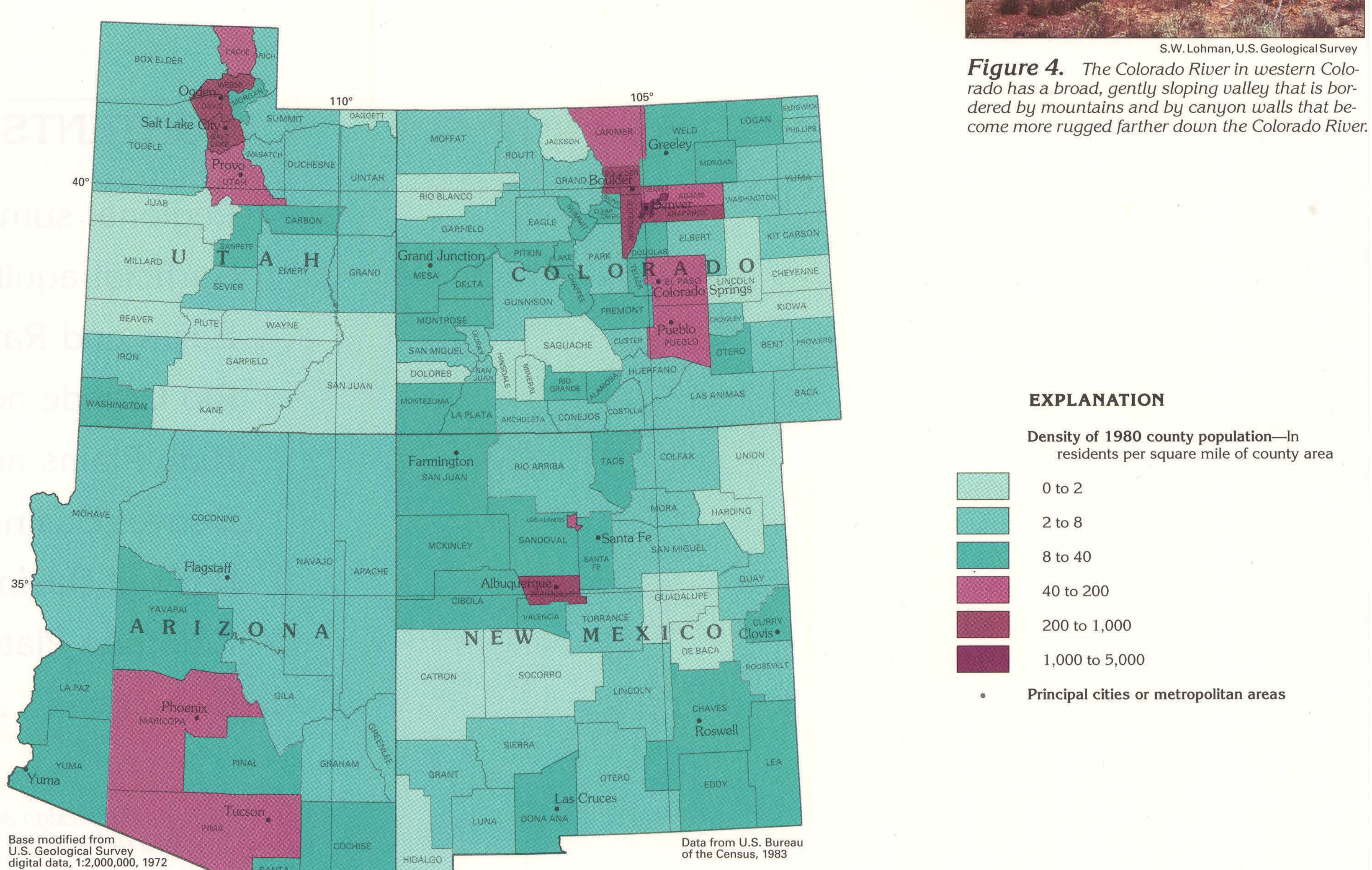


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Figure 7. *The population density of much of Segment 2 is less than eight persons per square mile. Counties containing large metropolitan areas such as Albuquerque, Denver, Phoenix, Salt Lake City, and Tucson are more densely populated.*

CLIMATIC CONDITIONS

The varied topographic settings of Segment 2 have differing climatic conditions primarily because of differences in altitude. Temperature generally decreases and precipitation generally increases with increased altitude. The Great Plains, Colorado Plateaus, and Basin and Range Physiographic Provinces generally have abundant sunshine, moderate to high wind, low relative humidity, a large daily range in temperature, and little precipitation. In the Great Plains, about 60 to 80 percent of the annual precipitation falls during April to September, primarily from thunderstorms. Winter generally is the driest season. Average annual precipitation ranges from about 12 to 16 inches on the plains of New Mexico and Colorado (**fig. 8**). The slightly larger rates of precipitation near the eastern boundaries of Colorado and New Mexico are part of a general eastward increase in precipitation across the central part of the Nation.

The rugged and varied topography of the Colorado Plateaus causes diverse climatic conditions. Alpine climatic conditions may occur in the higher altitudes of the various mountain ranges in the area. Less precipitation and higher temperatures prevail at lower altitudes, where semiarid, desert-like conditions are present in some valleys and basins. Average annual precipitation ranges from about 8 to 16 inches in the plateau areas (4,000–9,000 feet altitude) to more than 30 inches in the high mountain ranges (**fig. 8**). Most of the precipitation occurs in winter and spring and comes from eastward-tracking Pacific storm systems. Summer rainfall occurs almost entirely during brief, but often intense, thunderstorms that produce 20 to 40 percent of the annual precipitation. Winter storms produce either snow or rain in the southern valleys. In some mountainous areas, 10 feet or more of snowpack may accumulate.

Arid to semiarid climatic conditions exist in most of the Basin and Range area. Average annual precipitation ranges from less than 4 inches in southwestern Arizona to about 16 inches in western Utah. Rainfall occurs from widespread winter storms and from sporadic, local summer thunderstorms; many months may elapse between periods of measurable rainfall at a given location.

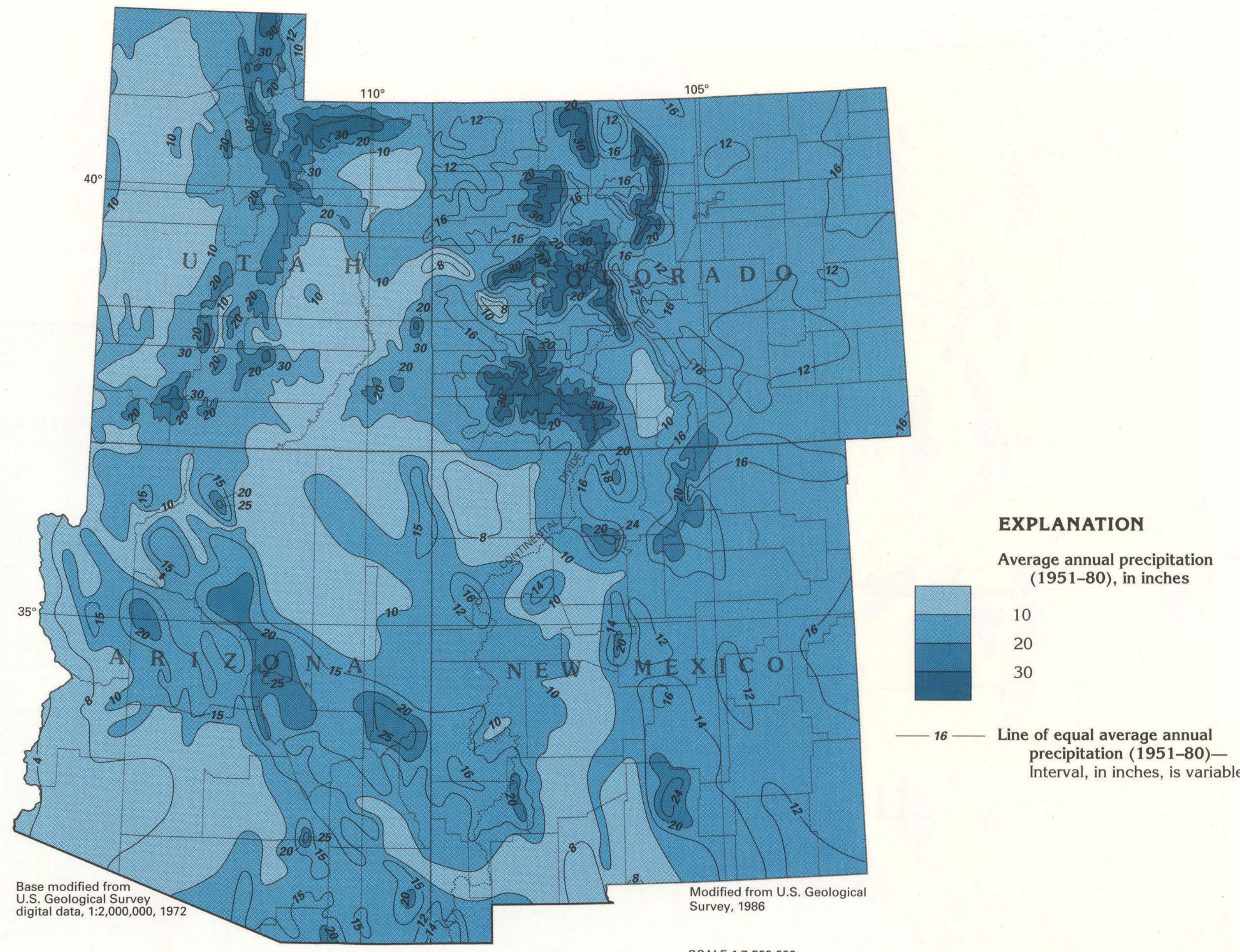


Figure 8. The average annual precipitation in Segment 2 ranges from less than 4 inches in the southwestern deserts of Arizona to more than 30 inches in parts of the Rocky Mountains.

The climate of the Rocky Mountains is greatly affected by differences in altitude and, to a lesser extent, by the orientation of the mountain ranges and valleys with respect to the general eastward movement of storm systems that originate in the Pacific Ocean. Rain-shadow effects tend to decrease precipitation on leeward slopes of mountain ranges and to increase precipitation on windward slopes. The eastern slopes of the Continental Divide generally receive less precipitation than do the western slopes because of this rain-shadow effect.

Precipitation in the Rocky Mountains is more evenly distributed throughout the year than that in the other physiographic provinces of Segment 2. Most precipitation occurs as snowfall from the large Pacific storm systems in the winter months; June is one of the driest months. Average annual snowfall ranges from about 5 feet to more than 35 feet. By August, most of the mountain snowpack has melted. Perennial snow fields occur in only a few small, sheltered areas at high altitude.

Some of the precipitation that falls on the land surface in Segment 2 flows directly into streams and rivers as direct runoff, and some precipitation infiltrates the soil and underlying aquifers and moves laterally to discharge into streams and rivers as baseflow. Direct runoff and baseflow are combined and mapped as average annual runoff in **figure 9**. In Segment 2, average annual runoff ranges from less than 0.2 inch to about 1 inch in most of the nonmountainous parts of the area. The small quantity of precipitation in this area, combined with relatively flat topography, porous soils, and general aridity causes most precipitation to be retained as soil moisture, which later evaporates or is transpired by plants rather than forming runoff. In the mountainous areas, the steeper topography, thin to nonexistent soils, cooler temperatures, and larger quantities of precipitation cause greater runoff, which can exceed 30 inches per year in some areas.

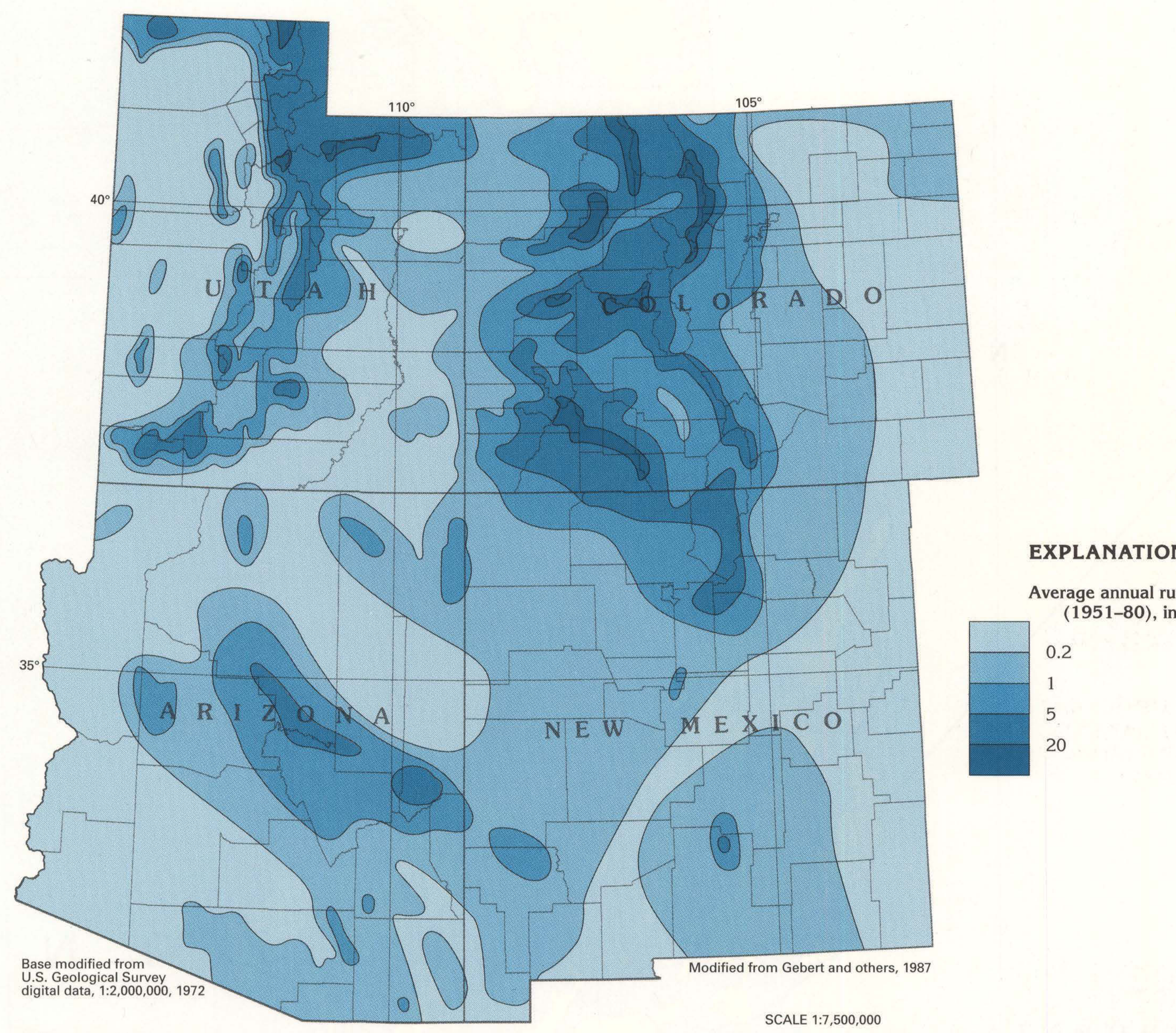


Figure 9. The average annual runoff in streams and rivers in the area varies greatly. The largest runoff occurs in mountainous areas; lesser runoff occurs in the eastern plains and desert basins.

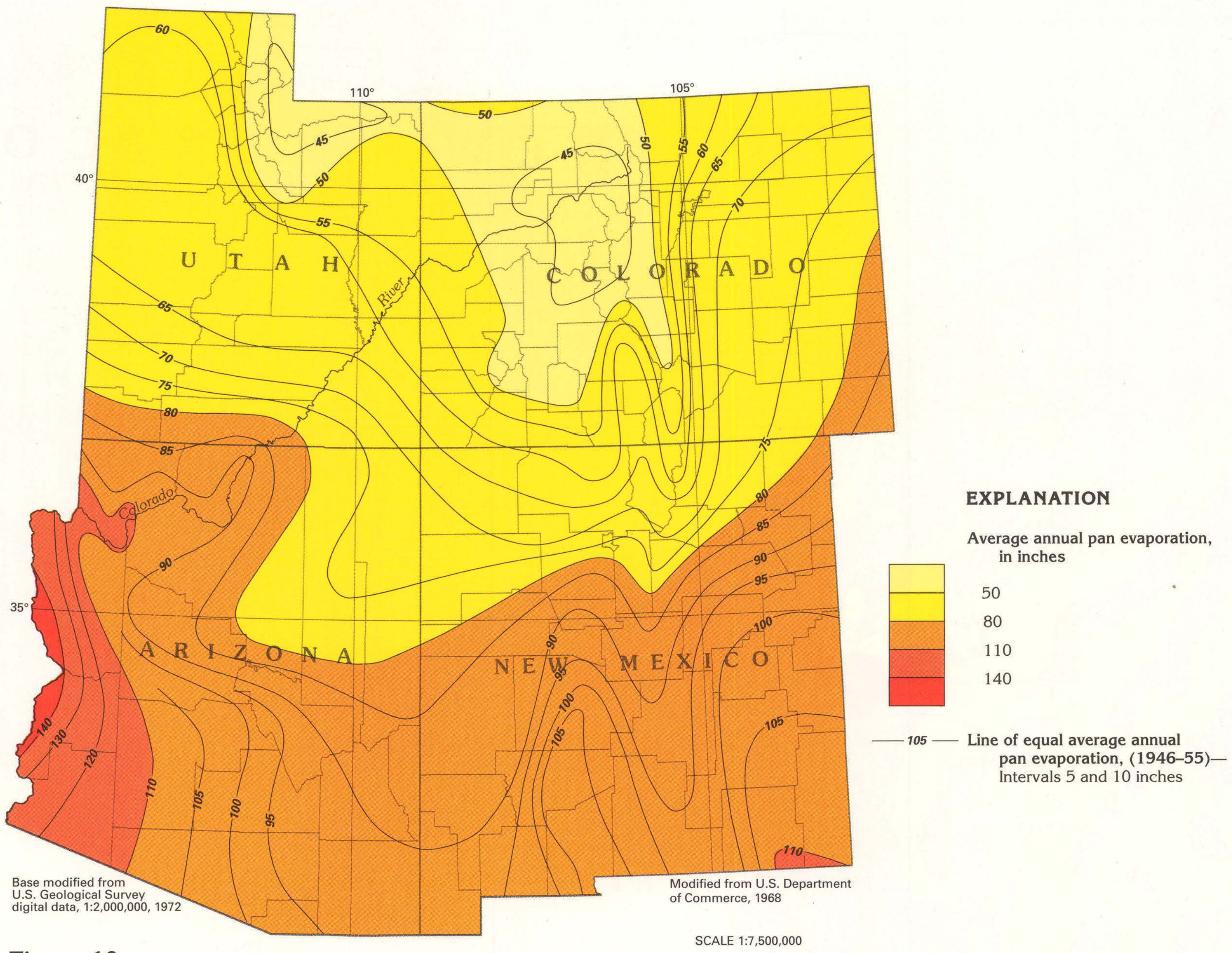


Figure 10. Average annual evaporation is as much as 140 inches in western Arizona and is in excess of annual precipitation in almost all of Segment 2.

The generally clear skies, low relative humidity, wind, and moderate to high temperatures in the nonmountainous parts of Segment 2 cause large rates of evaporation. Much of the precipitation returns to the atmosphere through evaporation of surface water, transpiration by plants, or sublimation of snow and ice. The annual rate of potential evaporation, as measured by standard evaporation pans, ranges from about 45 inches in the mountainous areas of northern Utah and north-central Colorado to about 140 inches in southwestern Arizona (**fig. 10**). Potential annual evaporation generally exceeds average annual precipitation throughout Segment 2 and is as much as 35 times the annual precipitation in a small area along the Colorado River in southwestern Arizona. Evaporation greatly in excess of precipitation removes most surface water and soil moisture before the water can percolate below the root zone of plants and recharge an underlying aquifer.

Recharge from precipitation generally ranges from zero to a few inches per year in Segment 2 but is extremely variable because it is affected by such factors as aquifer depth, thickness and permeability of soil and confining layers, land-surface slope and aspect, precipitation intensity and duration, and the temperature, wind, and relative humidity. In the parts of Segment 2 where geologic conditions do not preclude recharge, recharge from precipitation might only occur during the slow melting of a winter snowpack or during an extended period of winter rainfall when excess soil moisture is maintained.

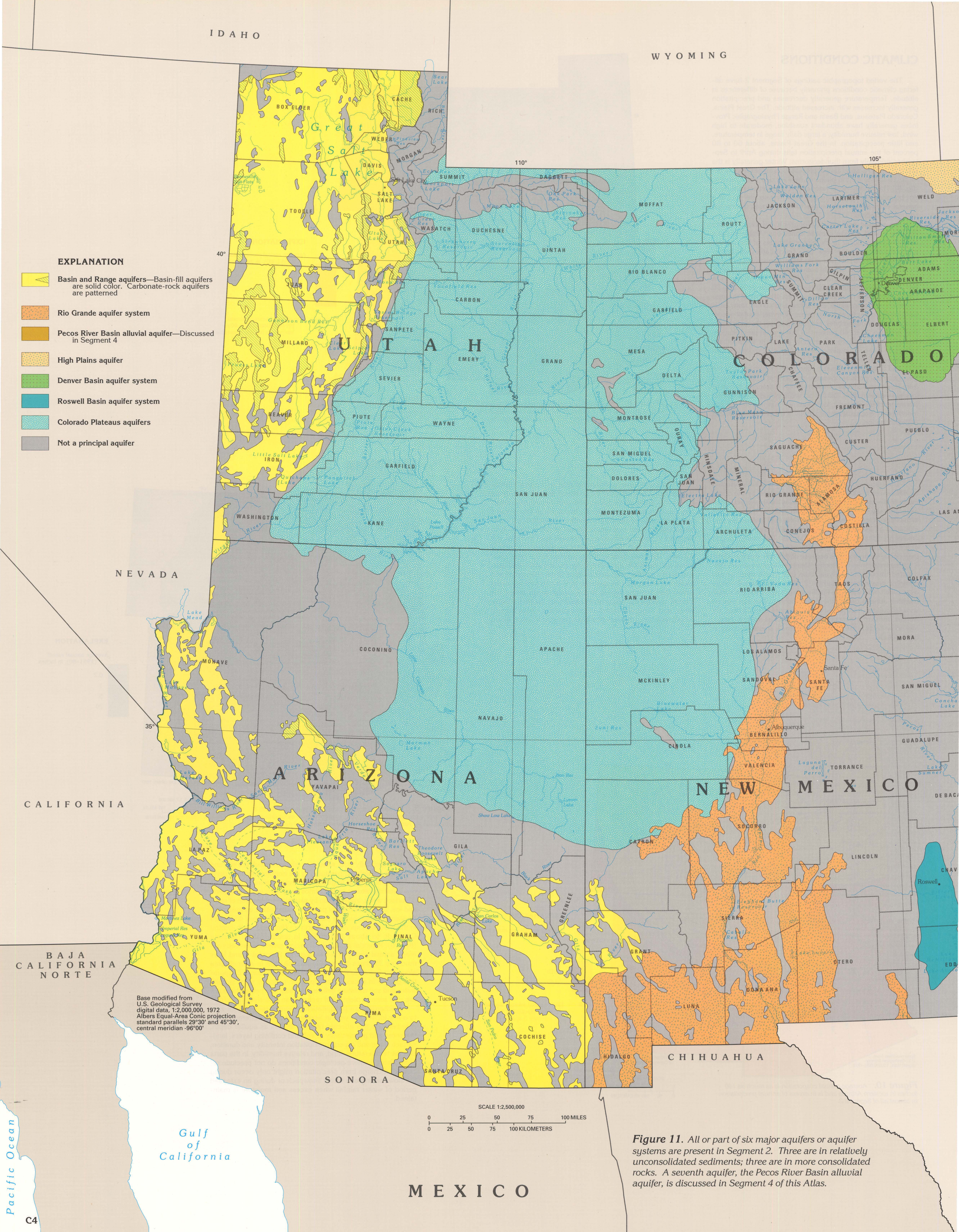
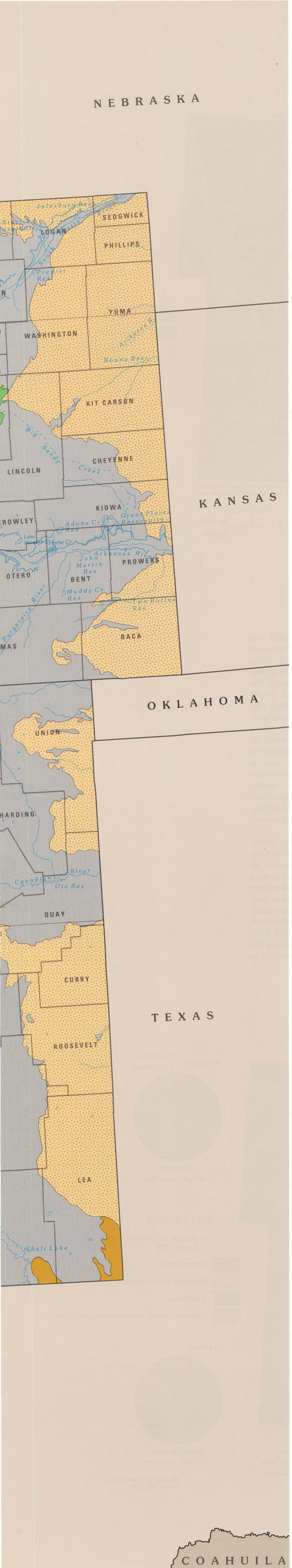


Figure 11. *All or part of six major aquifers or aquifer systems are present in Segment 2. Three are in relatively unconsolidated sediments; three are in more consolidated rocks. A seventh aquifer, the Pecos River Basin alluvial aquifer, is discussed in Segment 4 of this Atlas.*



MAJOR AQUIFERS AND AQUIFER SYSTEMS

All or part of six major aquifers or aquifer systems are present in Segment 2 (**fig. 11**). Four are in relatively unconsolidated sediments of Tertiary or Quaternary age; three are in more consolidated rocks of Tertiary, Cretaceous, or older age. A small part of a seventh aquifer, the Pecos River Basin alluvial aquifer, is present in extreme southeastern New Mexico, and extends into Texas. This aquifer is discussed in Chapter E of this Atlas, which pertains to Texas and Oklahoma.

The three aquifers or aquifer systems in unconsolidated sediments are (1) the Basin and Range aquifers in southern Arizona and western Utah, (2) the Rio Grande aquifer system of southern Colorado and central New Mexico, and (3) the High Plains aquifer of eastern Colorado and eastern New Mexico.

Surficial aquifers occur primarily at shallow depth in unconsolidated sediments along parts of major river valleys in Segment 2. Individual stream-valley aquifers mostly are small and separate from aquifers in other valleys or from distant aquifers in the same valley. Only in the valleys of the South Platte and the Arkansas Rivers of eastern Colorado are the aquifers large and continuous enough to form a major aquifer; therefore, the stream-valley aquifers are not mapped in **figure 11**.

The Basin and Range aquifers and the Rio Grande aquifer system generally consist of unconsolidated gravel, sand, silt, and clay, or partly consolidated sedimentary or volcanic materials. These materials have filled deep fault-block valleys formed by large vertical displacement across faults. Mountain ranges that generally consist of impermeable rocks separate adjacent valleys. When mountains encircle a valley, the aquifer in the valley is isolated, and ground water is contained within the valley. However, most valleys are interconnected, and ground water moves from valley to valley through the interconnected network of aquifers. The Basin and Range aquifers extend westward beyond Segment 2 into California, Nevada, Oregon, and Idaho; the Rio Grande aquifer system extends southward beyond Segment 2 into Texas and Mexico.

The High Plains aquifer of eastern Colorado and eastern New Mexico extends over a large area and is little affected by faulting and deformation. The primarily unconsolidated to poorly consolidated sediments of this aquifer extend beyond Segment 2 into parts of Kansas, Nebraska, Oklahoma, South Dakota, Texas, and Wyoming.

The three consolidated-rock aquifers or aquifer systems are (1) in the Denver Basin of east-central Colorado, (2) the Roswell Basin of southeastern New Mexico, and (3) the Colorado Plateaus of northern Arizona, western Colorado, northwestern New Mexico, and eastern Utah.

The Denver Basin aquifer system consists of a layered sequence of four aquifers in beds of permeable conglomerate, sandstone, and siltstone. Layers of relatively impermeable shale separate the aquifers and impede the vertical movement of ground water between the aquifers. The northern part of this aquifer system underlies the surficial aquifer of the South Platte River. Although the Denver Basin aquifer system and the surficial aquifer are hydraulically connected in part of this area, they primarily function as separate aquifer systems.

The Roswell Basin aquifer system consists of an underlying carbonate-rock aquifer and a hydraulically connected, overlying alluvial aquifer. The carbonate-rock aquifer primarily has been formed by solution openings in extensive limestone and dolomite formations of Permian age. The alluvial aquifer is in unconsolidated gravel, sand, silt, and clay that overlies the eastern part of the carbonate-rock aquifer. The alluvial aquifer hydraulically connects the carbonate-rock aquifer with surface flow in the Pecos River, which flows through the Roswell Basin.

The Colorado Plateaus aquifers are contained in a thick sequence of poorly to well-consolidated conglomerate, sandstone, siltstone, and shale. Volcanic rocks, carbonate rocks, and evaporite deposits in the area also can yield water to wells. Structural deformation, faulting, and lateral changes in the lithology of the rocks have produced a complex sequence of water-yielding layers.

Some parts of Segment 2 are shown in **figure 11** as having no major aquifers. In some areas, aquifers either do not exist or yield too little water to wells to be significant. In other areas, small aquifers yield sufficient water to supply local requirements but are not extensive enough to be classified as a major aquifer.

GEOLOGY

Aquifers in Segment 2 are present in geologic units that are varied and complex (**fig. 12**) primarily because of extensive deformation of the Earth's crust associated with the uplift of the Rocky Mountains. Prior to the mountain-building uplifts, most of the area was covered by an extensive layer of sediments that had been deposited during the previous millions of years. These layers of sediment were gradually buried and altered to form layers of rock. Today, the Great Plains area of eastern Colorado and eastern New Mexico is still underlain by a relatively flat and undeformed sequence of these rocks (**fig. 13**).

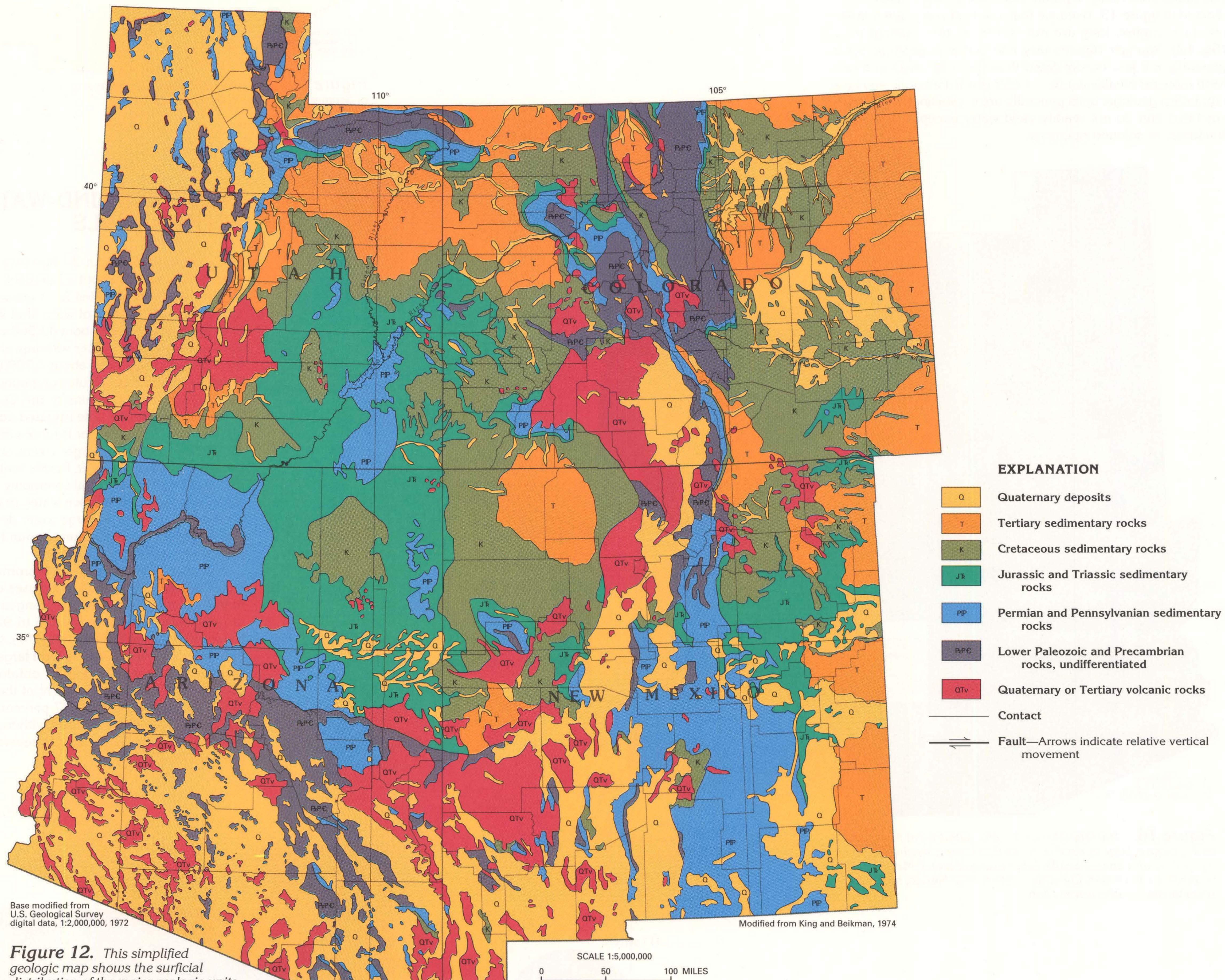


Figure 12. *This simplified geologic map shows the surficial distribution of the major geologic units in Segment 2.*

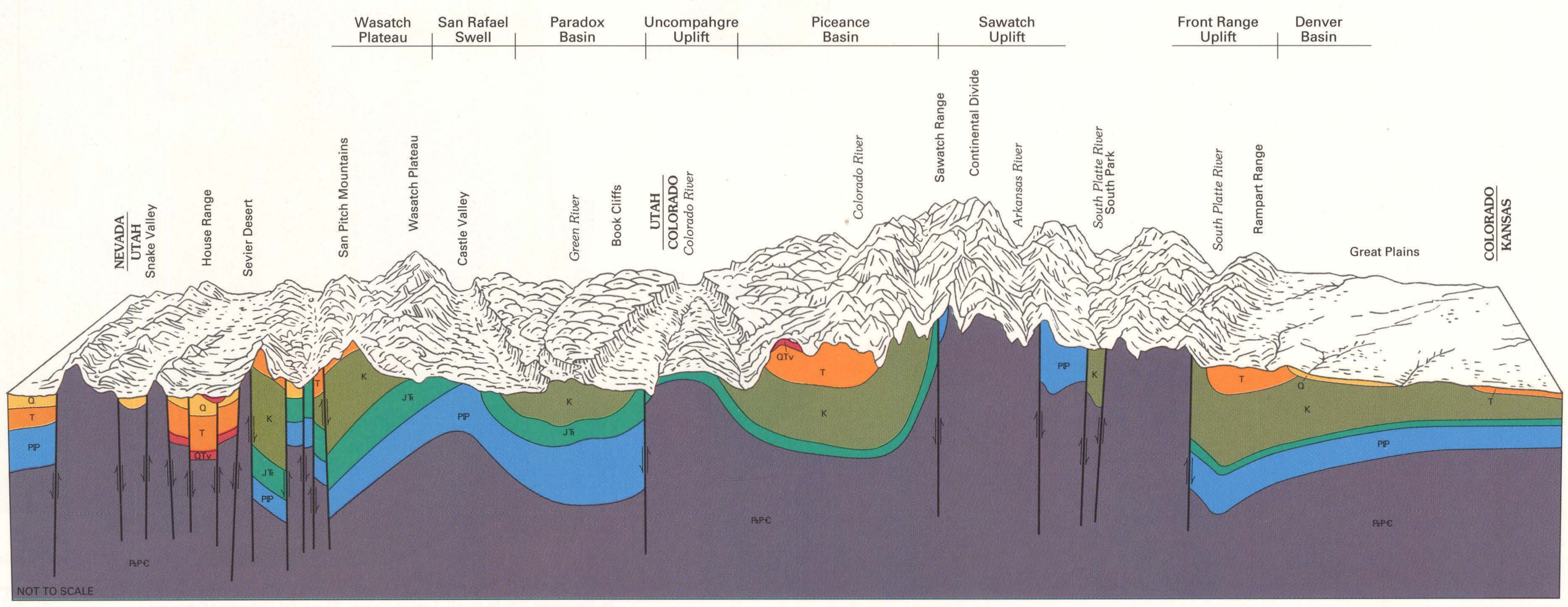


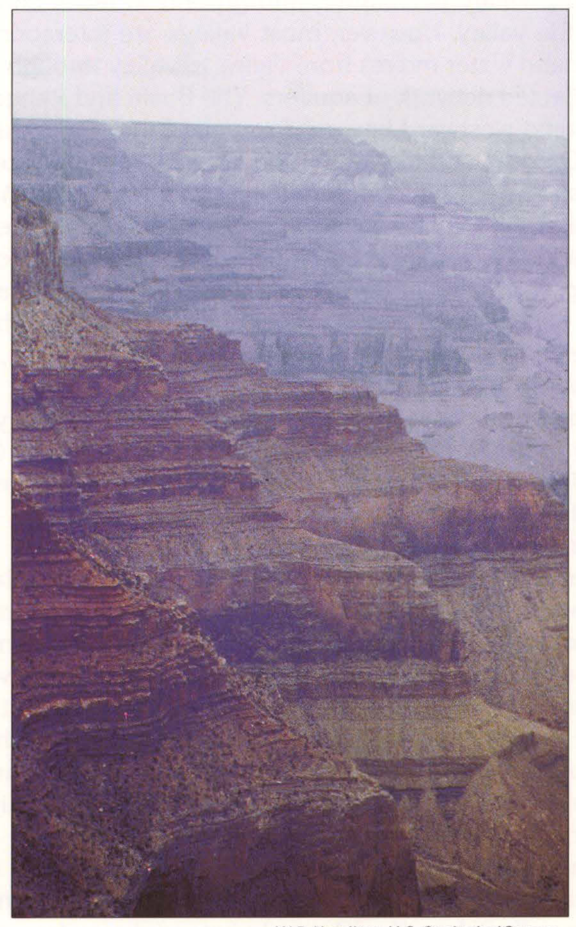
Figure 13. *This idealized block diagram of northern Utah and Colorado shows the configuration of the land surface and its relation to the generalized subsurface geology.*

GEOLOGY—Continued

The most recent uplift of the Rocky Mountains, which began about 70 million years ago, faulted, deformed, and elevated the land surface and the underlying ordered layers of rock. Faulting was prevalent, and a few faults developed more than 20,000 feet of vertical offset. As uplift continued, erosion removed the uppermost rocks and, in some areas, exposed the underlying crystalline-rock core of the mountains. Today these older crystalline rocks form many of the principal mountain ranges in Segment 2. Uplift of the Colorado Plateaus steepened stream gradients and accelerated the downcutting of the Colorado River and its principal tributaries. Downcutting of the Colorado River in the Grand Canyon has exposed thousands of feet of sedimentary rocks (**fig. 14**).

Broad structural basins, such as the Black Mesa, San Juan, Piceance, and Uinta (**fig. 15**), were formed between some of the uplifted areas. These basins generally contain an underlying, relatively undeformed sequence of rock that was deposited in the area prior to uplift and an overlying younger layer of rock and sediment that was derived from the erosion of nearby uplifted areas. Smaller basins formed by block faulting developed when a block of the Earth's crust was displaced downward with respect to adjacent uplifted blocks. Some of these basins contain older sedimentary rocks or volcanic rocks, and almost all contain a thick overlying sequence of Tertiary and Quaternary sediment derived from erosion of nearby uplifted blocks. The Basin and Range Province of western Utah and southern Arizona and the Rio Grande Rift area of central New Mexico contain many of these basins.

Rocks of various geologic age have a wide surficial distribution (**fig.12**) because of the depositional history and deformation of the area. Deformation caused extensive faulting, and faults commonly separate adjacent geologic units as indicated in **figure 13**. Because faults or fault zones are numerous or complex, they are not shown on the regional map (**fig. 12**). Younger (Quaternary and Tertiary) geologic units generally are less consolidated than the older units and can form more permeable aquifers. Older (Early Paleozoic and Precambrian) geologic units generally are crystalline or well consolidated and do not readily yield water except from faults, fractures, or solution openings.



W.B. Hamilton, U.S. Geological Survey

Figure 14. *Thousands of feet of layered sedimentary rocks such as these at Mohave Point are exposed in the rugged walls of the Grand Canyon.*

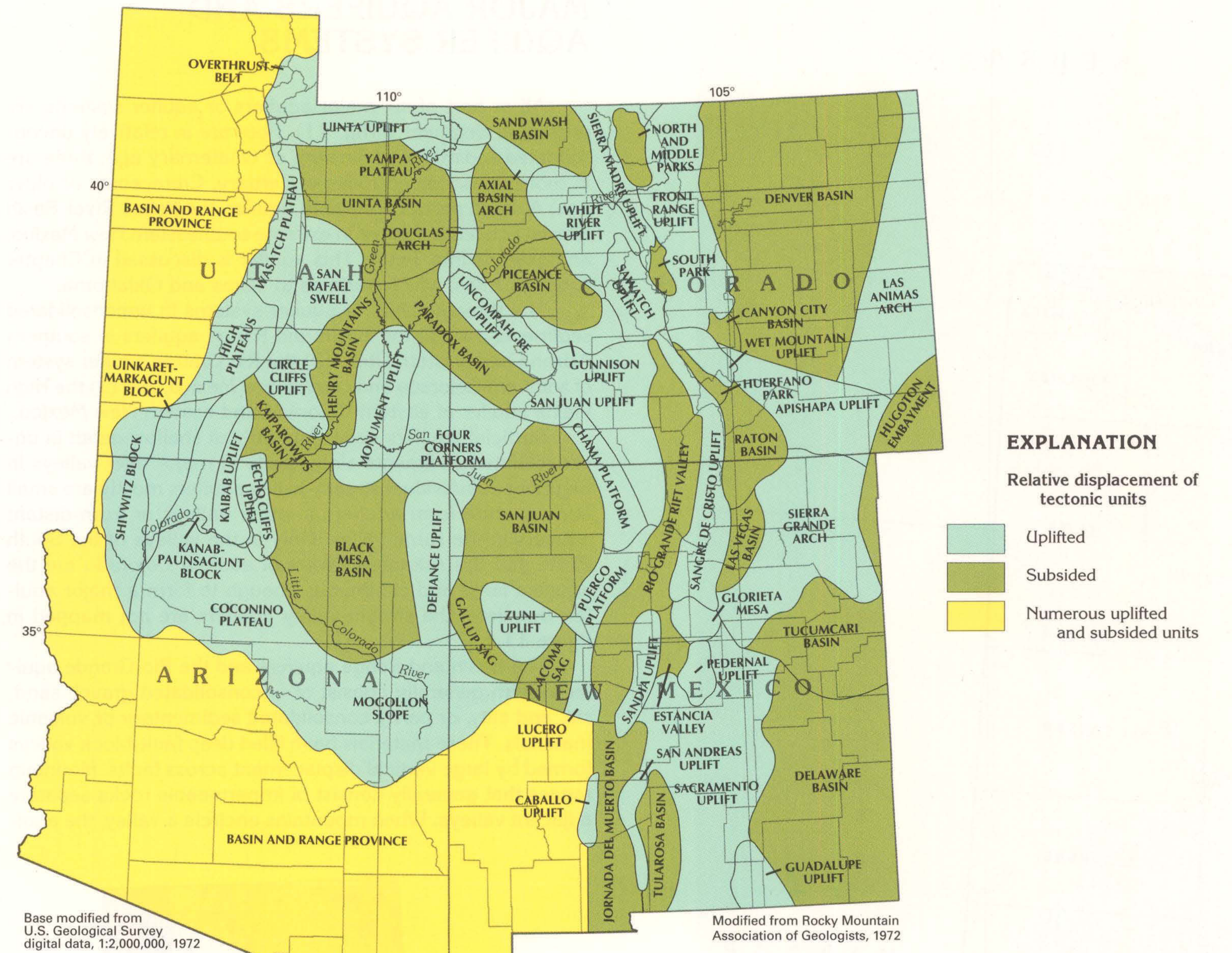


Figure 15. *The principal structural basins and uplifts of Segment 2 are shown on this map.*

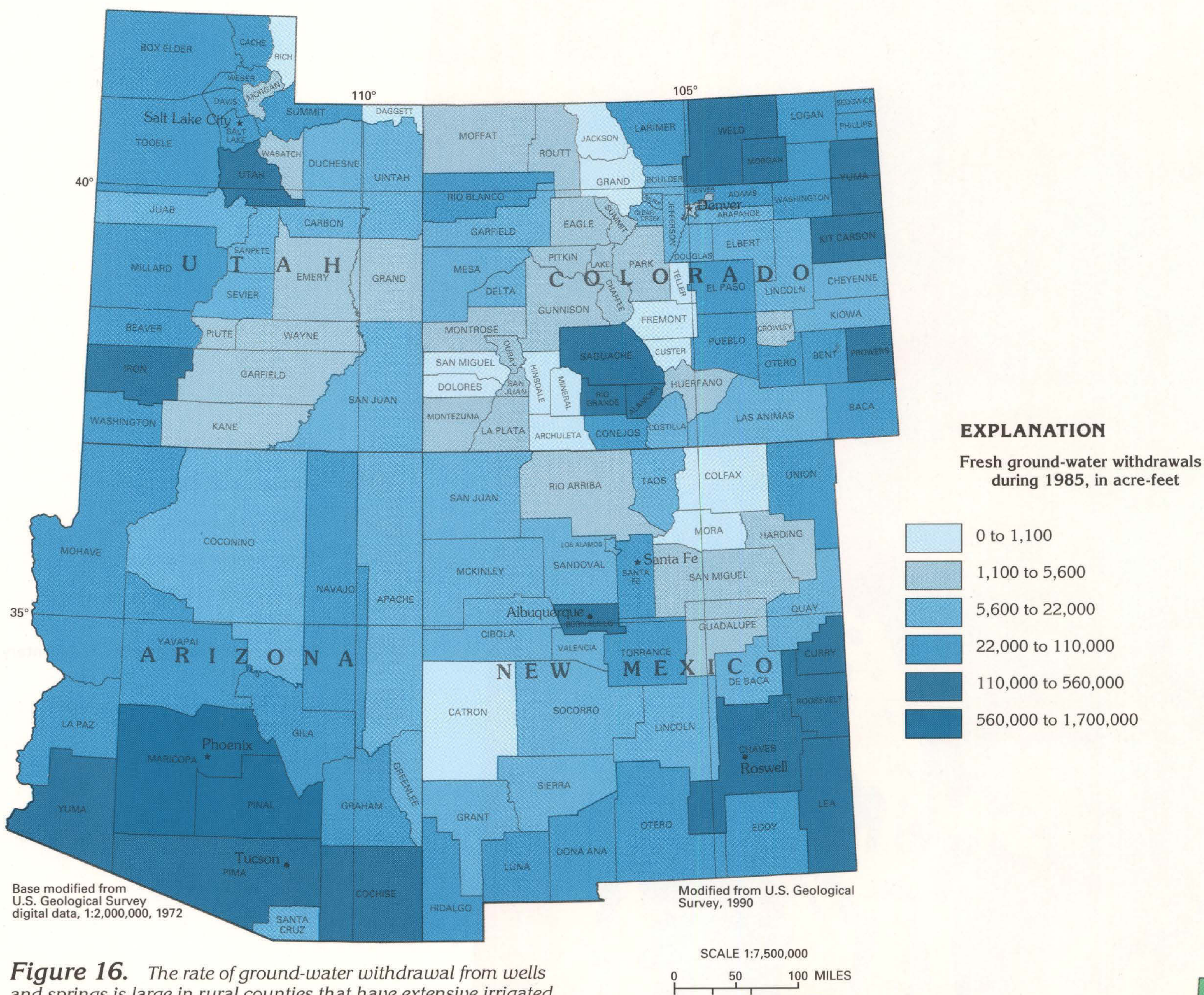


Figure 16. *The rate of ground-water withdrawal from wells and springs is large in rural counties that have extensive irrigated agriculture and limited availability of surface water. Many such counties are in eastern Colorado, eastern New Mexico, southern Arizona, and southwestern Utah.*

FRESH GROUND-WATER WITHDRAWALS

During 1985, about 31,000,000 acre-feet of water was used in Segment 2; about 28 percent, or 8,700,000 acre-feet of this water was supplied from ground-water sources. (One acre-foot is the volume of water that will cover 1 acre of land to a depth of 1 foot, or about 43,560 cubic feet of water.) The 1985 rate of ground-water withdrawal within counties ranged from less than 1,000 to about 1,700,000 acre-feet (**fig. 16**). Counties that had small rates of ground-water withdrawal generally are in the mountains or the Colorado Plateaus, where small population and little irrigated agriculture limit water requirements. Surface water is more commonly available in the mountains and along larger rivers, where it can be used in preference to ground water. Rural counties that have large rates of ground-water withdrawal commonly have extensive irrigated agriculture and little surface water. Irrigated agriculture and a large population cause large water demands in parts of Utah and southern Arizona, where ground water is an important source of supply.

During 1985, irrigation of commercial crops and watering of livestock were the largest uses of ground water in Segment 2. Such agricultural use ranged from 51 percent of all ground water withdrawn in Utah to 92 percent of all ground water withdrawn in Colorado (**fig. 17**). Ground water withdrawn for public supply was the second largest use of water in Segment 2. Public water supplies obtained from ground water constituted as little as 4 percent of the total 1985 withdrawal in Colorado to as much as 37 percent of the total 1985 withdrawal in Utah. Ground water withdrawn for mining was the principal use of ground water in several counties, primarily in mountainous areas.

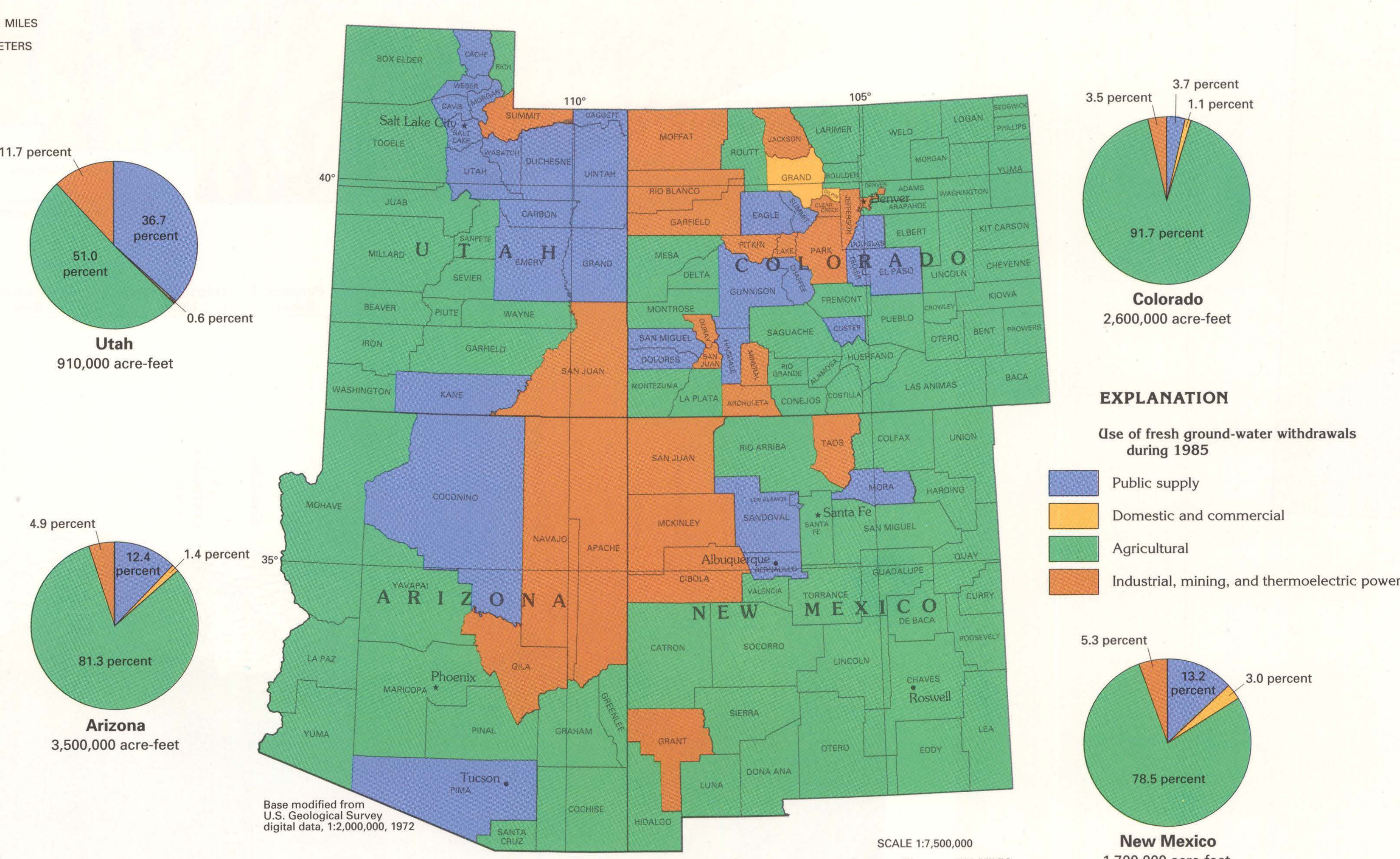


Figure 17. *The principal use of ground water in each county in Segment 2 is indicated on the map. The total volume of ground water withdrawn from wells and springs in each State during 1985 and the distribution of ground-water use in the State as a whole are shown in the pie charts.*

Surficial aquifers

INTRODUCTION

Surficial aquifers that are present in many parts of Segment 2 (**fig. 18**) generally contain the shallowest ground water in the Segment. These aquifers consist of Quaternary deposits of alluvial gravel, sand, silt, and clay or Quaternary deposits of eolian sand and silt. The alluvial and eolian deposits of the South Platte River Valley and the Arkansas River Valley are moderately thick and extensive and contain the two major surficial aquifers in Segment 2.

Surficial aquifers also are present in alluvial deposits of the Basin and Range, the Rio Grande Basin, and the Roswell Basin (**fig. 18**). However, these shallow aquifers are part of deeper and more extensive aquifers and are described with the deeper aquifers in subsequent sections of this Chapter.

In the more mountainous parts of Segment 2, much of the alluvium in the stream valleys is too thin, narrow, and discontinuous to be considered a major aquifer, even though some of the larger of the mountain alluvial deposits (**fig. 18**), such as those near the Sevier River in central Utah and in the Uinta Basin of northeastern Utah, contain locally important surficial aquifers. Alluvial or eolian deposits on the eastern plains of Colorado and New Mexico also contain surficial aquifers. However, these aquifers commonly are thin and little utilized, and the extent of the aquifers within these materials is poorly known.

The surficial aquifer along the South Platte River in Colorado was selected as an example of a stream-valley aquifer in Segment 2 because it is extensively used and well studied. Although the information presented is specific to this aquifer, the hydrologic processes described are similar in other stream-valley aquifers in the Segment.

The stream valley of the South Platte River is eroded into the surface of the underlying bedrock formations. In the mountains and plains, the bedrock generally is much less permeable than the alluvium and forms a relatively impermeable lower boundary to the alluvial aquifer. The sedimentary rocks that underlie the alluvium of the South Platte River Valley on the plains generally consist of shale and sandstone of Cretaceous age that are easily eroded to form broad, gently sloping valleys (**fig. 19**) that contain meandering streams (**fig. 20**) and moderately thick alluvium. The headwaters of the South Platte River flow on crystalline rocks of the Rocky Mountains. In this area, stream gradients are steep, the valley is narrow, and alluvium is thin and discontinuous (**fig. 21**).

The surficial aquifer of the South Platte River extends for about 200 miles in Segment 2. The aquifer ranges in thickness from about 20 to 200 feet and ranges in width from about 1 to 15 miles. Smaller deposits of alluvium or windblown sand extend up the valleys of numerous tributaries (**fig. 18**). Where saturated, these deposits form an aquifer that can yield moderate to large volumes of water to wells.

Figure 18. *Alluvial or eolian deposits are present in many parts of Segment 2. Where saturated, these deposits comprise surficial aquifers and can yield large volumes of water to wells.*

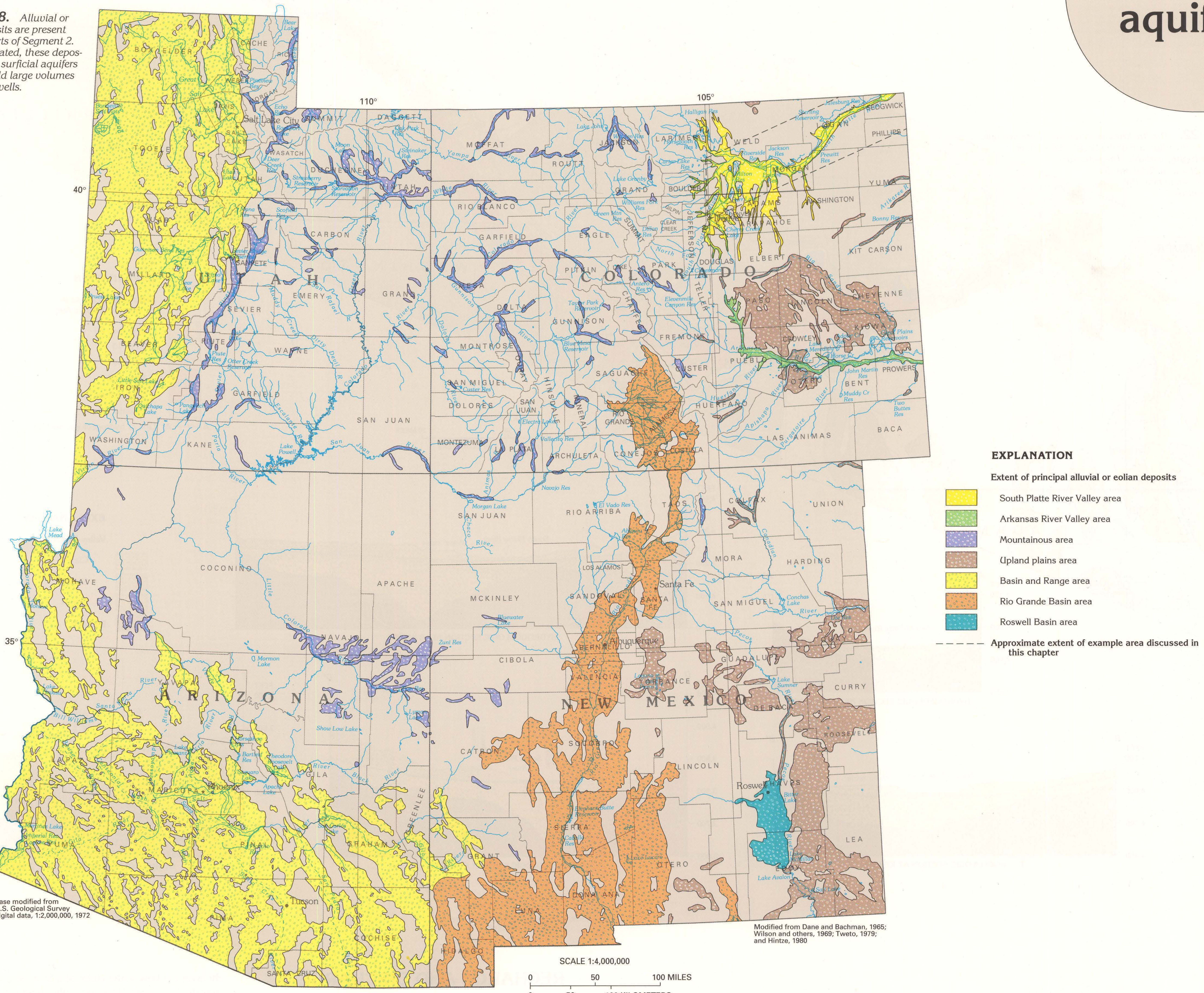


Figure 19. *The valley of the South Platte River near Fort Lupton, Colo., has a broad, sparsely wooded surface that slopes gently toward the river. The plains beyond the alluvial valley are more arid and treeless.*

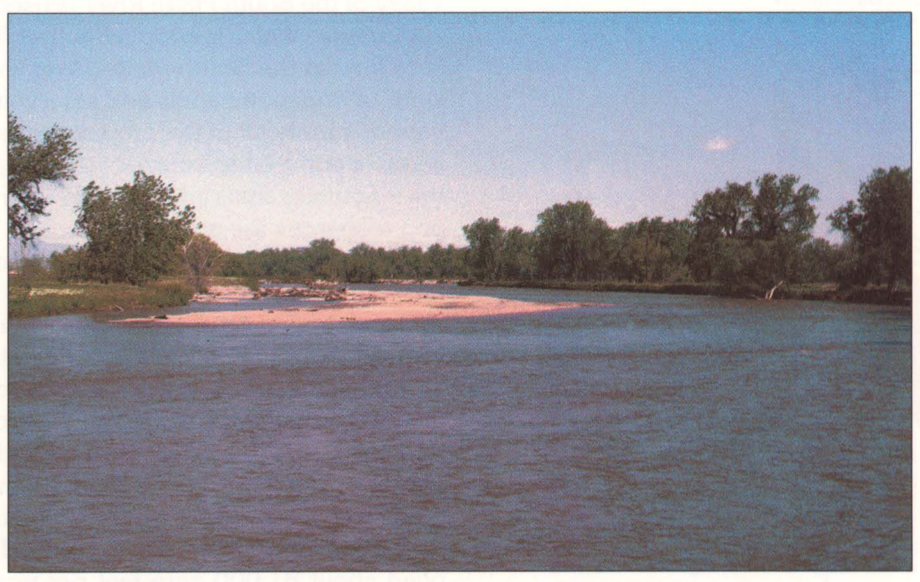


Figure 20. *The South Platte River on the plains of eastern Colorado generally is broad and shallow with numerous sand bars in the meandering channel.*

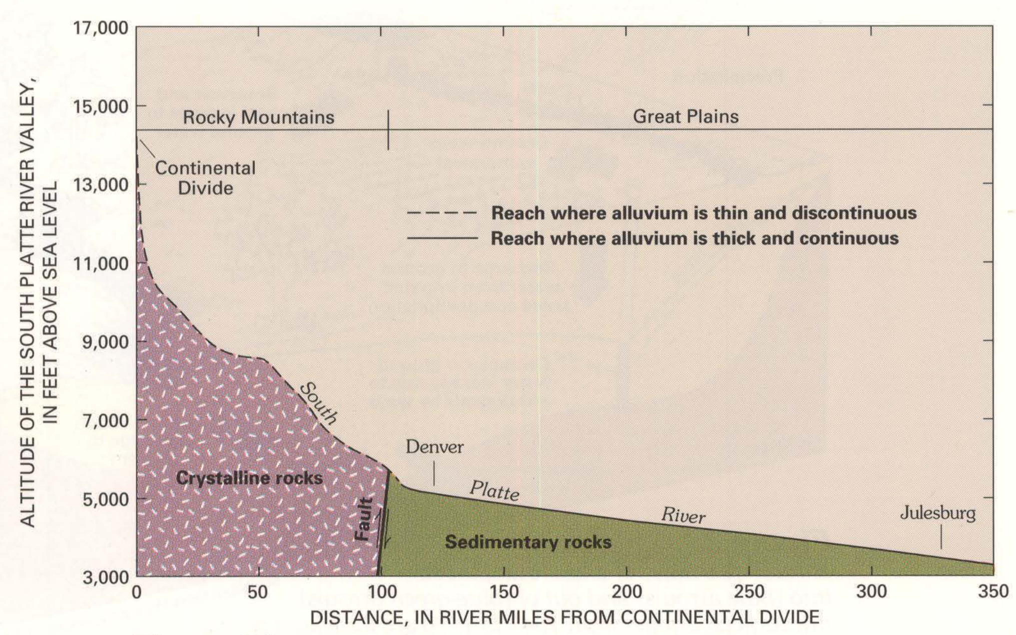


Figure 21. *In the Rocky Mountains, the valley of the South Platte River is steep and contains little alluvium. On the plains, the valley slopes gently and the alluvium is thick and continuous.*

HYDROGEOLOGIC UNITS

Alluvium in the valley of the South Platte River consists of poorly sorted mixtures of unconsolidated gravel, sand, silt, and clay or interlayered beds of relatively well-sorted sand, gravel, or silty clay. Beds may range in thickness from a few inches to 100 feet or more. Beds of gravel and cobbles occur at the base of the alluvium in the valley near Denver. Materials of finer grain size are more prevalent in the downstream parts of the valley in Colorado.

The thickness of the alluvium increases down the stream valley. In Denver, the alluvium has a maximum thickness of about 20 feet; 60 miles downstream (fig. 22, section *B–B'*), it has a maximum thickness of about 100 feet; about 100 miles farther downstream (fig. 22, section *D–D'*), the alluvium has a maximum thickness of about 200 feet. The thickness of the alluvium is more variable across than along the stream valley. In most areas, the alluvium thickens gradually from the margins of the valley to near the center (**fig. 22**, section *B–B'*). However, in some areas, the thickness of the alluvium is irregular because of the presence of two or more channels cut into the buried bedrock surface (**fig. 22**, section *E–E'*). Where tributary valleys enter the main valley, or where thick deposits of windblown sand overlie the alluvium, a thick part of the valley-fill material can extend beyond the valley margin (**fig. 22**, section *D–D'*).

The water table in the alluvium forms the top of the surficial aquifer. The saturated thickness of the aquifer is the distance from the water table to the base of the alluvium. Almost the entire thickness of the alluvium is saturated near the South Platte River where the water table is near land surface. Near the margins of the valley, the saturated thickness of the aquifer is generally much smaller than the thickness of the alluvium.

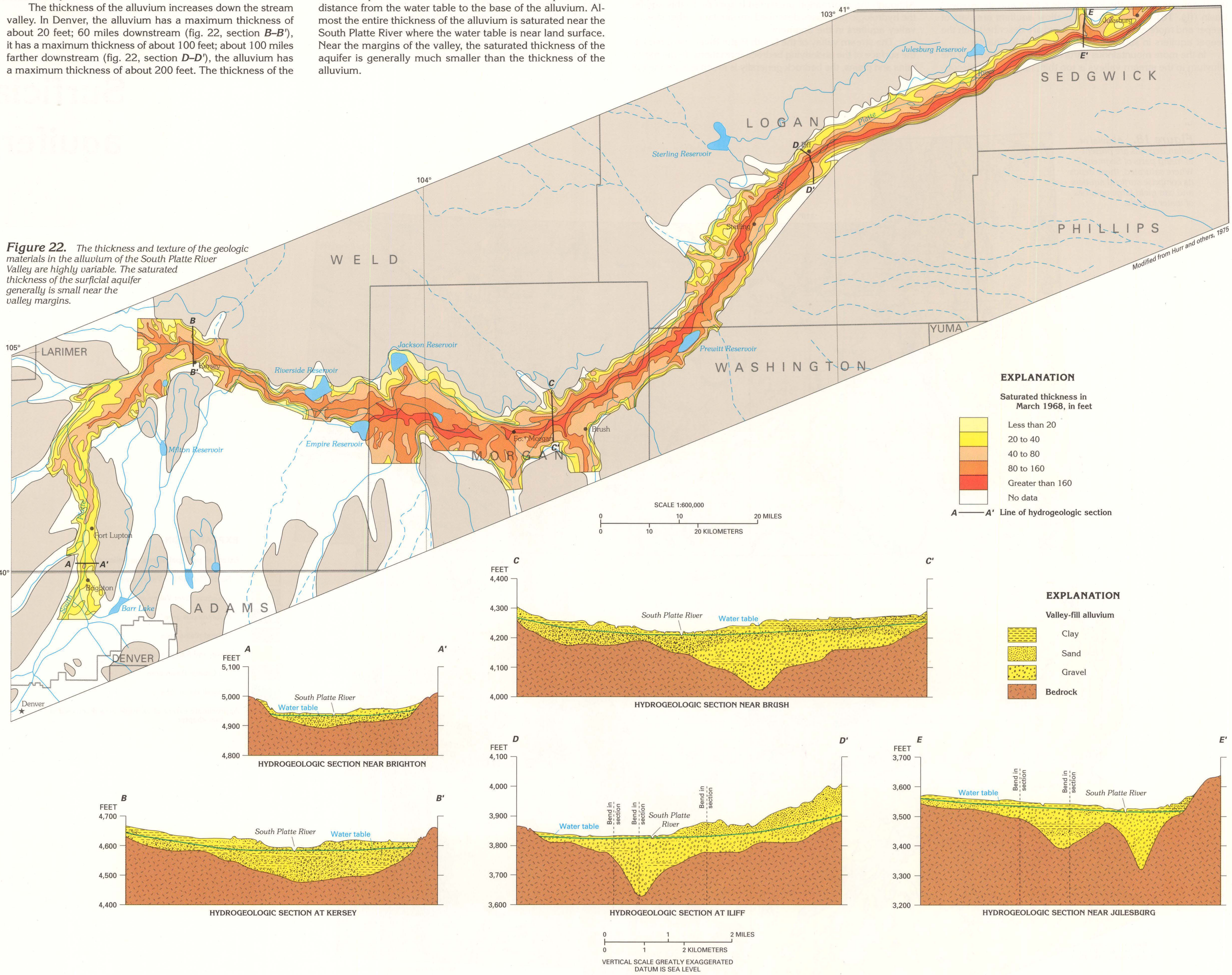


Figure 22. The thickness and texture of the geologic materials in the alluvium of the South Platte River Valley are highly variable. The saturated thickness of the surficial aquifer generally is small near the valley margins.

RECHARGE AND DISCHARGE

The volume of water that flows into (recharges) the surficial aquifer or flows out of (discharges from) the aquifer is affected by the presence of surface water in the area. The valley of the South Platte River is used extensively for irrigated agriculture. Water is supplied to the fields by diversion of surface flow in the South Platte River through an extensive network of irrigation canals and reservoirs. Surface water usually cannot supply all of the crop water requirements, and ground water is pumped to augment the supply. Some of the water in the irrigation canals and reservoirs percolates to the water table and recharges the aquifer. Between 1947 and 1970, the average rate of recharge from canals and reservoirs was about 104,000 acre-feet per year.

Part of the water applied to irrigated fields and part of the precipitation that falls in the valley also percolates downward and recharges the aquifer. Deep percolation of water applied to irrigated fields combined with precipitation supplied about 914,000 acre-feet per year of recharge to ground water (**fig. 23**). Recharge also occurs by inflow of water from adjacent alluvial aquifers, primarily in tributary valleys, or from underlying bedrock aquifers. Combined, these two sources supplied about 81,000 acre-feet per year of recharge. The total annual recharge to the surficial aquifer near the South Platte River was about 1,099,000 acre-feet.

Ground-water seepage to the channel of the South Platte River and withdrawal from wells are the most important forms of ground-water discharge. The South Platte River forms a natural drain for the surficial aquifer along almost all of the length of the river in eastern Colorado. During 1967–69, for example, the average rate of surface-water diversion from the river in Colorado was about 922,000 acre-feet per year, and the average flow at the Nebraska State line was about 340,000 acre-feet per year. However, only about 682,000 acre-feet per year of surface water flowed to the river from the headwaters and tributaries during this period. The difference in flow of about 580,000 acre-feet per year was supplied by ground-water discharge to the river.

In 1930, there were about 200 wells in the surficial aquifer near the South Platte River that were capable of yielding more than 100 gallons per minute. By 1970, the number of such wells had increased to slightly more than 3,200; the number of wells has remained relatively constant since 1970. The average yield of these wells was about 920 gallons per minute in 1970, but some well yields were as large as 3,200 gallons per minute. Most of these wells supply water for crop irrigation, but a few are used for the municipal supply of several communities in the valley. Numerous small-capacity wells supply water for domestic, stock, and other uses. Ground-water withdrawal for irrigation was about 388,000 acre-feet per year during 1951–60 and about 556,000 acre-feet per year during 1961–70.

During the period from 1947–70, discharge from the surficial aquifer by ground-water seepage to the South Platte River averaged about 515,000 acre-feet per year. Withdrawal by wells averaged about 420,000 acre-feet per year, evapotranspiration from vegetation growing in areas of shallow water table (phreatophytes) averaged about 163,000 acre-feet per year, and outflow to Nebraska of ground water in the alluvium averaged about 20,000 acre-feet per year (**fig. 23**). The total discharge from the surficial aquifer averaged about 1,118,000 acre-feet per year. Discharge exceeded recharge by about 19,000 acre-feet per year; as a result, there was a decline in the volume of ground water in storage in the surficial aquifer. For comparison, the total volume of ground water in storage in the aquifer was about 8,300,000 acre-feet.

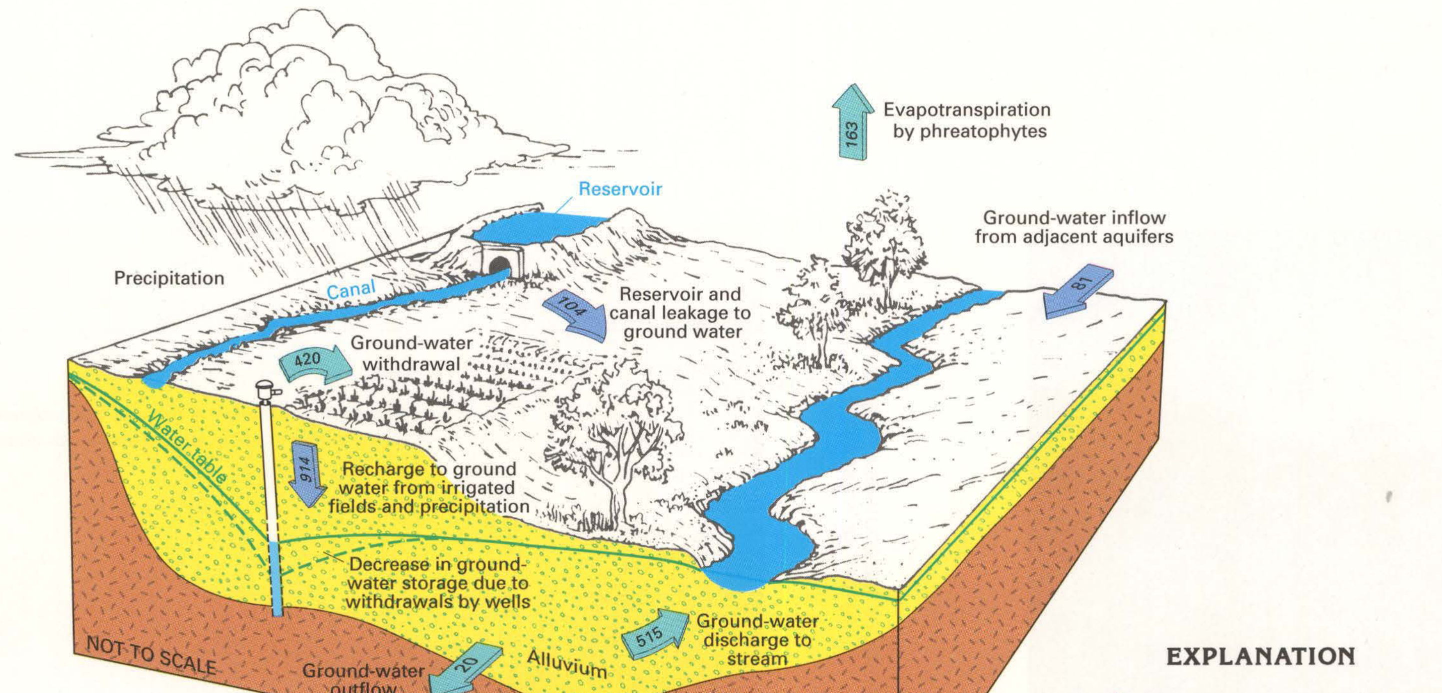


Figure 23. The average annual quantities of water that flowed into (blue arrows) and out of (blue-green arrows) the surficial aquifer from 1947 to 1970 are indicated on this diagram. Discharge exceeded recharge by about 19,000 acre-feet; this volume of water represents annual depletion of stored water in the aquifer.

WATER-LEVEL CONDITIONS

Ground water flows from areas of recharge, where the water-table altitude is higher, toward areas of discharge, where the water-table altitude is lower. Recharge areas associated with canals, reservoirs, and irrigated fields generally are located along the valley on either side of the South Platte River. As a result, the altitude of the water table near the margin of the valley generally is higher than in the central part of the valley (**fig. 22**, sections *A–A'* through *E–E'*). Because ground-water discharges readily to the river, the altitude of the river determines the altitude of the water table along the valley. For example, the altitude of the river and the water table both range from about 5,000 feet near Denver to about 3,500 feet at Julesburg, which is 200 miles downstream (**fig. 24**).

The direction of ground-water movement generally is down the valley and toward the South Platte River, as indicated by the arrows on **figure 24**. This movement and the practice of diverting surface water creates a cycle of water reuse. Ground water discharges to the river; part of the river's flow is diverted into irrigation canals, reservoirs, and irrigated fields where some of the water percolates back into the aquifer. Some of this recharge ultimately flows back into the river. As a result, some water that flows into this stream-aquifer system near Denver may be used and reused many times before it flows out of Colorado. Only in the spring when irrigation demands are small and streamflow is large as a result of runoff from the mountain snowpack can large volumes of surface water flow directly downstream and bypass the ground-water system.

Ground-water withdrawal lowers the water table and either decreases ground-water flow to the river or causes water to flow from the river into the aquifer near the site of withdrawal. Thus, ground-water withdrawal has an effect on streamflow similar to that of surface-water diversion. Both decrease streamflow, but the slow rate of ground-water movement delays the effect of withdrawal on streamflow, whereas the effect of surface-water diversion is immediate.

Between 1947 and 1970, water levels declined as much as 35 feet in parts of the surficial aquifer (**fig. 25**). The magnitude of annual decline varied from year to year, primarily in response to changes in the annual rate of ground-water withdrawal. Other factors that affect the decline include availability of surface water, precipitation during the growing season, and crop size and water requirements.

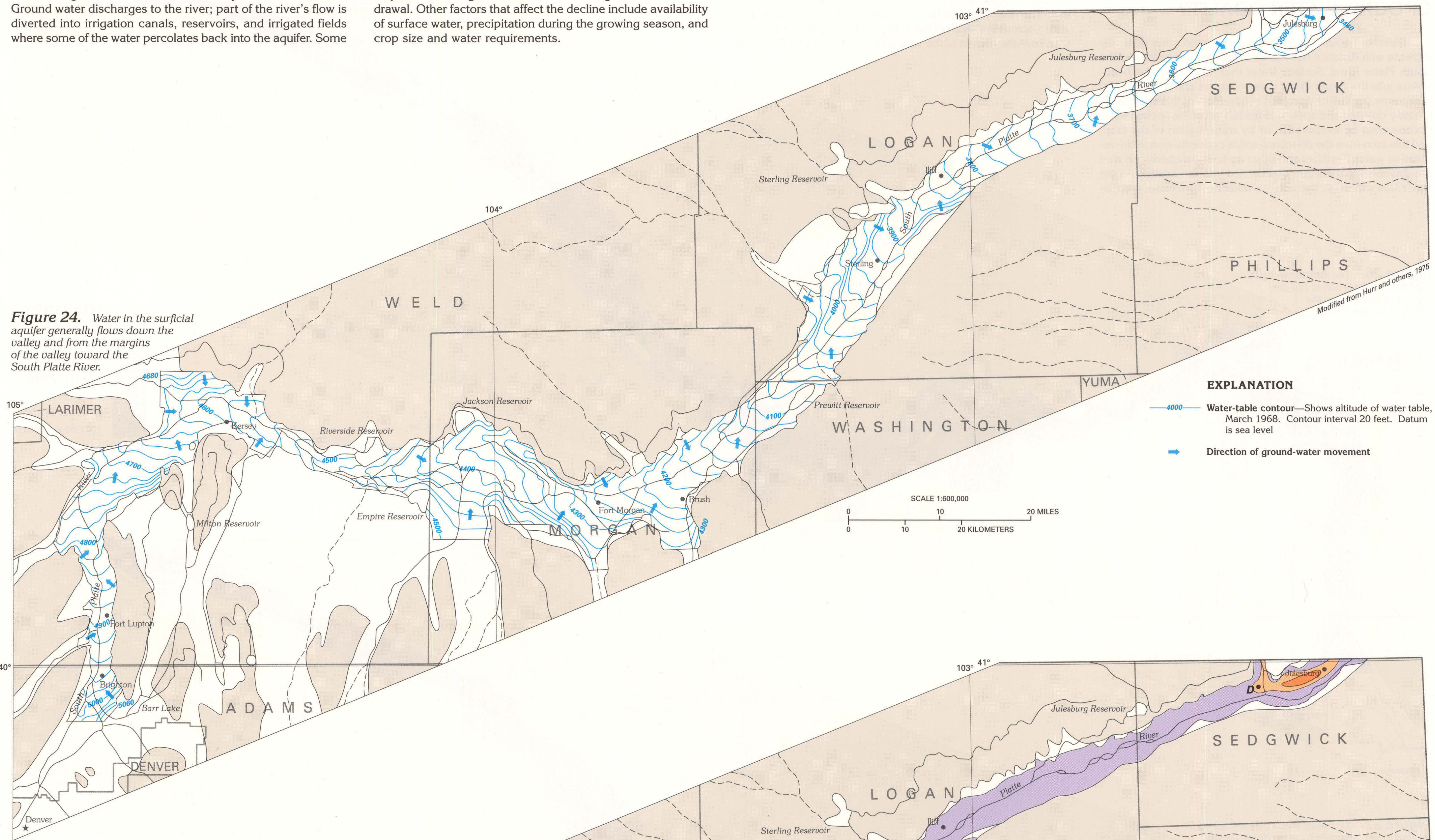


Figure 24. Water in the surficial aquifer generally flows down the valley and from the margins of the valley toward the South Platte River.

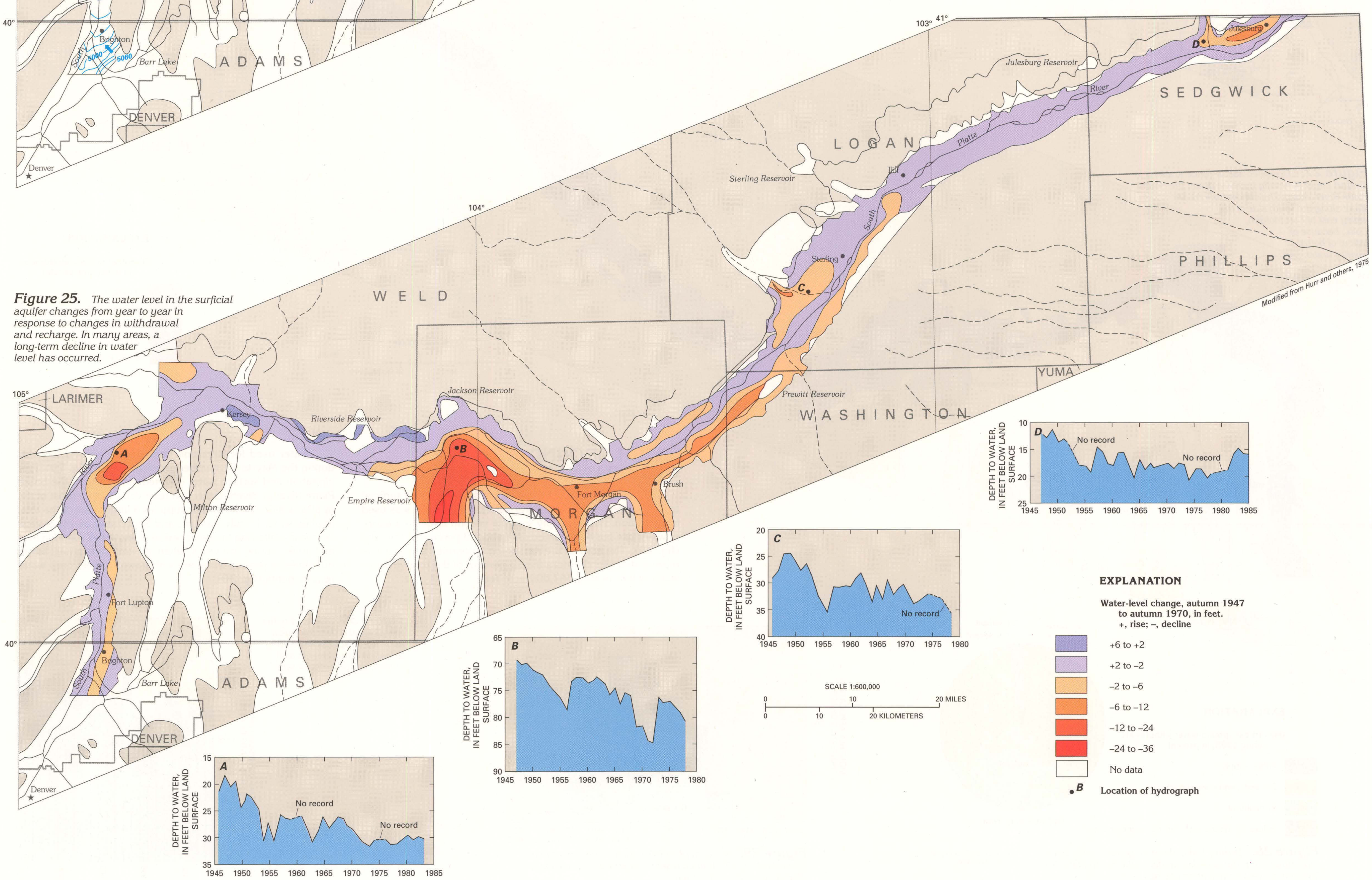


Figure 25. The water level in the surficial aquifer changes from year to year in response to changes in withdrawal and recharge. In many areas, a long-term decline in water level has occurred.

AQUIFER CHARACTERISTICS

The ability of the surficial aquifer to transmit water is determined by the thickness of the aquifer and by the uniformity and coarseness of the sand and gravel in the aquifer. This ability to transmit water is called transmissivity. An aquifer of large transmissivity easily transmits water and can yield larger volumes of water to wells than an aquifer of small transmissivity.

The transmissivity of the surficial aquifer along the South Platte River ranges from less than 1,000 feet squared per day along the margins of the aquifer where the aquifer is thin to more than 100,000 feet squared per day in a few areas near the central part of the lower valley where the aquifer is thick (**fig. 26**). The transmissivity of the aquifer also increases down the valley, primarily because of the increased thickness of alluvium downstream.

GROUND-WATER QUALITY

Dissolved-solids concentrations in ground water generally increase with distance downstream from the headwaters of the South Platte River. Surface water that flows from the headwaters into the South Platte River contains about 300 to 800 milligrams per liter of dissolved solids. Most of this water is ultimately diverted and applied to fields. Part of the applied water is consumed by evaporation or by transpiration of the crop, and this increases the dissolved-solids concentration in the remaining water. Fertilizer and other agricultural chemicals also are dissolved in the water and carried into the aquifer. As the water flows through the aquifer, additional minerals are dissolved from the alluvial material. As a result, the ground water that returns to the river contains larger concentrations of dissolved solids than the water that was diverted from the river. Downstream, water again is diverted from the river and applied to another field. This recycling of water causes an increase in dissolved-solids concentration in the ground water down the valley. In some areas, dilution occurs when recharge from precipitation and inflow from tributary aquifers and upland areas underlain by thin alluvium or sand dunes decreases the dissolved-solids concentrations in water in the surficial aquifer (**fig. 27**). These processes cause the decreases in dissolved-solids concentration evident along the south side of the valley downstream from Fort Morgan, Colo.

Dissolved-solids concentrations of ground water range from about 1,000 milligrams per liter near Denver to as much as 4,000 milligrams per liter at Sterling. Downvalley from near Sterling, the concentration decreases because of dilution by ground-water inflow. Near Julesburg, the dissolved-solids concentration of the ground water is about 1,500 milligrams per liter. The dissolved-solids concentration of ground water also varies across the width of the valley and is larger near the river than near the margin of the valley. Surface water that flows into Nebraska has an average dissolved-solids concentration of about 1,300 milligrams per liter.

Water in the surficial aquifer generally contains large concentrations of dissolved calcium, bicarbonate, and sulfate, and is classified as either a calcium bicarbonate- or calcium sulfate-type water. Calcium bicarbonate water is more prevalent in the upstream part of the valley; calcium sulfate water is more prevalent in the downstream part. Water is classified as very hard if it contains more than 200 milligrams per liter hardness, measured as calcium carbonate. Water from more than 90 percent of a sampling of 89 wells in the downvalley part of the aquifer had hardness in excess of 200 milligrams per liter, and 50 percent had hardness in excess of 600 milligrams per liter. Ground water in some areas also contains large concentrations of dissolved iron, nitrate, or chloride. Although many local residents and communities have long used the aquifer as a source of drinking water, the water is not well suited for this use because of its taste, large dissolved-solids concentrations, hardness, or localized large concentrations of iron, nitrate, or sulfate. In some areas, using ground water for irrigation may require careful management of the water to prevent buildup of crop-damaging salts in the soil. However, in most areas, the surficial aquifer yields water that is of suitable quality for irrigation, and the dissolved minerals that make the water objectionable for public or domestic uses are not a serious problem for irrigation use.

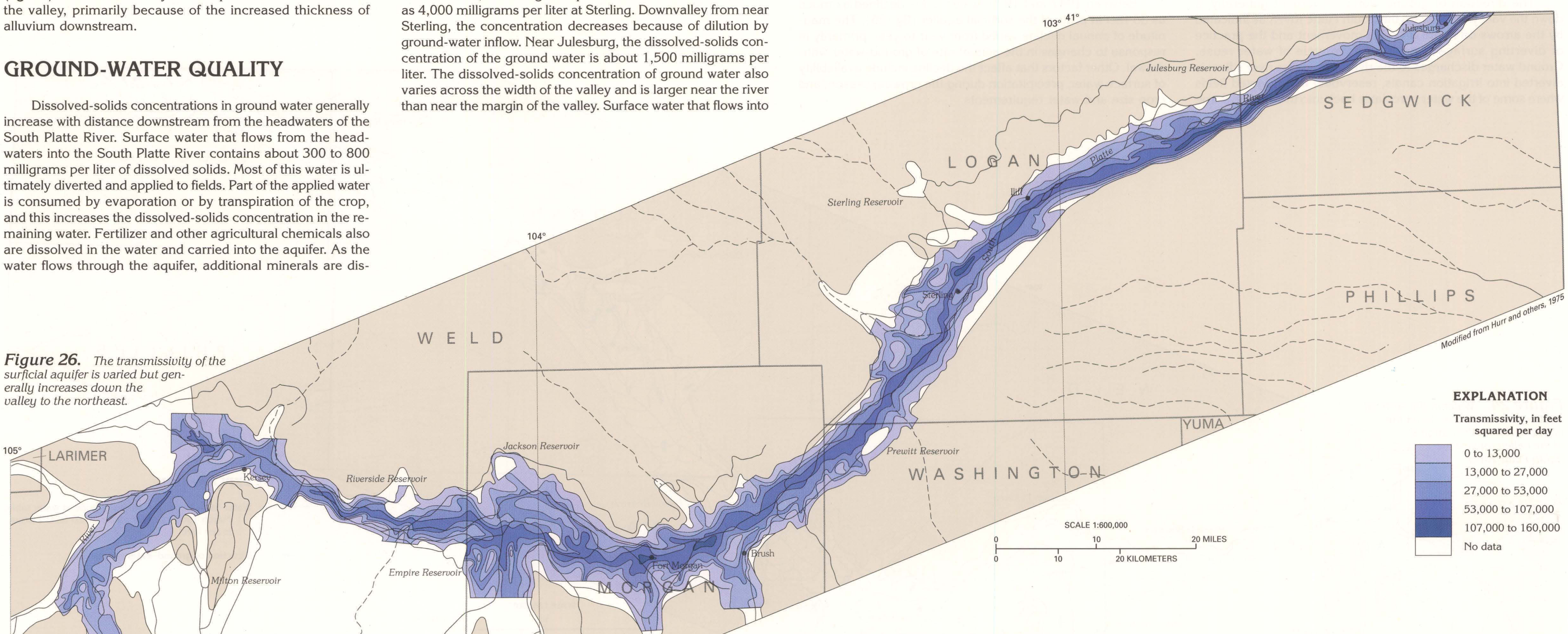


Figure 26. *The transmissivity of the surficial aquifer is varied but generally increases down the valley to the northeast.*

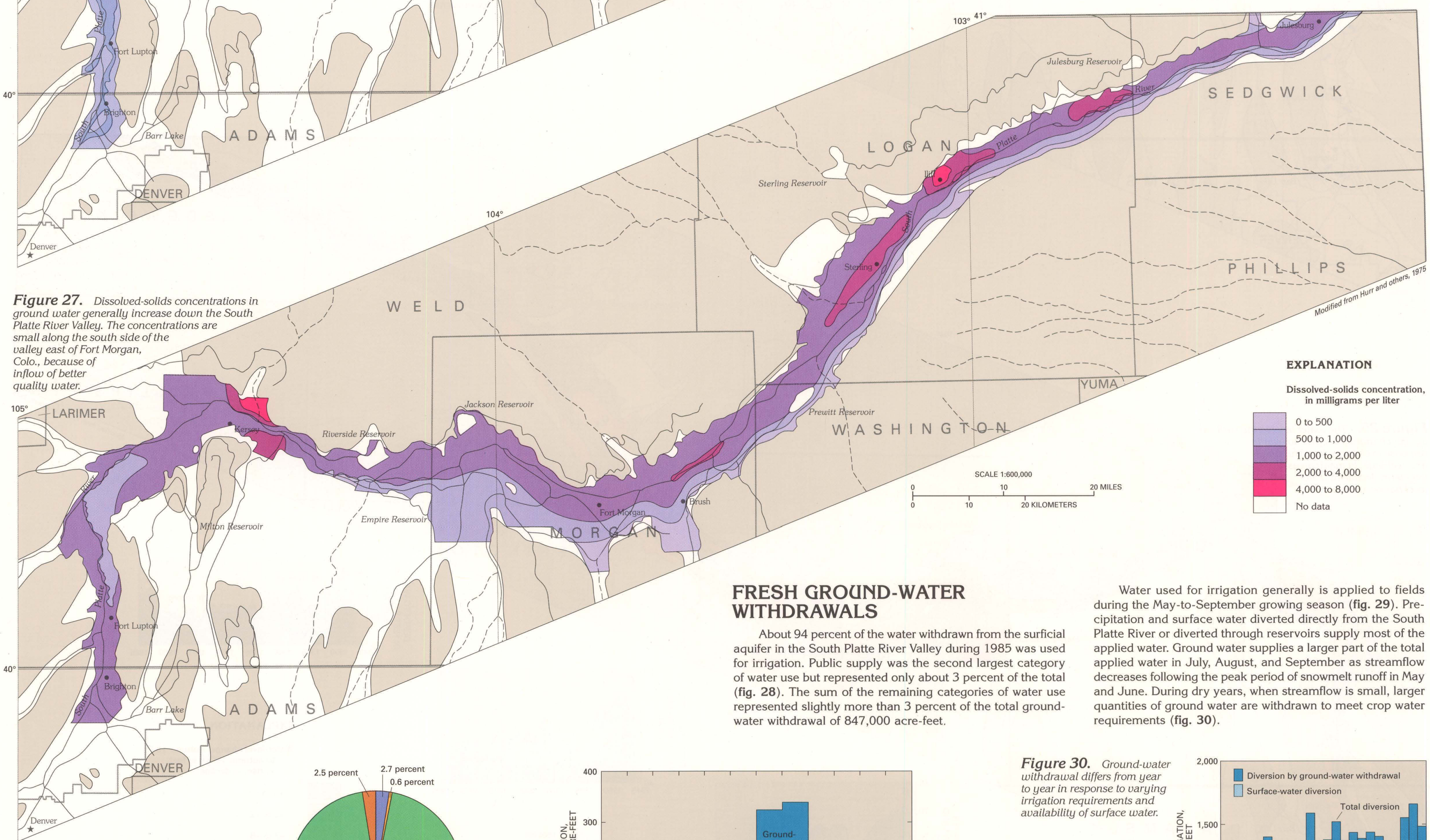


Figure 27. *Dissolved-solids concentrations in ground water generally increase down the South Platte River Valley. The concentrations are small along the south side of the valley east of Fort Morgan, Colo., because of inflow of better quality water.*

FRESH GROUND-WATER WITHDRAWALS

About 94 percent of the water withdrawn from the surficial aquifer in the South Platte River Valley during 1985 was used for irrigation. Public supply was the second largest category of water use but represented only about 3 percent of the total (**fig. 28**). The sum of the remaining categories of water use represented slightly more than 3 percent of the total ground-water withdrawal of 847,000 acre-feet.

Water used for irrigation generally is applied to fields during the May-to-September growing season (**fig. 29**). Precipitation and surface water diverted directly from the South Platte River or diverted through reservoirs supply most of the applied water. Ground water supplies a larger part of the total applied water in July, August, and September as streamflow decreases following the peak period of snowmelt runoff in May and June. During dry years, when streamflow is small, larger quantities of ground water are withdrawn to meet crop water requirements (**fig. 30**).

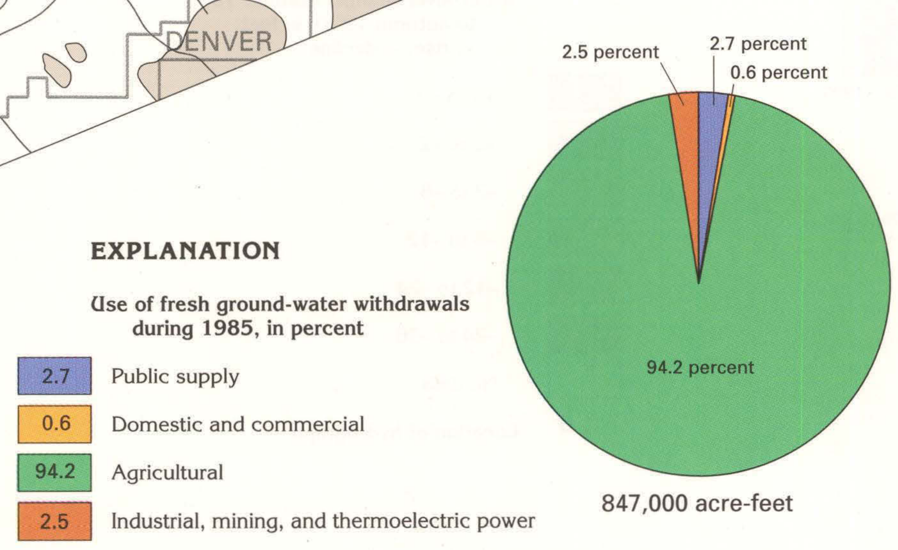


Figure 28. *During 1985, about 94 percent of the water withdrawn from the surficial aquifer in the South Platte River Valley was used for irrigation of crops.*

Data from U.S. Geological Survey files, 1990

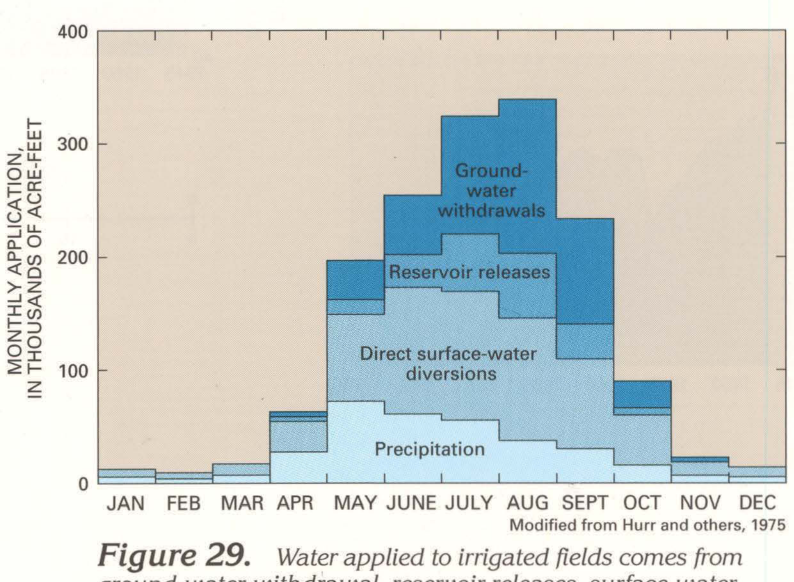
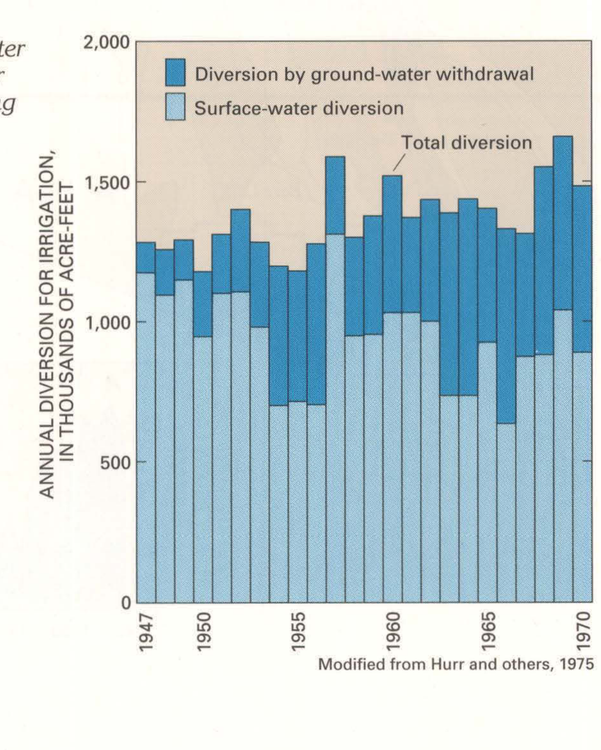


Figure 29. *Water applied to irrigated fields comes from ground-water withdrawal, reservoir releases, surface-water diversions, and precipitation.*

Figure 30. *Ground-water withdrawal differs from year to year in response to varying irrigation requirements and availability of surface water.*



Basin and Range aquifers

INTRODUCTION

The Basin and Range aquifers extend through about 200,000 square miles of the southwestern United States and underlie most of Nevada and parts of eastern California, southern Oregon and Idaho, western Utah, southern Arizona, and southwestern New Mexico. The aquifers extend through most of the Basin and Range Physiographic Province and consist of about 400 alluvial basins that contain aquifers. The eastern part of the Basin and Range aquifers as described in this report includes the aquifers of western Utah, southern Arizona, and southwestern New Mexico (**fig. 31**). The aquifers comprise the eastern part of the Great Basin aquifer system in Utah and the western part of the Southwest alluvial basins aquifer system in Arizona as defined by U.S. Geological Survey Regional Aquifer-System Analysis studies. The western part of the Basin and Range aquifers is described in Chapter B of this Atlas, which pertains to California and Nevada.

The Basin and Range aquifers are the principal sources of ground water in western Utah and southern Arizona. The aquifers are present in about 120 alluvium-filled basins interspersed between ranges of mountains. About 150,000,000 acre-feet of recoverable ground water is in storage in the upper 100 feet of the saturated sediments of these basins. The ground water in some basins is extensively utilized, and large water-level declines have occurred; in other basins, population is sparse, ground water is little utilized, and water levels are stable.

The 95,000-square-mile area of the aquifers in Segment 2 ranges in altitude from about 150 feet near Yuma in southwestern Arizona to more than 10,000 feet at the crests of a few desert mountain ranges. Most of the mountain ranges protrude 3,000 to 6,000 feet above the level of the surrounding basins and extend in a northerly or northwesterly direction for 10 to 50 miles. The land surface of the basins generally slopes gently from the adjacent mountain fronts toward the flat-lying central parts of the basins, where dry lake beds (playas) or shallow sandy stream channels are common (**fig. 32**). Some basins are topographically closed, and all surface water drains to a central lake or playa; other basins are topographically open, and surface water may flow between basins. The high temperatures and arid climate of the desert region result in minimal streamflow, and the stream channels and playas generally are dry.

HYDROGEOLOGIC UNITS

Structure and lithology are the principal geologic factors that affect the occurrence and movement of ground water in the Basin and Range aquifers. The principal aquifers are in thick deposits of basin fill in valleys bounded by mountain ranges formed mostly of relatively impermeable bedrock.

The structural deformation that produced the system of basins and ranges generally began in Tertiary time with block faulting along steeply dipping normal faults. Crustal extension produced horst and graben blocks in some places (**fig. 33*A***) and tilted blocks in others (**fig. 33*B***). The downthrown parts of the blocks became basins; the upthrown parts became mountain ranges. Vertical displacement across the fault zones exceeded 10,000 feet in some areas. Many of the resulting basins are asymmetrical because the grabens are not centered in the valleys. As the mountain blocks were uplifted and eroded, sediment was carried by streams into the basins, and alluvial fans were formed. The fans coalesced to produce broad surfaces that sloped gently to the center of the basins, where fine-grained sediments were deposited in lakes and playas (**fig. 34**). Coarse-grained sediments tended to be deposited near the steeper margins of the basins. Fault movement resulting from continuing deformation offset some of these older sediments. Deformation and sedimentation occurred at different rates throughout the area; as a result, the thickness, areal extent, and grain size of the basin fill are highly varied.

Basin fill (**fig. 35**) primarily consists of unconsolidated to moderately consolidated, well- to poorly sorted beds of gravel, sand, silt, and clay deposited on alluvial fans, pediments, flood plains, and playas. More cemented or compact sediments in the older basin fill and finer grained sediments near the center of the basin are less permeable than the coarser grained sediments near the margins of the basins. Evaporites, such as gypsum, anhydrite (calcium sulfate), and halite (rock salt) are present in the deeper fine-grained sediments of the central parts of some basins. Extrusive volcanic rocks also are interspersed with basin fill in some basins; volcanic rocks overlie basin fill in a few areas. The thickness of the basin fill is not well known in some basins but ranges from about 1,000 to 5,000 feet in many basins and may exceed 10,000 feet in a few deep basins in Utah and south-central Arizona.

Stream alluvium that consists of moderately to well-sorted, fine to coarse sand with gravel, silt, and clay is present along most of the larger stream channels. These deposits are about 100 feet thick and 1 to 2 miles wide along the Gila, Salt, and Santa Cruz Rivers in Arizona and exceed 700 feet in thickness along the lower Colorado River.

Extensive layers of sediments also were deposited on the bed of Lake Bonneville, a Pleistocene lake that covered about 20,000 square miles of western Utah (**fig. 36**) during the last phase of the great ice age (26,000–11,000 years ago). These sediments ranged in size from sand, gravel, and boulders near the margins of the ancient lake to clay, silt, and sand in the deeper parts of the lake. The lake level remained relatively stable for long periods of time as the volume of water that flowed into the lake was virtually balanced by the volume of outflow and evaporation. Eventually, the rate of evaporation exceeded the rate of inflow, and the lake level progressively declined; the once large body of freshwater was reduced to bodies of brine that today are the Great Salt Lake and Utah Lake. As Lake Bonneville retreated and evaporated, extensive deposits of halite (rock salt) were formed in the lakebed sediments and on the surface of what is now the Great Salt Lake Desert. Important sources of fresh ground water are present in coarser grained sediments near the mountain fronts of this area; saline ground water is common in shallow aquifers or in fine-grained sediments in the central parts of basins.

Bedrock is present in the uplifted blocks of the mountain ranges and beneath the basin fill in the valleys. Bedrock consists of consolidated carbonate rocks or metamorphic, igneous, and clastic rocks that are relatively impermeable unless extensively fractured (**fig. 35**). Fracturing in carbonate rocks (limestone and dolomite) may enable ground water to circulate through the fractures where the water can dissolve the slightly soluble rock and enlarge and increase the size and number of pathways for water movement through the rock. Such dissolution eventually can change a relatively impermeable carbonate rock into a permeable water-yielding unit. Carbonate rocks predominate in a 20,000- to 30,000-foot-thick sequence of Paleozoic and Lower Mesozoic rocks in an extensive area of western Utah in Segment 2 (**fig. 35**) and southern and eastern Nevada in Segment 1. The location of solution-altered zones of enhanced permeability within these carbonate rocks is poorly known. However, some data indicate that ground water might flow between basins through permeable carbonate rocks in the mountains of west-central Utah, and water might flow from recharge areas in the mountains to local basins through permeable carbonate rocks bordering the northeastern part of the aquifer system. Although extrusive igneous rocks (primarily basalt) can be permeable in local areas, most other types of consolidated rock are not sufficiently permeable to transmit large volumes of water, and bedrock generally forms a relatively impermeable boundary to the Basin and Range aquifers.

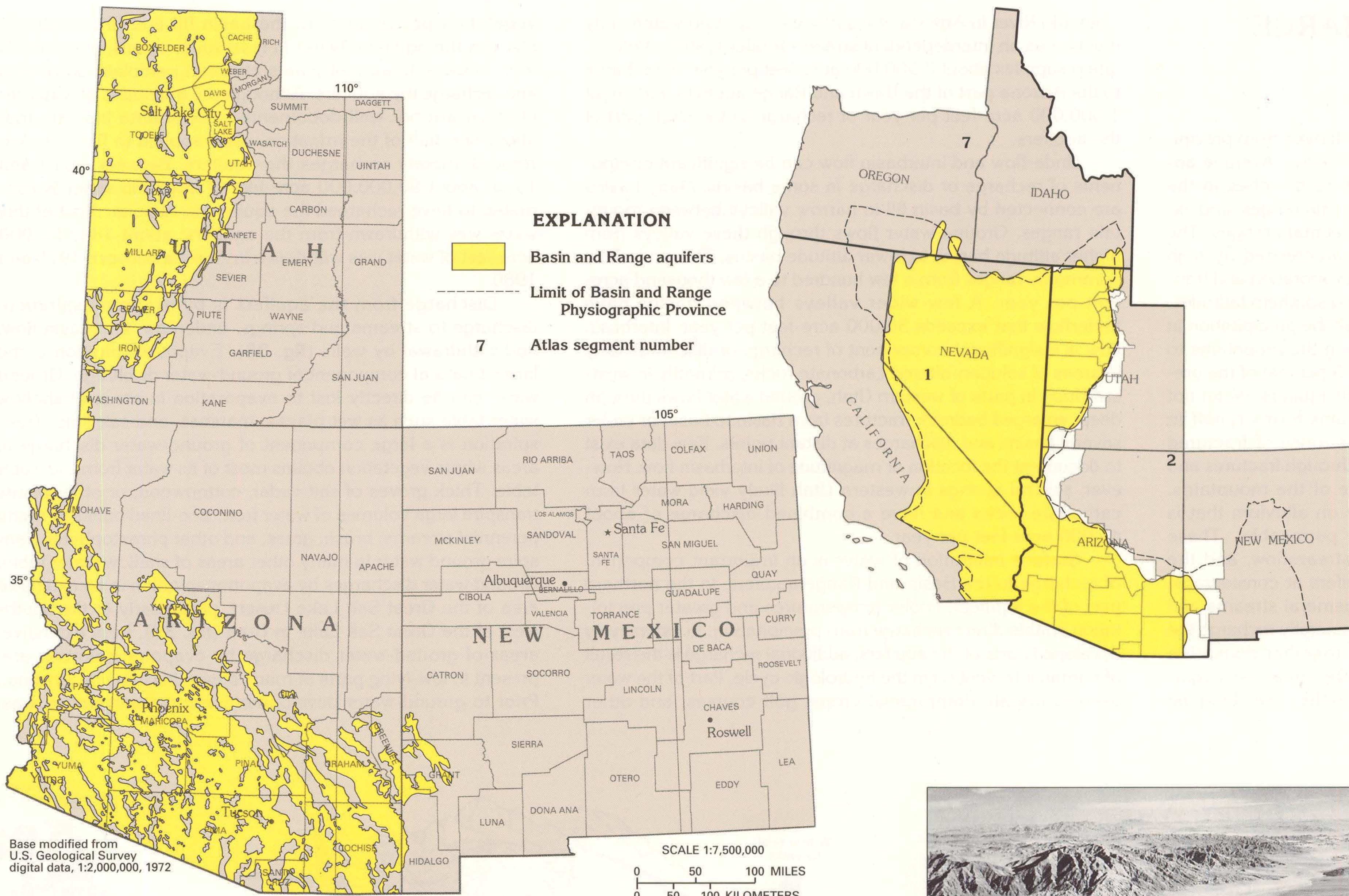


Figure 31. The Basin and Range aquifers extend through parts of seven States and are the principal sources of water in much of this desert area.

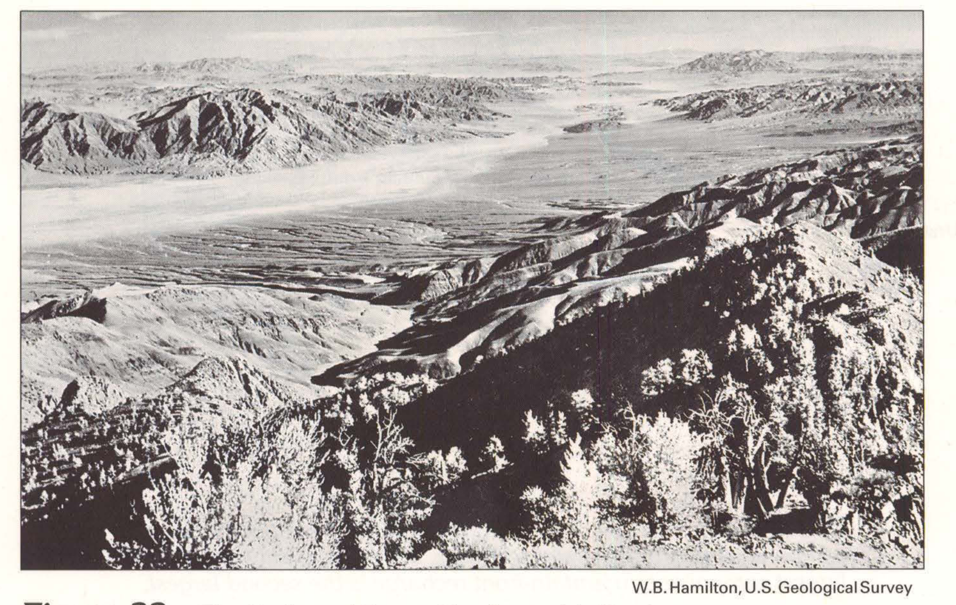


Figure 32. The Basin and Range Physiographic Province is a vast arid region that consists of broad alluvial basins in downfaulted blocks of the Earth's crust bordered by mountain ranges formed from upfaulted blocks of crust.

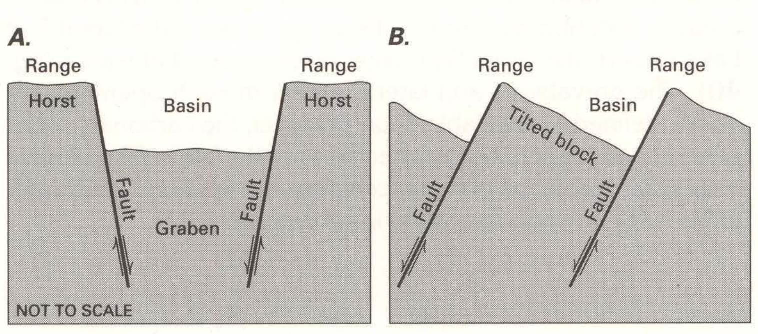


Figure 33. The alternating basins and ranges that characterize the topography of the area were formed during the past 17 million years by the gradual movement along faults. A, horst and graben blocks of the Earth's crust. B, tilting of blocks of the Earth's crust. The arrows indicate the relative direction of movement of rocks on either side of the faults.

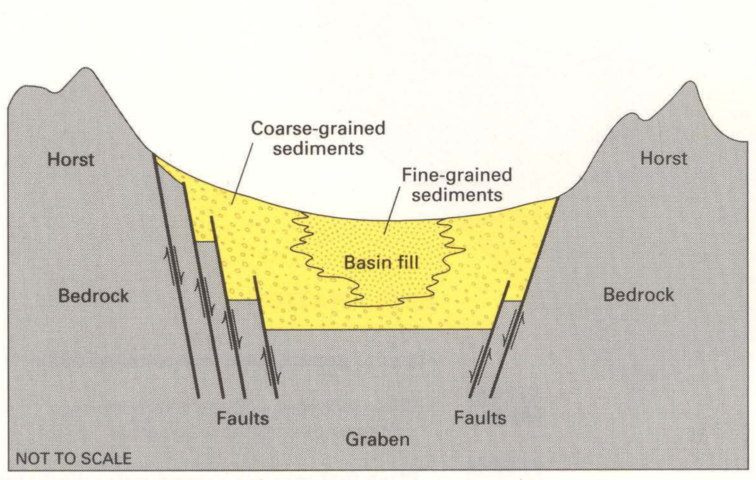
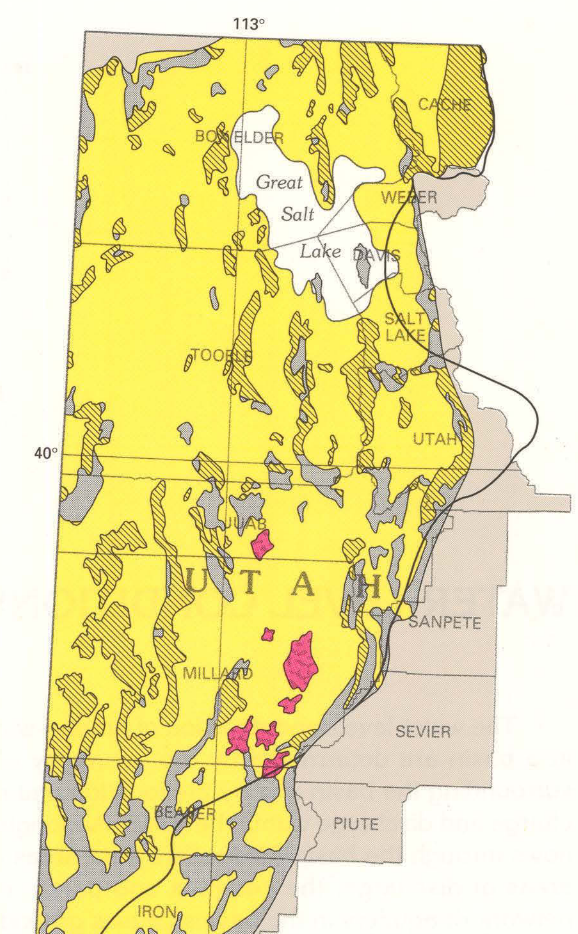


Figure 34. Basin fill is located between bedrock mountain blocks and contains fine-grained sediments near the center of the basin. Coarse-grained sediments were deposited near the basin margins, primarily as alluvial fans.



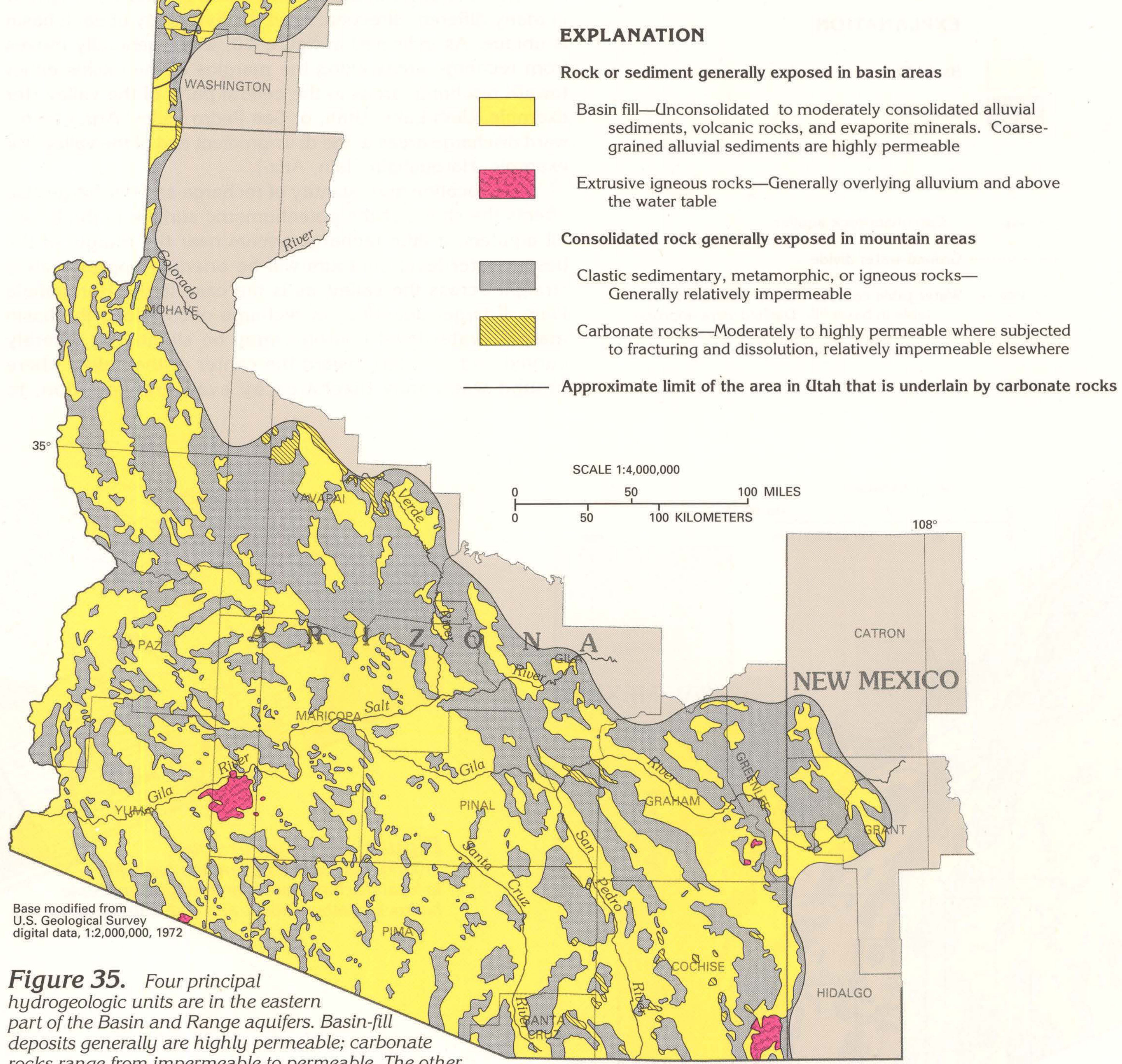


Figure 35. Four principal hydrogeologic units are in the eastern part of the Basin and Range aquifers. Basin-fill deposits generally are highly permeable; carbonate rocks range from impermeable to permeable. The other two units mapped here are relatively impermeable.

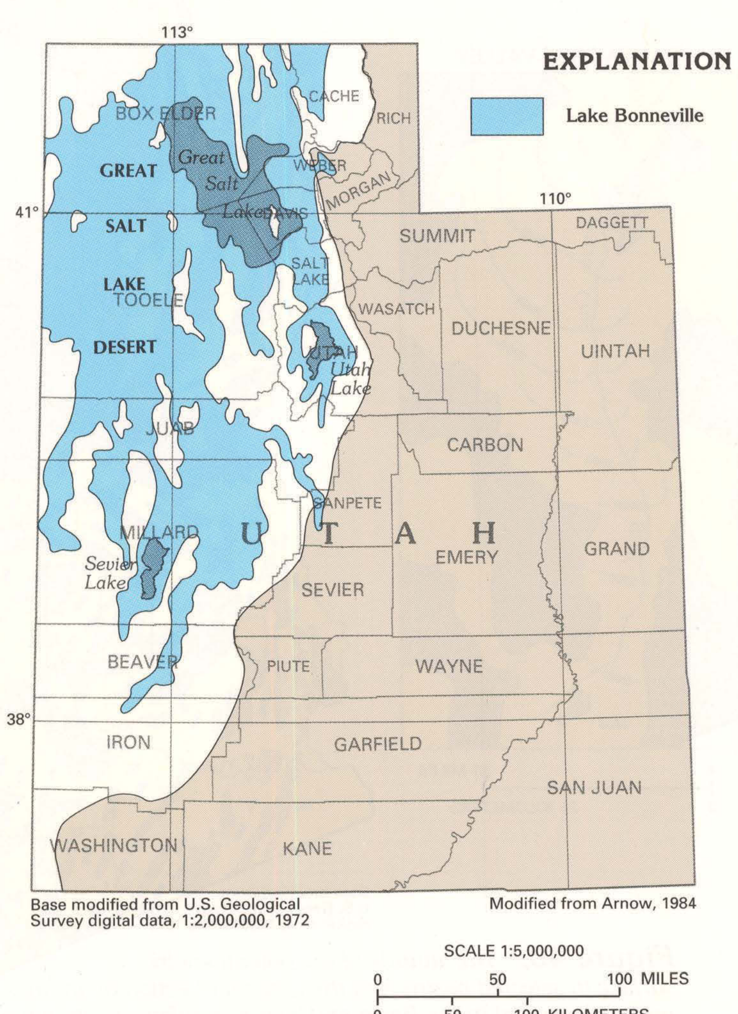


Figure 36. The Great Salt Lake, Utah Lake, and Sevier Lake are the remnants of a much larger lake (Lake Bonneville) that covered western Utah 11,000 to 26,000 years ago.

RECHARGE AND DISCHARGE

Recharge to the aquifers primarily is derived from precipitation in the mountains surrounding the basins. Average annual precipitation ranges from about 4 to 8 inches in the basins, is 16 inches or more in most mountain ranges, and exceeds 30 inches in a few of the higher mountain ranges. The generally arid climate of the area is characterized by high summer temperatures and large rates of evaporation and transpiration, particularly at lower altitudes and southern latitudes. These climatic conditions cause almost all the precipitation in the basins and most of the precipitation in the mountains to be lost to evapotranspiration. Only about 5 percent of the precipitation that falls recharges the basin-fill aquifers. Water not lost to evapotranspiration in the mountains forms runoff in streams or infiltrates the soil and upper zones of fractured bedrock, where it may flow to springs or through fractures and discharge into the basin fill at the base of the mountains. Larger streams in the basins often flow on alluvium that is unconsolidated, well sorted, and highly permeable. These sediments enable rapid infiltration of streamflow, and the streams may recharge the basin-fill aquifers at considerable distance from the mountains. Small ephemeral streams and water flowing through fractured bedrock generally recharge the aquifers near the mountain fronts and together constitute mountain-front recharge (**fig. 37**). When the stream and aquifer are in direct hydraulic connection, as is the case along the Colorado River in Arizona, the surface and ground waters may function as an interdependent stream-aquifer system. Precipitation supplies about 2,500,000 acre-feet per year of recharge to the Arizona part of the Basin and Range aquifers and about 1,500,000 acre-feet per year of recharge to the Utah part of the aquifers.

Underflow and interbasin flow can be significant components of recharge or discharge in some basins. Many basins are connected by basin fill in narrow valleys between mountain ranges. Ground water flows through these valleys from higher altitude basins to lower altitude basins. This underflow commonly ranges from a few hundred to a few thousand acre-feet per year. A few wider valleys between basins have underflow that exceeds 30,000 acre-feet per year. Interbasin flow is a significant component of recharge or discharge only in areas of solution-altered carbonate rocks, primarily in western Utah. In parts of western Utah, ground water flows through deep, enlarged bedrock fractures from basin to basin or under several basins and discharges at distant points. Few data exist to document the location or magnitude of interbasin flow. However, several springs in western Utah likely yield water from carbonate rocks and have a combined discharge of about 45,000 acre-feet per year.

Surface infiltration of water is an important component of recharge to the Basin and Range aquifers. In the northern part of the aquifers, cooler temperatures and greater precipitation enable direct recharge from precipitation. In extensively developed parts of the aquifers, additional recharge is the result of human intervention in the hydrologic cycle. Part of the water used to irrigate commercial crops, golf courses, and other vegetation percolates into the basin fill and ultimately recharges the aquifers. Water in reservoirs, canals, and outfalls from sewage-treatment plants also can percolate downward and recharge the aquifers. Although the quantities of water recharged are not well documented, some data indicate that about one-half of the irrigation water applied to fields in Arizona ultimately recharges the aquifers. Between 1915 and 1980, about 90,000,000 acre-feet of irrigation water is estimated to have recharged the aquifers in Arizona. Most of this water was withdrawn from the aquifers; about 184,000,000 acre-feet of water was withdrawn in Arizona between 1915 and 1980.

Discharge from the aquifers is by evapotranspiration, discharge to streams and springs, underflow, interbasin flow, and withdrawal by wells (**fig. 38**). Evapotranspiration is the largest natural component of ground-water discharge. Ground water can be directly lost to evaporation in areas of shallow water table such as wet playas, marshes, and salt flats. Transpiration is a large component of ground-water discharge in areas where vegetation obtains most of its water from the water table. Thick groves of salt cedar, cottonwood, or other plants transpire large volumes of water from tree-lined banks of many perennial streams; brush, grass, and other phreatophytes transpire ground water in many other areas of shallow water table. Ground water discharges by evapotranspiration in an extensive area of the Great Salt Lake Desert, particularly to the southwest of the Great Salt Lake in Utah (**fig. 39**). Less extensive areas of ground-water discharge by evapotranspiration are present in low-lying parts of many basins in Utah and Arizona. Prior to ground-water development, evapotranspiration was about 1,300,000 acre-feet per year along the lower Colorado River in Arizona, and about 700,000 acre-feet per year along the Gila River. Natural evapotranspiration can decrease when ground-water withdrawal lowers the water table under the phreatophytes.

Ground water discharges to streams or lakes in areas where the water level in the aquifer is higher than the level of the stream or lake bed (**fig. 38**). This situation can occur where a constriction in the width or thickness of the aquifer forces ground water to the surface, or where ground water flows toward a stream from aquifers of higher altitude on either side of the stream. In arid climates, perennial flow in streams that cross many miles of basin fill usually is maintained by ground-water discharge from underlying aquifers. Prior to ground-water development, the Gila River and its principal tributaries, the Salt, Verde, and San Pedro Rivers were perennial. These rivers and the Colorado River, which is still perennial, received ground-water discharge from aquifers in most of the basins they crossed. The banks of many perennial streams in Arizona are covered by trees and other vegetation that obtain water from the aquifers. Most perennial streams in Utah are located along the eastern margin of the aquifers and drain the Wasatch Range and Wasatch Plateau or other high mountains to the east; streamflow generally provides recharge to aquifers underlying the alluvial fans before the streams enter Great Salt Lake, Utah Lake, or other lakes.

Underflow and interbasin flow are additional components of ground-water discharge that can be large in some basins. The factors that relate to underflow and interbasin recharge discussed above also apply to discharge.

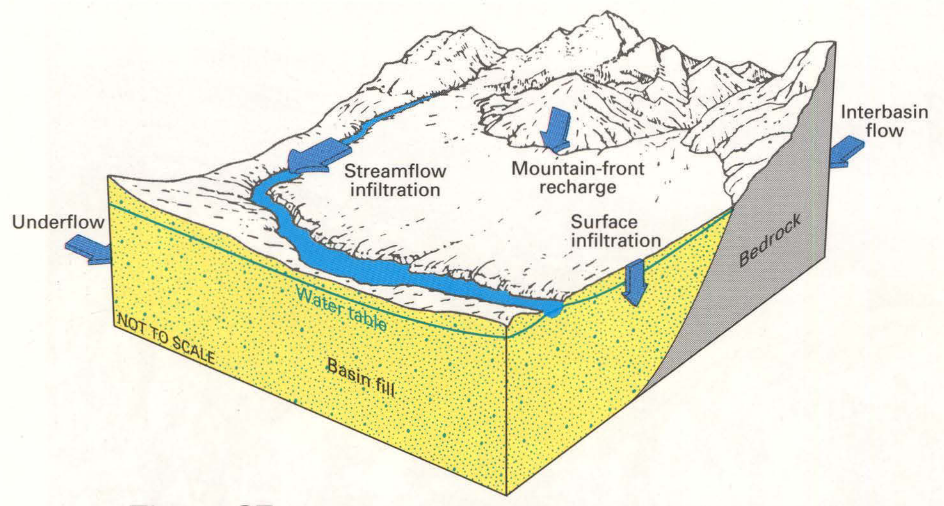


Figure 37. The Basin and Range aquifers have five principal components of ground-water recharge. Streamflow infiltration is the largest component; mountain-front recharge is the second largest.

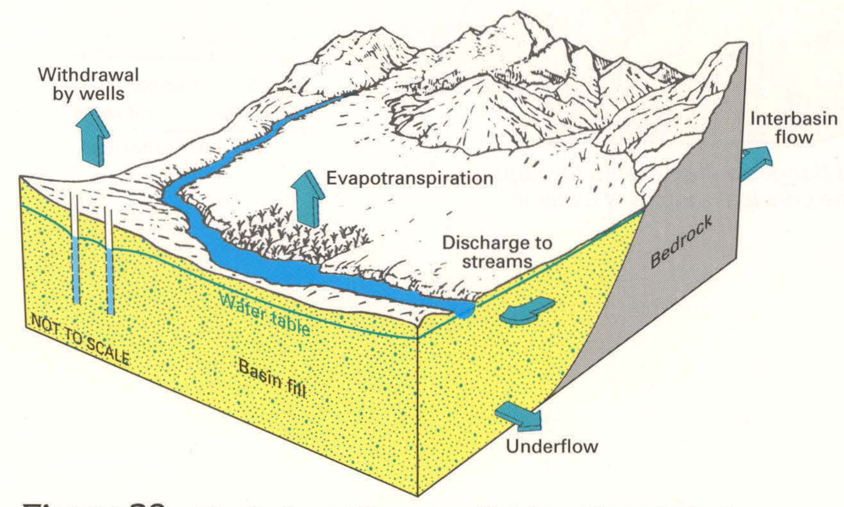


Figure 38. The Basin and Range aquifers have five principal components of ground-water discharge. Ground-water withdrawal from wells is the largest component in Arizona; evapotranspiration is the largest component in Utah.

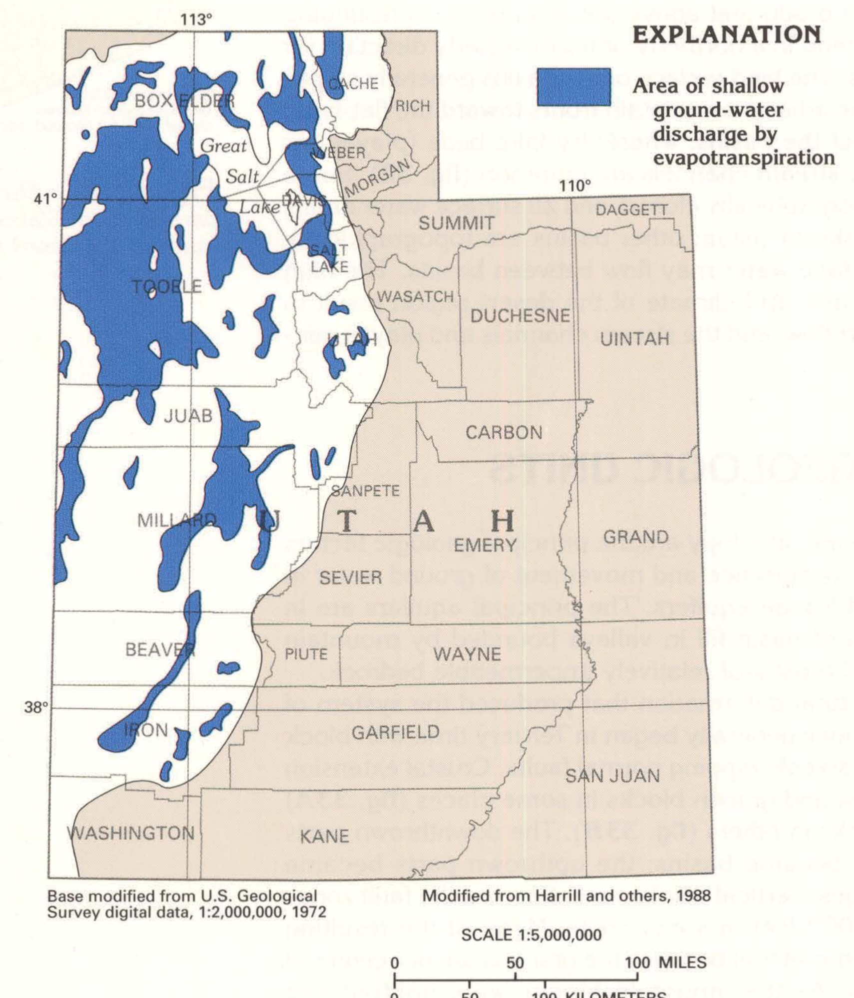


Figure 39. Ground water is discharged by evapotranspiration in large areas of western Utah, particularly to the southwest of the Great Salt Lake.

WATER-LEVEL CONDITIONS

The water levels and direction of ground-water movement in a basin are determined by the geometry of the bedrock surrounding the basin and by the location and quantity of recharge and discharge within the basin. Although ground water flows through the basin-fill aquifers from areas of recharge to areas of discharge, the complex and partly interconnected network of aquifers in the basins causes ground water to flow in many different directions, and the hydrology of each basin is unique. As indicated in **figure 40**, water generally moves from recharge areas along the margins of the basins either toward discharge areas in the central parts of the valley (for example, Utah Lake, Utah, or San Pedro Valley, Ariz.), or toward discharge areas at the downgradient end of the valley (for example, Harquahala Plain, Ariz.).

The location and quantity of recharge and discharge also affects the shape of the potentiometric surface in the basin-fill aquifers. If little recharge occurs near the margin of the basin, water-level contours will be oriented approximately straight across the valley, as is the case in the Harquahala Plain. If larger quantities of recharge occur along the basin margin, water-level contours may be slightly to severely curved and oriented toward the center of the valley where ground water may discharge by evapotranspiration, to streams, or from wells. Before extensive ground-water withdrawals, the recharge along the margin of the San Pedro Valley in Arizona was larger than the rate of underflow down the valley, and the excess water was discharged to the river, as indicated by a marked curvature of the water-level contours (**fig. 40**). In valleys where basin-margin recharge is moderate and discharge occurs as a varied combination of evapotranspiration, seepage to streams, withdrawal from wells, or underflow, the water-level contours may be of more varied shape, as in the Santa Cruz Valley of Arizona. Lakes and playas are the sole discharge areas in a few valleys. A closed ground-water basin can result if all ground-water recharge and discharge occurs within the valley, and little or no water moves beyond the valley.

Ground water also flows through fractures and solution openings (primarily in carbonate rocks) that underlie and border the basin fill in parts of western Utah (**fig. 41**). If the fractures and solution openings are numerous and extensive enough, ground water may flow through the permeable bedrock from basin to basin, or beneath basins from recharge areas in distant mountains to discharge areas in the Great Salt Lake Desert, the Great Salt Lake, or the Sevier Lake area (**fig. 40**). The prevalence and lateral extent of such openings are poorly defined by available data. However, the carbonate rocks generally are much less permeable than the basin fill and have only scattered zones of enhanced permeability that yield water to bedrock springs or enable interbasin flow.

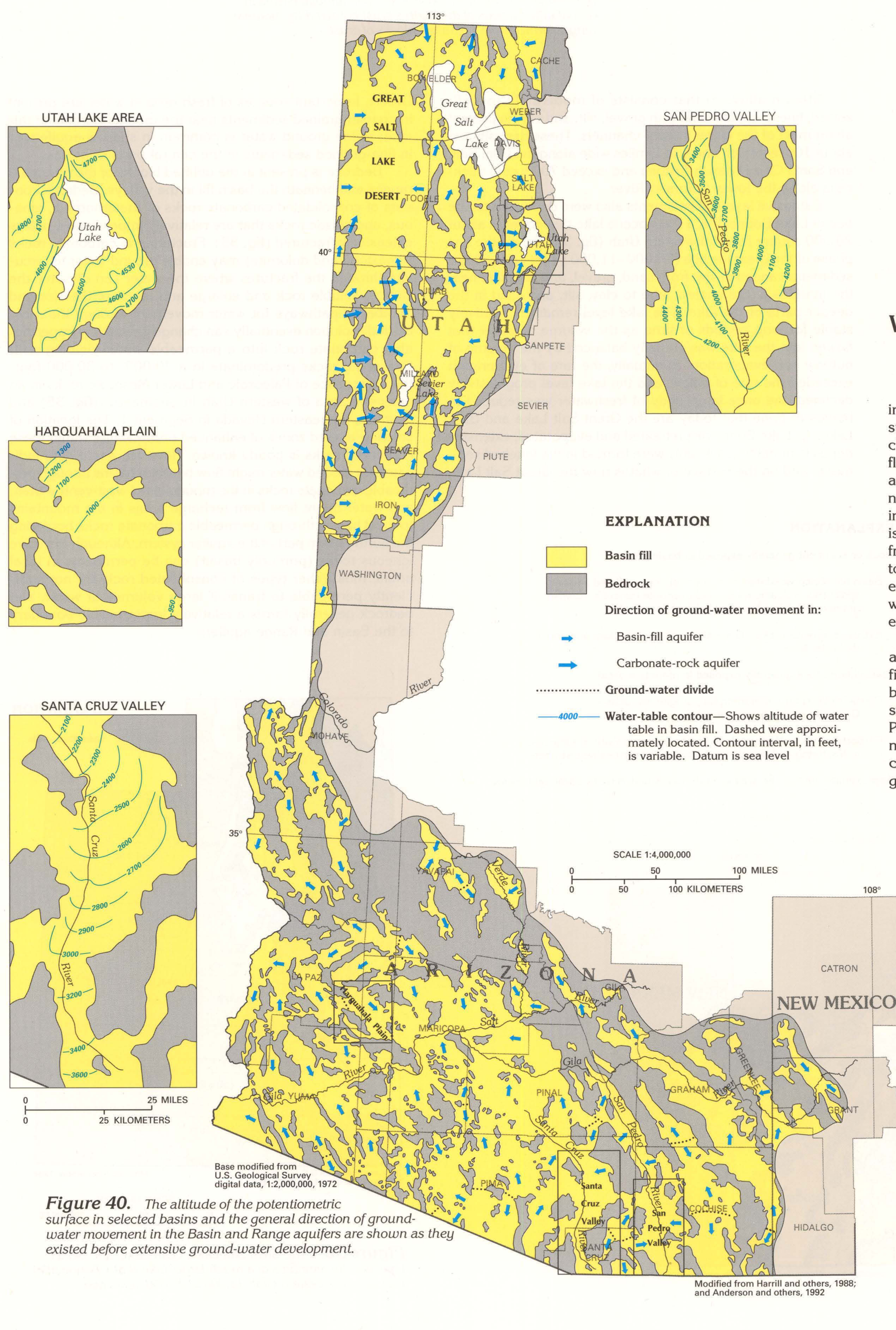


Figure 40. The altitude of the potentiometric surface in selected basins and the general direction of ground-water movement in the Basin and Range aquifers are shown as they existed before extensive ground-water development.

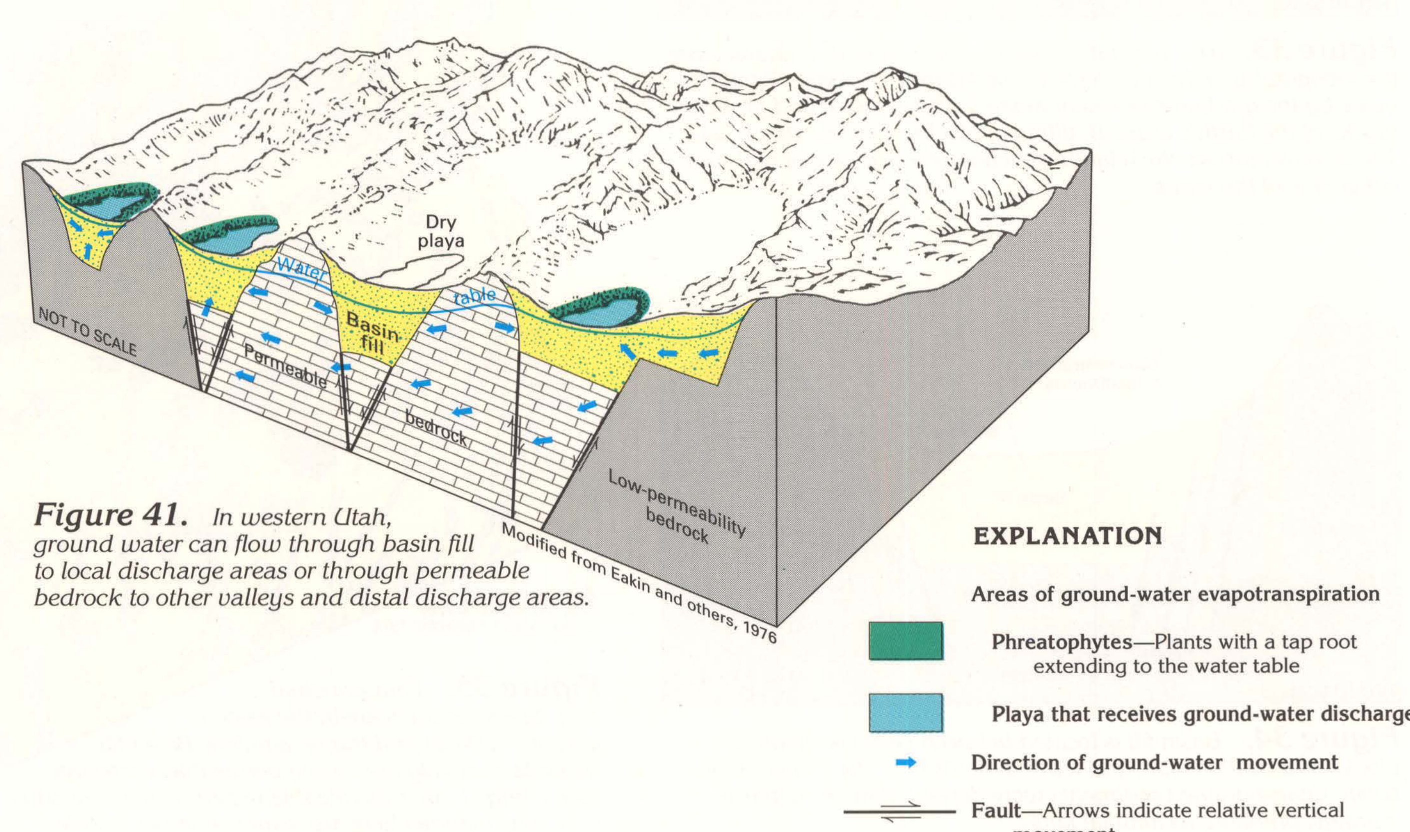


Figure 41. In western Utah, ground water can flow through basin fill to local discharge areas or through permeable bedrock to other valleys and distal discharge areas.

WATER-LEVEL CONDITIONS — Continued

Before extensive ground-water development, about 900,000,000 acre-feet of ground water was in storage in the upper 1,200 feet of the basin fill in the Arizona part of the aquifer system. Incomplete data from Utah indicate that about 800,000,000 acre-feet of water may have been in storage in the upper 1,200 feet of basin fill in this area. The volume of ground water in storage in basins in Arizona and Utah (**fig. 42**) is estimated to range from less than 5,000,000 acre-feet in areas along the northern margin of the aquifer system in Arizona to more than 70,000,000 acre-feet in three large areas of western Utah. The volumes of ground water in storage in the basins in Arizona and Utah are only approximations because little is known about the subsurface extent, thickness, and water-yielding character of the basin-fill sediments in many areas. Not all the ground water in storage is potable. Water in the deeper parts of most basins contains larger concentrations of dissolved solids than water in the shallower parts of the basin. Most of the ground water under the Great Salt Lake Desert and Sevier Lake area is saline or brine.

The volume of ground water in storage in the basin-fill aquifers is large in comparison to the annual rate of recharge to the basins. Most basins contain between 200 and 2,000 times as much water in storage as they receive from annual recharge. In basins that contain relatively small volumes of water in storage and have relatively large recharge (such as the basins near the Colorado River in Arizona), only about 5 times as much water may be in storage as received from recharge. In basins that have large volumes of water in storage and minimal recharge, about 14,000 times as much water may be in storage as received from recharge.

Large rates of ground-water withdrawal can cause areally extensive water-level declines, decrease natural discharge, and deplete streamflow. Persistent large rates of withdrawal can cause water-level declines to extend throughout a basin. As water levels are lowered, the natural ground-water gradients may be decreased or reversed in some areas, and ground-water flow toward areas of natural discharge may decrease; reversal of ground-water gradients may decrease or eliminate natural discharge. Discharge to streamflow and evapotranspiration also may decrease or cease if the shallow water table declines below the level of the streambed or below the root zone of phreatophytes. Ground-water withdrawal can cause depletion of streamflow by lowering the water table near the stream. The greater depth to ground water enables more streamflow to infiltrate, thereby inducing additional ground-water recharge.

Water-level declines are the largest and most widespread effect of ground-water withdrawal. Major agricultural regions have undergone the largest water-level declines, which exceeded 450 feet through 1980 in some areas in Arizona and exceeded 70 feet in parts of Utah (**fig. 43**). Arizona generally has had larger water-level declines than Utah, primarily because of the larger rates of withdrawal in Arizona. Water-level declines have been relatively small in basins along most of the eastern margin of the aquifer system in Utah, where withdrawal is small, and along the Colorado River, where induced recharge from streamflow and lakes has moderated declines. Along parts of the Colorado River, water levels rose following agricultural development because of infiltration of irrigation water supplied from surface-water diversions. In other areas where development has been moderate and the aquifers receive smaller rates of recharge, long-term water-level declines have ranged from 50 to 200 feet. In basins where no perennial streamflow occurs and recharge to the basin is small, even minor withdrawal can significantly decrease the volume of ground water in storage. The hydrology of these basins is relatively simple. Pumped water is derived from storage in the aquifer, and ground-water levels decline when pumping occurs. The largest water-level declines generally have occurred in basins of this type.

Differing rates of water-level decline occur in the aquifers. The rate of water-level decline in a few observation wells is shown in the hydrographs of **figure 43**. Water levels near pumping wells undergo greater long-term decline than those in distant parts of the basin, and withdrawal to supply crops causes large seasonal fluctuations in water level near the pumped wells. Water levels in most wells in Utah and in the less developed parts of Arizona generally declined at average rates of less than 5 feet per year. Average rates of decline exceeded 20 feet per year in a few extensively developed basins in south-central Arizona. The rate of water-level decline generally was larger for 10 to 20 years prior to 1970 than it has been since 1970. The reduction in the rate of decline is the result of decreased withdrawal, possibly caused by a combination of decreased irrigated acreage, greater irrigation efficiency, conversion to crops that require less water, larger amounts of precipitation, or greater availability of surface water for irrigation. In Utah, the last two factors are the primary causes of decreased rates of decline or water-level rises during 1982–86.

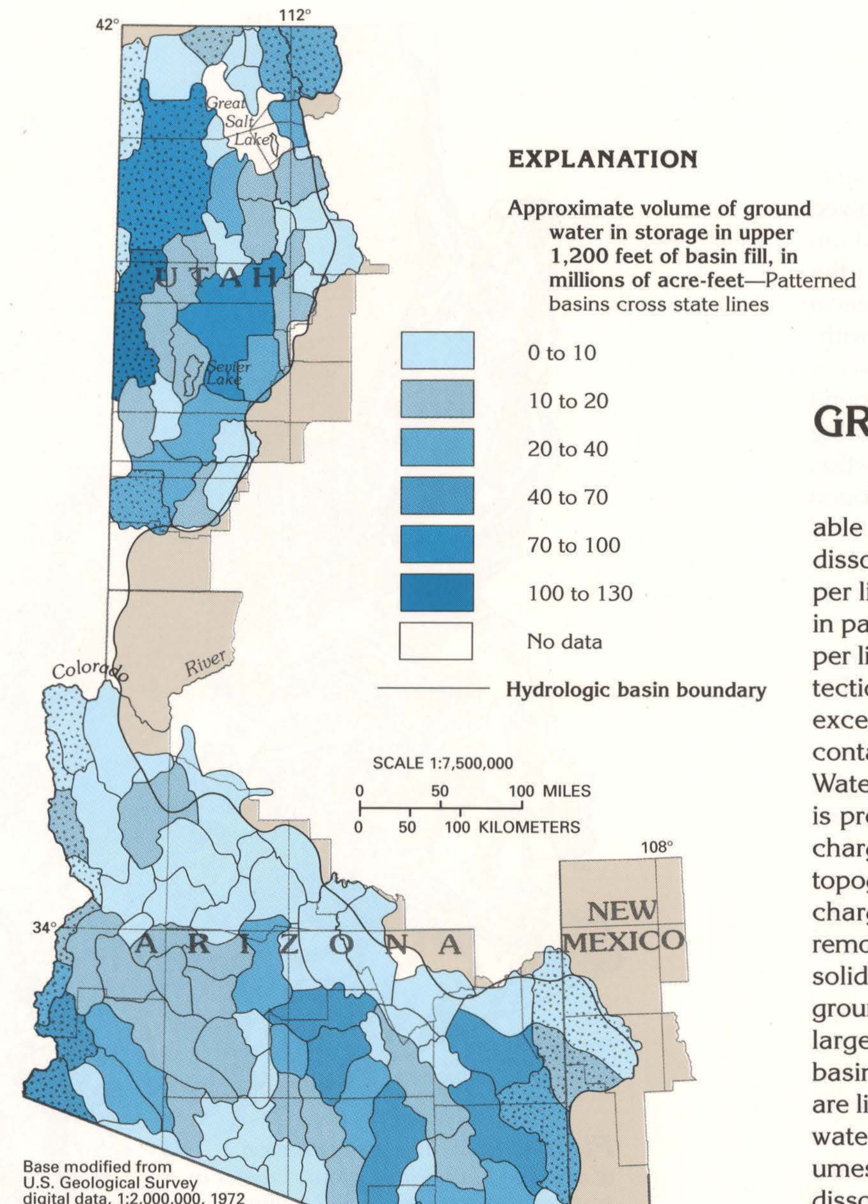


Figure 42. *Basin-fill aquifers within these geographic areas contain large volumes of ground water in storage. However, not all this water is potable.*

GROUND-WATER QUALITY

Ground water in the basin-fill aquifers generally is of suitable chemical quality for most uses; most ground water has a dissolved-solids concentration of less than 1,000 milligrams per liter. However, the dissolved-solids concentration of water in parts of some basins can be as large as 300,000 milligrams per liter (**fig. 44**). By comparison, the U.S. Environmental Protection Agency recommends that water for public supply not exceed 500 milligrams per liter of dissolved solids; seawater contains about 35,000 milligrams per liter of dissolved solids. Water that has small dissolved-solids concentration generally is present near the margins of the basins (**fig. 44**), where recharge from the nearby mountains enters the aquifers. In the topographically low parts of some basins, ground water is discharged by evaporation and transpiration. These processes remove some water from the aquifer and concentrate dissolved solids in the remaining water. If the volume of underflow or ground-water discharge to streamflow in the basin is relatively large, the accumulated salts are continually flushed from the basin, and the dissolved-solids concentrations in the aquifers are little affected. In basins that have less underflow or ground-water discharge to streamflow, or in basins where large volumes of irrigation water evaporate, salts may accumulate, and dissolved-solids concentrations in the ground water may range from 1,000 to more than 3,000 milligrams per liter. In basins that have no discharge by underflow or streamflow, salts can accumulate over long periods of time in the fine-grained sediments near the center of the basin, or can form extensive surface deposits of salt, such as the salt flats of the Great Salt Lake Desert in western Utah. Dissolved-solids concentrations in ground water near surface or subsurface deposits of saline minerals can be very large; concentrations commonly exceed 200,000 milligrams per liter in parts of western Utah.

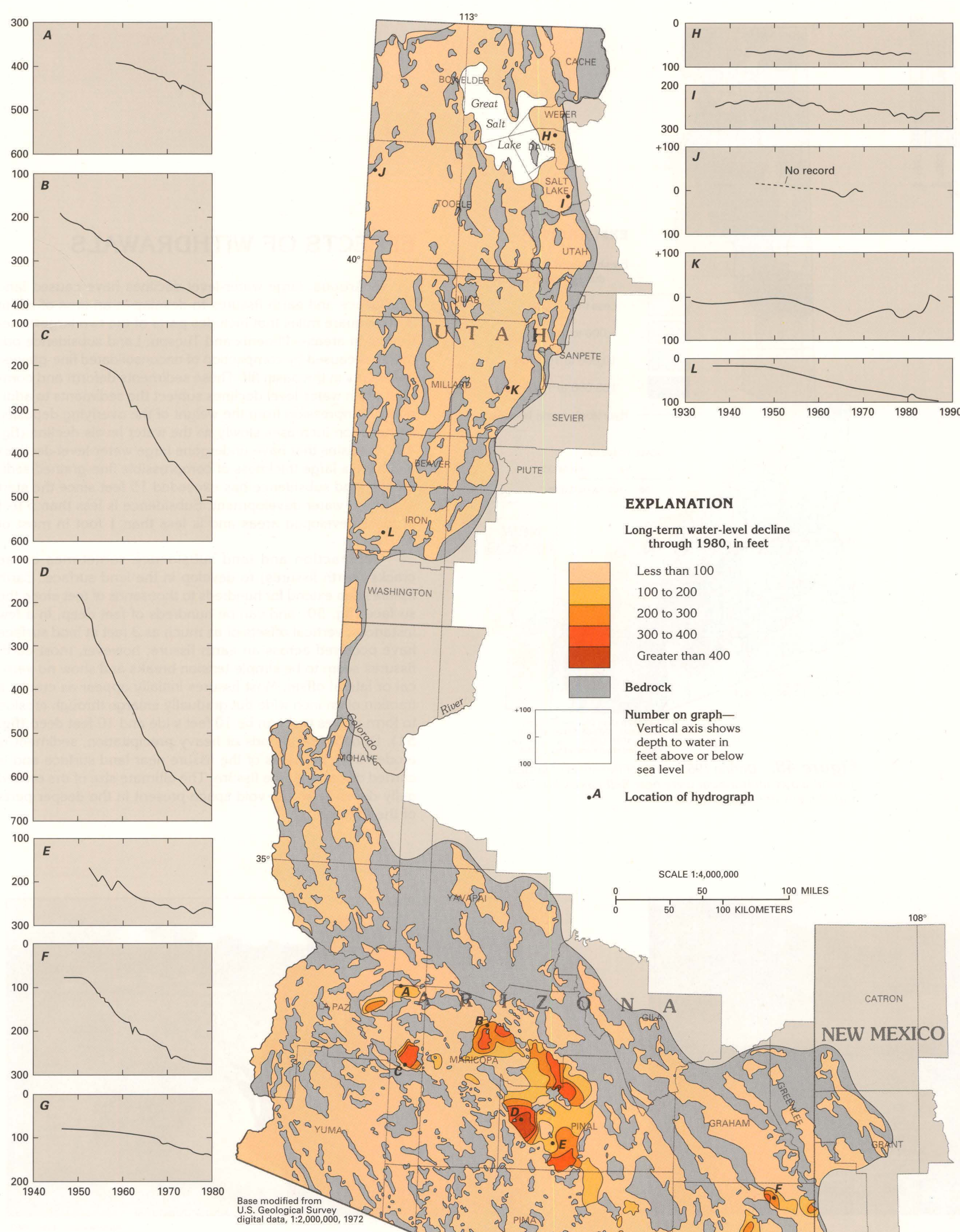


Figure 43. *Water-level declines in Utah generally are smaller and less extensive than the water-level declines in many basins in Arizona. The water-level hydrographs show that the rate and magnitude of water-level declines vary greatly at different places in the aquifers.*

Modified from U.S. Geological Survey, 1985; and Anderson and others, 1992

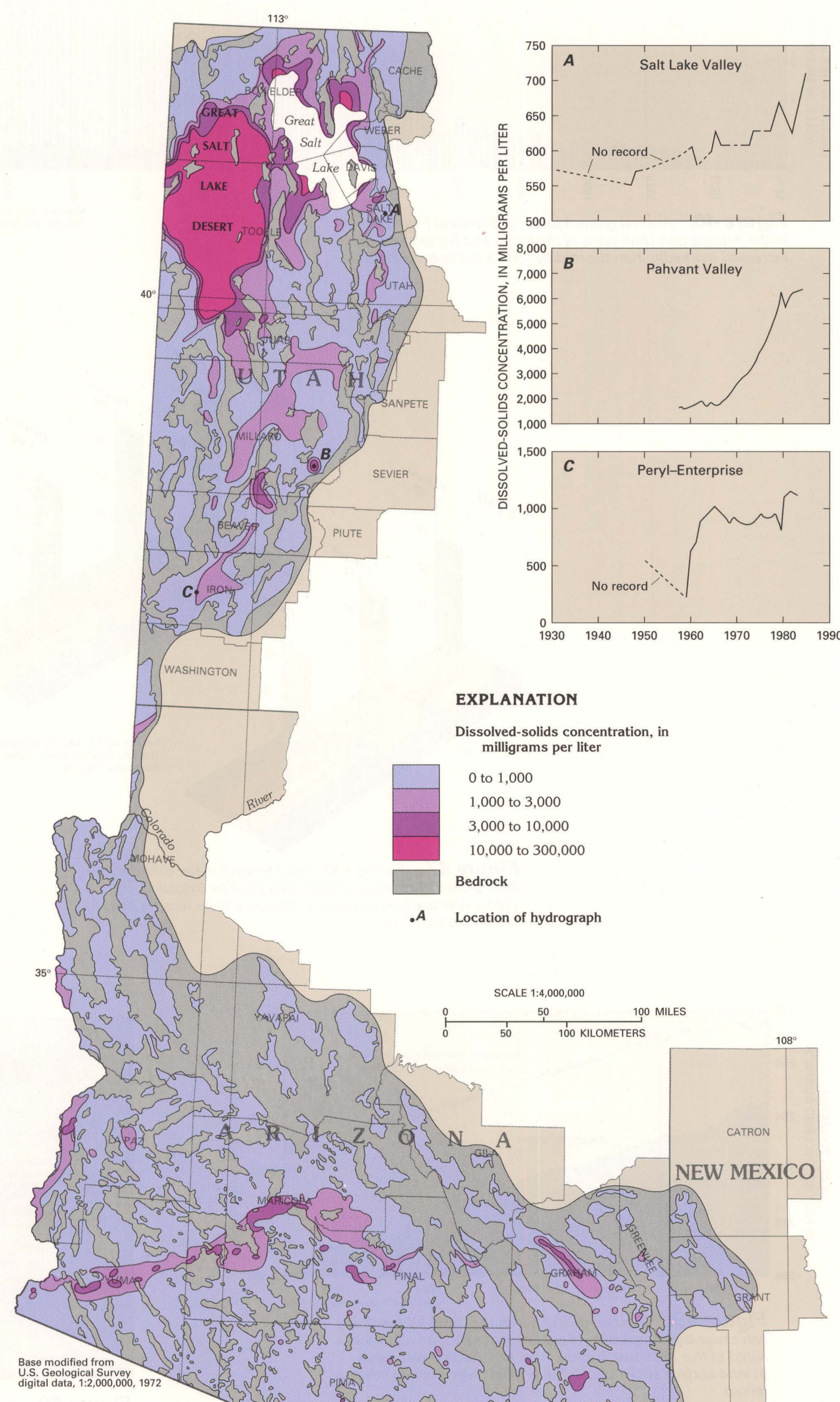


Figure 44. *The dissolved-solids concentrations in water in the basin-fill aquifers generally are less than 1,000 milligrams per liter but exceed 10,000 milligrams per liter in large areas of Utah in the Great Salt Lake Desert and near Great Salt Lake.*

GROUND-WATER QUALITY—
Continued

Ground-water quality can be degraded by water use. Some of the irrigation water applied to fields carries dissolved salts, which have been concentrated by evapotranspiration, through the soil and to the water table, thus increasing the salinity of the ground water. If the irrigated fields are near a pumping well, the degraded ground water may again be withdrawn and applied to the field, thereby creating a cycle of water reuse that can progressively increase the salinity of the ground water. Other factors that may degrade ground-water quality include leaching of salts from the unsaturated zone, migration of mineralized ground water into the cone of depression around a well, and application of mineralized irrigation water obtained from other sources. Most irrigated areas underlain by a shallow water table have undergone water-quality degradation. In areas of greater depth to the water table, particularly in parts of Arizona, water-quality degradation might be minimal because recharge from irrigation might not have percolated to the water table.

Ground water in the basin-fill aquifers is of varied chemical composition. Near the recharge areas along the margins of most basins, the water generally contains a preponderance of calcium or magnesium cations and bicarbonate or sulfate anions, and thus is classified as a calcium magnesium, bicarbonate sulfate water type. Near the central part of many basins and near the Great Salt Lake Desert and Great Salt Lake in Utah, the water is a sodium chloride type.

Concentrations of most dissolved constituents in ground water do not exceed U.S. Environmental Protection Agency Primary or Secondary Drinking Water Regulations. However, concentrations of dissolved fluoride are as much as 5 times greater than the Maximum Contaminant Level (4 milligrams per liter) for drinking-water supplies in parts of some basins in Arizona. Dissolved fluoride concentrations of more than 20 milligrams per liter are present in a few areas of Arizona (**fig. 45**). Smaller concentrations are measured in most parts of Arizona; concentrations in Utah generally are less than 2 milligrams per liter.

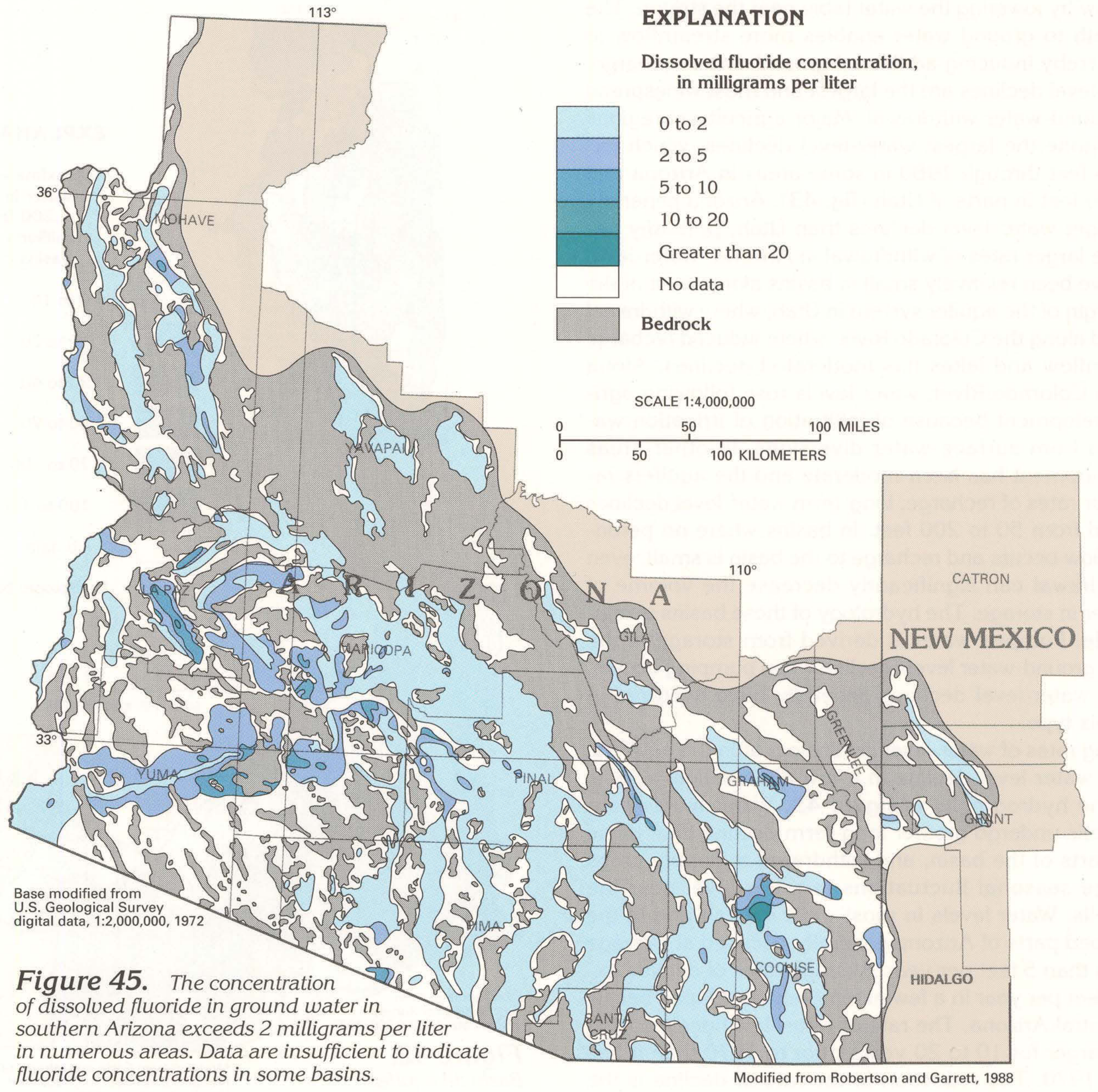


Figure 45. *The concentration of dissolved fluoride in ground water in southern Arizona exceeds 2 milligrams per liter in numerous areas. Data are insufficient to indicate fluoride concentrations in some basins.*

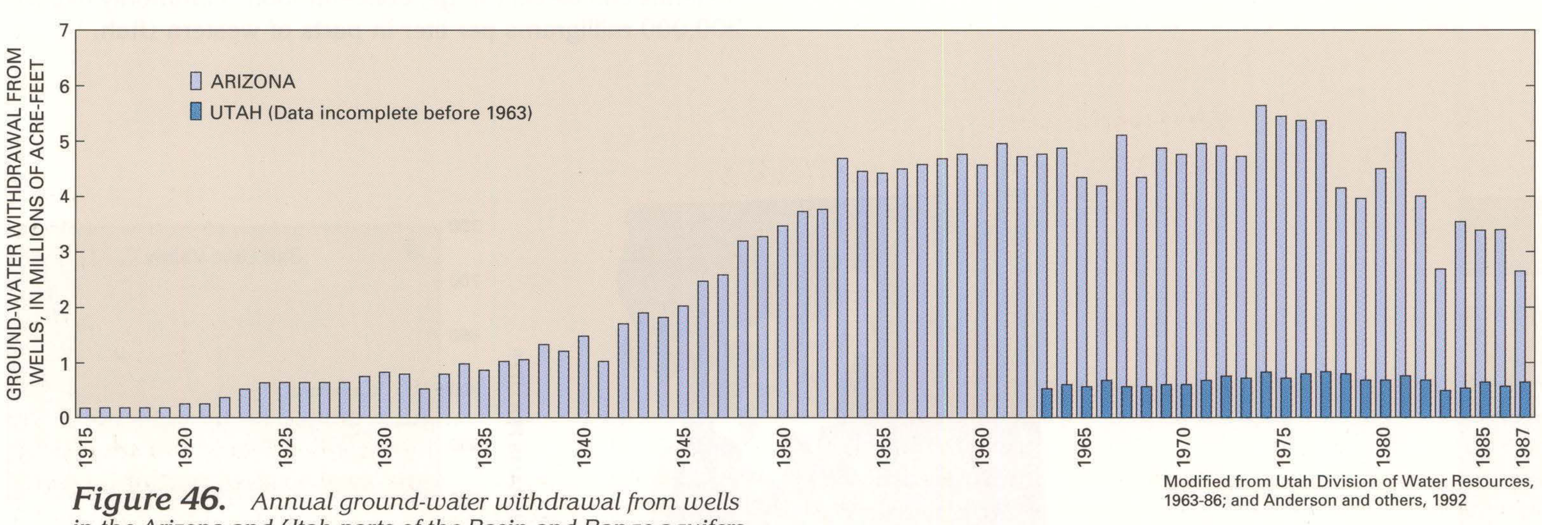


Figure 46. *Annual ground-water withdrawal from wells in the Arizona and Utah parts of the Basin and Range aquifers increased markedly from the middle 1940's to the early 1950's.*

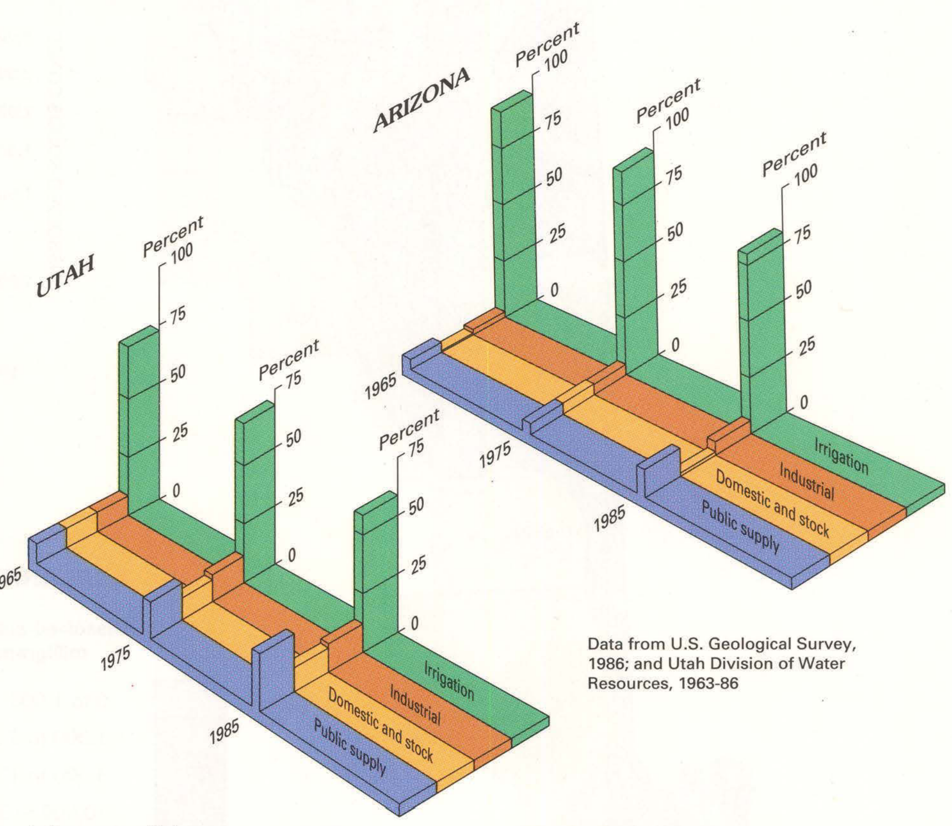


Figure 47. *During 1965, most ground water withdrawn from wells in Arizona and Utah was used for irrigation. By 1985 water use had changed in response to increasing need for water in urban areas.*

FRESH GROUND-WATER WITHDRAWALS

Ground-water withdrawal from wells is the largest component of discharge from the Basin and Range aquifers. In Arizona, for example, about 184,000,000 acre-feet of ground water was withdrawn from the basin-fill aquifers between 1915 and 1980 (**fig. 46**). In some extensively developed basins in Arizona, the rate of withdrawal is about 200 times the rate of recharge. About half the water withdrawn is lost to the atmosphere by evapotranspiration; the other half percolates through the soil and eventually returns to the aquifer as recharge. The evapotranspiration loss is called ground-water depletion. The 92,000,000 acre-feet of ground-water depletion in Arizona is about 10 percent of the original 900,000,000 acre-feet of ground water in storage.

Most ground water is withdrawn for irrigation of commercial crops. Withdrawal for irrigation began in the late 1800's when settlers realized that the water needed for their expanding agriculture exceeded the water available from surface-water sources. In Arizona, withdrawal was small until near the start of World War II when advances in pump technology, availability of inexpensive sources of electricity, and wartime demands combined to create a rapid increase in acreage irrigated by wells. Withdrawal has been less than 1,000,000 acre-feet per year in Utah from 1963 through 1987; withdrawal data are incomplete before 1963 (**fig. 46**). In Arizona, withdrawal has undergone a general decline since the late 1970's in response to use of more water-efficient irrigation systems, introduction of crops that use less water, and reduction in acreage irrigated by wells.

An increase in population has caused an increase in ground-water use in urban areas and the conversion of agricultural land to urban use. During 1965, about 93 percent of the ground water withdrawn in Arizona and about 72 percent of the ground water withdrawn in Utah was used for irrigation; by 1985, these percentages had decreased to about 80 percent in Arizona and 58 percent in Utah. During this 20-year period, ground-water withdrawal for public supply increased from less than 5 percent to more than 12 percent in Arizona and from less than 10 percent to more than 22 percent in Utah (**fig. 47**). Some of the largest rates of ground-water withdrawal are near the rapidly expanding metropolitan areas of Salt Lake City and Provo, Utah, and Phoenix and Tucson, Ariz. (**fig. 48**). Irrigation is still a principal water use in these areas, although the populations of Tucson and many smaller communities in Arizona and Utah are increasing and ground water is the principal source of supply.

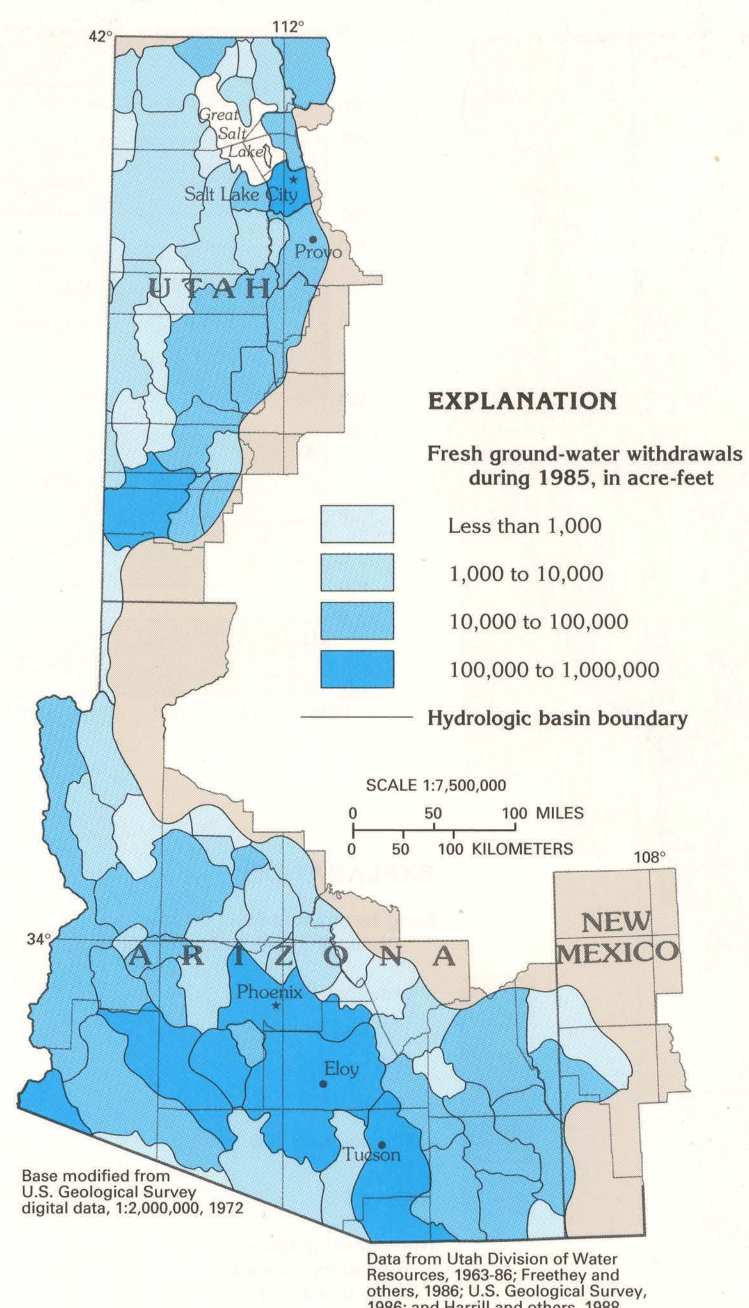


Figure 48. *During 1985, some of the areas of largest ground-water withdrawal were near Salt Lake City, Utah, and Phoenix and Tucson, Ariz.*

EFFECTS OF WITHDRAWALS

In Arizona, large water-level declines have caused land subsidence and earth fissures to develop in an area of about 3,000 square miles that includes parts of the two largest metropolitan areas—Phoenix and Tucson. Land subsidence primarily is caused by compaction of unconsolidated fine-grained sediments in the basin fill. These sediments deform and compact when water-level declines subject the sediments to additional compression from the weight of the overlying deposits; compaction increases slowly as the water levels decline (**fig. 49**). In basins that have undergone large water-level declines and have a large thickness of compressible fine-grained sediments, land subsidence has exceeded 15 feet since the start of ground-water development. Subsidence is less than 5 feet in most developed areas and is less than 1 foot in most of Arizona.

Compaction and land subsidence sometimes cause cracks (earth fissures) to develop in the land surface. Earth fissures can extend for hundreds to thousands of feet along the surface (**fig. 50**) and can be hundreds of feet deep. In a few instances, vertical offsets of as much as 3 feet at land surface have occurred across an earth fissure; however, most earth fissures seem to be simple tension breaks and show no vertical or lateral offset. Most fissures initially appear as cracks a fraction of an inch wide but gradually enlarge through erosion to form gullies that can be 10 feet wide and 10 feet deep (**fig. 51**). During rare periods of heavy precipitation, sediment is eroded from the sides of the fissure near land surface and is carried deeper into the fissure. The ultimate size of the eroded gully depends on the void space present in the deeper parts of the fissure.

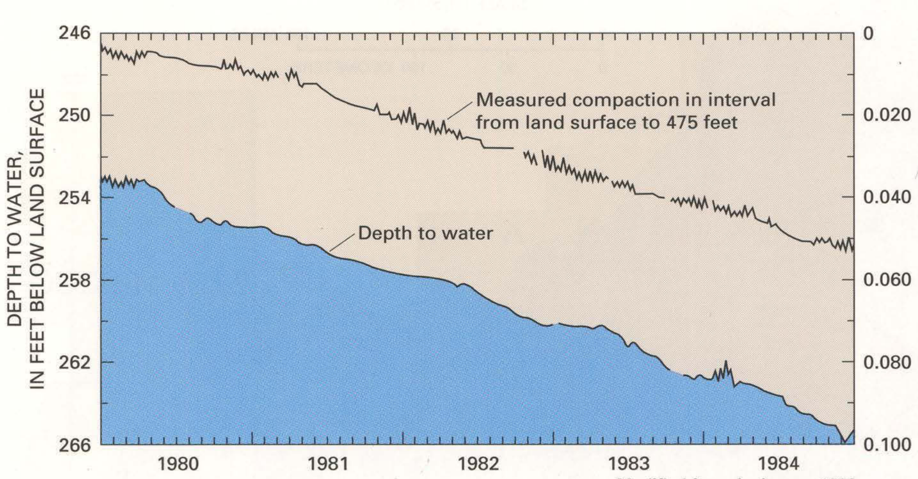
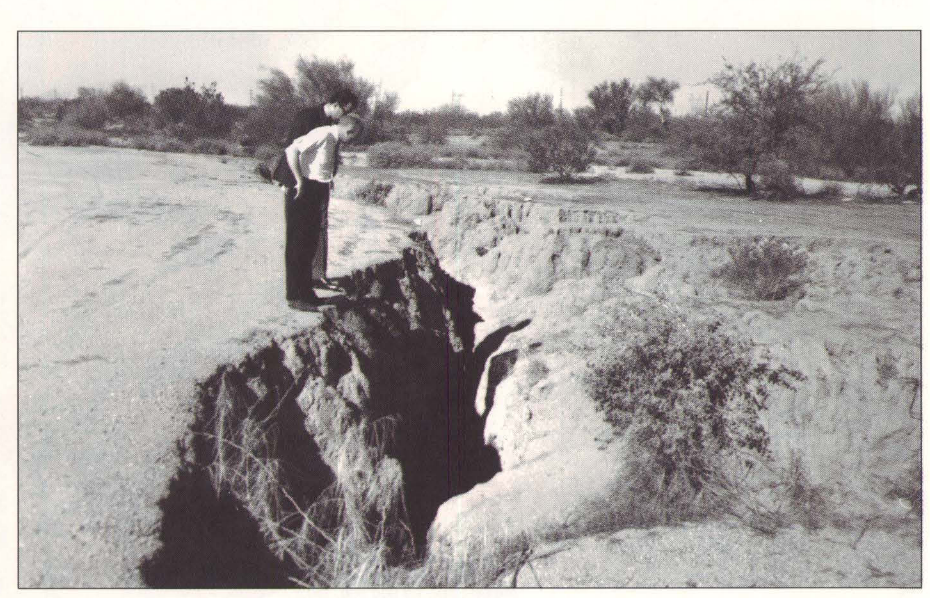


Figure 49. *Water-level declines in the basin-fill aquifers can cause the compaction of aquifer sediments, as was measured in this well near Tucson, Ariz. Compaction can result in land-surface subsidence. The broken lines indicate missing record.*

Anderson and others, 1992

Figure 50. *Earth fissures caused by compaction and subsidence generally are located near the margins of the basins and can extend for many miles, as shown in this photograph near Eloy, Ariz.*



Anderson and others, 1992

Figure 51. *Earth fissures are enlarged by erosion when precipitation washes sediment deeper into the fissure.*

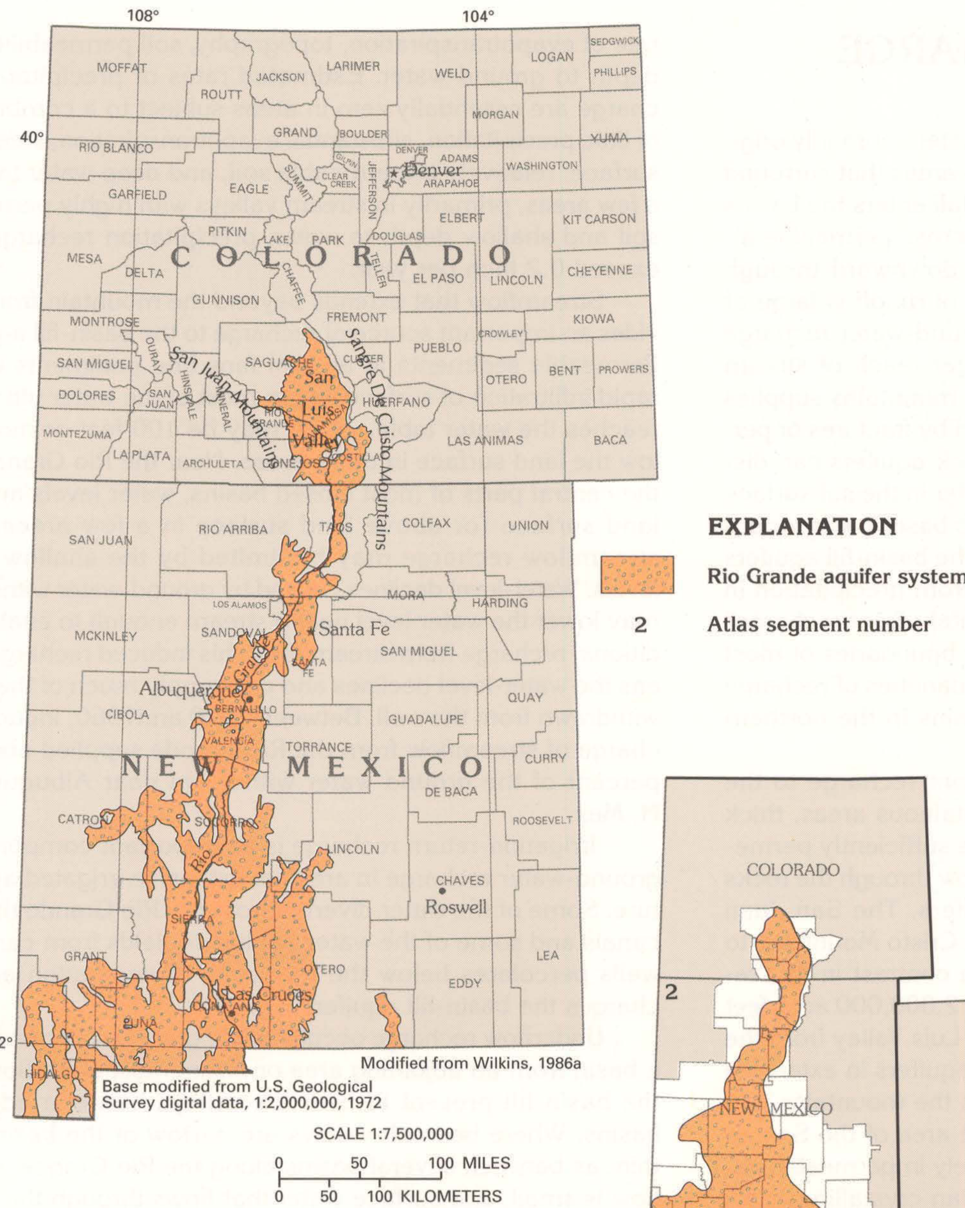
INTRODUCTION

The Rio Grande aquifer system is the principal aquifer in a 70,000-square-mile area of southern Colorado, central New Mexico, and western Texas (**fig. 52**). The aquifer system consists of a network of hydraulically interconnected aquifers in basin-fill deposits located along the Rio Grande Valley and nearby valleys. The aquifer system corresponds to the eastern part of the Southwest alluvial basins aquifer system, as defined by U.S. Geological Survey Regional Aquifer-System Analysis studies, and is located in the Southern Rocky Mountains and Basin and Range Physiographic Provinces. The part of the aquifer system in Colorado and New Mexico is described in this report; the part in Texas is described in Chapter E of this Atlas.

The basin-fill aquifers of the Rio Grande aquifer system are present in intermountain basins between discontinuous mountain ranges in southern New Mexico and between mountains and tablelands in northern New Mexico. High mountains border the aquifers in southern Colorado. The mountains and edges of the tablelands slope steeply, almost precipitously in some areas, toward the basins. Coalescing alluvial fans lie near the base of many mountains and form the intervening slopes located between the mountains and the relatively flat basin floor. The altitudes of the basin floors range from about 4,000 feet near El Paso, Tex., to about 8,000 feet in the San Luis Valley of Colorado. The altitude of some mountains exceeds 14,000 feet in Colorado, although mountains commonly are 8,000 to 10,000 feet in altitude and extend only 1,000 to 4,000 feet above the basin floors in most of the area.

The arid climate of the Rio Grande Valley and adjacent basins provides insufficient precipitation for growth of most commercial crops; consequently, irrigation is required. In the San Luis Valley, for example, the average rate of pan evaporation during a month may be as much as 10 times larger than the average rate of precipitation for the month (**fig. 53**). The rate of evaporation is large in most of the valley areas because of high summer temperatures, low relative humidity, abundant sunshine, and frequent wind. Much of the area is sparsely vegetated.

The Rio Grande is the largest river in the area and has perennial flow through most of its length in Colorado and New Mexico. The river flows across the broad basin-fill deposits in the San Luis Valley in Colorado and then flows through about 100 miles of deep canyon (**fig. 54**) and small intermountain basins in northern New Mexico. South of Santa Fe, N. Mex., the river flows through a series of broad basins and narrow valleys to the State line in southern New Mexico. Most basins along the Rio Grande have surface drainage to the river and are topographically open basins. The northern end of the San Luis Valley and most other basins distant from the river have internal surface-water drainage and are topographically closed basins that generally do not contribute streamflow to the Rio Grande or its tributaries (**fig. 55**). Much of the streamflow in the more mountainous northern part of the Rio Grande is derived from snowmelt runoff in the mountains. Streamflow in the southern part of the river system is derived from upstream flow, ground-water discharge, and runoff from summer thunderstorms.



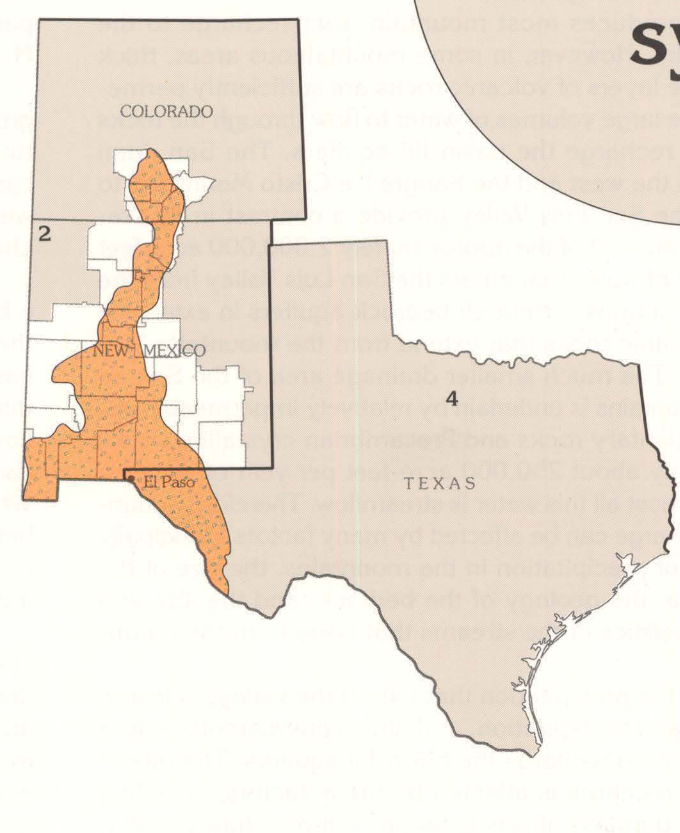


Figure 52. *The Rio Grande aquifer system extends through parts of three States and is the principal source of ground water in much of the area.*

Rio Grande aquifer system

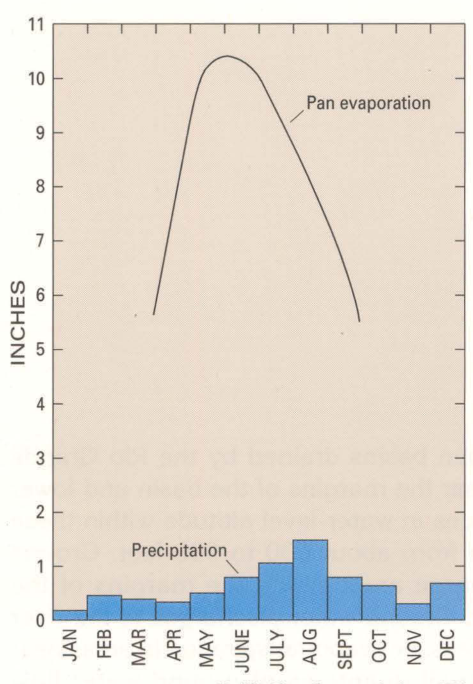


Figure 53. *The average monthly pan evaporation is much larger than the average monthly precipitation near the center of the San Luis Valley in Colorado. Data are for 1961 to 1967.*

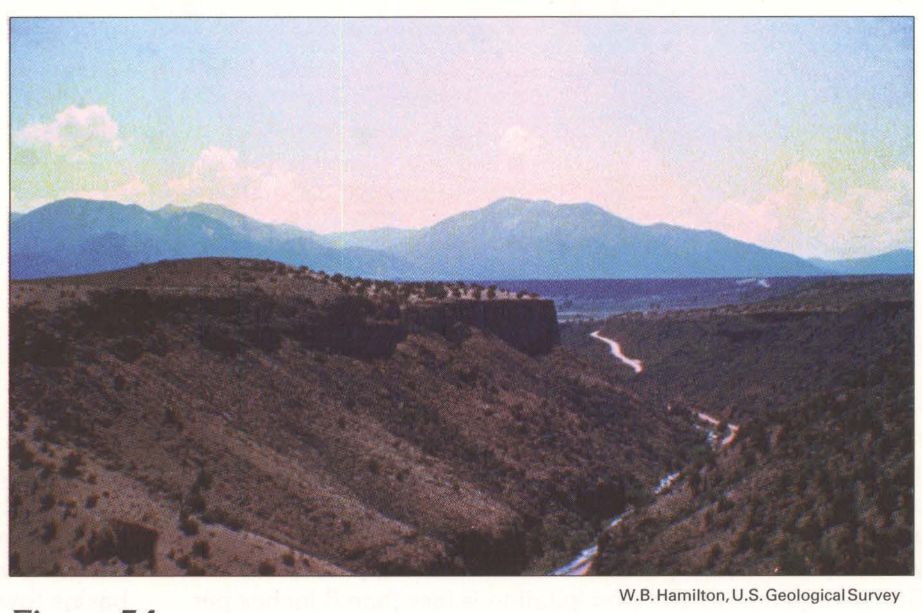


Figure 54. *The Rio Grande is entrenched in a canyon below the level of the tableland in the area north of Santa Fe, N. Mex.*

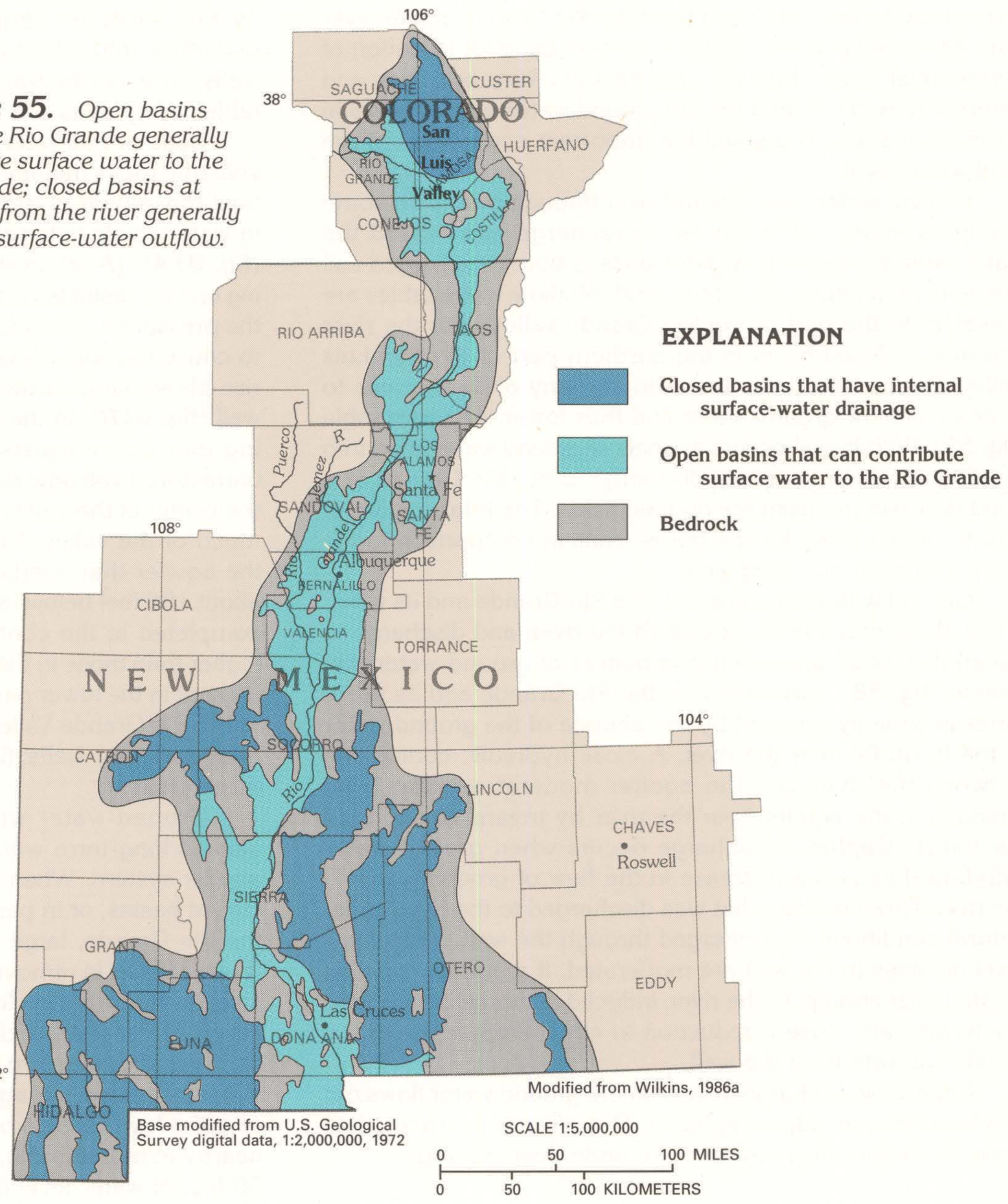


Figure 55. *Open basins along the Rio Grande generally contribute surface water to the Rio Grande; closed basins at distance from the river generally have no surface-water outflow.*

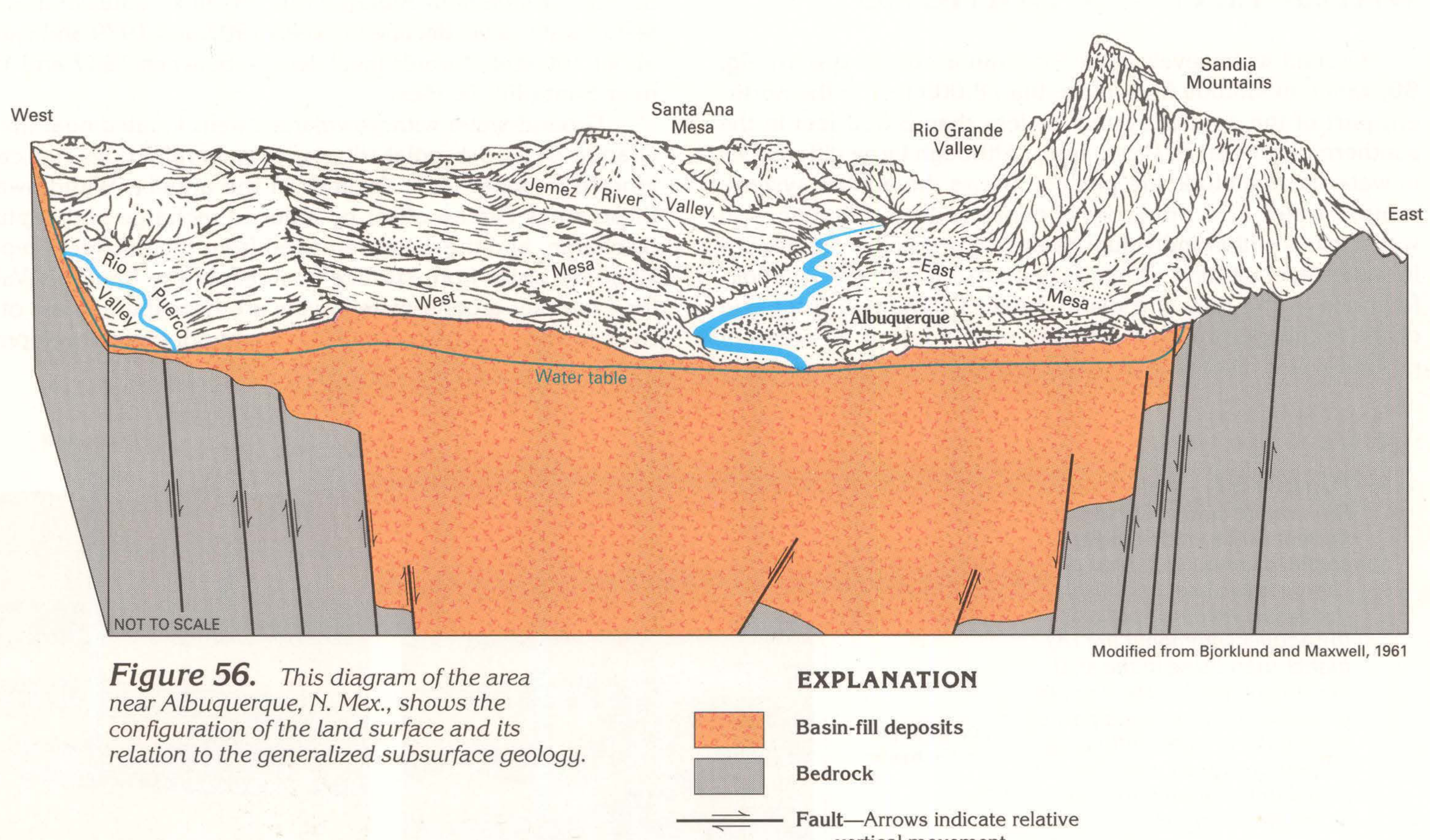


Figure 56. *This diagram of the area near Albuquerque, N. Mex., shows the configuration of the land surface and its relation to the generalized subsurface geology.*

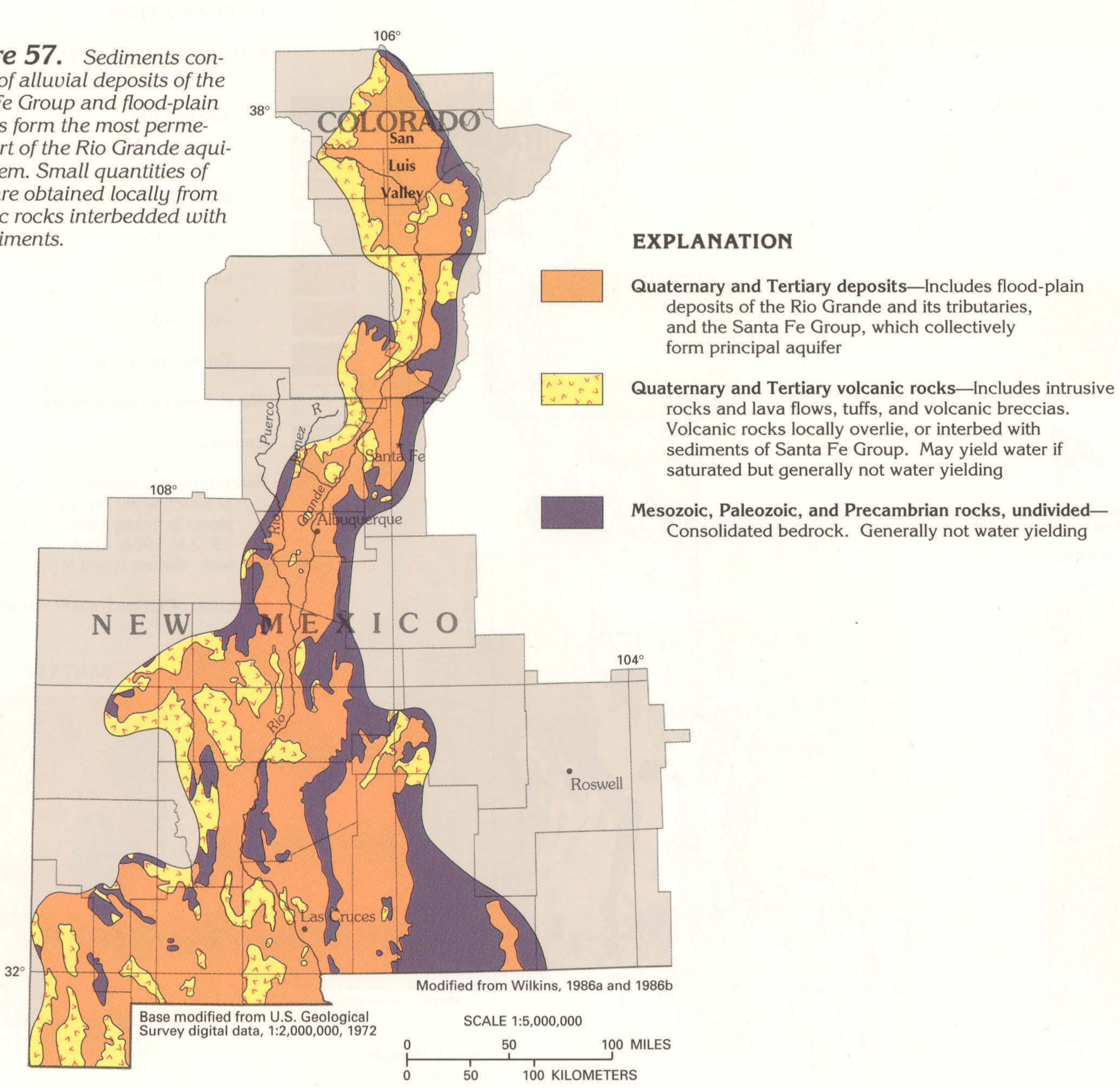
HYDROGEOLOGIC UNITS

The Rio Grande Rift is the principal geologic feature of the area. The rift affected the configuration of the bounding highlands, which in turn has affected precipitation, runoff, ground-water recharge, source material of the basin fill, aquifer characteristics, and water quality. The rift is a northward-trending series of interconnected, downfaulted and rotated blocks located between uplifted blocks to the east and west. Various blocks have been displaced downward thousands of feet, and most of the rift has been filled with alluvium and volcanic rocks (basin fill). The thickness of the basin fill is unknown in most areas but is estimated to be as much as 30,000 feet in the San Luis Valley, about 20,000 feet near Albuquerque, N. Mex. (**fig. 56**), and about 2,000 feet near El Paso, Tex. Total vertical displacement across some faults that border the rift exceeds 20,000 feet from the crest of the nearby mountains to the top of the equivalent rocks in the rift. Most basins of the rift are bounded on the north and west by Tertiary and Quaternary volcanic rocks. Igneous, metamorphic, and sedimentary rocks of Precambrian, Paleozoic, and Mesozoic ages (**fig. 57**) form the eastern boundary of most basins.

The bedrock formations that bound the basins generally consist of granite, quartzite, schist, and gneiss of Precambrian age; marine carbonates, volcanics, and clastic sedimentary rocks of Paleozoic age; and clastic sedimentary rocks and volcanic rocks of Mesozoic and Cenozoic age. Most of the many geologic formations present in this thick sequence of rock are relatively impermeable. Although some volcanic rocks, solution-altered carbonate rocks, or extensively fractured beds can yield water in local areas, the bedrock as a whole has minimal permeability and is considered to form an impermeable base to the Rio Grande aquifer system.

Older and younger basin fill are the principal water-yielding materials in the system. Older basin fill consists of the Santa Fe Group in most of the area and its lateral equivalent, the Gila Conglomerate, in the southwestern part of the aquifer system in New Mexico. The Santa Fe Group is a Tertiary and Quaternary rock-stratigraphic unit that consists of unconsolidated to moderately consolidated lenticular deposits of gravel, sand, and clay interbedded in some areas with andesitic and rhyolitic lava flows, tuffs, and breccias. Younger basin fill consists of unconsolidated, poorly to well-sorted, interbedded Quaternary gravel, sand, silt, and clay. Alluvial fans and pediment-cover deposits near the mountains generally grade imperceptibly into, and intertongue with, either fine-grained playa deposits in valleys or medium- to coarse-grained fluvial deposits. Terrace deposits that consist of gravel, sand, and silt extend 30 to 175 feet above the level of the present flood plain. During late Quaternary time, the Rio Grande was entrenched as much as 60 to 130 feet below the present level of the flood plain. The younger basin fill is similar in appearance and composition to the underlying older basin fill from which the younger unit was largely derived. The contact between the two units generally is about 100 feet below the flood plain and is characterized by subtle changes in lithology and consolidation.

Figure 57. *Sediments consisting of alluvial deposits of the Santa Fe Group and flood-plain deposits form the most permeable part of the Rio Grande aquifer system. Small quantities of water are obtained locally from volcanic rocks interbedded with the sediments.*



RECHARGE AND DISCHARGE

Recharge to the Rio Grande aquifer system primarily originates as precipitation in the mountainous areas that surround the basins. Runoff from snowmelt or rainfall enters the basins and generally flows for short distances across permeable alluvial fans before the water percolates downward through streambeds or evaporates. If the volume of runoff is large or becomes part of a perennial stream, ground-water recharge can be distributed through a much longer reach of stream channel. Some of the precipitation in the mountains supplies water to bedrock aquifers that were formed by fractures or permeable layers in the bedrock. The bedrock aquifers can discharge water directly to the basin-fill aquifer in the subsurface at the mountain front or discharge water to base flow in mountain streams that subsequently recharge the basin-fill aquifers near the mountain front. Such recharge from precipitation in mountainous areas is here termed "mountain-front recharge" and is distributed along the mountainous boundaries of most basins, as indicated in **figure 58**. Larger quantities of recharge generally occur along the higher mountains in the northern parts of the aquifer system.

Runoff produces most mountain-front recharge to the aquifer system. However, in some mountainous areas, thick and extensive layers of volcanic rocks are sufficiently permeable to enable large volumes of water to flow through the rocks and directly recharge the basin-fill aquifers. The San Juan Mountains to the west and the Sangre De Cristo Mountains to the east of the San Luis Valley provide a contrast in this regard. About one-half of the approximately 2,000,000 acre-feet per year flow of water that enters the San Luis Valley from the San Juan Mountains is through bedrock aquifers in extensive layers of volcanic rocks that extend from the mountains into the basin fill. The much smaller drainage area of the Sangre De Cristo Mountains is underlain by relatively impermeable Paleozoic sedimentary rocks and Precambrian crystalline rocks and yields only about 250,000 acre-feet per year of water to the valley; almost all this water is streamflow. Therefore, mountain-front recharge can be affected by many factors, principally the quantity of precipitation in the mountains, the size of the drainage area, the geology of the bedrock, and the size and flow characteristics of the streams that issue from the mountains.

Most of the precipitation that falls in the valleys is lost to evaporation and transpiration, and little water percolates to a depth sufficient to recharge the basin-fill aquifers. The rate of precipitation recharge is affected by many factors, including quantity and duration of precipitation, soil-moisture content, rate of evapotranspiration, topography, soil permeability, and depth to ground water. Estimated rates of precipitation recharge are essentially zero in areas subject to a combination of little precipitation, substantial evapotranspiration, steep land surface, relatively impermeable soil, and deep water table. In a few areas, primarily in stream valleys with highly permeable soil and shallow depth to water, precipitation recharge may exceed 0.2 inch per year.

Streamflow that extends beyond the mountain front provides an important source of recharge to the basin-fill aquifers. Permeable sediments in alluvial fans and pediments enable rapid infiltration of surface water; most of the water ultimately reaches the water table, which may be 100 feet or more below the land surface in these areas. Near the Rio Grande and the central parts of most closed basins, water levels are near land surface (or above land surface in a few areas), and streamflow recharge may be limited by the shallow water levels. Water-level declines caused by ground-water withdrawal may lower the water level near a stream enough to enable additional recharge from streamflow. This induced recharge lessens the water-level declines and can supply much of the water withdrawn from the well. Between 1920 and1960, induced recharge of streamflow from the Rio Grande supplied about 80 percent of the ground water withdrawn near Albuquerque, N. Mex.

Irrigation-return recharge is an important component of ground-water recharge in areas of extensive irrigated agriculture. Some of the water diverted from the Rio Grande through canals and some of the water applied to fields from canals or wells percolates below the root zone of vegetation and recharges the basin-fill aquifers.

Underflow recharge occurs where ground water flows into a basin from an adjoining area or basin. Most underflow is in the basin fill present in bedrock valleys between adjacent basins. Where bedrock valleys are narrow or the basin fill is thin, as between several basins along the Rio Grande, underflow is small, and surface water that flows through the valley can provide the principal hydraulic connection between basins. Where bedrock valleys are broad or the basin fill is thick, as between most basins remote from the Rio Grande, underflow recharge may constitute a large part of the total recharge to a basin.

Ground water discharges from the Rio Grande aquifer system by evapotranspiration, withdrawal from wells and drains, discharge to streamflow, and underflow (**fig. 58**). In the arid climate of New Mexico and southern Colorado, rates of evapotranspiration are large, and water is readily lost by evaporation from moist soil and water surfaces and by transpiration from vegetation. Evapotranspiration annually removes about 2,800,000 acre-feet per year of water from the 1,700,000-acre area of the San Luis Valley. This is a rate of about 20 inches per year from an area that receives about 10 inches per year of precipitation and has about 60 inches per year of potential evaporation. In the southern part of the Rio Grande aquifer system, precipitation is less than 8 inches per year, and potential evaporation exceeds 100 inches per year; the rate of evapotranspiration likely is greater than 20 inches per year in this area.

Ground-water withdrawal primarily occurs as discharge from pumping wells. In 1985, about 1 million acre-feet of ground water was withdrawn from the Rio Grande aquifer system. About 90 percent of this water was used for irrigation of commercial crops. Public water supplies for most cities and communities in the area rely on ground water, and withdrawal for municipal use is a small but important component of the total withdrawal.

Ground water also is withdrawn through drains. In some low-lying areas, irrigation-return recharge has caused the water table to rise so near land surface that waterlogged soil prevents agricultural use of the land. Shallow water tables are prevalent in the part of the Rio Grande Valley near the river and in the closed basin in the northern part of the San Luis Valley. Drains have been installed in many of these areas to remove shallow ground water and thus lower the water table (**fig. 59**). Withdrawal occurs as shallow ground water flows into the drains, which generally discharge to the Rio Grande at a point downstream from the drained fields. The total volume of flow to drains is not known but is small in comparison to the discharge from pumping wells.

Ground water discharges to the Rio Grande and its tributaries along much of the length of the river, and discharge to streamflow is an important component of ground-water discharge (**fig. 58**). Streamflow in the Rio Grande and its tributaries is strongly affected by the altitude of the ground water in the basin fill near the river. A close hydraulic connection between the river and the aquifer moderates water-level changes in the aquifer near the river by means of captured discharge. Captured discharge occurs when ground-water withdrawal causes a decrease in the flow of ground water to the river. Ground water that was discharged to the river under natural conditions is discharged through the well, and water-level declines in the well are moderated. If a pumping well is located near enough to the river, induced recharge or captured discharge can cause a reduction in streamflow equal to the withdrawal rate from the well.

Underflow discharge occurs where ground water flows out of a basin into an adjoining basin. This discharge process has been discussed above in terms of underflow recharge.

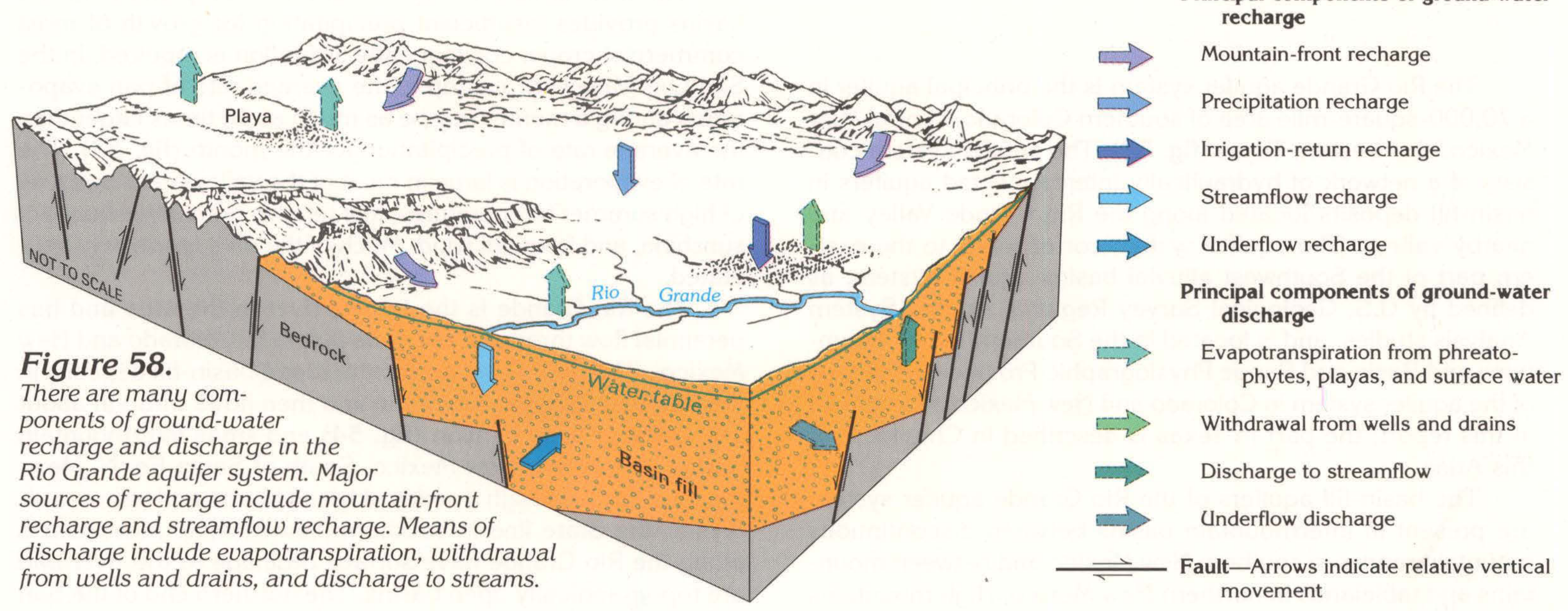


Figure 58. *There are many components of ground-water recharge and discharge in the Rio Grande aquifer system. Major sources of recharge include mountain-front recharge and streamflow recharge. Means of discharge include evapotranspiration, withdrawal from wells and drains, and discharge to streams.*

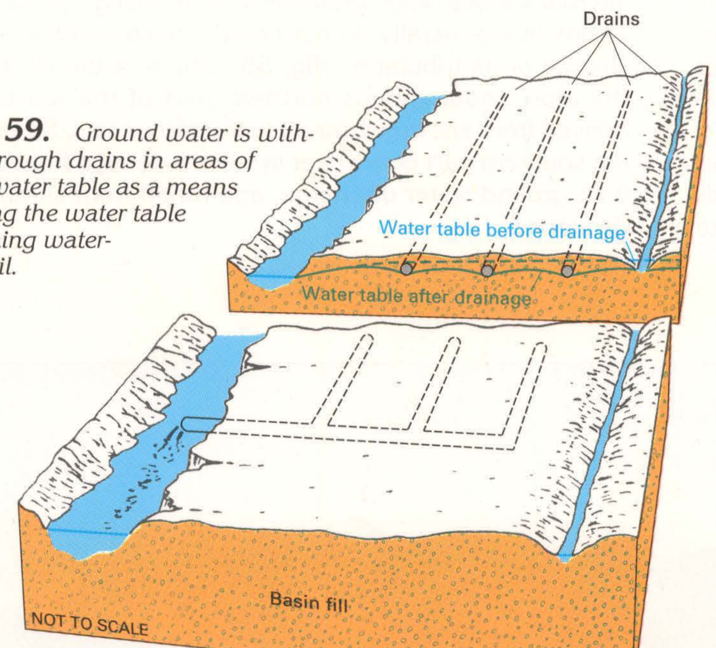


Figure 59. *Ground water is withdrawn through drains in areas of shallow water table as a means of lowering the water table and draining waterlogged soil.*

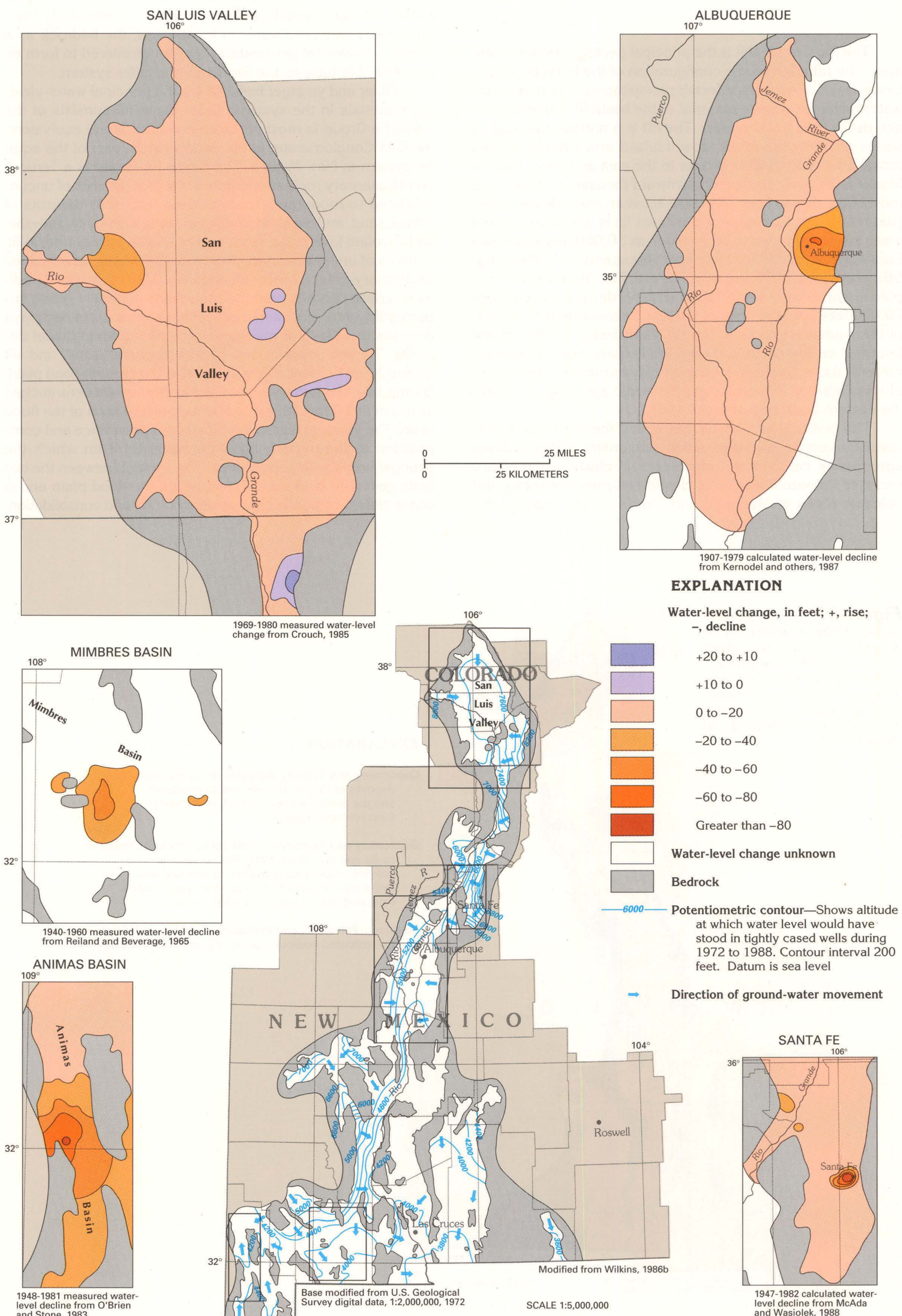


Figure 60. *The altitude of the water level in the Rio Grande aquifer system is highest in the San Luis Valley, Colo., and lowest near El Paso, Tex. Water-level declines generally have occurred in areas distant from the Rio Grande and in closed basins.*

Water levels in open basins drained by the Rio Grande generally are higher near the margins of the basin and lower near the river. Differences in water-level altitude within these basins generally range from about 600 to 800 feet. Ground water flows from recharge areas along the margins of the basins toward the river. Ground water discharges to the river or to drains, pumping wells, or evapotranspiration in areas near the river. Relatively small quantities of ground water flow through the narrow basin-fill areas between the broader basins.

Water levels in closed basins tend to be higher near the margins of the basin, but these basins do not have a stream as a principal discharge area. Discharge generally occurs as underflow into adjoining basins, as discharge from pumping wells, or as evapotranspiration from areas of shallow water table near playas.

Beds of relatively impermeable clay, silt, or unfractured volcanic rocks that are thick and extensive can retard the vertical movement of ground water and cause large differences in water levels between shallow wells and nearby deep wells (**fig. 61*A***). (A relatively impermeable bed is called a confining unit if it restricts vertical movement of water.) In some areas, the pressure in the confined part of the aquifer is large enough to cause the water level in a well completed in the aquifer to rise above land surface, and water may flow freely from the well (**fig. 61*B***). In the San Luis Valley, for example, a confining unit that consists of interlayered clay, silt, sand, and unfractured volcanic rocks is in excess of 1,200 feet thick near the center of the valley and extends through the subsurface of much of the valley. Depth to water in an unconfined part of the aquifer that overlies the confining unit ranges from 0 to about 100 feet below land surface. Water levels in the few wells completed in the confined part of the aquifer generally are higher than those in the unconfined part, and flowing wells are present in the lower part of the valley. In many low-lying areas of the Rio Grande Valley, water levels are higher in deep wells than in shallow wells; flowing wells also are present in some of these areas.

Ground-water withdrawal from wells and drains has caused long-term water-level declines in some parts of the aquifer system. When rates of withdrawal have been large in closed basins, or in parts of open basins that are remote from the Rio Grande, large water-level declines have occurred as ground water is removed from storage in the aquifers. Extensive development in closed basins generally has produced the largest water-level declines. For example, in the Animas basin near the southwestern margin of the aquifer system, ground-water withdrawal for agricultural use caused more than 80 feet of water-level decline between 1948 and 1981 (**fig. 60**). In the nearby Mimbres basin, similar water use caused as much as 50 feet of water-level decline between 1940 and 1960. Withdrawal for irrigation and municipal use in areas remote from the Rio Grande near Albuquerque, N. Mex., caused about 60 feet of water-level decline between 1907 and 1979 and caused about 100 feet of water-level decline between 1947 and 1982 near Santa Fe, N. Mex.

Ground-water withdrawal from wells located near the Rio Grande or its perennial tributaries generally does not cause long-term water-level declines in the aquifer. Withdrawal in these areas can be offset by induced recharge and captured discharge, and the water removed from the aquifer is replenished by water from surface sources. In the San Luis Valley, for example, ground-water levels are near the altitude of the river and have changed little since ground-water development began.

WATER-LEVEL CONDITIONS

Ground-water levels in the Rio Grande aquifer system (**fig. 60**) range in altitude from more than 8,000 feet in the northern part of the aquifer system to less than 3,800 feet in the southern part (near El Paso, Tex.). Although large differences in water-level altitude are present across the aquifer system, ground-water flow primarily is controlled by differences in water levels within individual basins. The smaller differences in water-level altitude within basins are the result of local differences in the rate and distribution of mountain-front and other recharge, and the altitude and location of areas of ground-water discharge, such as the Rio Grande.

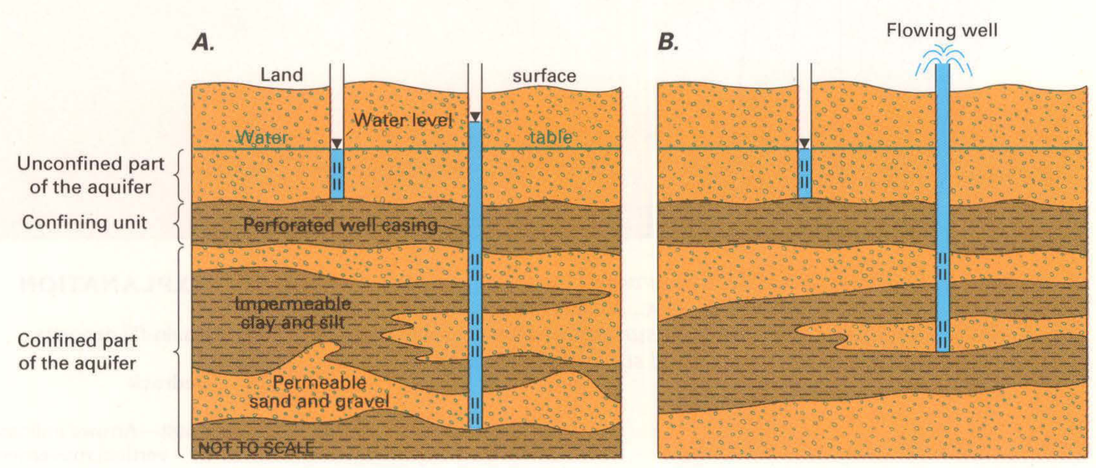


Figure 61. *Water in the Rio Grande aquifer system is present under unconfined and confined conditions. Near discharge areas, water levels in the deeper confined parts of the aquifer commonly are (A) higher than those in the shallow, unconfined parts. If water pressure in the confined part of the aquifer is sufficiently large (B), an uncapped well completed in the aquifer will flow at the land surface.*

GROUND-WATER QUALITY

The chemical composition and dissolved-solids concentration of water in the Rio Grande aquifer system are affected by the quality of the water that enters the aquifer, the type and solubility of minerals present in the basin fill, and the quantity of water lost by evaporation and transpiration. Soluble minerals present in the rocks of the mountains adjacent to the basins affect the quality of the water draining from the mountains, which, in turn, affects the quality of the recharge entering the aquifers. Water in the aquifer system is of varied chemical composition (**fig. 62**), in part because of the varied geology of the nearby mountains. Surface water in the Rio Grande in the reach from the headwaters to near Albuquerque, N. Mex., generally has a small dissolved-solids concentration and contains a preponderance of calcium, bicarbonate, and sulfate ions. This water is classified as a calcium bicarbonate or calcium sulfate type. Ground water near recharge areas and in the northern part of the aquifer system generally is a calcium or magnesium bicarbonate type. Streamflow in the Jemez River (to the northwest of Albuquerque) and the Rio Puerco (to the southwest of Albuquerque) has a larger dissolved-solids concentration than the Rio Grande, and the water is a sodium chloride or sodium sulfate type. Ground water near these streams is of similar chemical composition. Differences in chemical composition and dissolved-solids concentration of recharge to other parts of the aquifer system produce most of the areal differences in ground-water quality shown in **figures 62 and 63**.

As ground water flows through the basin fill, soluble minerals such as calcite and dolomite (calcium and magnesium carbonates), gypsum (calcium sulfate), halite (rock salt), and many other minerals are dissolved from the sediments. This dissolution increases the dissolved-solids concentration of the ground water and may alter the chemical composition of the water. Clay minerals also may alter the composition of the water through the process of cation exchange. This natural water-softening process involves the exchange of calcium or magnesium ions in solution for sodium ions that are bound to some clay minerals. For example, a hard water of a calcium bicarbonate type could exchange the calcium ions for the sodium ions on the clay and become a soft water of a sodium bicarbonate type. The calcium ions that formerly were in solution become immobile on the clay minerals. Water in the aquifer system ranges locally from soft to very hard, but softer water is more prevalent in parts of the aquifer system in Colorado and northern and southwestern New Mexico.

Water loss to evapotranspiration has an important effect on ground-water quality in areas of irrigated agriculture, near playas, and other areas of shallow water table. Evapotranspiration removes water from the aquifer or the soil but does not remove the minerals that formerly were dissolved in the water. These minerals can accumulate in the soil to form alkali deposits or salt flats or can be flushed from the soil by infiltration of precipitation or irrigation water. Additional dissolved solids carried into the aquifer from such surficial sources can form a zone of degraded ground-water quality at the top of the aquifer (**fig. 64**). In the lower Rio Grande Valley near Las Cruces, N. Mex., infiltration of irrigation water has produced a slightly saline zone (1,000–3,000 milligrams per liter dissolved solids) that is about 100 to 150 feet thick at the top of the aquifer. A transition zone of intermediate salinity (500–1,000 milligrams per liter dissolved solids) that is 50 to 100 feet thick separates the slightly saline zone from the underlying freshwater zone (300–500 milligrams per liter dissolved solids) that extends to depths of 1,000 to 1,500 feet. A second transition zone separates the freshwater zone from the deep saline zone where dissolved-solids concentrations can exceed 3,000 milligrams per liter. Similar stratification of water of large dissolved-solids concentration in the upper part of the aquifer occurs in the San Luis Valley, much of the lower Rio Grande Valley near the river, and in many of the closed basins where playas are major discharge areas.

Evapotranspiration and tributary inflow produce a general downgradient increase in dissolved-solids concentrations along the valley of the Rio Grande (**fig. 63**). The dissolved-solids concentration of most ground water is about 230 milligrams per liter north of Santa Fe, N. Mex., and about 410 milligrams per liter in the reach from Santa Fe to about 50 miles south of Albuquerque; farther south, most ground water contains between 600 and 700 milligrams per liter dissolved solids.

Figure 62. *Ground water in the Rio Grande aquifer system is of varied chemical composition. Generally, calcium bicarbonate or magnesium bicarbonate type water is more prevalent in the northern part of the system, and sodium bicarbonate or sodium sulfate type water is more prevalent in the southern part of the system along the Rio Grande.*

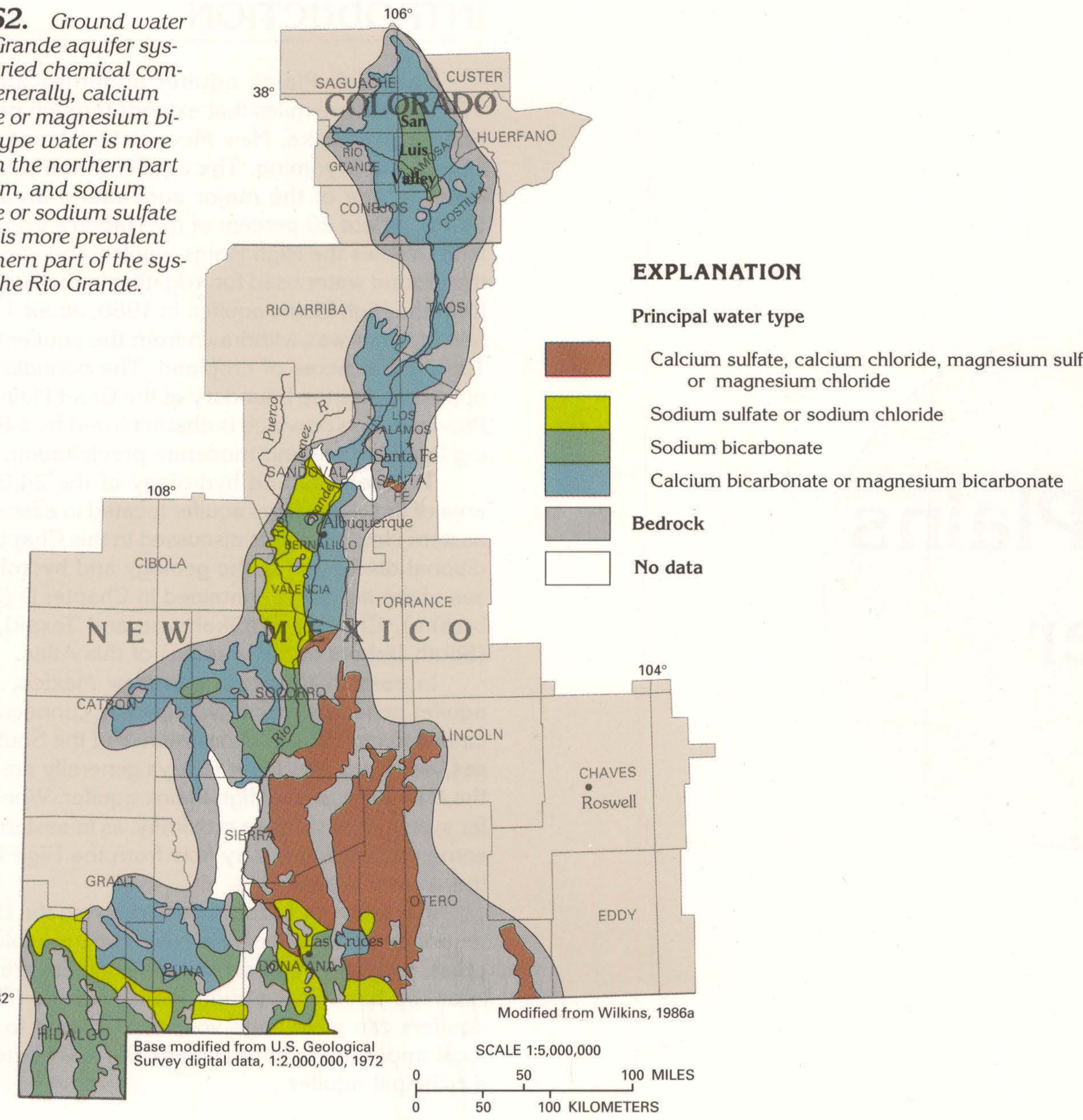
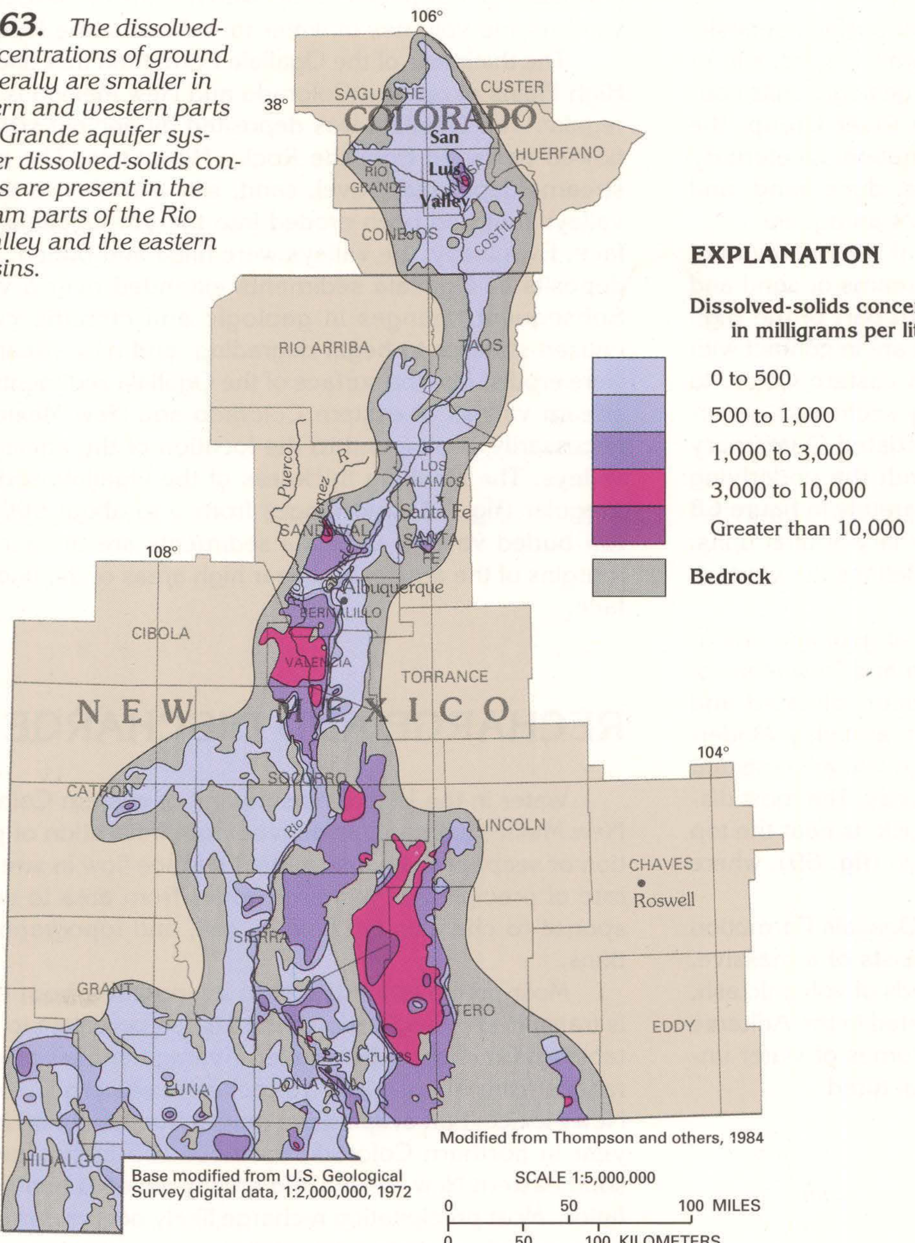


Figure 63. *The dissolved-solids concentrations of ground water generally are smaller in the northern and western parts of the Rio Grande aquifer system. Larger dissolved-solids concentrations are present in the downstream parts of the Rio Grande valley and the eastern closed basins.*



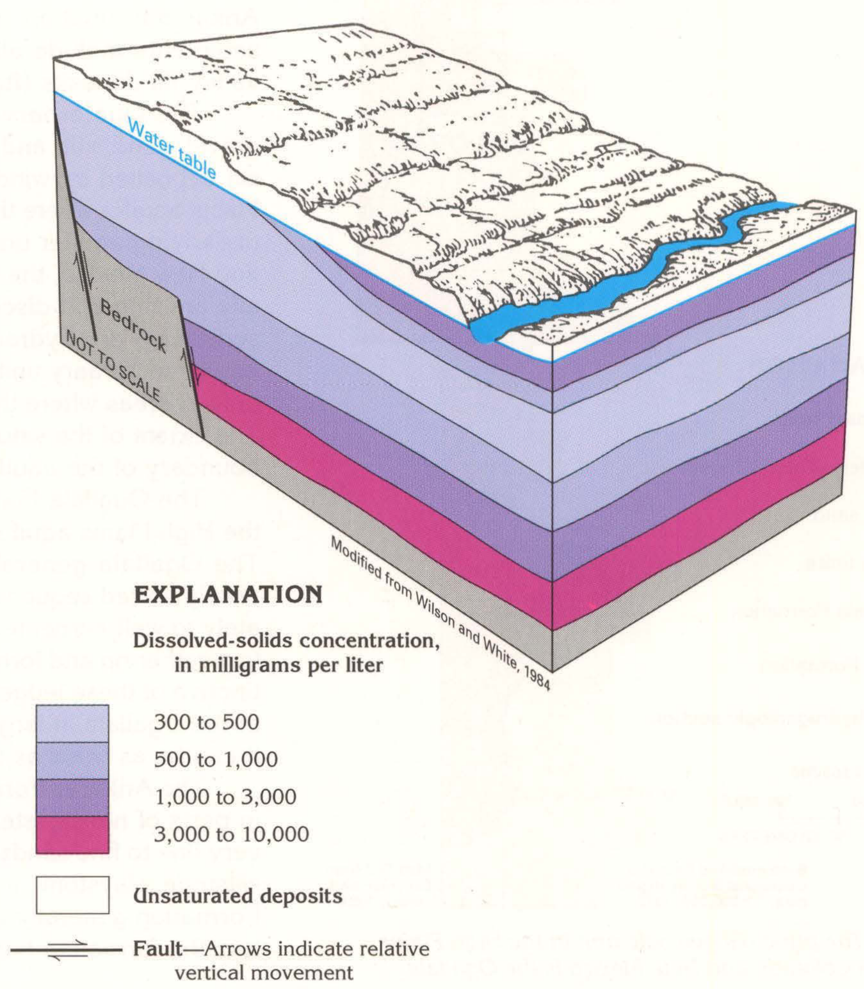


Figure 64. *Applied irrigation water can flush salts from the soil, recharging the aquifer with slightly saline water. Near Las Cruces, N. Mex., and elsewhere along the Rio Grande, this recharge has created a zone of slightly saline water that overlies freshwater.*

FRESH GROUND-WATER WITHDRAWALS

Withdrawals of freshwater from the Rio Grande aquifer system totaled about 1,200,000 acre-feet during 1985. Agriculture used about 900,000 acre-feet or about 77 percent of the ground water withdrawn (**fig. 65**). Public supply, primarily for the cities of Albuquerque, Las Cruces, and Santa Fe, N. Mex., used about 180,000 acre-feet or about 15 percent of the ground water withdrawn. Domestic and commercial, and industrial, mining, and thermoelectric power uses constituted the remaining approximately 8 percent.

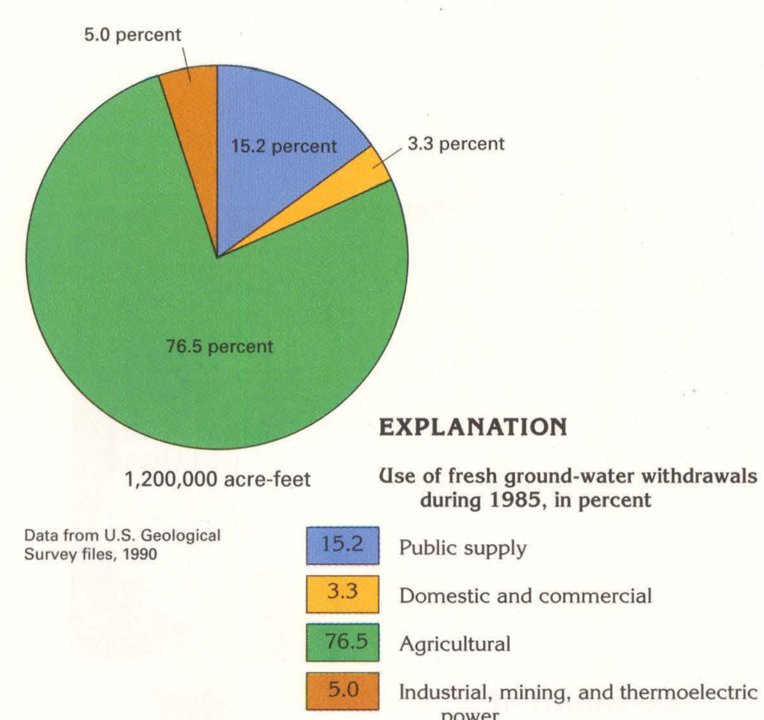


Figure 65. *Most of the freshwater withdrawn from the Rio Grande aquifer system during 1985 was used for agricultural purposes.*

High Plains aquifer

INTRODUCTION

The High Plains aquifer underlies an area of about 174,000 square miles that extends through parts of Colorado, Kansas, Nebraska, New Mexico, Oklahoma, South Dakota, Texas, and Wyoming. The aquifer is the principal source of water in one of the major agricultural areas of the United States. About 20 percent of the Nation's irrigated agricultural land overlies the High Plains aquifer, and about 30 percent of the ground water used for irrigation in the Nation is withdrawn from the High Plains aquifer. In 1980, about 17,800,000 acre-feet of water was withdrawn from the aquifer to irrigate about 13,000,000 acres of cropland. The boundary of the aquifer approximates the boundary of the Great Plains Physiographic Province. The province is characterized by a flat to gently rolling land surface and moderate precipitation.

The geology and hydrology of the 24,000-square-mile area of the High Plains aquifer located in eastern Colorado and eastern New Mexico is discussed in this Chapter (**fig. 66**). Additional discussion of the geology and hydrology of other areas of the aquifer is contained in Chapter D (Kansas and Nebraska), Chapter E (Okahoma and Texas), and Chapter I (South Dakota and Wyoming) of this Atlas.

In eastern Colorado and New Mexico, the High Plains aquifer generally is not hydraulically connected to other principal aquifers. The alluvial aquifers of the South Platte, Arkansas, and Canadian River valleys generally are located beyond the boundary of the High Plains aquifer. Where the two aquifer systems are in close proximity, as in easternmost Colorado, some ground water may flow from the High Plains aquifer to the alluvial aquifers.

The bedrock formations underlying the High Plains aquifer primarily consist of relatively impermeable shale. In some areas, water-yielding sandstone is interlayered with the shale near the base of the High Plains aquifer. These sandstone aquifers can yield large volumes of water to wells and be of local importance, but they lack the areal extent to constitute a principal aquifer.

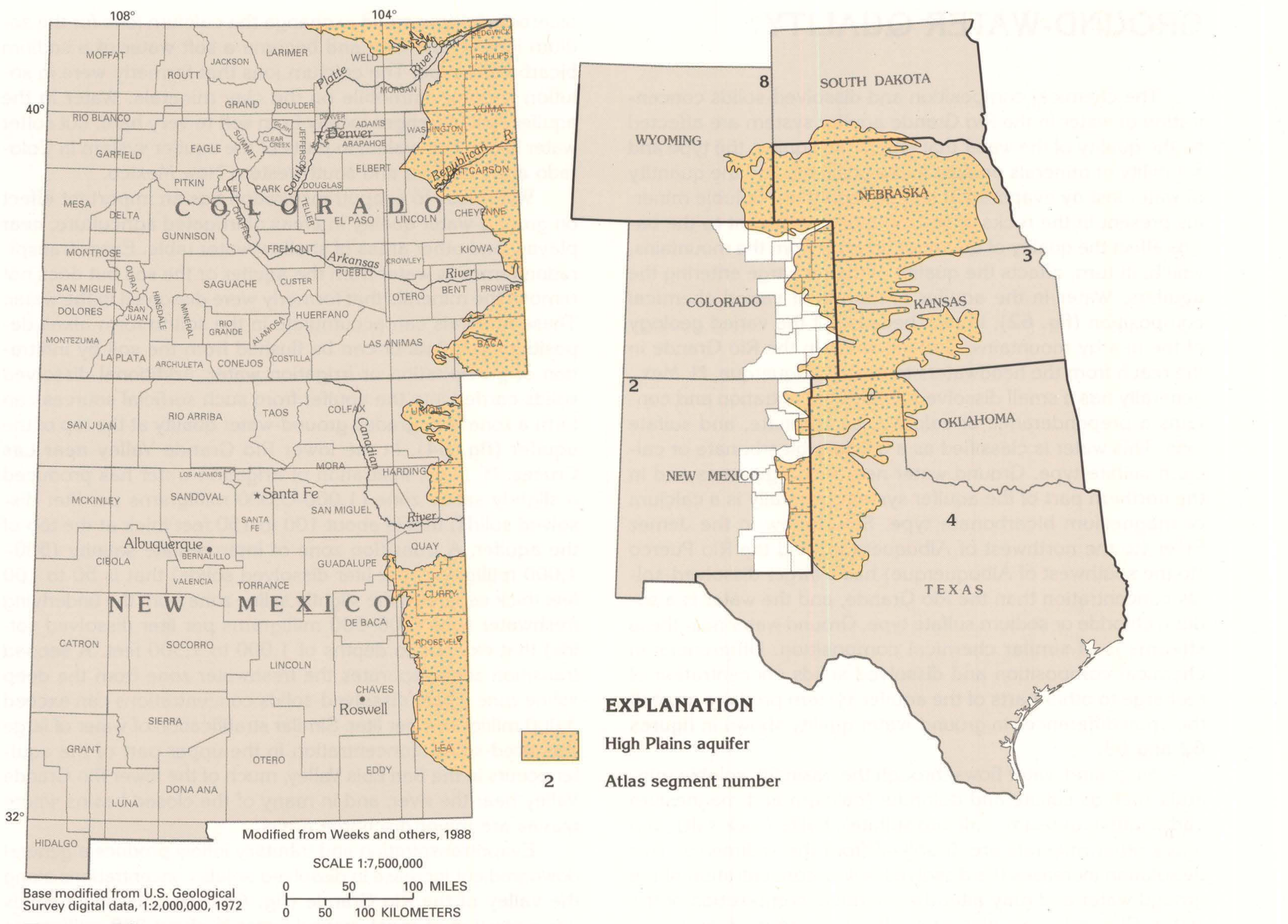


Figure 66. *The High Plains aquifer extends through an area of about 174,000 square miles in parts of eight States.*

Era	System	Series	Stratigraphic unit	Hydrogeologic unit	Physical characteristics
Cenozoic	Quaternary	Holocene and Pleistocene	Alluvial deposits, valley-fill deposits and dune sand	High Plains aquifer	Gravel, sand, silt, and clay
	Tertiary, Upper	Miocene	Ogallala Formation		Unconsolidated, poorly sorted gravel, sand, silt, and clay
			Arikaree Formation		Sandstone, fine to very fine. Local beds of volcanic ash, siltstone, claystone, and marl
	Tertiary, Lower	Oligocene	Brule Formation (White River Group)		Siltstone with sandstone as beds and channel deposits
			Chadron Formation (White River Group)	Confining unit	Clay and silt

Modified from Gutentag and others, 1984

Figure 67. *Geologic units ranging in age from Oligocene to Quaternary compose the High Plains aquifer. The permeable units consist of sand, sandstone, and gravel.*

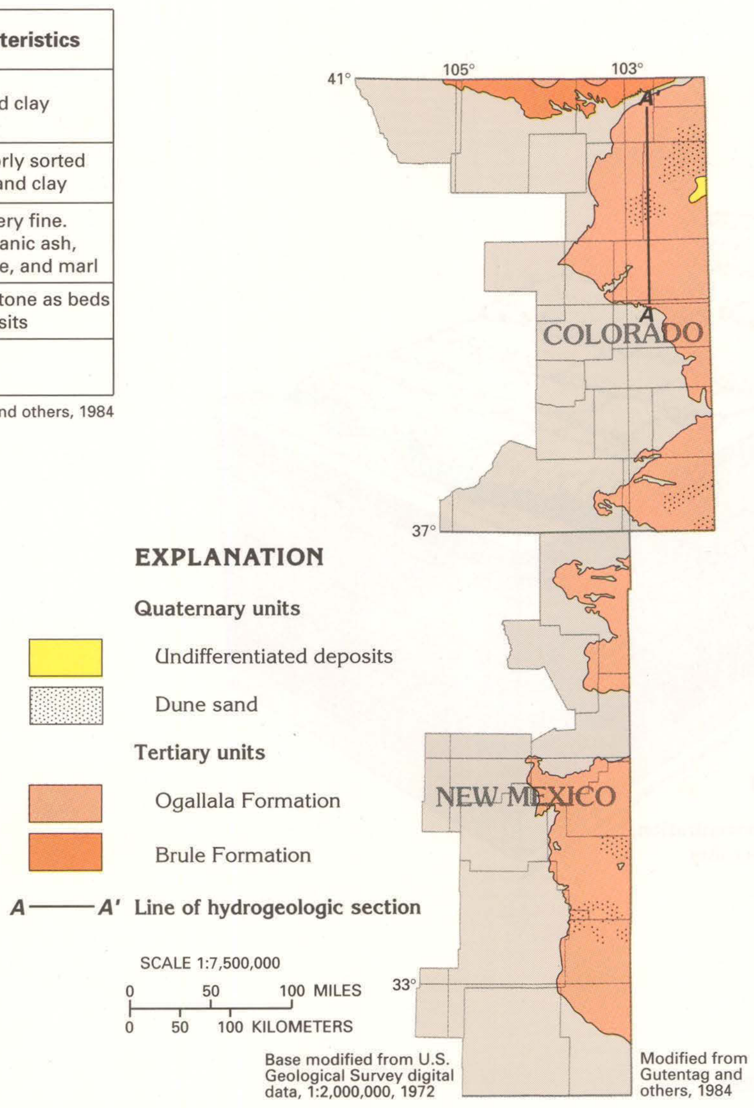


Figure 68. *The principal geologic unit in the High Plains aquifer of eastern Colorado and New Mexico is the Ogallala Formation.*

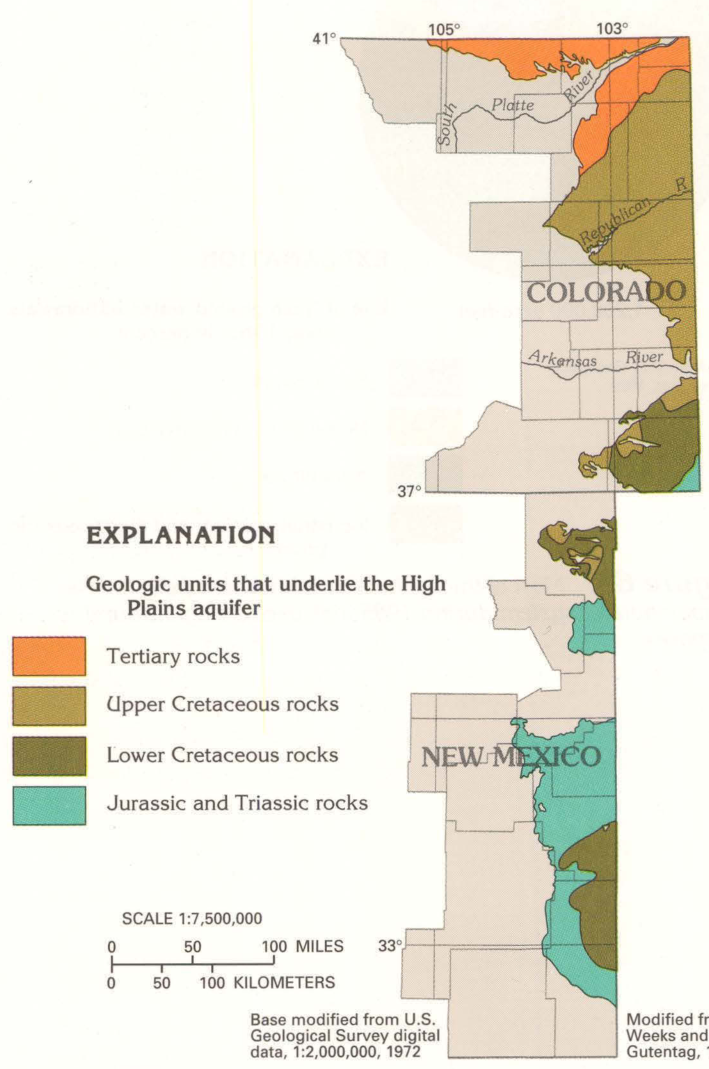
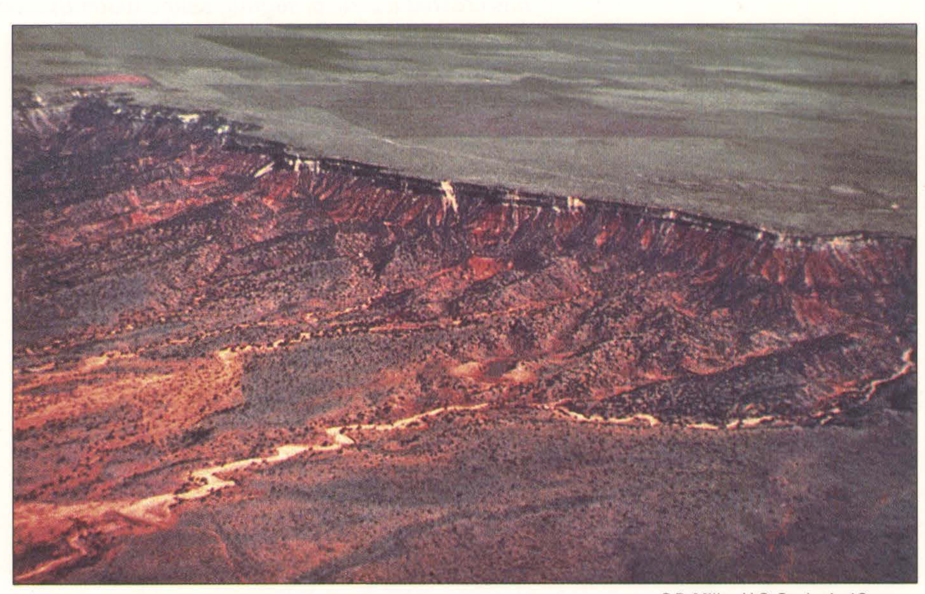


Figure 70. *The High Plains aquifer overlies older rocks that generally are much less permeable than the rocks of the aquifer. The older rocks form the base of the aquifer.*



C.D. Miller, U.S. Geological Survey

Figure 69. *The Mescalero Escarpment forms the margin of the distant High Plains to the south of the Canadian River in eastern New Mexico. The escarpment has formed on a cemented zone (the Ogallala cap rock) near the top of the Ogallala Formation.*

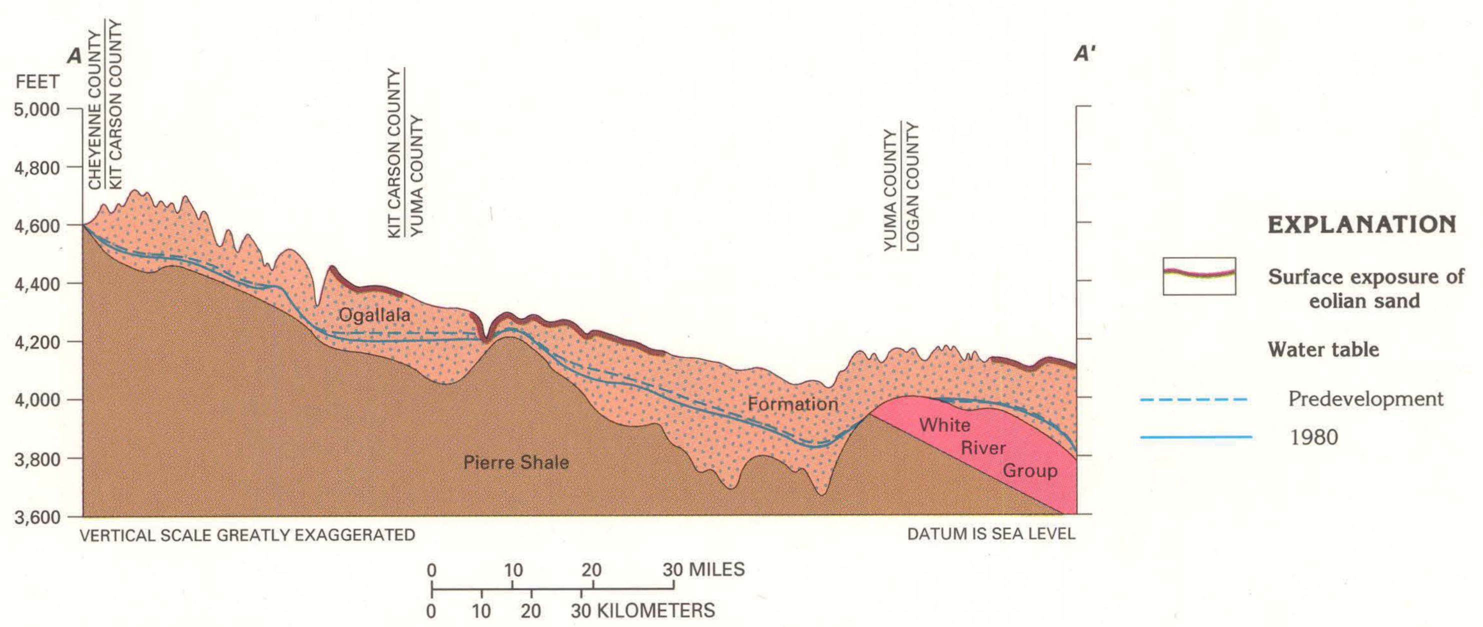


Figure 71. *The thickness of the High Plains aquifer varies greatly in northeastern Colorado. The aquifer is thickest where the Ogallala Formation fills ancient stream channels in the bedrock. The line of the section is shown in figure 68.*

HYDROGEOLOGIC UNITS

The High Plains aquifer consists of near-surface deposits of unconsolidated or partly consolidated gravel, sand, silt, or clay of Tertiary or Quaternary age. Tertiary geologic units consist of the Brule Formation of the White River Group, the Arikaree Formation, and the Ogallala Formation. Quaternary sediments include alluvial deposits, loess, dune sand, and valley-fill deposits (**fig. 67**); the loess is not an aquifer.

The Quaternary sediments consist of unconsolidated gravel, sand, silt, and clay deposited by streams or sand and silt deposited by wind. These deposits form part of the High Plains aquifer where they are saturated and are in contact with underlying aquifer units of Tertiary age. In eastern Colorado and New Mexico, the saturated Quaternary sediments generally are thin and discontinuous. Undifferentiated Quaternary sediments are hydraulically connected with the underlying aquifer in Tertiary units and are shown separately in **figure 68** only in areas where they do not overlie Tertiary aquifer units. The extent of the saturated Tertiary units defines the western boundary of the aquifer.

The Ogallala Formation is the principal geologic unit in the High Plains aquifer in eastern Colorado and New Mexico. The Ogallala generally consists of an unconsolidated and poorly sorted sequence of gravel, sand, silt, and clay. Moderately to well-cemented zones within the Ogallala are resistant to weathering and form ledges in outcrop areas. The most distinctive of these ledges, the Ogallala cap rock, is near the top of the Ogallala in large areas of New Mexico (**fig. 69**), where it can be as thick as 60 feet.

The Arikaree Formation underlies the Ogallala Formation in parts of northeastern Colorado and consists of a massive, very fine to fine sandstone and localized beds of volcanic ash, siltstone, claystone, and marl. Wells completed in the Arikaree Formation generally do not yield large volumes of water unless the formation has been extensively fractured.

The Brule Formation of the White River Group underlies the Arikaree Formation and crops out in northeastern Colorado north of the South Platte River. The Brule primarily consists of massive siltstone containing beds of sandstone, volcanic ash, claystone, and fine sand. It is relatively impermeable and is not part of the High Plains aquifer except in those areas where the rock has been extensively fractured. In northeastern Colorado, the Brule typically yields less than 300 gallons per minute to wells; however, yields as large as 1,500 gallons per minute have been reported.

Rocks that directly underlie the High Plains aquifer consist of the White River Group of Tertiary age (including the Brule Formation where it is unfractured and relatively impermeable) and rocks of Cretaceous, Jurassic, and Triassic age (**fig. 70**) that are mostly shale and interlayered sandstone, claystone, limestone, and dolomite. The Pierre Shale of Late Cretaceous age forms an impermeable base to the High Plains aquifer in much of Colorado. The thickness of the Pierre Shale increases from zero near the Arkansas River to more than 4,000 feet near the western edge of the aquifer in central Colorado. In southeastern Colorado and eastern New Mexico, the High Plains aquifer is underlain by shale and interlayered sandstone, claystone, limestone, and dolomite of relatively low permeability. Permeable zones within some of these rocks may yield usable volumes of water to wells in some local areas.

The thickness of the Ogallala Formation that underlies the High Plains aquifer of eastern Colorado and New Mexico is highly irregular. The Ogallala was deposited by ancient streams that flowed eastward from the Rocky Mountains. The aggrading streams deposited gravel, sand, silt, and clay in the stream valleys that had been eroded into the pre-Ogallala land surface. Eventually, the valleys were filled and buried, and thick deposits of Ogallala sediments extended over a vast area. Subsequent changes in geologic and climatic conditions caused streams to begin degrading, and new stream valleys were eroded into the surface of the Ogallala sediments. Present stream valleys of eastern Colorado and New Mexico do not necessarily correspond to the location of the ancient stream valleys. The resulting thickness of the Ogallala sediments is irregular (**fig. 71**) and ranges from 0 to about 500 feet in a few buried valleys. Ogallala sediments are thinner near the margins of the aquifer and near high areas of the bedrock surface.

RECHARGE AND DISCHARGE

Water in the High Plains aquifer of eastern Colorado and New Mexico primarily is derived from infiltration of precipitation or seepage from intermittent surface flow in streams. The rate of precipitation recharge varies from area to area in response to changes in climatic, soil, and topographic conditions.

Most precipitation is lost to evaporation from the soil or is transpired by vegetation before it can percolate to the water table and recharge the aquifer. Average annual precipitation ranges from about 12 to 16 inches in eastern Colorado and New Mexico; pan evaporation ranges from about 60 inches per year in northern Colorado to about 105 inches per year in southeastern New Mexico and greatly exceeds annual precipitation. Most precipitation recharge likely occurs during periods of snowmelt or prolonged rainfall when water is available for percolation and evapotranspiration rates are small.

Soil conditions affect recharge rates by impeding downward water movement. In deep, well-sorted, sandy soils or sand dunes, water can readily percolate to depths sufficient to prevent further loss to evapotranspiration. Most of this water will ultimately recharge the aquifer; thus, sandy soils are important sources of recharge. In clayey soils, percolation is slow, and most soil water is held at shallow depth, where it can be lost to evapotranspiration. In some areas of eastern Colorado and southeastern New Mexico, shallow, consolidated rocks or caliche deposits consisting of a well-cemented layer of subsoil retard deep percolation and hinder recharge.

Intense precipitation produces surface runoff that can accumulate in local depressions or contribute to streamflow. High flow in the ephemeral streams that cross the aquifer in Colorado and New Mexico is an important source of recharge. Most of the stream channels are located above the water table, and the streams lose water first to the sandy alluvial deposits in the channels, and subsequently to the underlying High Plains aquifer. High flow in the normally dry stream channels is uncommon, and significant recharge in these areas generally occurs only at intervals of many years.

The average rate of recharge to the High Plains aquifer in Colorado and New Mexico was about 196,000 acre-feet per year before development of irrigated agriculture. About two-thirds of the recharge occurred in the larger area of the aquifer in Colorado. Recharge rates of about 0.07 inch per year were common in much of the area, but as much as 0.8 to 1.0 inch of water per year might have recharged the aquifer near some streams and between the South Platte and Republican Rivers, where flat-lying sandy soil and larger rates of annual precipitation are common. After agricultural development, infiltration of irrigation water applied to fields that overlie the aquifer became another source of recharge.

Most ground water is discharged from the aquifer by underflow, withdrawal, and evapotranspiration. Water discharges from the aquifer in Colorado and New Mexico by subsurface flow (underflow) into Nebraska, Kansas, Oklahoma, and Texas. Ground-water withdrawal from wells is another major means of discharge. Evapotranspiration by phreatophytes in areas of shallow water table is a minor means of discharge in eastern Colorado and New Mexico.

WATER-LEVEL CONDITIONS

The altitude and configuration of the water table in the High Plains aquifer is most affected by the altitude and configuration of the underlying bedrock surface, the transmissivity of the aquifer, and the rate and distribution of recharge and discharge. As indicated in **figure 72**, large areas of the aquifer are not continuously saturated or are saturated only in isolated channels in the bedrock surface. In other parts of the aquifer, the water table is continuous and slopes eastward at gradients of about 10 to 40 feet per mile. The water-table altitude ranges from about 6,800 feet at the northern edge of the aquifer in Colorado to about 3,600 feet at the eastern boundary of Colorado and New Mexico. Ground water generally moves downgradient at right angles to the water-table contours. A general eastward movement of water is indicated in the figure, but the local direction of movement is complex due to local changes in the bedrock surface and recharge and discharge. Depth to water in the aquifer is less than 50 feet in parts of New Mexico but generally ranges from 50 to 400 feet (**fig. 73**).

Water levels in the High Plains aquifer have declined in most places since irrigation withdrawal became widespread. Predevelopment to 1980 water-level declines exceeded 25 feet in parts of Colorado and 100 feet in parts of New Mexico (**fig. 74**). Water-level changes were between 10 feet of rise and 10 feet of decline over about 78 percent of the aquifer in Colorado and about 62 percent of the aquifer in New Mexico.

The decrease in saturated thickness caused by the water-level declines produced a reduction of about 15,000,000-acre-feet in the volume of ground water in storage in the aquifer in Colorado and New Mexico. In Colorado, about 5 percent of the original volume of ground water in storage was withdrawn between predevelopment and 1980. In New Mexico, about 16 percent of the original volume of ground water in storage was withdrawn between predevelopment and 1980. The decrease in saturated thickness exceeded 25 percent of the original saturated thickness in some south-central parts of eastern New Mexico. Saturated-thickness decreases generally did not exceed 25 percent in eastern Colorado and were less than 10 percent in most areas. The saturated thickness of the aquifer ranges from less than 1 foot, primarily near the western aquifer boundary, to about 400 feet in eastern Colorado (**fig. 75**). The saturated thickness is irregular because of the uneven surface of the base of the aquifer (**fig. 71**) and differences in water-table altitude (**fig. 72**).

In 1990, the part of the High Plains aquifer in Colorado contained about 108,000,000 acre-feet of recoverable ground water in storage. In New Mexico, the aquifer contained about 47,000,000 acre-feet. The 155,000,000 acre-feet in eastern Colorado and eastern New Mexico was only about 5 percent of the total recoverable water in the High Plains aquifer in the eight-State area where it is present.

AQUIFER CHARACTERISTICS

Hydraulic conductivity is a measure of the ease with which sediments can transmit water. The average hydraulic conductivity of the High Plains aquifer in Colorado and New Mexico is about 60 feet per day and ranges from less than 1 to more than 100 feet per day (**fig. 76**). Differences in hydraulic conductivity are the result of differences in the particle size, shape, sorting, and cementation of the aquifer materials. The High Plains aquifer in eastern Colorado and eastern New Mexico has an average saturated thickness of about 75 feet; the average transmissivity of the aquifer is about 4,500 feet squared per day.

GROUND-WATER QUALITY

The agricultural economy of the High Plains is based on the availability of large quantities of ground water of quality suitable for irrigation. Large concentrations of dissolved solids in the water can retard plant growth. Most crops can tolerate water with as much as 500 milligrams per liter dissolved solids and can tolerate water with 500 to 1,500 milligrams per liter or more if the soils are well drained.

The dissolved-solids concentration of water in the aquifer in eastern Colorado and eastern New Mexico generally is less than 500 milligrams per liter but exceeds 1,000 milligrams per liter in a small area of Colorado (**fig. 77**). Concentrations less than 250 milligrams per liter in northeastern Colorado are the result of relatively large rates of recharge in areas of sandy soil that contains few soluble minerals. The area with large dissolved-solids concentrations north of the Arkansas River is likely caused by dissolution of gypsum (calcium sulfate) in the Upper Cretaceous bedrock that underlies the aquifer. Ground water in this area contains more than 1,000 milligrams per liter of dissolved sulfate. The area with large dissolved-solids concentrations in southeastern New Mexico is likewise due to the effects of the underlying bedrock. In this area, Lower Cretaceous, Jurassic, and Triassic rocks that underlie the High Plains aquifer contain highly mineralized water that may discharge into and degrade the water quality of the High Plains aquifer.

Most ground water in the aquifer is a calcium magnesium bicarbonate type. In parts of southeastern New Mexico and in the area with large dissolved-solids concentration north of the Arkansas River in Colorado, dissolved sulfate is the principal anion, and local ground water is either a calcium magnesium sulfate or a bicarbonate sulfate type.

Figure 72. The water table in the High Plains aquifer slopes gently eastward, and water moves from Colorado and New Mexico toward Nebraska, Kansas, Oklahoma, and Texas.

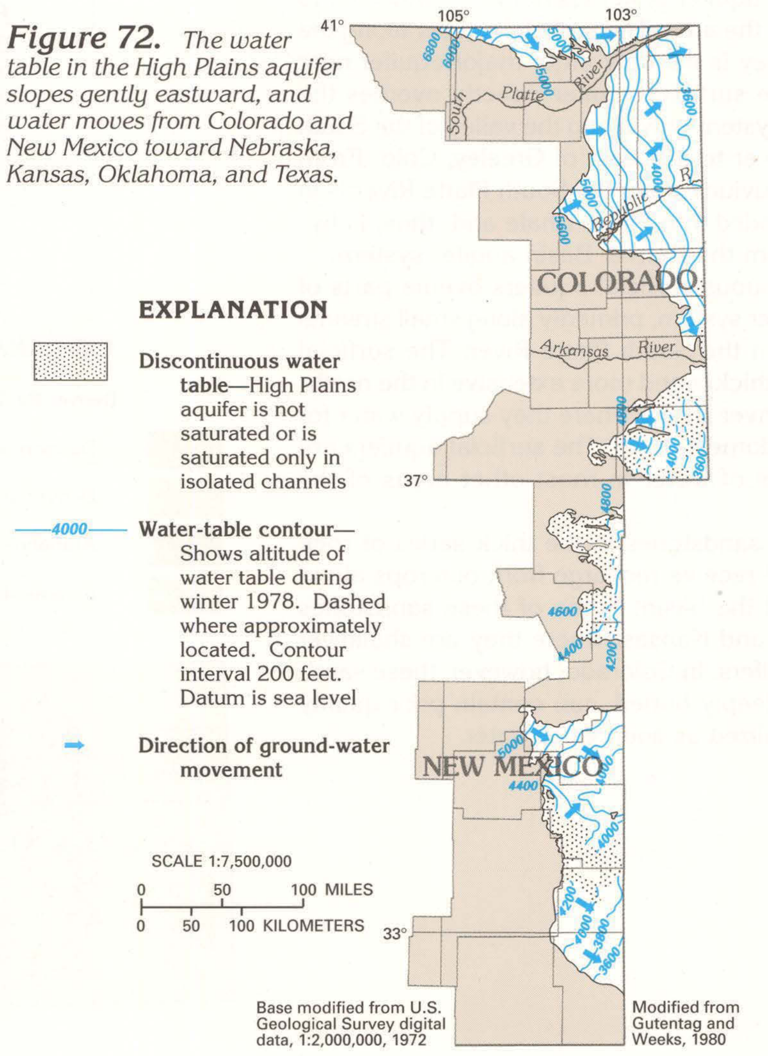


Figure 73. The depth to water in the High Plains aquifer in 1980 generally was less than 200 feet but exceeded 300 feet in some areas.

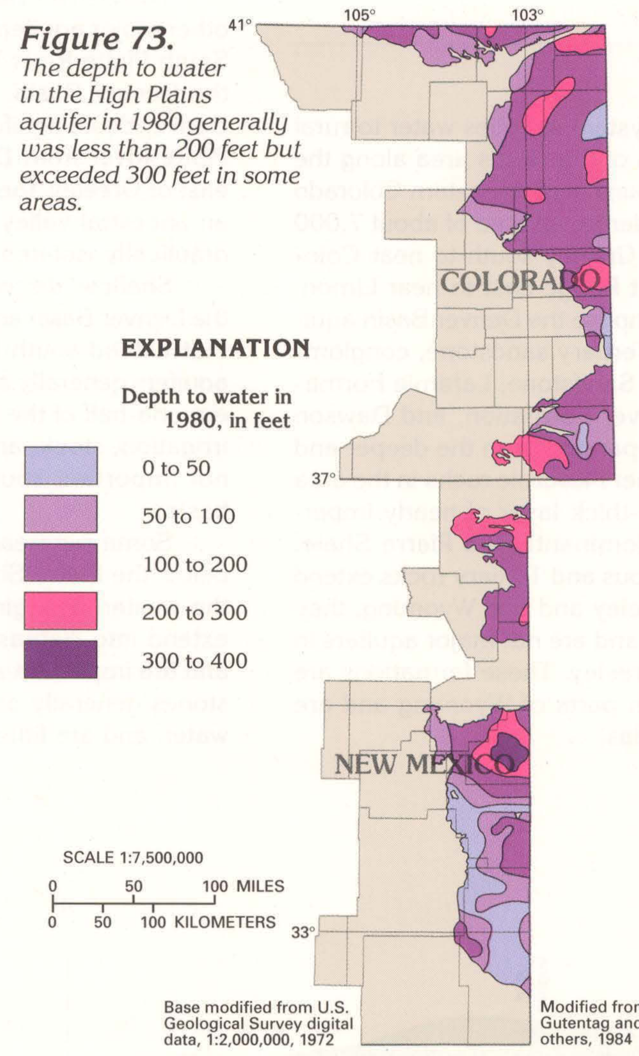
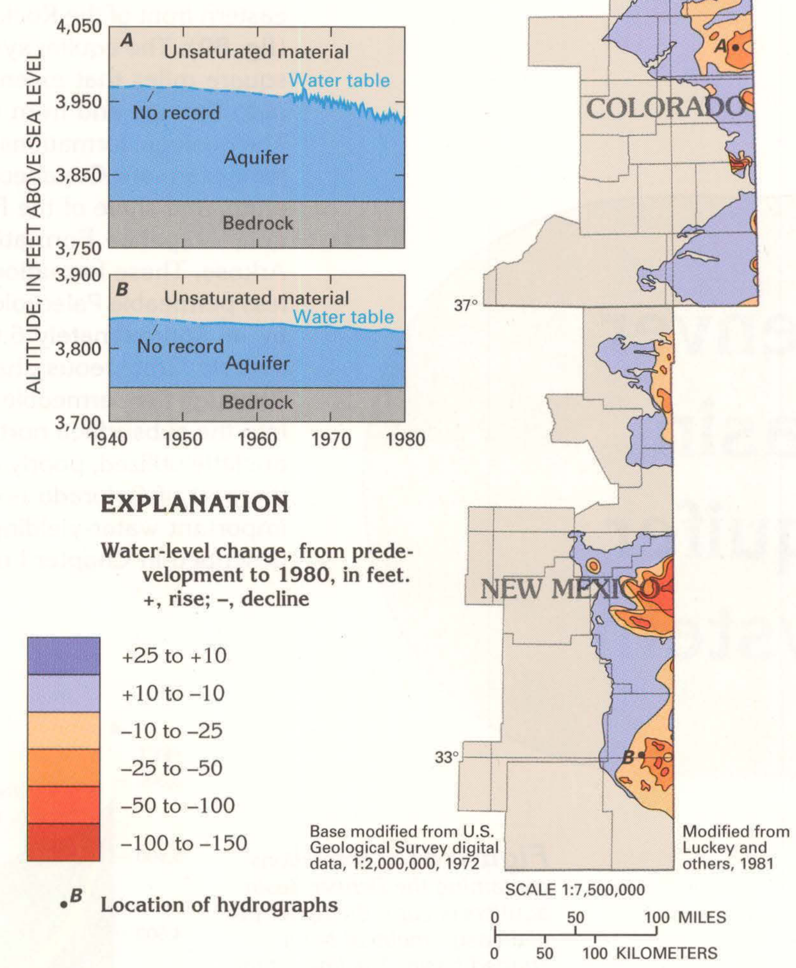


Figure 74. Water-level declines in the High Plains aquifer in eastern Colorado and eastern New Mexico generally ranged from 0 to 100 feet from predevelopment to 1980.



FRESH GROUND-WATER WITHDRAWALS

The development of irrigation on the High Plains has transformed the area into one of the principal agricultural regions of the Nation. Irrigation with ground water began in the 1800's with the use of windmill-powered pumps; large-scale ground-water withdrawal did not begin until after the drought of the 1930's. Drought, technological advances in well drilling and pumping equipment, and favorable economic factors led to the large-scale irrigation of nearly level land overlying a shallow water table. By 1949, from 0 to 10 percent of the land that overlies the aquifer in Colorado and New Mexico (**fig. 78**) was irrigated with ground water. Further advances in drilling and pumping technology and the increasing availability of low-cost energy enabled development of ground water in areas of greater depth to water. The introduction of center-pivot irrigation systems in the 1960's made irrigation of rolling terrain and sandy soils practical. In 1978, the increase in irrigated acreage in the northern part of the High Plains (**fig. 78**) primarily was the result of center-pivot technology. By 1980, the High Plains aquifer in Colorado supplied water to about 770,000 acres from about 4,000 wells. In New Mexico, the aquifer supplied water to about 320,000 acres from about 6,000 wells. Ground-water withdrawal during 1980 was about 985,000 acre-feet in Colorado and about 519,000 acre-feet in New Mexico. About 99 percent of the water withdrawn from the High Plains aquifer was used for irrigation in 1985.

The 1,504,000 acre-feet of water withdrawn from the High Plains aquifer during 1980 in Colorado and New Mexico greatly exceeded the 196,000 acre-feet per year of natural recharge to the aquifer in these States. Annual withdrawal probably has exceeded annual natural recharge since the mid-1960's. Long-term withdrawal from an aquifer that exceeds recharge will cause a long-term decline in water levels and a decrease in saturated thickness.

Figure 75. In 1980, the saturated thickness of the aquifer was less than 100 feet in most areas of eastern Colorado and eastern New Mexico.

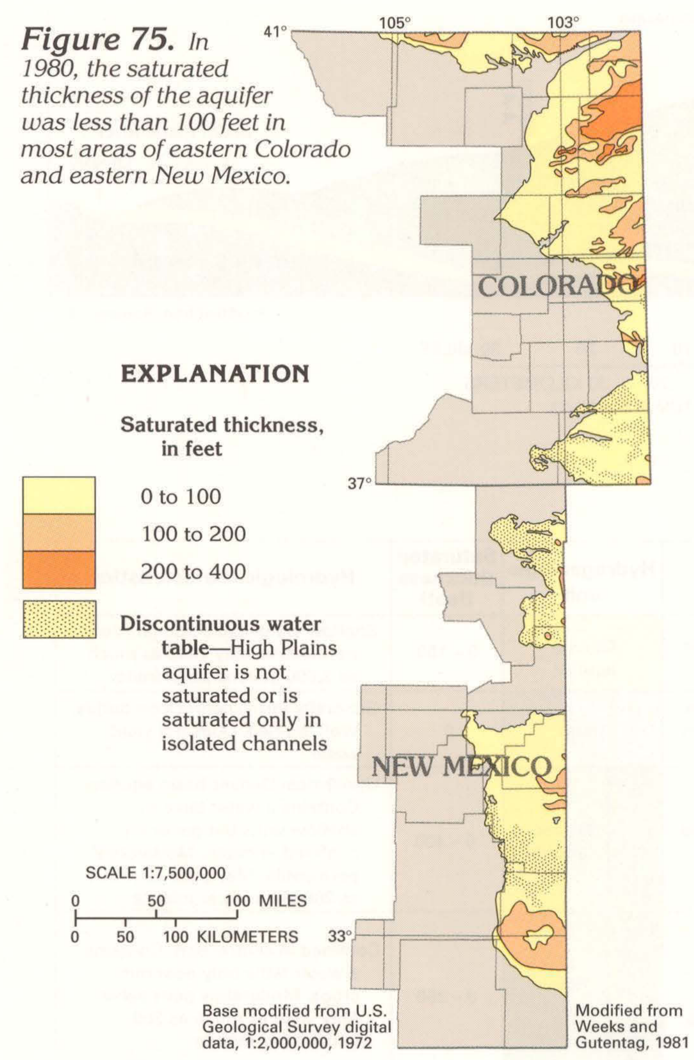
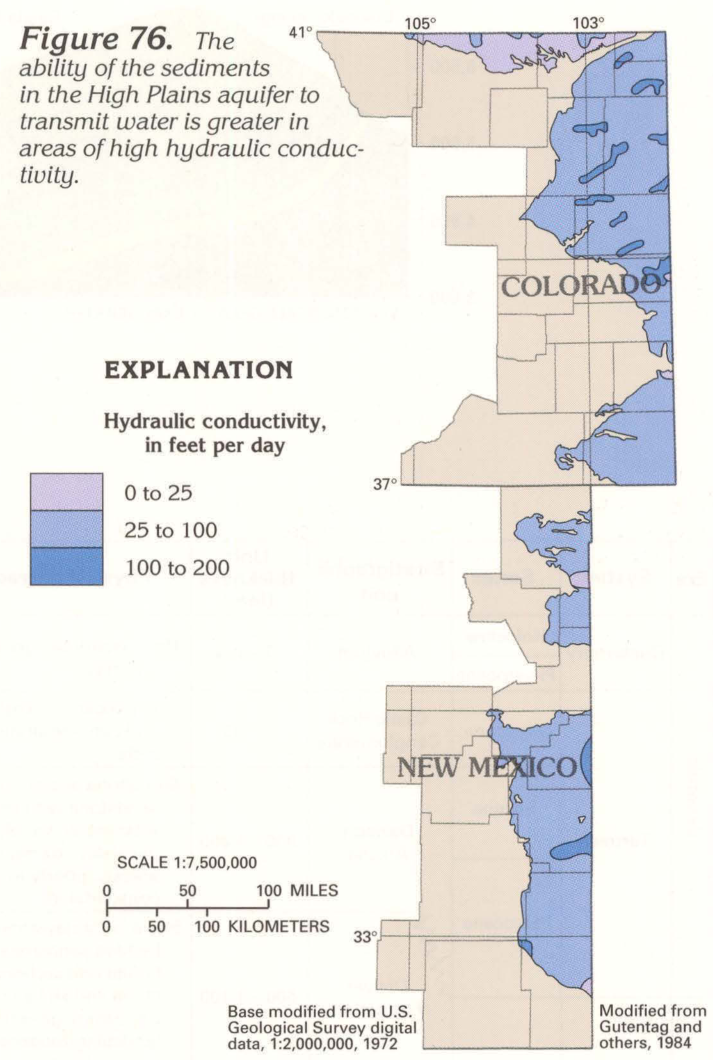


Figure 76. The ability of the sediments in the High Plains aquifer to transmit water is greater in areas of high hydraulic conductivity.



EFFECTS OF WITHDRAWAL

Because water-level declines in the High Plains aquifer have been large, they have substantially decreased the saturated thickness of the aquifer in some areas. As water levels decline, costs to obtain water increase as the result of the need for deeper wells, larger pumps, and larger energy expenditure to lift the water to the surface. As saturated thickness decreases, well yield also decreases, and additional wells can be required to maintain a constant rate of withdrawal. Although several factors are involved, a well capable of producing 250 gallons per minute can irrigate about 40 acres; a well capable of producing 750 gallons per minute can irrigate about 160 acres and effectively operate a center-pivot irrigation system on a quarter section of land. Potential well yields of more than 750 gallons per minute may be obtainable in parts of eastern Colorado and eastern New Mexico (**fig. 79**). As the cost of water increases, irrigated agriculture becomes less economical, and the future of this important agricultural area could become uncertain.

Figure 77. The dissolved-solids concentration of water in the High Plains aquifer in eastern Colorado and eastern New Mexico generally ranges from less than 250 to about 1,000 milligrams per liter.

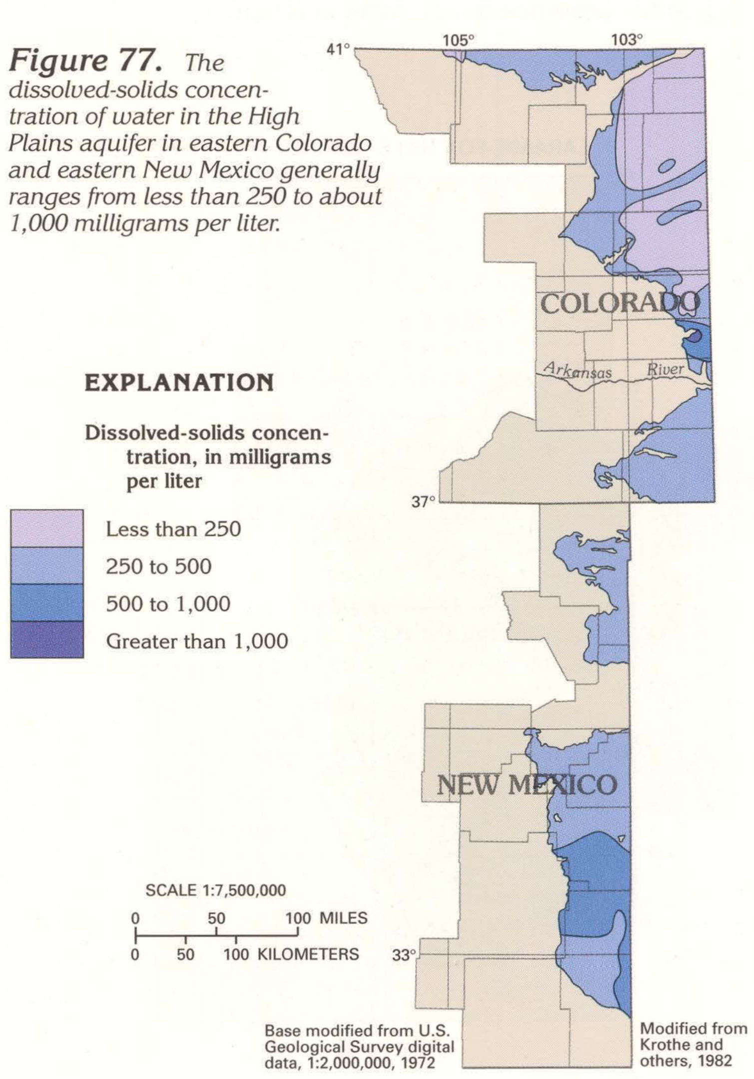


Figure 78. Land irrigated with ground water from the High Plains aquifer increased markedly between 1949 and 1978.

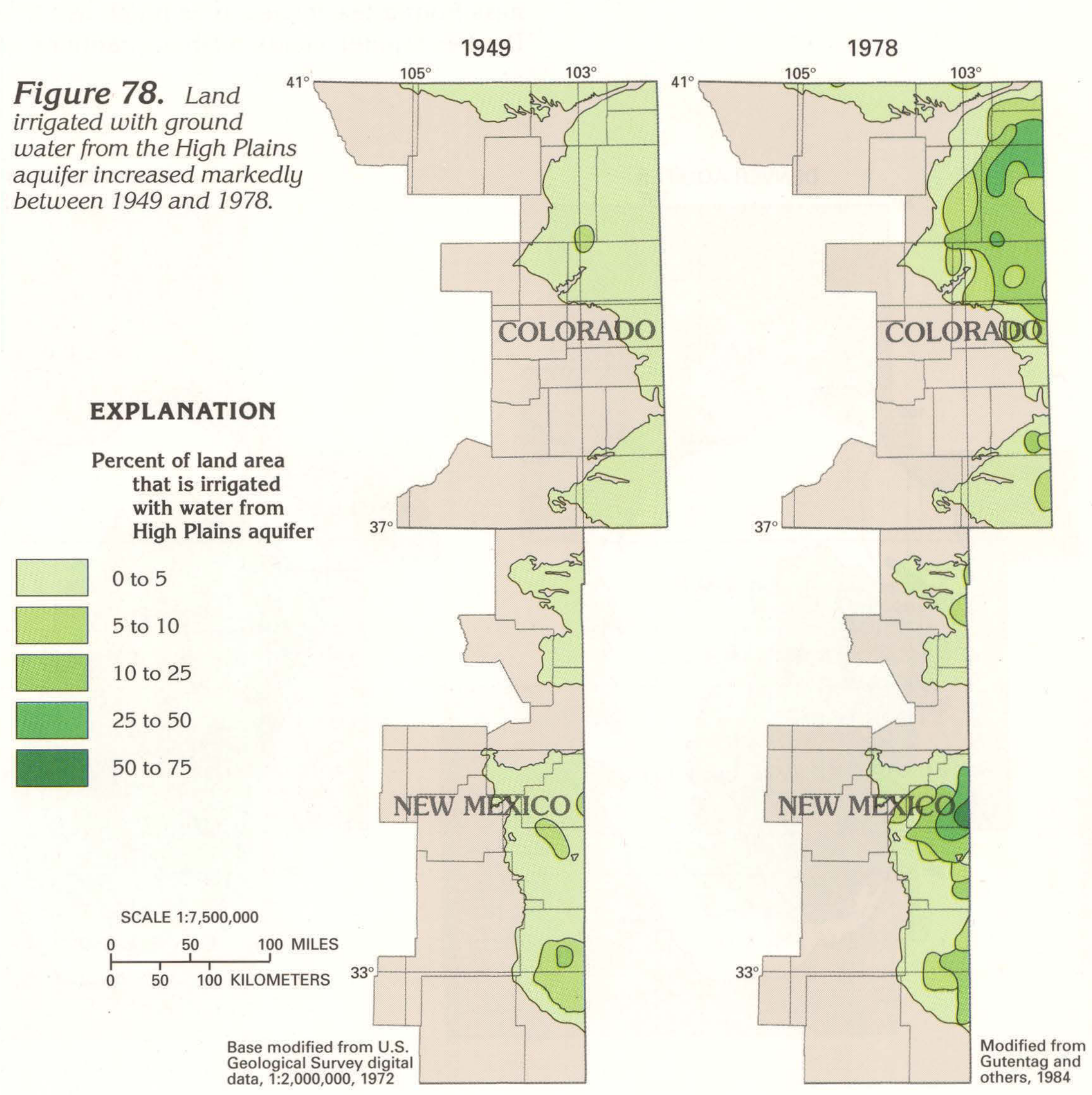
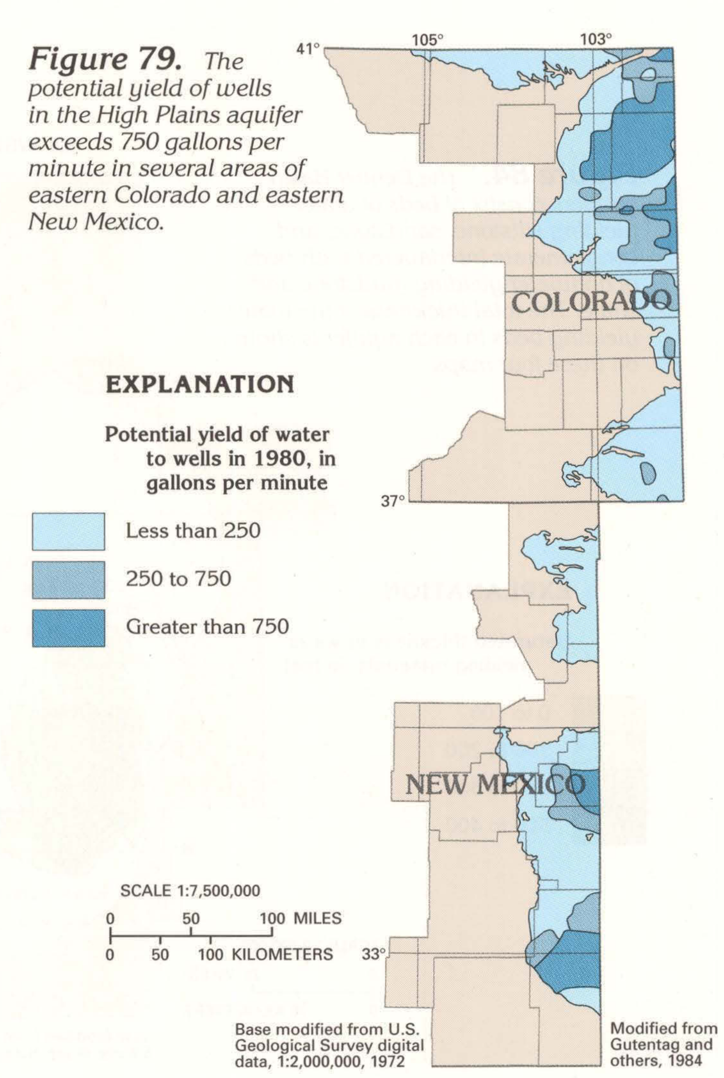


Figure 79. The potential yield of wells in the High Plains aquifer exceeds 750 gallons per minute in several areas of eastern Colorado and eastern New Mexico.



Denver Basin aquifer system

INTRODUCTION

The Denver Basin aquifer system supplies water to rural and suburban residents of much of the plains area along the eastern front of the Rocky Mountains in northeastern Colorado (**fig. 80**). The aquifer system underlies an area of about 7,000 square miles that extends from Greeley south to near Colorado Springs and from the Front Range east to near Limon. The geologic formations that compose the Denver Basin aquifer system are Cretaceous and Tertiary sandstone, conglomerate, and shale of the Fox Hills Sandstone, Laramie Formation, Arapahoe Formation, Denver Formation, and Dawson Arkose. These formations are separated from the deeper and less permeable Paleozoic and other Mesozoic rocks in the area by an approximately 6,000-foot-thick layer of nearly impermeable Cretaceous shale, predominantly the Pierre Shale. Although the permeable Cretaceous and Tertiary rocks extend into the subsurface north of Greeley and into Wyoming, they are little utilized, poorly defined, and are not major aquifers in the part of Colorado north of Greeley. These formations are important water-yielding units in parts of Wyoming and are described in Chapter I of this Atlas.

The Denver Basin aquifer system is not well connected to other major aquifers in the area. The surficial aquifer along the South Platte River Valley is the only other major aquifer near the Denver Basin. The surficial aquifer directly overlies the Denver Basin aquifer system only along the valley of the South Platte River from Denver to just east of Greeley, Colo. From east of Greeley, the alluvium along the South Platte River is in an ancestral valley eroded into Pierre Shale and, thus, is hydraulically isolated from the Denver Basin aquifer system.

Shallow, discontinuous surficial aquifers overlie parts of the Denver Basin aquifer system, primarily along small streams that extend south from the South Platte River. The surficial aquifers are thicker and more extensive in the northern one-half of the Denver Basin, where they supply water for irrigation, stock, and domestic use. The surficial aquifers are not important sources of water in most other areas of the basin.

Some permeable sandstones in the thick series of rock below the Pierre Shale receive recharge from outcrops along the western margin of the basin. Some of these sandstones extend into Nebraska and Kansas, where they are shallower and are important aquifers. In Colorado, however, these sandstones generally are deeply buried, can contain poor-quality water, and are little utilized as sources of water.

Figure 80. *The Denver Basin aquifer system consists of four aquifers that underlie the plains of Colorado to the east of the Rocky Mountains.*

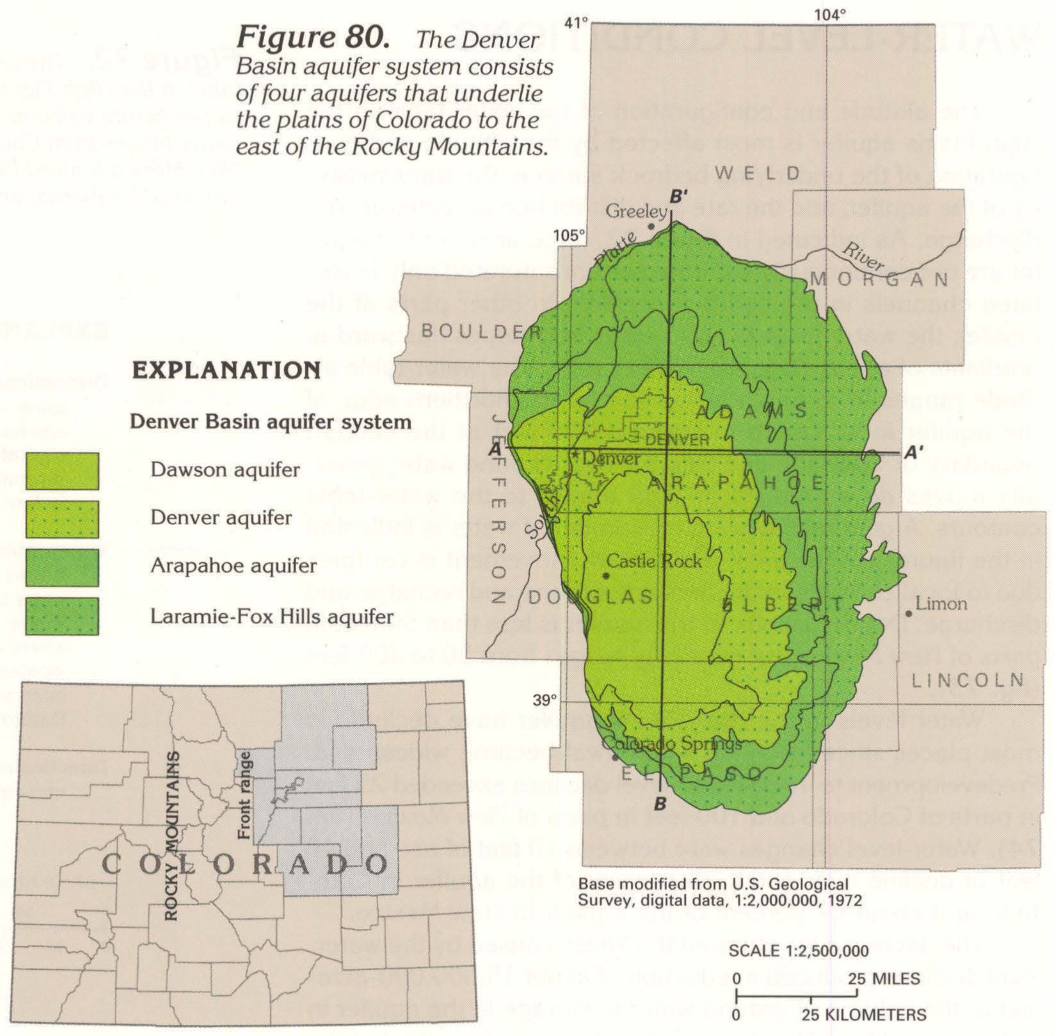


Figure 81. *Formations containing the Denver Basin aquifers occupy the upper part of an asymmetrical bowl-shaped basin. The lines of the sections are shown in figure 80.*

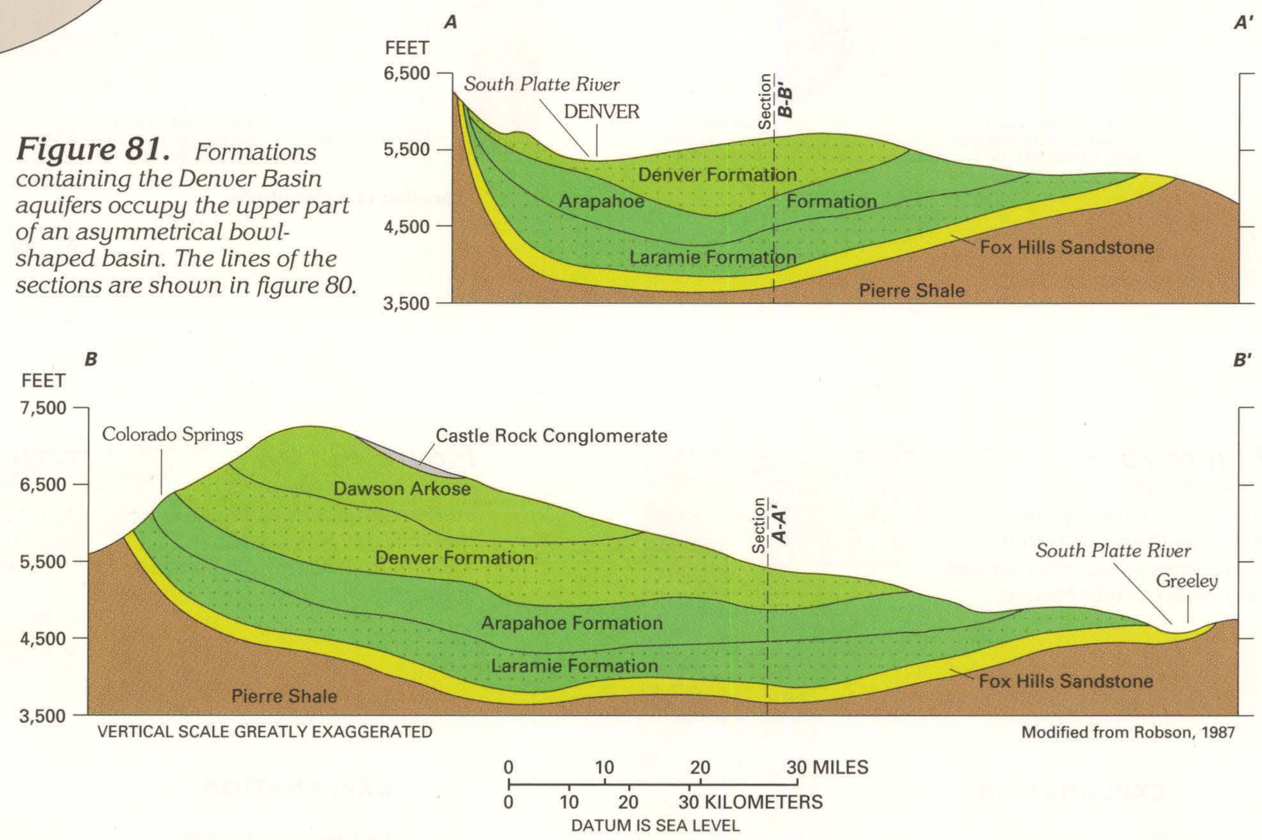


Figure 82. *Bedrock formations dip steeply into the subsurface along the margin of the Denver Basin west of Denver, Colo. These variously colored beds were originally deposited as horizontal layers, and have been bent upward into their present position by the uplift of the Rocky Mountains.*

Robson, 1989

Era	System	Series	Stratigraphic unit	Unit thickness (feet)	Physical characteristics		Hydrogeologic unit	Saturated thickness (feet)	Hydrologic characteristics
Cenozoic	Quaternary	Holocene / Pleistocene	Alluvium	0 – 125	Unconsolidated gravel, sand, silt, and clay		Alluvial aquifer	0 – 100	Shallow water-table aquifer. Very permeable. May yield as much as 3,000 gallons per minute
	Tertiary	Oligocene	Castle Rock Conglomerate	0 – 50	Fine to coarse arkosic sandstone and conglomerate. Exposed in cliffs		none	0	Generally forms caprock on buttes. Well drained. Does not yield water
		Eocene	Dawson Arkose	800 – 1,400	Sandstone and conglomeratic sandstone with interbedded siltstone and shale. Sandstone generally coarse, quartzose, arkosic, poorly to well consolidated	Denver Basin aquifer system	Dawson aquifer	0 – 400	Uppermost Denver basin aquifer. Contains a water table in shallow units but generally confined at depth. Moderately permeable. May yield as much as 200 gallons per minute
		Paleocene	Denver Formation	600 – 1,100	Shale, silty claystone, and interbedded sandstone. Beds of lignite and carbonaceous siltstone and shale common. Sandstone generally andesitic, lenticular, moderately consolidated		Denver aquifer	0 – 350	Confined in central part. Contains a water table only near outcrops. Moderately permeable. May yield as much as 200 gallons per minute
Mesozoic	Cretaceous	Upper Cretaceous	Arapahoe Formation	400 – 700	Sandstone, conglomeratic sandstone, and interbedded shale and siltstone. Sandstone generally quartzose, fine to coarse, poorly to well consolidated		Arapahoe aquifer	0 – 400	Confined in central part. Contains a water table only near outcrops. Most permeable of Denver basin aquifers. May yield as much as 700 gallons per minute
			Laramie Formation	100 – 600	Upper part shale, silty shale, siltstone, and interbedded fine sandstone. Bituminous coal seams		Laramie confining unit	0 – 400	Shale is impermeable
					Lower part sandstone and shale. Sandstone fine to medium, friable, carbonaceous		Laramie-Fox Hills aquifer	0 – 300	Lowermost Denver basin aquifer. Confined in central part. Contains a water table only near outcrops. Moderately permeable. May yield as much as 300 gallons per minute
			Fox Hills Sandstone	100 – 200	Sandstone and siltstone interbedded with shale. Sandstone generally very fine to fine, poorly consolidated				
			Pierre Shale	5,000 – 7,000	Shale, silty, dense, calcareous, fossiliferous		Lower confining unit	0	Impermeable. Forms base of Denver basin aquifer system

Figure 83. *The geologic and hydrologic characteristics of the four Denver Basin aquifers are summarized in this diagram.*

HYDROGEOLOGIC UNITS

The Denver Basin aquifer system consists of four aquifers (**fig. 80**) that are present in five geologic formations. As shown in the geologic sections in **figure 81**, the five formations form a layered sequence of rock in an elongated, asymmetrical bowl-shaped structural depression. The structure of the formations is asymmetrical because rocks near the western edge of the basin dip more steeply than rocks near the eastern edge of the basin (**figs. 81, 82**). The differences in dip and the overall shape of the basin are the result of the uplift of the Rocky Mountains, which followed deposition of most of the Cretaceous formations in the Denver Basin.

The Dawson Arkose (**fig. 83**) contains the Dawson aquifer and is the uppermost and least extensive water-yielding formation in the Denver Basin. The Dawson aquifer extends over an area of about 1,200 square miles between Denver and Colorado Springs. The sediments that form the Dawson aquifer primarily consist of coarse-grained, poorly to well-consolidated sandstones interbedded with conglomerate, siltstone, and shale. Individual conglomerate or sandstone beds commonly are lens-shaped and range in thickness from a few inches to as much as 200 feet. Saturated thickness of the aquifer is 300 to 400 feet in the central part of the formation (**fig. 84**).

The Denver Formation contains the Denver aquifer (**fig. 83**), which extends through an area of about 3,000 square miles and underlies the city of Denver, Colo. The Denver Formation is a 600- to 1,100-foot-thick sequence of moderately consolidated, interbedded shale, claystone, siltstone, and sandstone, in which coal and fossilized plant remains are common. Water-yielding layers of sandstone and siltstone occur in poorly defined irregular beds that are dispersed within relatively thick sequences of claystone and shale. Individual sandstone and siltstone layers commonly are lens-shaped and range in thickness from a few inches to as much as 50 feet. Although the Denver aquifer yields usable quantities of water to wells, claystone and shale are prevalent in this unit and tend to form a leaky confining layer between the overlying Dawson aquifer and the underlying Arapahoe aquifer.

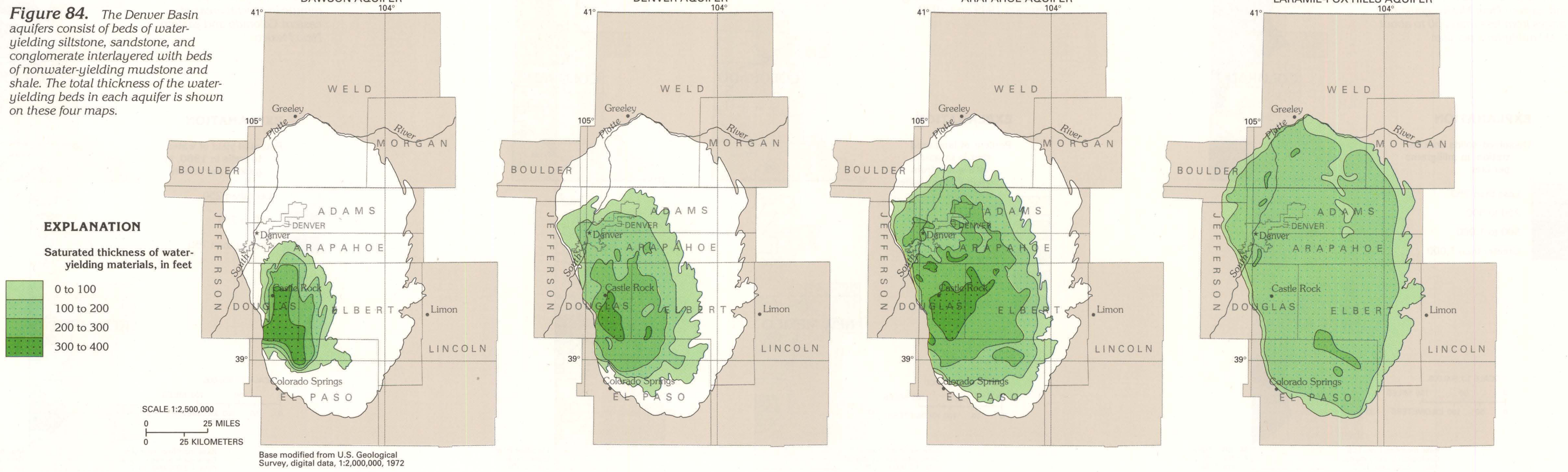
The Arapahoe Formation consists of a 400- to 700-foot-thick sequence of interbedded conglomerate, sandstone, siltstone, and shale. It contains the Arapahoe aquifer (**fig. 83**), which extends over an area of about 4,300 square miles or about two-thirds the area of the Denver Basin aquifer system (**fig. 80**). The top of the Arapahoe aquifer is defined by the base of shale beds in the lower part of the overlying Denver Formation; the base of the Arapahoe aquifer is defined by the top of the shale, coal seams, and thin beds of sandstone and siltstone in the upperpart of the underlying Laramie Formation. The upper part of the Laramie Formation forms a nearly impermeable confining layer that is 300 to 400 feet thick and impedes water movement between the Arapahoe aquifer and the underlying Laramie–Fox Hills aquifer.

In some areas the Arapahoe aquifer can be divided into an upper, somewhat shalier part and a lower, somewhat sandier part. Individual conglomerate and sandstone beds in the aquifer commonly are lens-shaped, moderately consolidated, and range in thickness from a few inches to 30 or 40 feet. The beds are so closely spaced that they form a single hydrologic unit that is 200 to 300 feet thick in some areas.

The Fox Hills Sandstone and sandstones in the lower part of the Laramie Formation form the Laramie–Fox Hills aquifer (**fig. 83**). The Laramie–Fox Hills aquifer underlies all of the approximately 7,000-square-mile Denver Basin. The aquifer is underlain by the nearly impermeable Pierre Shale, which forms the base of the aquifer system. The thickness of the Laramie–Fox Hills aquifer ranges from 0 to about 300 feet (**fig. 84**).

The Laramie–Fox Hills aquifer generally consists of beds of fine- to very fine-grained sandstone or siltstone interbedded with shale. Deeply buried beds of sandstone and siltstone generally are friable and light to medium gray. In outcrops, sandstone and siltstone range from friable to very hard, depending on the presence of iron mineralization.

Figure 84. *The Denver Basin aquifers consist of beds of water-yielding siltstone, sandstone, and conglomerate interlayered with beds of nonwater-yielding mudstone and shale. The total thickness of the water-yielding beds in each aquifer is shown on these four maps.*



RECHARGE AND DISCHARGE

The Denver Basin has a semiarid climate in which potential annual evaporation is about five times larger than annual precipitation. Most precipitation that falls on the land surface runs off in streams, is evaporated from the soil surface, or is consumed by vegetation. However, a small part of the precipitation usually percolates downward and recharges the ground-water system. In the Denver Basin, most recharge occurs in the highland areas between stream channels in the topographically higher southern part of the basin. Precipitation is greater in this area, and the permeable soils derived from the Dawson Arkose enable deep percolation. Recharge in this area can occur on a local and a regional scale. On a local scale, water moves from the highland recharge areas through shallow sandstone beds and discharges in nearby stream valleys. Because it follows short flow paths, this water primarily affects ground-water levels only in a local area. On a regional scale, water moves from the recharge areas into deeper parts of the aquifer and can move great distances through the aquifer before reaching a point of discharge many miles away.

Recharge and discharge also can result from water moving upward or downward through confining units of mudstone or shale located within or between the aquifers. In the central part of the basin, water levels in shallower wells generally are higher than those in deeper wells, creating the potential for water to move downward from the shallower aquifers to the deeper aquifers. Such water-level differences between adjacent aquifers indicate the direction ground water can move vertically, but do not indicate how much movement, if any, is actually occurring. Mudstone and shale beds in the Dawson, Denver, and Arapahoe aquifers allow some vertical water movement through or around the beds. In contrast, shale in the upper part of the Laramie Formation is thick, relatively impermeable, and areally extensive; water generally does not move vertically across this confining unit even if water-level differences exist.

The principal means of ground-water discharge from the Denver Basin aquifers are withdrawal from wells and interaquifer movement of water from the bedrock to overlying alluvial aquifers. Estimated ground-water withdrawal from the bedrock aquifers increased from about 14,000 acre-feet per year during 1960 to about 29,000 acre-feet per year during 1980. Water discharged to alluvial aquifers can contribute to the flow in those aquifers or streams adjacent to them or can be lost to evapotranspiration.

On average, about 5,000,000 acre-feet of water falls as precipitation each year on the Denver Basin (**fig. 85**). About 4,960,000 acre-feet of this water is lost to evaporation, transpiration by plants, or surface runoff. The remaining water, about 40,000 acre-feet, recharges the four Denver Basin aquifers. During 1958–78, about 45,000 acre-feet of water per year was discharged from the aquifers—26,000 acre-feet by natural discharge to alluvial aquifers, springs, and evaporation and about 19,000 acre-feet to pumping wells. A net quantity of about 4,800 acre-feet of water flowed from the Dawson aquifer to the Denver aquifer; a net quantity of about 8,200 acre-feet of water flowed from the Denver aquifer to the Arapahoe aquifer. Thick shale beds prevent most water movement across the upper part of the Laramie Formation and the Pierre Shale.

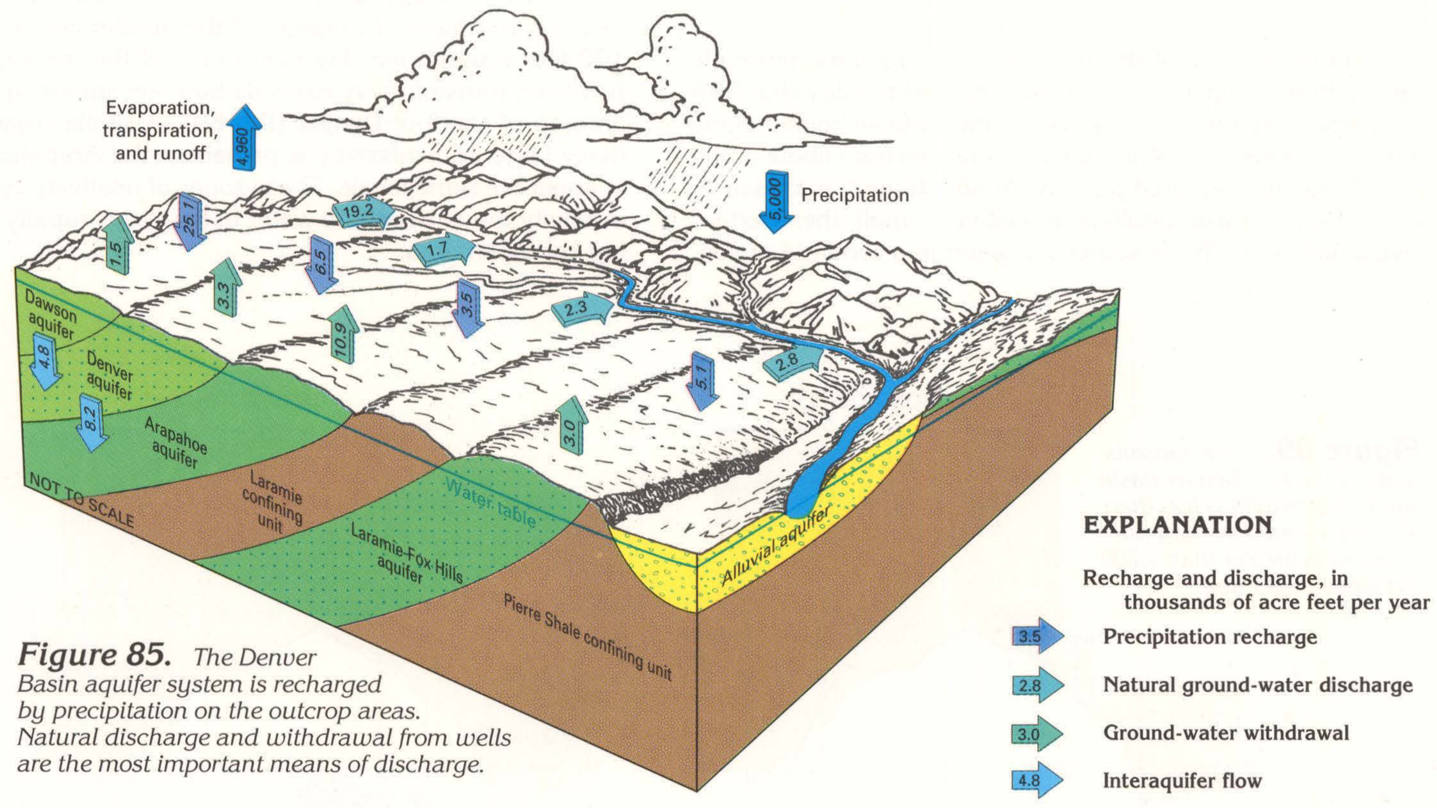


Figure 85. The Denver Basin aquifer system is recharged by precipitation on the outcrop areas. Natural discharge and withdrawal from wells are the most important means of discharge.

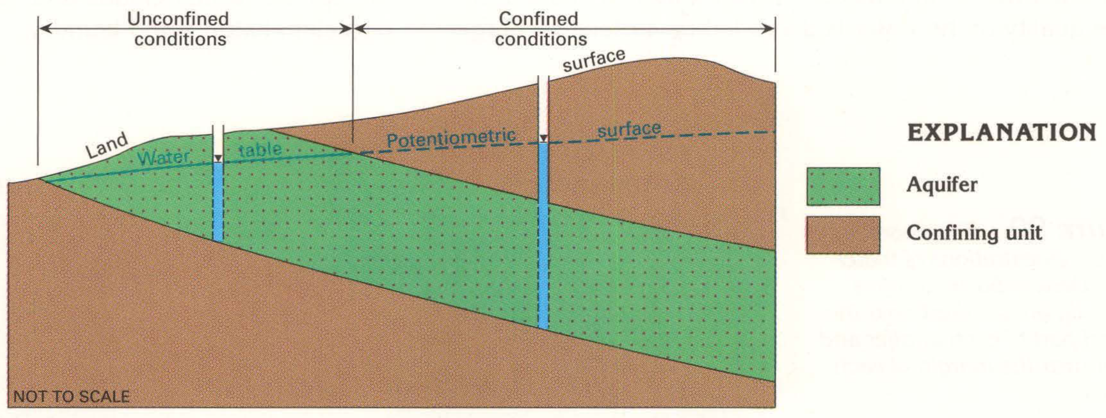


Figure 86. Unconfined conditions exist near the outcrops of water-yielding units in the Denver Basin. Confined conditions exist in the deeper parts of the basin.

WATER-LEVEL CONDITIONS

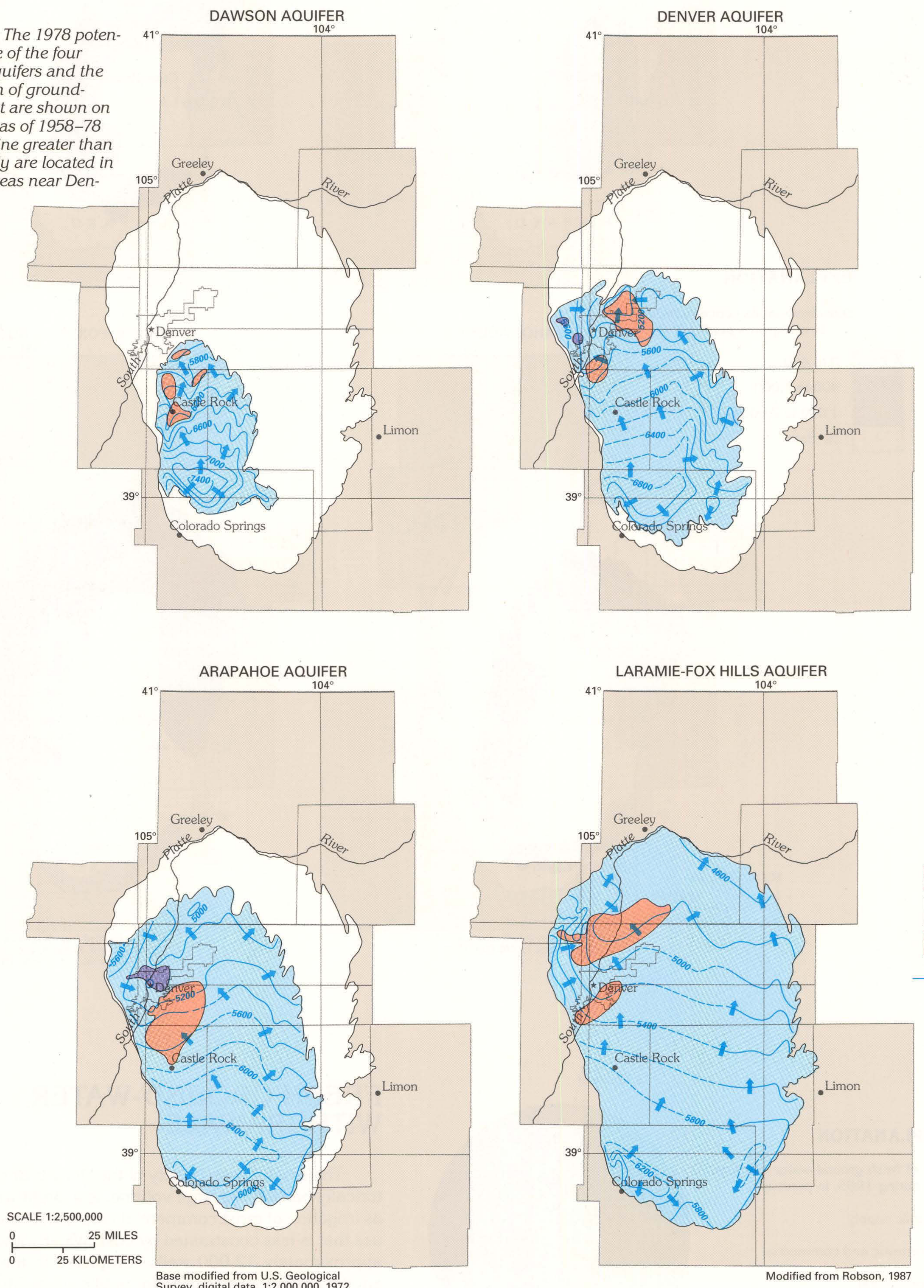
Unconfined and confined conditions are present in the bedrock aquifers of the Denver Basin. The aquifers are unconfined primarily near outcrops (**fig. 86**), where the water table may be at relatively shallow depth. The aquifers are confined in the deeper parts of the basin where confining units restrict vertical movement of water and cause water levels in wells to rise above the top of the aquifer. The surface defined by the water level in wells in either confined or unconfined parts of an aquifer is called a potentiometric surface. (In an unconfined aquifer, the potentiometric surface also is called a water table.)

Depth to water in wells completed in the Denver Basin aquifers generally ranges from 0 to 250 feet, but water levels are 500 to 1,300 feet below land surface in parts of the deeper aquifers between Denver and Castle Rock. The large depth to water and the prevalent mudstone and shale layers in the bedrock formations restrict the effect of surface water on water levels in the deep aquifers. As a result, the potentiometric surface in the deeply buried central part of the Laramie–Fox Hills and Arapahoe aquifers is of uniform shape and slopes gently to the north in most areas (**fig. 87**). The shallow parts of the aquifers are recharged more easily from the land surface, and the potentiometric surface in the Dawson aquifer, for example, is irregular because of recharge from highland areas between stream valleys and discharge to the alluvial aquifers or streams.

Arrows on the potentiometric-surface maps of the Denver Basin aquifers indicate the general direction of ground-water movement. Three major features of water movement in the aquifers are the general northward movement of water from recharge areas in the southern part of the basin toward discharge areas in the northern part of the basin, the convergent flow of water toward the discharge area near the South Platte River Valley north of Denver, and the movement of water into depressions in the potentiometric surface caused by ground-water withdrawal from wells near the Denver metropolitan area.

Denver has a long history of water-level decline in wells. Ground-water withdrawal has caused these water-level declines and the resulting decrease in the volume of ground water in storage. Between 1884 and 1980, about 750,000 acre-feet of water was withdrawn from the basin aquifers in the Denver metropolitan area. Although this quantity is only about 0.3 percent of the approximately 270,000,000 acre-feet of recoverable ground water in storage in the aquifers, it is about 30 percent of the 3,800,000 acre-feet of recharge from precipitation for the entire basin that occurred during this period. Thus, in the metropolitan area, withdrawal greatly exceeded the recharge, and water-level declines have exceeded 500 feet in some wells. The only long-term hydrograph of water-level change in the basin indicates that the water level declined about 400 feet in the Arapahoe aquifer near downtown Denver between 1884 and 1960 (**fig. 88**). Decreased withdrawal has caused a moderate recovery in the water level in this well since 1960. Between 1958 and 1978, water-level declines exceeded 150 feet in parts of the Dawson aquifer, were more than 200 feet in parts of the Denver aquifer, and exceeded 250 feet in parts of the Arapahoe and Laramie–Fox Hills aquifers.

Figure 87. The 1978 potentiometric surface of the four Denver Basin aquifers and the general direction of ground-water movement are shown on these maps. Areas of 1958–78 water-level decline greater than 100 feet primarily are located in the suburban areas near Denver.



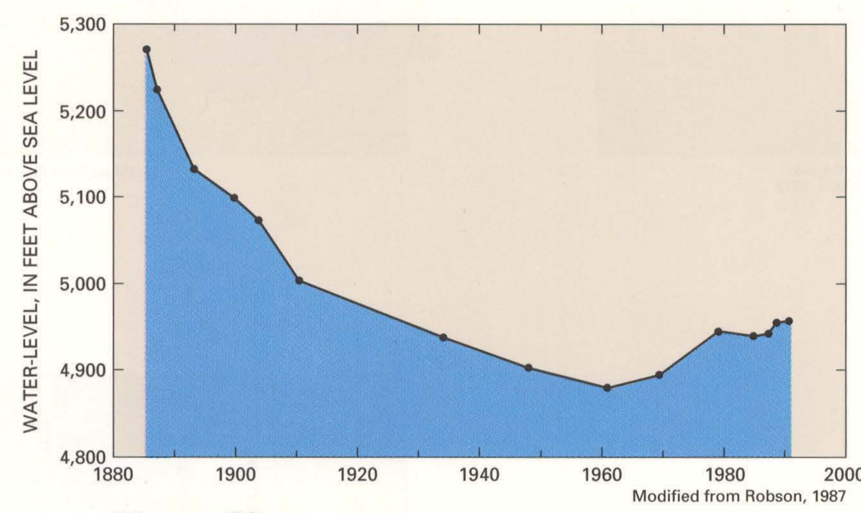


Figure 88. The water level in the Arapahoe aquifer near downtown Denver declined about 400 feet between 1884 (when wells were first drilled in the area) and 1960. Since 1960, decreased local withdrawal has caused the water level to rise.

AQUIFER CHARACTERISTICS

Transmissivity of the Denver Basin aquifers generally ranges from about 10 to 1,000 feet squared per day (**fig. 89**). In comparison, the transmissivity of the surficial aquifer along the South Platte River Valley generally ranges from about 1,000 to 100,000 feet squared per day. Although the transmissivity of the Denver Basin aquifers is relatively small, these extensive aquifers are major sources of water in a semiarid region.

Large differences in transmissivity are present in each Denver Basin aquifer. The transmissivity of the Laramie–Fox Hills aquifer, for example, is less than 5 feet squared per day near the northwestern margin of the aquifer and is less than 100 feet squared per day over much of the remaining area. However, transmissivity exceeds 600 feet squared per day in an area of south of Denver (**fig. 89**). A similar zone of relatively large transmissivity is present in the Arapahoe aquifer in about the same locale. These zones of relatively large transmissivity are of benefit to water users in this rapidly developing suburban area.

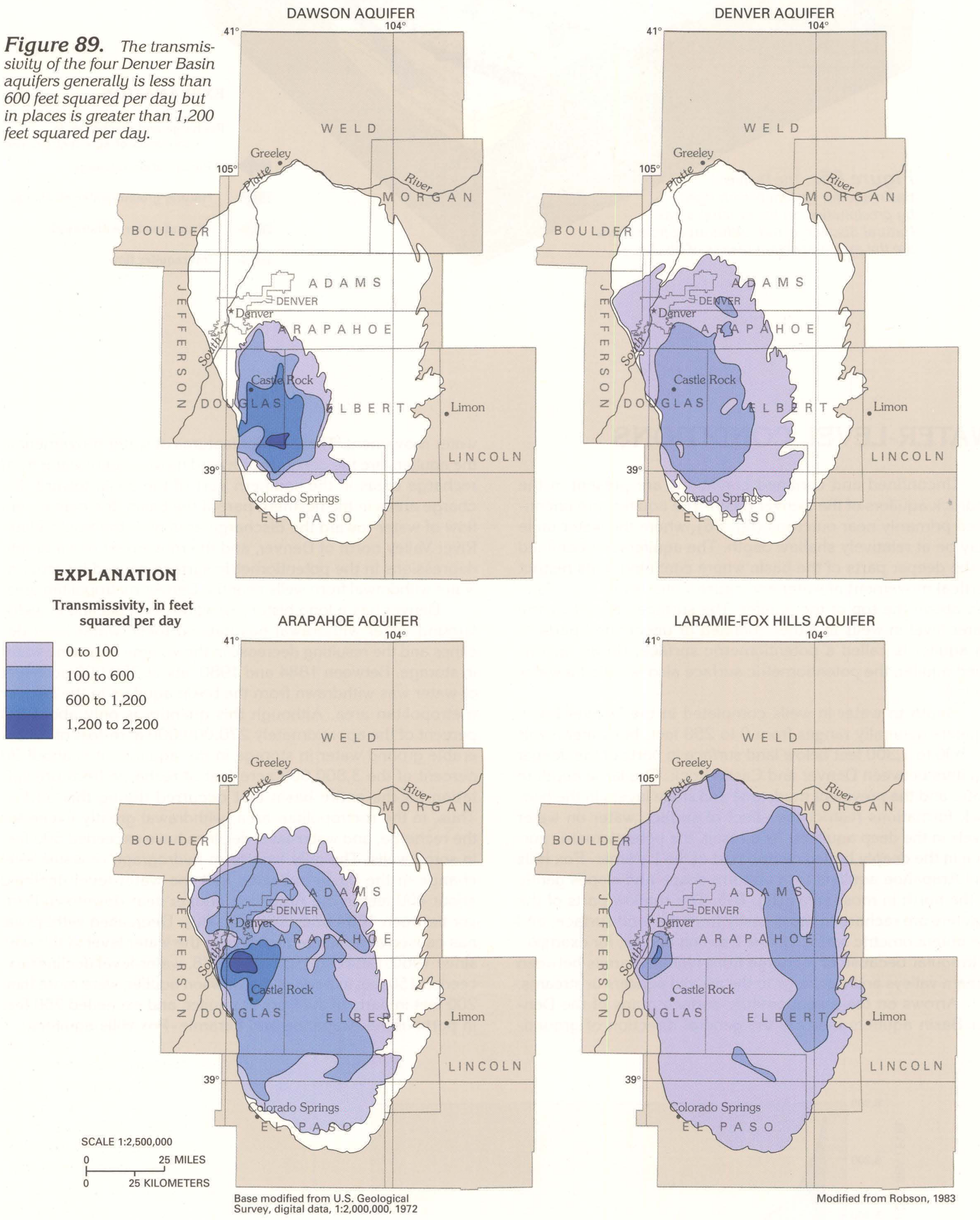
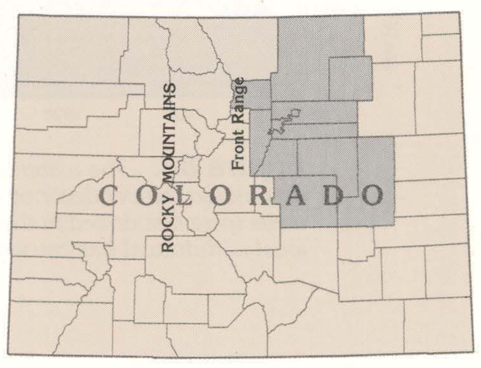


Figure 89. *The transmissivity of the four Denver Basin aquifers generally is less than 600 feet squared per day but in places is greater than 1,200 feet squared per day.*



GROUND-WATER QUALITY

Water in the Denver Basin aquifer system generally has a small dissolved-solids concentration and, in most areas, meets drinking-water regulations established by the U.S. Environmental Protection Agency for public water supplies. The concentration of dissolved solids in ground water ranges from less than 100 milligrams per liter in the Dawson aquifer to about 2,000 milligrams per liter in the Laramie–Fox Hills aquifer. Maps of dissolved-solids concentrations (**fig. 90**) indicate a general trend of increasing concentrations with distance along ground-water flow paths. In the Dawson aquifer, flow paths generally are short, and water is locally recharged by direct infiltration of precipitation. As a result, this aquifer generally contains water with a small dissolved-solids concentration. Water in the Dawson aquifer is a calcium bicarbonate type, has small sulfate concentrations, and is moderately hard. Some water in the Dawson aquifer moves downward and affects the water quality in the underlying Denver aquifer.

Water in the Denver aquifer generally contains about 100 to 1,000 milligrams per liter of dissolved solids, and is a calcium bicarbonate type near the center of the aquifer and a sodium bicarbonate or sodium sulfate type near the margins of the aquifer. Near the center of the Denver aquifer, the water quality is similar to that in the center of the Dawson aquifer. As the calcium bicarbonate water moves through the Denver and underlying aquifers, the water is naturally softened by cation exchange of calcium ions for sodium ions on the surface of clay minerals in the formations. This cation exchange process increases the dissolved-sodium concentration in the water while decreasing the dissolved-calcium concentration. As a result, water in the Denver Basin aquifers generally is softer at greater depth.

As water moves laterally toward the margins of the aquifers, the quality of the water is degraded by surface recharge that contains dissolved sulfate and other chemical constituents leached from soluble minerals in the overlying soil and rock. The increase in sulfate concentrations is particularly evident near the margins of the Denver, Arapahoe, and Laramie–Fox Hills aquifers.

The processes of ion exchange and dissolution cause water in the Arapahoe aquifer to be somewhat softer and have larger sulfate concentrations than water in the Denver aquifer. Water in the Arapahoe aquifer generally is a sodium bicarbonate or sodium sulfate type. The dissolved-solids concentrations of the water generally range from 200 to 1,400 milligrams per liter (**fig. 90**).

Water in the Laramie–Fox Hills aquifer also is a sodium bicarbonate or sodium sulfate type and is soft in the central parts of the aquifer and hard to very hard near the margins of the aquifer. The dissolved-solids concentrations of water in this aquifer range from about 200 to 2,000 milligrams per liter; larger concentrations are near the aquifer margins. Reducing (oxygen-deficient) conditions present in some parts of the Laramie–Fox Hills aquifer allow hydrogen sulfide and methane gasses to exist in the aquifer. When these gasses are present in sufficient concentrations, water pumped from the aquifer may effervesce, have a putrid odor, and be of marginal value for many uses.

Dissolved-iron concentrations generally are between 20 and 200 micrograms per liter in water from the Denver Basin aquifers. However, much larger concentrations (ranging from 7,000 to 85,000 micrograms per liter) have been reported in some water samples; these concentrations differ considerably from well to well, apparently in response to the chemical environment of the aquifer near the well. When water that contains large concentrations of dissolved iron is pumped from a well and exposed to oxygen in the air, the dissolved iron reverts to an insoluble form that is visible as a black to reddish-brown precipitate, which clouds and discolors the water and stains porcelain fixtures and laundry.

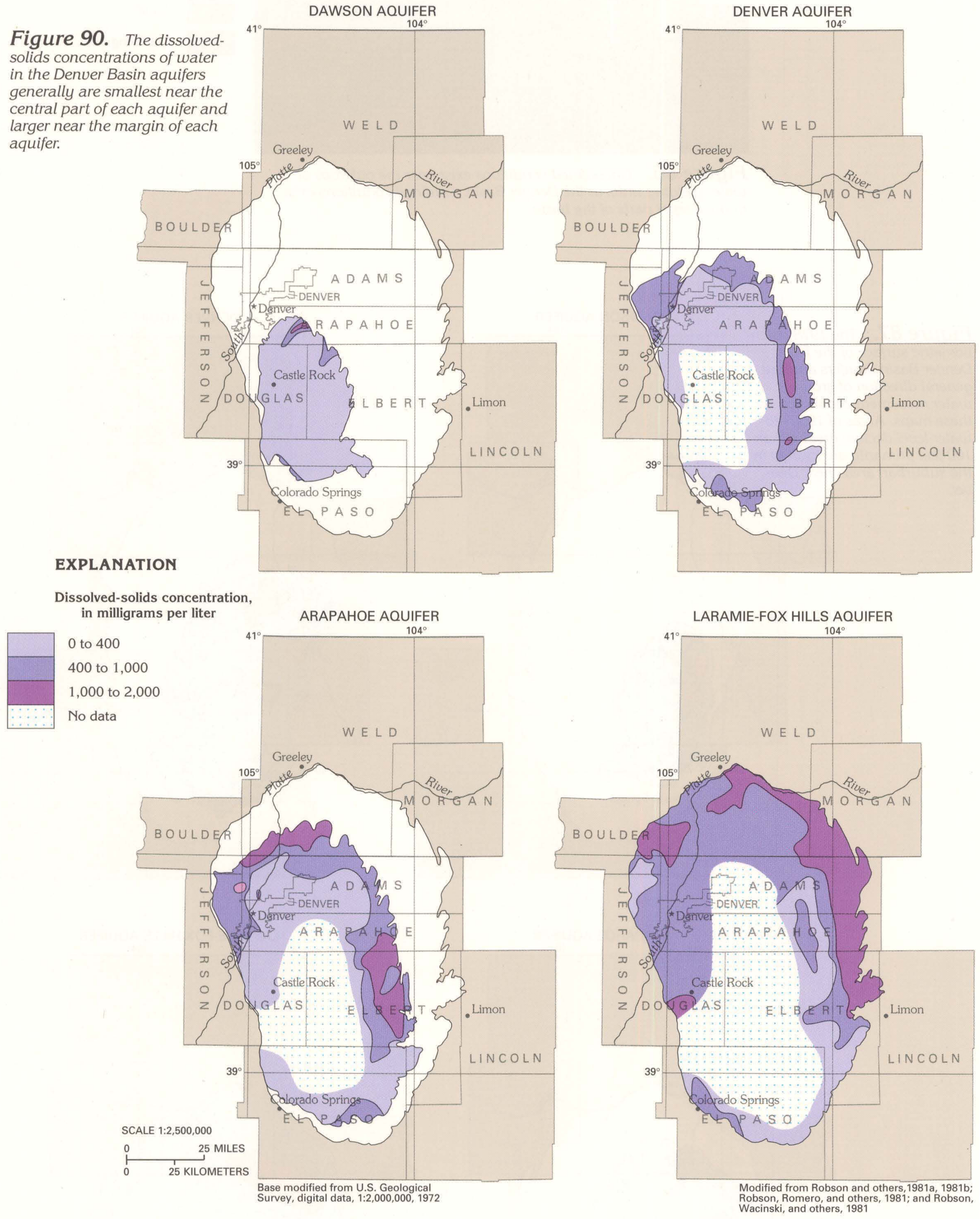


Figure 90. *The dissolved-solids concentrations of water in the Denver Basin aquifers generally are smallest near the central part of each aquifer and larger near the margin of each aquifer.*

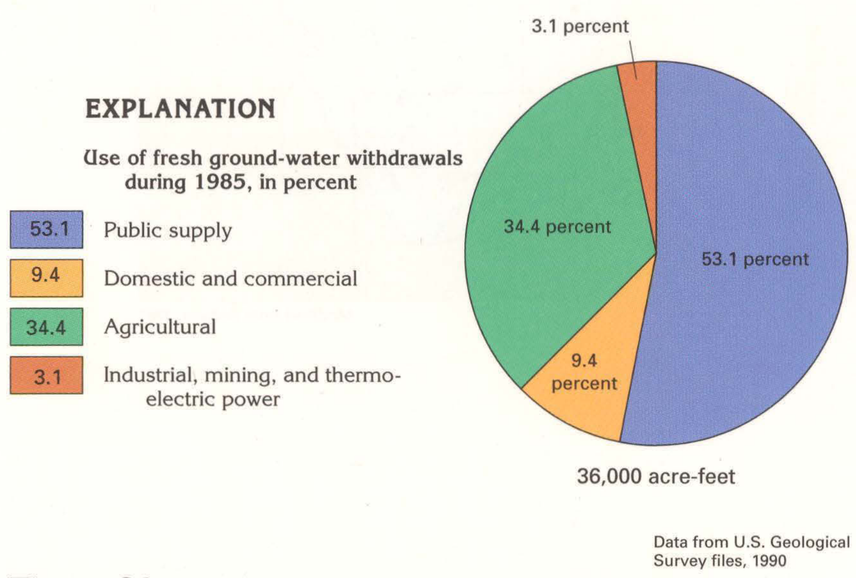


Figure 91. *During 1985, most of the freshwater withdrawn from the Denver Basin aquifers was used for public supply.*

FRESH GROUND-WATER WITHDRAWALS

The small transmissivity of the Denver Basin aquifers historically has limited large volume, low-profit water uses, such as irrigation of most commercial crops, and has enabled water use that is less constrained by cost. Water withdrawn from the approximately 12,000 wells completed in the Denver Basin aquifers primarily was used for public supply during 1985 (**fig. 91**). About 53 percent of the fresh ground water withdrawn was used for public supply; about 34 percent was used for agriculture. Total withdrawals from the four aquifers were 36,000 acre-feet.

Roswell Basin aquifer system

INTRODUCTION

Large volumes of ground water are withdrawn from an alluvial aquifer and an underlying carbonate-rock aquifer in the Roswell Basin of southeastern New Mexico. These two aquifers form the Roswell Basin aquifer system (**fig. 92**). The aquifer system underlies part of the Pecos River and extends through an area of about 2,200 square miles from north of Roswell to northwest of Carlsbad, N. Mex. Although the alluvium covers an area of approximately 1,200 square miles, it is an important aquifer only in about 740 square miles, primarily along the western side of the Pecos River. Ground water in Permian carbonate rocks is present in openings formed by dissolution of part of the limestone, dolomite, and gypsum that are prevalent in the rock. Carbonate rocks underlie an area of about 12,000 square miles between Vaughn, New Mexico, and the New Mexico–Texas State line. The most permeable and extensively utilized aquifer in the carbonate rocks is in the Roswell Basin. This aquifer and the hydraulically connected aquifer in the alluvium have been studied extensively because of the importance of this source of ground water. Other aquifers in the 12,000-square-mile area of carbonate rocks are discontinuous and yield large volumes of water only in local areas or yield slightly saline water due to dissolution of saline minerals that are prevalent in some areas. Although the following description pertains to the geohydrology of the Roswell Basin, the geohydrologic processes described likely are characteristic of the numerous other small carbonate-rock aquifers in the 12,000-square-mile area.

Ground water that flows through joints, fractures, or faults in soluble rocks composed of carbonate (limestone or dolomite) or evaporite (gypsum or halite) minerals can dissolve the surrounding rock and enlarge the openings. Over geologic time, a vast network of interconnected openings can develop in the rock, and large volumes of ground water can flow through the openings. Large caverns also can form; Carlsbad Caverns, which is about 20 miles southwest of Carlsbad, N. Mex., is an example of large solution caverns in a limestone formation. When solution-altered rock becomes sufficiently porous, it may lack the strength to support the weight of overlying materials, and a broad gradual collapse of the rock can occur. The collapse reduces the thickness of the formation and creates additional fractures that are subject to further dissolution. A local, and sometimes rapid, collapse of the rock can produce sinkholes and create a crater like appearance known as karst topography on the land surface.

Unfractured and unaltered carbonate rocks and evaporite minerals generally have low permeability and do not readily yield water to wells. However, dissolution of these rocks can create extremely large or numerous solution openings; altered rocks of this type can be among the most permeable water-yielding formations. A more comprehensive discussion of the hydrology of carbonate rocks is contained in Chapters G and K of this Atlas.

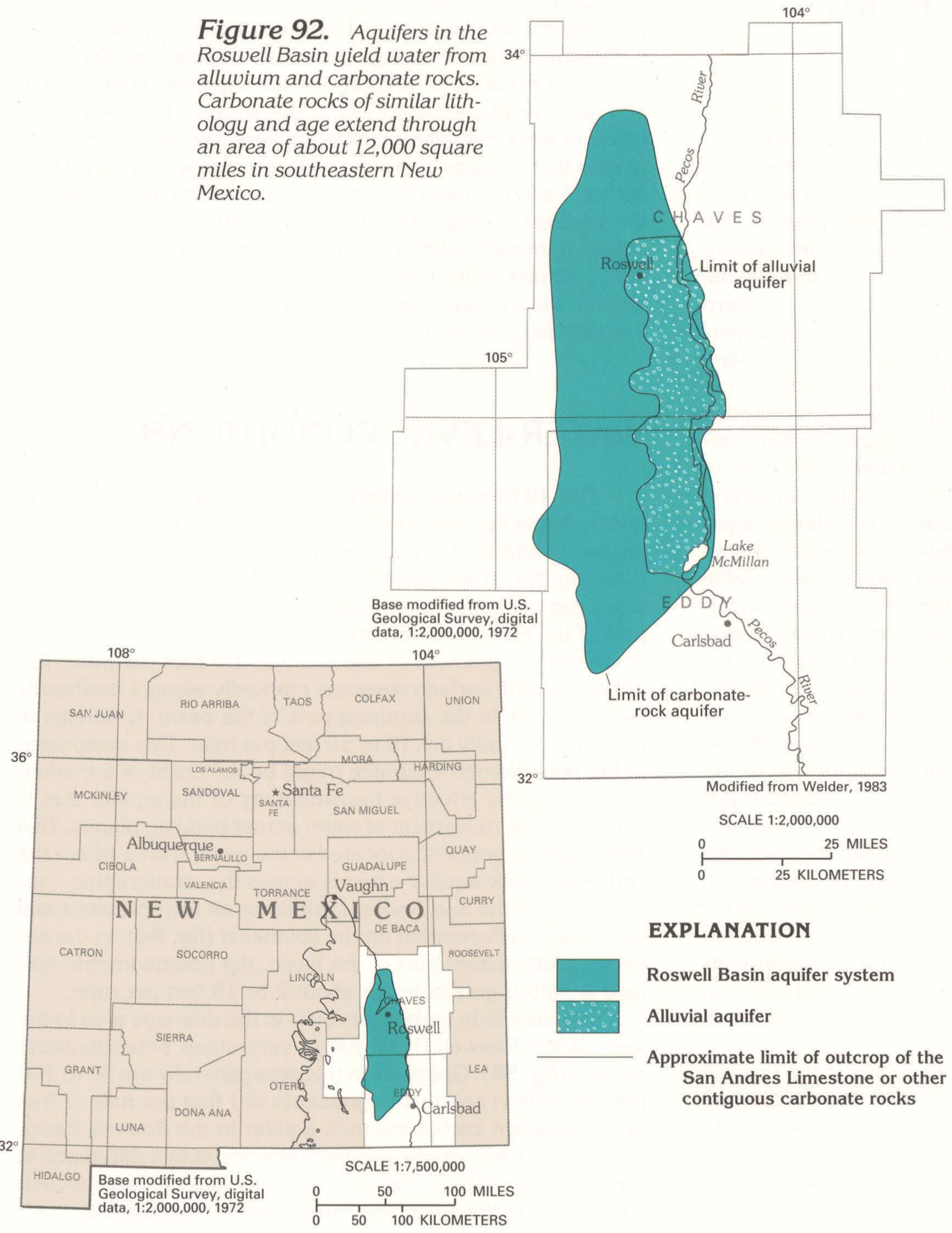


Figure 92. *Aquifers in the Roswell Basin yield water from alluvium and carbonate rocks. Carbonate rocks of similar lithology and age extend through an area of about 12,000 square miles in southeastern New Mexico.*

Era	System	Series	Stratigraphic unit	Unit thickness (feet)	Physical characteristics	Hydrogeologic unit		Saturated thickness (feet)	Hydrologic characteristics
Cenozoic	Quaternary	Holocene / Pleistocene	Alluvium	0 – 300	Unconsolidated gravel, sand, silt, and clay	Roswell Basin aquifer system	Alluvial aquifer	0 – 300	Water-table aquifer. Very permeable. Wells may yield more than 2,000 gallons per minute
Paleozoic	Permian	Upper	Tansil, Yates, and Seven Rivers Formations, undivided	900 – 1,200	Dolomite, limestone, and gypsum interbedded with sandstone and siltstone	Roswell Basin aquifer system	none	—	Generally does not yield water to wells. Where permeable, may yield saline water
Paleozoic	Permian	Upper	Queen and Grayburg Formations, undivided	400 – 800	Dolomite and sandstone interbedded with siltstone and gypsum	Roswell Basin aquifer system	Upper confining unit	0 – 800	Generally low permeability except where fractured or where dissolution of gypsum has created solution openings
Paleozoic	Permian	Upper / Lower	San Andres Limestone and Glorieta Sandstone, undivided	700 – 1,500	Limestone, dolomite, sandstone and gypsum	Roswell Basin aquifer system	Carbonate aquifer	0 – 500	Very permeable aquifer present in solution openings of middle to upper part of San Andres Limestone and lower part of Grayburg Formation. Wells may yield more than 3,000 gallons per minute
Paleozoic	Permian	Lower	Yeso Formation	1,200 – 2,400	Sandstone, siltstone, dolomite, and gypsum	Lower confining unit		0 – 2,400	Lower, unaltered part, of San Andres Limestone, Glorieta Sandstone and Yeso Formation are much less permeable than the carbonate aquifer and form lower confining layer of the aquifer

Modified from Welder, 1983

Figure 93. *The Roswell Basin aquifer system contains two aquifers. An alluvial aquifer that consists of Quaternary sediments overlies a more extensive carbonate-rock aquifer that primarily consists of the San Andres Limestone.*

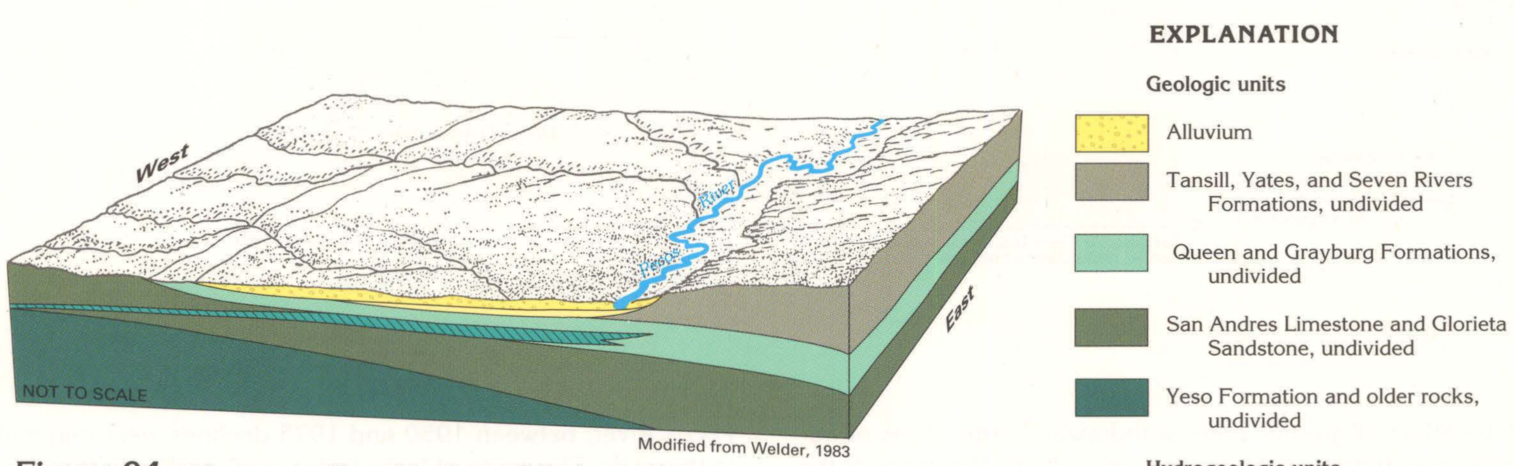


Figure 94. *The two aquifers of the Roswell Basin aquifer system are separated by a thick confining unit. The deeper carbonate-rock aquifer is mostly within the San Andres Limestone but extends upward into the overlying Grayburg Formation to the east.*

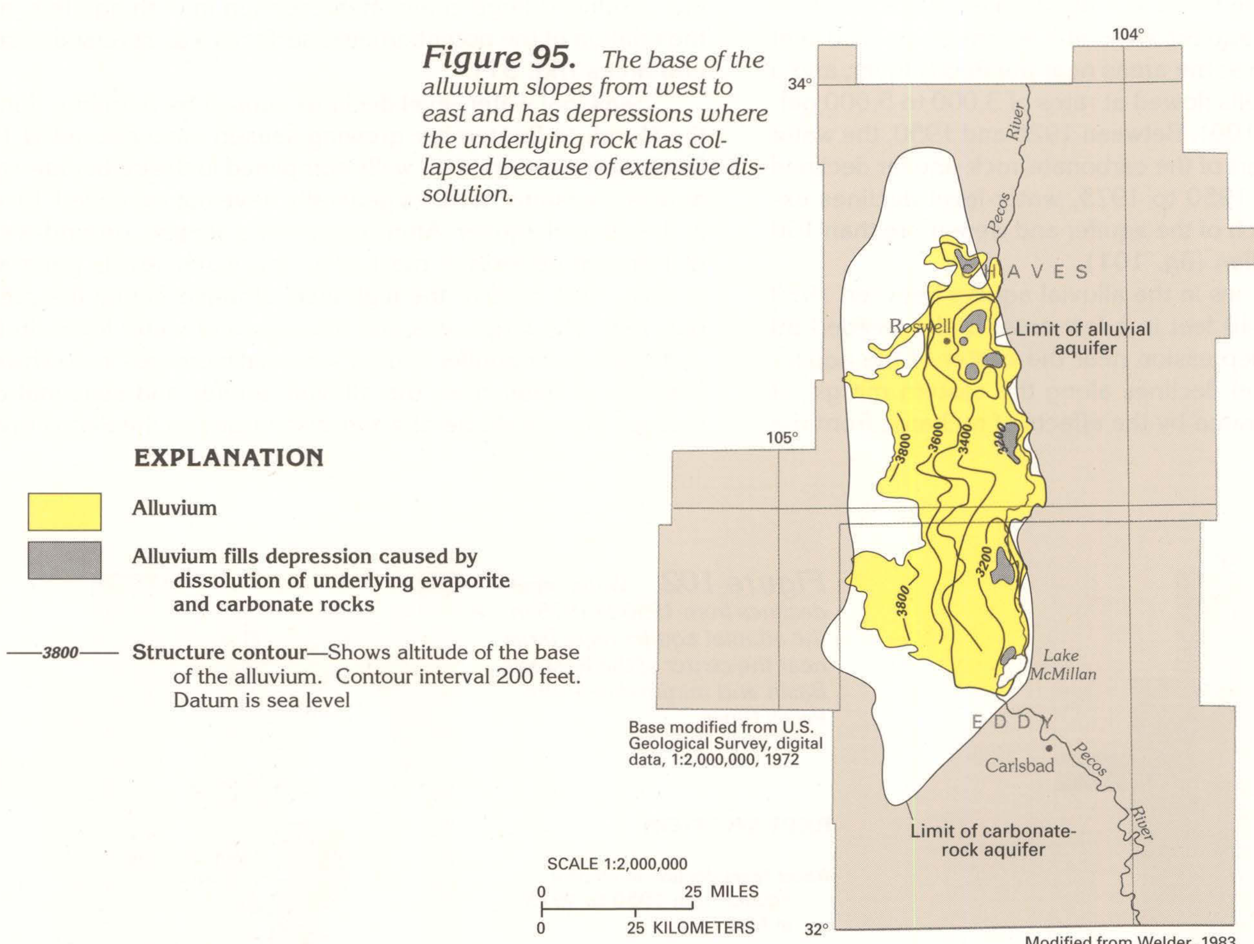


Figure 95. *The base of the alluvium slopes from west to east and has depressions where the underlying rock has collapsed because of extensive dissolution.*

HYDROGEOLOGIC UNITS

The Yeso Formation is the oldest geologic unit that has hydrologic significance in the Roswell Basin (**fig. 93**). The formation consists of sandstone and interbedded siltstone, dolomite, and gypsum and is 1,200 to 1,400 feet thick in the Roswell Basin. Near the land surface, dissolution of dolomite and gypsum locally can produce some permeability in the Yeso, but at depth, dissolution is minimal, and sandstone likely is the principal permeable unit. The permeability of the formation at depth is much smaller than that of the overlying carbonate rocks, and the Yeso functions as a lower, leaky, confining unit for the carbonate-rock aquifer.

Ground water in the carbonate-rock aquifer in the Roswell Basin primarily is present in solution-altered zones in the San Andres Limestone and the overlying Queen and Grayburg Formations (**fig. 94**). These formations dip to the southeast and thin to the northwest. Strata of the Tansil, Yates, and Seven Rivers Formations overlie the Queen Formation on the eastern margin of the Roswell Basin. The San Andres Limestone is at depths ranging from 300 to 1,300 feet along the eastern margin and crops out in a broad band along the western margin of the basin. Limestone and dolomite are the principal rocks in the San Andres, although a 100- to 200-foot-thick sandstone (Glorieta Sandstone) is near the base of the formation. The San Andres Limestone is 1,200 to 1,500 feet thick along the eastern margin of the basin but thins to 700 to 1,000 feet in the northwestern part of the basin where dissolution and collapse have decreased the thickness of the unit.

The Grayburg and Queen Formations overlie the San Andres Limestone. The Grayburg Formation predominantly consists of dolomite and gypsum with interbedded sandstone and shale. The Queen Formation consists of fine-grained sandstone and siltstone with interbedded gypsum. The two formations are 400 to 800 feet thick along the eastern margin of the basin and thin westward as the result of erosion and dissolution. In the southern one-half of the Roswell Basin, the lower part of the Grayburg Formation has been solution altered, and the carbonate-rock aquifer extends upward into this unit.

The San Andres Limestone and the Grayburg and Queen Formations were subject to extensive erosion prior to the deposition of the much younger alluvium in the Roswell Basin. Extensive dissolution of near-surface parts of these carbonate formations likely occurred during this period and may have continued after deposition of the alluvium. A map of the altitude of the base of the alluvium (**fig. 95**) reveals the presence of depressions in the buried surface of the bedrock that are similar to those in areas of karst topography.

The Seven Rivers, Yates, and Tansil Formations overlie the Queen Formation and crop out at higher altitude than the alluvium in the northern part of the basin. In the central and southern parts of the basin, the three formations laterally abut the alluvium or underlie the eastern part of the alluvium. The three formations consist of dolomite, limestone, and gypsum, with interbedded sandstone and siltstone. In the northern part of the basin, dissolution of these formations has been minimal. In the extreme southeastern part of the basin, the three formations have undergone dissolution and contain a local carbonate-rock aquifer.

Quaternary alluvium that consists of unconsolidated gravel, sand, silt, and clay unconformably overlies Permian rocks in the Roswell Basin. Alluvium and associated terrace deposits form a 10- to 20-mile-wide band, primarily to the west of the Pecos River. The deposits extend from about 10 miles north of Roswell to about 10 miles north of Carlsbad (**fig. 92**). The alluvium generally is 150 to 300 feet thick near the Pecos River and thins to the west, but the thickness is irregular because of the collapse of solution openings in the underlying rocks.

The alluvium contains water under unconfined (water table) conditions in the eastern one-half of its areal extent; elsewhere, the alluvium is unsaturated. The alluvial aquifer is hydraulically connected to the underlying carbonate-rock aquifer by leakage through the upper confining unit of the carbonate-rock aquifer. The water table in the alluvial aquifer is at or near land surface along most of the Pecos River, and the alluvial aquifer is recharged by, or discharges to, the river.

Solution-altered zones in the San Andres Limestone and the Grayburg Formation form the principal carbonate-rock aquifer in the Roswell Basin. The carbonate-rock aquifer is 200 to 500 feet thick in the eastern one-half of the basin and thins northward and westward. The lower boundary of the aquifer is formed by the unaltered lower part of the San Andres Limestone or the Glorieta Sandstone, or the underlying Yeso Formation, all of which are much less permeable than the aquifer. The upper part of the Grayburg and the Queen Formations generally are little altered and have low permeability. The zones of low permeability in these two formations form the upper confining unit of the carbonate-rock aquifer and separate it from the overlying alluvial aquifer. In the western part of the basin, water in the carbonate-rock aquifer generally is under unconfined conditions (**fig. 96**), and recharge readily percolates from the land surface to the water table. In the eastern part of the basin, water in the carbonate-rock aquifer is under confined conditions, and flowing wells (**fig. 96*A***, well C) are present in some low-lying areas where the potentiometric surface is above land surface.

RECHARGE AND DISCHARGE

Aquifers in the Roswell Basin primarily are recharged from infiltration of precipitation in the outcrop areas of the San Andres Limestone and the alluvium (**fig. 96*B***). These two units are exposed at the land surface throughout most of an approximately 8,400-square-mile drainage area to the west of the Pecos River. In much of this area, ground water moves eastward toward the Roswell Basin through thin permeable layers in the basal part of the San Andres Limestone or through sandstone beds in the Glorieta Sandstone or the Yeso Formation. Much of the water that recharges the aquifers in the western part of the drainage area probably moves laterally and upward into the western margin of the much more permeable carbonate-rock aquifer. The carbonate-rock aquifer receives additional recharge from direct infiltration of precipitation, from surface water in streams and ponds, and from water applied to irrigated fields. In local areas where the potentiometric surface in the alluvial aquifer is higher than that of the carbonate-rock aquifer, additional recharge can occur by downward leakage of water from the alluvial aquifer through the upper confining unit and into the carbonate-rock aquifer.

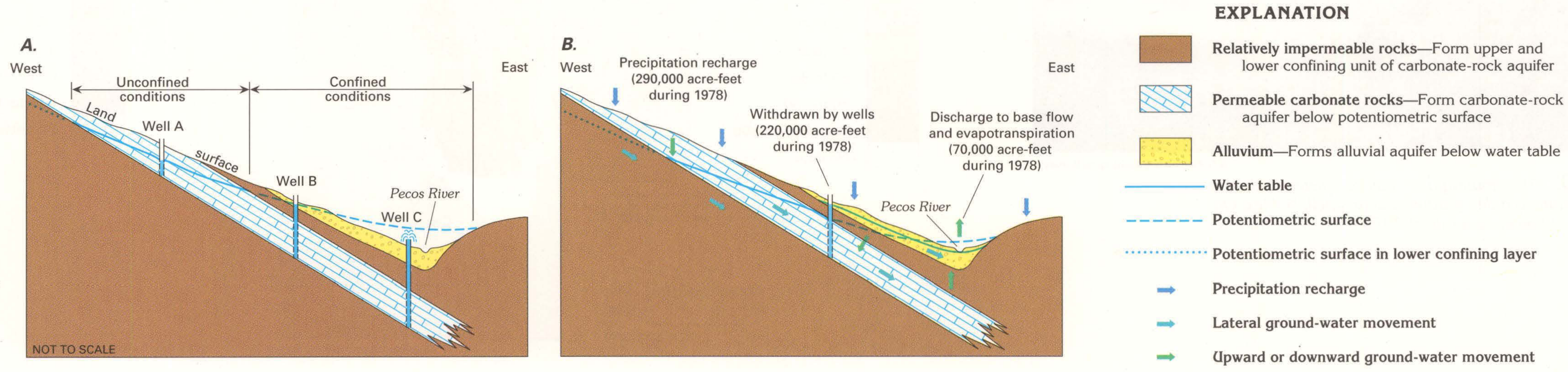


Figure 96. *The potentiometric surface of the carbonate-rock aquifer slopes from west to east. A, the carbonate-rock aquifer contains water under unconfined and confined conditions. Wells will flow where the potentiometric surface of the aquifer is higher than land surface. B, Recharge to the alluvial and carbonate-rock aquifers primarily is from infiltration of precipitation. Discharge primarily is by withdrawal from wells, but some water moves from the aquifers to the Pecos River.*

RECHARGE AND DISCHARGE—
Continued

In most of the Roswell Basin, the potentiometric surface in the carbonate-rock aquifer was higher than that of the alluvial aquifer in 1975 (**fig. 97**). In low-lying areas near the Pecos River, the potentiometric surface in the carbonate-rock aquifer was above land surface, and flowing wells were present. The higher heads in the carbonate-rock aquifer caused the upward discharge of water from that aquifer either into the alluvial aquifer or to the land surface. Rocks beyond the eastern boundary of the carbonate-rock aquifer are relatively impermeable, and minimal discharge occurs across this boundary.

The alluvial aquifer receives recharge from infiltration of precipitation but also is readily recharged from water applied to the approximately 120,000 acres of irrigated fields that overlie the aquifer. The Pecos River flows on alluvium for most of its length in the Roswell Basin and is in hydraulic connection with the alluvial aquifer in most of this area. The alluvial aquifer generally discharges to the Pecos but can receive recharge from the river if ground-water withdrawals lower the water table near the river. The aquifer also discharges by evapotranspiration from areas where the water table is at shallow depth.

Ground-water withdrawals are the principal means of discharge from the aquifer system. Water-level declines caused by withdrawal have altered water-level relations between the alluvial and carbonate-rock aquifers and between the alluvial aquifer and the Pecos River. Before the development of the area's ground-water resources began in 1891, ground water moved from the western recharge areas through the carbonate-rock aquifer upward into the alluvial aquifer and then into the Pecos River. Water-level declines caused by withdrawal from the carbonate-rock aquifer locally have decreased, halted, or reversed the normal upward discharge from that aquifer. Withdrawals from the alluvial aquifer can have the same effect on the natural discharge to the Pecos River and also can induce upward discharge of water from the carbonate-rock aquifer.

The rate of natural recharge to the aquifers in the Roswell Basin has been estimated to range from 240,000 to 280,000 acre-feet per year on the basis of measurements made in 1926 and 1953. More recent measurements (1978) of base-flow gain in the Pecos River and measurements of ground-water withdrawal allow additional (although possibly no more accurate) estimates of the water budget of the basin. During 1978, the base flow in the Pecos River increased by about 20,000 acre-feet in the Roswell Basin reach of the river as the result of ground-water discharge to the river. Phreatophytes grow in part of a 64-square-mile area along the Pecos River and annually consume about 50,000 acre-feet of ground water. The combined ground-water discharge to phreatophytes and the Pecos River totaled about 70,000 acre-feet. During 1978, about 380,000 acre-feet of ground water was withdrawn; about 252,000 acre-feet was withdrawn from the carbonate-rock aquifer, and about 126,000 acre-feet was withdrawn from the alluvial aquifer through approximately 1,500 large-capacity irrigation, commercial, and industrial wells. This water primarily was used for irrigation, and about 160,000 acre-feet returned to the aquifer through infiltration. Effective withdrawal, which totaled 220,000 acre-feet, is the volume of water actually removed from the aquifers (**fig. 96B**). The total effective discharge from the aquifers in the Roswell Basin during 1978 was about 290,000 acre-feet. If the volume of ground-water in storage does not change, then this estimate of discharge also is an estimate of recharge to the aquifers. During 1978 and for several preceding years, the annual change in water levels in wells completed in the aquifers was small, which indicates that changes in ground-water storage were small. Thus, 290,000 acre-feet was the approximate volume of recharge to the aquifers during 1978. Because most of the recharge is supplied through infiltration of precipitation and runoff in the 8,400-square-mile drainage area to the west of the alluvial aquifer, the average rate of precipitation recharge was about 0.6 inch per year in 1978.

WATER-LEVEL CONDITIONS

The 1975 potentiometric surface of the carbonate-rock aquifer in the Roswell Basin sloped gently to the southeast and ranged in altitude from about 3,550 to 3,250 feet above sea level (**fig. 98**). The potentiometric surface can be divided into three distinct zones on the basis of gradient. In the northern part of the basin, the potentiometric surface has a nearly flat southeasterly gradient of about 0.5 to 1.0 foot per mile. The potentiometric surface steepens markedly along a northeast-trending zone in the southern part of the basin. Gradients in this area generally are 10 to 50 feet per mile. This steepening of the potentiometric surface might be the result of a marked decrease in the effective transmissivity of the aquifer that is caused by the movement of water across bedding planes. This zone of steep gradient is located in the general area where the carbonate-rock aquifer extends across the stratigraphic contact between the San Andres Limestone on the northwest and the Grayburg Formation on the southeast (**fig. 94**). In the extreme southeastern part of the basin, the potentiometric surface is flatter; gradients are about 2 to 15 feet per mile.

The aquifer in carbonate rocks in the drainage area to the west of the Roswell Basin has a very steep potentiometric surface (**fig. 98**). Gradients in this area generally are 50 to 100 feet per mile in contrast to gradients of 1 foot per mile or less in the adjacent carbonate-rock aquifer in the Roswell Basin. This difference in gradient is the likely result of a difference in the transmissivity of the two aquifers; the carbonate-rock aquifer within the basin has much larger transmissivity.

The 1975 potentiometric surface of the alluvial aquifer had a general slope from west to east toward the Pecos River (**fig. 99**), but the shape of the surface was irregular because of the effect of ground-water withdrawal. Large cones of depression near groups of pumping wells altered the shape of the potentiometric surface.

The Roswell Basin has a long history of ground-water use. The first wells were drilled in the basin in 1891, and by 1937, about 1,300 large-capacity irrigation, municipal, and industrial wells had been constructed. In 1926, when the first study of the hydrology of the basin was undertaken, water levels in the carbonate-rock aquifer were still as much as 100 feet above land surface in some areas near the Pecos River, and a few large-capacity wells flowed at rates of 3,000 to 5,000 gallons per minute (**fig. 100**). Between 1926 and 1950, the water level in the eastern part of the carbonate-rock aquifer declined 10 to 30 feet. From 1950 to 1975, water-level declines exceeded 40 feet in much of the aquifer and were more than 100 feet near Lake McMillan (**fig. 101**).

Water-level declines in the alluvial aquifer between 1950 and 1975 exceeded 40 feet in a few areas and exceeded 80 feet in one cone of depression near the center of the aquifer (**fig. 102**). Water-level declines along the eastern margin of the aquifer are moderated by the effects of recharge from the Pecos River; between 1950 and 1975 declines were minimal in this area. The areas of large water-level decline in the alluvial aquifer do not coincide with the areas of large water-level decline in the carbonate-rock aquifer. In 1950, the potentiometric surface in the carbonate-rock aquifer was higher than the potentiometric surface of the alluvial aquifer in most of the area of the alluvial aquifer. By 1975, ground-water withdrawal had produced large cones of depression in both aquifers, and the relation of the potentiometric surfaces was reversed in several areas (**fig. 97**).

Seasonal water-level declines caused by pumping during the March to September growing season have exceeded 120 feet in some observation wells completed in the carbonate-rock aquifer. Seasonal declines generally have not exceeded 30 feet in the alluvial aquifer. After the growing season, ground-water withdrawal decreases markedly, and water levels generally recover back to near the high level attained during the previous year. The large seasonal drawdown of water levels in the carbonate-rock aquifer causes seasonal increases in downward leakage of water from the alluvial aquifer and seasonal decreases in the volume of water discharged to the Pecos River.

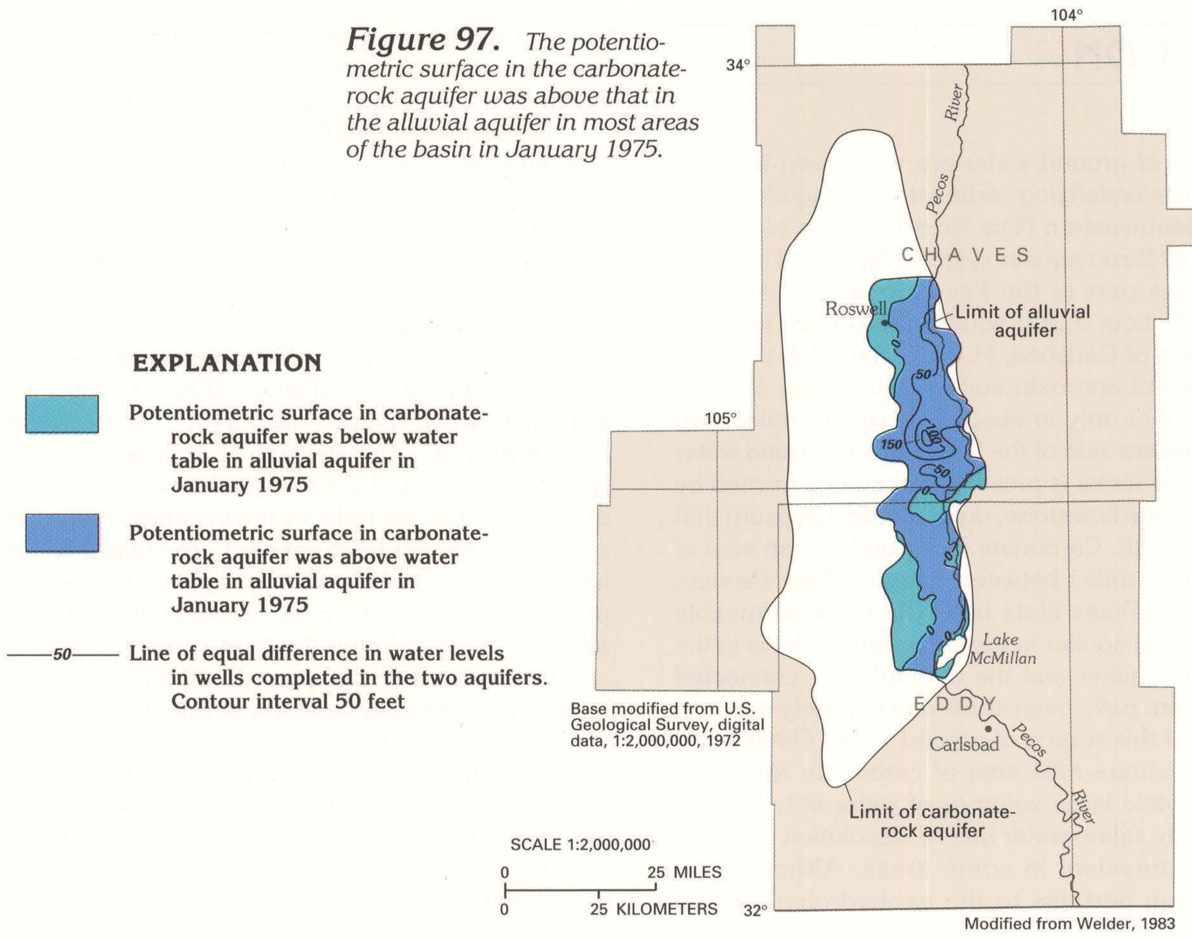


Figure 97. *The potentiometric surface in the carbonate-rock aquifer was above that in the alluvial aquifer in most areas of the basin in January 1975.*

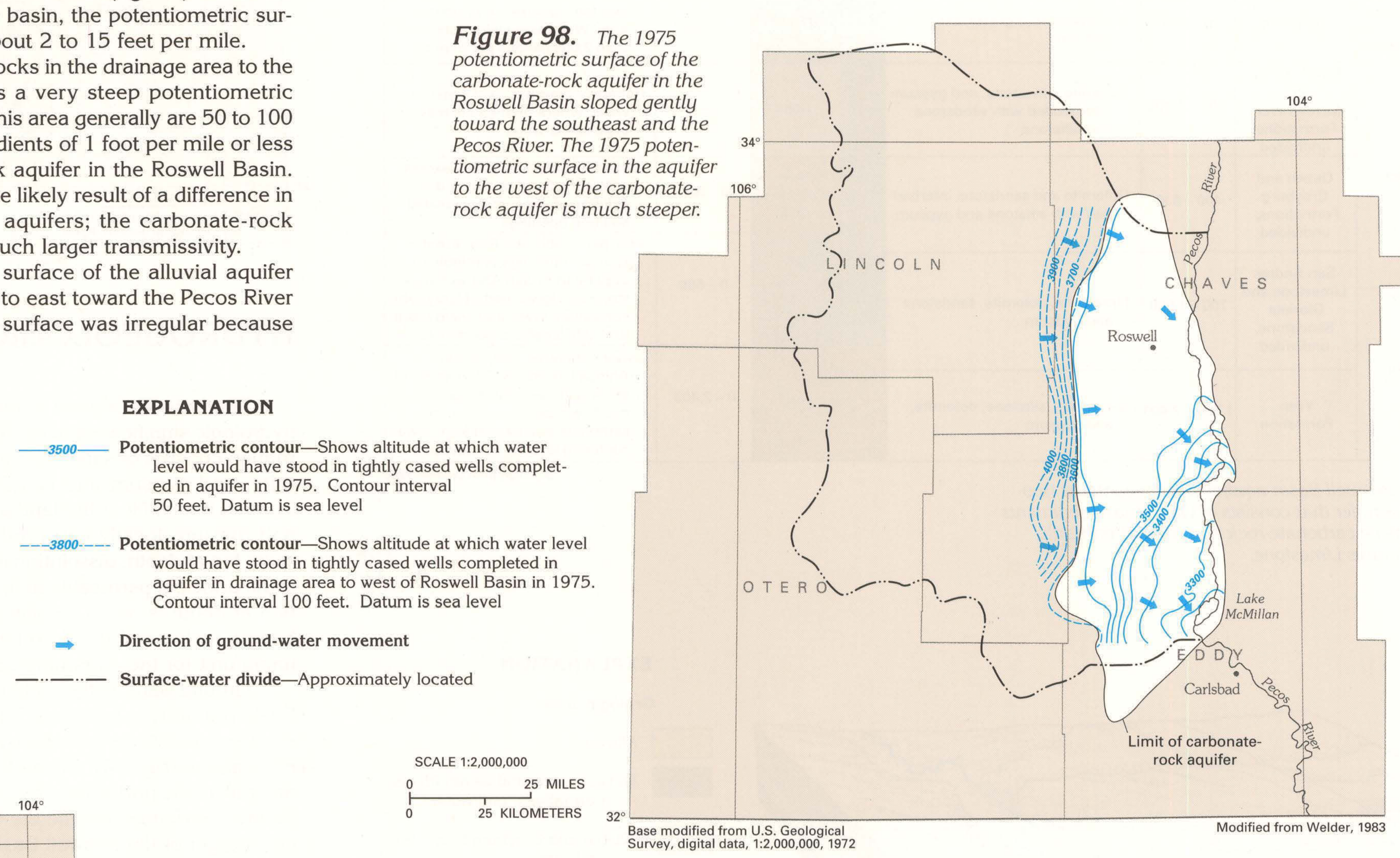


Figure 98. *The 1975 potentiometric surface of the carbonate-rock aquifer in the Roswell Basin sloped gently toward the southeast and the Pecos River. The 1975 potentiometric surface in the aquifer to the west of the carbonate-rock aquifer is much steeper.*

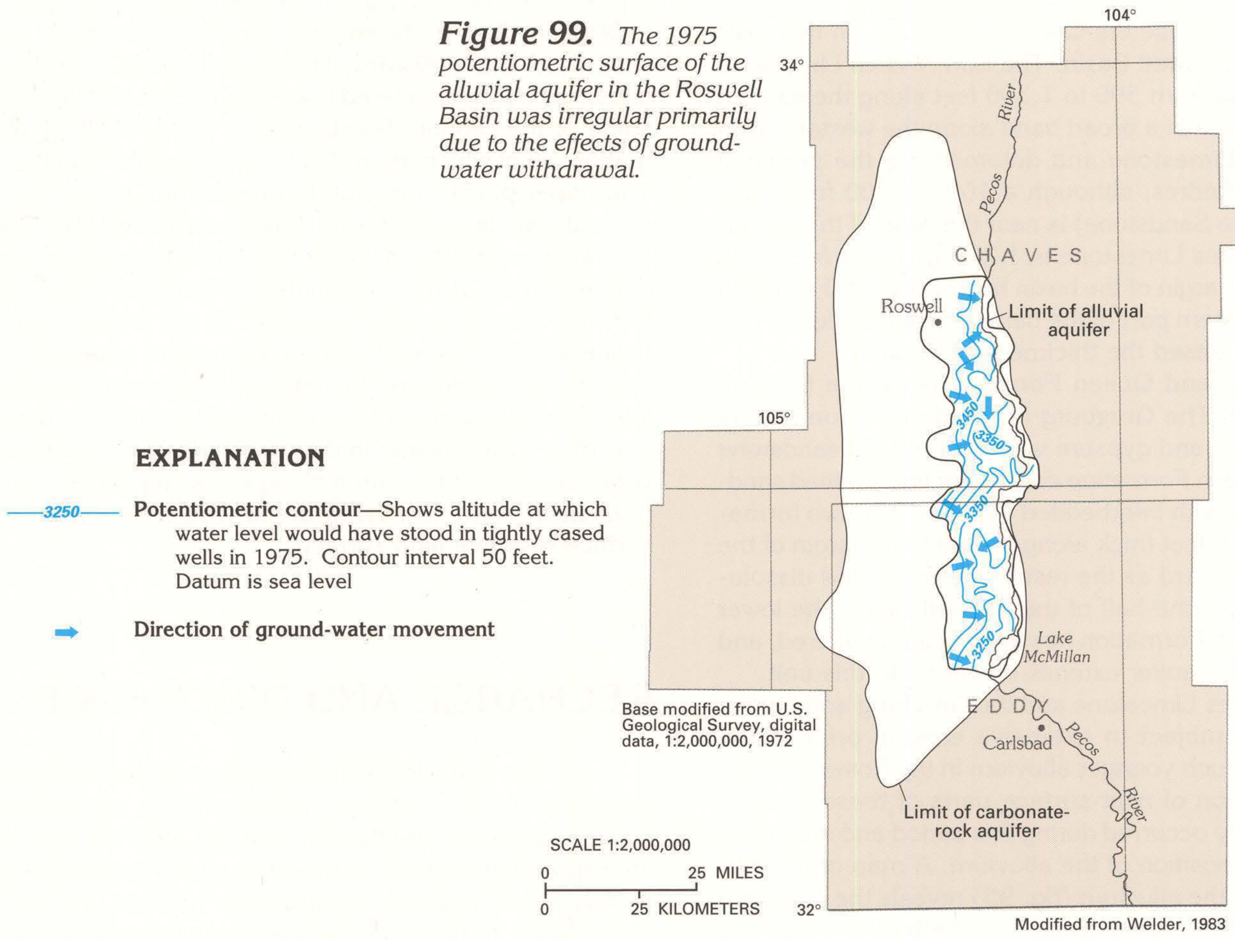


Figure 99. *The 1975 potentiometric surface of the alluvial aquifer in the Roswell Basin was irregular primarily due to the effects of ground-water withdrawal.*

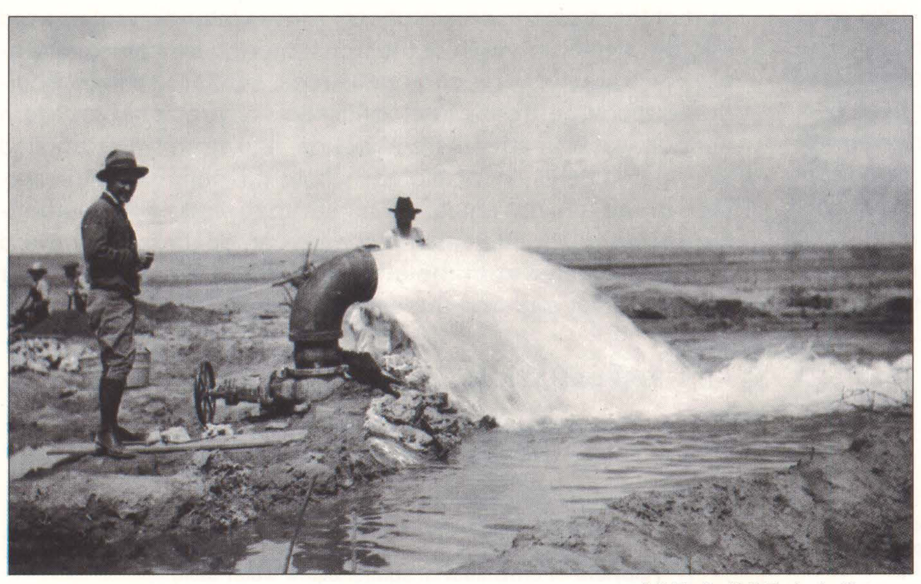

A.G. Fiedler, U.S. Geological Survey

Figure 100. *This flowing well was completed in the carbonate-rock aquifer 8 miles southeast of Roswell, N. Mex. On April 21, 1926, it flowed at a rate of 3,190 gallons per minute.*

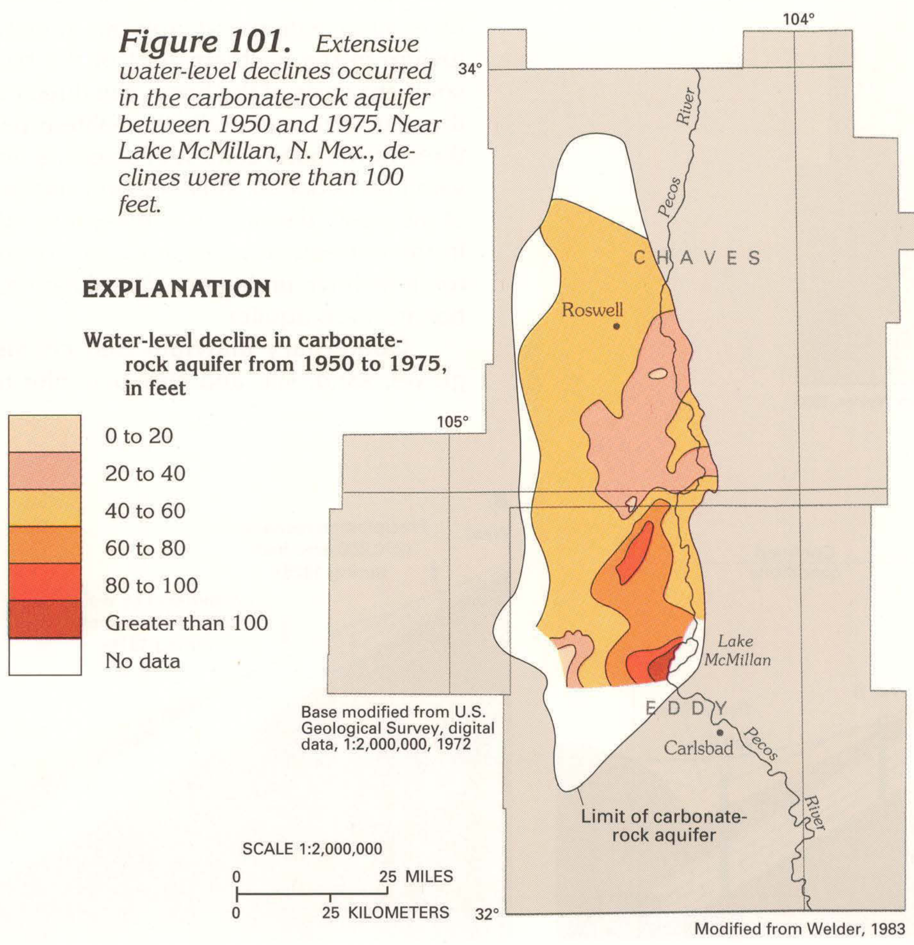


Figure 101. *Extensive water-level declines occurred in the carbonate-rock aquifer between 1950 and 1975. Near Lake McMillan, N. Mex., declines were more than 100 feet.*

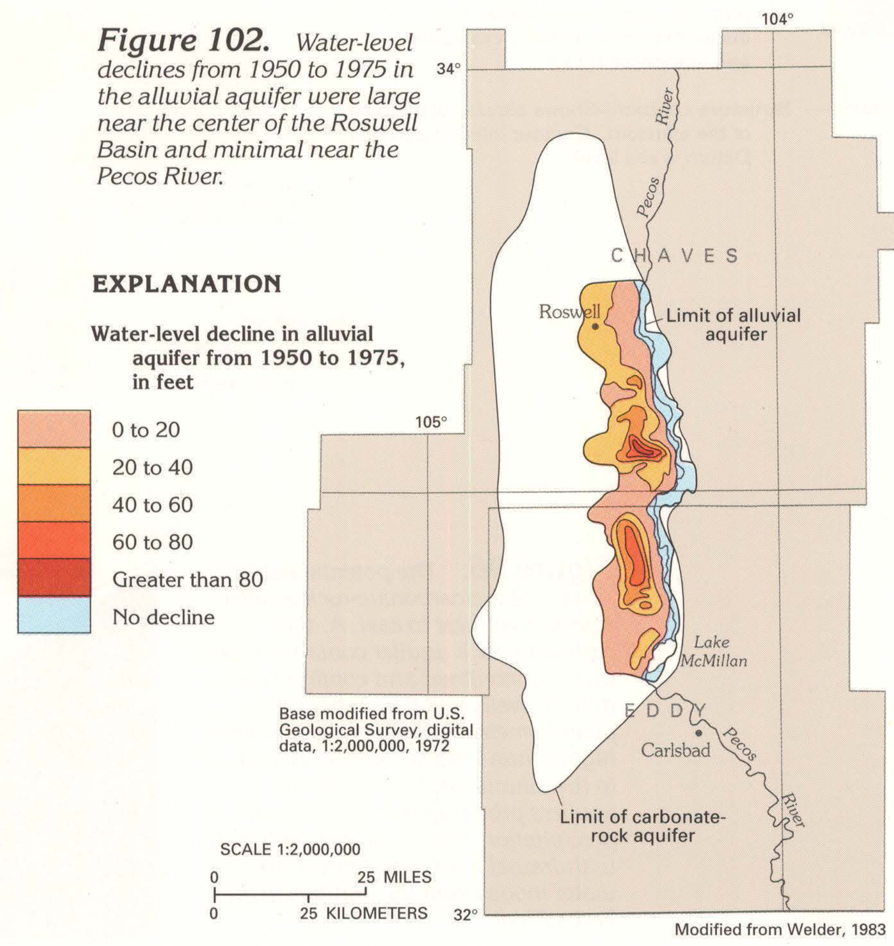


Figure 102. *Water-level declines from 1950 to 1975 in the alluvial aquifer were large near the center of the Roswell Basin and minimal near the Pecos River.*

AQUIFER CHARACTERISTICS

The transmissivity of an aquifer is a measure of the relative ease with which the aquifer transmits water. An aquifer of large transmissivity can sustain larger well yields, commonly contains a flatter potentiometric surface, and transmits larger volumes of water than an aquifer of small transmissivity.

Large differences in transmissivity are present in the aquifers of the Roswell Basin. In the northern part of the carbonate-rock aquifer where the potentiometric surface is relatively flat, transmissivity generally ranges from 30,000 to 50,000 feet squared per day, but values greater than 300,000 and less than 2,000 feet squared per day have been calculated from aquifer-test data for a few wells. These large differences in transmissivity likely are caused by the uneven distribution of solution openings in the aquifer. The transmissivity of the zone of steep potentiometric gradient in the southern part of the carbonate-rock aquifer is poorly documented but probably is small. In the southeastern part of the carbonate-rock aquifer, transmissivity ranges from about 8,000 to 20,000 feet squared per day. These values are smaller than those in the northern part of the aquifer, and the potentiometric gradients in this southeastern area are steeper than those in the northern part of the aquifer.

The transmissivity of the alluvial aquifer is more nearly uniform and generally is smaller than that of the carbonate-rock aquifer. Transmissivity of the alluvial aquifer averages about 13,000 feet squared per day and ranges from about 5,000 to 15,000 feet squared per day.

GROUND-WATER QUALITY

Ground water in the western part of the carbonate aquifer in the Roswell Basin generally contains a preponderance of dissolved calcium, magnesium, and sulfate and is classified as either a calcium sulfate or a calcium magnesium sulfate type water. Calcium concentrations generally range from 100 to 500 milligrams per liter, magnesium concentrations generally range from 50 to 130 milligrams per liter, and sulfate concentrations generally range from 300 to 1,400 milligrams per liter. The water is of similar chemical composition to that in other carbonate-rock aquifers where active dissolution of limestone, dolomite, and gypsum is occurring. The water is classified as very hard. Dissolved-solids concentrations generally range from 700 to 2,600 milligrams per liter.

Along the northeastern margin of the carbonate-rock aquifer, dissolved sodium and chloride concentrations in the water can be large; consequently, the water is classified as a sodium chloride type. Sodium concentrations in this area generally range from 1,500 to 3,000 milligrams per liter, and chloride concentrations range from 2,000 to 5,000 milligrams per liter (**fig. 103**). The water in this area is classified as very hard. Dissolved-solids concentrations range from 7,000 to 12,000 milligrams per liter.

Water of large sodium chloride (salt) content is of particular concern in the Roswell Basin because most water is used for irrigation, and many crops can be damaged by excessive salt in the water and soil. The source of the large chloride concentrations in the carbonate-rock aquifer is uncertain but might be brine that moved across the relatively impermeable eastern boundary of the aquifer. Seasonal water-level declines in the carbonate-rock aquifer might temporarily reverse the direction of ground-water movement across the eastern boundary and enable brines in the deeper parts of the San Andres Limestone to move westward into the carbonate-rock aquifer. Chloride concentrations in water in the eastern part of the aquifer generally are larger near the end of the pumping season when water-level declines are large; concentrations decrease in the winter and early spring when water levels have returned to nonpumping levels. Large chloride concentrations in water samples from the bottom of some wells indicate that these concentrations are larger at greater depth in water in the eastern part of the carbonate-rock aquifer (**fig. 104**).

When water with large chloride concentration is deep in the carbonate-rock aquifer (**fig. 104*A***), it has little effect on the water quality in shallow parts of the aquifer, and water pumped from wells is of relatively uniform quality. However, if the water with large chloride concentration is drawn farther into the aquifer (**fig. 104*B***), then wells close to the eastern boundary can be severely affected (well C), and more westerly wells might be unaffected or only moderately affected (wells A and B), depending on well location and depth. Water in the carbonate-rock aquifer to the east of Roswell has undergone a marked increase in chloride concentration. Between 1959 and 1978, chloride concentrations increased by 1,000 to 2,000 milligrams per liter in water from some wells in this area. Increases in 1959–78 chloride concentrations generally have been less than 100 milligrams per liter along the southern one-half of the eastern margin of the aquifer.

Water in the southern one-half of the alluvial aquifer generally is a calcium sulfate type. In the northern one-half of the aquifer, and at a few points along the southeastern margin of the aquifer, the water generally is a mixed calcium sodium sulfate chloride type. The water is very hard throughout the aquifer; dissolved-solids concentrations range from about 500 to 5,000 milligrams per liter. Chloride concentrations range from about 50 milligrams per liter along the western margin of the aquifer to about 2,000 milligrams per liter in a few areas along the eastern margin of the aquifer (**fig. 105**).

In the eastern part of the alluvial aquifer, chloride concentrations can be large in ground water near the upper or lower parts of the aquifer. Large concentrations in the upper part of the aquifer probably are caused by infiltration of water with large chloride concentration from local canals or from wells completed in more saline zones in the carbonate-rock aquifer (**fig. 104*B***). Evapotranspiration by phreatophytes also concentrates dissolved minerals in the soil and shallow water table near the Pecos River. Water with large chloride concentration in the lower part of the alluvial aquifer likely is caused by upward movement of more saline water through the upper confining layer of the carbonate-rock aquifer. Both processes have caused water-quality degradation in the alluvial aquifer. Between about 1957 and 1978, chloride concentrations increased from 30 to 1,000 milligrams per liter in water from some wells.

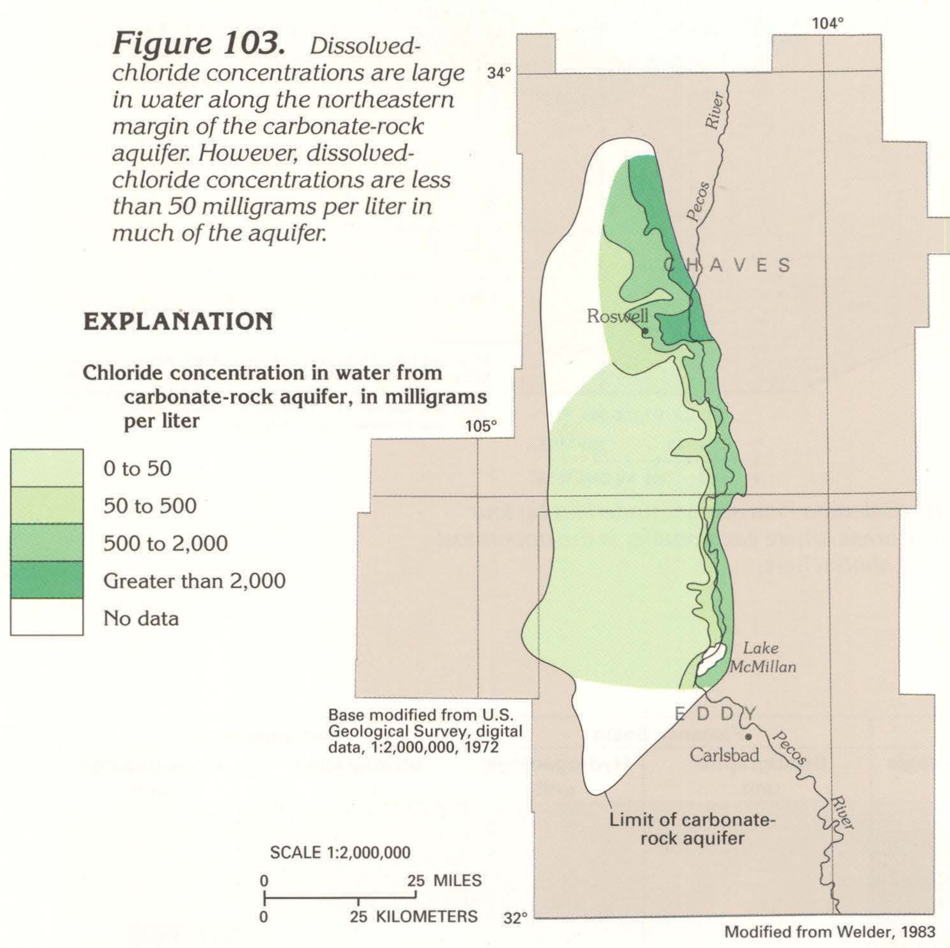


Figure 103. *Dissolved-chloride concentrations are large in water along the northeastern margin of the carbonate-rock aquifer. However, dissolved-chloride concentrations are less than 50 milligrams per liter in much of the aquifer.*

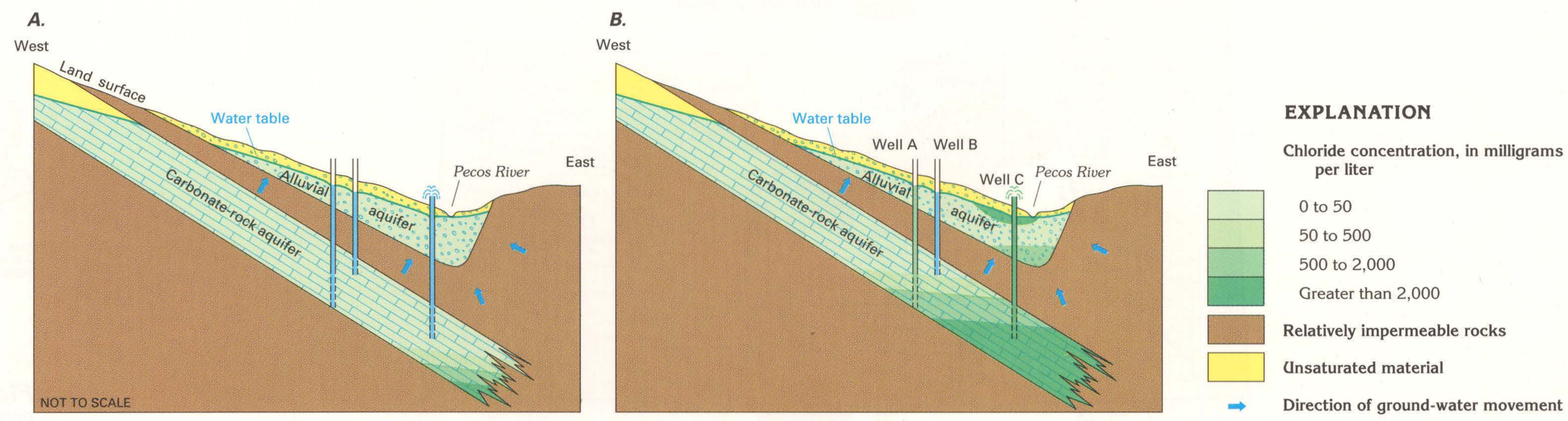


Figure 104. *A, During winter months when water levels in the aquifer are high, water with large chloride concentration can be present in the deep parts of the carbonate-rock aquifer. B, As water levels decline during the growing season, water with large chloride concentration can move upward and degrade the quality of water in some wells and in the alluvium.*

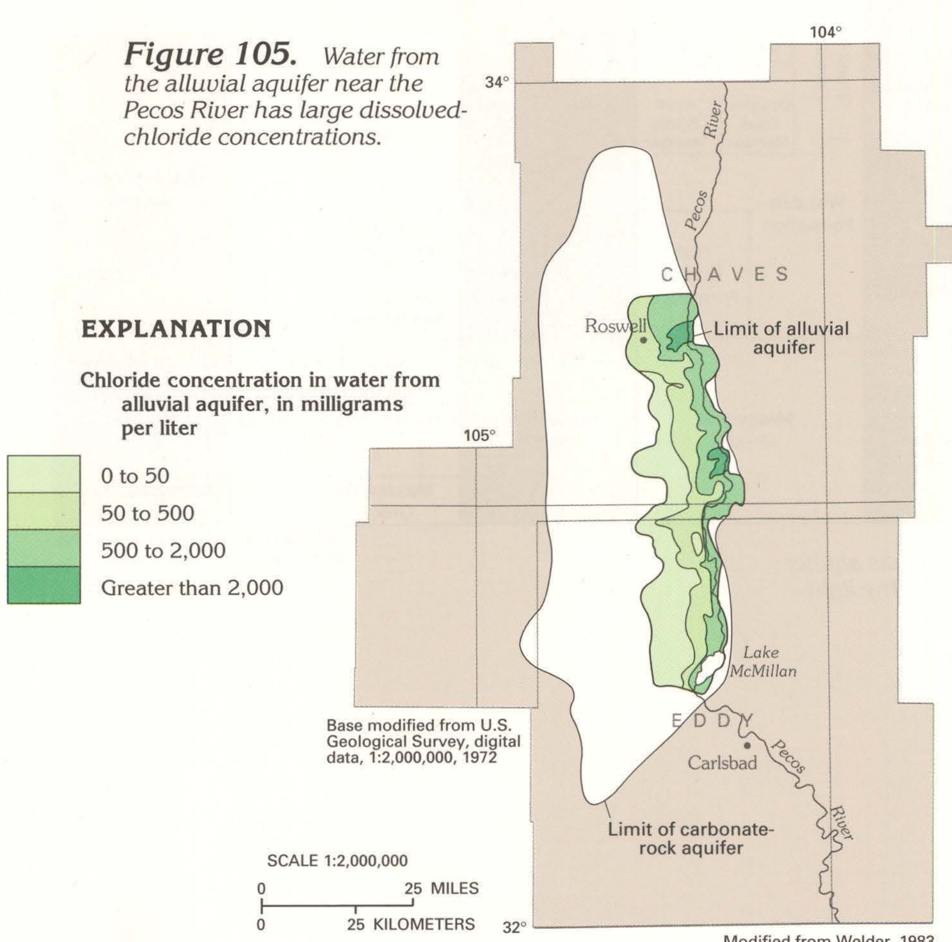


Figure 105. *Water from the alluvial aquifer near the Pecos River has large dissolved-chloride concentrations.*

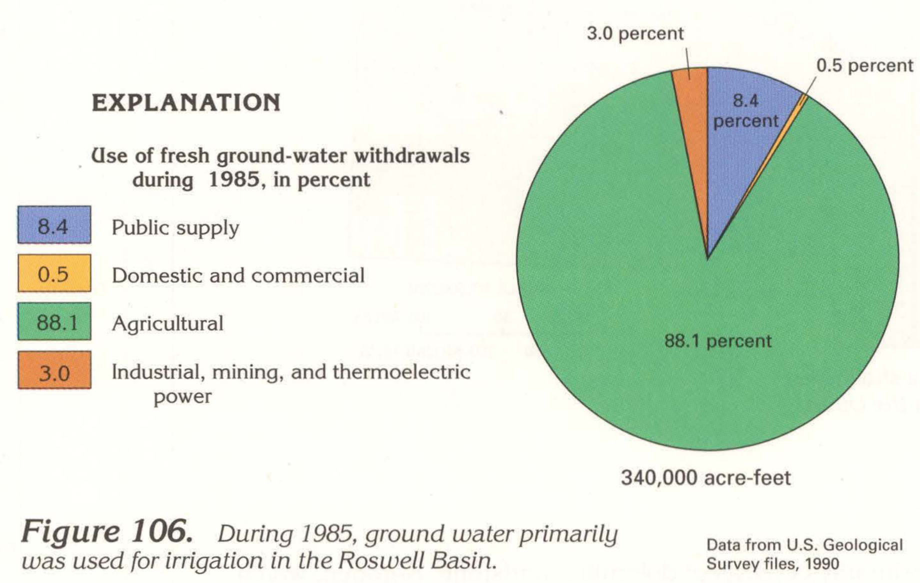


Figure 106. *During 1985, ground water primarily was used for irrigation in the Roswell Basin.*

FRESH GROUND-WATER WITHDRAWALS

In 1978, about 95 percent of the 380,000 acre-feet of ground water withdrawn from the Roswell Basin was used for irrigation. In 1985, about 88 percent of the 340,000 acre-feet of ground water withdrawn from the basin was used for irrigation; about 8 percent of the withdrawal was used for public supply (**fig. 106**). Large-capacity wells completed in the carbonate-rock aquifer generally yield in excess of 3,000 gallons per minute. Smaller capacity wells commonly yield 500 to 1,500 gallons per minute. Large-capacity wells completed in the alluvial aquifer generally yield about 2,000 gallons per minute, and smaller capacity wells yield 300 to 700 gallons per minute.

Colorado Plateaus aquifers

INTRODUCTION

The Colorado Plateaus aquifers underlie an area of approximately 110,000 square miles in western Colorado, northwestern New Mexico, northeastern Arizona, and eastern Utah (**fig. 107**). This area is approximately coincident with the Colorado Plateaus Physiographic Province. The distribution of aquifers in the Colorado Plateaus is controlled in part by the structural deformation and erosion that has occurred since deposition of the sediments that compose the aquifers. The principal aquifers in younger rocks are present only in basins such as the Uinta, Piceance, and San Juan Basins (**fig. 108**). In uplifted areas, such as the Monument and Defiance Uplifts and the Coconino Plateau, younger rocks have been eroded away, and aquifers are present in older rocks that underlie more extensive parts of the Colorado Plateaus area. Although the quantity and chemical quality of water in the Colorado Plateaus aquifers are extremely variable, much of the land in this sparsely populated region is underlain by rocks that contain aquifers capable of yielding usable quantities of water of a quality suitable for most agricultural or domestic use.

In general, the aquifers in the Colorado Plateaus area are composed of permeable, moderately to well-consolidated sedimentary rocks. These rocks range in age from Permian to Tertiary and vary greatly in thickness, lithology, and hydraulic characteristics. The stratigraphic relations of the rocks are complicated in places, and the stratigraphic nomenclature consequently is diverse. Many water-yielding units have been identified in these rocks, and most publications that pertain to the hydrogeology of the area describe only a few of the units or pertain to only part of the Colorado Plateaus. In this Chapter, the many water-yielding units in the area have been grouped into four principal aquifers for purposes of discussion. The principal aquifers are the Uinta–Animas aquifer, the Mesaverde aquifer, the Dakota–Glen Canyon aquifer system, and the Coconino–De Chelly aquifer (**fig. 107**). Most widespread and productive water-yielding units are included in these aquifers; however, some locally productive water-yielding units have been excluded.

Water-yielding units excluded from the principal aquifers can form aquifers of local importance, but these units either are not extensive enough or not productive enough to be considered as principal aquifers for the purposes of this Altas. In general, these rocks are considered to be confining units containing minor water-yielding units.

Relatively impermeable confining units separate each of the four principal aquifers in the Colorado Plateaus. The two thickest units are the Mancos confining unit, which immediately underlies the Mesaverde aquifer, and the Chinle–Moenkopi confining unit, which immediately underlies the Dakota–Glen Canyon aquifer system. Thinner and less extensive confining units separate some water-yielding zones within the principal aquifers; however, these units generally form less effective barriers to ground-water movement than the confining units between the principal aquifers. Where the intra-aquifer confining units are thin or absent, water can move between adjacent water-yielding zones within an aquifer.

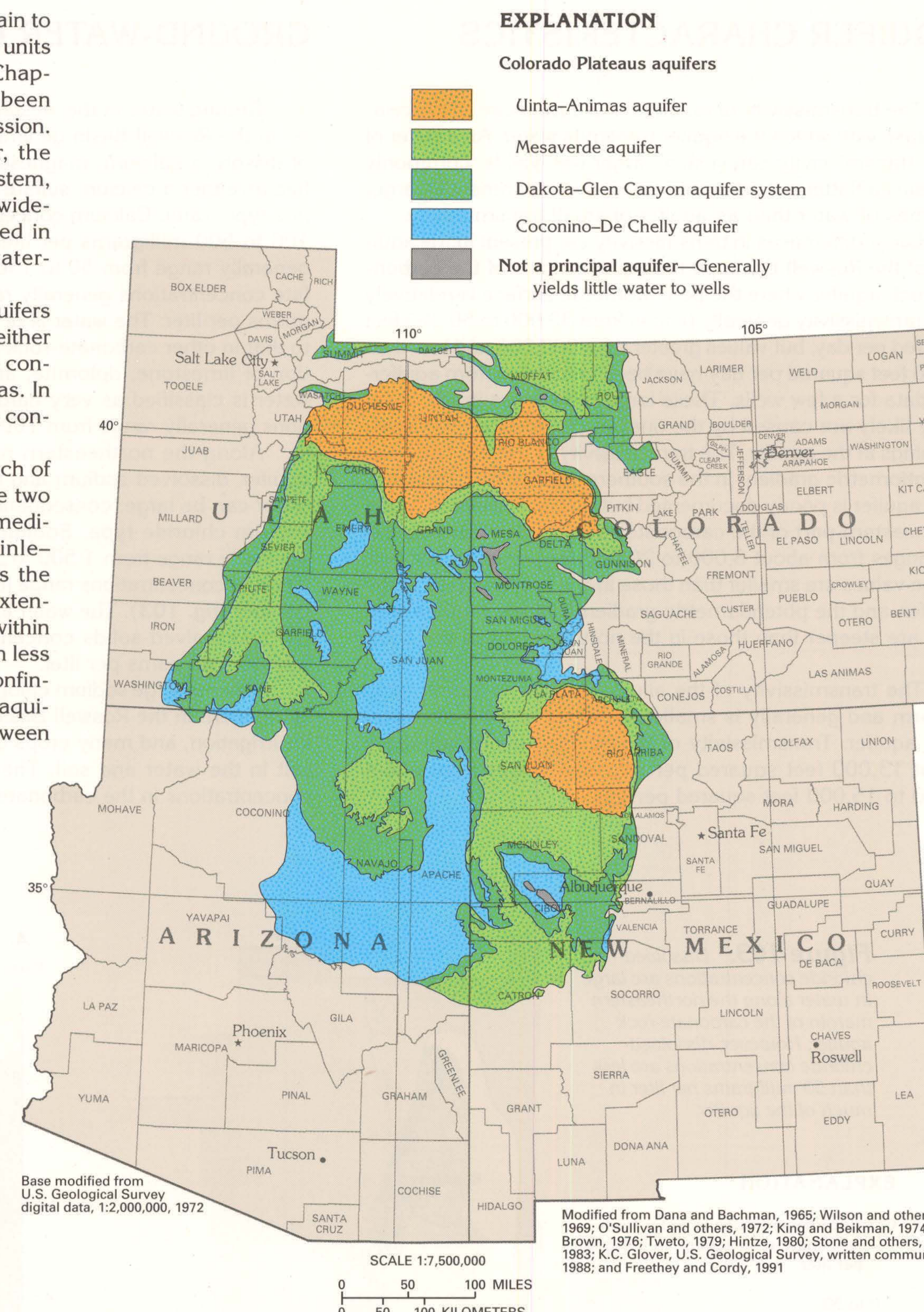


Figure 107. The Colorado Plateaus are underlain by four principal aquifers. The areas where each aquifer is the uppermost water-yielding unit are shown here.

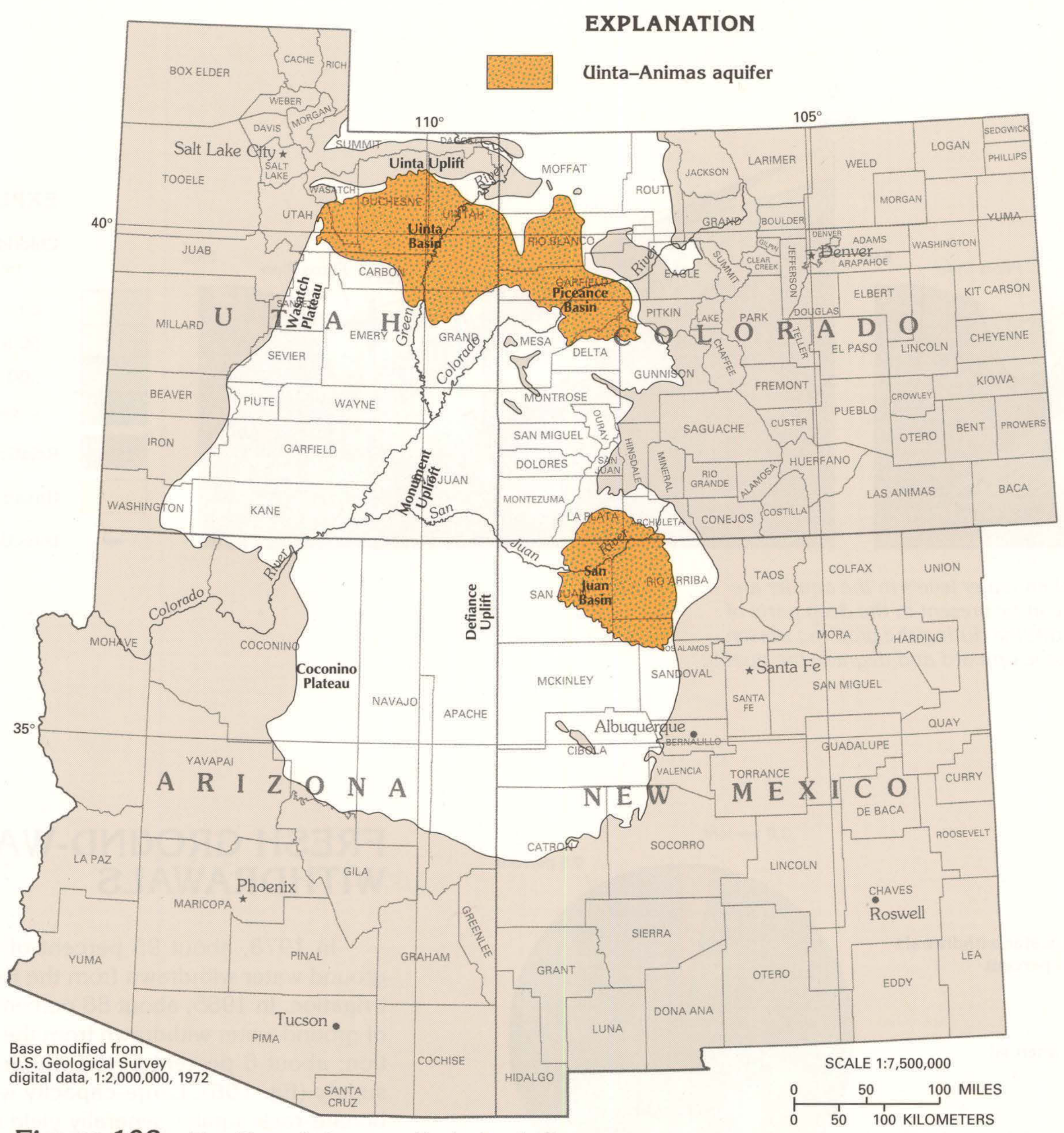


Figure 108. The Uinta–Animas aquifer is the shallowest of the Colorado Plateaus aquifers and is present in the Uinta, Piceance, and San Juan Basins.

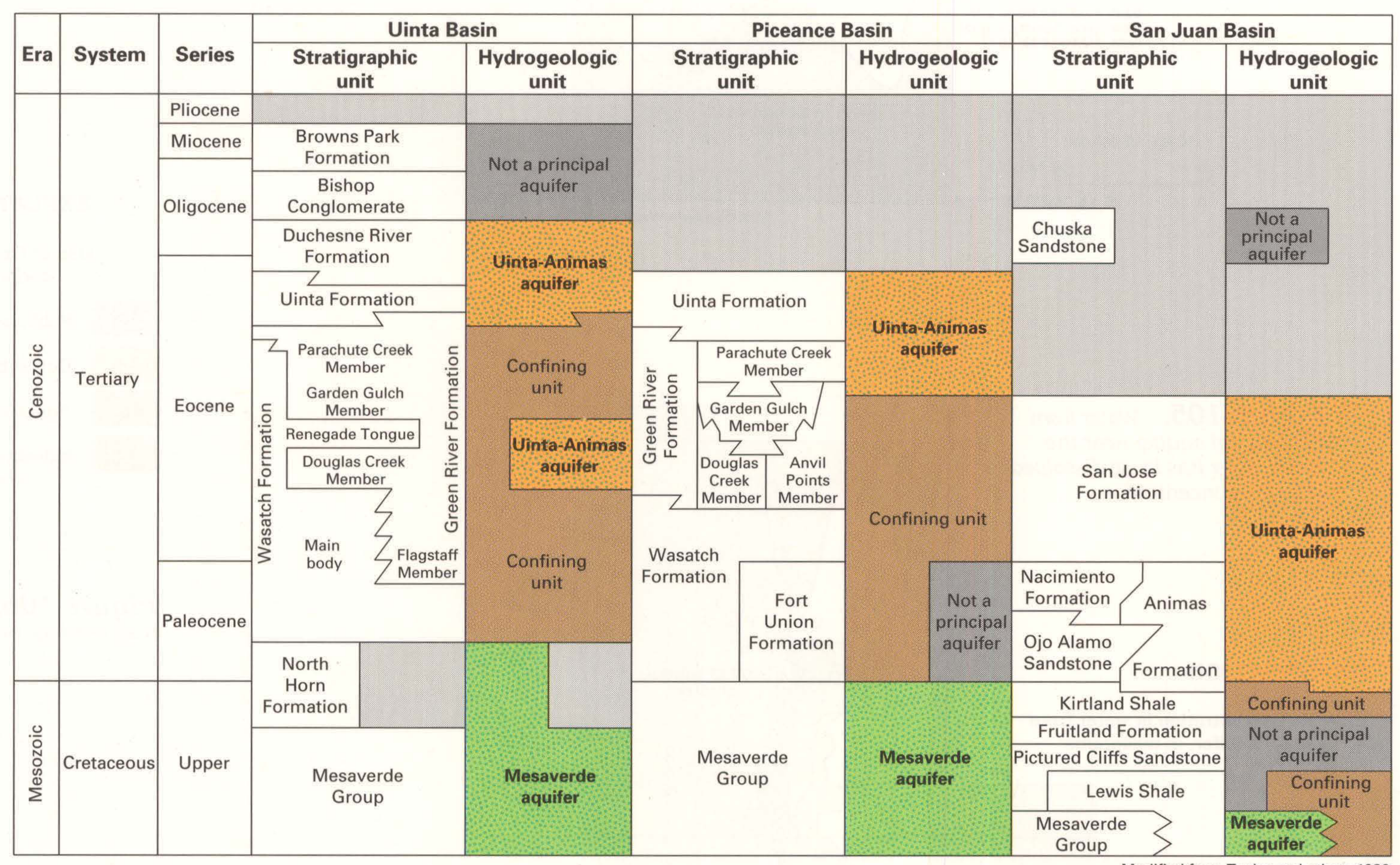


Figure 109. Rock units that contain the Uinta–Animas aquifer are in different stratigraphic intervals in the three basins. The light gray areas represent missing rocks.

UINTA–ANIMAS AQUIFER

The Uinta–Animas aquifer primarily is composed of Lower Tertiary rocks in the Uinta Basin of northeastern Utah, the Piceance Basin of northwestern Colorado, and the San Juan Basin of northwestern New Mexico (**fig. 108**). Aquifers in each basin are present in different parts of the stratigraphic section (**fig. 109**). Some formations are considered to be an aquifer in more than one basin; however, some formations vary so much in their hydraulic characteristics that they are considered to be an aquifer in one basin and a confining unit in another.

Hydrogeologic Units

The Uinta–Animas aquifer in the Uinta Basin is present in water-yielding beds of sandstone, conglomerate, and siltstone of the Duchesne River and Uinta Formations, the Renegade Tongue of the Wasatch Formation, and the Douglas Creek Member of the Green River Formation (**fig. 109**). The Duchesne River Formation consists mostly of permeable fluvial sandstone and conglomerate. Grain size of these sediments decreases with distance from the Uinta Uplift, and relatively impermeable shale is common in the center of the basin. The Uinta Formation consists of permeable, poorly sorted, fine to coarse sandstone with some siltstone and mudstone. These rocks become more coarse grained and permeable toward the top of the formation. Coarse-grained rocks adjacent to the Uinta Uplift and the Wasatch Plateau grade into finer grained sediments away from the uplifted areas. The Renegade Tongue of the Wasatch Formation and the Douglas Creek Member of the Green River Formation contain an aquifer along the southern and eastern margins of the basin where the rocks primarily consist of fluvial, massive, irregularly bedded sandstone and siltstone. Water-yielding units in the Uinta–Animas aquifer in the Uinta Basin commonly are separated from each other and from the underlying Mesaverde aquifer by units of low permeability composed of claystone, shale, marlstone, or limestone.

The Uinta–Animas aquifer in the Piceance Basin consists of the Uinta Formation and the Parachute Creek Member of the Green River Formation. The Uinta Formation consists of silty sandstone, siltstone, and marlstone. Much of the intergranular space in these rocks has been filled by sodium and calcium bicarbonate cements, but fractures are numerous and produce substantial permeability. The Parachute Creek Member primarily consists of dolomitic marlstone. Kerogen, which is a waxlike hydrocarbon, is present in some parts of the member in the Piceance and Uinta Basins. Marlstone that contains large concentrations of kerogen is known as oil shale and generally is less fractured than marlstone that contains smaller concentrations of kerogen (lean marlstone). Fractures and dissolution openings along fractures in the lean marlstone form the principal pathways for water movement in the aquifer. Oil shale generally is less permeable and forms confining units. The Mahogany zone in the Piceance Basin is an example of one such confining unit (**fig. 110**). In the central part of the Piceance Basin, a saline zone in the marlstone contains the minerals nahcolite and halite, is not extensively fractured, and forms part of the relatively impermeable lower confining unit of the aquifer. The lower part of the Green River Formation and the Wasatch Formation form most of the lower confining unit of the aquifer.

The Uinta–Animas aquifer in the San Juan Basin consists of the San Jose Formation, the underlying Animas Formation and its lateral equivalent, the Nacimiento Formation, and the Ojo Alamo Sandstone. The San Jose Formation is the uppermost significant bedrock formation in the San Juan Basin and primarily consists of permeable, coarse, arkosic sandstone interlayered with mudstone. The Animas and Nacimiento Formations and the Ojo Alamo Sandstone primarily consist of permeable conglomerate and medium to very coarse sandstone interlayered with relatively impermeable shale and mudstone.

The thickness of the Uinta–Animas aquifer generally increases toward the central part of each basin. In the Uinta Basin, for example, the part of the aquifer in the Duchesne River and Uinta Formations ranges in thickness from 0 feet at the southern margin of the aquifer to as much as 9,000 feet in the north-central part of the aquifer. The part of the aquifer in the Renegade Tongue and Douglas Creek Member in the Uinta Basin is about 500 feet thick. In the Piceance Basin, the Uinta–Animas aquifer is as much as 2,000 feet thick in the central part of the basin. In the northeastern part of the San Juan Basin, the maximum thickness of the Uinta–Animas aquifer is about 3,500 feet.

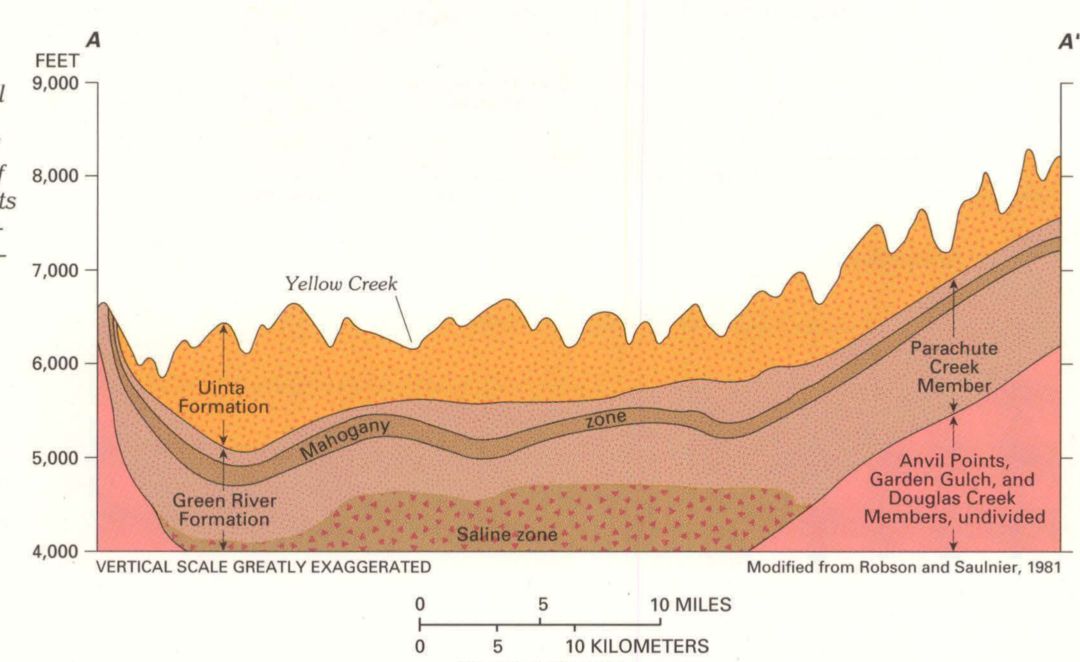


Figure 110. In the central part of the Piceance Basin, a saline zone in the lower part of the Parachute Creek Member of the Green River Formation limits the thickness of the freshwater-yielding materials in the Uinta–Animas aquifer. The line of section is shown in figure 112.

Recharge and Discharge

Ground-water recharge to the Uinta–Animas aquifer generally occurs in the areas of higher altitude along the margins of each basin. Ground water is discharged mainly to streams, springs, and by transpiration from vegetation growing along stream valleys.

In the Uinta Basin, the part of the aquifer in the Duchesne River and Uinta Formations has about 200,000 acre-feet per year of recharge. The rate of ground-water withdrawal is small, and natural discharge is approximately equal to recharge. In the Renegade Tongue and Douglas Creek Member part of the aquifer, recharge and discharge also are approximately equal and total about 1,000 acre-feet per year. Recharge occurs near the southern margin of the aquifer, and discharge occurs near the White and Green Rivers.

The Uinta–Animas aquifer in the Piceance Basin receives about 24,000 acre-feet per year of recharge, primarily in the upland areas near the margins of the aquifer. Discharge is approximately equal to recharge and primarily occurs in the valleys of Piceance Creek and other tributaries to the White River or in the valley of the Colorado River and its tributaries.

In the San Juan Basin, water recharges the Uinta–Animas aquifer in the higher altitude areas that nearly encircle the basin. Ground water generally flows toward the San Juan River and its tributaries where it is discharged to streamflow, to the alluvium that locally is present in the valleys, or to evapotranspiration. During 1985, about 28,000 acre-feet of ground water was withdrawn from the aquifer in the San Juan Basin.

Water-Level Conditions

The potentiometric surface of the Uinta–Animas aquifer generally ranges from about 100 feet above land surface to about 500 feet below land surface; the surface generally is near or above land surface in valleys in areas of ground-water discharge. Large depths to water are more common in highland areas that are remote from streams or other sources of recharge.

The potentiometric surfaces in the three basins containing the Uinta–Animas aquifer are similar in that the surfaces are higher near the margins of the basins and lower near one or two principal streams draining the basins. In the Uinta Basin, the potentiometric surface ranges in altitude from about 5,000 to 8,000 feet, and ground water primarily flows toward the discharge area along the Strawberry River (**fig. 111**). In the Piceance Basin, the potentiometric surface ranges in altitude from about 6,000 to 8,500 feet, and ground water primarily flows toward the discharge areas along Piceance and Yellow Creeks (**fig. 112**). In the San Juan Basin, the potentiometric surface is incompletely known but ranges in altitude from about 5,500 to 7,000 feet in the southern part of the basin (**fig. 113**). The valley of the San Juan River forms the principal area of ground-water discharge in this basin.

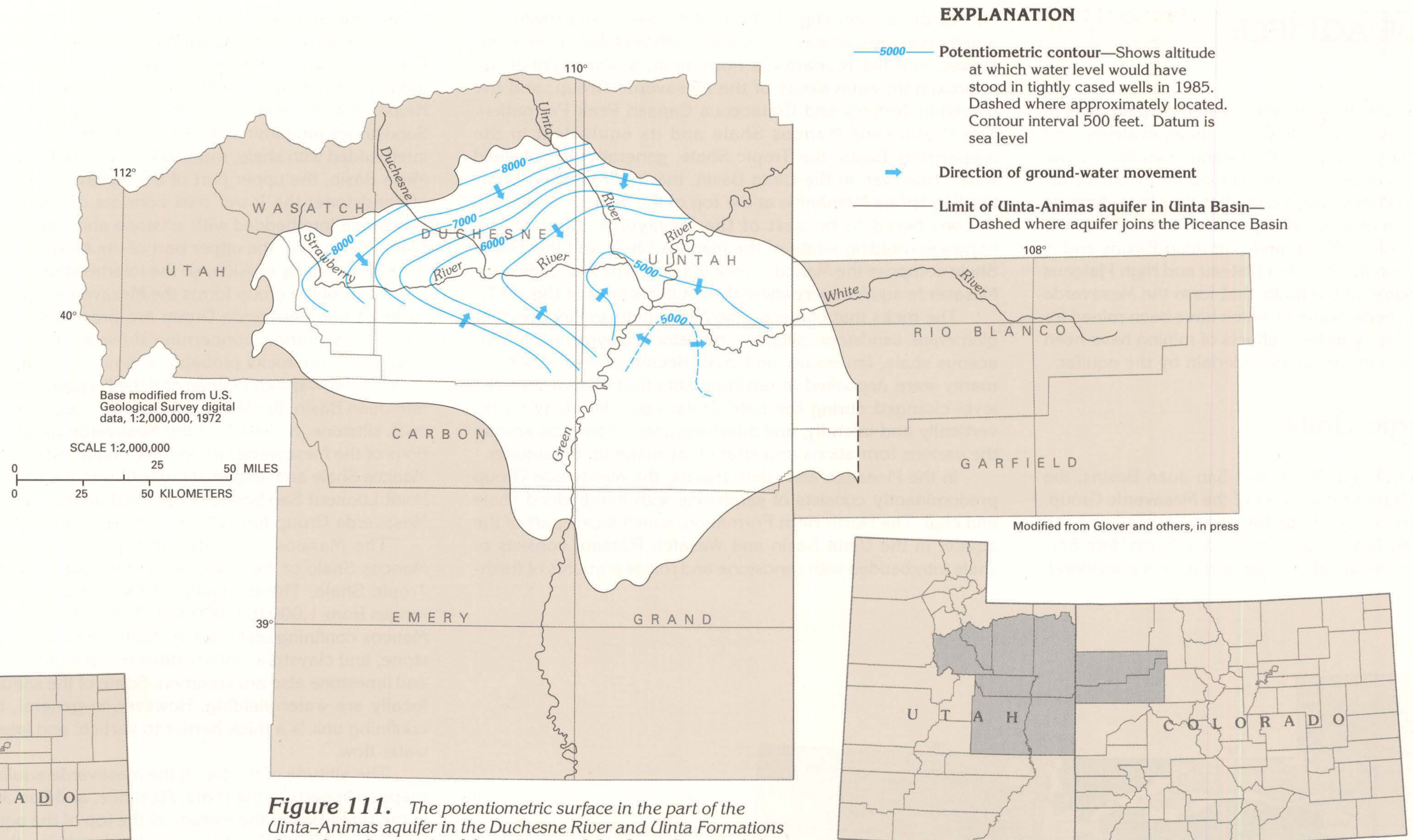


Figure 111. The potentiometric surface in the part of the Uinta–Animas aquifer in the Duchesne River and Uinta Formations slopes from the margin of the unit toward the Green River and its tributaries in the Uinta Basin.

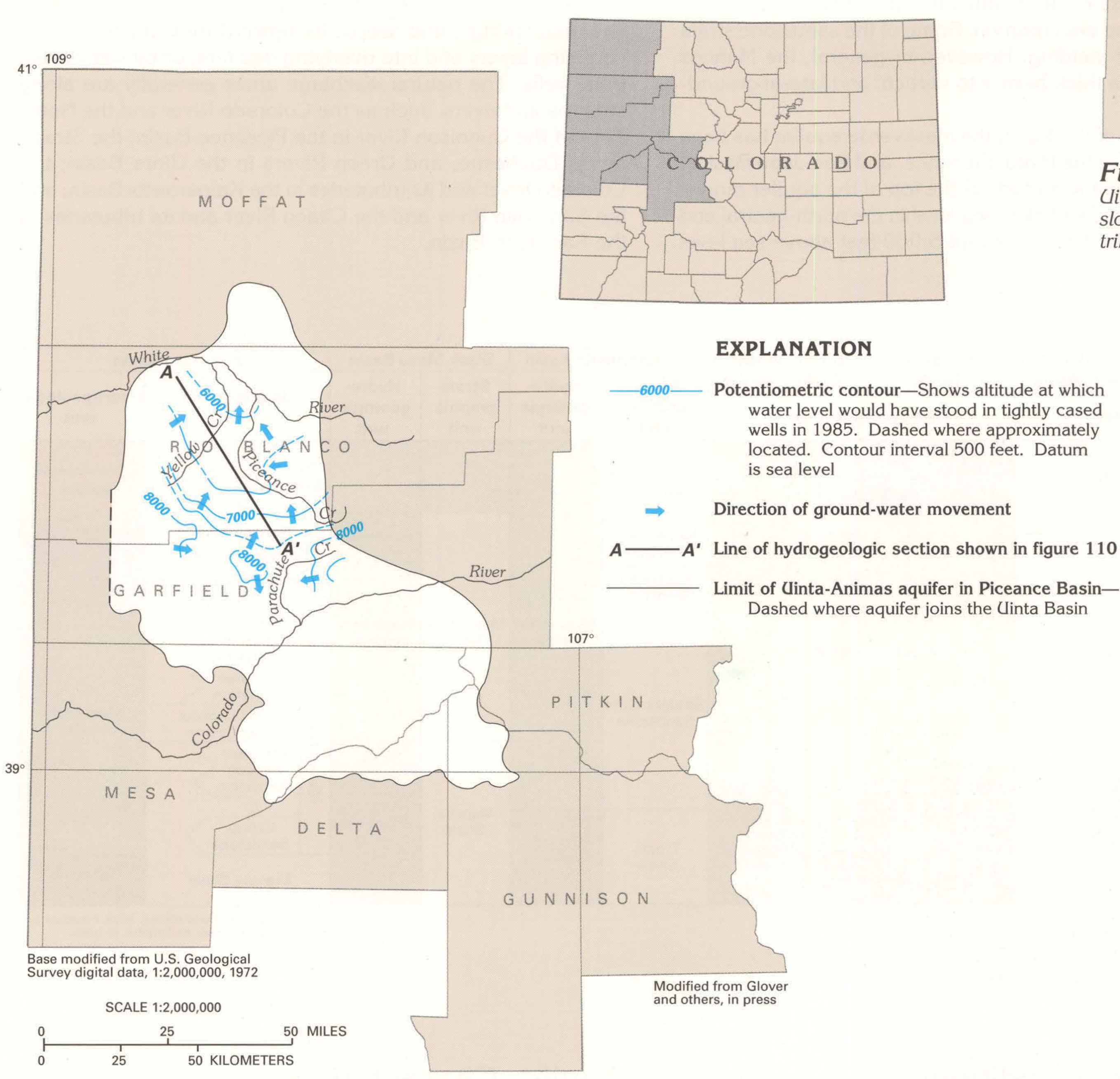


Figure 112. Ground water in the upper part of the Uinta–Animas aquifer (upper part of Parachute Creek Member of Green River Formation) flows toward Piceance Creek, Yellow Creek, and tributaries of the Colorado River in the Piceance Basin.

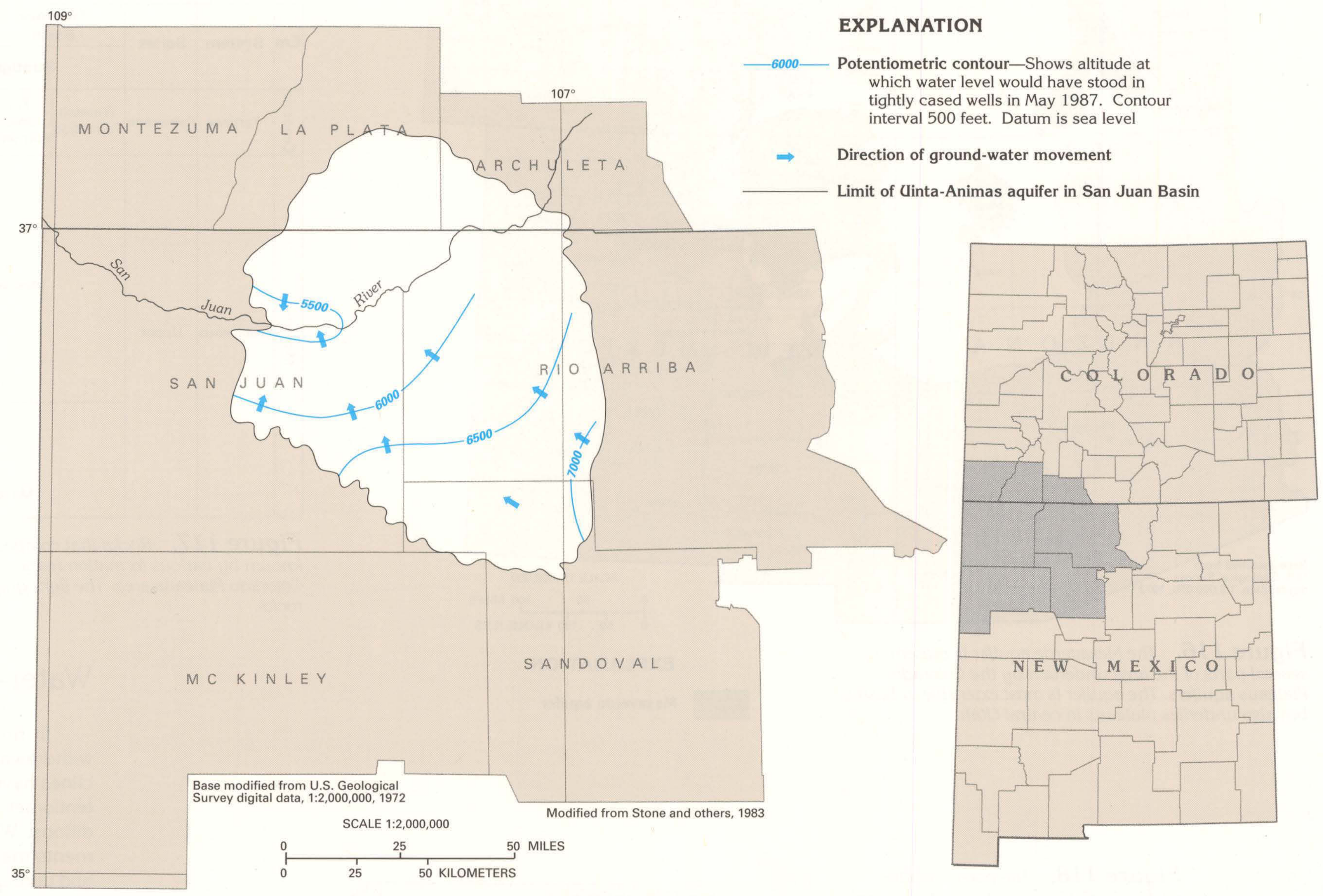


Figure 113. Ground water in the Uinta–Animas aquifer in the Ojo Alamo Sandstone in the San Juan Basin flows toward the San Juan River.

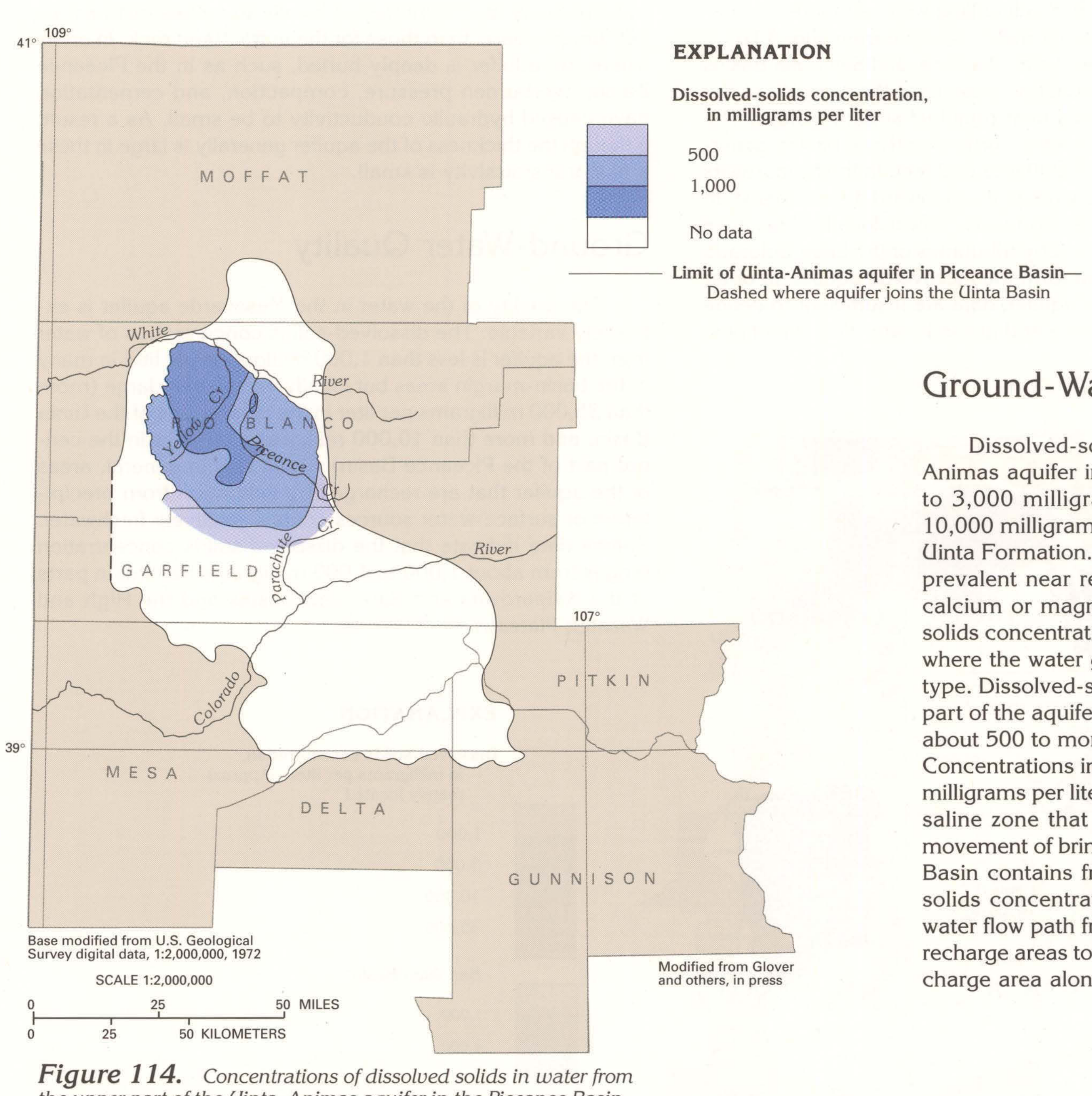


Figure 114. Concentrations of dissolved solids in water from the upper part of the Uinta–Animas aquifer in the Piceance Basin generally increase toward the northern part of the basin.

Ground-Water Quality

Dissolved-solids concentrations in water in the Uinta–Animas aquifer in the Uinta Basin generally range from 500 to 3,000 milligrams per liter; concentrations can exceed 10,000 milligrams per liter in some of the deeper parts of the Uinta Formation. Smaller dissolved-solids concentrations are prevalent near recharge areas where the water usually is a calcium or magnesium bicarbonate type. Larger dissolved-solids concentrations are more common near discharge areas where the water generally is a sodium bicarbonate or sulfate type. Dissolved-solids concentrations in water from the upper part of the aquifer in the Piceance Basin generally range from about 500 to more than 1,000 milligrams per liter (**fig. 114**). Concentrations in the lower part of the aquifer exceed 10,000 milligrams per liter (**fig. 115**) where extensive fracturing of the saline zone that underlies the aquifer has enabled upward movement of brine. The Uinta–Animas aquifer in the San Juan Basin contains fresh to moderately saline water. Dissolved-solids concentrations generally increase along the ground-water flow path from less than 1,000 milligrams per liter near recharge areas to about 4,000 milligrams per liter near the discharge area along the valley of the San Juan River.

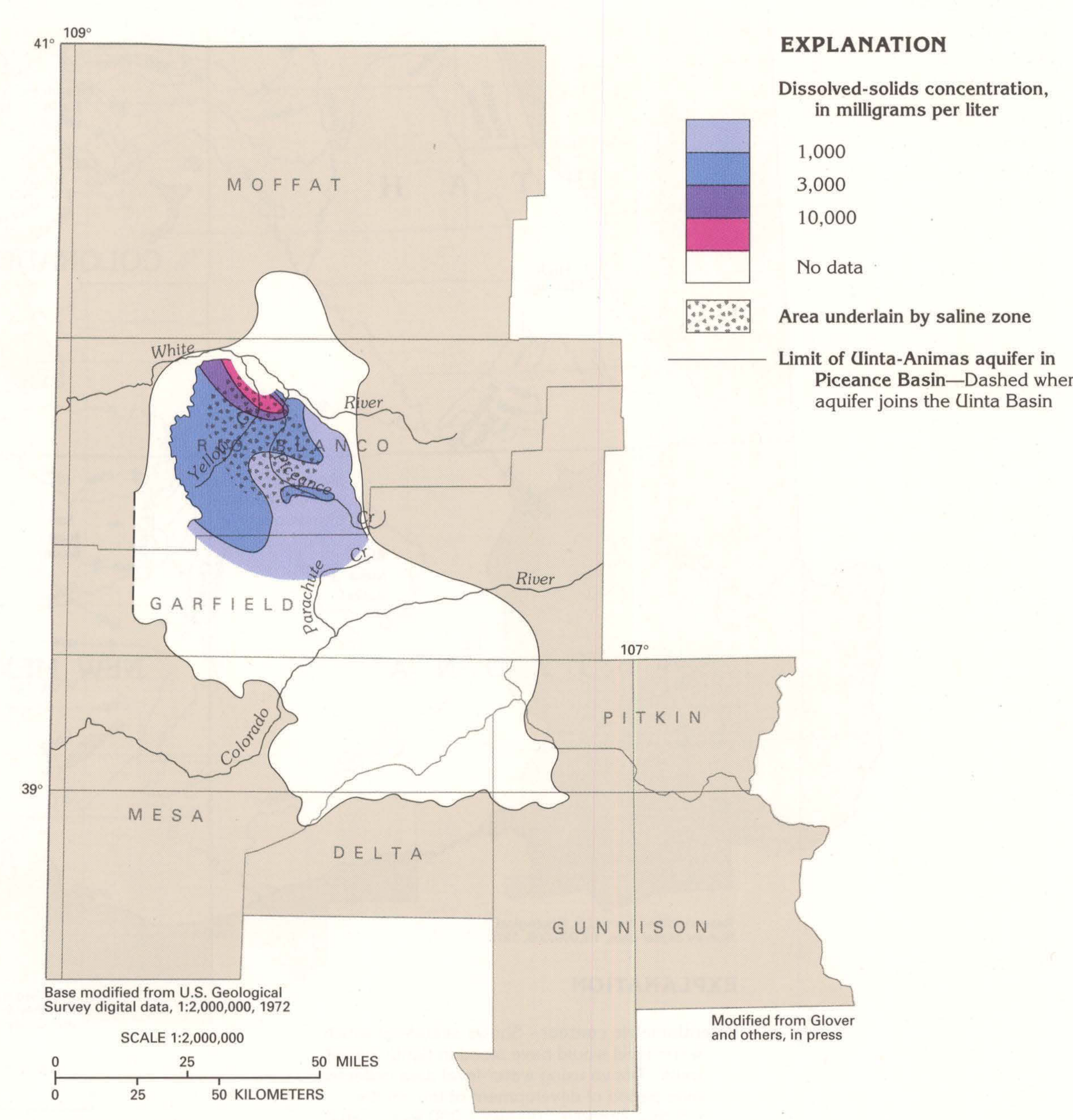


Figure 115. Concentrations of dissolved solids in water from the lower part of the Uinta–Animas aquifer in the Piceance Basin are large. Fractures in the underlying saline zone enable upward flow of saline water that has degraded the water quality in the northern part of the aquifer.

MESAVERDE AQUIFER

The Mesaverde aquifer comprises water-yielding units in the Upper Cretaceous Mesaverde Group, its equivalents, and some adjacent Tertiary and Upper Cretaceous formations. The Mesaverde aquifer is at or near land surface in extensive areas of the Colorado Plateaus and underlies the Uinta–Animas aquifer. The aquifer is of regional importance in the Piceance, Uinta, Kaiparowits, Black Mesa, and San Juan Basins and is of lesser importance in the Wasatch Plateau and High Plateaus areas (**fig. 116**). Some of the rocks that form the Mesaverde aquifer contain coal beds, some of which have been mined for at least a century. The hydrologic effects of mining have been of increasing concern in the areas underlain by the aquifer.

Hydrogeologic Units

In the Piceance, Black Mesa, and San Juan Basins, the Mesaverde aquifer is present in rocks of the Mesaverde Group. In the western part of the Uinta Basin and in parts of the Wasatch Plateau, the Tertiary and Cretaceous North Horn Formation overlies the Mesaverde Group and also is considered part of the aquifer (**fig. 117**). In the Kaiparowits Basin, the aquifer is in the Cretaceous Straight Cliffs and Wahweap Sandstones, and the Kaiparowits Formation, which together are approximate equivalents of the Mesaverde Group, and the overlying Tertiary and Cretaceous Canaan Peak Formation. The Cretaceous Mancos Shale and its equivalent in the Kaiparowits Basin, the Tropic Shale, generally do not yield water. However, in the Uinta Basin, the water-yielding Frontier Sandstone Member is at the top of the Mancos Shale and is considered to be part of the Mesaverde aquifer. The nonwater-yielding strata of the Mancos Shale and the Tropic Shale compose the Mancos confining unit, which underlies the Mesaverde aquifer everywhere the aquifer is present (**fig. 117**).

The rocks that compose the Mesaverde aquifer are conglomerate, sandstone, siltstone, mudstone, claystone, carbonaceous shale, limestone, and coal. Because these rocks primarily were deposited in environments that changed as sea level changed during the Late Cretaceous, lithology varies vertically and laterally, and intertonguing is common among the various formations and strata that make up the aquifer.

In the Piceance and Uinta Basins, the Mesaverde Group predominantly consists of sandstone with interbedded shale and coal. The North Horn Formation, which forms part of the aquifer in the Uinta Basin and Wasatch Plateau, consists of shale interbedded with sandstone and minor amounts of freshwater limestone and conglomerate. In the Kaiparowits Basin, the upper part of the Mesaverde aquifer is in the Canaan Peak Formation, which mainly consists of conglomerate and conglomeratic sandstone with minor amounts of mudstone. The Kaiparowits Formation and the Wahweap and Straight Cliffs Sandstones predominantly consist of fine to coarse sandstone interbedded with shale, mudstone, and coal beds. In the Black Mesa Basin, the upper part of the Mesaverde Group consists of sandstone; the lower part consists of sandstone or silty sandstone interbedded with siltstone and coal. In most of the Black Mesa area, the upper part of the Mesaverde Group has been removed by erosion, so the interbedded sequence of the lower part of the group forms the Mesaverde aquifer. Although rocks of the Mesaverde Group are present on the High Plateaus, information concerning these rocks is sparse. The lithology of the rocks probably is similar to that of equivalent rocks in the Wasatch Plateau and the Kaiparowits Basin. In the San Juan Basin, the Mesaverde aquifer consists of sandstone, coal, siltstone, and shale of the Mesaverde Group. The formations of the Mesaverde Group intertongue extensively with the Mancos Shale and, to a lesser extent, with the Lewis Shale. The Point Lookout Sandstone is the most areally extensive of the Mesaverde Group formations in the San Juan Basin.

The Mancos confining unit generally comprises the Mancos Shale or its equivalent in the Kaiparowits Basin, the Tropic Shale. The thickness of the confining unit typically ranges from 1,000 to 6,000 feet. The rocks that compose the Mancos confining unit predominantly are marine shale, mudstone, and claystone; interbedded minor sandstone, siltstone, and limestone also are common. Some of the sandstone strata locally are water-yielding. However, in general, the Mancos confining unit is a thick barrier to vertical and lateral groundwater flow.

The altitude of the top of the Mesaverde aquifer has been mapped in parts of the Uinta, Piceance, and San Juan Basins. In the Uinta Basin, the altitude of the top of the aquifer ranges from about 10,000 feet below sea level in the north-central and deepest part of the basin to about 5,000 feet above sea level near the margins of the basin. In the Piceance Basin, the top of the aquifer ranges in altitude from about sea level in the central part of the basin to between 5,000 and 7,500 feet above sea level near the margins of the basin. In the San Juan Basin, the top of the aquifer is about 2,500 to 5,000 feet above sea level. In the Piceance and Uinta Basins, the thickness of the Mesaverde aquifer generally is between 2,000 and 4,000 feet. However, the thickness exceeds 7,000 feet locally in the eastern part of the Piceance Basin and is less than 1,000 feet near the margins of the basins. In the San Juan Basin, the Mesaverde aquifer has a maximum thickness of about 4,500 feet in the southern part of the basin.

Recharge and Discharge

Water generally recharges the Mesaverde aquifer in upland areas that receive more precipitation than lower altitude areas. In the Piceance Basin, recharge occurs on the northern flanks of the West Elk Mountains, in the area near Grand Mesa, and along the Roan Plateau. Ground water in the Uinta Basin is recharged near the basin margins. Interbasin flow from the Piceance Basin contributes water to the Uinta Basin. Ground-water flow directions in much of the west-central part of the Uinta Basin are poorly defined by available data. The available data in the San Juan Basin indicate recharge in the area of the Zuni Uplift, Chuska Mountains, and in northern Sandoval County, N. Mex.

Ground water discharges from the aquifer directly to streams, springs, and seeps, by upward movement through confining layers and into overlying aquifers, or by withdrawal from wells. The natural discharge areas generally are along streams and rivers, such as the Colorado River and the North Fork of the Gunnison River in the Piceance Basin; the Strawberry, Duchesne, and Green Rivers in the Uinta Basin; the Colorado River and its tributaries in the Kaiparowits Basin; and the San Juan River and the Chaco River and its tributaries in the San Juan Basin.

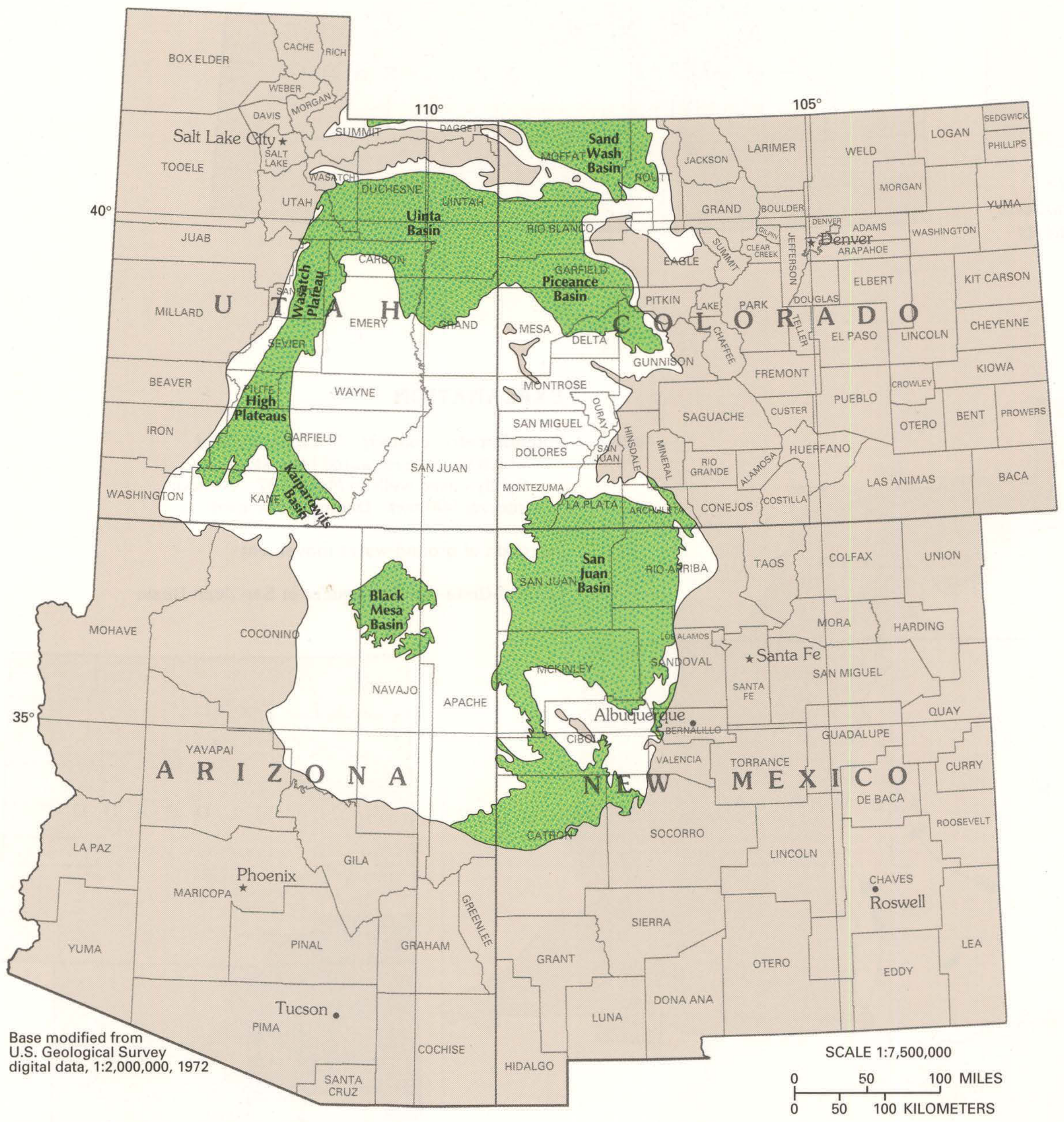


Figure 116. *The Mesaverde aquifer is present in several parts of the area underlain by the Colorado Plateaus aquifers. The aquifer is most extensive in basins but also underlies plateaus in central Utah.*

EXPLANATION

Mesaverde aquifer

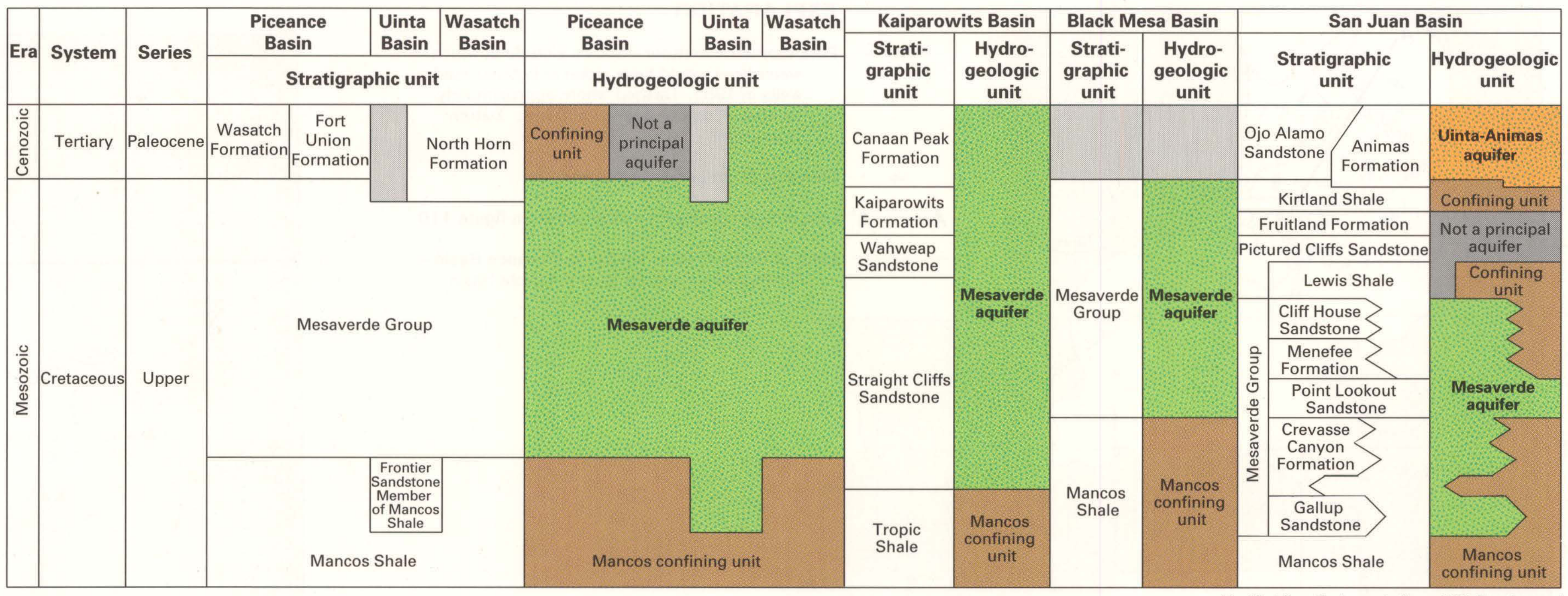


Figure 117. *Rocks that compose the Mesaverde aquifer are known by various formation names in the principal basins of the Colorado Plateaus area. The light gray areas represent missing rocks.*

Figure 118. *The potentiometric surface in the Mesaverde aquifer is poorly defined in some areas. Where data are available to define the surface, it ranges in altitude from about 5,000 feet to about 8,000 feet.*

EXPLANATION

6000 — **Potentiometric contour**—Shows altitude at which water level would have stood in tightly cased wells. Drawn using water-level data collected over period of development of the aquifer. Contour intervals 500 and 1,000 feet. Datum is sea level

Direction of ground-water movement

Limit of Mesaverde aquifer

Water-Level Conditions

In most areas of the Mesaverde aquifer, ground-water withdrawals have been small. Consequently, water-level declines have been limited to localized areas; elsewhere, the potentiometric surface generally represents predevelopment conditions. Water-level measurements and reports of measurements made during the period of development of the aquifer and during oil and gas test-well drilling were combined to generate a generalized potentiometric-surface map (**fig. 118**).

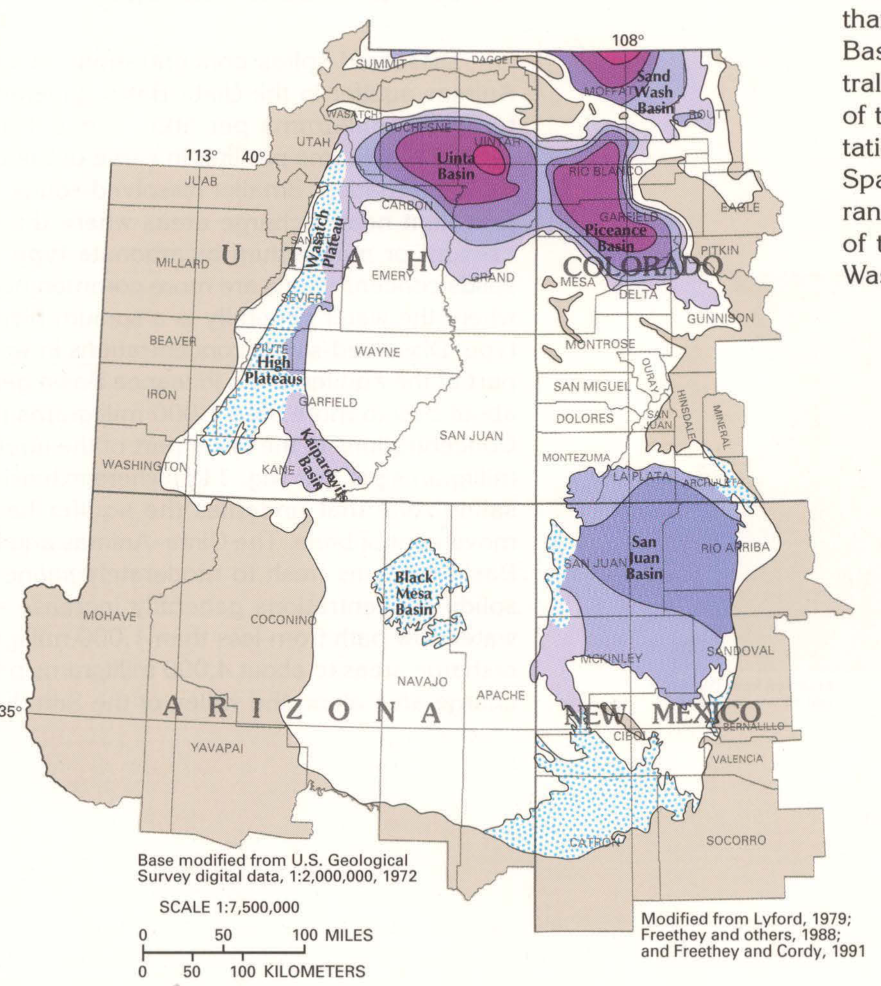
Ground water in the Uinta, Piceance, and San Juan Basins generally flows from recharge areas near the margins of the basins to discharge areas near principal stream valleys. The altitude of the potentiometric surface in these basins generally ranges from about 5,000 to 8,000 feet. In the Kaiparowits Basin, ground-water flow generally is toward the southeast. In the Black Mesa Basin, ground-water flow is localized because of the shallow canyons cut by tributaries of the Little Colorado River into the rocks that form the Mesaverde aquifer. In other areas of the Mesaverde aquifer, data are insufficient to define the potentiometric surface and ground-water flow directions.

Aquifer Characteristics

Transmissivity of the Mesaverde aquifer is less than 50 feet squared per day in large areas of the Colorado Plateaus but exceeds 2,000 feet squared per day locally in the western part of the Uinta Basin and the eastern part of the Wasatch Plateau. Fracturing of rocks that form the Mesaverde aquifer locally increases the secondary permeability; as a result, the transmissivity also is increased locally to values as much as 100 times greater than those for the unfractured rock. In areas where the aquifer is deeply buried, such as in the Piceance Basin, overburden pressure, compaction, and cementation have caused hydraulic conductivity to be small. As a result, although the thickness of the aquifer generally is large in these areas, transmissivity is small.

Ground-Water Quality

The quality of the water in the Mesaverde aquifer is extremely variable. The dissolved-solids concentration of water from the aquifer is less than 1,000 milligrams per liter in many of the basin-margin areas but locally can be very large (more than 35,000 milligrams per liter in the central part of the Uinta Basin, and more than 10,000 milligrams per liter in the central part of the Piceance Basin) (**fig. 119**). In general, areas of the aquifer that are recharged by infiltration from precipitation or surface-water sources contain relatively freshwater. Sparse data indicate that the dissolved-solids concentration ranges from about 1,000 to 4,000 milligrams per liter in parts of the Kaiparowits and San Juan Basins and the High and Wasatch Plateaus.



EXPLANATION

Dissolved-solids concentration, in milligrams per liter—Approximately located

1,000
3,000
10,000
35,000

San Juan Basin

1,000
4,000

No data

Mesaverde aquifer absent

Figure 119. *Concentration of dissolved solids in water from the Mesaverde aquifer is extremely variable; it is less than 1,000 milligrams per liter near many basin margins and exceeds 35,000 milligrams per liter near the center of the Uinta Basin.*

DAKOTA–GLEN CANYON AQUIFER SYSTEM

Water-yielding rocks ranging in age from late Cretaceous to Triassic underlie most of the Colorado Plateaus area. These rocks contain a series of aquifers and confining units, which, for the purposes of this chapter, are referred to as the Dakota–Glen Canyon aquifer system. In much of the area underlain by the aquifer system (**fig. 120**), the great depth to the aquifers or poor water quality make the aquifers unsuitable for development. However, in areas where an aquifer is near land surface, the aquifer may be an important source of water.

Rocks that compose the Dakota–Glen Canyon aquifer system are older than the Mancos and Tropic Shales, which form the overlying Mancos confining unit; and are younger than the Chinle, Ankareh, or Moenkopi Formations, which form the underlying Chinle–Moenkopi confining unit. In general, both confining units are thick, low-permeability zones that severely restrict vertical flow between the Dakota–Glen Canyon aquifer system and overlying and underlying aquifers.

The Dakota–Glen Canyon aquifer system includes four permeable zones that herein are referred to as the Dakota aquifer, the Morrison aquifer, the Entrada aquifer, and the Glen Canyon aquifer. The units that form the bulk of these aquifers are, respectively: (1) The Dakota Sandstone and adjacent water-yielding rocks; (2) water-yielding rocks generally of the lower part of the Morrison Formation; (3) the Entrada Sandstone and its equivalent in the western part of the Uinta Basin, the Preuss Sandstone; and (4) the Glen Canyon Sandstone or Group and its equivalent in the western part of the Uinta Basin, the Nugget Sandstone. These rocks are at land surface or at reasonable drilling depths below land surface primarily on the flanks of the San Rafael Swell, White River, and Circle Cliffs Uplifts, in the Henry Mountains Basin, in parts of the Paradox Basin, Uncompahgre Uplift, and Four Corners Platform, in the Black Mesa Basin, and in the Acoma Sag (**fig. 120**). The stratigraphic relations among the formations that contain these aquifers and the adjacent confining units are shown in **figure 121**.

Sandstone, conglomerate, and conglomeratic sandstone are the major water-yielding materials in this series of aquifers. The aquifers commonly also contain interbedded siltstone. Mudstone, claystone, siltstone, shale, and limestone generally form the confining units that separate these aquifers (**table 1**).

The aquifers described in this section are grouped together as an aquifer system because they are separated everywhere from overlying and underlying aquifers by thick confining units and because some hydraulic connection exists between each of the aquifers in the system at some point in the Colorado Plateaus area. For example, in the Black Mesa Basin, the Morrison and Curtis–Stump confining units are missing; as a result, the Dakota, Morrison, and Entrada aquifers are in direct contact (**fig. 122**). This contact likely allows interaquifer flow among these three aquifers, although the rate of interaquifer flow may be limited by low-permeability zones within the aquifers. The confining units in the aquifer system generally are not as thick as the more substantial Mancos and Chinle–Moenkopi confining units, and interaquifer flow is more likely among the aquifers of the Dakota–Glen Canyon aquifer system than between these aquifers and those that overlie or underlie the aquifer system.

In a regional context, recharge areas, discharge areas, ground-water flow directions, and water quality are similar among the four aquifers. The uppermost aquifer (the Dakota) and the lowermost aquifer (the Glen Canyon) are best defined by data, and these two aquifers are discussed here as examples of the hydrogeology near the top and bottom of the aquifer system.

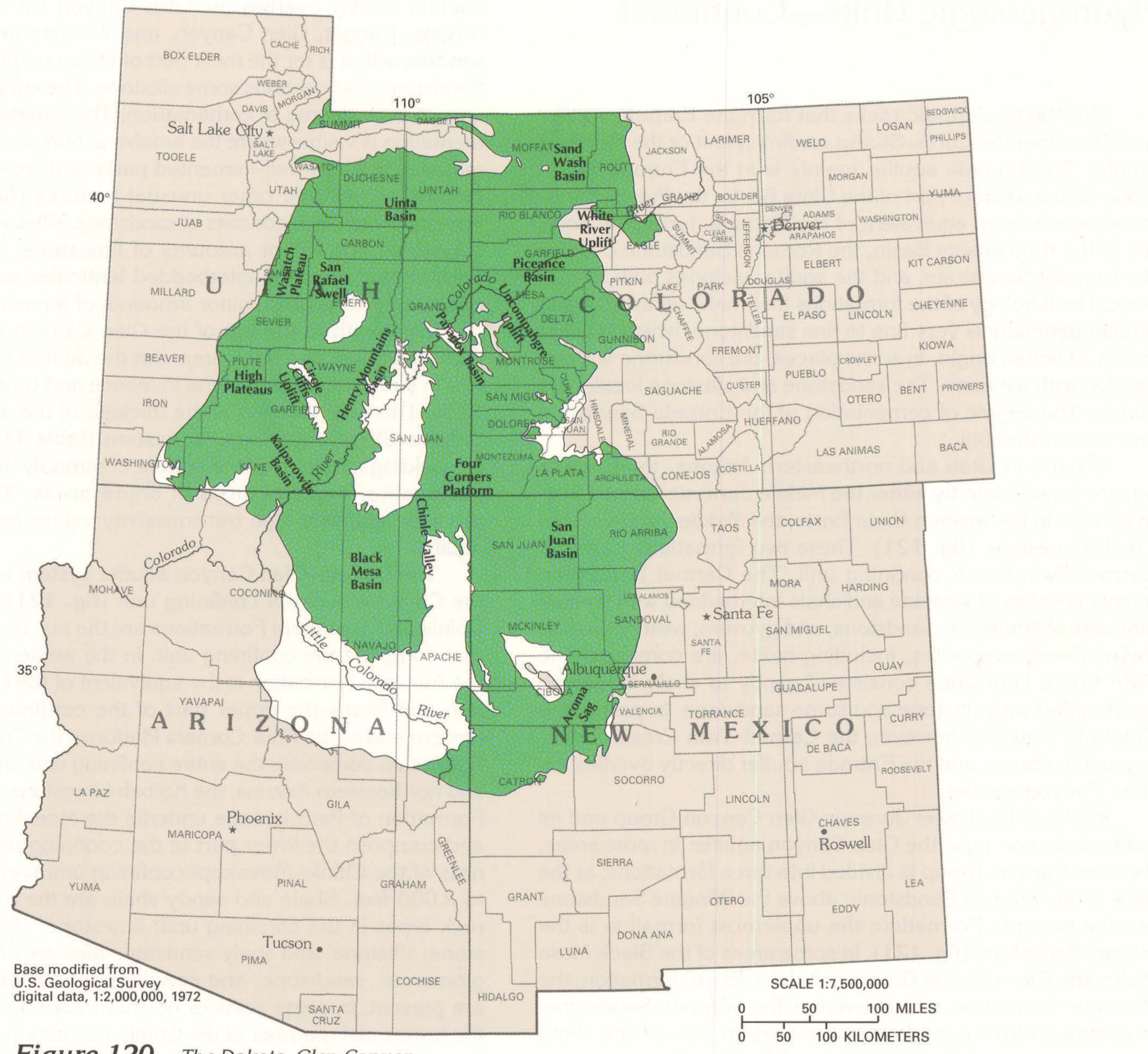


Figure 120. *The Dakota–Glen Canyon aquifer system underlies most of the Colorado Plateaus area.*

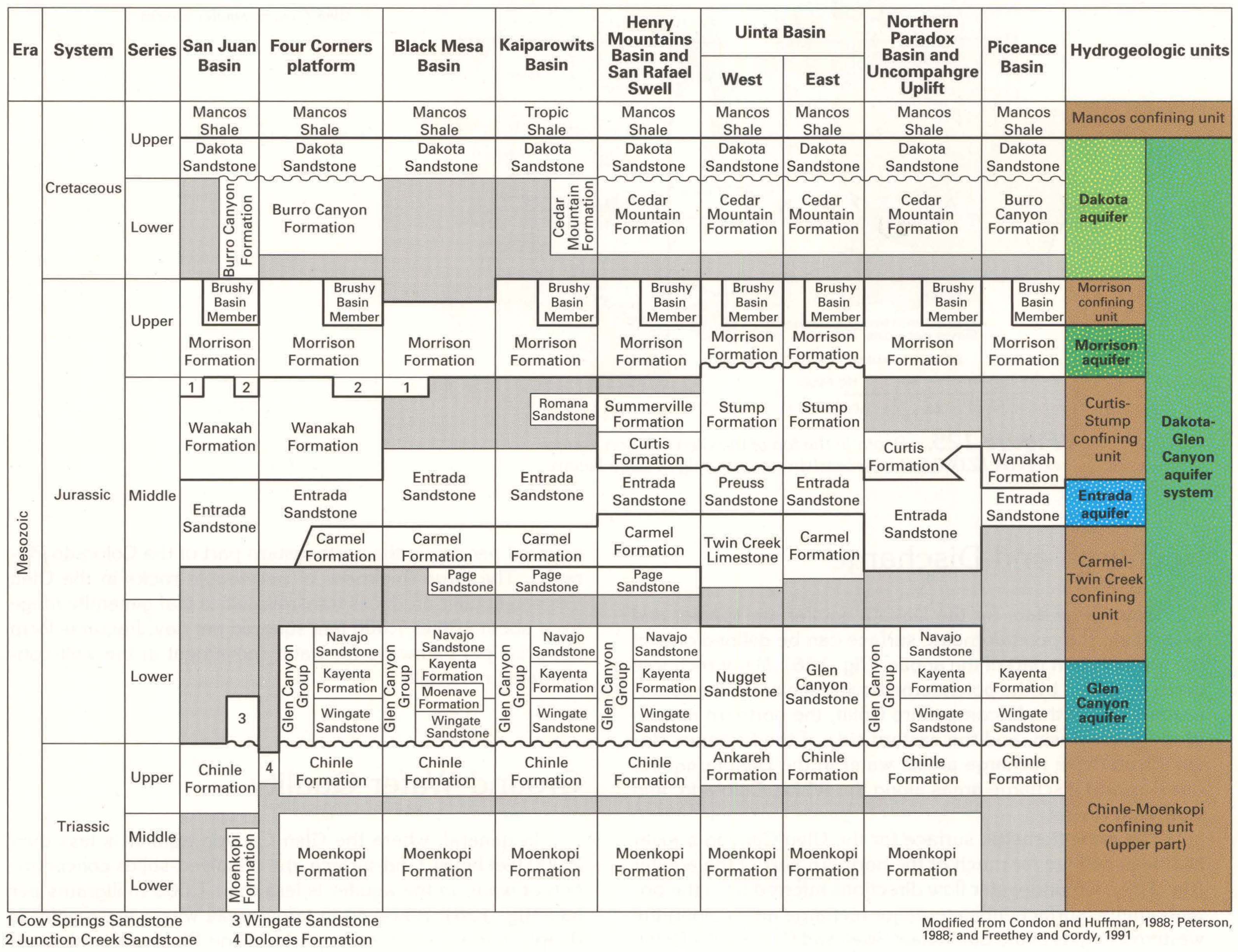


Figure 121. *The four aquifers and three confining units of the Dakota–Glen Canyon aquifer system are in several geologic formations; some formations are absent in some areas. The gray area represents missing rock.*

Table 1. *The lithologic character and thickness of aquifers and confining units in the Dakota–Glen Canyon aquifer system are variable. The aquifers commonly consist of sandstone and conglomerate; the confining units commonly consist of siltstone and shale*

[Modified from Freethey and Cordy, 1991]

Geohydrologic unit	Principal lithology	Typical thickness of unit, in feet (queried where thickness in doubt): Northwestern Colorado and northeastern Utah	East-central Utah	South-central Utah	South-western Colorado	North-western New Mexico	North-eastern Arizona
Dakota aquifer	Sandstone Conglomerate Mudstone	200	300	Missing	200	200	100
Morrison confining unit	Siltstone Mudstone Claystone	400	250	Missing	300	185	Not divided
Morrison aquifer	Sandstone Conglomeratic sandstone Siltstone Mudstone	300	400	Missing	450	250	200 to 600
Curtis–Stump confining unit	Siltstone Shale Sandstone	250	400	Missing	150	500	0 to 275 (?)
Entrada aquifer	Sandstone Siltstone	500	550	600	150	300	50 to 350
Carmel–Twin Creek confining unit	Siltstone Shale Limestone	500	350	160	Missing	Missing	0 to 300
Glen Canyon aquifer	Sandstone	750	1,100	2,100	600	Missing	200 to 2,300

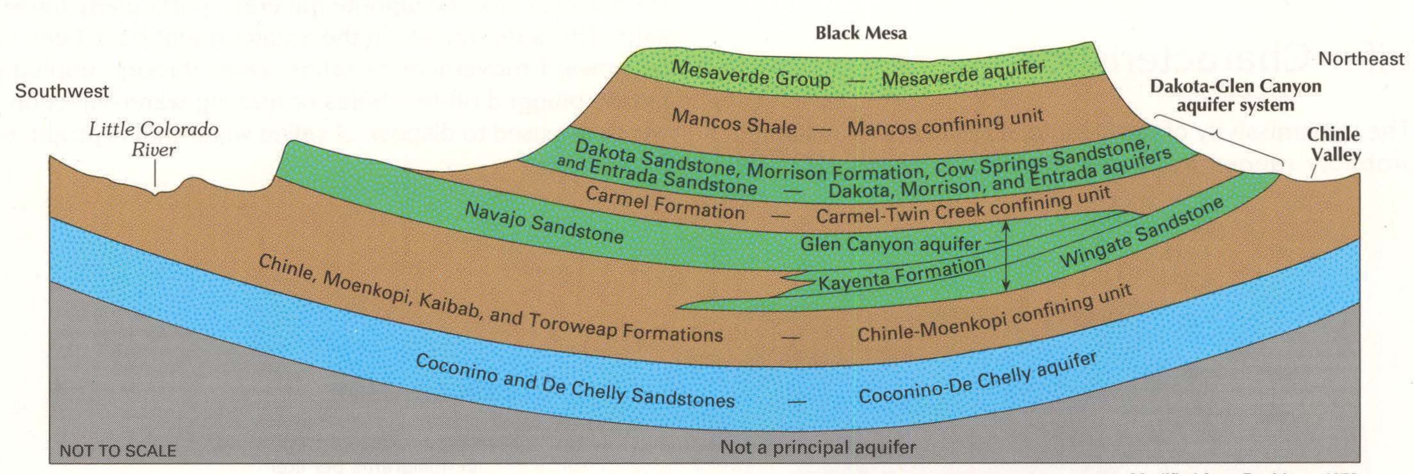


Figure 122. *This generalized hydrogeologic section through the Black Mesa Basin shows the relation of the Dakota–Glen Canyon aquifer system to the overlying and underlying hydrogeologic units.*

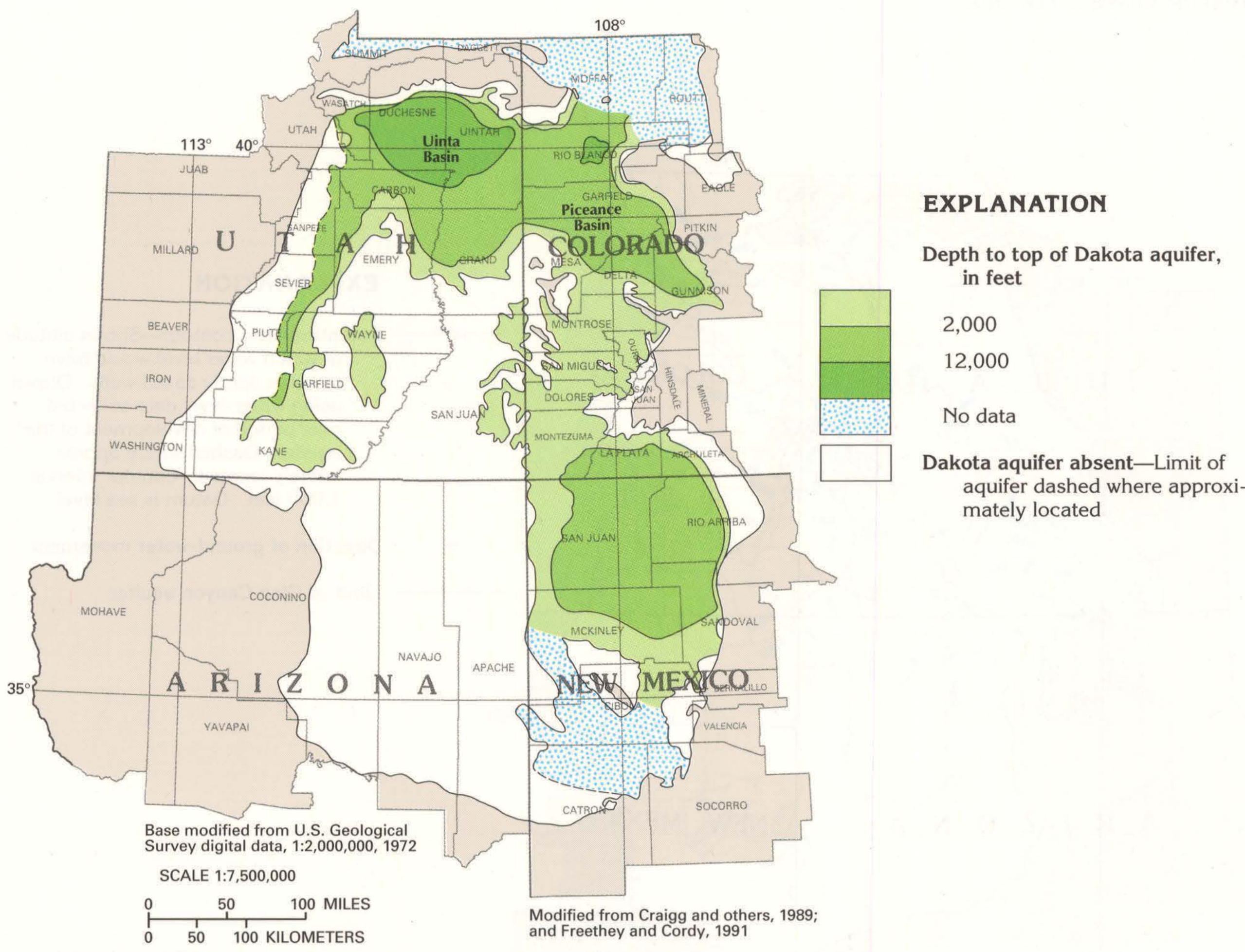


Figure 123. *Depth to the top of the Dakota aquifer is less than 2,000 feet in extensive areas of Colorado, New Mexico, and Utah.*

Hydrogeologic Units

The Dakota aquifer is in the Upper Cretaceous Dakota Sandstone and underlying Lower Cretaceous Burro Canyon and Cedar Mountain Formations (**fig. 121**). The lithology of the Dakota Sandstone varies widely and includes conglomerate, sandstone, siltstone, mudstone, carbonaceous shale, and coal. Three units can be recognized over a large area: a basal conglomeratic sandstone; a middle sequence of interbedded carbonaceous shale, impure coal, and lenticular sandstone and siltstone; and an upper, massive, fine to medium sandstone. Sandstone, which is commonly interbedded with thin mudstone beds, constitutes about one-half of the total thickness of the Burro Canyon Formation; in some places, the sandstone forms a single, thick bed. Minor chert and limestone beds also are present in the formation. The lithology of the Cedar Mountain Formation is similar to that of the Burro Canyon Formation, except that sandstone generally composes less than 30 percent of the thickness of the Cedar Mountain. In some places, the Cedar Mountain Formation includes a basal conglomeratic sandstone unit. The Dakota aquifer is present in the Piceance and Uinta Basins, along the Wasatch and High Plateaus, in the Kaiparowits, Henry Mountains, Black Mesa, and San Juan Basins, in the eastern part of the Four Corners Platform, and in parts of the Paradox Basin and Uncompahgre Uplift (**fig. 120**). The depth to the top of the aquifer is less than 2,000 feet in many areas but exceeds 12,000 feet in parts of the Piceance and Uinta Basins (**fig. 123**).

The Upper Jurassic Morrison Formation underlies the Dakota aquifer in the Colorado Plateaus (**fig. 121**). In most parts of the area, the Morrison Formation includes an upper, nonwater-yielding unit called the Brushy Basin Member, which forms the Morrison confining unit. This member mainly consists of relatively impermeable siltstone, mudstone, and claystone. The member is absent in the Black Mesa Basin.

The middle and lower parts of the Morrison Formation consist of interbedded fine to medium sandstone, siltstone, and mudstone. This sequence is called the Morrison aquifer, although only the coarser grained strata generally can be expected to yield water. In the Four Corners Platform and San Juan and Black Mesa Basins, the Morrison aquifer includes two underlying water-yielding sandstone units, the Middle Jurassic Cow Springs and Junction Creek Sandstones.

In most places in the Colorado Plateaus, the Morrison aquifer is underlain by nonwater-yielding Middle Jurassic rocks that form the Curtis–Stump confining unit. The formations that make up the Curtis–Stump confining unit are the Curtis, Summerville, Stump, and Wanakah Formations. These formations predominantly consist of siltstone with interbedded shale and sandstone. Minor amounts of limestone and gypsum also are present.

Hydrogeologic Units—Continued

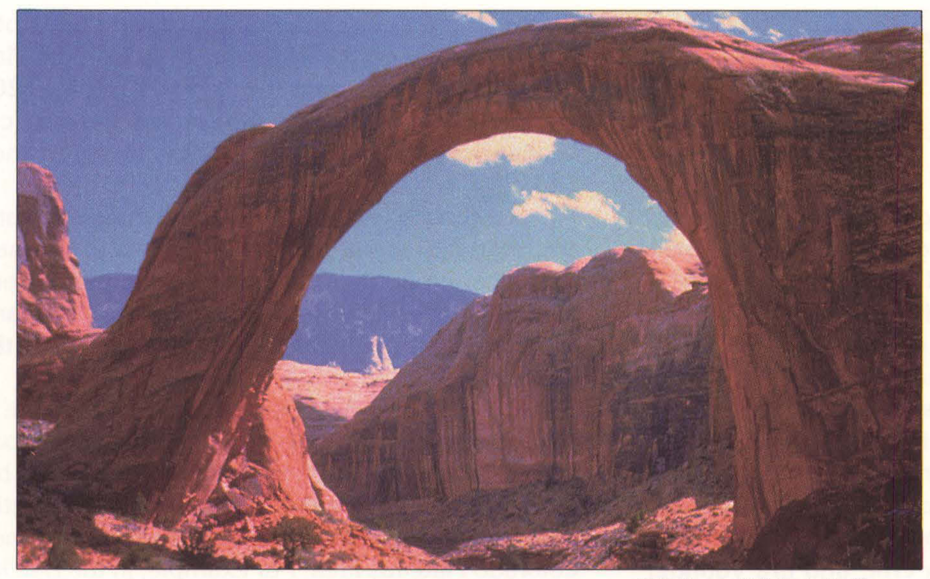
The Middle Jurassic rocks that form the Entrada aquifer underlie either the Curtis–Stump confining unit or the Morrison aquifer. The Entrada aquifer mainly is in the Entrada Sandstone: in the western part of the Uinta Basin, the Preuss Sandstone, which is an equivalent of the Entrada, forms the aquifer. In the Kaiparowits Basin, the Romana Sandstone overlies the Entrada Sandstone, and the aquifer includes both formations. The lithology of the formations that make up the Entrada aquifer generally is very fine to fine sandstone, which is commonly of eolian origin. In some places, the sandstone is interbedded with siltstone. The sandstone and siltstone locally are clayey. The degree of cementation of the Entrada Sandstone varies considerably.

In parts of Utah and northeastern Arizona, the Entrada aquifer is underlain by either the Middle Jurassic Carmel Formation or, in the western Uinta Basin, the Middle Jurassic Twin Creek Limestone (**fig. 121**). These two formations form the Carmel–Twin Creek confining unit. The Carmel Formation mainly consists of siltstone and shale interbedded with smaller amounts of limestone, sandstone, and gypsum; west of the San Rafael Swell, evaporites, including halite, are common. The Twin Creek Limestone consists of sandy to shaly limestone interbedded with siltstone and some sandstone. In part of the Colorado Plateaus, however, the Carmel–Twin Creek confining unit is absent, and the Entrada aquifer directly overlies the Glen Canyon aquifer.

Rocks of the Lower Jurassic Glen Canyon Group and its equivalents compose the Glen Canyon aquifer. In most areas, the Glen Canyon Group is divided into three formations; at the base is the Wingate Sandstone; above the Wingate Sandstone lies the Kayenta Formation; the uppermost formation is the Navajo Sandstone (**fig. 121**). In some areas of the Black Mesa Basin, the Glen Canyon Group includes a fourth formation, the Moenave Formation, which overlies the Wingate Sandstone. In northwestern Colorado and the eastern part of the Uinta Basin, the stratigraphic equivalent of the Glen Canyon Group is the Glen Canyon Sandstone, and, in the western Uinta Basin, the equivalent is the Nugget Sandstone. From the San Rafael Swell to the Black Mesa Basin, the Glen Canyon aquifer includes the Middle Jurassic Page Sandstone, which unconformably overlies the Glen Canyon Group. The Page, Navajo, Nugget, Glen Canyon, and Wingate units consist of sandstone that is for the most part of eolian origin; the Wingate Sandstone also contains some siltstone. The eolian sandstones vary in their degree of cementation. The variability of the cementation is visible where the erosive action of water and wind removes the less well-cemented parts of a rock outcrop and creates arches and other unusual features (**fig. 124**). The Kayenta Formation consists of sandstone, siltstone, mudstone, claystone, and minor amounts of limestone. The Moenave Formation comprises interbedded lenticular sandstone, siltstone, claystone, and minor amounts of limestone.

The depth to the top of the Glen Canyon aquifer is less than 2,000 feet in a large area, but the depth exceeds 12,000 feet in substantial parts of the Piceance and Uinta Basins (**fig. 125**). The Glen Canyon is the thickest of the aquifers of the Dakota–Glen Canyon aquifer system (**table 1**), and the water-yielding materials in the aquifer commonly are well sorted, permeable, and fractured in some areas. These factors produce relatively high transmissivity values for much of the aquifer.

The Dakota–Glen Canyon aquifer system is underlain by the Chinle–Moenkopi confining unit (**fig. 121**). The Triassic Chinle and Moenkopi Formations are the two main formations that compose the confining unit. In the western Uinta Basin, the Ankareh Formation is the equivalent of the Chinle Formation and forms the upper part of the confining unit. In the eastern end of the Four Corners Platform, the Triassic Dolores Formation composes the entire confining unit. In eastern Utah and northeastern Arizona, the Kaibab Limestone and Toroweap Formation of Permian age underlie the Moenkopi Formation and compose the lower part of the confining unit. The thickness of the Chinle–Moenkopi confining unit typically is 1,000 to 2,000 feet. Shale and sandy shale are the most prevalent rock types in the confining unit; limestone, claystone, mudstone, siltstone, and shaly sandstone also are common. Conglomerate, sandstone, and conglomeratic sandstone locally are present. In some parts of northern Arizona, sandstone in the lowermost member of the Chinle Formation or the Kaibab Limestone yields small amounts of water to wells. Elsewhere, the formations generally do not yield water. Overall, the Chinle–Moenkopi confining unit is an effective barrier to interaquifer ground-water flow and forms the base of the Dakota–Glen Canyon aquifer system.



W.R. Hansen, U.S. Geological Survey

Figure 124. *Well-cemented parts of the Navajo Sandstone at Rainbow Bridge National Monument, Utah, have been preserved as a natural bridge. Less-cemented parts of the sandstone have been eroded away.*

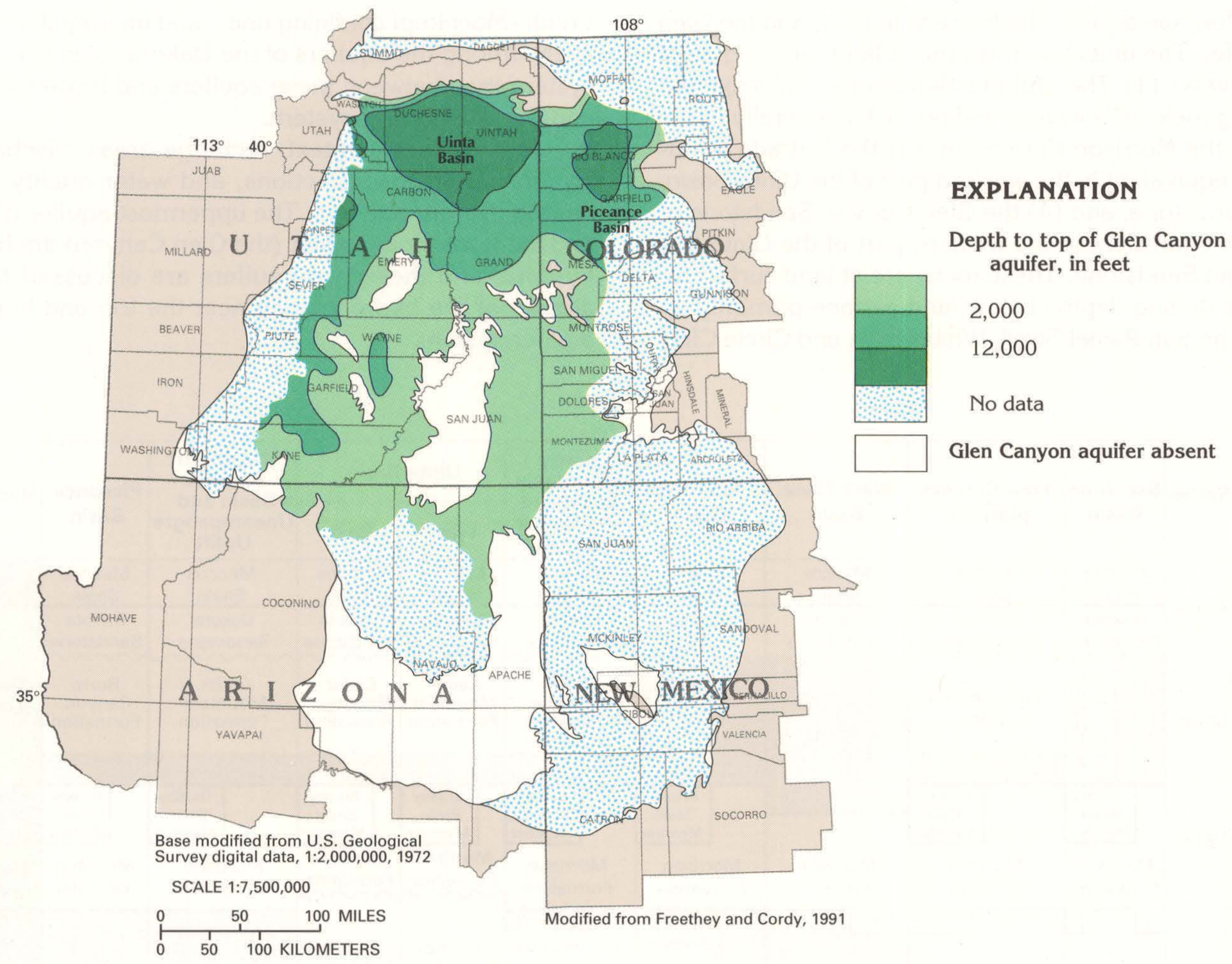


Figure 125. *Depth to the top of the Glen Canyon aquifer exceeds 12,000 feet in several areas, primarily in large basins.*

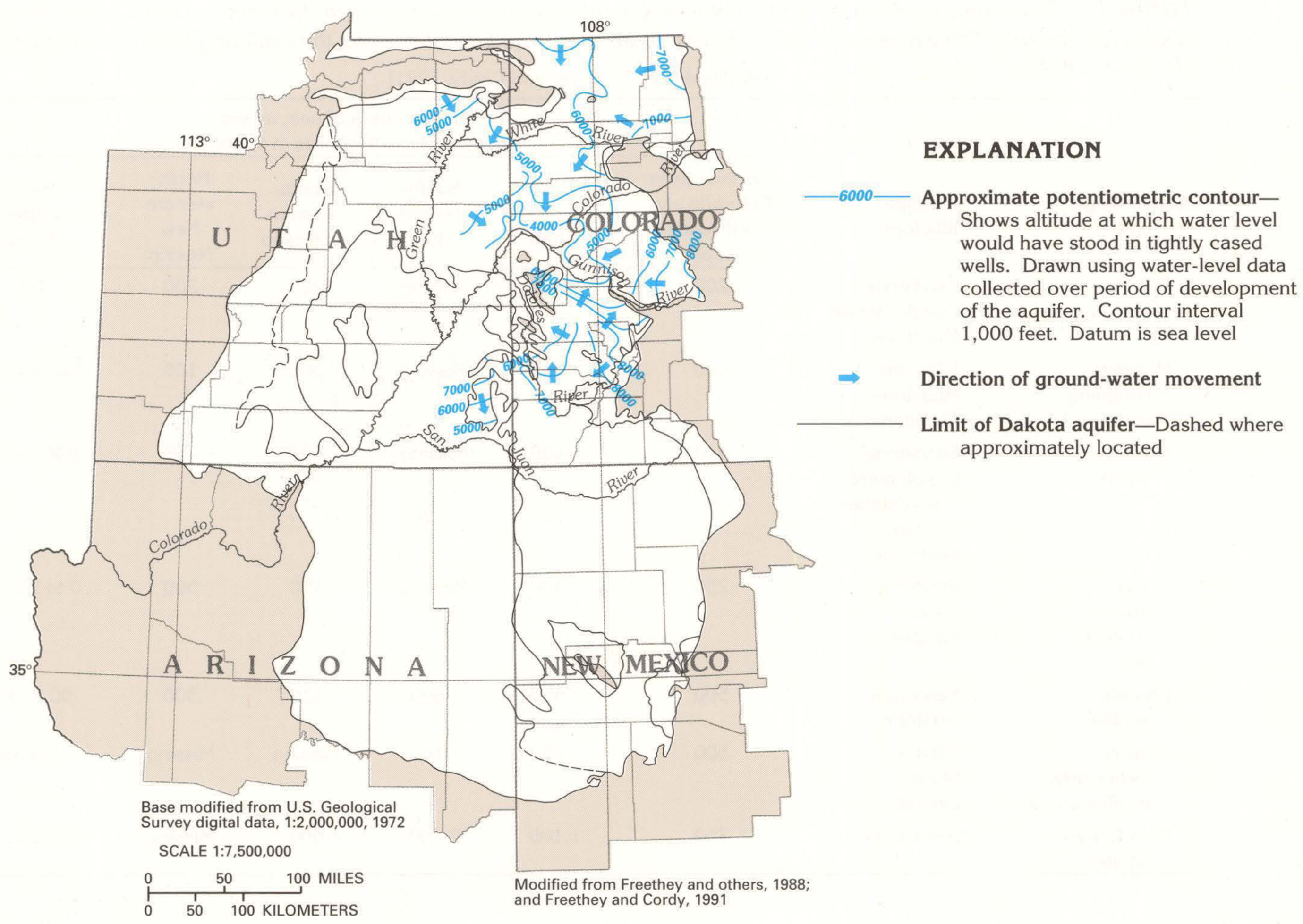


Figure 126. *Ground water in the Dakota aquifer generally flows from outcrops along the basin margins toward the major streams in the Colorado Plateaus.*

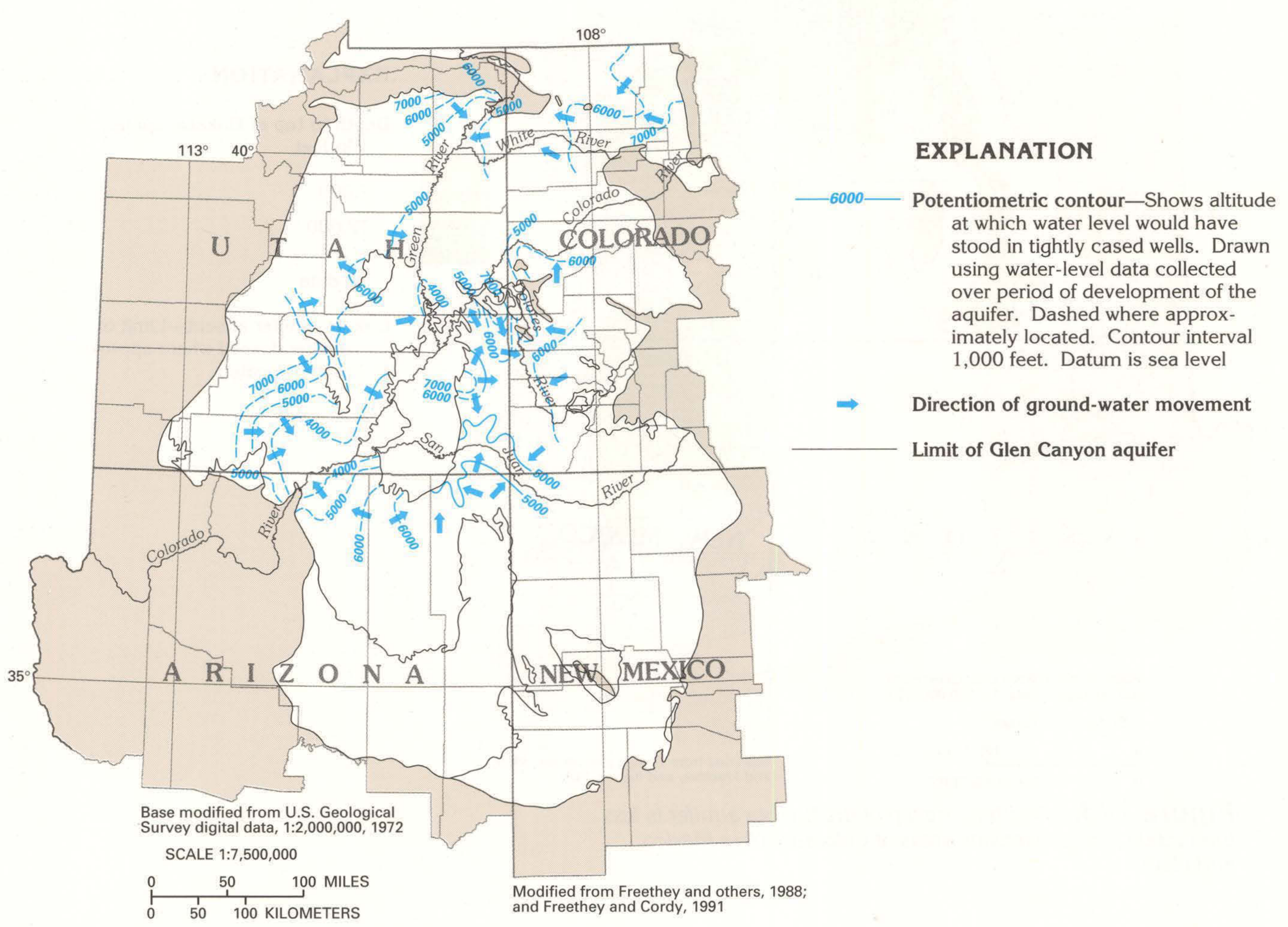


Figure 127. *Ground water in the Glen Canyon aquifer generally flows from recharge areas toward the Colorado River and its main tributaries.*

Recharge and Discharge

Water-level data for the Dakota aquifer are sparse, and as a result, the potentiometric surface can be defined only in the northeastern part of the aquifer (**fig. 126**). Major recharge areas indicated by the potentiometric surface are in the southeastern end of the Uncompahgre Uplift, the northern margin of the Uinta Basin, and the eastern side of the Piceance Basin. From these recharge areas, water in the Dakota aquifer flows toward discharge areas along the White, Colorado, and Gunnison Rivers.

The potentiometric surface for the Glen Canyon aquifer has been defined for much of the northern part of the aquifer (**fig. 127**). Ground-water flow directions inferred from the potentiometric surface indicate major recharge areas along the western margins of the San Rafael Swell and Circle Cliffs Uplift, in the northern part of the Four Corners Platform, in the southeastern parts of the Uncompahgre Uplift and Paradox Basin, at the eastern margin of the Piceance Basin, and at the northeastern margin of the Uinta Basin. Ground-water flow in the Glen Canyon aquifer is toward major discharge areas along the Green, Colorado, Dolores, and San Juan Rivers.

Aquifer Characteristics

The transmissivity of the Dakota aquifer is poorly defined but probably ranges from less than 10 to about 100 feet squared per day in the northeastern part of the Colorado Plateaus. The large thickness of permeable rocks in the Glen Canyon aquifer produces transmissivities that generally range from about 100 to 1,000 feet squared per day; fractures form the principal pathways for water movement in the well-consolidated materials.

Ground-Water Quality

In general, where the Glen Canyon aquifer is less than 2,000 feet below land surface, the dissolved-solids concentration of water in the aquifer is less than 1,000 milligrams per liter (**fig. 128**). However, in large areas where the aquifer is deeply buried, such as in parts of the Piceance and Uinta Basins, the dissolved-solids concentration exceeds 35,000 milligrams per liter. In an area in extreme southeastern Utah where oil and gas exploration and production are concentrated, water in the Glen Canyon aquifer is highly mineralized. Analysis of the water chemistry indicates that the source of the mineralized water likely is deeper strata, which contain substantial deposits of evaporite minerals, particularly halite (rock salt). The water quality in the aquifer might have been caused by upward movement of saline water through unplugged or poorly plugged oil-test holes or leaking water-injection wells, which are used to dispose of saline water that is produced with oil and gas.

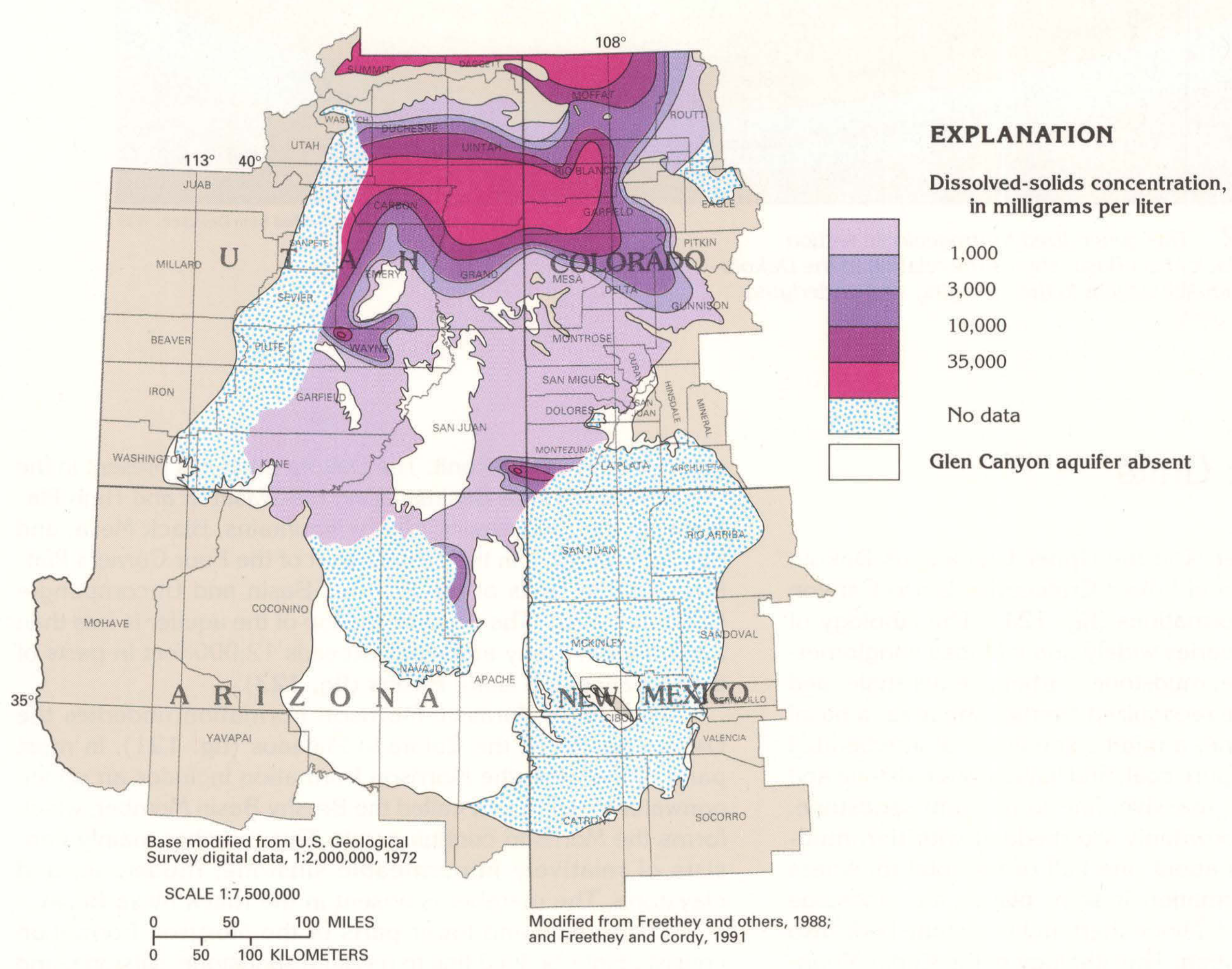


Figure 128. *Although the concentration of dissolved solids in water from the Glen Canyon aquifer is less than 1,000 milligrams per liter in large areas, the concentration exceeds 35,000 milligrams per liter in the deeper parts of the Uinta and Piceance Basins.*

COCONINO–DE CHELLY AQUIFER

Water-yielding rocks of Early Permian age underlie the southern part of the Colorado Plateaus. In this chapter these rocks are referred to as the Coconino–De Chelly aquifer (**fig. 129**).

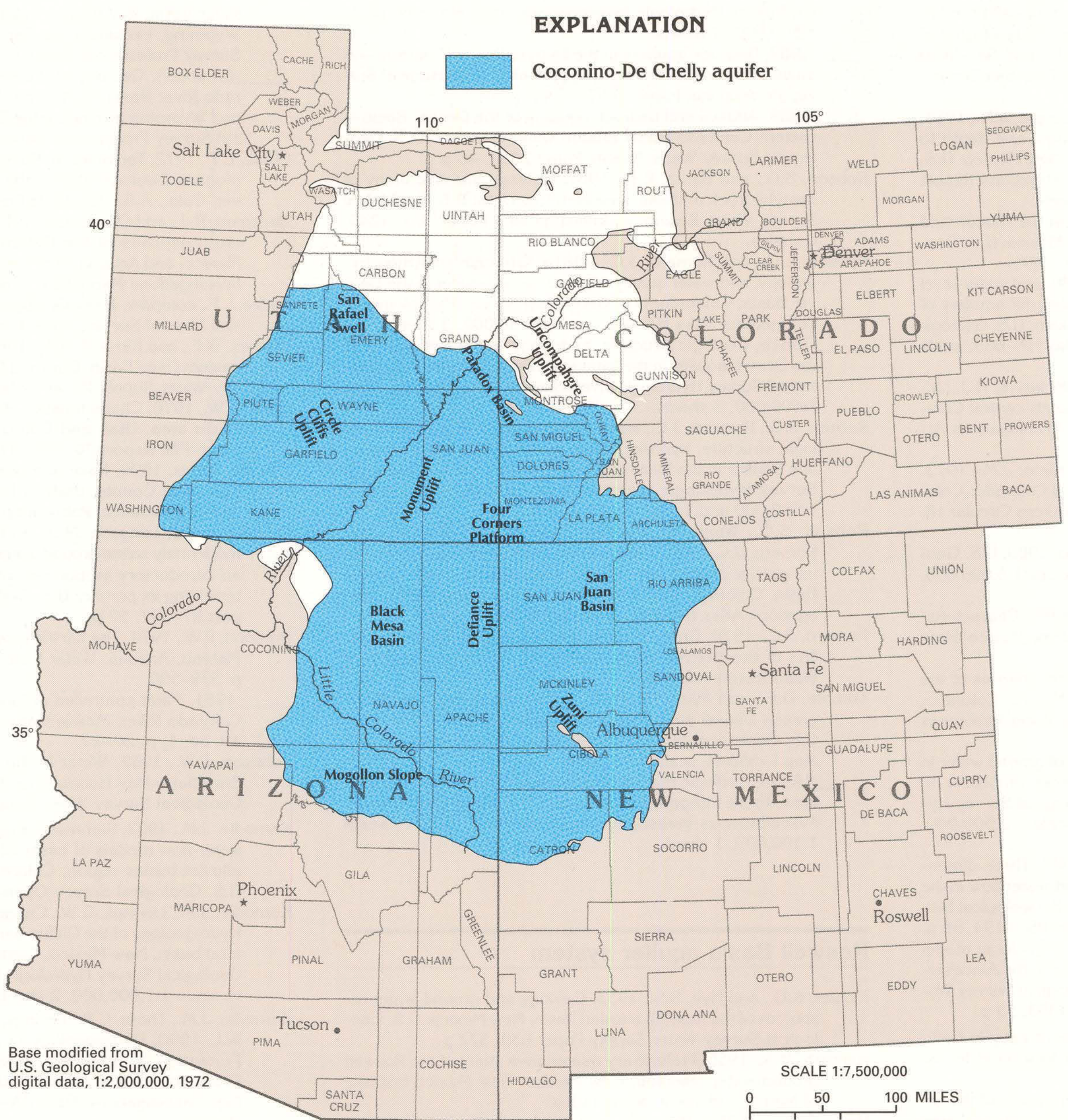


Figure 129. *The Coconino–De Chelly aquifer underlies the southern part of the Colorado Plateaus.*

Hydrogeologic Units

The formations that comprise the Coconino–De Chelly aquifer are the Coconino, De Chelly, and Glorieta Sandstones; the San Andres Limestone; and the Yeso and Cutler Formations (**fig. 130**). The Coconino and De Chelly Sandstones generally consist of well-sorted quartz sandstone with thin interbeds of siltstone, mudstone, and carbonates. The Glorieta Sandstone consists of well-sorted, well-cemented, fine to medium quartz sandstone. The San Andres Limestone consists of dolostone, limestone, and fine-grained clastic rocks. The carbonate rocks in the San Andres Limestone are characterized by solution openings, which substantially increase the hydraulic conductivity of the formation. The Yeso Formation consists of interbedded sandstone, siltstone, limestone, anhydrite, and gypsum and forms a low-permeability zone in the aquifer. The Cutler Formation consists of shale, siltstone, sandstone, arkose, and arkosic conglomerate.

In most areas near the Grand Canyon, the Coconino Sandstone probably does not yield water because of the proximity to the canyon, where the formation has been truncated and drained (**fig. 131**). Fractures and associated solution openings in underlying rocks in the vicinity of the Grand Canyon allow water to discharge from the Coconino Sandstone. In much of the northern part of the Colorado Plateaus, rocks equivalent to those included in the aquifer are present, but the water in these rocks generally has dissolved-solids concentrations in excess of 10,000 milligrams per liter. The hydrogeology of the aquifer in this area is not described in this chapter because of the salinity of the water.

Era	System	Series	Western Black Mesa Basin	Eastern Black Mesa Basin and Defiance Uplift	San Juan Basin	Four Corners platform	San Rafael Swell	Hydrogeologic unit
Paleozoic	Permian	Upper			San Andres Limestone			Chinle-Moenkopi confining unit (lower part)
		Lower	Kaibab Formation				Kaibab Formation	
			Toroweap Formation		Glorieta Sandstone	Cutler Formation	Cutler Formation; Coconino Sandstone	Coconino-De Chelly aquifer
			Coconino Sandstone	De Chelly Sandstone	Yeso Formation			
			Hermit Shale		De Chelly Sandstone			
				Supai Group	Abo Formation	Rico Formation		
	Pennsylvanian	Upper	Supai Group		Hermosa Formation; Madera Formation	Hermosa Formation	Hermosa Formation	Not a principal aquifer
		Middle		Hermosa Formation				
		Lower		Molas Formation	Molas Formation	Molas Formation		
	Mississippian	Upper	Redwall Limestone	Redwall Limestone	Leadville Limestone	Leadville Limestone	Redwall Limestone	
		Lower						
	Devonian	Upper	Temple Butte Limestone	Ouray Limestone	Ouray Limestone	Ouray Limestone	Ouray Limestone	
				Elbert Formation	Elbert Formation	Elbert Formation	Elbert Formation	

Figure 130. *Rocks that comprise the Coconino–De Chelly aquifer are known by different formation names in various parts of the Colorado Plateaus. The light gray areas represent missing rocks.*

Modified from Repenning, 1959; Molenaar, 1977; Lindner-Lunsford and others, 1989; and Geldon, in press

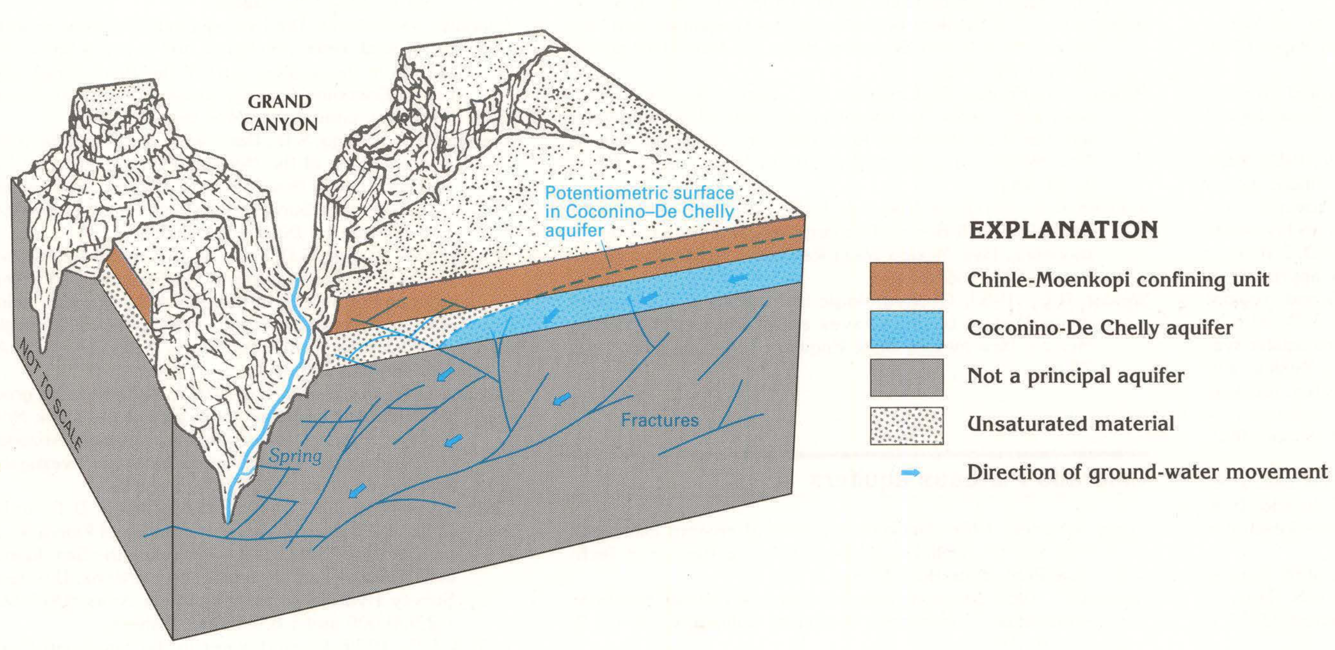


Figure 131. *Fractures in the vicinity of the Grand Canyon act as conduits that allow ground water to drain from the Coconino–De Chelly aquifer. The water emerges from underlying rocks at springs in the Grand Canyon and canyons of tributaries of the Colorado River.*

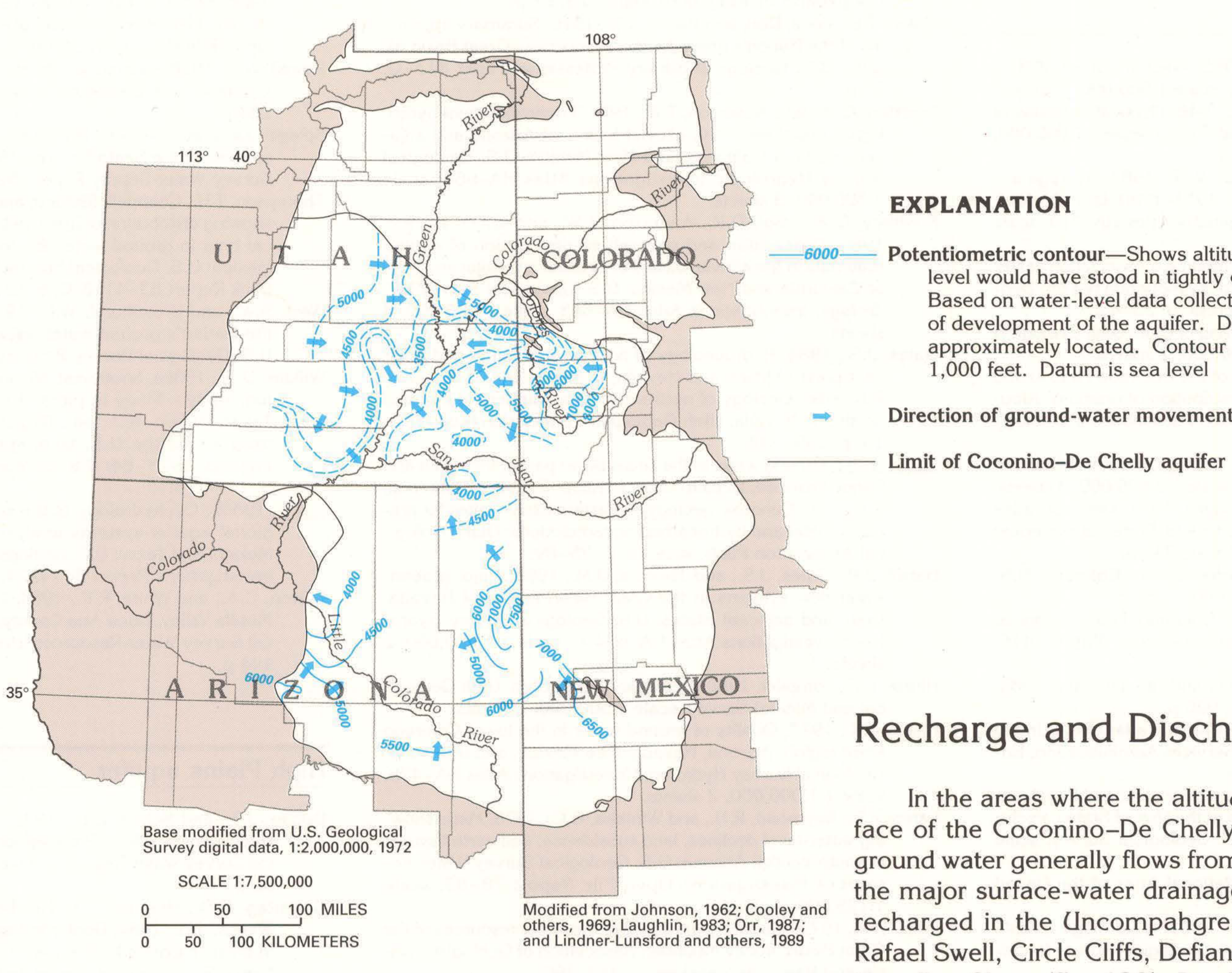
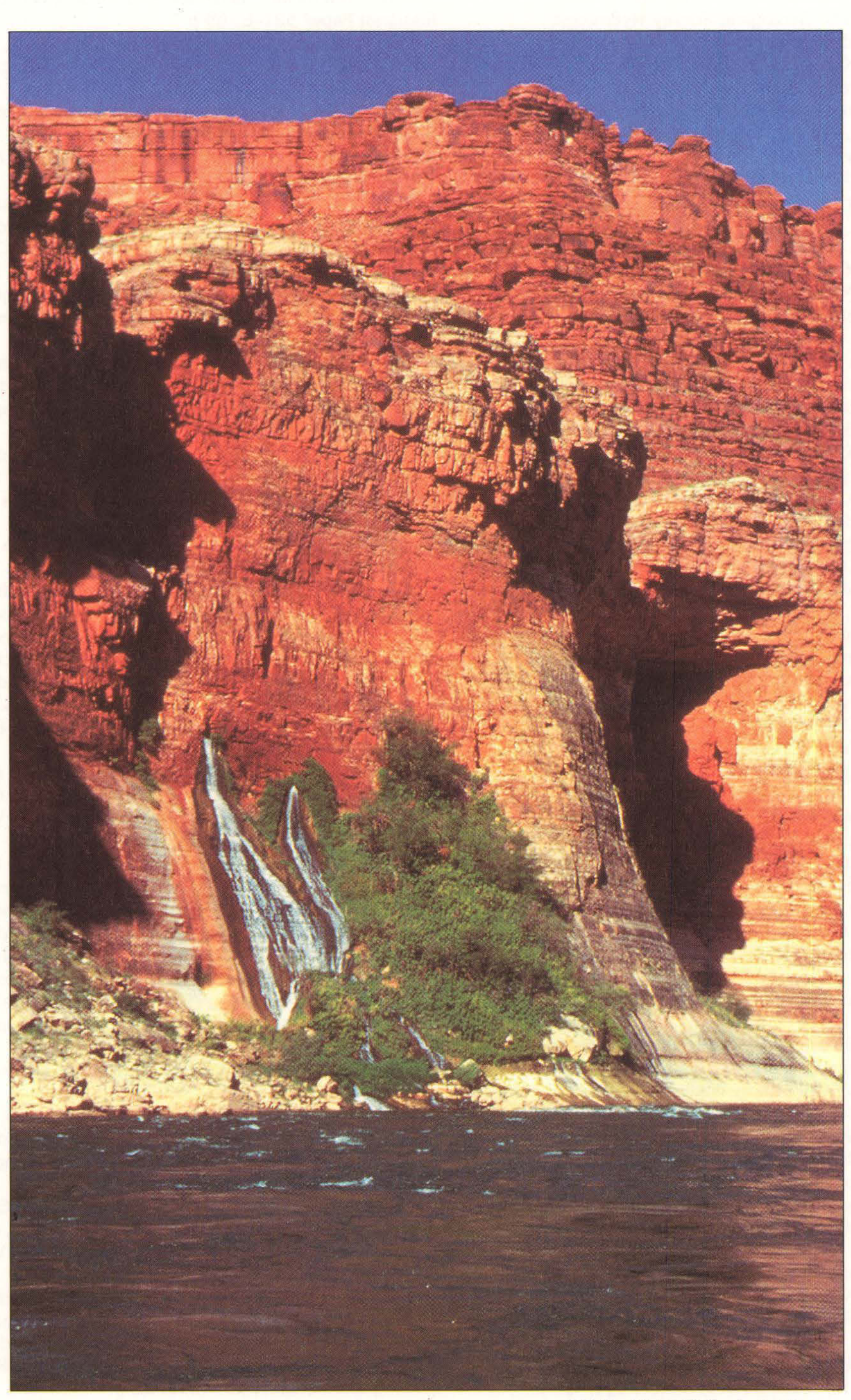


Figure 132. *Ground water in the Coconino–De Chelly aquifer generally flows toward discharge areas near the Colorado River and its tributaries.*

Recharge and Discharge

In the areas where the altitude of the potentiometric surface of the Coconino–De Chelly aquifer has been mapped, ground water generally flows from the structural uplifts toward the major surface-water drainages (**fig. 132**). The aquifer is recharged in the Uncompahgre Uplift, Paradox Basin, San Rafael Swell, Circle Cliffs, Defiance, and Zuni Uplifts, and the Mogollon Slope (**fig. 129**). Discharge mainly is to the Colorado and Green Rivers. Water in the Coconino–De Chelly aquifer near the Black Mesa Basin generally flows northwestward toward a discharge area near the mouth of the Little Colorado River. In the Grand Canyon, a series of springs issuing from the Mississippian Redwall Limestone (**fig. 133**) discharges water derived in part from the Coconino–De Chelly aquifer. Fractures and solution channels in the Redwall Limestone and the rocks separating the Redwall Limestone from the Coconino Sandstone provide conduits for the ground water. Similar processes affect the ground-water flow system elsewhere in the vicinity of the Grand Canyon.



R.D. MacNish, U.S. Geological Survey

Figure 133. *Ground water discharges from springs in the Redwall Limestone and cascades into the Colorado River at Vasey's Paradise in the Grand Canyon.*

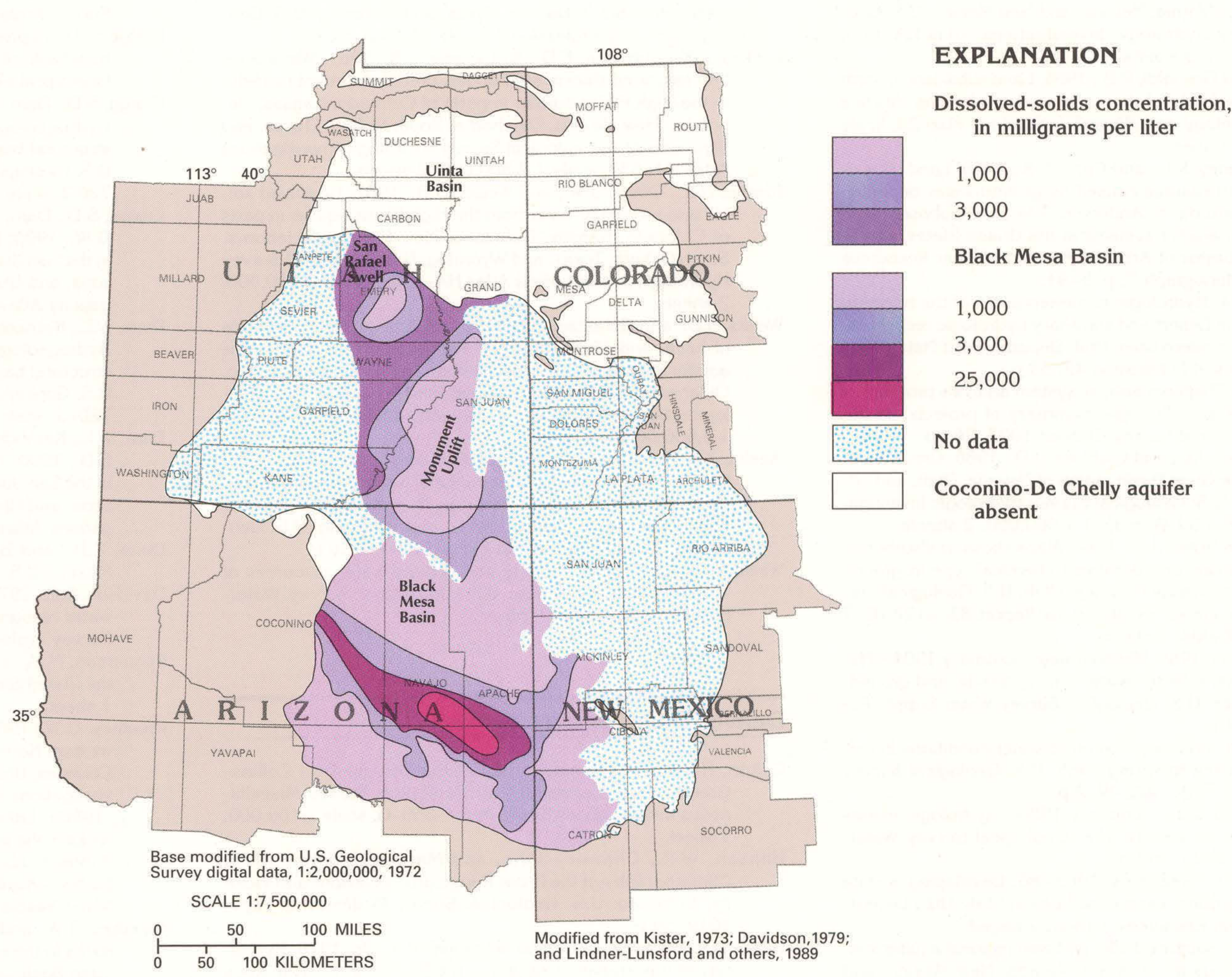


Figure 134. *The dissolved-solids concentration of water in the Coconino–De Chelly aquifer is less than 1,000 milligrams per liter in large areas of northeastern Arizona and southeastern Utah. Little is known of the dissolved-solids concentrations in some parts of the aquifer.*

Ground-Water Quality

In Utah, the dissolved-solids concentration in water from the Coconino–De Chelly aquifer ranges from less than 1,000 milligrams per liter in the San Rafael Swell and Monument Uplift to 10,000 milligrams per liter along the margin of the Uinta Basin (**fig. 134**). In northeastern Arizona and west-central New Mexico, the dissolved-solids concentration of water in the aquifer generally is less than 1,000 milligrams per liter. However, in an area near the southeastern margin of the Black Mesa Basin, the dissolved-solids concentration exceeds 25,000 milligrams per liter. The northwestward regional movement of ground water near the Black Mesa Basin may have produced the elongated distribution of the more mineralized water in that area.

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