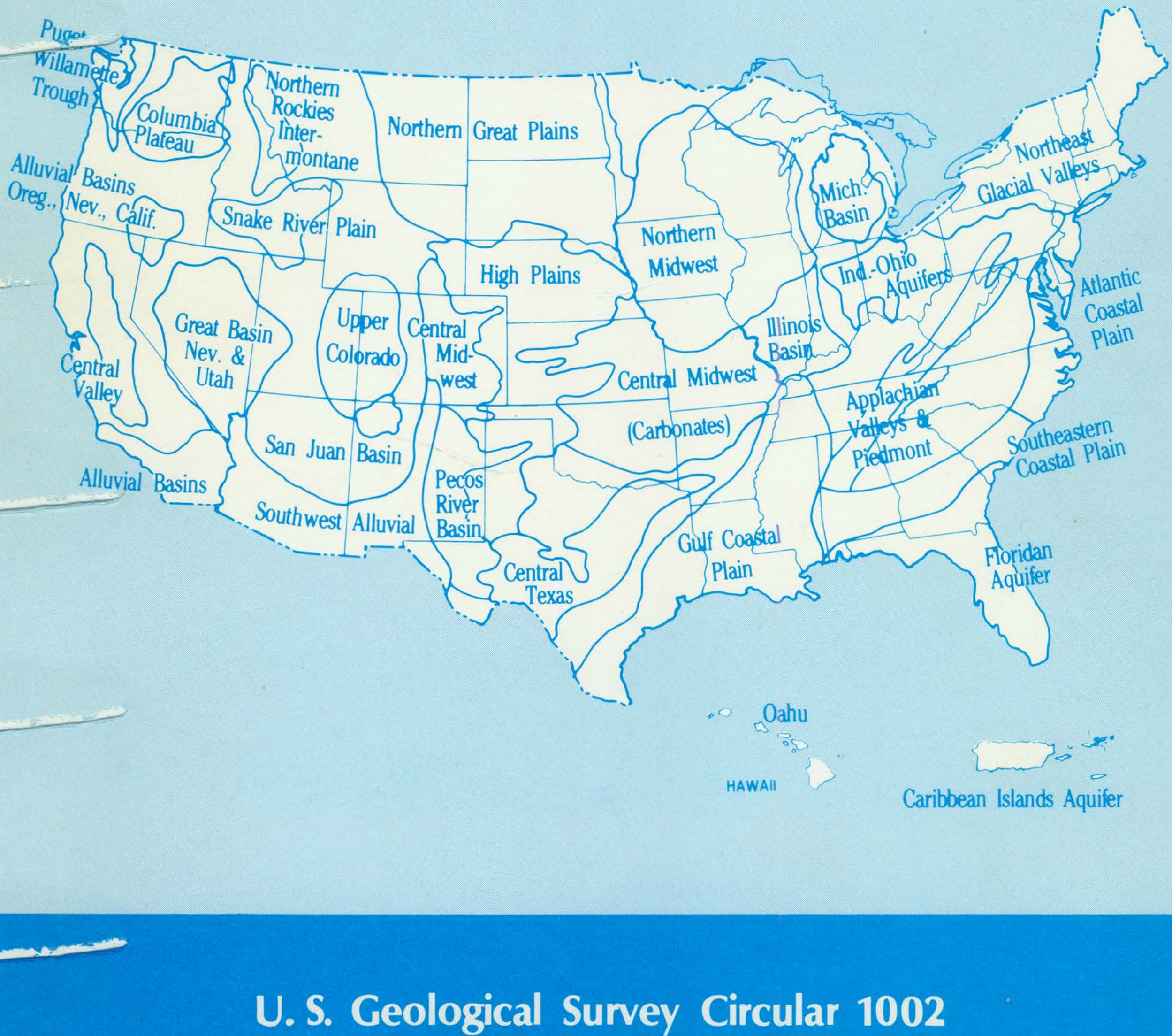
REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

OF THE

U. S. GEOLOGICAL SURVEY

SUMMARY OF PROJECTS, 1978-84



U. S. Geological Survey Circular 1002

REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM
OF THE
U.S. GEOLOGICAL SURVEY
SUMMARY OF PROJECTS, 1978-84

Edited by
Ren Jen Sun

U.S. Geological Survey Circular 1002

UNITED STATES DEPARTMENT OF THE INTERIOR
DONALD PAUL HODEL, *Secretary*

U.S. GEOLOGICAL SURVEY
Dallas L. Peck, *Director*

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ABBREVIATIONS AND CONVERSION FACTORS

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

Multiply	By	To obtain
	Length	
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
acre	4,047	square meter (m^2)
acre	0.4047	hectare
square foot (ft^2)	0.09294	square meter (m^2)
square mile (mi^2)	2.59	square kilometer (km^2)
	Volume	
acre foot (acre-ft)	1,233	cubic meter (m^3)
cubic foot (ft^3)	0.02832	cubic meter (m^3)
gallon (gal)	3.785	liter (L)
gallon (gal)	0.003785	cubic meter (m^3)
cubic mile (mi^3)	4.168	cubic kilometer (km^3)
	Flow	
foot per day (ft/d)	0.3048	meter per day (m/d)
inch per year (in./yr)	25.4	millimeter per year (mm/yr)
square foot per day (ft^2/d)	0.0929	square meter per day (m^2/d)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
gallon per minute (gal/min)	0.06308	liter per second (L/s)
gallon per day (gal/d)	0.003785	cubic meter per day (m^3/d)
billion gallons per day (Bgal/d)	43.81	cubic meter per second (m^3/s)
million gallons per day(Mgal/d)	0.04381	cubic meter per second (m^3/s)

U.S. Geological Survey Data Base

Two U.S. Geological Survey's data bases frequently mentioned in the report are:

GWSI: Ground-Water Sites Inventory
WATSTORE: Water Data Storage and Retrieval System

REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM OF THE U.S. GEOLOGICAL SURVEY—SUMMARY OF PROJECTS, 1978-84

Edited By
Ren Jen Sun, Reston, Virginia

ABSTRACT

The Regional Aquifer-System Analysis Program of the U.S. Geological Survey was initiated in 1978 as a result of specifications of the appropriations bill of the 95th Congress, prompted by the 1977 drought. The purpose of this program is to define the regional hydrology and geology and to establish a framework of background information of geology, hydrology, and geochemistry of the Nation's important aquifer systems. This information is critically needed to develop an understanding of ground-water flow systems, and to support better ground-water resources management.

As of 1984, investigations of seven regional aquifer systems were completed, nine regional aquifer systems were still being studied, and three new studies were started. This report summarizes the status of each investigation of the regional aquifer systems under the program from 1978 through 1984. The nature of the summaries differs somewhat from study to study. For those studies which either have been completed or are near completion, summaries of results are presented. For projects that are not near completion or have just been started, discussions may be brief and focus on problem issues or hydrogeologic conditions.

All reports resulting from the study as of 1984 are listed at the end of each summary. A list of project chiefs and their offices is also included in the report for those who are interested in obtaining additional information.

INTRODUCTION

Water is essential for plant, animal, and human life. The source of the world's freshwater supply is precipitation, which is unevenly distributed in both time and space. Part of the precipitation falling on the Earth's surface is evaporated or transpired by plants to the atmosphere, while part flows directly into rivers, streams, and lakes, and eventually discharges to the sea or returns to the atmosphere by evaporation. The remainder of the precipitation infiltrates past the root zone and gradually moves downward to the zone of saturation. Water in the zone of saturation is termed "ground water."

In much of the western United States, where precipitation ranges from sparse to moderate, ground water forms a vital part of the water supply. Even in the humid East, where precipitation normally is plentiful, ground water is used to meet water needs during droughts, or as a supplement to surface water. In 1980, about 90 billion gallons of ground water were pumped daily in the United States. About 5 percent of this pumpage was used for rural water supply, 13 percent for public supply, 67 percent for irrigation, and 15 percent for industrial use. Ground water contributed about 20 percent of the national water supply—450 billion gallons per day (Bgal/d) in 1980. About 34 percent of the water used for public supply and 79 percent for rural domestic and livestock use in 1980 were supplied by ground water (Solley and others, 1983).

Increases in ground-water use are expected to occur during the next decade. These increases will be associated with irrigation, urban expansion, and energy production, including such activities as power generation and refining of oil shales. In addition, the Nation will look increasingly to ground water as available sites for construction of new surface reservoirs become more limited, and as ground-water becomes more widely recognized as a major component to the Nation's water supply during droughts. These projected increases in the use of ground water will have widespread impacts on the Nation's aquifer systems.

The withdrawal of water through wells initially results in removal of ground water from aquifer storage with an associated lowering of water levels. As water levels decline, new hydraulic gradients are established in the ground-water system which may either reduce natural ground-water discharge to streams, or induce ground-water recharge. When pumpage is greater than the amount of water that can be obtained by increases in recharge or reductions in natural discharge, sustained withdrawal from aquifer storage and continued lowering of water levels occur. In turn, this increases pumping lifts and, in water-table aquifers, causes progressive reduction of the saturated thickness of the aquifer. The result is an increase in pumping costs, a decrease in well yields, and eventually a depletion of the ground-water resource. In areas where aquifers are covered with thick, compressible, fine-grained sediments of low permeability, such as clay and silt, land subsidence may occur due to the sustained withdrawal of water from aquifer storage thus causing compaction of the overlying thick fine-grained sediments by the overburden pressure.

In irrigated areas, where return flow from irrigation water recharges the aquifer, constituents such as salts leached from soil or applied pesticides, herbicides, and fertilizers may move into the aquifer system, thus degrading the ground water quality. Water-quality problems may also develop in industrial or urban areas, especially as a result of waste disposal. Irrigation return flow resulted from surface-water irrigation may cause ground-water levels to rise, creating water-logging problems.

For several decades, ground-water investigations both within and outside the Geological Survey were conducted largely on a local scale within county or State boundaries. In evaluating the effects of development on ground-water resources on the basis of such local investigations, questions have often arisen regarding the extent of the development effects beyond a study area, and the correct way to represent conditions along the boundaries of a study area. Moreover, hydrologists beginning a local ground-water investigation often find little or no background information on the regional geology, hydrology, or water chemistry, and thus have to begin their studies from a base of very little knowledge. These facts underscore the need for broad regional evaluations of the Nation's ground-water resources.

In 1973, the U.S. National Water Commission recognized that the depletion of ground-water storage had reached serious proportions in certain places, and that critically needed information on aquifer systems was not readily available to water-management officials. The Commission recommended to the President and to the Congress of the United States that the U.S. Geological Survey conduct continuing intensive investigations of significant aquifer systems giving priority to those with falling water levels and deteriorating water quality (U.S. National Water Commission, 1973, p. 245). In a report dated June 21, 1977, to the Congress of the United States, the Comptroller General of the United States stated that, "Although much ground-water data have been collected by Federal and State agencies and others, we were told during our study that substantially more geological and hydrological data—primarily of a more specific and detailed nature—will be needed to provide for the orderly development, proper use, and conservation of ground-water resources. According to Federal and State officials, the lack of such specific and detailed data is a major constraint to improving ground-water management" (U.S. Comptroller General, 1977, p. 30).

In 1977 the Committee on Appropriations of the U.S. House of Representatives, in a report of the appropriations for the Department of the Interior and related agencies for fiscal year 1978, stated that "The Committee is providing funds to initiate a program to identify the water resources of the major aquifer systems within the United States. This program will establish the aquifer boundaries, the quantity and the quality of the water within the aquifer, and the recharge characteristics of the aquifer. Although this initiative comes too late to help the present drought (1977), the action taken will develop an inventory of the major ground-water systems of the United States so that in the event of another drought we may be able to draw upon the ground-water reservoir system without doing irreparable damage to the system. The Committee expects the Survey to press this program vigorously" (U.S. House of Representatives, 1977, p. 36-37).

As a result of the specifications of the appropriations bill, and in response to Federal and State needs for information to support better ground-water management, the Geological Survey instituted the Regional Aquifer-System Analysis (RASA) Program in 1978. On the basis of a series of reports presenting the summary appraisals of the Nation's ground-water resources by the Geological Survey (fig. 1), and economic and hydrologic considerations, 28 regional aquifer systems were identified for study under the RASA Program (fig. 2).

This report reviews the reasons for initiating the Regional Aquifer-System Analysis Program and discusses the purpose, approach, and status of the program. Findings and accomplishments of each project are discussed. Reports produced by each project are also listed.

As of 1984, 19 RASA projects had been started. The geographic locations and status of these studies are shown in figure 3. Studies of the following seven regional aquifer systems have been completed as of 1984: (1) Central Valley, California; (2) Floridan; (3) High Plains; (4) Northern Great Plains; (5) Northern Midwest; (6) Snake River Plain; and (7) Southwest Alluvial basins. In some projects, where major technical problems or data deficiencies were identified which could not be addressed adequately within the context of the initial investigation (phase I study), followup studies (termed RASA phase II study) have been undertaken. For example, during the study in the Central Valley, Calif., it was found that ground-water quality is being degraded by irrigation return flow recharging the aquifer; similarly, during the High Plains study, it was found that more accurate information on pumpage and on recharge from irrigation return flow was necessary for a full understanding of the ground-water system and its use for irrigation. RASA phase II studies were initiated for both the Central Valley and the High Plains to address these issues. Except for the Northern Great Plains regional aquifer system study, the other six completed studies have followup projects. In addition to the followup studies, there are eight on-going studies within the continental United

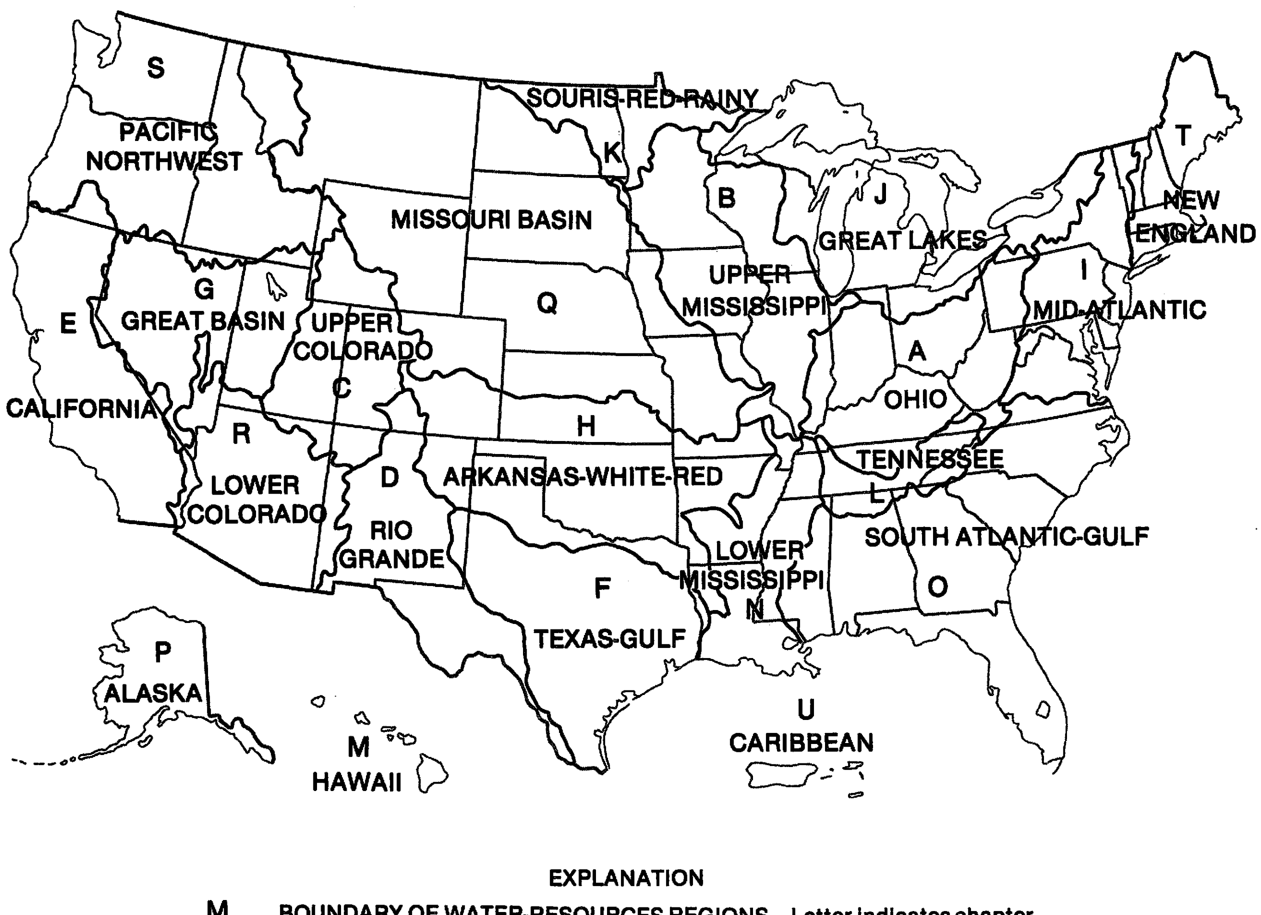
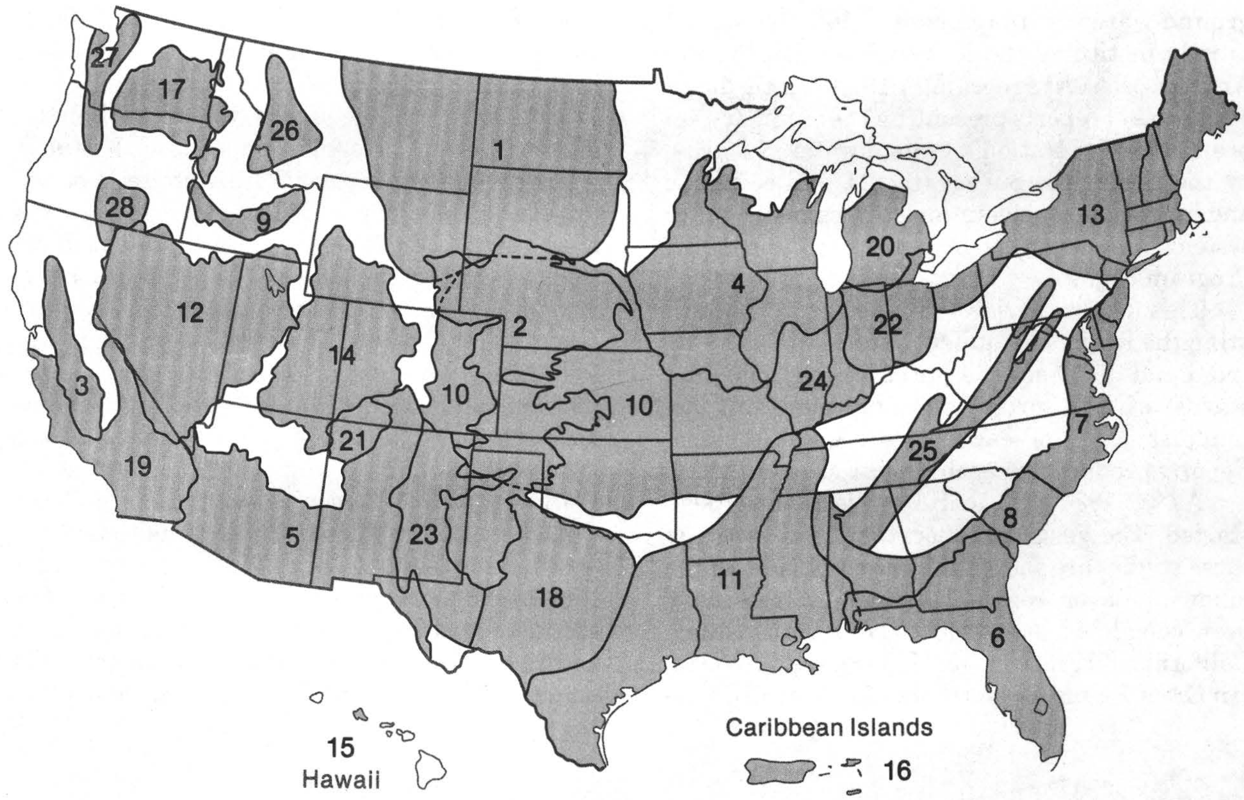


Figure 1.—Location of reports of "Summary Appraisals of the Nation's Ground-Water Resources" by the U.S. Geological Survey.

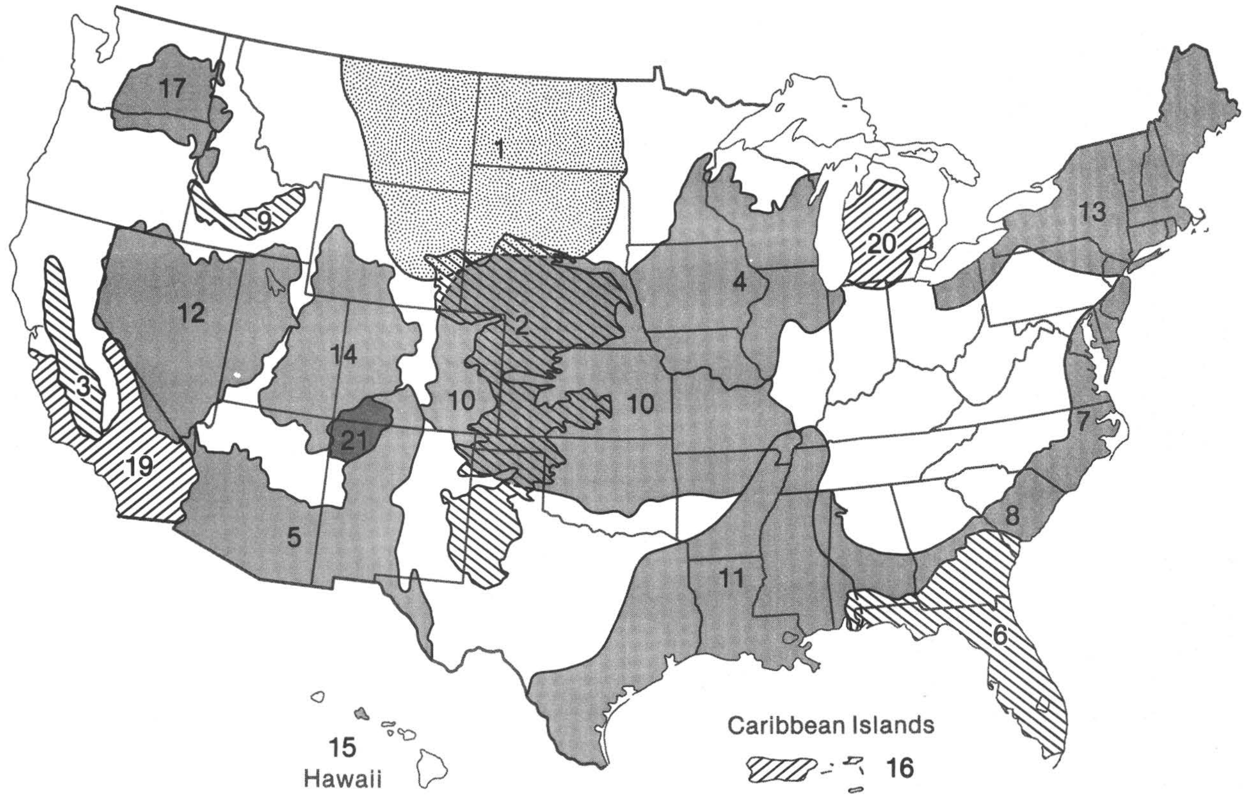


EXPLANATION

10 REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM -- Numbering system for identification purposes only, not intended to imply priority

1, Northern Great Plains
2, High Plains
3, Central Valley, California
4, Northern Midwest
5, Southwest alluvial basins
6, Floridan aquifer
7, Northern Atlantic Coastal Plain
8, Southeastern Coastal Plain
9, Snake River Plain
10, Central Midwest
11, Gulf Coastal Plain
12, Great Basin
13, Northeast glacial valleys
14, Upper Colorado River Basin
15, Oahu Island, Hawaii
16, Caribbean Islands
17, Columbia Plateau
18, South-Central Texas
19, Southern California alluvial basins
20, Michigan Basin
21, San Juan Basin
22, Ohio-Indiana glacial and carbonates
23, Pecos River Basin
24, Illinois Basin
25, Appalachian Valleys and Piedmont
26, Northern Rockies Intermontane basins
27, Puget-Willamette Trough
28, Alluvial basins, Oregon, California, and Nevada

Figure 2.—Location of regional aquifer systems identified for study under the Regional Aquifer-System Analysis Program of the U.S. Geological Survey.



EXPLANATION

STATUS OF THE REGIONAL AQUIFER-SYSTEMS ANALYSIS PROGRAM in 1984
[Actual or planned duration shown by span of years]

STUDIES COMPLETED

PHASE I STUDIES COMPLETED
PHASE II STUDIES UNDERWAY, FY 1984

STUDIES UNDERWAY, FY 1984

STUDIES INITIATED, FY 1984

STUDIES PLANNED, FY 1985

1, Northern Great Plains; FY 1978-82
2, High Plains; FY 1978-82; Phase II study
3, Central Valley, California; FY 1978-82; Phase II study
4, Northern Midwest; FY 1979-84
5, Southwest alluvial basins; FY 1979-84
6, Floridan aquifer; FY 1979-82; Phase II study
7, Northern Atlantic Coastal Plain; FY 1980-85
8, Southeastern Coastal Plain; FY 1980-86
9, Snake River Plain; FY 1980-84; Phase II study
10, Central Midwest; FY 1981-86
11, Gulf Coastal Plain; FY 1981-88
12, Great Basin; FY 1981-85
13, Northeast glacial valleys; FY 1982-86
14, Upper Colorado River Basin; FY 1982-86
15, Oahu Island, Hawaii; FY 1982-86
16, Caribbean Islands; FY 1984-87
17, Columbia Plateau; FY 1983-86
19, Southern California alluvial basins; FY 1984-87
20, Michigan Basin; FY 1984-87
21, San Juan Basin; to be iniated in FY 1985

Figure 3.—Location of areas being investigated during 1978-84 under the Regional Aquifer-System Analysis Program of the U.S. Geological Survey.

States and one in Oahu Island, Hawaii. Three new regional aquifer system studies were initiated during fiscal year 1984, as follows: (1) Caribbean, (2) southern California alluvial-basins, and (3) Michigan Basin.

A regional aquifer system as defined in the RASA Program may be of two general types: (1) it may be comprised of an extensive set of aquifers and confining units which locally may be discontinuous, but which act hydrologically as a single aquifer system on a regional scale—examples include, aquifer systems underlying the Great Plains, High Plains, Gulf Coastal Plain, and Atlantic Coastal Plain; or (2) it may represent a set of virtually independent aquifers which share so many common characteristics that investigation of a few of these aquifers can establish common principles and hydrogeologic factors controlling the occurrence, movement, and quality of ground water throughout the aquifer systems. Examples of the second type include the alluvial basins of Arizona, and New Mexico, and the glacial aquifers of the northeastern United States (Bennett, 1979).

The objective of each RASA study is to define the regional hydrology and geology and to establish a framework of background information—geologic, hydrologic, and geochemical—that can be used for regional assessment of ground-water resources and in support of more detailed local studies. Each RASA project is designed to fit the particular needs of the study area; however, every project utilizes simulation (computer modeling) to help understand and analyze ground-water flow patterns and to provide the ability to evaluate effects of development on ground-water resources. Available information on quality of water throughout the aquifer system is also assembled and analyzed in all studies. Although the studies rely primarily on existing data, some new data are collected in each investigation, and some exploratory drilling has been carried out.

The delineation and continuity of aquifers and confining units are dependent on the scale of the problem under study. For example, the schematic section in figure 4 shows three major aquifers, A1, A2, and A3, separated by two confining units, C1 and C2, viewed on a regional scale. On a more local scale, however, aquifers A2 and A3 might be subdivided further, treating the discontinuous lenses of low permeability material within those aquifers as confining units. Similarly, the permeable sand lenses within confining units C1 and C2 could be viewed as aquifers if their size were significant on a local scale. Regional studies normally combined local aquifers into a relatively small number of regional aquifers. However, many of the studies under this program also have considered the ground-water system at subregional and even local scales, through the establishment of subprojects, to address specific areas of interest. Normally, these subprojects utilize greater resolution in the delineation of aquifers and confining units than do the regional studies.

Delineation and correlaion of aquifers and confining units in the regional studies is based on the hydraulic characteristics of the rocks rather than on their geologic age. Thus, an aquifer delineated in a RASA study may contain several different geologic formations or time-stratigraphic units.

Simulation, using computer-based numerical models of ground-water flow, is used in all studies to provide quantitative understanding of the aquifer systems. However, the regional simulation models developed in the regional studies may not always be appropriate for evaluation of effects of development on ground-water resources at a local level. The regional simulations are intended primarily for regional assessments, to improve understanding of regional flow patterns, and to provide boundary conditions for subregional or local simulation studies. For example, during the northern Atlantic Coastal Plain regional aquifer-system study, four sub-regional flow models were constructed for the Coastal Plain aquifers underlying the States of New Jersey, Delaware and Maryland, Virginia, and North Carolina. A regional-flow model was also constructed representing the entire Coastal Plain aquifer system from Long Island, N.Y., through North Carolina. The regional-flow model supplies the boundary conditions for each of the subregional flow models, and provides an overall view of the flow system and its evaluation; the subregional flow models provide the resolution needed for more detailed analysis of the effects of development on the aquifer system.

The results of RASA investigations are released through publications of the Survey as Hydrologic Investigations Atlases (HA), Open-File Reports (OFR), Water-Resources Investigations Reports (WRIR), Water-Supply Papers (WSP), and Professional Papers (PP), or are published in outside scientific journals. Upon completion of each RASA project, Professional

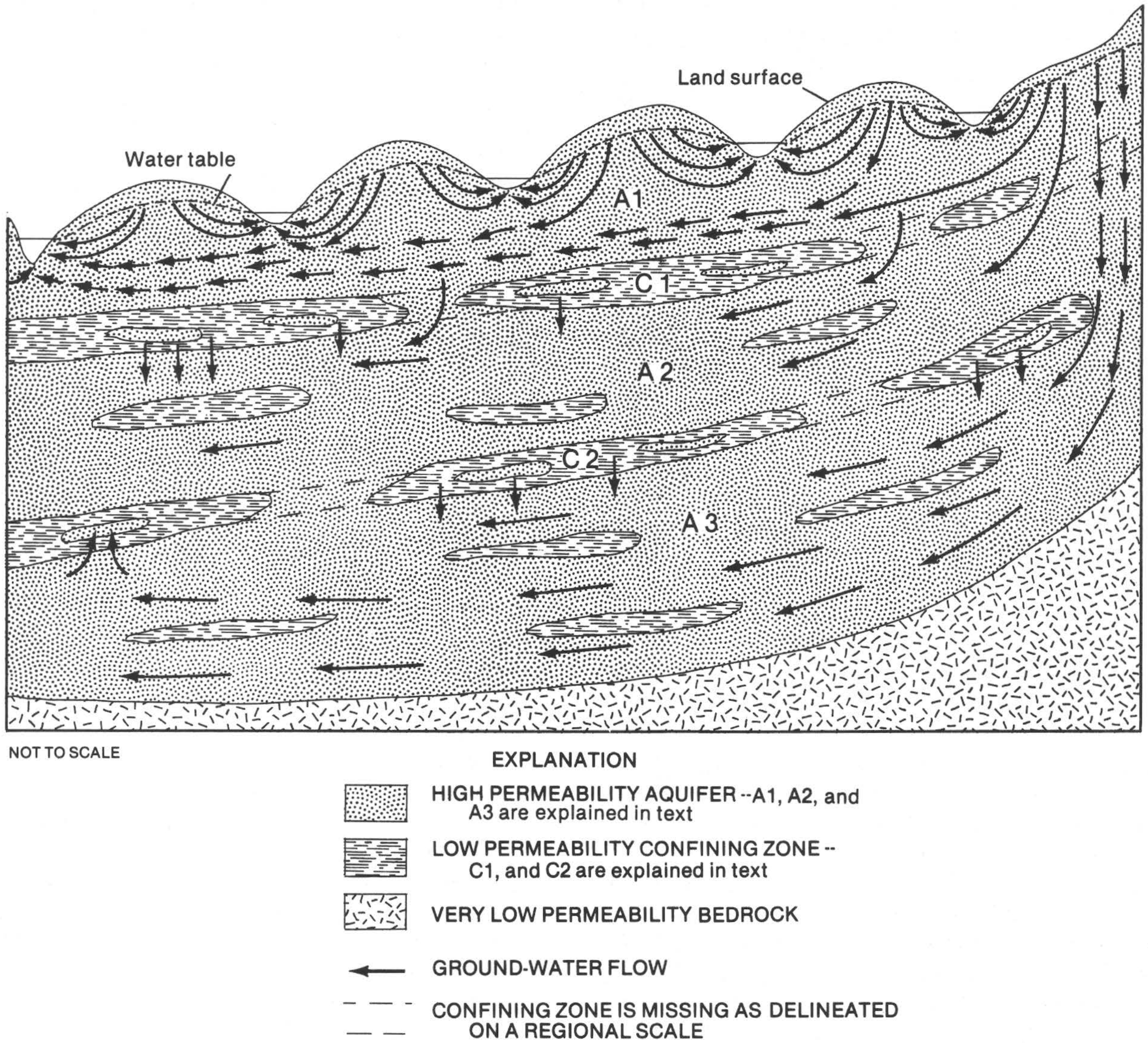


Figure 4.—Schematic section showing aquifers and confining units.

Papers are published that summarize and synthesize the results of the project. The Professional Paper associated with each project is divided into chapters designated by letters, with the letter A reserved for a general chapter summarizing and integrating the major project findings. For example, the Professional Paper describing the Floridan aquifer system is designated Professional Paper 1403. "Summary of the Hydrology of the Floridan Aquifer System in Florida and in parts of Georgia, South Carolina, and Alabama" is Professional Paper 1403-A; "Hydrogeologic Framework of the Floridan Aquifer System in Florida and in parts of Georgia, South Carolina, and Alabama" is Professional Paper 1403-B; and so on. All of the Professional Papers derived from RASA studies will be designated by a Professional Paper number between 1400 and 1428.

During the RASA studies, technical problems for which there are no readily available solutions are often encountered; for example, analyzing hydrologic data from wells that are open to many aquifers, or simulating the movement of ground water of different densities in an extensive three-dimensional system. To address such problems, RASA project personnel have devoted considerable effort to develop new techniques. For example, a multiaquifer-well simulation technique was developed during the Northern Midwest RASA study (Bennett and others, 1982). An improved method of simulating variable-density ground-water flow also has been developed by Weiss (1982) and Kuiper (1983). The RASA Program also supports research by specialists within the Geological Survey in areas of geohydrology and geochemistry. Numerous reports have been prepared by researchers under this support, and have contributed not only to the RASA Program, but also to the overall understanding of ground-water hydrology, geology and geochemistry. However, only reports prepared by hydrologists assigned to RASA projects are listed in this report.

Summaries of each RASA project are presented in the following four groups: (1) completed phase I projects, (2) active phase I projects, (3) newly initiated phase I projects, and (4) active phase II projects (followup studies). A list of reports released, as of 1984, by each project is also included at the end of each summary. The nature of the summaries differs somewhat from project to project. For those projects which either have been completed or are near completion, summaries of the results are presented. For projects that are not near completion or have just been started, discussion may be brief.

It should be noted that aquifer names presented in this summary are provisional; these names may change when more suitable names are selected during investigations.

For those who are interested in obtaining additional information, a list of project chiefs and their offices is included in the Appendix.

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COMPLETED PHASE I REGIONAL AQUIFER-SYSTEM ANALYSIS PROJECTS

CENTRAL VALLEY REGIONAL AQUIFER-SYSTEM STUDY, CALIFORNIA

By **Gilbert L. Bertoldi, Sacramento, California, and Ren Jen Sun, Reston, Virginia**

INTRODUCTION

The Central Valley of California occupies about 12 percent of the total land area of the State of California. It is a large alluvium-filled structural basin occupying approximately 20,000 mi^2 of the flatland lying between the Coastal Ranges and Valleys to the west and Sierra Nevada Range to the east (fig. 5). The aquifer system of the Central Valley is composed of a heterogeneous mixture of continental alluvial materials derived from the surrounding mountains. Thickness of the sediments averages about 2,900 feet in the San Joaquin Valley and 1,500 feet in the Sacramento Valley.

The climate of the Central Valley is arid to semiarid, with average annual precipitation ranging from 14 to 20 inches in the Sacramento Valley, and 5 to 14 inches in the San Joaquin Valley. Under predevelopment conditions, there was a water deficiency (potential evapotranspiration exceeding precipitation) of between 0 to 40 in./yr in the Central Valley (fig. 6).

In order to support an agricultural economy, large quantities of water are needed for irrigation. It is estimated that, including delivery losses, an average of 36 inches of water per acre is used annually. During 1961-77, approximately 22 million acre-ft of water was used in an average year. Of this 22 million acre-ft of water, about 50 percent of the water was supplied by surface water and the remaining 50 percent was ground water.

In addition to the agricultural use of ground water, nearly every city in the San Joaquin Valley uses ground water as the principal source for municipal and industrial supplies.

The natural distribution of water in California is very erratic (fig.6). The Central Valley has an average annual water deficiency, under natural conditions, as great as 40 inches; whereas parts of the bordering Sierra Nevada and Coastal Ranges have an average annual surplus of water. However, because of flat, low land with fertile soil, agricultural development and human population tend to be concentrated in the valleys that are deficient in precipitation. Fortunately, ground water is available everywhere in the Central Valley, even where little rain normally falls or little surface water exists.

Since the last half of the 1800's, ground water in the Central Valley has been developed for irrigation. Early records indicate that estimated pumpage was about 250,000 acre-ft in the San Joaquin Valley in 1912 (Harding and Robertson, 1912) and about 112,000 acre-ft in the Sacramento Valley in 1913 (Bryan, 1923). The annual pumpage in the Central Valley has increased from about 362,000 acre-ft in 1912-13 to about 15 million acre-ft in the drought year, 1977; however, the average pumpage in the mid-1970's was about 12 million acre-ft.

Ground water withdrawn from aquifers by pumping may or may not be replenished. In some parts of the Central Valley, the ground water withdrawn can be replenished annually during the nonirrigation season by recharge from precipitation and streams; in other areas, replenishment of aquifers occurs only in years of abundant precipitation. In parts of the Central Valley, pumping has caused continuous water-level declines; in parts of the San Joaquin Valley and Tulare Basin, water levels have declined nearly 400 feet, depleting the stored ground water.

Page (1984) has shown that throughout the Central Valley, about 50 percent of the thickness of continental sediments is composed of fine-grained sediments (clay or silt). As water levels decline in aquifers, water stored in the pores of the layers consisting of fine-grained materials

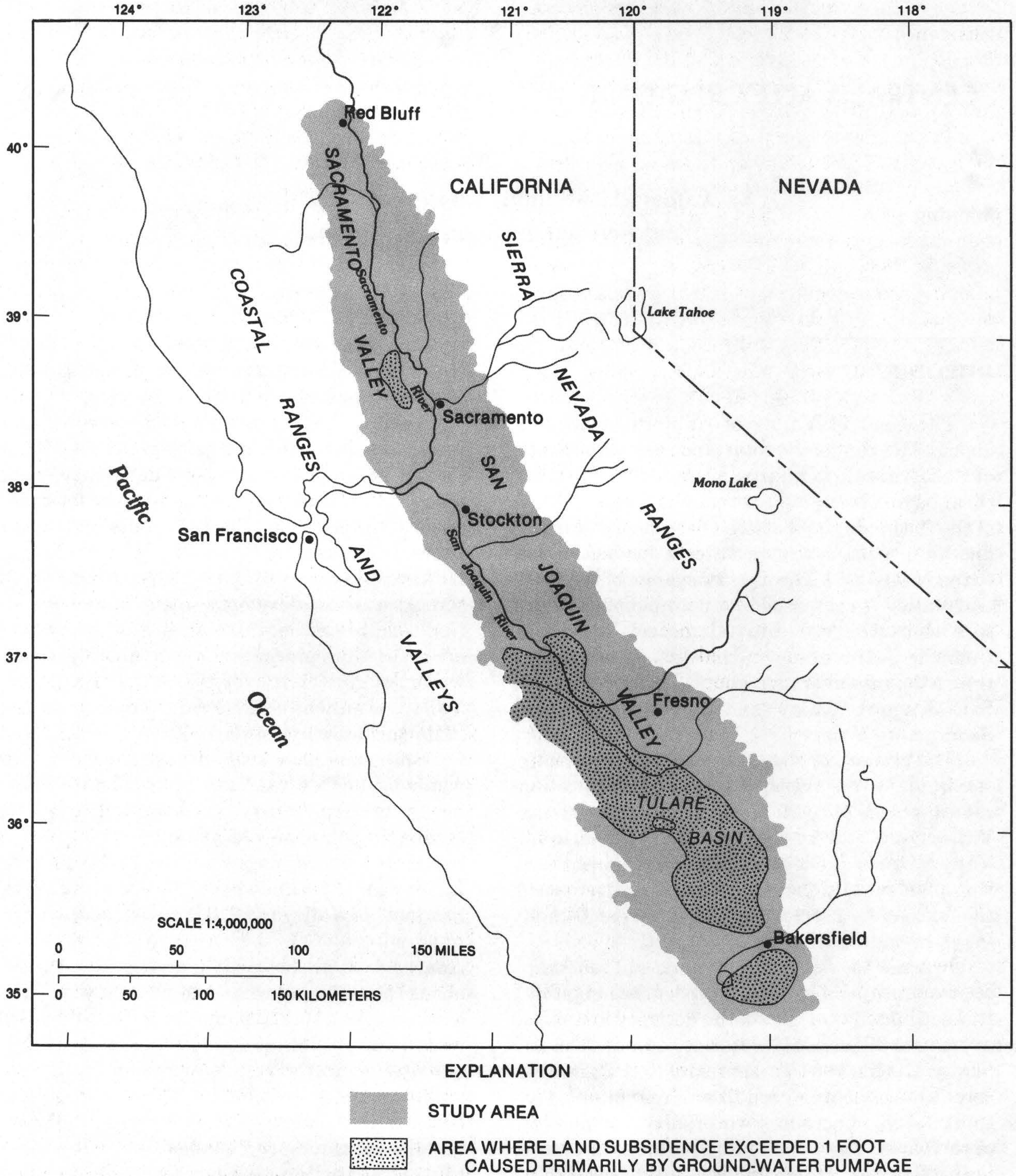


Figure 5.—Location of the Central Valley of California and the principal areas of land subsidence (from Williamson and Prudic, 1984).

starts to drain into the adjacent aquifers where heads have been reduced.

As a result, the land surface may subside due to compaction of the overlying fine-grained sediments. Subsidence in the Central Valley began in the mid-1920's as ground water was developed for irrigation. By 1970, the maximum subsidence exceeded 29 feet at one point in the San Joaquin Valley. Over 5,000 mi^2 of land surface in the Central Valley had subsided more than 1 foot by 1970. Figure 7 illustrates the areas of land subsidence observed during 1961-75 and the corresponding 1961 pumpage per unit area.

The rate of subsidence in the most severe pumping areas in the San Joaquin Valley has been decreasing since delivery of surface water began in 1968 (fig. 8); however, it may recur if pumping reduces water levels below the previous maximum lows. For example during the 1976-77 drought, water levels declined a maximum of nearly 200 feet in 8 months; the subsidence, which had nearly ceased after 1972, was as much as 0.5 foot in 1977.

Volume of water released from compacting sediments is much greater than that available from artesian storage. However, the storage of the compacted sediments cannot be restored to its original volume; therefore, water is available only during compaction as water levels are lowered to a given depth. As compaction ceases, water can be withdrawn from or returned to artesian storage only. Subsidence has also caused damage to structures such as aqueducts, roads, bridges, buildings, and well casings. In low areas, subsidence also increases the potential for flooding and seawater encroachment, which may become a problem in the Sacramento Valley where subsidence is only beginning.

Except at a few localities, quality of ground water is good for nearly all purposes in the Sacramento Valley; dissolved solids range from 200 to 1,200 milligrams per liter (mg/L). In the San Joaquin Valley, the shallow water body in the west is not extensively used because of high dissolved solids which range from 1,000 to 5,000 mg/L. Chloride and sulfate concentrations average about 400 and 440 mg/L, respectively. Prior to the importation of surface water, ground water for irrigation was pumped from a confined aquifer below the water-table aquifer. Originally, water levels in the confined aquifer were above those in the shallow aquifer. However, irrigation pumping caused a reversal of heads, and a downward hydraulic gradient developed from the shallow aquifer (with poor water quality) to the confined aquifer. This downward leakage could cause contamination of the lower confined aquifer by poor-quality water from the shallow aquifer. In addition, pesticide and fertilizer have been applied to the land and minerals in soil such as selenium and boron could be leached by irrigation water. All those human activities can contaminate ground-water in water-table and confined aquifers. Dibromochloropropane (DBCP), a suspected carcinogen and widely used nematocide, was found in ground water in every county in the San Joaquin Valley and Tulare Basin and has been found in public-water supplies of the city of Fresno.

Because of these problems, water managers need accurate and consistent information on the behavior of the ground-water flow system in the Central Valley and on the potential impact of developments on the aquifer system. Previous analyses of ground-water flow were conducted within limited geographic areas or were limited to definition of only a part of the ground-water flow system. Ground-water problems in the Central Valley, however, are not strictly limited to local areas and in many cases affect the entire valley. For example, an increase in pumpage in the city of Stockton may have an effect as far away as Sacramento.

In order to provide such needed information on a regional scale, the Geological Survey started a 4-year study of the regional ground-water flow system in the Central Valley in 1978 (Bertoldi, 1979). The study compiled and analyzed geologic, hydrologic, and geochemical information and established a valley-wide ground-water data base that could be used to construct ground-water flow and land-subsidence simulation models. The study was completed in 1982. Twenty-two reports have been published; several more reports are in review and preparation.

For the first 18 months of the study, data were gathered from 600 published reports and from files of 150 Federal, State, county, and local agencies. These data were coded, stored, and analyzed, using various computer programs for areal distribution, validity, and suitability. Data needed for model simulation generally are available from existing information. However, geologic information, for the part of the aquifer from 100 feet below land surface to the top of Tehama Formation of Pliocene age, about 2,500 feet below land surface, was not available in much of the Sacramento Valley. Most wells in the Sacramento Valley are less than 200 feet deep and therefore provide no information on the

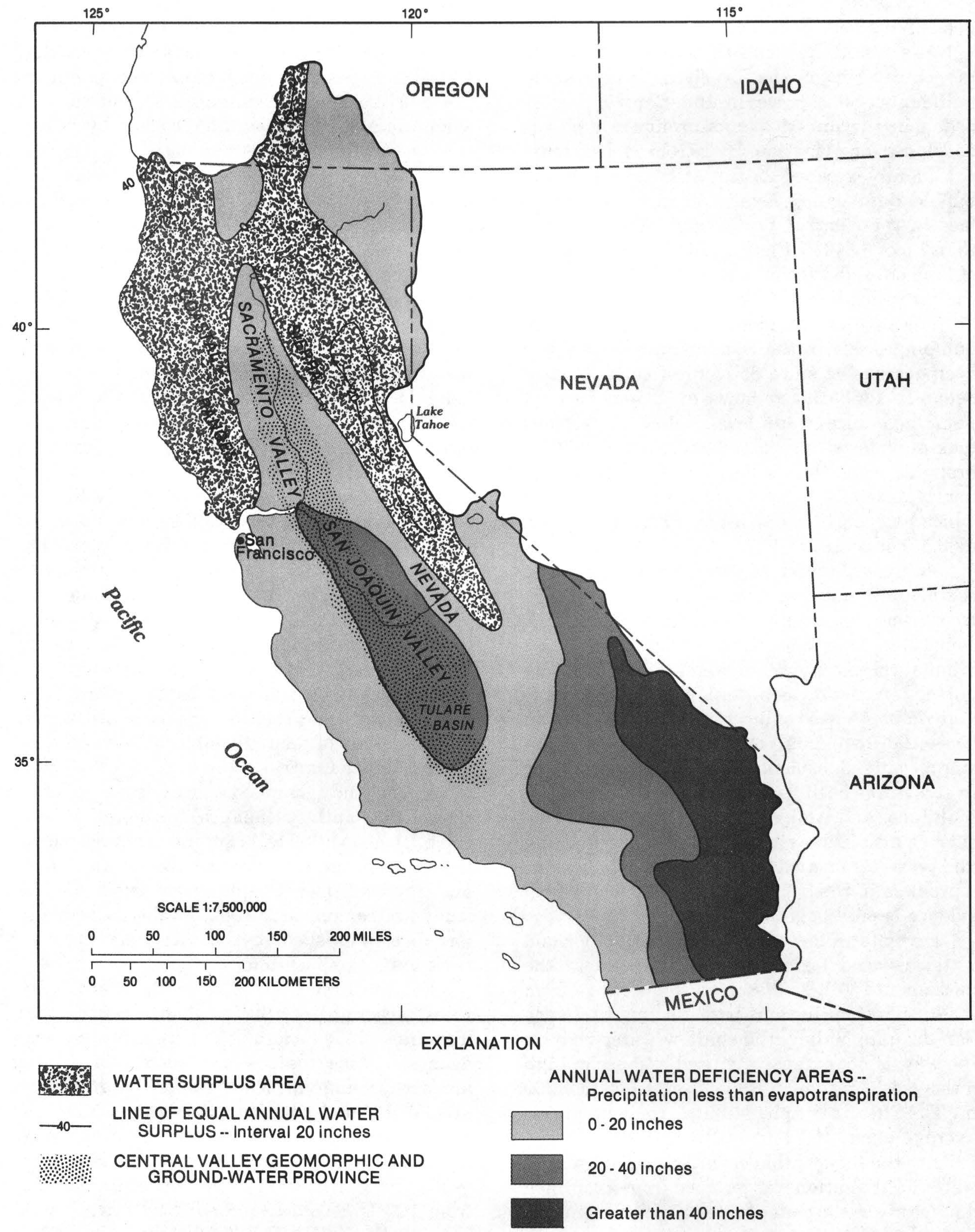
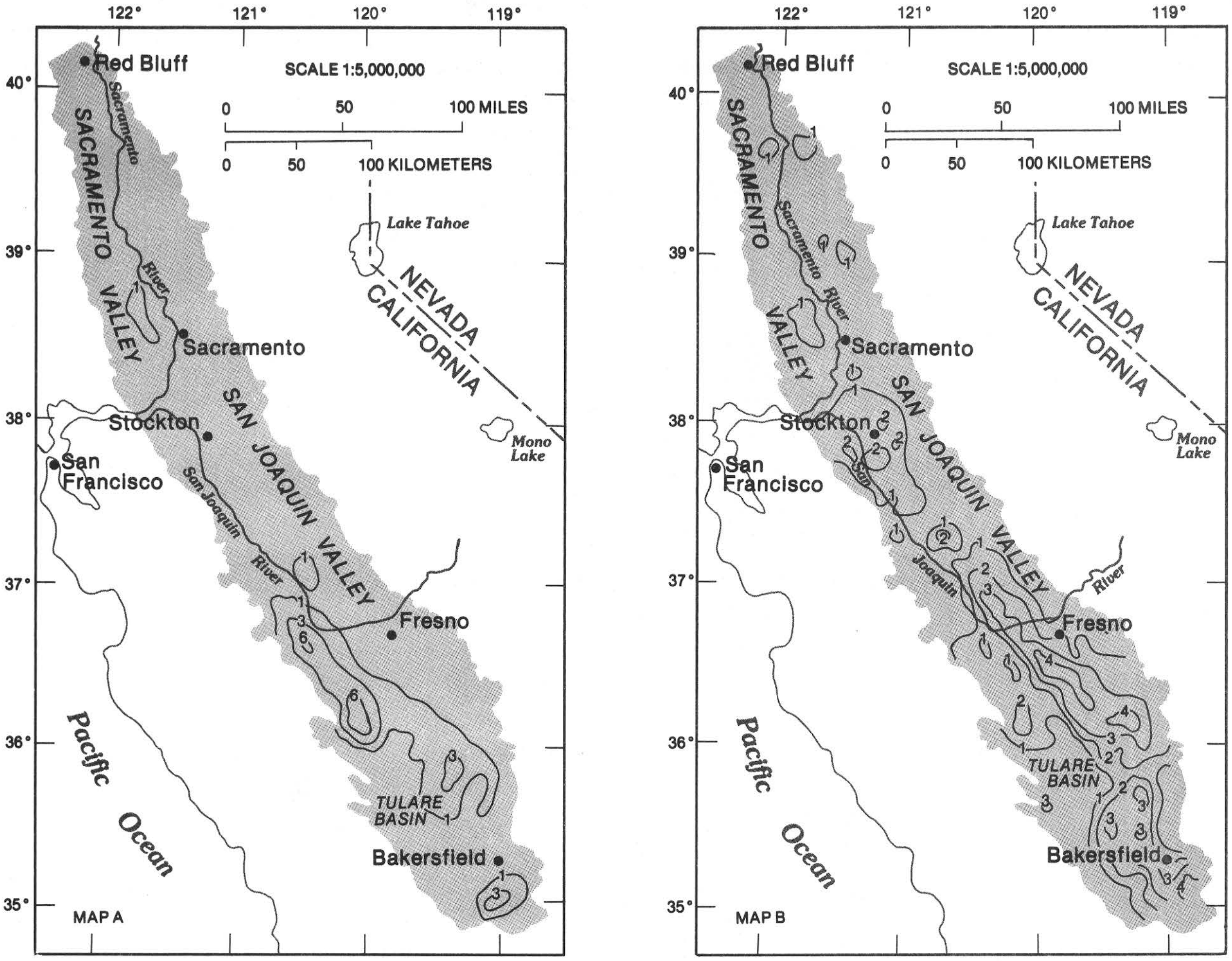


Figure 6.—Water-surplus and water-deficient areas in California (modified from Thomas and Phoenix, 1976).



EXPLANATION

MAP A

—6— LINE OF EQUAL LAND SUBSIDENCE, 1961 - 75 -- Interval in feet, is variable

MAP B

—4— LINE OF EQUAL PUMPAGE PER UNIT AREA, 1961 -- Interval in feet, is variable

STUDY AREA

Figure 7.—Areas of A, land subsidence observed during 1961-75 and, B, the corresponding 1961 pumpage per unit area (modified from Williamson and Prudic, 1984).

aquifer below that depth. To remedy this data gap, seven deep exploratory wells were drilled during the study for the purpose of obtaining: (1) geologic logs, (2) core samples, (3) geophysical logs, (4) sonic logs, (5) water samples, and (6) water-level measurements.

From examination of the available data, it was found that water-quality information was inadequate for the San Joaquin Valley and nonexistent for an area of 400 mi^2 in the Sacramento Valley. A one-time sampling was done to fill the gap in the water-quality data base.

Data for water levels, precipitation, soil types, pumpage, land use, and streamflow were adequate for initial modeling. A finite-difference ground-water flow model incorporating subsidence was developed from the data compiled.

SIGNIFICANT RESULTS

Many significant findings that emerged from the Central Valley regional aquifer study may have application to the general knowledge of ground-water hydrology. For example, knowledge of the extent of changes in aquifer characteristics caused by permanent deformation of sediments during subsidence may lead to new concepts in ground-water hydrology. Only a few of the findings are listed below:

•Discovery of a compressible clay in the Sacramento Valley that is similar to the Corcoran

Clay Member of the Tulare Formation, a major confining unit in the San Joaquin Valley. This clay bed occurs in an area where little ground water is being used. Hence, there is considerable potential for land subsidence should ground water be developed in that area because the clay has a very high compression index and is over 50 feet thick. This discovery is very important to water managers when plans to develop ground-water resources in that area are considered.

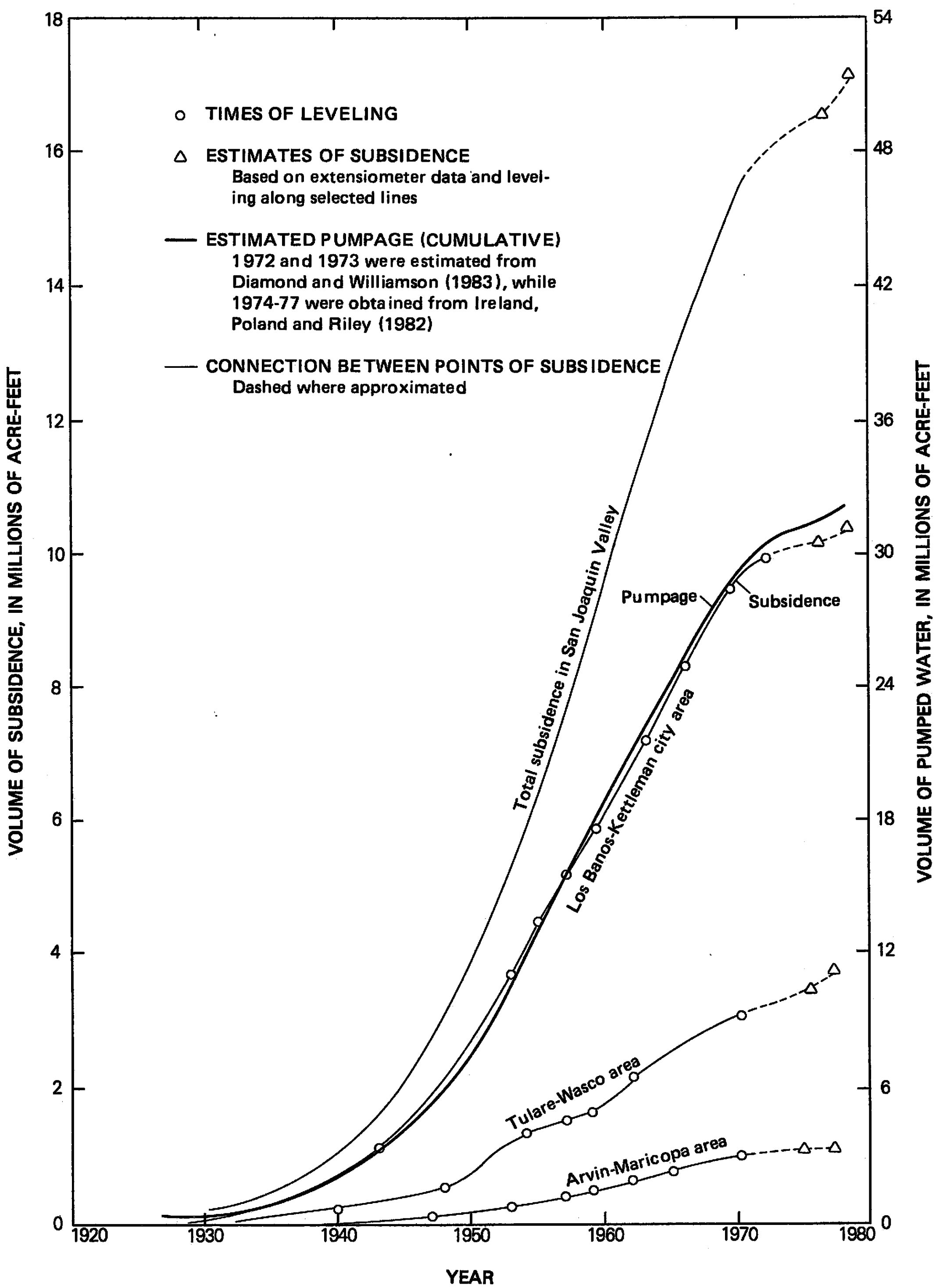


Figure 8.—Volume of land subsidence in major pumping areas in San Joaquin Valley and volume of pumped water in the west part of the San Joaquin Valley (from Williamson and Prudic, 1984).

•Natural geochemical controls and mechanisms were defined for the Sacramento Valley thus establishing baseline information on water quality (Hull, 1984). Future water-quality changes in the Sacramento Valley can be evaluated by comparison to the baseline data.

•Areas where ground water has high concentrations of boron were mapped. This information is important for future planning of boron-sensitive crops.

•Estimates of ground-water storage, hydraulic conductivity, porosity, and potential land subsidence in the Sacramento Valley were made on the basis of information resulting from more than 10,000 wells augmented by the seven deep exploratory test wells. This information was not available before this study.

•Prior to development, the aquifers were recharged by precipitation and stream seepage in upland and discharged to streams, lakes or topographic depressions and by evapotranspiration in the central part of the Valley. The total regional circulation through the ground-water flow system was estimated to be approximately 200,000 acre-ft/yr.

•Since development, about 64 million acre-ft of ground water has been removed from aquifer storage; 40 million acre-ft of water was removed due to lowering of water table, 17 million acre-ft was removed due to compaction of fine-grained sediments accompanied by land subsidence, and 7 million acre-ft was removed from the elastic storage of the confined aquifer.

•Simulation indicates that during 1961-77, ground-water discharge was about 11.8 million acre-ft/yr, of which 94 percent was for irrigation (11.1 million acre-ft/yr), 3 percent was for municipal water supplies (350,000 acre-ft/yr), and 3 percent discharged to streams, lakes and topographic depressions (350,000 acre-ft/yr). The simulation also indicates that the ground-water recharge was about 11 million acre-ft/yr, of which 81 percent was irrigation return flow (9.0 million acre-ft/yr), 14 percent was percolated from precipitation (1.5 milion acre-ft/yr), and 5 percent was induced from streams and lakes (0.5 million acre-ft/yr). The difference between annual ground-water discharge and recharge was 0.8 million acre-ft/yr. About 50 percent of this difference was derived from the aquifer as the result of water-table declines (0.4 million acre-ft/yr) and another 50 percent was derived from elastic storage and compaction of sediments.

•The average horizontal hydraulic conductivity of the valley sediments is about 6 ft/d. The horizontal hydraulic conductivity of sediments in the Sacramento Valley is about twice that for sediments in the San Joaquin Valley.

•The average thickness of the continental deposits in the Central Valley is about 2,400 feet, and increases from north to south with the maximum thickness of about 9,000 feet near Bakersfield.

•Ground-water quality has the potential of being degraded by poor quality irrigation return flow. However, due to time and budget constraints, this problem was not fully explored during the initial phase of this study. A second unexpected problem was the potential migration of an extensive body of saline water lying beneath the freshwater aquifer system. Both problems are to be studied during phase II (a followup) study.

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FLORIDAN REGIONAL AQUIFER-SYSTEM STUDY

By **Peter W. Bush and Richard H. Johnston, Atlanta, Georgia**

INTRODUCTION

The Floridan aquifer system is one of the major sources of ground-water supplies in the United States. This highly productive aquifer system underlies all of Florida, southeastern Georgia, and small parts of adjoining Alabama and South Carolina, for a total area of about 100,000 mi^2. A total of about 3 Bgal/d is withdrawn from the aquifer system, and, in many areas, the Floridan aquifer system is the sole source of freshwater.

During 1978-83, the Survey conducted a regional assessment of the Floridan aquifer system that involved the review and synthesis of many previous studies, the acquisition of new data in selected areas, and the extensive use of computer-based models to simulate the ground-water flow.

The Floridan aquifer system includes several Tertiary carbonate formations that are hydraulically connected in various degrees to form a regional aquifer system. However, locally there are significant differences in hydrologic properties, water chemistry, and flow; and development has proceeded unevenly with large withdrawals concentrated in a few areas.

The approach to studying the Floridan aquifer system was to focus on (and document) local differences while tying together, in a regional analysis, the individual segments of the aquifer system.

A series of regional geohydrologic, geochemical, and potentiometric surface maps was prepared. Eleven of these maps were published during the course of the study (Johnston and others, 1980, 1981; Miller, 1982 a-e; Sprinkle, 1982 a-d).

A data-collection program was undertaken to fill the data gaps. This work involved a program of exploratory drilling, aquifer tests, seismic surveys (onshore and offshore), selective geochemical sampling, and mass measurement of water levels and artesian pressures. A notable example of these activities was the collection of hydrologic and geochemical data from an abandoned oil exploratory well 55 miles offshore from the east Florida coast (Johnston and others, 1982).

Computer simulation involved the design and calibration of a "coarse-mesh" regional flow model and four subregional flow models. The goal of the regional flow model was to understand the major features of the flow system. The design of the regional flow model and the results of simulated predevelopment conditions are described by Bush (1982). The four subregional flow models focus on the areas of greatest ground-water development. Preliminary reports describing model design and the results of simulated predevelopment conditions were prepared for three subregions: southeast Georgia, including small adjacent parts of South Carolina and Florida (Krause, 1982); west-central Florida (Ryder, 1982); and east-central Florida (Tibbals, 1981). Final reports of the Floridan regional aquifer system study comprise nine chapters in the U.S. Geological Survey Professional Paper 1403, designated 1403-A through I. As of September 1984, three of the nine chapters have been released and are in press, five are in colleague review, and one is still in preparation.

Throughout the study, the central theme was to translate the hydrogeologic framework into, first, a conceptual model, and then into a flow model that could be used to quantify aquifer properties and components of the flow system. Figure 9 illustrates the results of this process for a typical area in central Florida.

HYDROGEOLOGIC FRAMEWORK

The aquifer system is a sequence of hydraulically connected carbonate rocks (principally limestone with some dolomite) that range in age from late Paleocene to early Miocene. The rocks vary in thickness from a featheredge in outcrop areas to more than 3,500 feet in coastal areas. The aquifer system generally consists of an upper and lower aquifer separated by a less-permeable confining unit of highly variable properties. In parts of northern Florida and southwestern Georgia, there is little permeability contrast within the aquifer system. Thus, in these areas, the Floridan acts effectively as one continuous aquifer. The upper and lower aquifers are defined on the basis of permeability contrast and their boundaries locally do not coincide with either time-stratigraphic or rock-stratigraphic units.

Prior to this study, several stratigraphic units and two aquifer names were generally associated with the carbonate rocks now defined as the Floridan aquifer system. In Florida, the term "Floridan aquifer" was widely applied although its thickness and extent differed among investigators. In Georgia and South Carolina, the term "principal artesian aquifer" was applied to these carbonate rocks. A major contribution of this study was the regional definition of the Floridan aquifer system, its component aquifers and confining units, and their relation to associated stratigraphic units. The regional definition resulted from subsurface hydrogeologic studies that used geophysical and lithologic logs, faunal data, and hydrologic data. Based on analyses of these data, a series of structure contours, isopach maps and sections were prepared for the component aquifers and the seven principal stratigraphic units that comprise all or parts of the Floridan aquifer system.

Throughout the coastal areas, the effective lower boundary of the aquifer system is a transition from circulating freshwater to underlying saline water. The configuration of this transition zone agrees, in general, with the principles outlined by Hubbert (1940) for the case of an interface between moving freshwater and static saline water. That is, the interface is deepest beneath inland areas; it rises seaward and ultimately intersects the top of the aquifer system. In inland and updip areas, however, the base of the Floridan aquifer system is considered as the contact surface below which rocks have very low permeability. This surface generally coincides with a facies change to fine-grained clastic rocks or bedded anhydrite.

Low-permeability clastic rocks overlie the Floridan aquifer system in much of its area of occurrence. The lithology, thickness, and integrity of these low-permeability rocks have a controlling effect on the flow in the Floridan aquifer system locally. A surficial sand aquifer overlies these low-permeability rocks in many places.

CONCEPTUAL MODEL

A conceptual model of the Floridan aquifer system, as shown in figure 9, incorporates the major features of the flow system. The Upper and Lower Floridan aquifers are separated by a less-permeable middle confining unit. The low-permeability clastic rocks separating the Upper Floridan from the surficial aquifer are considered an upper confining unit. A freshwater- saltwater interface forms the base and lateral boundary of the system in coastal areas. Fine-grained clastic rocks or bedded anhydrite act as a lower confining unit and thus are treated as the base of the aquifer system in inland and updip areas.

DIGITAL FLOW MODEL

The Trescott-Larson quasi-three-dimensional finite-difference model (Trescott, 1975; Trescott and Larson, 1976) was used to simulate the regional and subregional flow systems. The basic structure of the model is shown in figure 9, although the subregional models depart from the regional model in some degree to accommodate local conditions or features. The Upper and Lower Floridan aquifers were simulated as active layers; the surficial aquifer was treated as a source-sink bed; and the lower confining unit and the freshwater-saltwater interface were assumed to be no-flow boundaries. However, the Upper Floridan, in a strip of area adjacent to its updip limit in Georgia, was bounded below by a constant-head boundary which simulated a small amount of upward leakage from the sand aquifer system that underlies the Upper Floridan. The other part of the updip limit of the Floridan aquifer system was considered as a no-flow boundary, because it is generally a pinchout of the carbonate rocks.

The Flow System

The major features of the flow system are illustrated and summarized by a potentiometric surface map of the Upper Floridan aquifer constructed from more than 2,700 water-level or pressure-head measurements made in May 1980 (fig. 10). Superimposed on figure 10 is a delineation of areas where the aquifer system is unconfined or loosely confined (the definition of loosely

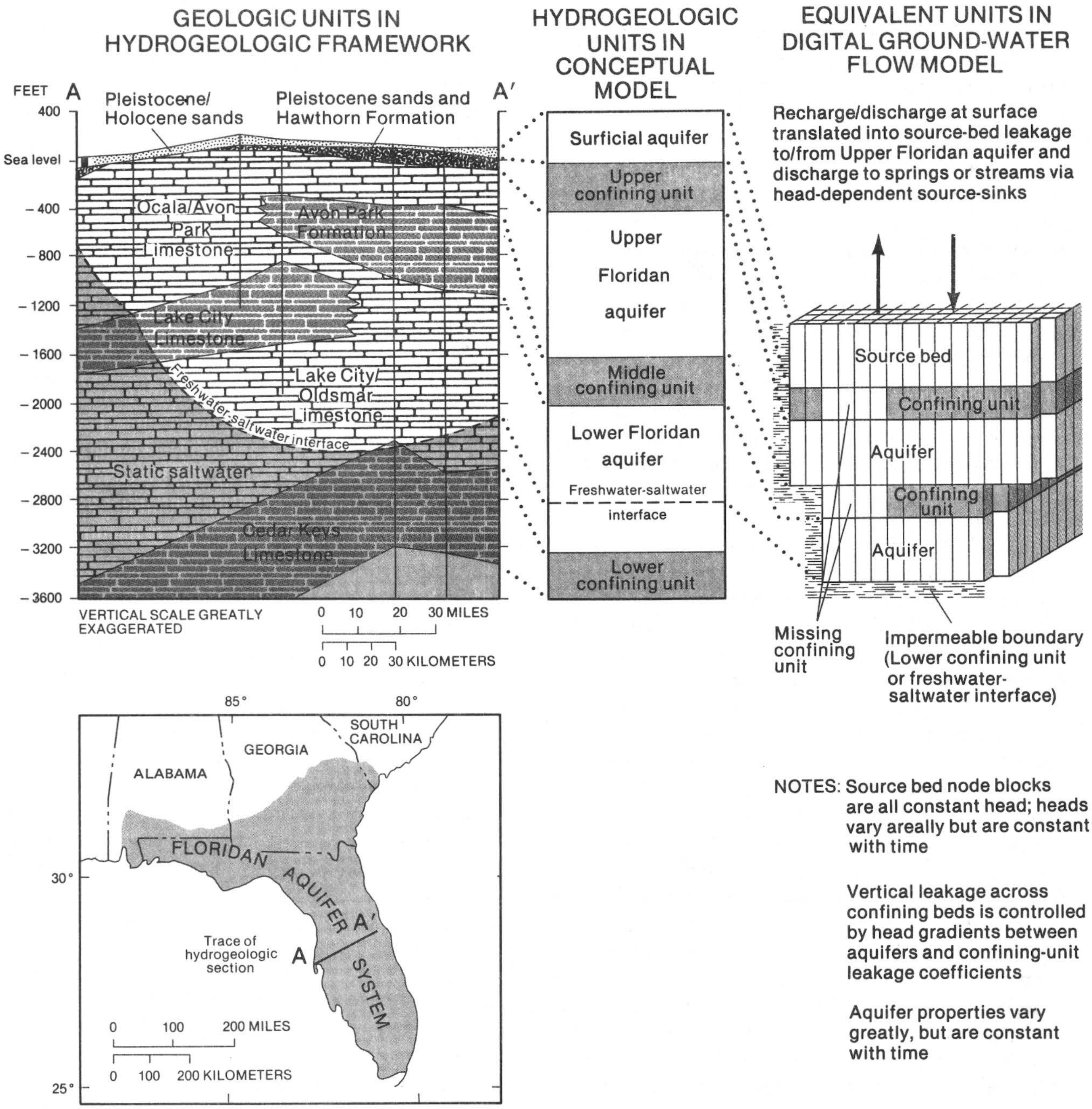


Figure 9.—Translation of hydrogeologic framework into conceptual and digital ground-water flow model, Floridan aquifer system.

confined hereafter used in this report is that the confining unit is less than 100 feet thick, breached, or both).

The configuration of the potentiometric surface indicates that, in South Carolina and Georgia, the direction of flow is generally east and southeast from the topographically high outcrop areas toward the Atlantic coast and Florida. In Alabama and western Florida, flow is generally south from the outcrop areas toward the Gulf coast. In peninsular Florida, the general flow direction is from the central inland areas toward the Gulf and Atlantic coasts. Thus, regional recharge occurs primarily in northern outcrop and peninsular inland areas, and regional discharge occurs in coastal areas.

The degree of confinement of the Upper Floridan is the characteristic that most strongly influences the distribution of natural recharge, flow, and discharge. Most of the natural recharge, flow, and discharge occurs in the unconfined and loosely confined areas. Potentiometric contours that are distorted as they cross streams indicate the interaction between streams and the Upper Floridan aquifer, and typify unconfined and loosely confined conditions. Smoother, less distorted contours are associated with tightly confined (the definition of tightly confined hereafter used in this report is that the confining unit is more than 100 feet thick) parts of the system that have less interaction with the surface drainage features.

Before development, the flow system was in a state of dynamic equilibrium in which natural recharge to the Floridan aquifer system was balanced by natural discharge. The predevelopment recharge area is estimated to be about 67,000 mi^2; the discharge area is estimated to be about 55,000 mi^2, of which about 27,000 mi^2 was land discharge area. The total predevelopment recharge, and therefore discharge, simulated by the regional flow model was about 21,500 ft^3/s. This is equivalent to 4.4 in./yr of water distributed evenly over the entire estimated recharge area.

The dominant feature of the Floridan aquifer system, both before and after development, is discharge through springs. The locations of the springs are shown in figure 10. Nearly all of the springs occur in unconfined and loosely confined areas. In the early 1980's, the average total discharge from about 300 known Upper Floridan springs was about 13,000 ft^3/s. In addition to discharge from known springs, point discharges to streams and lakes in unconfined and loosely confined areas is appreciable. About 7,000 ft^3/s discharge to streams and lakes, in addition to spring flow, was simulated by the regional flow model. Spring flow and point discharge to streams and lakes accounted for a very high percentage of the total aquifer discharge. Before ground-water development, simulated spring flow and point discharge to surface-water bodies was 88 percent of the 21,500 ft^3/s simulated discharge, or about 19,000 ft^3/s. Diffuse upward leakage, which occurs primarily in the confined areas, accounted for the remaining fraction of the total simulated predevelopment discharge, about 12 percent or 2,500 ft^3/s.

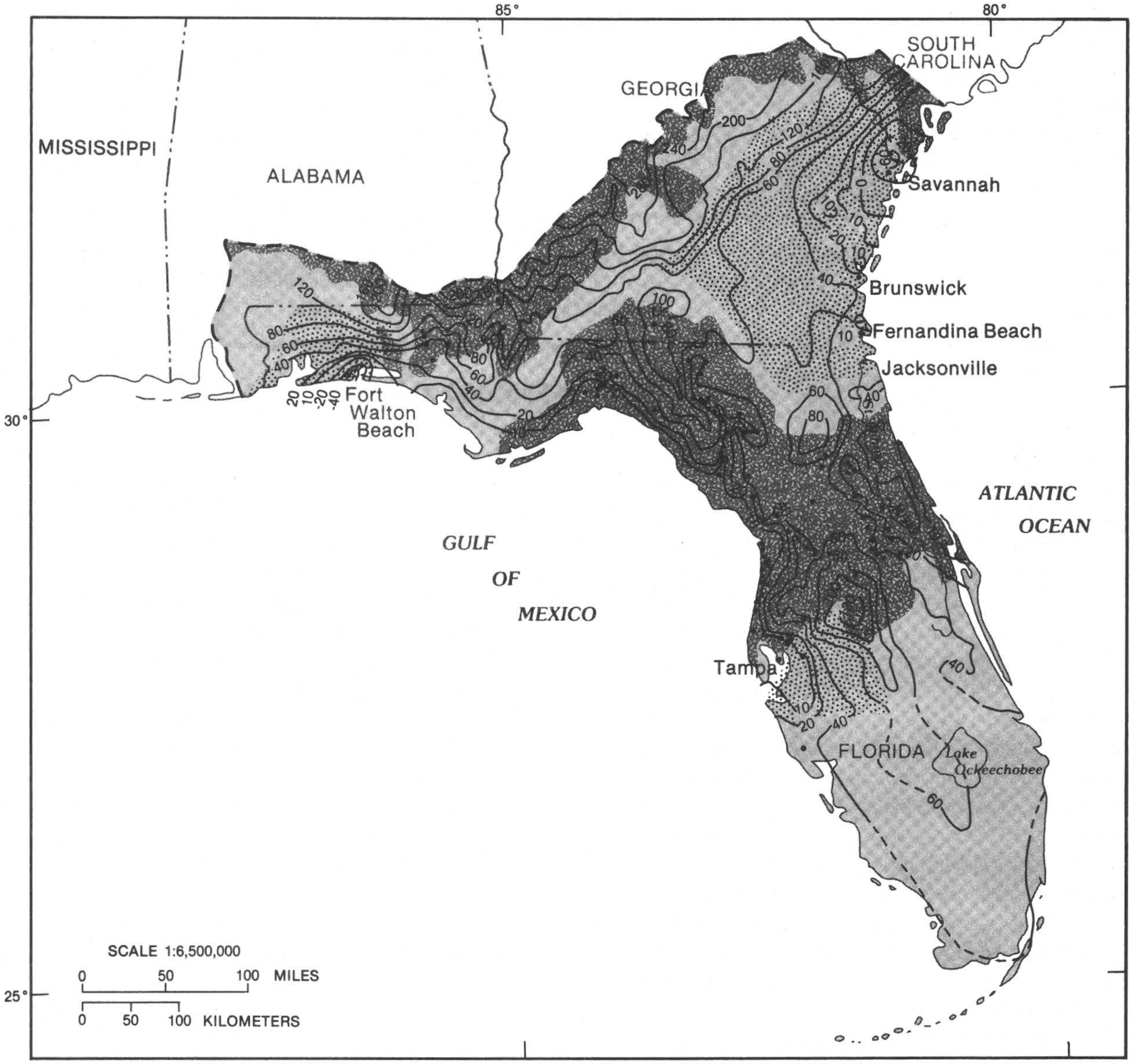
Most of the recharge necessary to sustain spring flow and aquifer discharge to streams and lakes occurs relatively close to the discharge points. Recharge to the Upper Floridan is highest in unconfined and loosely confined spring areas and averages 10 to 20 in./yr. The proximity of high recharge to high discharge implies a vigorous and well developed shallow flow system in the unconfined and loosely confined parts of the Upper Floridan aquifer.

The distribution of transmissivity is generally related to the degree of confinement. All of the areas of very high transmissivity (greater than 1,000,000 ft^2/d) and much of the area of high transmissivity (250,000 to 1,000,000 ft^2/d) occur where the aquifer is either unconfined or loosely confined.

Areas of very high transmissivity are characterized by extensive development of secondary permeability (solution openings) in the carbonate rock. Solution openings ranging from slightly widened joints to caves and large sinkholes characterize the upper part of the aquifer system. The resulting permeability distribution is extremely complex and marked differences in transmissivity occur within short distances.

In contrast to ground-water flow in unconfined and loosely confined areas, flow in parts of the aquifer that are deeply buried and tightly confined is very sluggish. These areas are in southeastern Georgia and northeastern Florida, south Florida, and western panhandle Florida. Springs and points of discharge to streams and lakes are practically nonexistent in these areas; natural discharge occurs almost exclusively by diffuse upward leakage through thick confining units. As would be expected in areas of sluggish flow, rates of recharge and discharge are considerably lower than the rates in unconfined and loosely confined areas. Recharge and diffuse upward leakage average less than 5 in./yr in tightly confined areas. Transmissivity is also generally lower in deeply buried and tightly confined areas.

Comparatively little is known about the Lower Floridan aquifer because in most areas wells obtain sufficient supplies from the Upper Floridan; thus, there is little reason to drill into the Lower Floridan. Aquifer-test data or well performance information are therefore minimal. Freshwater flow is considerably less in the Lower Floridan than in the Upper Floridan.The lack of field data precludes precise areal definition of transmissivity of the Lower Floridan aquifer. The Lower Floridan aquifer in south Florida includes a highly permeable cavernous unit known as the "Boulder Zone." This zone contains saline water and has been used for underground disposal of treated sewage through injection wells.



EXPLANATION

UNCONFINED OR LOOSELY CONFINED AREAS OF THE UPPER FLORIDAN AQUIFER -- Unconfined-upper confining unit is absent or very thin. Loosely confined-upper confining unit is generally less than 100 feet thick, breached, or both

AREAS WHERE LONG-TERM WATER-LEVEL DECLINE IS GREATER THAN 10 FEET IN THE UPPER FLORIDAN AQUIFER

—40— - POTENTIOMETRIC CONTOUR -- Shows altitude of potentiometric surface of the Upper Floridan aquifer. Dashed where approximately located. Contour interval, in feet, is variable. Hachures indicate depressions. Datum is sea level

— — — APPROXIMATE UPDIP LIMIT OF THE UPPER FLORIDAN AQUIFER

• SPRING OR GROUP OF SPRINGS IN THE UPPER FLORIDAN AQUIFER

STUDY AREA

Figure 10.—Potentiometric surface of the Upper Floridan aquifer, May 1980; occurrence of unconfined or loosely confined conditions; locations of springs discharging from the Upper Floridan aquifer; and areas of long-term water-level decline.

To quantify and summarize the areal distribution of predevelopment flow, the estimated predevelopment discharge from the major ground-water areas of the Upper Floridan aquifer is shown in figure 11. Regionally, and in every area except south Florida, the predominance of spring discharge and aquifer discharge to surface-water bodies over diffuse upward leakage is apparent. The five areas that are predominantly unconfined or loosely confined (Dougherty Plain-Apalachicola, Thomasville-Tallahassee, Suwannee, west-central Florida, and east-central Florida), although comprising only about 50 percent of the occurrence of the Floridan aquifer system, contribute nearly 90 percent of the simulated total predevelopment discharge. The Suwannee area is the most active part of the aquifer system in terms of ground-water flow. More than one-fourth of the simulated total predevelopment discharge, close to 6,000 ft^3/s, occurred there.

In tightly confined panhandle Florida, diffuse upward leakage occurred over the major part of the area. But confinement is lacking in the eastern third of the panhandle and along the outcrop, allowing direct aquifer discharge to streams. Similarly, in the southeast Georgia-northeast Florida and southwest South Carolina area, about three-quarters of the simulated predevelopment discharge leaked to four major rivers crossing the areally small northern outcrop. Diffuse upward leakage occurred over a much larger area, but accounted for a minor part of the total discharge in the area. Only in south Florida, where the Floridan aquifer is tightly confined, was diffuse upward leakage the major part of predevelopment discharge. The simulated 100 ft^3/s predevelopment discharge from south Florida represents less than 1 percent of the total regional discharge.

Pumping Effects

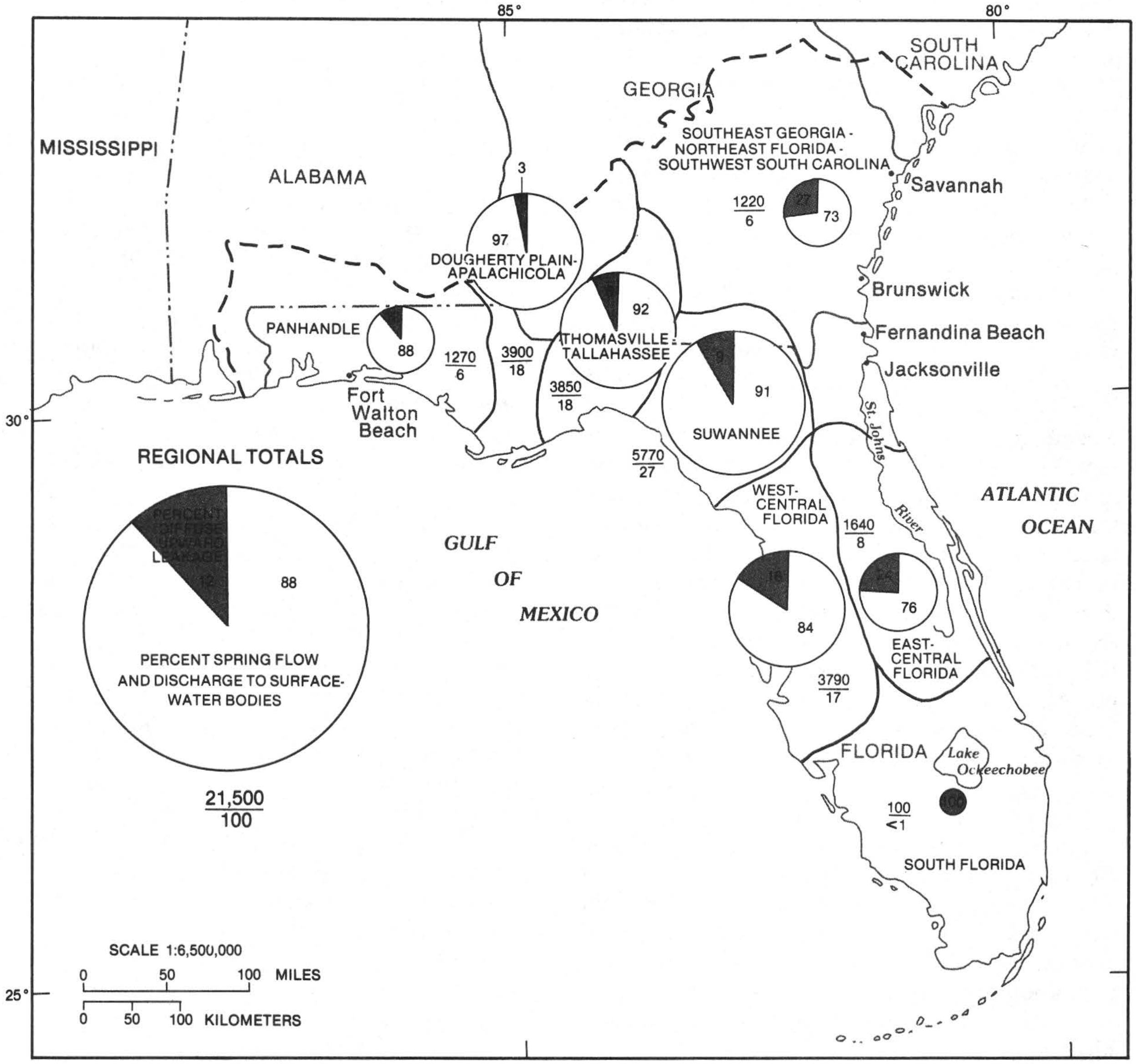
The general characteristics of the flow system described herein have not been altered appreciably by development. However, similarity of the 1980 flow system to the predevelopment flow system does not mean that ground-water development has not brought significant changes. In 1980, about 3 Bgal/d of water were pumped from the aquifer system (almost all from the Upper Floridan) for all uses. This is equal to about 20 percent of the estimated predevelopment recharge or discharge. This pumpage has resulted in long-term regional water-level declines of more than 10 feet in three broad areas (fig. 10): (1) coastal Georgia, adjacent South Carolina and northeastern Florida; (2) west-central Florida; and (3) panhandle Florida. The effect of ground-water development on the potentiometric surface is particularly evident at Savannah and Brunswick, Ga., and at Fernandina Beach and Fort Walton Beach, Fla., where deep cones of depression have formed. Saltwater encroachment as a result of pumping has occurred in some coastal areas, but its extent has been local.

Pumpage has been and continues to be balanced primarily by the diversion of natural discharge from the aquifer system and by induced recharge rather than by depletion of aquifer storage. The transient response of the aquifer system to changes in pumping rates dissipates fairly rapidly (days or weeks) in most areas. Thus, on the average (that is, excluding the effects of seasonal changes in stresses), the aquifer system is considered to be approximately at equilibrium (steady-state conditions) except during short periods following sustained increases in pumpage.

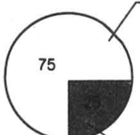
As of 1980, about 470 Mgal/d, mostly for industrial use, were being withdrawn from the coastal strip of southeastern Georgia and northeastern Florida. As a result, what began as a trough-like depression of water levels in the 1940's in coastal Georgia and northeastern Florida has spread inland to become the largest area of significant water-level declines in the Floridan aquifer system.

The overall extent of this broad area of regional water-level decline reflects the potential for increasing recharge, and diverting natural discharge to wells. Figure 10 shows that in the area of southeast Georgia and northeast Florida the regional water-level decline is coincident with a tightly confined part of the aquifer. Thus, there is little potential for increasing recharge or reducing upward leakage near the pumping centers. Therefore, the area of pumping influence in northeast Florida and southeast Georgia has spread southwestward to the region of loosely confined or unconfined areas where natural recharge is greater.

In west-central Florida, about 1 Bgal/d is pumped. The phosphate industry and irrigation account for the major part of the pumpage. As a result, a regional depression exists in the potentiometric surface southeast of Tampa. Although the pumping rate is greater in west-central Florida than in southeast Georgia and northeast Florida, the area of influence and magnitude of water-level decline is smaller in west-central



EXPLANATION



PERCENT SPRINGFLOW AND DISCHARGE TO SURFACE-WATER BODIES

PERCENT DIFFUSE UPWARD LEAKAGE

– – – APPROXIMATE UPDIP LIMIT OF THE UPPER FLORIDAN AQUIFER

——— BOUNDARY OF MAJOR GROUND WATER AREAS

21,500 — CUBIC FEET PER SECOND TOTAL DISCHARGE

100 — PERCENT OF TOTAL DISCHARGE -- Values associated with largest circle are regional totals; others are area totals

Figure 11.—Estimated predevelopment discharge from major ground-water areas of the Upper Floridan aquifer.

Florida. This is mainly because the potential for increasing recharge, and diverting natural discharge to wells, is greater in west-central Florida than in southeast Georgia and northeast Florida. Figure 10 shows that most of the area of net water-level decline in west-central Florida occurs in places where the aquifer system is under tightly confined conditions. However, the center of heaviest pumping in west-central Florida is in the northern part of area of the regional water-level declines. This center is close to a region of loosely confined or unconfined areas, thus enhancing the potential for inducing recharge.

In the Fort Walton Beach area of panhandle Florida, pumping of about 15 Mgal/d has caused a regional depression in the potentiometric surface. This relatively low pumping rate has caused a regional depression because, in this area, transmissivity is much lower than that in either southeast Georgia and northeast Florida, or west-central Florida. In much of the area of water-level declines in panhandle Florida, transmissivity is less than 10,000 ft^2/d. Confinement in the vicinity of Fort Walton Beach is at least as tight as in any other part of the aquifer system. The area of appreciable water-level declines appears to have spread eastward to a loosely confined area. However, the very low transmissivity limits the ability of the aquifer to transmit water that is available in the loosely confined area to wells. Because of continually increasing pumpage, the area of water-level declines in panhandle Florida continues to enlarge. Because of the proximity of loosely confined conditions to the east, the panhandle area of water-level declines is expected to grow less rapidly in that direction than in other directions.

Simulation of 1980 pumpage by the regional flow model indicates that development has caused the total recharge area to expand from about 67,000 mi^2 before development to about 76,000 mi^2 in 1980. Recharge to or discharge from the simulated aquifer system increased from a predevelopment total of about 21,500 ft^3/s to about 24,100 ft^3/s in 1980. The recharge area has increased by about the same percentage as the recharge rate due to development. Therefore, the 1980 recharge rate per unit area was nearly the same as the predevelopment rate, about 4 to 5 in./yr.

If steady-state conditions are assumed, about 60 percent of 1980 pumpage was supplied by increased recharge. About half of the remaining 40 percent of the pumpage was obtained from reduction in spring flow and discharge to surface-water bodies, and the other half was obtained from reduction in upward leakage. Because the total flow rate of spring discharge and discharge to surface-water bodies is much greater than the total flow rate of diffuse upward leakage, ground-water development has reduced the total spring flow and discharge to surface-water bodies less than 5 percent, however, the reduction in total diffuse upward leakage is about 30 percent.

To quantify and summarize the areal distribution of 1980 flow conditions, figure 12 shows the discharge from the major ground-water areas of the Upper Floridan aquifer. Regionally, spring discharge and point discharge to surface-water bodies were the predominant forms of discharge from the 1980 flow system, as they were from the predevelopment flow system; simulations suggest that about three-fourths of 1980 discharge was spring flow or discharge to surface-water bodies. The remaining fourth was the sum of the pumpage (17 percent) and the diffuse upward leakage (8 percent). Ground-water development has not appreciably affected the ground-water divides that separate the eight major ground-water areas. The percentage of total discharge that occurred in each area under 1980 conditions was not significantly different from that simulated before development.

Central Florida is the most heavily developed part of the aquifer system. Roughly half of the total 1980 pumpage was withdrawn from the two central Florida areas—east-central and west-central Florida. These two areas accounted for about 29 percent of the total 1980 discharge, an increase of about 3 percent over the predevelopment discharge (figs. 11 and 12). Figure 12 shows that in central Florida, about 60 percent of the aquifer discharge is spring flow or direct discharge to surface-water bodies; about 30 percent is pumpage, and about 10 percent is diffuse upward leakage.

Pumpage in the area of southeast Georgia, northeast Florida, and southwest South Carolina accounted for about 25 percent of the total 1980 pumpage; this area is the only one in which spring flow and direct discharge to surface-water bodies has been supplanted by pumpage (figs. 11 and 12).

Figure 12 shows that in each of the four contiguous areas west of the southeast Georgia-northeast Florida-southwest South Carolina area, pumpage is a minor fraction of the ground-water discharge in the area. Even in the Dougherty

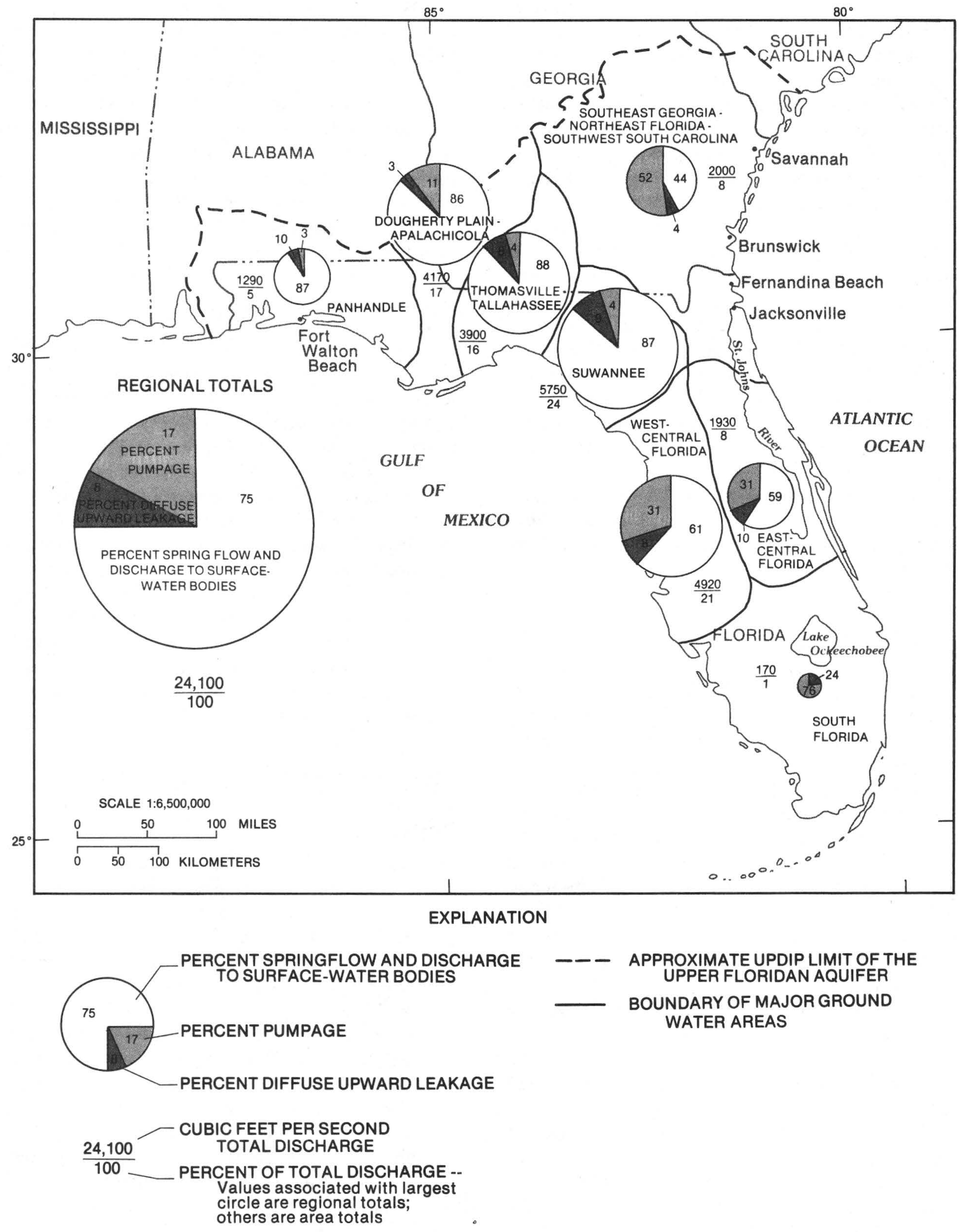


Figure 12.—Estimated 1980 discharge from major ground-water areas of the Upper Floridan aquifer.

Plain-Apalachicola area, where irrigation withdrawals have soared in recent years, pumpage represents only about 11 percent of the average ground-water discharge in the area (fig. 12). Spring flow and discharge to surface-water bodies constitute about 87 percent of the 1980 discharge in each of the four areas.

In terms of flow, the south Florida area, under 1980 conditions, is an insignificant part of the Floridan aquifer system, as it was during predevelopment time. Discharge from the area is only about 1 percent of the total discharge from the Floridan aquifer system.

In summary, the major part of the 1980 flow system is largely unchanged from predevelopment conditions. Large-discharge springs are still the dominant feature of the aquifer system. Although pumping has caused recharge rates to increase locally, the greatest recharge still occurs in areas near springs. Even after development, ground-water flow remains sluggish in areas where the aquifer is deeply buried relative to flow in areas where the aquifer is close to land surface or loosely confined.

GEOCHEMISTRY

The water chemistry in the Upper Floridan is generally related to flow and proximity to the saltwater-freshwater interface. In the unconfined or loosely confined areas, where flow is vigorous, concentrations of dissolved solids are low. Where the system is more tightly confined and flow is more sluggish; concentrations of dissolved solids are high. In Florida, south of Lake Okeechobee and in parts of the St. Johns River valley, residual saltwater remains unflushed from the system and concentrations of dissolved solids are high. Concentrations of dissolved solids also are higher in coastal areas near the saltwater-freshwater interface.

Several distinct hydrochemical facies characterize the water chemistry in the Upper Floridan. The principal chemical processes leading to the development of the hydrochemical facies are:

•Dissolution of aquifer minerals toward equilibrium as ground water moves from recharge to discharge areas.

•Mixing of fresh ground water with saltwater along the freshwater-saltwater transition zone in coastal areas; or with residual saltwater where the saltwater is unflushed by freshwater in areas comprised of low-permeability rocks.

•Cation exchange between water and aquifer minerals. In the unconfined or loosely confined areas, dissolution of calcite is the principal process and a calcium-bicarbonate hydrochemical facies occurs. In coastal areas or where the system is more tightly confined, mixing of freshwater with saltwater or unflushed residual saltwater produces a sodium-chloride facies. In western panhandle Florida, cation exchange has produced a unique occurrence of a sodium-bicarbonate facies in the Upper Floridan.

POTENTIAL FOR DEVELOPMENT

A considerable area of the Floridan aquifer system remains highly favorable for development of large ground-water supplies. This area is largely inland and is characterized by high transmissivity as well as minimal development as of 1980. Simulation suggests that new pumping centers in parts of southeast Georgia and north-central Florida can supply large withdrawals, while relatively shallow cones of depression will develop. The major constraint on future development is degradation of water quality rather than limitation of water quantity. The possibility of saltwater encroachment in coastal areas and upconing of deep saltwater in some inland areas are important factors to consider for future development.

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HIGH PLAINS REGIONAL AQUIFER-SYSTEM STUDY

By John B. Weeks, Denver, Colorado

INTRODUCTION

The High Plains regional aquifer system underlies about 174,000 mi² in parts of Colorado, Kansas, Nebraska, New Mexico, Oklahoma, South Dakota, Texas, and Wyoming. The aquifer system is the shallowest and most abundant source of water in one of the major agricultural areas in the United States. About 20 percent of the irrigated land in the United States is in the High Plains, and about 30 percent of the ground water used for irrigation in the United States is pumped from the High Plains aquifer system. In 1980, about 170,000 wells pumped about 18 million acre-ft of water to irrigate nearly 14 million acres.

The irrigated-agricultural economy of the High Plains is dependent on the aquifer system for continued growth and prosperity. However, declining water levels and decreasing water supplies threaten the future of irrigation using ground water in parts of the High Plains.

National concern about the economic impact of declining water supplies in the High Plains was responsible for the initiation of a regional study of the High Plains aquifer system in 1978. The regional study was completed in 1982.

GEOHYDROLOGY

Previous hydrologic studies of the region have not kept pace with ground water development. Many water-resource studies of small areas such as individual counties or groups of counties have been made, but few studies of larger areas within the High Plains have been made. Previous investigations provide a wealth of information, but without continuity across political boundaries. In developing the geohydrologic framework of the High Plains aquifer system, hydrologists assembled, organized, and assimilated the hydrogeologic data available from many previous studies.

Physical Setting

The High Plains occupies the southern part of the Great Plains physiographic province, which lies east of the Rocky Mountains. The region extends from southern South Dakota to northwestern Texas. The High Plains is characterized by flat to gently rolling terrain, which is a remnant of a vast plain formed by sediments deposited by streams flowing eastward from the Rocky Mountains.

Most of the High Plains has a middle-latitude dry continental climate with abundant sunshine, moderate precipitation, frequent winds, low humidity, and a high rate of evaporation. Mean annual precipitation increases eastward across the High Plains by about 1 inch every 25 miles from less than 16 inches near the western side of the study area to about 28 inches in eastern Nebraska and central Kansas. Typically, about 75 percent of the precipitation falls as rain during the growing season. However, much of the rain falls during local thunderstorms; therefore, large variations in rainfall can occur from place to place and year to year.

Persistent wind and high summer temperature cause high rates of evaporation. Mean annual evaporation from Class A pans ranges from about 60 inches in northern Nebraska and southern South Dakota to about 105 inches in western Texas and southeastern New Mexico. Because of these conditions, most of the water that enters the soil is returned to the atmosphere by evapotranspiration, and little precipitation is available to recharge the ground-water system, except in sand-dune areas where water can readily percolate to the water table.

Aquifer System

The High Plains aquifer system consists mainly of near-surface deposits of late Tertiary or Quaternary age (Gutentag and Weeks, 1980). The principal geologic units that comprise the High Plains aquifer system are shown in figure 13. The Tertiary rocks include part of the Brule Formation of the White River Group, the Arikaree Group, and the Ogallala Formation. The Quaternary deposits consist of alluvial, dune-sand, and valley-fill deposits. Except for dune sand, the Quaternary deposits are combined and are shown in figure 13 only in areas where they do not overly Tertiary aquifer units. In northern Texas, some collapse structures, filled with Triassic, Jurassic, and Lower Cretaceous rocks that have secondary permeability, are considered part of the aquifer system; however, they are minor and are not shown in figure 13.

The Ogallala Formation is the principal geologic unit of the High Plains aquifer system and underlies about 134,000 mi^2 (fig. 13). Maximum thickness of the Ogallala is about 700 feet. During deposition of the Ogallala, aggrading streams filled valleys eroded into pre-Ogallala rocks. Braided streams flowed eastward from the mountains, transporting rock debris, which was deposited as a heterogeneous sequence of clay, silt, sand, and gravel.

Dune-sand deposits, consisting predominantly of very fine to medium grained wind blown sand, are part of the High Plains aquifer system. Of the 174,000 mi^2 underlain by the High Plains aquifer system, about 19 percent is covered by dune sand. The most extensive area of dune sand is in west-central Nebraska, where the deposits cover an area of 20,000 mi^2 with a maximum thickness of about 300 feet. Large areas also are covered by dune-sand deposits south of the Arkansas River in Kansas (fig. 13). Throughout the High Plains, the areas covered by dune sands are important recharge areas of the aquifer system.

Bedrock that underlies the High Plains aquifer system ranges in age from Permian to Tertiary. The areal extent and age of the bedrock are shown in figure 14. Some bedrock units may yield water to wells; although, the bedrock units generally have low permeability. Bedrock of marine origin contains evaporites and saline water that affect the quality of water in the overlying aquifer. This is particularly true in areas underlain by Permian bedrock and areas south of the Canadian River underlain by Lower Cretaceous rocks.

Flow System

Climate and geology control the hydrology of the High Plains aquifer system. Precipitation, temperature, and soil characteristics limit the amount of recharge. Lithology, structure, and topography control ground-water movement and discharge.

Recharge to the aquifer system, which is generally unconfined, is entirely from precipitation and seepage from streams. Because evapotranspiration greatly exceeds precipitation, little or no recharge occurs from precipitation except in areas covered by sandy soil which have high permeability, low field capacity, and rapid infiltration rates.

Recharge on the High Plains is highly variable. Estimated recharge ranges from 0.024 in./yr in parts of Texas to 6 in./yr in sand-dune areas in Kansas and Nebraska (Gutentag and others, 1984). The volume of recharge to the aquifer system has been estimated by model simulation (Luckey and others, 1985). The simulated average annual recharge to the aquifer system was about 5.7 million acre-ft or 0.6 in./yr over the entire area of the aquifer system. About 80 percent of the recharge occurs north of 39 degrees-latitude where evapotranspiration is low and dune sands are prevalent.

Ground water in the High Plains aquifer system generally flows from west to east in response to the slope of the water table. Based on regional hydraulic gradient and aquifer properties, the velocity of water moving through the aquifer system is about 1 ft/d.

Hydraulic conductivity and specific yield are the two principal properties that control ground-water flow in an unconfined aquifer. Both properties, depending on the characteristics of the sediments, can vary horizontally and vertically. The average hydraulic conductivity of the High Plains aquifer system is about 60 ft/d. For most (68 percent) of the aquifer area, hydraulic conductivity ranges from 25 to 100 ft/d, although it may be as much as 300 ft/d for individual lithologic units. The specific yield of most (76 percent) of the aquifer area is 10 to 20 percent, although the values range from 3 to 35 percent for individual lithologic units. The average specific yield is about 15 percent.

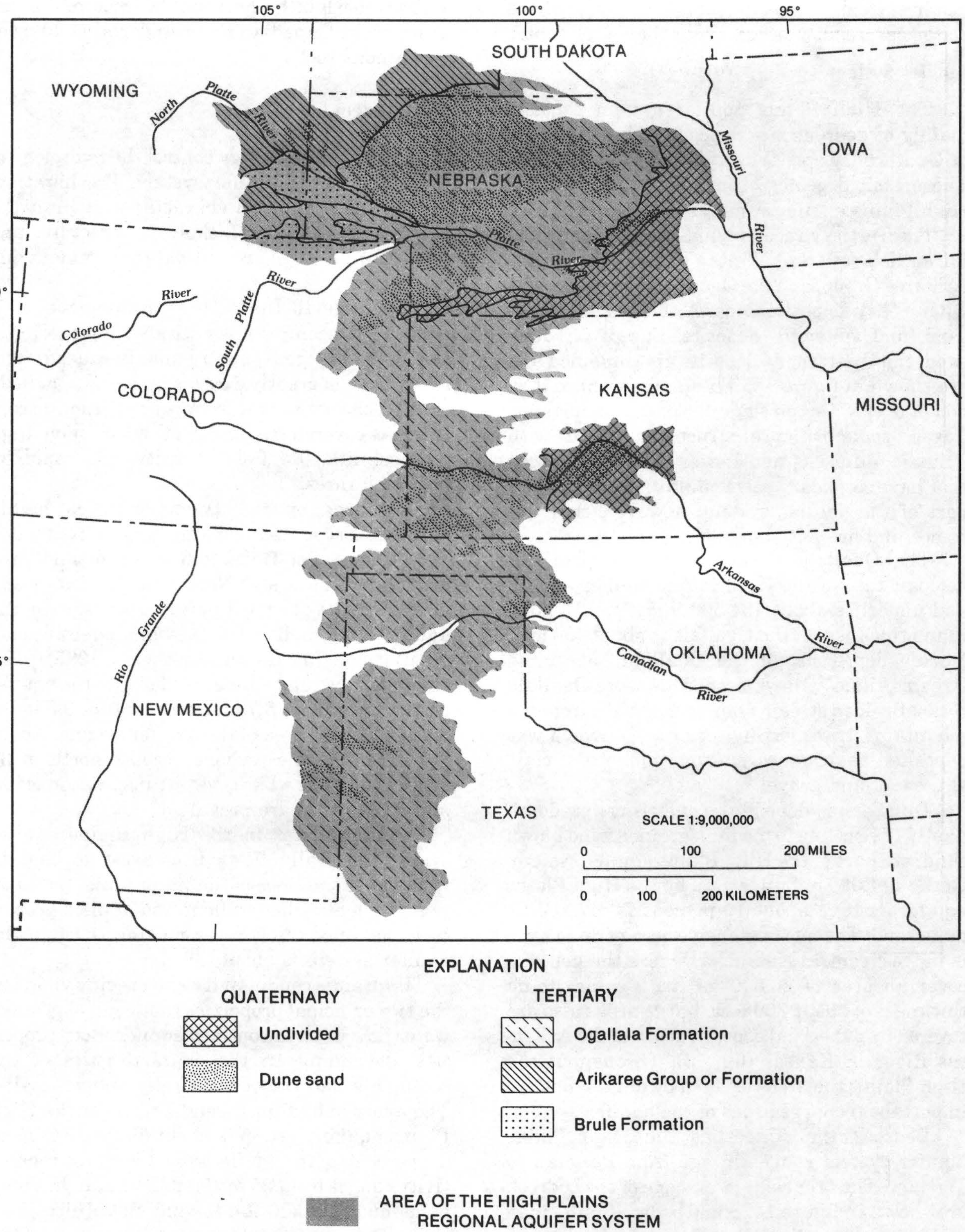


Figure 13.—Principal geologic units of the High Plains regional aquifer system.

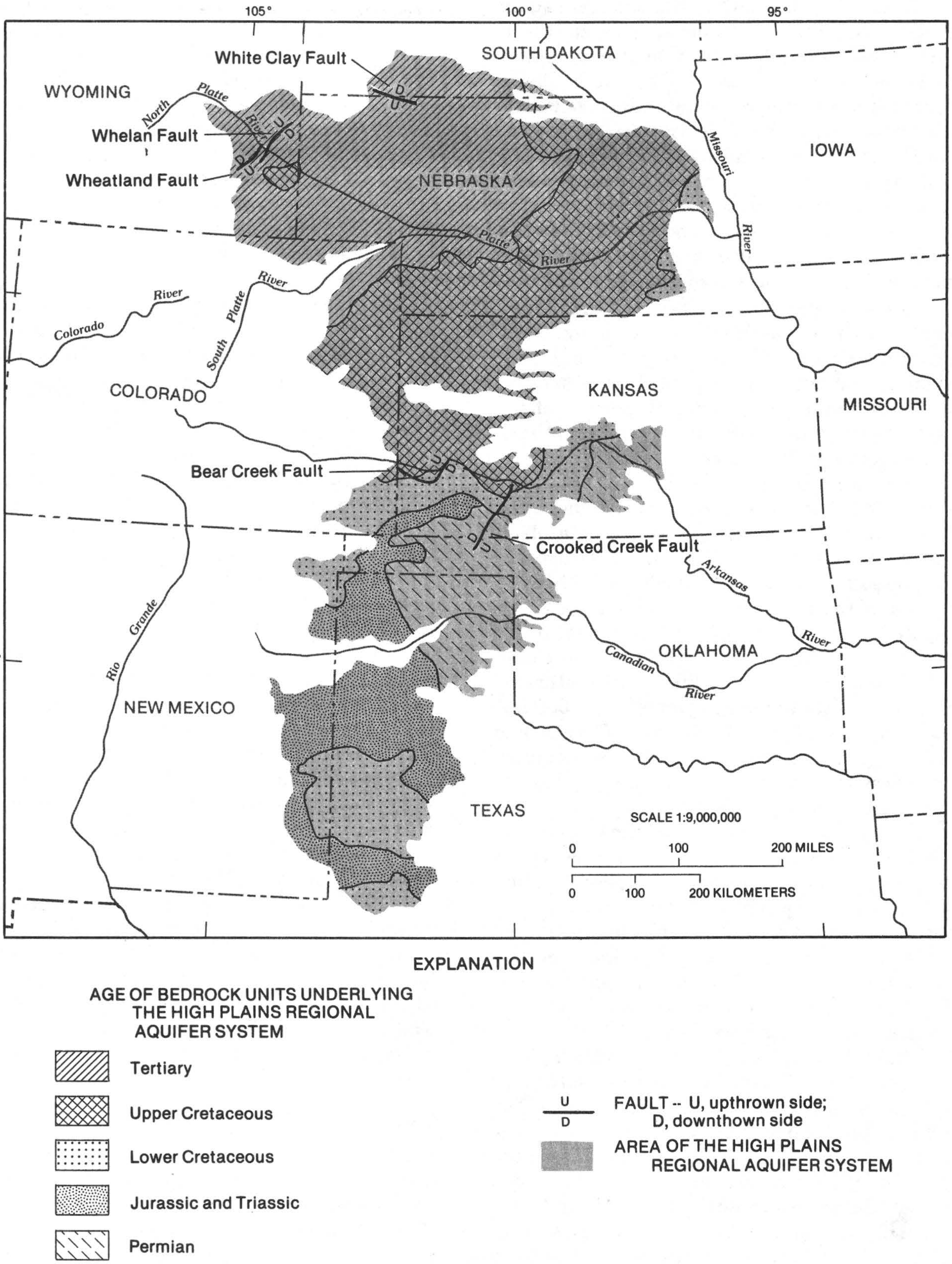


Figure 14.—Areal extent and age of bedrock units underlying the High Plains regional aquifer system.

Ground water from the High Plains aquifer system discharges to streams, springs, and seeps, and to the atmosphere by evapotranspiration. Streams that originate in the High Plains generally are ephemeral in upstream reaches and perennial where channels are incised into the water table. Ground-water discharge by evapotranspiration is greatest in stream valleys in which phreatophytes grow. In sand hills in Nebraska, evaporation from numerous lakes which are hydraulically connected to the aquifer system may equal or exceed seepage to streams.

Ground-water discharge from the High Plains aquifer system has been estimated by model simulation (Luckey and others, 1985). The simulated annual discharge to streams or lakes is 2.3 million acre-ft. About 85 percent of the discharge occurs north of 39 degrees-latitude where recharge is greatest.

Saturated thickness of the aquifer system ranges from zero, where the deposits comprising the aquifer system are unsaturated, to about 1,000 feet in west-central Nebraska. The average saturated thickness is about 200 feet. About 46 percent of the area underlain by the High Plains aquifer system has less than 100 feet of saturated thickness, whereas only 5 percent has more than 600 feet of saturated thickness. The saturated thickness of the aquifer is greater than 600 feet only in Nebraska and Wyoming (Weeks and Gutentag, 1981). Areas where the saturated thickness of the aquifer system exceeded 200 feet in 1980 are shown in figure 15.

Salt dissolution has affected the thickness of the High Plains aquifer system. Many evaporite beds in the Permian rocks underlying the High Plains are susceptible to dissolution by ground water. Sinkholes associated with salt dissolution have been forming since Permian time. Collapse structures have been filled by younger material in large areas in southwestern Kansas, Oklahoma, and northern Texas.

The volume of water stored in the High Plains aquifer system depends on the saturated thickness and specific yield of the aquifer. The total volume of drainable water in storage (product of specific yield, saturated thickness, and area) in 1980 was estimated to be 3.25 billion acre-ft (Gutentag and others, 1984). About 65 percent of the water in aquifer storage is in Nebraska, where recharge and saturated thickness are greatest because of the presence of dune sands. About 12 percent of the water in aquifer storage is in Texas, 10 percent in Kansas, 4 percent in Colorado, 3.5 percent in Oklahoma, 2 percent each in South Dakota and Wyoming, and 1.5 percent in New Mexico.

The quality of water in the High Plains aquifer system generally is suitable for irrigation; however, the water does not meet U.S. Environmental Protection Agency drinking water standards in many places. The concentrations of dissolved solids, fluoride, chloride, and sulfate exceed recommended limits for drinking-water in parts of the aquifer in all States (Krothe and others, 1982).

Most of the water in the High Plains aquifer system contains 250 to 500 milligrams per liter (mg/L) of dissolved solids and is a calcium bicarbonate type water. Ground water containing more than 500 mg/L dissolved solids generally is a mixed type of water, with calcium, sodium, sulfate, and chloride the most prevalent ions. The concentration of dissolved solids in water from the High Plains aquifer system is shown in figure 16. About 62 percent of the aquifer area contains water with 250 to 500 mg/L dissolved solids; only about 3 percent of the area contains water exceeding 1,000 mg/L dissolved solids, most of which is in Texas. Generally, concentrations of dissolved solids are least in areas covered by sand, because recharge is relatively high and the sand contains few readily soluble minerals.

Bedrock underlying the aquifer system affects the chemical composition of water in the aquifer in parts of all states in the High Plains. In some areas, water at the base of the aquifer in contact with bedrock dissolves minerals contained in the bedrock and transports them through the aquifer by diffusion or ground-water flow in response to hydraulic-head differences. In some areas, dissolution of minerals from reworked bedrock in the aquifer sediments has affected water chemistry. In most areas of the High Plains aquifer system, where the concentrations of dissolved solids exceed 1,000 mg/L, the chemical composition of the water is affected by the underlying bedrock.

GROUND-WATER PUMPAGE

Water from the High Plains aquifer system is the principal source of supply for irrigation in the area. Rapid development of irrigation in the High Plains in recent years has made the area one of the major agricultural regions of the United States. About 95 percent of all water pumped from the aquifer system is used for irrigation.

Ground-water irrigation in the High Plains was sporadic until the drought of the 1930's. By 1949, about 2 million acres of land, mostly in southern High Plains, were irrigated. Center-pivot irrigation systems, which were adaptable to sandy soil and rolling terrain, were developed during the 1960's. Center pivots made land available for irrigation that previously was not suitable, particularly in the northern parts of the High Plains. About 13 million acres of land were irrigated in 1978.

Historical Pumpage

Historical information on ground-water pumpage for irrigation was needed for a ground-

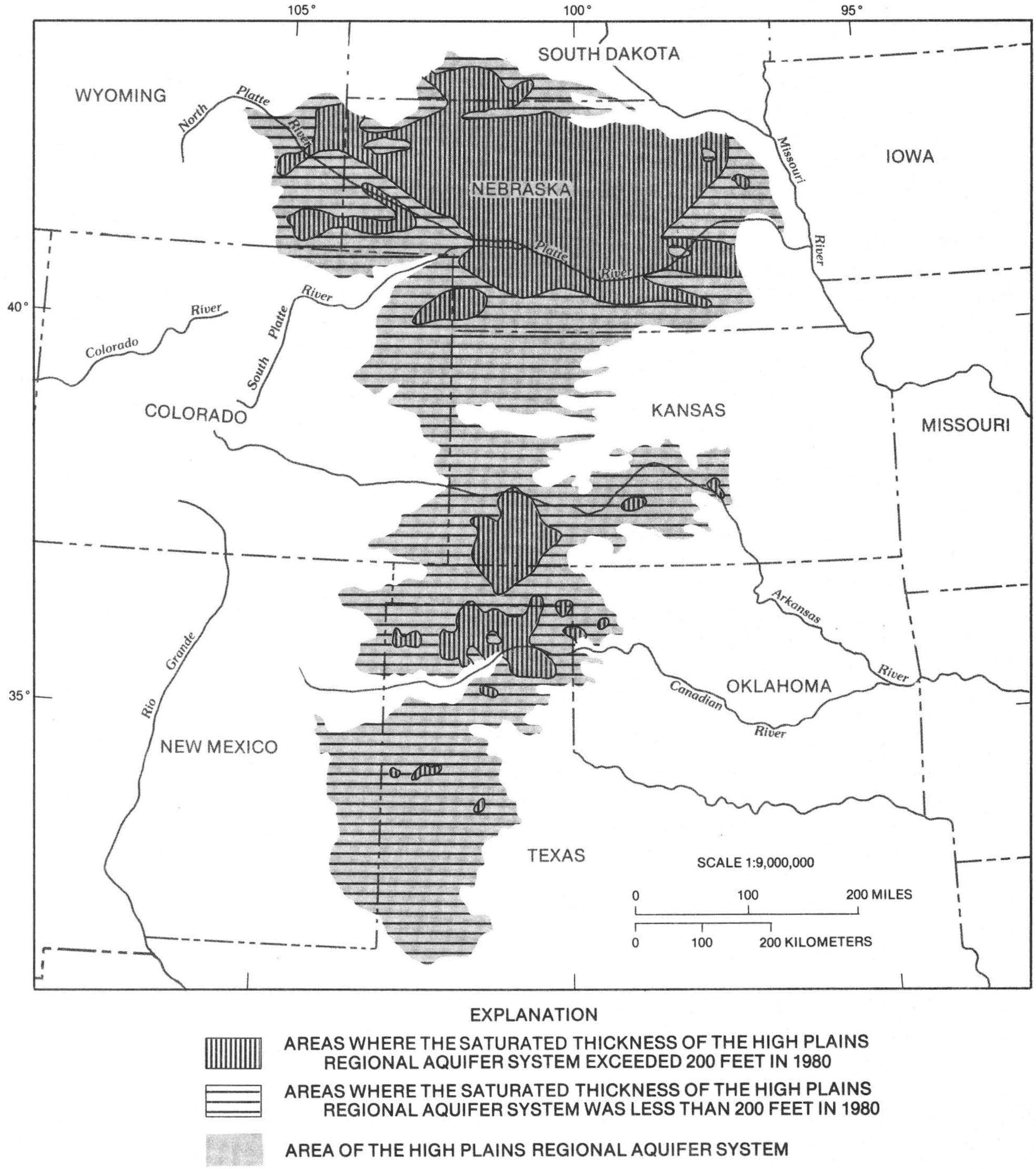


Figure 15.—Areas where saturated thickness of the High Plains regional aquifer system exceeded 200 feet in 1980.

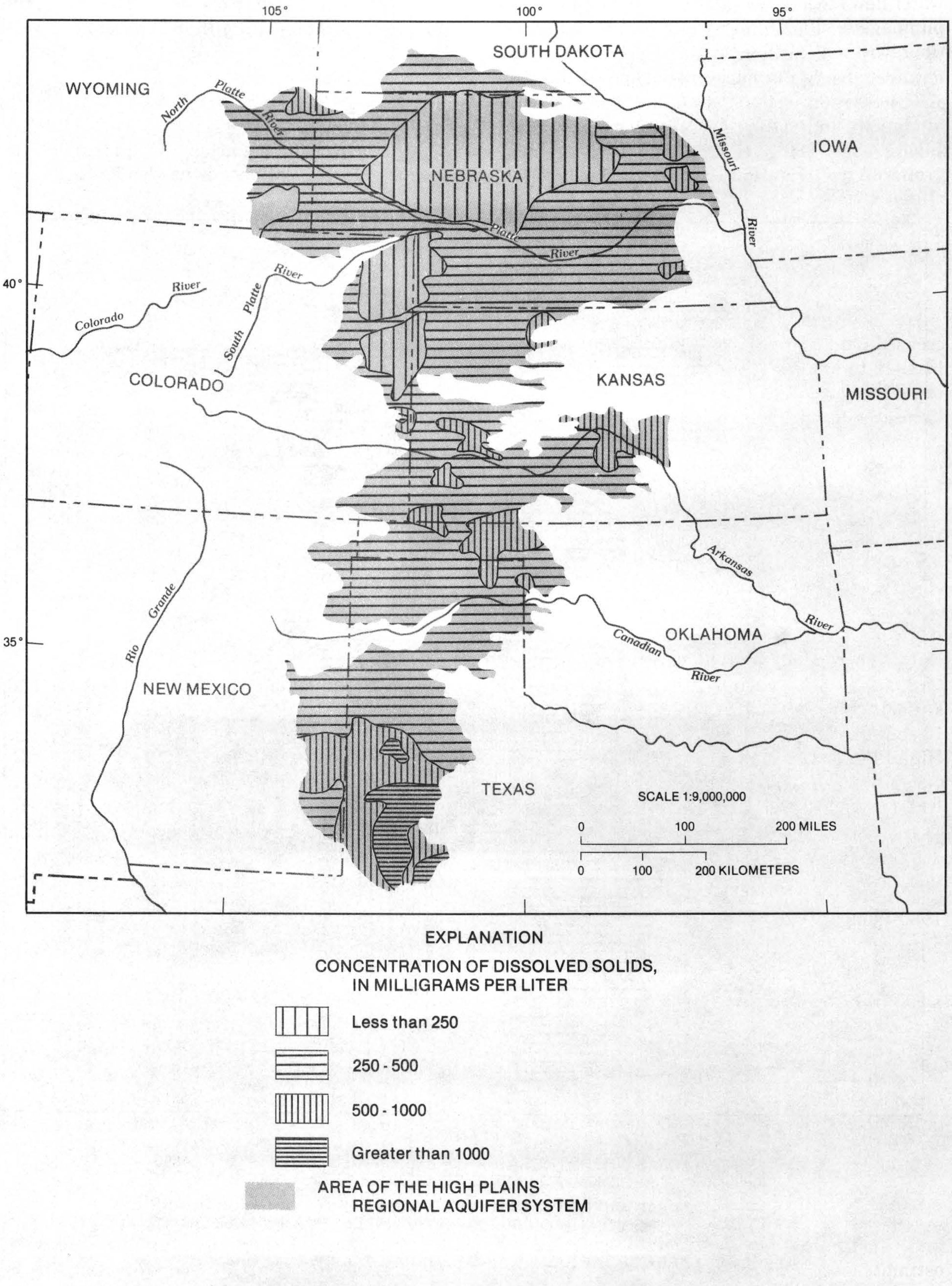


Figure 16.—Dissolved-solids concentrations in ground water in the High Plains regional aquifer system.

water flow model. Adequate data on ground-water pumpage in the High Plains were not available. Therefore, a method for estimating pumpage from irrigation requirement was developed (Heimes and Luckey, 1982). Irrigation requirement is defined as the volume of water required in addition to precipitation for all of the irrigated crops grown in a given area, on the basis of average climatic conditions.

Irrigated acreage in the High Plains, based on the census of agriculture of the U.S. Department of Commerce (1949-78), is shown in figure 17. The graph shows that the total acreage irrigated by ground water in the High Plains increased from slightly more than 2 million acres in 1949 to about 13 million acres in 1978. This rapid increase in irrigated acreage has resulted in a corresponding increase in ground-water pumpage.

The Blaney-Criddle formula (U.S. Department of Agriculture, 1967) was selected to estimate irrigation requirement on the High Plains. Ground-water pumpage was estimated by adjusting irrigation requirement for the efficiency of irrigation. The efficiency of irrigation is defined as the ratio of irrigation requirement to the water pumped. The efficiency of irrigation generally ranges from about 40 to 90 percent and depends on many factors, such as the type of irrigation system, crop and soil types, and climate. Figure 17 shows estimated ground-water pumpage in the High Plains assuming an average irrigation efficiency of 65 percent for period 1949-78. Pumpage for irrigation during 1978 was estimated at about 23 million acre-ft, which is almost 20 million acre-ft more than the estimated pumpage for irrigation during 1949.

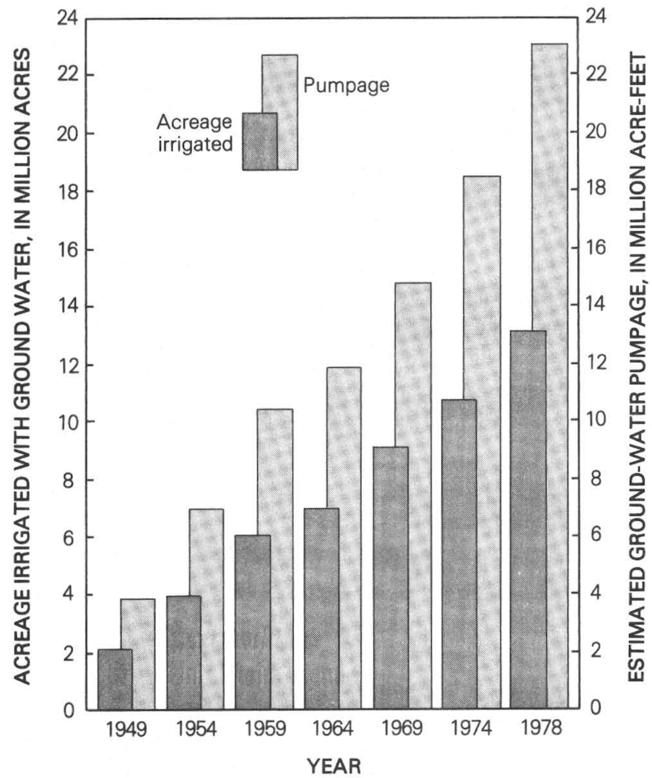


Figure 17.—Acreage irrigated by ground water and estimated ground-water pumpage in the High Plains, 1949-78.

1980 Pumpage Study

The objective of the 1980 pumpage study was to provide an estimate of the quantity of ground water withdrawn for irrigation during 1980. Historical pumpage estimates based on irrigation requirement were made because measured data on historical ground-water withdrawals generally are not available. Calculations of irrigation requirement give reasonable estimates of historical withdrawals, but changes in irrigation techniques caused by increased energy costs and decreasing water availabililty are likely to reduce the accuracy of these estimates. Consequently, a sampling program was designed to provide an estimate of ground-water withdrawal for irrigation during 1980 based on measured data. A pilot program was conducted in 1979 to test methods and instrumentation for application to the 1980 pumpage-sampling program (Luckey and others, 1980; Heimes and Luckey, 1980).

The 1979 pilot program was designed to: (1) develop a statistical approach to sampling ground-water pumpage for irrigation, (2) test instrumentation and develop procedures for measuring the annual volume of ground water pumped from selected irrigation wells, (3) determine the relation between the annual volume of ground water pumped and the acreage irrigated, and (4) develop a suitable approach to map irrigated cropland for the entire High Plains in order to extend the sampled information.

For the pilot program, the stratified random sampling technique was used to select 250 sites in two areas. Sites were stratified by crop and by area. The annual volume of ground water pumped at each site was computed by multiplying the pumping rate by the total time of pumping. During the pilot program, various approaches were evaluated for mapping the irrigated cropland.

For 1980, 15 counties, strategically located throughout the High Plains, were selected for data collection, and sampling sites were selected randomly within the counties (Heimes and Luckey, 1983). Discharge from wells, time-of-pumping, crop type, and crop-acreage data were collected at each of the sampling sites during the growing season. These data were used to compute the average depth of water applied at each site. The relation between irrigation requirement calculated using the Blaney-Criddle formula, and the sampled water application were used to estimate water application in unsampled areas of the High Plains. Estimates of application were combined with 1980 irrigated-acreage data to calculate the volume of water pumped during the 1980 growing season.

Irrigated acreage for the 1980 growing season was compiled using Landsat imagery. Fifty-nine Landsat scenes covering the High Plains were analyzed by computer to separate irrigated cropland from other land-cover categories. Estimates of irrigated acreage, derived from the computer classification of Landsat data, subsequently were aggregated into cells of 1-minute latitude by 1-minute longitude (1-minute cells). This aggregation made the irrigated-acreage data compatible with other hydrologic and geologic data compiled for this study. Data for the 1-minute cells were aggregated to estimate irrigated acreage for 1-degree cells.

The volume of ground water pumped for irrigation was computed by multiplying the water application by irrigated acreage. The volume of ground water pumped for irrigation was then estimated for each 1-degree cell, and summed up by State. The estimated volume of water pumped in 1980 for irrigation by State is summarized in table 1. The estimated total volume of water pumped during the 1980 growing season was 17,817,000 acre-ft which was applied to 13,385,000 acres. Three States (Kansas, Nebraska, and Texas) accounted for about 88 percent of this pumpage. Texas, which accounts for 20 percent of the area and 29 percent of the total pumpage, was the most densely irrigated area. South Dakota, which accounts for 2.7 percent of the area, but only 0.1 percent of the total pumpage, virtually was unirrigated.

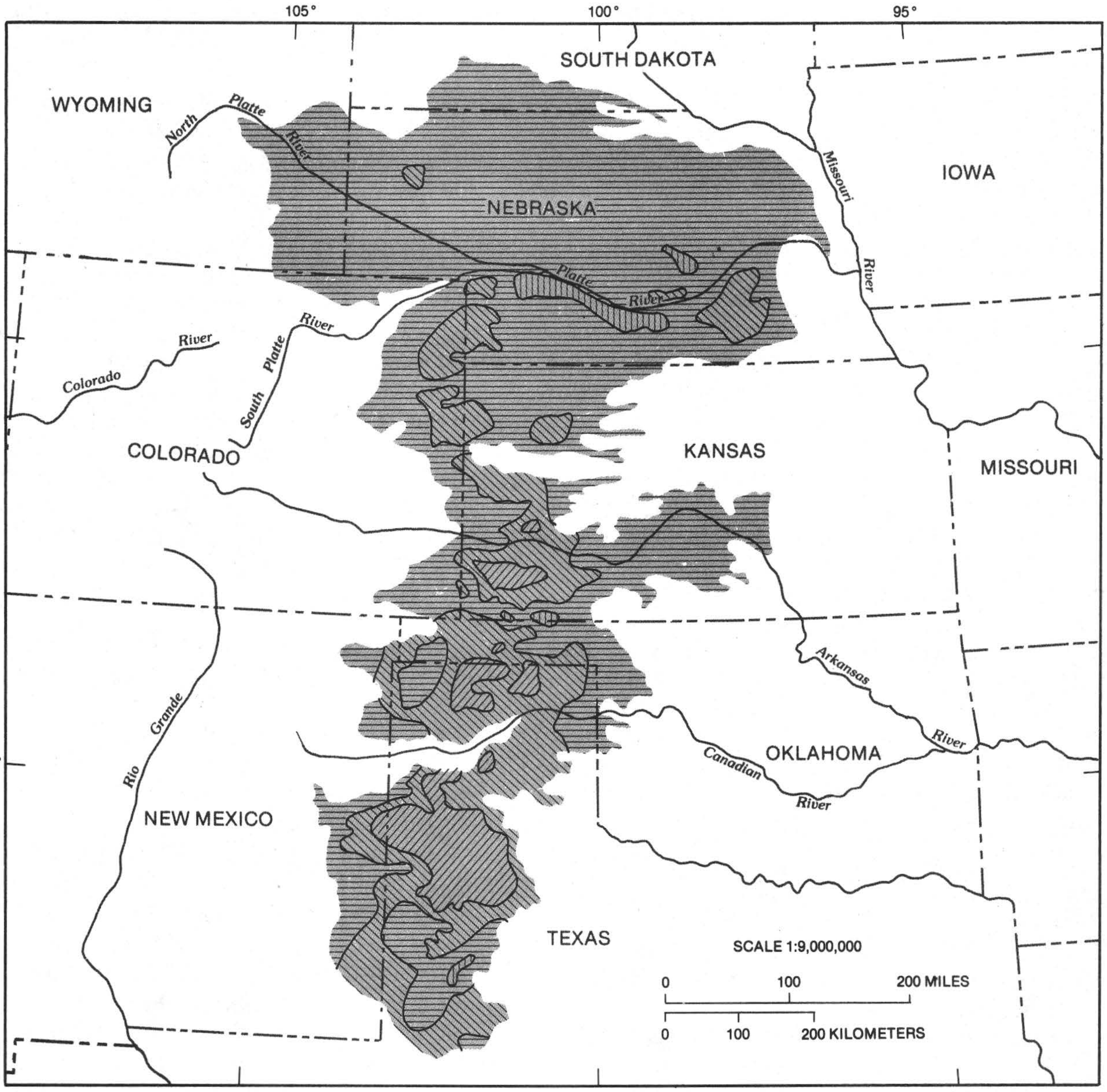
Irrigation requirement in the High Plains generally is much larger than the amount of ground-water recharge. Consequently, pumpage for irrigation in large parts of the High Plains greatly exceeds the recharge rate, and water levels in these areas are declining.

Water-Level Changes

Water-level changes from predevelopment to 1980 in the High Plains aquifer system are shown in figure 18. Areas of water-level decline caused by irrigation pumpage are found in all States except South Dakota, where irrigation development, relative to other High Plains States, is sparse. The largest area of water-level declines exceeding 50 feet occurred south of the Canadian River in New Mexico and Texas. The maximum water-level decline in the High Plains of nearly 200 feet, occurred south of the Canadian River in Floyd County, Texas (Luckey and others, 1981).

Table 1.—Estimated irrigated acreage and volume of ground water pumped for irrigation in areas of the High Plains during the 1980 growing season

State	Area of High Plains aquifer system within State (square miles)	Irrigated acreage during 1980 (acres)	Volume of ground-water pumped (acre-feet)
Colorado	14,900	767,000	985,000
Kansas	30,500	2,795,000	4,215,000
Nebraska	63,650	5,101,000	6,240,000
New Mexico	9,450	325,000	519,000
Oklahoma	7,350	389,000	540,000
South Dakota	4,750	20,000	24,000
Texas	35,450	3,878,000	5,169,000
Wyoming	8,000	110,000	125,000
Total	**174,050**	**13,385,000**	**17,817,000**



EXPLANATION

WATER-LEVEL CHANGES, IN FEET

Rise more than 10

10 to − 10

− 10 to − 50

Decline more than 50

AREA OF THE HIGH PLAINS REGIONAL AQUIFER SYSTEM

Figure 18.—Water-level changes, from predevelopment to 1980, in the High Plains regional aquifer system.

North of the Canadian River, water levels have declined more than 50 feet in parts of Texas, Oklahoma, and Kansas. Water levels have declined 10 to 50 feet in large areas in eastern Colorado, western Kansas, and southern Nebraska. Water-level declines in these areas are less severe, primarily because irrigation started later.

Water-level rises in Nebraska (fig. 18) are due to recharge from surface-water irrigation. In Kansas and Oklahoma, water-level rises probably represent recovery from abnormally low water levels during the drought of 1933-40. In Texas, the water-level rise is attributed to the clearing sandy soil of native vegetation for cultivation, resulting in increased recharge.

Significant declines in water levels (more than 10 feet) have occurred in 29 percent of the area underlain by the aquifer system, and significant rises in water level (more than 10 feet) have occurred in only 1 percent of the aquifer area. A total of 50,000 mi^2 (32 million acres) of the aquifer area have had water-level declines in excess of 10 feet, and more than 12,000 mi^2 (7.7 million acres) have had water-level declines in excess of 50 feet.

Saturated-Thickness Changes

Changes in saturated thickness of the High Plains aquifer system (in percent) from predevelopment to 1980 are shown in figure 19. Changes in saturated thickness are important because large changes in saturated thickness may affect well yield and pumping cost, or cause partially penetrating wells to go dry.

Two areas with significant increases in saturated thickness (greater than 10 percent) occur in Nebraska (fig. 19). In each of these areas, recharge to the aquifer system from surface-water irrigation has resulted in an increase in saturated thickness of more than 10 percent. Saturated thickness has increased more than 10 percent in one area in Texas. In this area, recharge to the aquifer system has increased because of land-use changes.

Areas of significant decreases in saturated thickness generally are found where the aquifer system has been pumped for irrigation for many years. Saturated thickness has decreased more than 25 percent in one-quarter of the High Plains of New Mexico and Texas where large-scale irrigation began in the 1930's. Saturated thickness also has decreased more than 25 percent in parts of Kansas and Oklahoma.

In Colorado, northwestern Kansas, Nebraska, and Wyoming, most of the irrigation has taken place since 1960. Saturated thickness decreased 10 to 25 percent in areas of Colorado and Kansas, and in a smaller area of Nebraska (fig. 19). In these areas, ground-water pumpage is beginning to reduce saturated thickness significantly, and well yields may decline in the future.

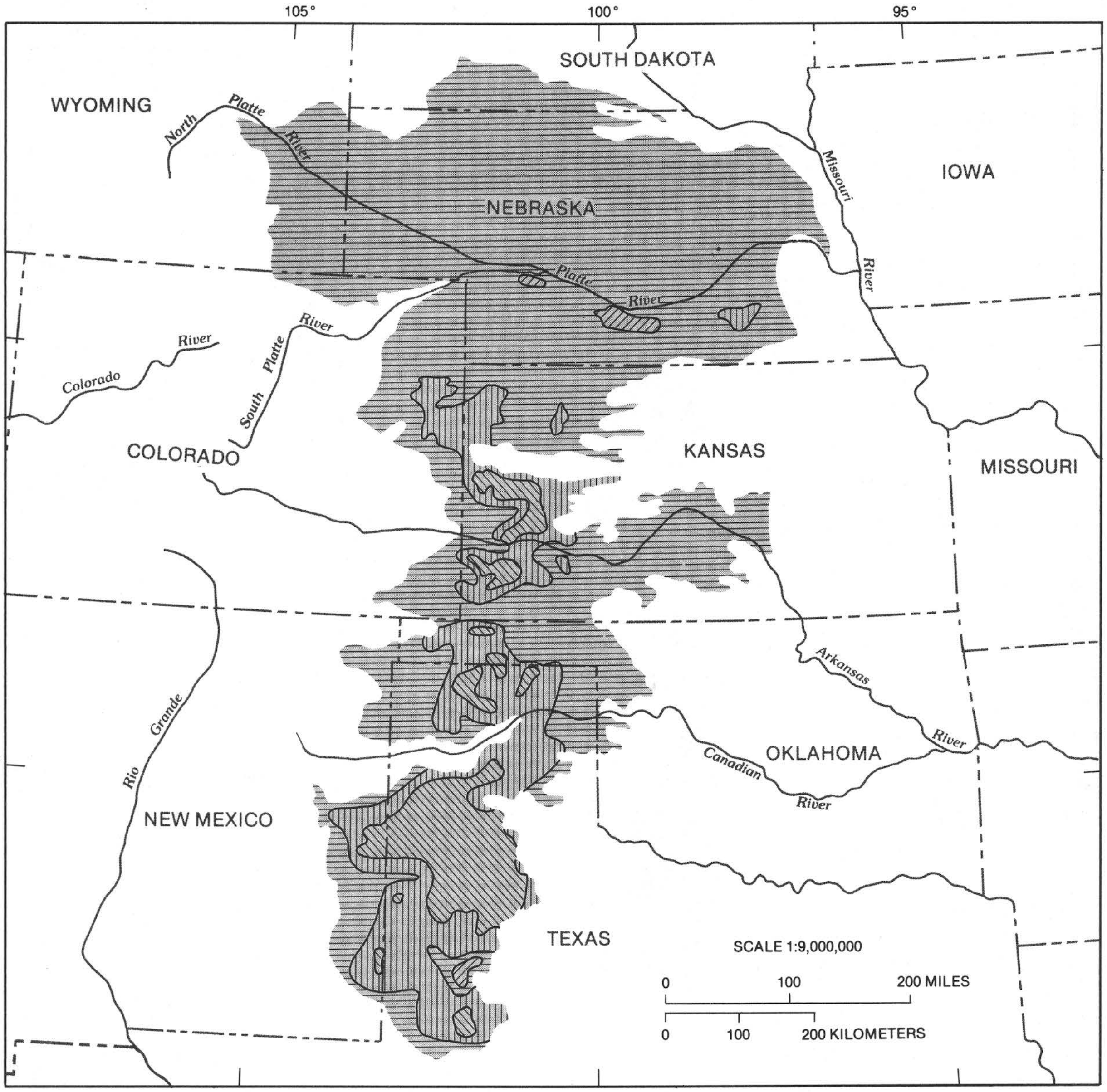
Significant (greater than 10 percent) increases in saturated thickness have occurred in only 1 percent of the aquifer area mostly in Nebraska (about 1,700 mi^2 or 1.1 million acres). Significant decreases (more than 10 percent) in saturated thickness have occurred in 25 percent of the area, (about 44,000 mi^2 or 28 million acres); saturated thickness has decreased more than 25 percent in 14,000 mi^2 (9 million acres). About 166 million acre-ft of water have been removed from aquifer storage since pumping began. About 16 percent of this decrease in aquifer storage has occurred in Kansas and 70 percent has occurred in Texas. No significant regional changes in saturated thickness of the High Plains aquifer system have occurred in South Dakota.

FLOW MODELS

The High Plains aquifer system was divided into three parts for simulation by digital models of ground-water flow as shown in figure 20. The southern High Plains subdivision includes most of the area (29,000 mi^2) south of approximately 35 degrees-latitude, the central High Plains subdivision includes the area (48,500 mi^2) between 35 and 39 degrees-latitude, and the northern High Plains subdivision includes the area (96,500 mi^2) north of 39 degrees-latitude. A narrow strip, approximately 12 miles wide, joins the southern and central High Plains subdivisions; another strip, approximately 32 miles wide, joins the central and northern High Plains subdivision (fig. 20).

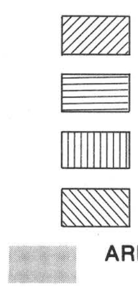
Model Development

The flow models used in this study can simulate two-dimensional ground-water flow in a heterogeneous, isotropic, unconfined aquifer. Two-, rather than three-, dimensional ground-water flow models were used because, on a regional scale, the vertical components of flow in the aquifer system were so small that they could be neglected. The models have a network of nodes spaced 10 miles apart in both the north-south and east-west directions. Four categories of informa-



EXPLANATION

SATURATED-THICKNESS CHANGES, IN PERCENT



Increase more than 10

10 to – 10

– 10 to – 25

Decrease more than – 25

AREA OF THE HIGH PLAINS REGIONAL AQUIFER SYSTEM

Figure 19.—Saturated thickness changes, from predevelopment to 1980, in the High Plains regional aquifer system.

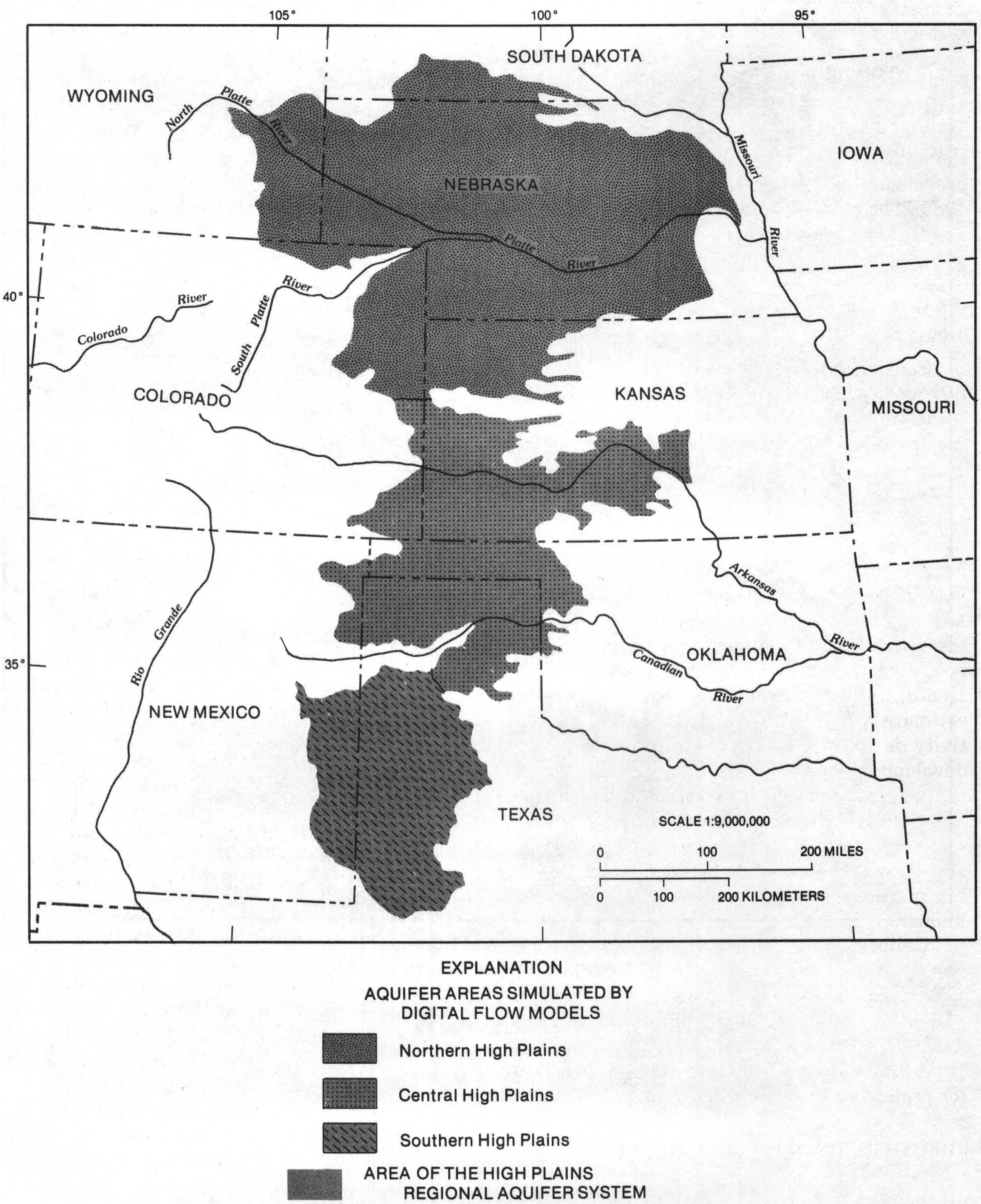


Figure 20.—Subdivisions used in modeling the High Plains regional aquifer system.

tion were needed for the flow models: (1) aquifer geometry (vertical and areal extent of the aquifer system); (2) boundary conditions; (3) aquifer parameters (hydraulic conductivity and specific yield); and (4) stresses on the aquifer (recharge and discharge). In addition, some simulations required an initial water-level configuration. All the data were stored in a data base designed to provide input to the model (Luckey and Ferrigno, 1982). The models integrate all of the input data and generate a water-level configuration and water budget for the aquifer system (Luckey and others, 1985).

Model Sensitivity

Sensitivity analyses were performed on the flow models to determine how changes in different model input parameters affect the simulation results (Luckey and others, 1985). The sensitivity analyses for the predevelopment-period models indicated that the models were about equally sensitive to changes in recharge and changes in hydraulic conductivity. The analyses further indicated that these two model inputs are highly interrelated and as long as an increase in one is accompanied by a similar increase in the other, the simulations will have a small mean residual between observed and calculated water levels. Hence, the models could be used to accurately estimate recharge only if the hydraulic conductivity is know accurately or vice versa. The development-period models are sensitive to changes in net pumpage and specific yield and insensitive to hydraulic conductivity. Specific yield and net pumpage are also highly interrelated and one cannot be accurately determined through a model analysis unless the other is accurately known.

Calibration of the predevelopment-period models provided the initial conditions for the calibration of the development-period models. Likewise, calibration of the development-period models for the southern, central, and northern High Plains can be used as the initial conditions for projecting future water levels.

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NORTHERN GREAT PLAINS REGIONAL AQUIFER-SYSTEM STUDY

By George A. Dinwiddie, Reston, Virginia, and Joe S. Downey, Denver, Colorado

INTRODUCTION

The study area of the Northern Great Plains regional aquifer system is about 250,000 mi² and includes North Dakota and parts of South Dakota, Montana, Wyoming, and Nebraska (fig. 21). It is bounded on the west by the central and northern Rocky Mountains, on the east by the Red River of the North, on the south by the central High Plains, and on the north by the United States-Canadian border. The Northern Great Plains mostly is underlain by sandstone, shale, and some evaporite deposits. The principal aquifers generally crop out along the flanks of the Williston and Powder River basins and along other major structural features (fig. 22).

Ground water generally flows northeastward across the study area. The source of recharge is precipitation in topographically high areas near the Black Hills uplift, Bighorn Mountains, the eastern flank of the northern Rocky Mountains, and other major structurally high areas. Ground water mostly discharges into topographically low areas of eastern North Dakota and South Dakota and occurs as outflow into Canada. Some ground-water discharge also occurs as diffuse upward leakage into overlying aquifers.

Flow characteristics vary significantly between the dominantly carbonate aquifers of Paleozoic age (such as the Madison Limestone) and the dominantly clastic aquifers of Mesozoic age (such as the Dakota Sandstone) and Cenozoic age. Potential for flow moving among aquifers exists near recharge and discharge areas and in the interior part of the study area, where hydraulic heads vary significantly between aquifers.

Development of ground-water supplies in the Northern Great Plains will be needed for the growth of energy resources, power generation, industrial expansion, increasing irrigation, and domestic and municipal water use. Streamflow has historically satisfied many of the water needs; however, surface water is fully appropriated in much of the area and is not always a dependable source because streamflows are extremely variable. Long-term, large-scale water needs require the development of ground water.

The Madison Limestone aquifer study, a federally funded program initiated in 1975, was designed to address the problems of water supplies associated with development of coal resources and the proposed coal-slurry pipelines in the Fort Union coal region (fig. 21). The Northern Great Plains regional aquifer system study, started in 1978 for the purpose of studying the aquifer system regionally underlying the Northern Great Plains; subsequently, the Madison Limestone aquifer study was merged into the Northern Great Plains regional aquifer system study which was completed in 1982. Significant results of the study are briefly discussed below.

GEOHYDROLOGY AND MODEL SIMULATION

The Northern Great Plains aquifer system is one of the largest confined systems in the United States, extending more than 600 miles from mountainous recharge areas in Montana, Wyoming, and South Dakota to discharge areas in eastern Dakotas and the Canadian Province of Manitoba. Five major aquifer systems were identified in the Northern Great Plains. In descending order, they are: the Upper Cretaceous aquifer system which mostly includes sediments of Late Cretaceous age but also includes the surficial glacial and Tertiary sediments; the Lower Cretaceous aquifer system; the Pennsylvanian

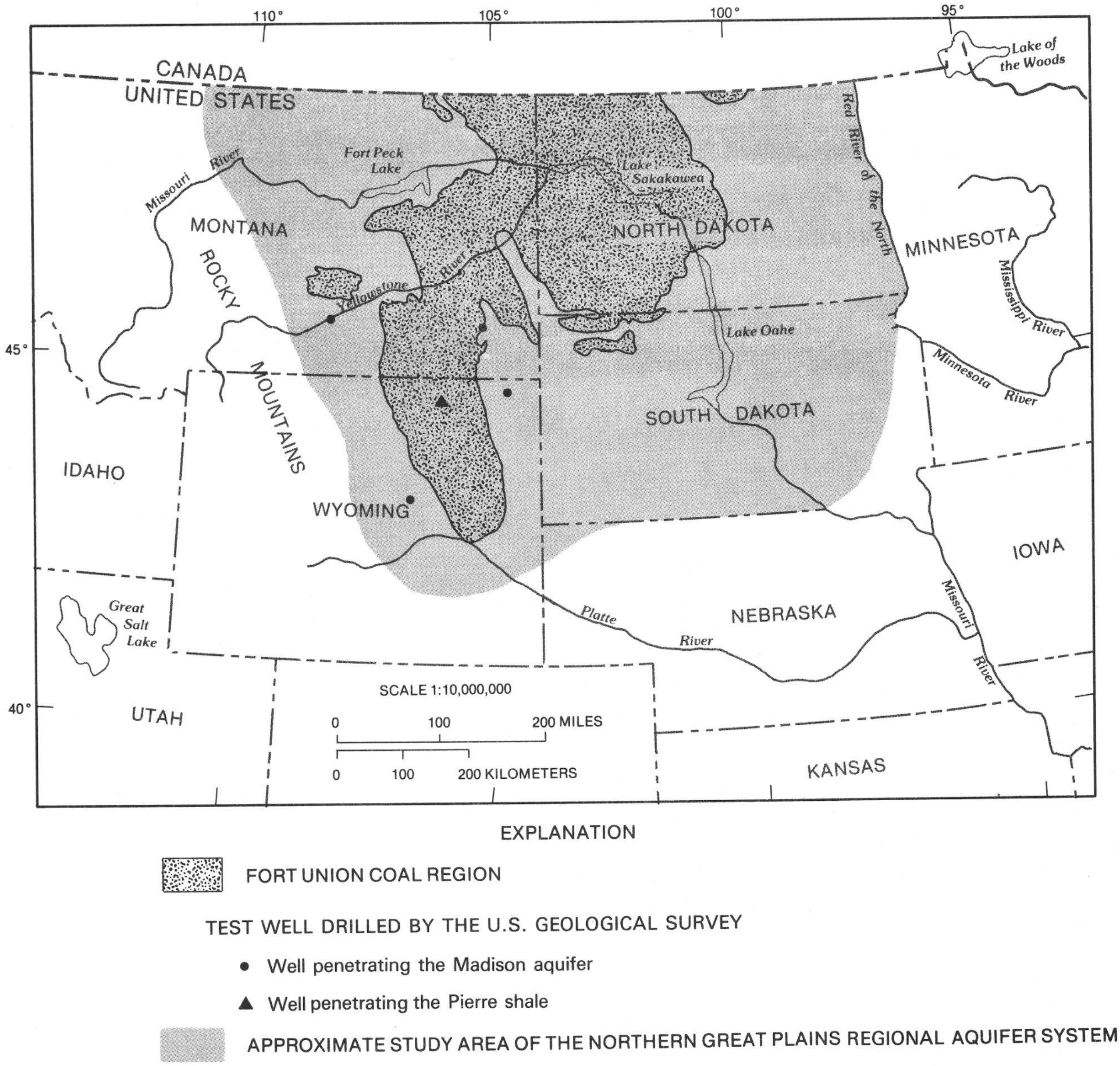
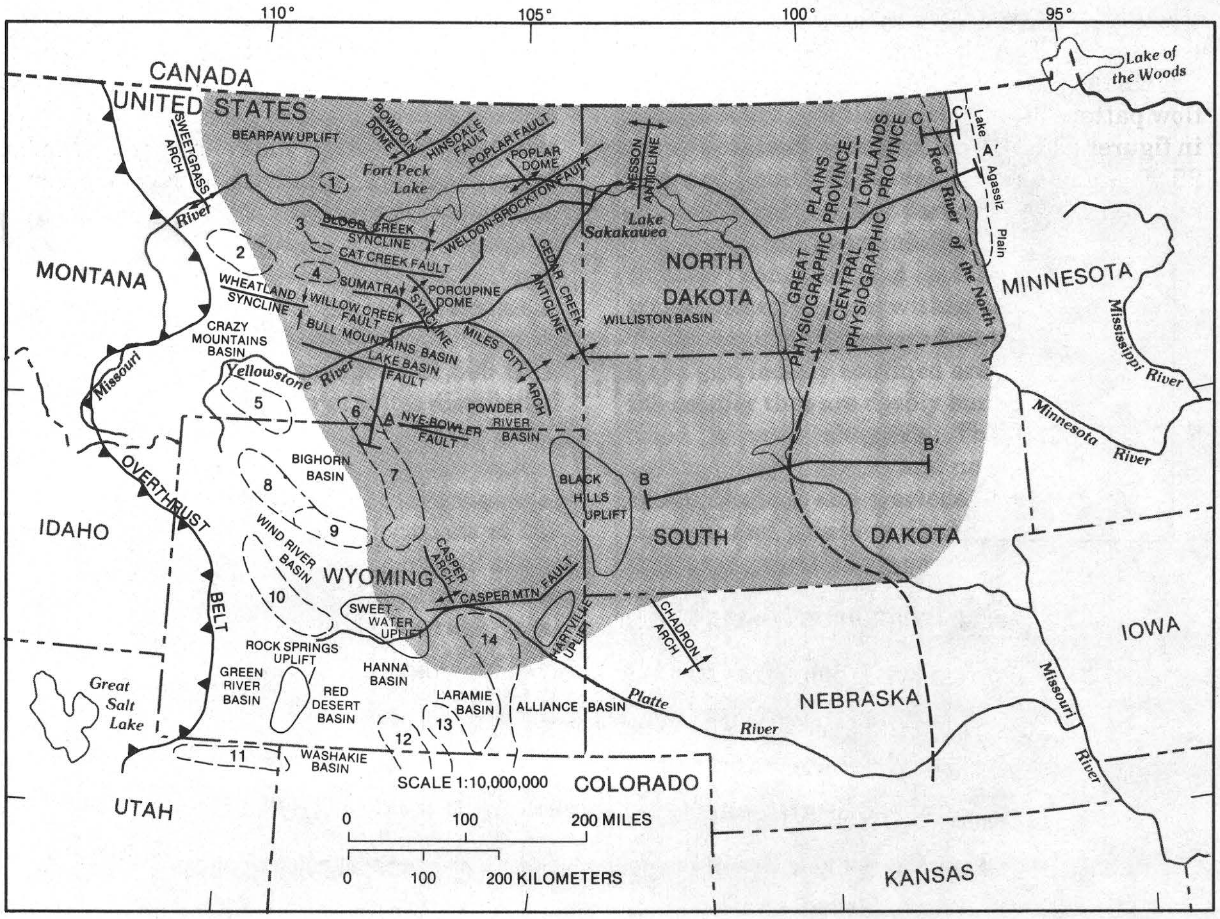


Figure 21.—Study area of the northern Great Plains regional aquifer system.

aquifer system; the Mississipian (Madison Limestone) aquifer system; and the Cambrian-Ordovician aquifer system. The relation between the aquifer systems is shown in figure 23. All aquifer systems crop out and receive recharge in the highlands in the western part of the study area. Recharge also occurs in outcrop areas around the Black Hills uplift in South Dakota. The major recharge area for the Mississippian aquifer system (Madison Limestone) in the Black Hills uplift area is a plateau on the west flank of the Black Hills uplift where the limestone contains many caves and sinkholes. The Wyoming State Engineer's Office (1974) states that the recharge in this plateau area (187,000 acres) is about 6.8 in./yr or about 146 ft^3/s. Virtually all streams lose a part of flow as they cross the aquifer-system outcrop areas (Swenson, 1968; Wyoming State Engineer's Office, 1974).

Streamflow measurements on several streams located at the east side of the Black Hills uplift indicate that as much as 10 ft^3/s of flow is lost from streams crossing the outcrop of the Mississippian (Madison Limestone) aquifer system (Swenson, 1968). Prior to sealing stream channels in 1937, about 100 ft^3/s of estimated streamflow loss was reported by Powell (1940). Based on similar aquifer lithology and degree of weathering, it is reasonable to assume that most streams at the western mountainous areas, such as the Bighorn Mountains, could lose nearly the same amount of flow, 100 ft³/s, as the streams cross the comparable outcrop areas of rocks of Paleozoic age.

Recharge enters the Lower Cretaceous aquifer system either by precipitation in the outcrop areas or by upward leakage from underlying aquifers. Miller and Rahn (1974) reported that about 0.8 in./yr of water was recharged in outcrop areas of Lower Cretaceous sandstone near the Black Hills uplift. Assuming an outcrop area for the Lower Cretaceous aquifer system of about 334 mi² (H. L. Case, U. S. Geological Survey, written commun., 1982), approximately 20 ft³/s of water could recharge the Lower Cretaceous aquifer system. Recharge from streamflow loss probably is small in the Black Hills uplift area. Brown (1944) noted that no loss in streamflow could be detected from gaged streams that cross the outcrop of the Cretaceous rocks, in contrast to water



EXPLANATION

MOUNTAIN RANGES OF THE EASTERN FLANK OF THE NORTHERN ROCKY MOUNTAINS

1	Little Rocky	6	Pryor	11	Uinta
2	Little Belt	7	Bighorn	12	Sierra Madre
3	Judith	8	Absaroka	13	Medicine Bow
4	Big Snowy	9	Owl Creek	14	Laramie
5	Beartooth	10	Wind River		

APPROXIMATE STUDY AREA OF THE NORTHERN GREAT PLAINS REGIONAL AQUIFER SYSTEM

Figure 22.—Structural and physiographic features of the northern Great Plains (modified from Peterson, 1981).

losses to the Paleozoic rocks; therefore, most of the recharge of the Cretaceous aquifer system probably is from the underlying Paleozoic rocks.

Although surface-water data indicate that large quantities of water enter the five aquifer systems along outcrop areas in the western highlands, not all this water recharges the deep, regional flow system and flows eastward to discharge areas. A large part of the recharged water discharges within short distances through springs and seeps along the flanks of the mountains (Swenson, 1968; Rahn and Gries, 1973; and Hodson, 1974) the remainder eventually enters the regional flow system.

On the basis of simulation, the conceptualized flow pattern of the five aquifer systems is shown in figures 24-30. Recharge rates shown in figures 25-29 are of the water that recharges the deep regional flow system. Water in the Upper Cretaceous aquifer discharges toward the Missouri River or upward to the overlying surficial aquifers (figs. 24). Discharge from the Lower Cretaceous aquifer system is mainly upward to overlying aquifers in eastern North Dakota and South Dakota and to the subcrop of the Lower Cretaceous aquifer system in the glacial Lake Agassiz basin of North Dakota (figs. 25 and 29). Discharge from aquifer systems consisting of rocks of Paleozoic age—the Pennsylvanian aquifer system, the Mississippian (Madison Limestone) aquifer system, and the Cambrian-Ordovician aquifer system—principally is upward to the overlying aquifers and to the Red River of the North. However, in eastern North Dakota, ground water in the Pennsylvanian aquifer system flows downward into the Cambrian-Ordovician aquifer system and then discharges to the Red River of the North or to land surface

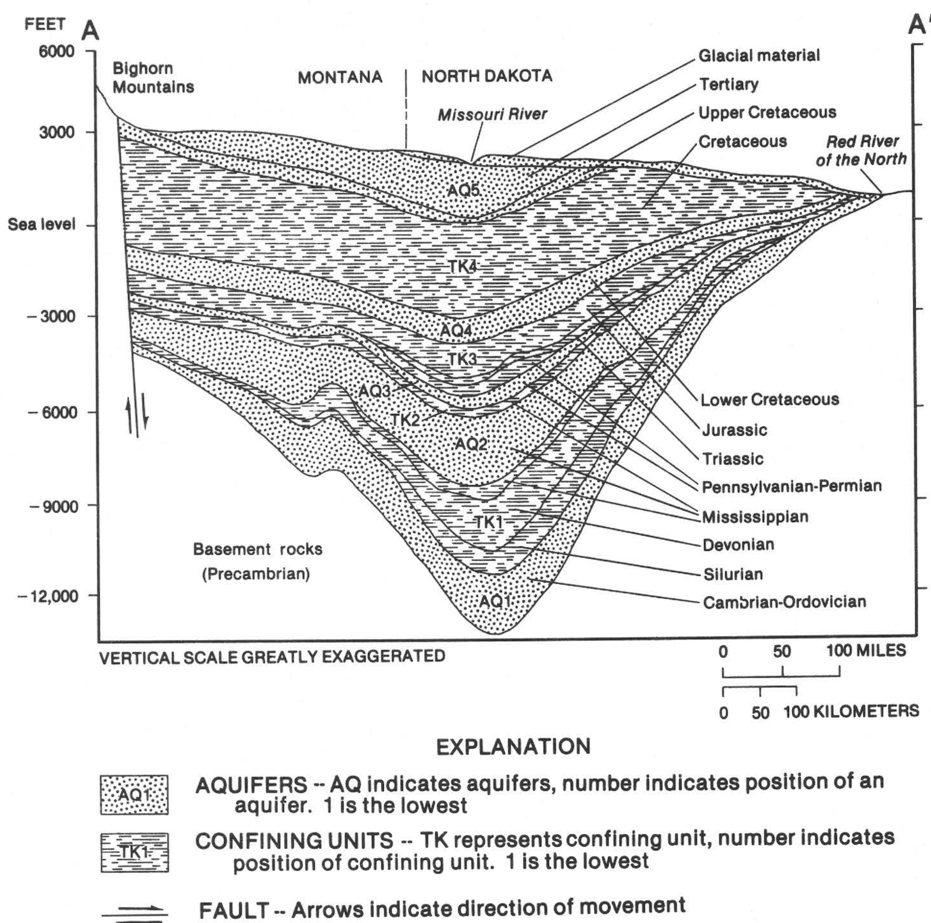
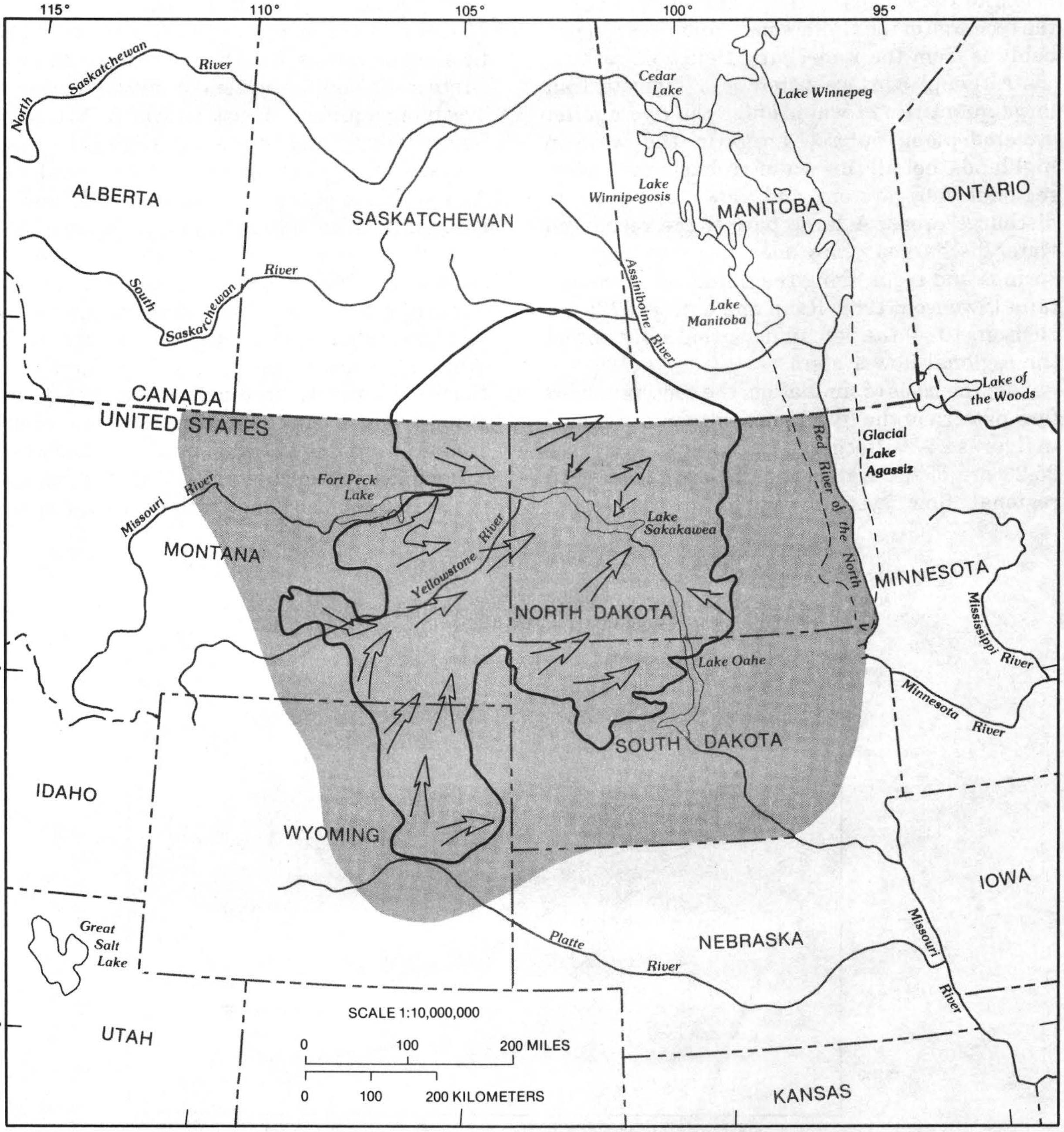


Figure 23.—Diagrammatic cross section from ground-water recharge area to discharge area showing relation of aquifers and confining units underlying the northern Great Plains, from south-central Montana to east North Dakota.



EXPLANATION

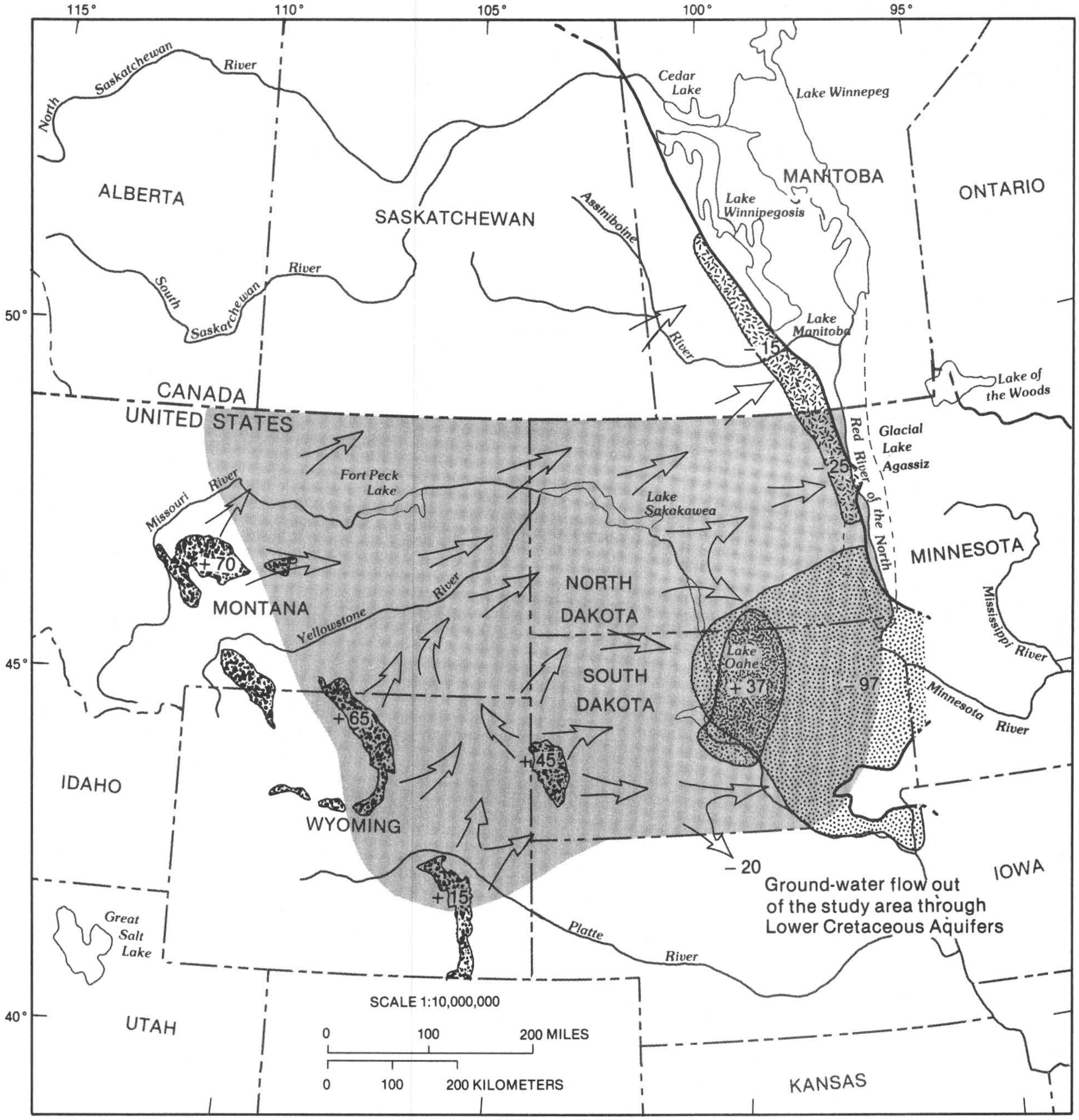
APPROXIMATE LIMIT OF UPPER CRETACEOUS ROCKS

GENERAL DIRECTION OF GROUND-WATER FLOW

APPROXIMATE STUDY AREA OF THE NORTHERN GREAT PLAINS REGIONAL AQUIFER SYSTEM

NOTE: Recharge to the aquifer is generally in outcrop areas and from shallow glacial deposits. Discharge from the aquifer is to rivers and streams, by evapotranspiration, and through wells

Figure 24.—Regional flow in the Upper Cretaceous aquifer system (AQ5) underlying the northern Great Plains.



EXPLANATION

RECHARGE AREA -- Generally areas of ground-water recharge. Numbers represent rate of recharge in cubic feet per second

+15 Recharge in highlands

+37 Recharge from underlying Mississippian and Pennsylvanian aquifers

APPROXIMATE LIMIT OF LOWER CRETACEOUS ROCKS

APPROXIMATE STUDY AREA OF THE NORTHERN GREAT PLAINS REGIONAL AQUIFER SYSTEM

DISCHARGE AREA -- Approximate areas of principal ground-water discharge. Numbers represent rate of discharge in cubic feet per second

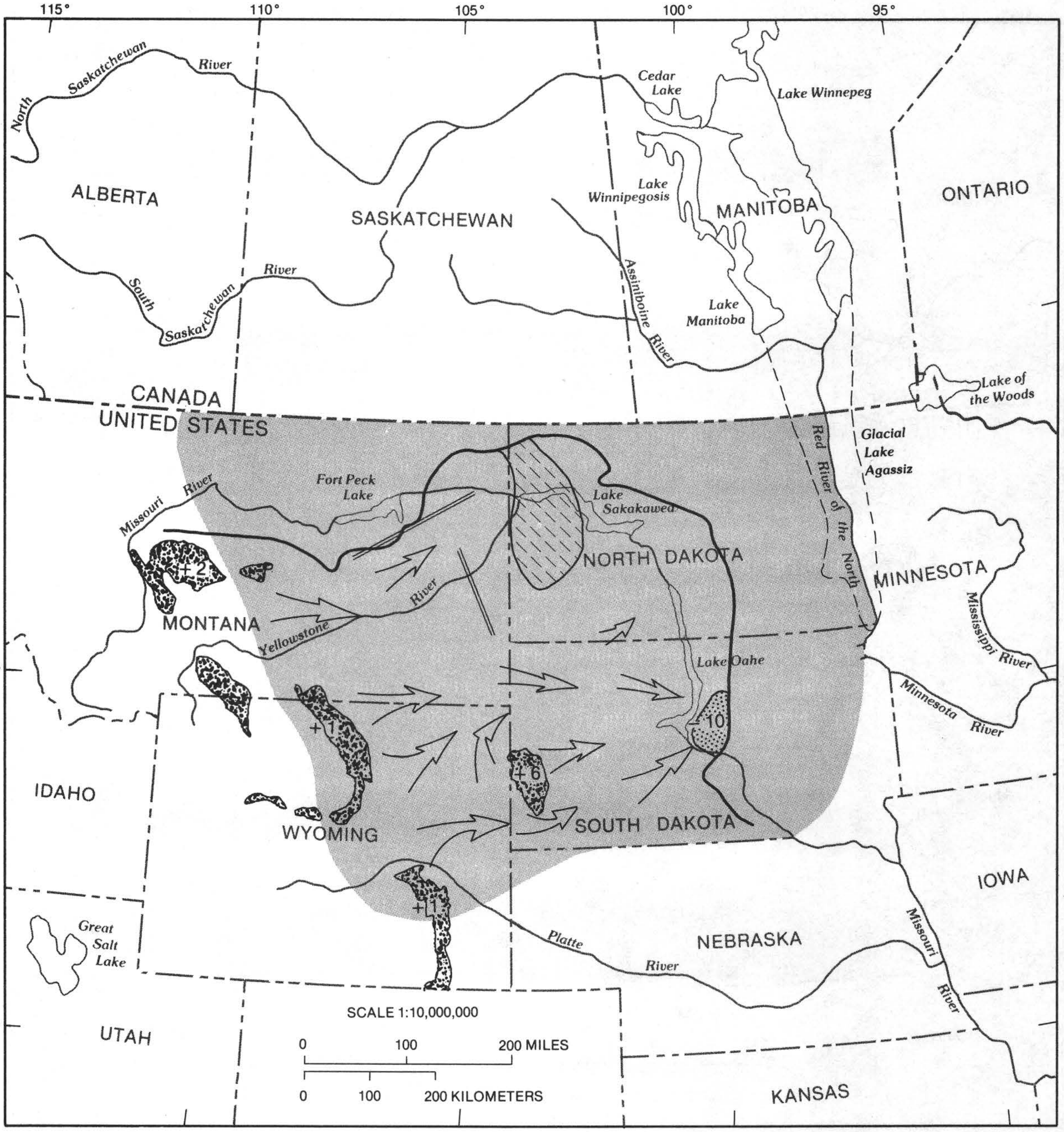
-25 Upward leakage to overlying glacial deposits

-97 Upward leakage to overlying Cretaceous aquifers and through wells

GENERAL DIRECTION OF GROUND-WATER FLOW

NOTE: Generally there is upward leakage from the Lower Cretaceous aquifer to overlying aquifers at a rate of approximately 75 cubic feet per second distributed evenly throughout the area

Figure 25.—Regional flow in the Lower Cretaceous aquifer system (AQ4) underlying the northern Great Plains.



EXPLANATION

RECHARGE AREA -- Generally areas of ground-water recharge. Numbers represent rate of recharge in cubic feet per second

+15 Recharge in highlands

APPROXIMATE STUDY AREA OF THE NORTHERN GREAT PLAINS REGIONAL AQUIFER SYSTEM

GENERAL DIRECTION OF GROUND-WATER FLOW

APPROXIMATE LIMIT OF PENNSYLVANIAN AND PERMIAN ROCKS

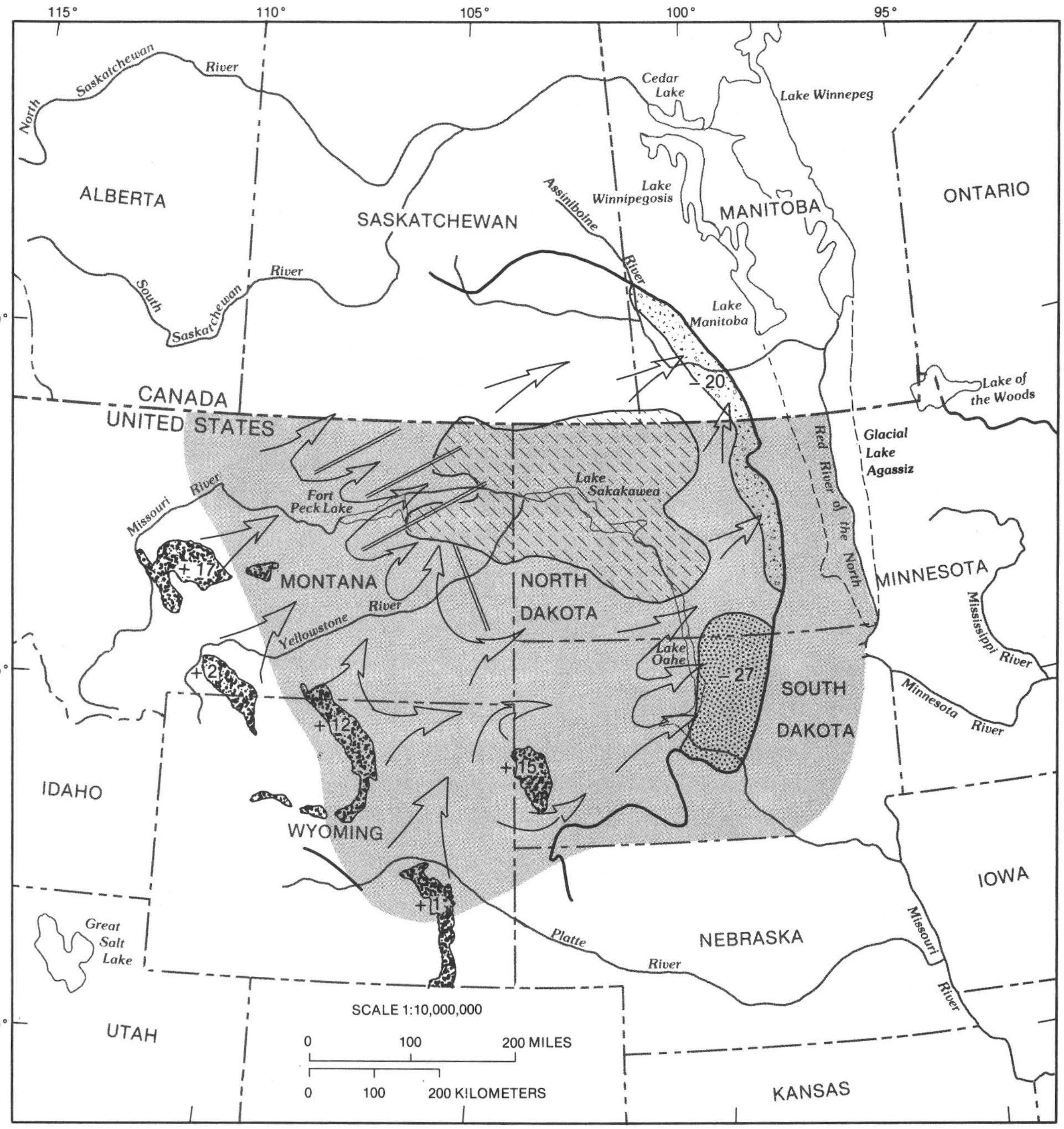
DISCHARGE AREA -- Approximate areas of principal ground-water discharge. Numbers represent rate of discharge in cubic feet per second

−10 Discharge to Lower Cretaceous aquifers

GEOLOGIC STRUCTURE -- Locations of paleostructures that may affect the flow of ground water

BRINE AREA -- Areas within which the total dissolved solids in ground water is greater than 100,000 milligrams per liter

Figure 26.—Regional flow in the Pennsylvanian aquifer system (AQ3) underlying the northern Great Plains.



EXPLANATION

RECHARGE AREA -- Generally areas of ground-water recharge. Numbers represent rate of recharge in cubic feet per second

+15 Recharge in highlands

BRINE AREA -- Areas within which the total dissolved solids in ground water is greater than 100,000 milligrams per liter

GENERAL DIRECTION OF GROUND-WATER FLOW

APPROXIMATE STUDY AREA OF THE NORTHERN GREAT PLAINS REGIONAL AQUIFER SYSTEM

DISCHARGE AREA -- Approximate areas of principal ground-water discharge. Numbers represent rate of discharge in cubic feet per second

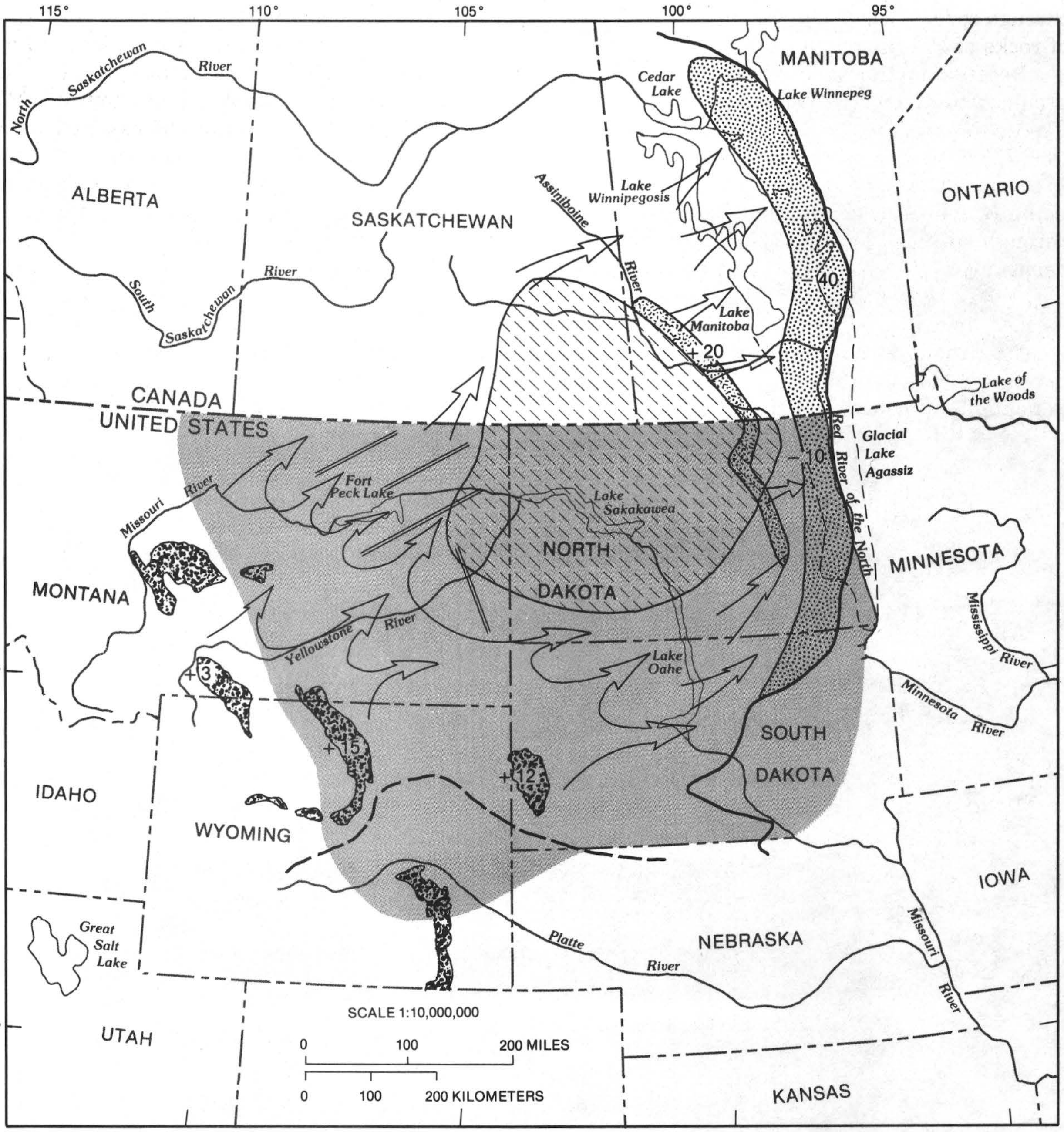
-27 Discharge to Lower Cretaceous aquifers

-20 Discharge to underlying Cambrian and Ordovician aquifers

GEOLOGIC STRUCTURE -- Locations of paleostructures that may affect the flow of ground water

APPROXIMATE LIMIT OF MADISON LIMESTONE

Figure 27.—Regional flow in the Mississippian (Madison Limestone) aquifer system (AQ2) underlying the northern Great Plains.



EXPLANATION

RECHARGE AREA -- Generally areas of ground-water recharge. Numbers represent rate of recharge in cubic feet per second

+15 Recharge in highlands

+20 Recharge from aquifers overlying Mississippian

BRINE AREA -- Areas within which the total dissolved solids in ground water is greater than 100,000 milligrams per liter

APPROXIMATE STUDY AREA OF THE NORTHERN GREAT PLAINS REGIONAL AQUIFER SYSTEM

DISCHARGE AREA -- Approximate areas of principal ground-water discharge. Numbers represent rate of discharge in cubic feet per second

-27 Discharge to land surface

GEOLOGIC STRUCTURE -- Locations of paleostructures that may affect the flow of ground-water

GENERAL DIRECTION OF GROUND WATER FLOW

APPROXIMATE LIMIT OF CAMBRIAN-ORDOVICIAN ROCKS

APPROXIMATE LIMIT OF ORDOVICIAN ROCKS

Figure 28.—Regional flow in the Cambrian-Ordovician aquifer system (AQ1) underlying the northern Great Plains.

through springs where aquifer systems consisting of rocks of Paleozoic age crop out.

Because of the slow rate of movement of ground water indicated by model simulation (figs. 31 and 32) and the short length of time (12,000 to 14,000 years) since ice covered southern Manitoba of Canada, it is possible that the water being discharged from the Paleozoic rocks through springs is a mixture of brine from the deeper part of the aquifer systems and glacial meltwater that recharged the aquifer systems while it was covered by glacial ice. The range in concentrations of dissolved solids indicates that the water may be a mixture of three flow components: (1) flow from the brine area, (2) water recharged from the Pleistocene glacial ice, and (3) freshwater flowing from the south along a flow path from the Black Hills uplift (fig. 22). Van Everdingen (1968) estimated that about 0.1 ft^3/s of water having concentrations of dissolved solids ranging from 29,000 to 63,000 milligrams per liter (mg/l) was discharging from springs along the outcrop of the Paleozoic rocks. The dominant ions are sodium and chloride.

A body of brine is located at the eastern flank of the Williston basin (figs. 22, 26-28 and 33) where movement of ground water is sluggish (figs. 31 and 32). Three hypotheses concerning movement of ground water and the brine are:

(1) The brine is static.

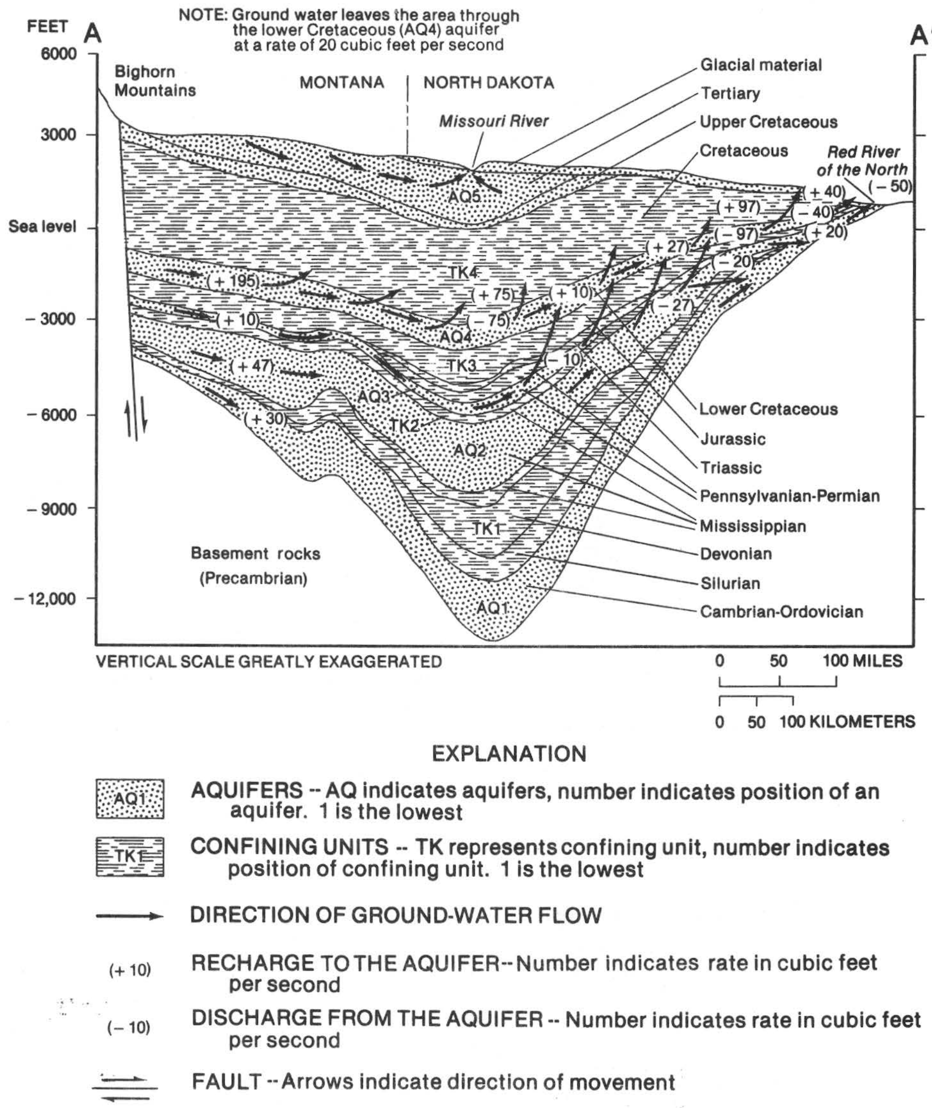


Figure 29.—Generalized hydrogeologic cross section showing general ground-water flow patterns in the northern Great Plains regional aquifer system.

(2) The brine is a very slow-moving segment of the regional ground-water flow.
(3) The brine is moving as part of the adjustment of the flow system to the melting of the Pleistocene ice.

The first hypothesis is similar to that described by Hubbert (1969) in which freshwater flowing through a synclinal structure comes into contact with static, dense brine along a sharp fluid interface. Hubbert shows that a body of saline water under these conditions does not lie uniformly in the deepest part of the structure but, rather, is displaced upward along the base of the outflow flank. That is, while the inflow flank is occupied by moving freshwater, the outflow flank contains static brine in the lower part of the

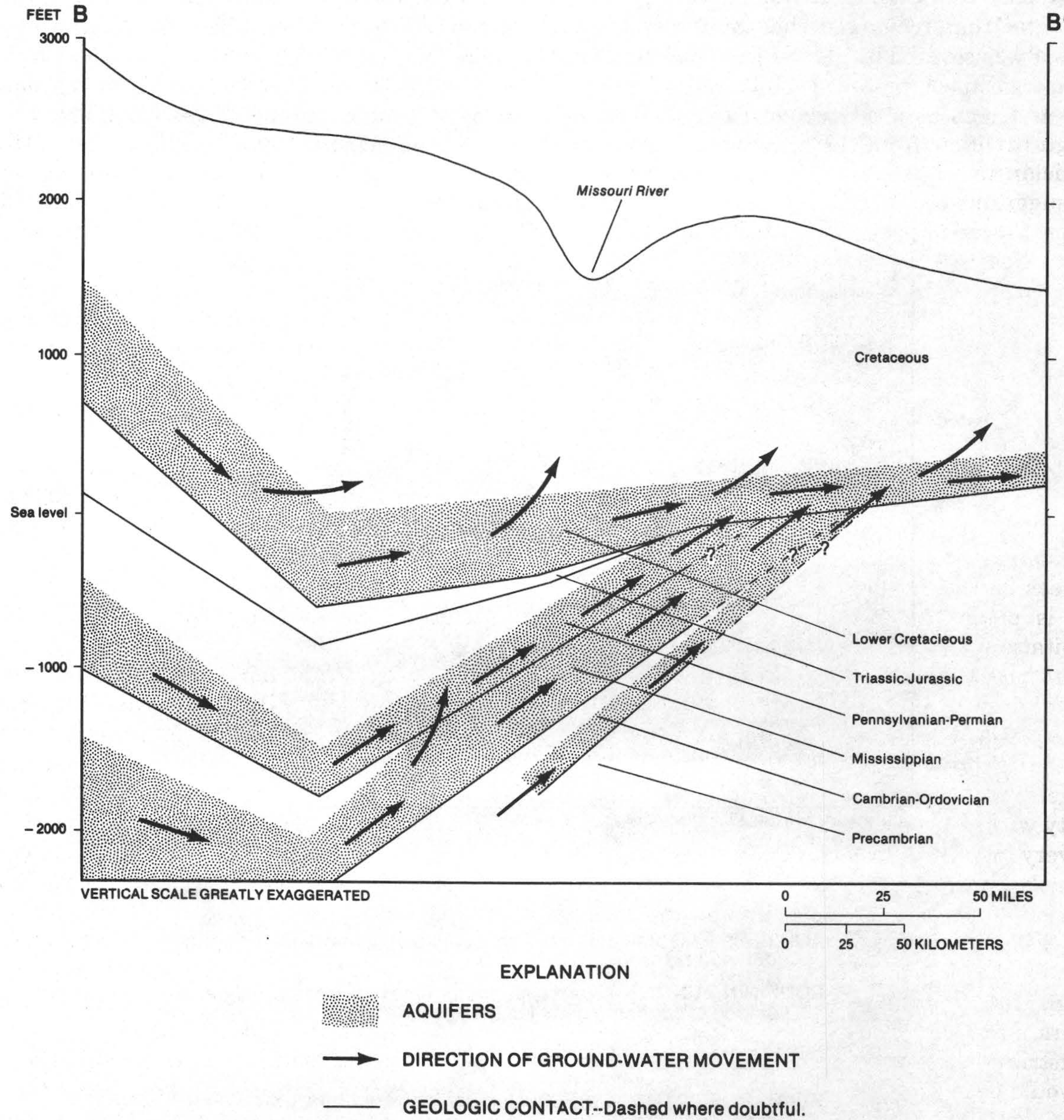


Figure 30.—Generalized hydrogeologic cross section showing ground-water discharge from aquifers consisting of rocks of Paleozoic to Mesozoic age underlying the northern Great Plains (modified from Swenson, 1968). Line of section shown in figure 22.

aquifer and moving freshwater above. The flow of freshwater in the Williston basin is around the dense brine area as well as above, reflecting that the structure is actually a basin rather than a simple syncline. The flow above the brine apparently is by upward leakage to the overlying aquifers rather than within the upper part of the Mississippian aquifer system. However, simulations of the flow system containing variable-density fluids indicate that the brine in the Mississippian aquifer system is not completely static as assumed in Hubbert's hypothesis.

Simulation shows that consistent flow with low velocity flows eastward and northeastward, through the brine body, in both the Cambrian-Ordovician and Mississippian aquifer systems. This suggests that a small component of the regional flow actually moves directly across the Williston basin from west to east, through the areas containing the brine, thus supporting the second hypothesis. This interpretation probably explains the origin of the brine, which can be attributed to dissolution of salt from beds of halite as ground water flows through the Williston basin, as described by Grossman (1968). The process of dissolution of halite is enhanced by increasing water temperature with depth. Water having higher salinity usually is found in regions of high temperature. Reduction in salinity in updip areas on the eastern flank of the Williston basin is presumably due at least in part to precipitation of halite associated with cooler temperature, although dilution by freshwater undoubtedly also is a factor. Precipitation of halite reduces permeability gradually over a long period of time and tends to decrease the ground-water flow velocity. Although some flow exists, flow velocity within the Williston brine area appears to be very low and nearly static. Flow velocity is extremely low relative to those elsewhere in the aquifer system, and most of the freshwater flow seems to be deflected around the body of the brine to the north or south or discharges upward through the confining units into overlying aquifers. Thus, both the hydraulics and the density distribution seem to be fairly close to those that would be expected in a system containing a body of static brine.

Concerning the third hypothesis regarding brine movement, the brine would have been set in motion at the end of the Pleistocene glaciation, seeking a new equilibrium with recharge and discharge. This readjustment could still be in progress, thus contributing to the apparent movement of the brine. However, flow velocity computed from simulation suggests that a simple flow across the Williston basin is more likely than the adjustment of the brine body to hydrologic changes of the Pleistocene glaciation.

In summary, the second hypothesis, that the brine represents a sluggish segment of the regional ground-water flow across the Williston basin, seems to agree best with simulations and field data. The origin of the brine appears to have been the dissolution of halite. As the density of the brine increases, freshwater is diverted around the brine to the north and south or discharges upward into overlying aquifers. Although the second hypothesis seems reasonable, the other hypotheses probably also can explain the field situation to some degree. The configuration of the brine on the outflow side of the Williston basin, and the generally low velocity, approximate the static conditions of the brine described by Hubbert (1969), even though the brine is not thought to be completely static. Distribution of water having high salinity may shift slightly in response to hydrologic changes at the end of the Pleistocene glaciation.

Saline water is found also in other parts of the study area—for example, in deeper parts of the Powder River basin—although not at the concentrations encountered in the Williston basin. However, similar processes presumably control the distribution and movement of the saline water.

Because the highland recharge areas were not covered by major ice during the Pleistocene (ice was generally limited to the area north and east of the Missouri River), recharge continued to be available to the underlying aquifers. Therefore, dissolution of halite along the western edge of the Williston basin probably existed during the glaciation period. Because of the short geologic time periods and low flow velocity, the brine resulting from dissolution of halite probably would not leave the hydrologic system but would tend to remain in the same general location shown in figures 27 and 28.

It has been noted that recharge and discharge may have been significantly different during the Pleistocene glaciation. Such hydrologic variation may affect the regional ground-water flow pattern. Evidence of the presence of water at the base of the continental ice has been discussed by Robin (1955), McGinnis (1968), Weertman (1972), and Downey (1983). The water is the result of melting at the base of the ice because of geothermal and

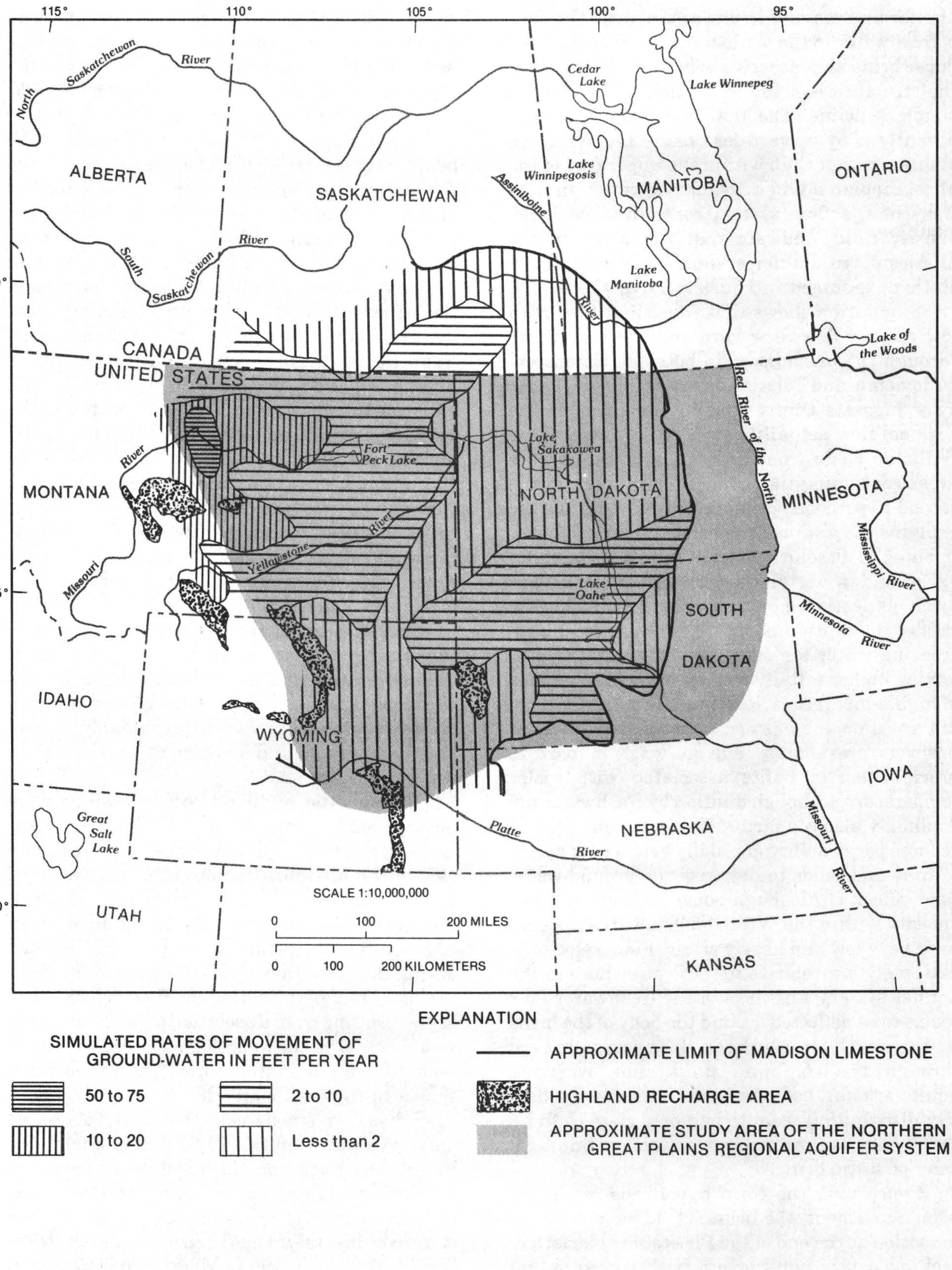


Figure 31.—Simulated rates of movement of ground water in the Mississippian (Madison Limestone) aquifer system (AQ2) underlying the northern Great Plains.

frictional heat. McGinnis (1968) estimated that the heat available to a temperate ice sheet from the geothermal and frictional heat could produce about 0.32 ft^3/yr of meltwater per square foot of ice surface. During the late Pleistocene, the continental ice covered an area of about 121,500 mi^2 in the Northern Great Plains. Based on 0.32 ft^3/yr per square foot of ice surface, about 7 mi^3/yr of water were available to recharge the underlying aquifers. The water would be under significant hydrostatic pressure from the weight of the overlying ice.

Widespread recharge to aquifers must have existed in the areas covered by the continental ice. The configuration of ice over the Northern Great Plains varied during the Pleistocene (Flint, 1971), and the resulting distribution of recharge similarly would also have varied. Highlands to the west of the study area that form the present recharge areas would presumably also be the recharge areas prior to glaciation, as well as during glacial and interglacial periods of the Pleistocene. However, during periods of strong glacial advance, recharge would also occur over the northern and eastern parts of the study area. The regional ground-water flow under these conditions must have been to the south toward the only available discharge areas. With successive retreats and advances of the continental ice, directions of regional ground-water flow probably varied accordingly, generating a complex history of ground-water movement. However, dissolution of halite in the deeper parts of the Williston basin must have occurred throughout the flow history, and velocity of ground-water movement in the areas of brine accumulation must have been low regardless of the variation of the flow directions. Hypothetical calculation shows that repeated changes in flow directions would probably have no major effects on the location of the high-salinity water. In terms of the deep brine, the glacial changes probably caused some minor transient disturbances but no major redistributions of the brine.

In summary, the flow pattern prior to glaciation was presumably similar to that at present, sustained by recharge in highlands to the west. Glaciation produced repeated variation of flow directions, but, in general, the glaciation did not cause major changes in distribution of the brine.

The Cambrian-Ordovician aquifer system, like the Mississippian (Madison Limestone) aquifer system, contains dense brine on the eastern flank of the Williston basin (fig. 27) and has freshwater flowing around the brine to the northeast. The Cambrian-Ordovician aquifer system apparently discharges partly to a number of saline lakes in eastern North Dakota. The component of flow to the north and east of the Williston basin is, accordingly, larger than in the Mississippian (Madison Limestone) aquifer system. The hypothesis of discharge to saline lakes is supported by several factors. The lakes lie in the eastern discharge area of the Cambrian-Ordovician aquifer system and are associated with depressions that overlie deposits of fine sand and gravel (fig. 34). The origin of these lake depressions has been attributed to artesian water discharging from deep aquifer systems (Laird, 1944). A supporting hypothesis (Downey, 1969, p. 12) suggests that, during Pleistocene glaciation, meltwater from the base of the ice sheet (Gow and others 1968; McGinnis, 1968) was forced into the underlying Mesozoic-Paleozoic aquifer systems by hydrostatic pressure. Upon deglaciation, the hydrostatic pressure was removed, allowing large quantities of water to move rapidly out of the systems. This rapid movement of water through overlying sediments resulted in erosion of the overlying sediments, forming the depressions in which the lakes exist today.

Geologic evidence and water chemistry suggest that these lakes now function as drains for the regional ground-water flow system. Test drilling indicates that thick deposits of glacial sand and gravel underlie the depressions and are hydraulically connected with the underlying Paleozoic aquifer systems (Downey, 1973). Chemical analyses of water samples taken from the test holes and lakes (Downey, 1971) indicate a chemical similarity to the water taken from the Cambrian-Ordovician aquifer system. These analyses indicate that the water is able to move upward from the Paleozoic aquifer systems through the glacial sand and gravel deposits to discharge points at bottoms of the lakes.

Vertical leakage through confining units are major contributors to ground-water discharge. Leakage occurs both through the confining units and along fractures associated with lineaments in the confining units (Weimer and others, 1982). Confining units are not present everywhere, having been removed by erosion or absent because of nondeposition. In those areas where a confining unit is absent, such as in eastern South Dakota, good hydraulic connection exists between aquifer systems, and one aquifer system may discharge to an adjoining aquifer system at a rate

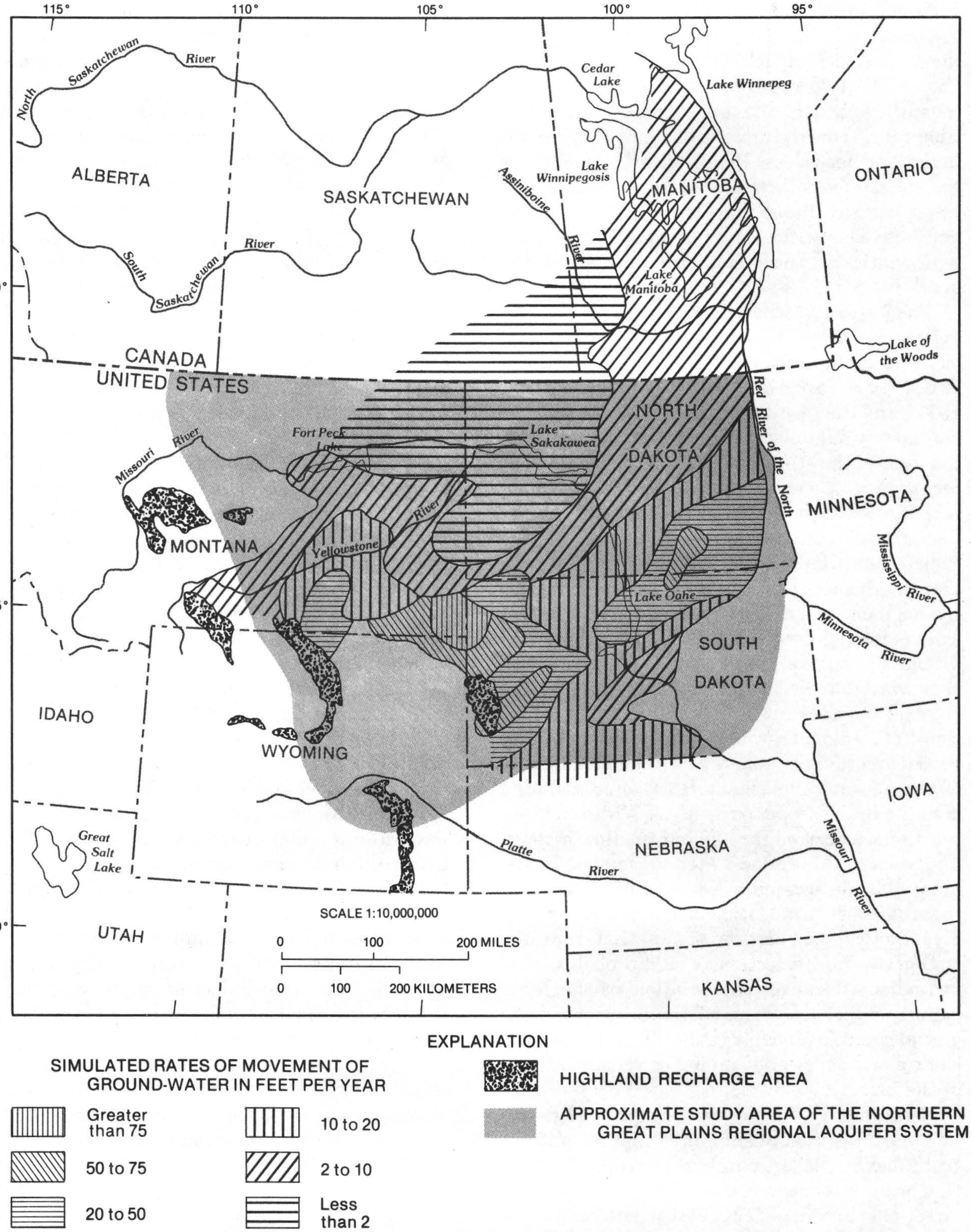


Figure 32.—Simulated rates of movement of ground water in the Cambrian-Ordovician aquifer system (AQ1) underlying the northern Great Plains.

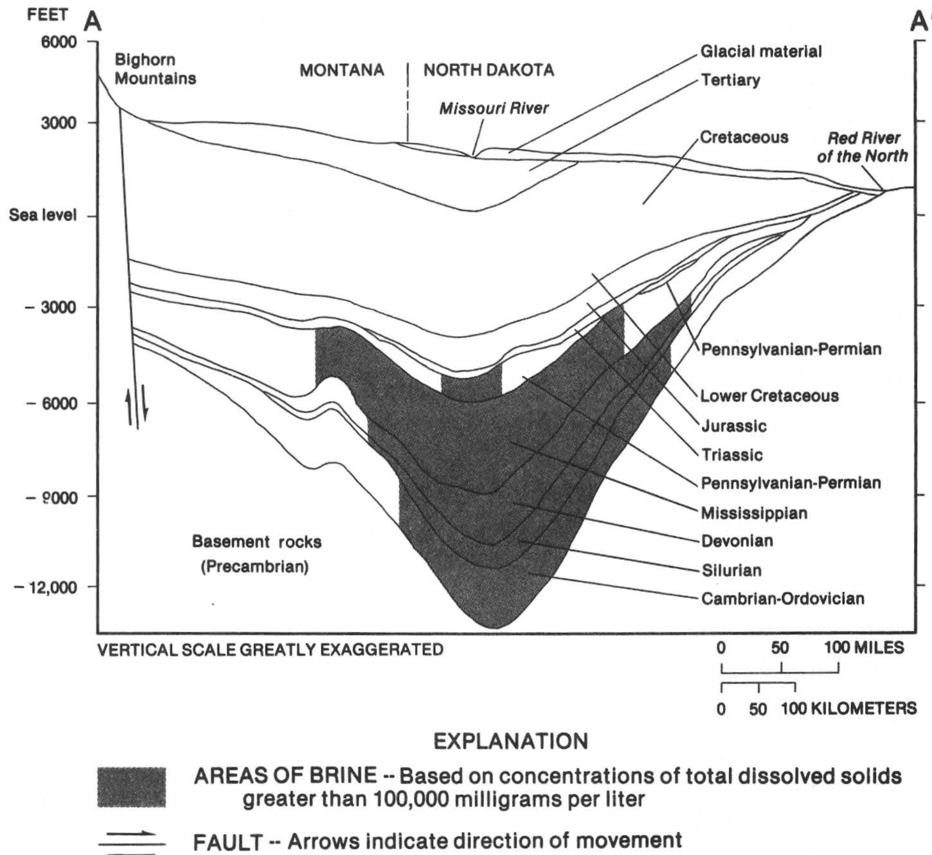


Figure 33.—Diagrammatic cross section showing general altitudes of brine in the Williston basin of the northern Great Plains. Line of section, shown in figure 22.

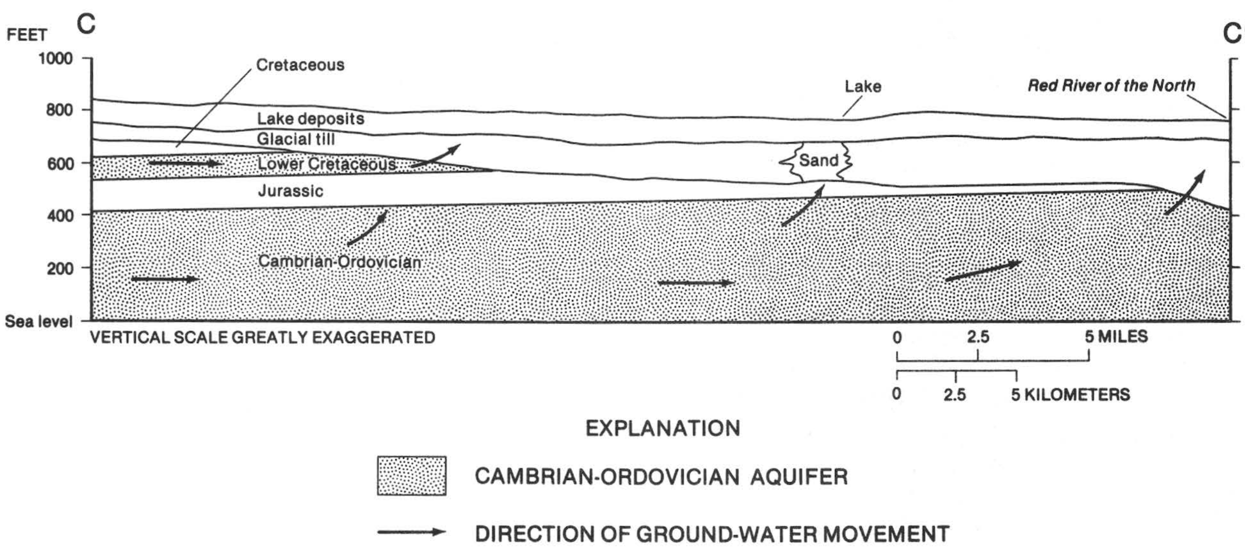


Figure 34.—Generalized hydrogeologic cross section showing discharge area of the Cambrian-Ordovician aquifer system (AQ1) underlying the northern Great Plains (cross section modified from Downey, 1978). Line of section, shown in figure 22.

that is dependent on head differences between the two aquifers and their vertical hydraulic conductivities. Where a confining unit is thick and unfractured, leakage through the confining unit will be low. Leakage along fractures is dependent on the degree of fracturing, cross-sectional area of the fractures, and interconnection among fractures.

Geochemical facies maps for the Cambrian-Ordovician, Mississippian, Pennsylvanian, and Lower Cretaceous aquifer systems have been used to indicate areas where upward leakage is occurring. Similar water types at the same location in adjoining aquifers suggest that water is able to move between the two aquifers through confining units. This type of geochemical data is of value in adjusting vertical leakage of the confining units in simulation.

Except where they are fractured, beds of halite are considered to be impermeable. Geochemical evidence (J. F. Busby, U. S. Geological Survey, written commun., 1982) indicates that there is extensive leakage between the Cambrian-Ordovician, Mississippian, and Pennsylvanian aquifer systems and that low vertical permeability of Triassic and Jurassic formations limits the upward leakage between the Paleozoic aquifer systems and the Mesozoic aquifer systems. This suggests that extensive development of the Paleozoic aquifer systems probably would not affect the Mesozoic aquifer systems in most of the study area within a reasonable time frame (40 years).

Future development of the regional aquifer system in the Northern Great Plains should take into account that part of the water withdrawn from wells may come from storage in the confining unit except where a confining unit is absent or highly fractured. The quality of water from the confining units may be entirely different from the quality of water from the aquifer systems.

Leakage between the Mississippian aquifer system and the Lower Cretaceous aquifer system in eastern South Dakota has been noted in several studies (Swenson, 1968) and was the basis for Swenson's theory of recharge to the artesian basin of the Dakotas. He proposed that water enters the Madison Limestone in the Black Hills uplift area, moves generally eastward across approximately two-thirds of the State of South Dakota, and is discharged vertically through confining units to the Lower Cretaceous aquifer system.

In the area of high upward leakage from the underlying aquifers to the Lower Cretaceous aquifer system, the confining unit between the underlying aquifer system and the Lower Cretaceous aquifer system is thin or absent. Geochemical facies maps indicate that water from the Pennsylvanian aquifer system is similar to water in the underlying Mississippian aquifer system and that vertical leakage is occurring between these two aquifer systems. Swenson (1968) indicates that the chemistry of the water from the Lower Cretaceous (Dakota) aquifer system is also similar to waters in the Mississippian and Pennsylvanian aquifer systems in the area of high upward leakage.

Simulation indicates that the theory advanced by Swenson (1968) concerning the ground-water flow of the Lower Cretaceous aquifer system in South Dakota is basically correct. Additional geochemical data (K. D. Peter, U.S. Geological Survey, written commun., 1982) also support the conclusion that water is moving upward from the Mississippian and Pennsylvanian aquifer systems into the Lower Cretaceous aquifer system in eastern South Dakota and North Dakota.

Geologic structure appears to be an important control (Weimer and others, 1982) on the rate and direction of ground-water movement in the Northern Great Plains. For example, the Casper Mountain fault (fig. 22) appears to prevent ground-water flow to the south from the Powder River basin, and the major fault system bounding the Bighorn Mountains on the east limits recharge to the Powder River basin. Recharge from the Bighorn Mountains appears to be channeled by geologic structures associated with major zones of lineaments and moves northeastward (figs. 26-28) across the northern part of the Powder River basin, south of the Cedar Creek anticline, to join the flow system recharged from the Black Hills uplift area. This flow system continues around the southern part of the Williston basin northeastward to the discharge area in northeastern North Dakota and eastern Manitoba of Canada. The Weldon-Brockton fault zone (fig. 22) appears to act as a major channel for ground water movement (figs. 27-28) from the Big Snowy Mountains and associated highlands in Montana to discharge areas in Canada north of the Williston basin.

Simulated drawdowns in selected aquifers after a hypothetical pumping for 5.9 years at a rate of 27.9 ft^3/s (18 Mgal/d) from the Mississippian aquifer system (Madison Limestone) with an assumed uniform storage coefficient of 2.0 x 10^{-6} indicate the degree of hydrologic connection among the aquifer systems. The pumping of the Mississippian aquifer (Madison Limestone) system results in large drawdowns in the overlying Pennsylvanian aquifer system and much larger drawdowns in the underlying Cambrian-Ordovician aquifer system, (figs. 35-36).

The digital model developed for the Northern Great Plains regional aquifer system may be used to evaluate the regional effects of planned development schemes (Downey and Paulson, 1974). However, the accuracy of evaluations depends on the accuracy of the data used in the model. During the Northern Great Plains regional aquifer system study, the accuracy of hydraulic-head data varied greatly from one area to another; therefore, the error range of the simulations is quite large. The other aquifer parameters such as hydraulic conductivity and storage coefficient were all estimated.

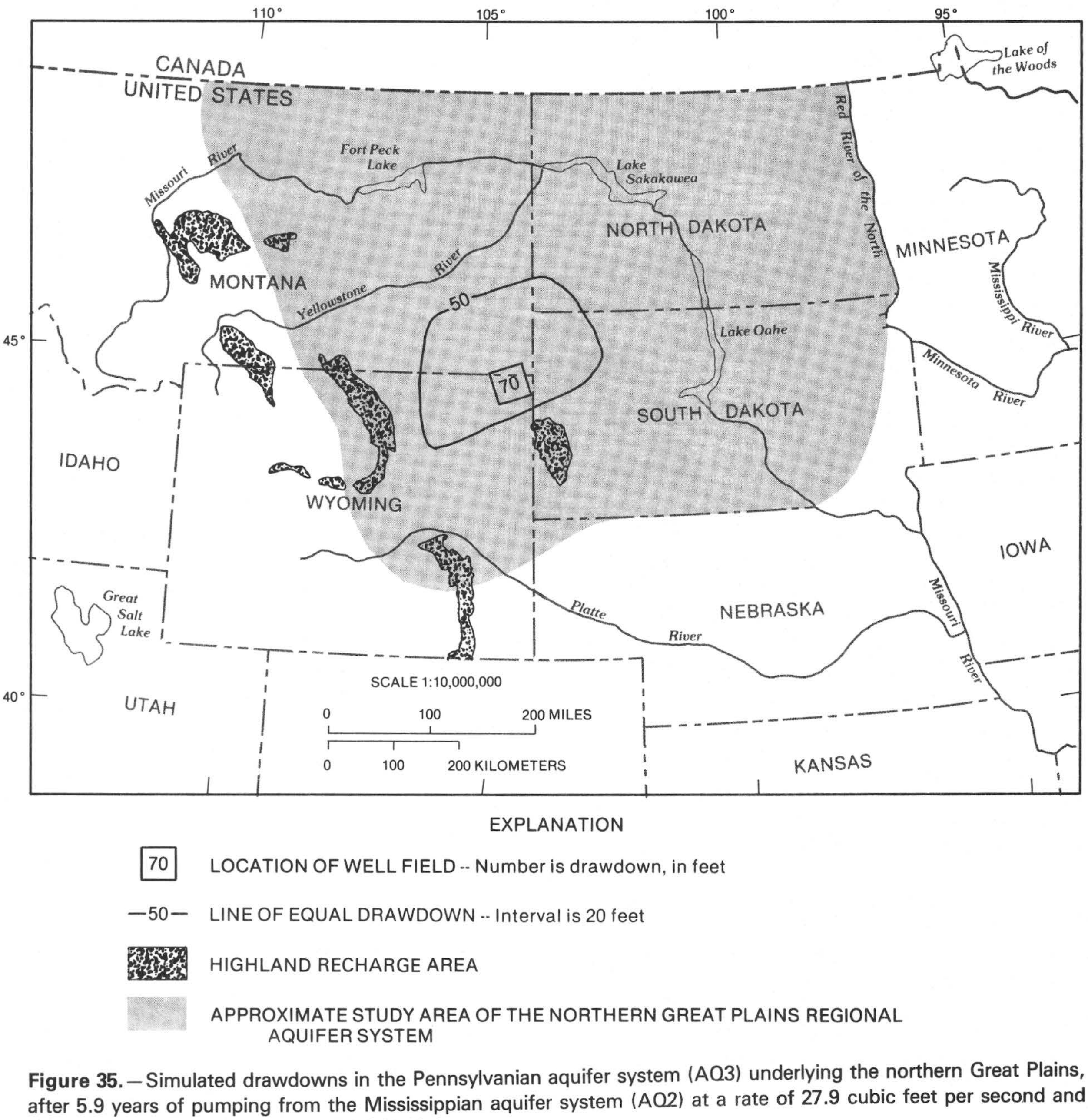


Figure 35.—Simulated drawdowns in the Pennsylvanian aquifer system (AQ3) underlying the northern Great Plains, after 5.9 years of pumping from the Mississippian aquifer system (AQ2) at a rate of 27.9 cubic feet per second and an assumed storage coefficient of 2.0 X 10^{-6}.

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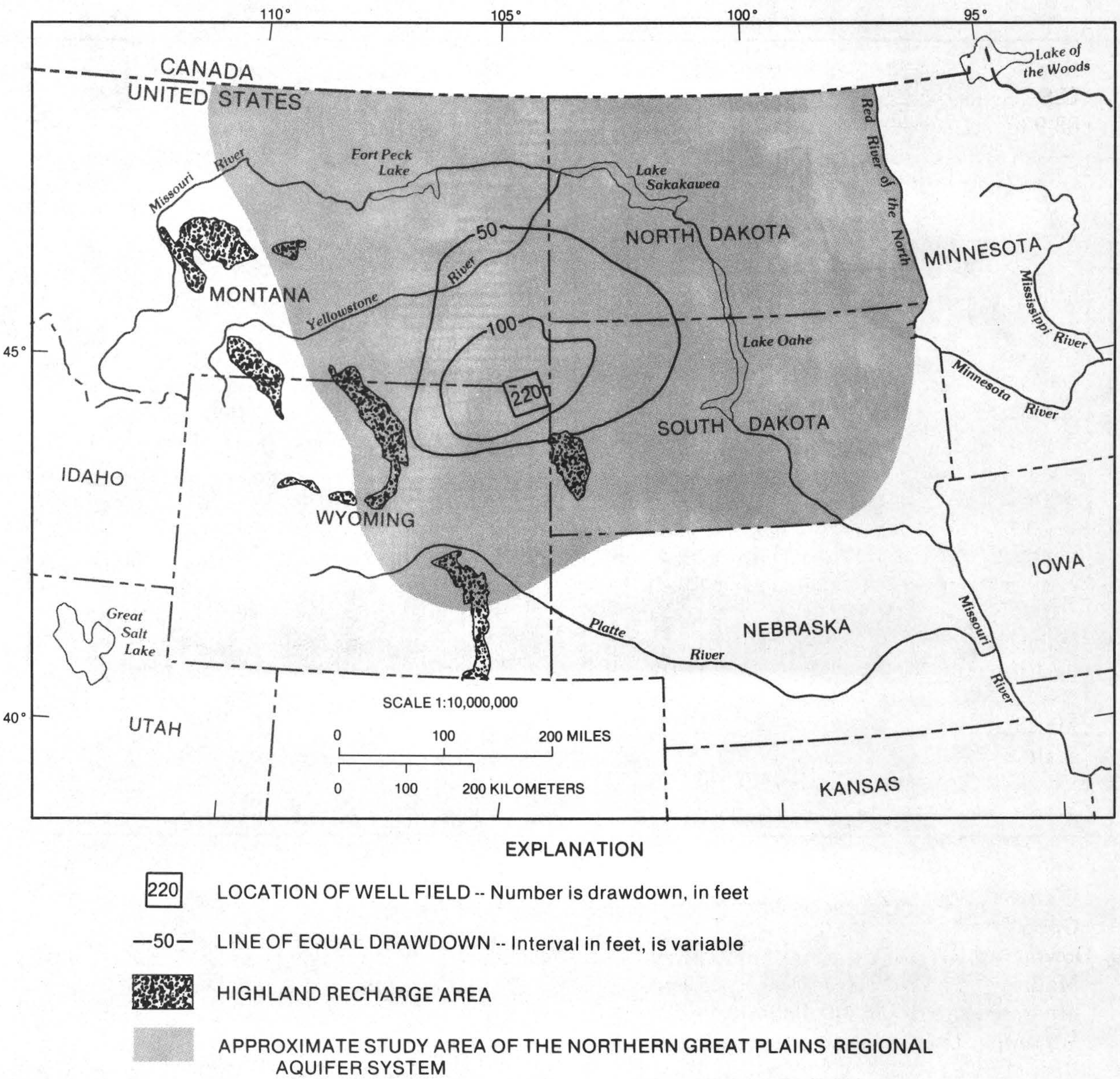

Figure 36.—Simulated drawdowns in the Cambrian-Ordovician aquifer system (AQ1) underlying the northern Great Plains, after 5.9 years of pumping from the Mississippian aquifer system (AQ2) at a rate of 27.9 cubic feet per second and an assumed storage coefficient of 2.0×10^{-6}.

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NORTHERN MIDWEST REGIONAL AQUIFER-SYSTEM STUDY

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INTRODUCTION

The Northern Midwest regional aquifer-system study was started in 1978 and completed in 1984. The study was designed to investigate the hydrogeology, ground-water availability, and chemical quality of the ground water in an aquifer system consisting of rocks of Cambrian and Ordovician age, in parts of Illinois, Indiana, Iowa, Minnesota, Missouri, and Wisconsin (fig. 37), and to describe the regional interaction of all components of the aquifer system (Steinhilber and Young, 1979). This aquifer system is referred to as the Cambrian-Ordovician aquifer system in this report. The rocks forming the aquifer system are mainly sandstone and dolomite. The aquifer system is used extensively for industrial and rural water supplies in the six states and is the primary source of water for many metropolitan areas. The study area is about 161,000 mi^2 and its border delimits either the natural physical or hydrologic boundaries of the aquifer system.

The aquifer strata were deposited in shallow seas that encroached on the Precambrian rock surface, which slopes generally southward from Minnesota and Wisconsin, but eastward from eastern Wisconsin. Structural basins are located in southwestern Iowa, south-central Illinois, and Michigan. The aquifer system crops out in an arc around the Wisconsin Arch and thickens toward the basins where it is deeply buried beneath younger sedimentary rocks (fig. 37). The aquifer system contains three distinct aquifers—St. Peter-Prairie du Chien-Jordan aquifer, Ironton-Galesville aquifer, and Mount Simon aquifer. The aquifers consist primarily of sandstone and are separated by shale, shaly dolomite, or siltstone confining units, as shown in figure 38. Where the Maquoketa Shale is present, it and the underlying dolomite and shale strata of the Galena, Decorah, Platteville, and Glenwood Formations form a major confining unit that overlies the three aquifers.

The Cambrian-Ordovician aquifer system is a leaky-artesian system; and movement of ground water is partly controlled by internal confining units of low permeability. In the outcrop area, water-table conditions prevail in shallow parts of the aquifer system and where the system is thin. Much of the recharge in upland areas discharges to streams through local flow systems, which are no more than a few miles in length. The remainder of the recharge moves slowly downward to deeper formations and downgradient to the regional ground-water flow system.

Regional ground-water movement in the confined part of the system is generally away from the structural highs in the north toward the structural lows (basins) in the south and east (fig. 39). The rate of ground-water movement is very slow and the flux along flow paths into the basins decreases due to a reduction in permeability and a progressive loss of water from the continuous, although small, upward leakage.

In the late 1800's, when wells first were drilled into the confined areas of the aquifer system in eastern Wisconsin, northeastern Illinois, and along stream valleys in eastern Iowa, the wells were flowing. The hydraulic head at that time was as much as 140 feet above Lake Michigan in Milwaukee and 130 feet above the Lake in Chicago. Flowing wells are still common in valleys of the Mississippi River and its major tributaries.

The Cambrian-Ordovician aquifer system supplies a major part of the water needs in the study area. Many metropolitan areas depend on it for all or part of their water supplies. Hydraulic heads in the aquifer system have declined hundreds of feet since the late 1800's in the heavily pumped Chicago-Milwaukee area (fig. 40) and to a somewhat lesser extent in other major metropolitan areas. Projections of future water needs indicate continuing or increasing demands and, therefore, continuing water-level declines are expected. The Illinois State Water Survey has monitored pumpage and water-level declines and has reported on aquifer characteristics and water availability and quality (Schicht and others, 1976; Sasman and others, 1982). In addition, the Illinois State Geological Survey has studied the geologic framework and geochemistry of the aquifer system. These agencies have contributed to the planning of the Illinois Division of Water Resources (1980) for its allocation of water from Lake Michigan. Control of the allocation process gives the Division indirect control of deep-well pumpage.

The aquifer system contains highly mineralized water in several places, especially in its deepest parts, which generally coincide with regional discharge areas or structurally low areas. These areas are mainly in the southwestern, southern, and eastern parts of the study area. Water from highly mineralized zones may be induced into freshwater zones by large withdrawals of freshwater, such as those presently occurring in northeastern Illinois, southeastern Wisconsin, and central Iowa.

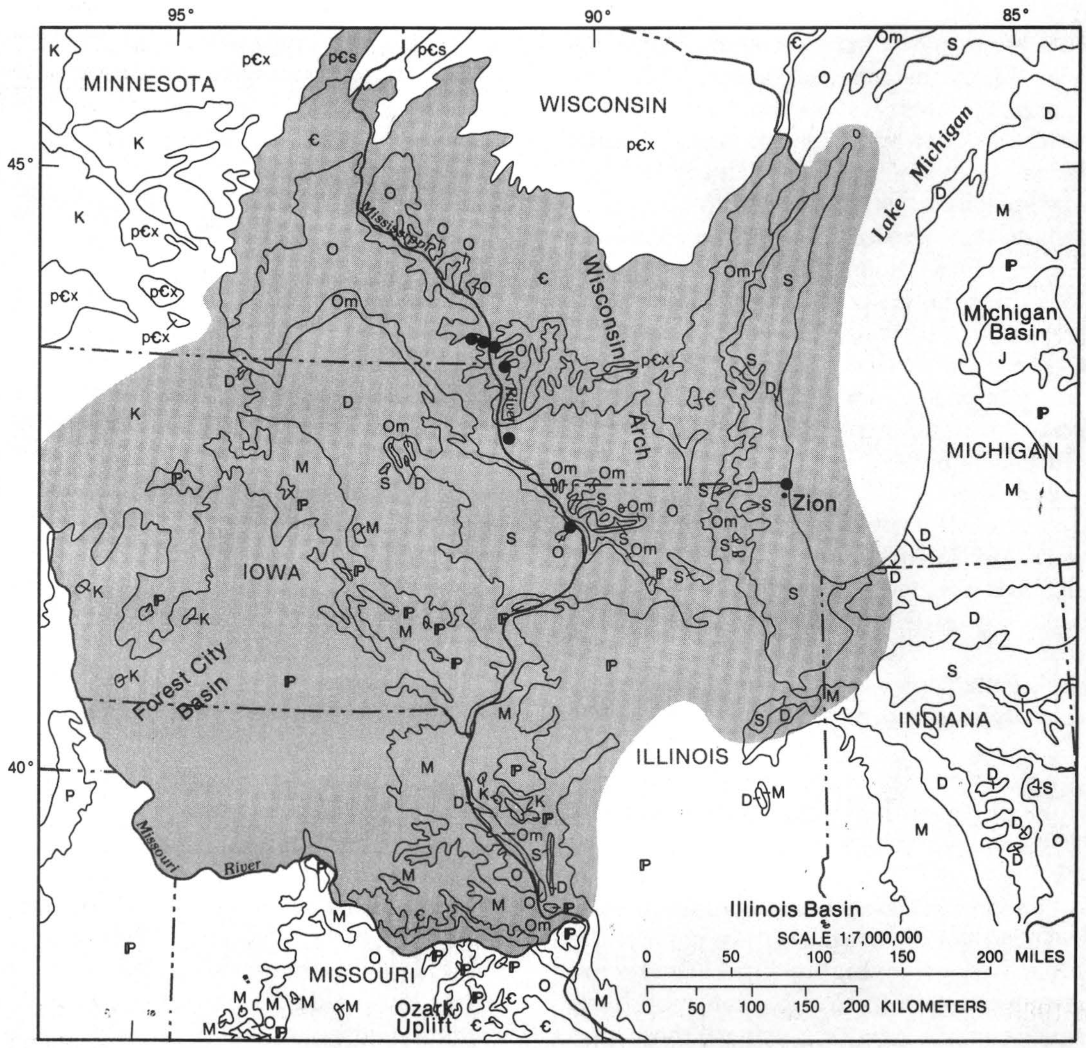
GEOLOGIC AND HYDROLOGIC DATA

State agencies and the District Offices of the U.S. Geological Survey provided the geologic data for the study, consisting of contour and isopach maps of the top of each aquifer and confining unit. The state maps were combined into regional maps and converted into a computerized data base (Kontis and Mandle, 1980) that provided easy input to flow models. Structure contour maps of the tops of ten units described in figure 38 were produced: (1) the Precambrian basement surface; (2) the Mount Simon, (3) Eau Claire, and (4) Ironton Sandstones; (5) the St. Lawrence Formation of Cambrian age; (6) the St. Peter Sandstone; (7) the Galena Dolomite and (8) Maquoketa Shale of Ordovician age; (9) the combined Silurian and Devonian rocks; and (10) the combined Mississippian and Pennsylvanian rocks. Isopach maps were made of the total section between successive structural surfaces.

Although geologic data are available for each formation, very little information exists on hydraulic properties, head distribution, and water quality within the aquifer system, except in outcrop areas. In most of the study area, wells are open to more than one formation. To obtain data on individual aquifers at various points, deep test wells were drilled at seven sites (fig. 37). Inflatable packers, pressure transducers, and submersible pumps were used to obtain hydraulic conductivity, head, and water quality in individual aquifers. After testing was completed, piezometers were installed in three to four discrete zones within the test wells. These piezometers are now part of the U.S. Geological Survey's ground-water monitoring network and provide long-term benefit for monitoring water quality and head in the Cambrian-Ordovician aquifer system.

The most important of the seven deep well test sites was the 3,475-foot well near Zion, IL, (Nicholas and others, 1984). The well reached Precambrian basement at a depth of 3,435 feet (2,849 feet below sea level) and encountered saline water below a depth of about 1,800 feet. Piezometers were installed at different depths. The observed head in the Mount Simon was about 50 and 55 feet higher than the heads in the St. Peter-Prairie du Chien-Jordan aquifer and the Ironton-Galesville aquifer, respectively. Head data from the test well indicated that the head in the Ironton-Galesville was the lowest. This fact indicates that ground water flows into the Ironton-Galesville aquifer from both the overlying and the underlying aquifers, resulting from regional pumping stress and the significantly higher transmissivity of the Ironton-Galesville aquifer.

Head data for predevelopment conditions are sparse in many areas, especially for deeper aquifers, however, they are sufficient for primary studies. The preparation of the potentiometric surface map for the St. Peter-Prairie du Chien-Jordan aquifer shown in figure 39 was based on those sparse data. The predevelopment heads in the deeper Ironton-Galesville and Mount Simon aquifers probably were similar to those shown in figure 39. The broadly spaced, smooth contours in Iowa, northern Missouri, and Illinois mark areas of strong confinement beneath the Maquoketa Shale and younger strata. Regional flow is southeastward to the Illinois Basin.



EXPLANATION

K	CRETACEOUS FORMATIONS	S	SILURIAN FORMATIONS
J	JURASSIC FORMATIONS	Om	ORDOVICIAN-AGE MAQUOKETA SHALE
P	PERMIAN FORMATIONS	O	ORDOVICIAN FORMATIONS OLDER THAN THE MAQUOKETA SHALE
ℙ	PENNSYLVANIAN FORMATIONS	€	CAMBRIAN FORMATIONS
M	MISSISSIPPIAN FORMATIONS	p€x	PRECAMBRIAN-AGE SANDSTONE
D	DEVONIAN FORMATIONS	p€s	PRECAMBRIAN-AGE CRYSTALLINE ROCK

——— CONTACT

● LOCATION OF TEST WELL DRILLED FOR THIS PROJECT

▒ STUDY AREA

Figure 37.—General bedrock geology of the northern Midwest and locations of test wells drilled during the northern Midwest regional aquifer-system study. (Modified from maps of individual states produced by state agencies, dated from 1966 to 1982).

Geologic system	Hydrogeologic unit	Designation of layers used in the regional flow model: Aquifer	Designation of layers used in the regional flow model: Confining unit
Quaternary	Drift	5	
Cretaceous	Cretaceous aquifer—only in Minnesota and Iowa		
Pennsylvanian Mississippian Devonian	Pennsylvanian-Mississippian-Devonian confining unit		4-5
Silurian	Silurian aquifer—includes basal Devonian in Illinois and Indiana	4	
Ordovician	Maquoketa confining unit—consists of the Maquoketa Shale and the Galena, Decorah, Platteville, and Glenwood formations or equivalents, where overlain by the Maquoketa. Equivalents in Missouri and southern Illinois are the Maquoketa Shale through the Joachim Dolomite		3-4
Ordovician	St. Peter-Prairie du Chien-Jordan aquifer—in Missouri and Illinois, includes the St. Peter through Eminence formations	3	
Cambrian	St. Lawrence-Franconia confining unit—in Missouri and Illinois, includes the Potosi through Davis formations		2-3
Cambrian	Ironton-Galesville aquifer—not present in southern Illinois, southwestern Iowa, and Missouri	2	
Cambrian	Eau Claire confining unit—Bonneterre Formation is the equivalent in Missouri and western Iowa		1-2
Cambrian	Mount Simon aquifer: Includes overlying Elmhurst Sandstone Member of the Eau Claire Formation in Illinois	1	
Cambrian	Mount Simon aquifer: Mount Simon Sandstone is main component. Equivalent in Missouri is the Lamotte Sandstone		
Precambrian	Mount Simon aquifer: Includes underlying Hinckley Sandstone in Minnesota		

Figure 38.—Stratigraphic configuration of the regional aquifers and confining units delineated for the northern Midwest regional aquifer-system study.

The 1980 potentiometric surface for the St. Peter-Prairie du Chien-Jordan aquifer shows little change in head in the recharge areas, but significant declines in most of the confined area (figs. 39 and 41). The area of most significant head decline is in the Chicago-Milwaukee area, where the potentiometric surface of the Cambrian-Ordovician aquifer system has declined as much as 850 feet in the vicinity of Chicago (Sasman and others, 1982) and 375 feet in the vicinity of Milwaukee (fig. 40). Water levels in some Chicago area wells are 100 to 150 feet below sea level. Pumpage has increased almost steadily to about 180 Mgal/d in the Chicago area and about 33 Mgal/d in southeastern Wisconsin in 1980.

FLOW MODELS

The complexities of the regional ground-water flow system were studied by development and calibration of a regional digital flow model and three subregional models (fig. 42). The models simulate the flow systems and can be used to estimate water-level changes based on projected future pumping. A quasi three-dimensional flow model (Trescott, 1975) was used to simulate the five aquifers and four confining units in the regional flow model. A uniform node spacing of 16 miles was used for the regional flow model. The three lower aquifers—St. Peter-Prairie du Chien-Jordan aquifer, Ironton-Galesville aquifer, and Mount Simon aquifer—are of primary interest; the upper two aquifers—drift and Cretaceous aquifer and Silurian aquifer—serve as sources and sinks for the lower three aquifers. Two significant modifications to the Trescott (1975) model code were necessary to simulate the Cambrian-Ordovician aquifer system. Saline water, including brine in the deep structural basins, is represented in the model by the addition of a function to simulate flow in ground water with variable densities. Simulation of multiaquifer well effects (Bennett and others, 1982) was also added to the model to incorporate the variation in yield from, and hydraulic head of, each aquifer penetrated by the well.

In addition to the regional flow model, subregional flow models were constructed for areas of local interest. The subregional flow models are: (1) a two-dimensional flow model of northeastern Missouri (Imes, 1984), (2) a quasi three-dimensional flow model of northeastern Wisconsin (Emmons, in press), and (3) a two-dimensional flow model of the Jordan aquifer in Iowa (M.R. Burkart and R.C. Buchmiller, U.S. Geological Survey, written commun., 1984). Although unrelated to the regional aquifer-system study, two U.S. Geological Survey cooperative program studies to evaluate ground-water flow benefited from joint data-collection efforts. The studies are

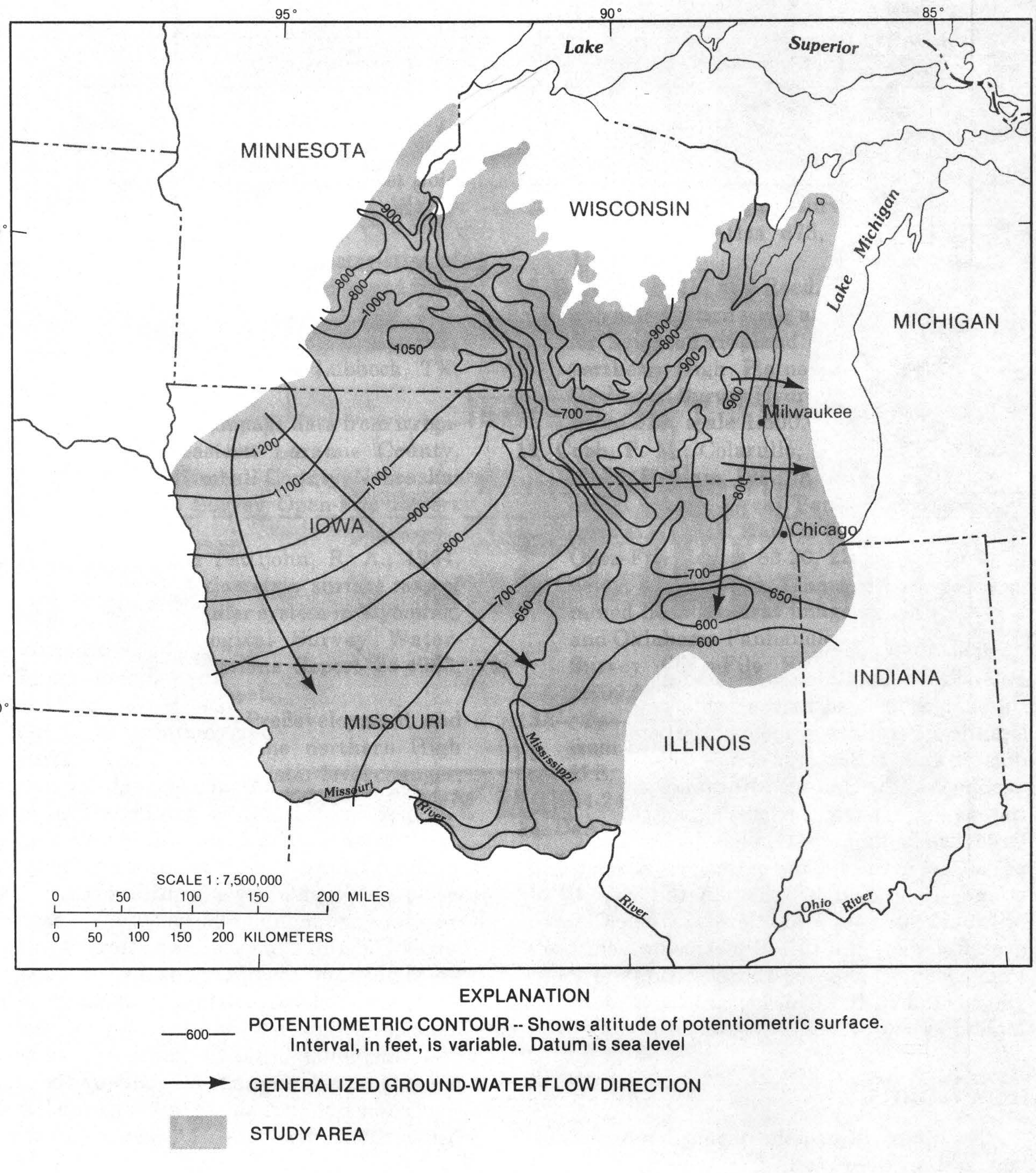
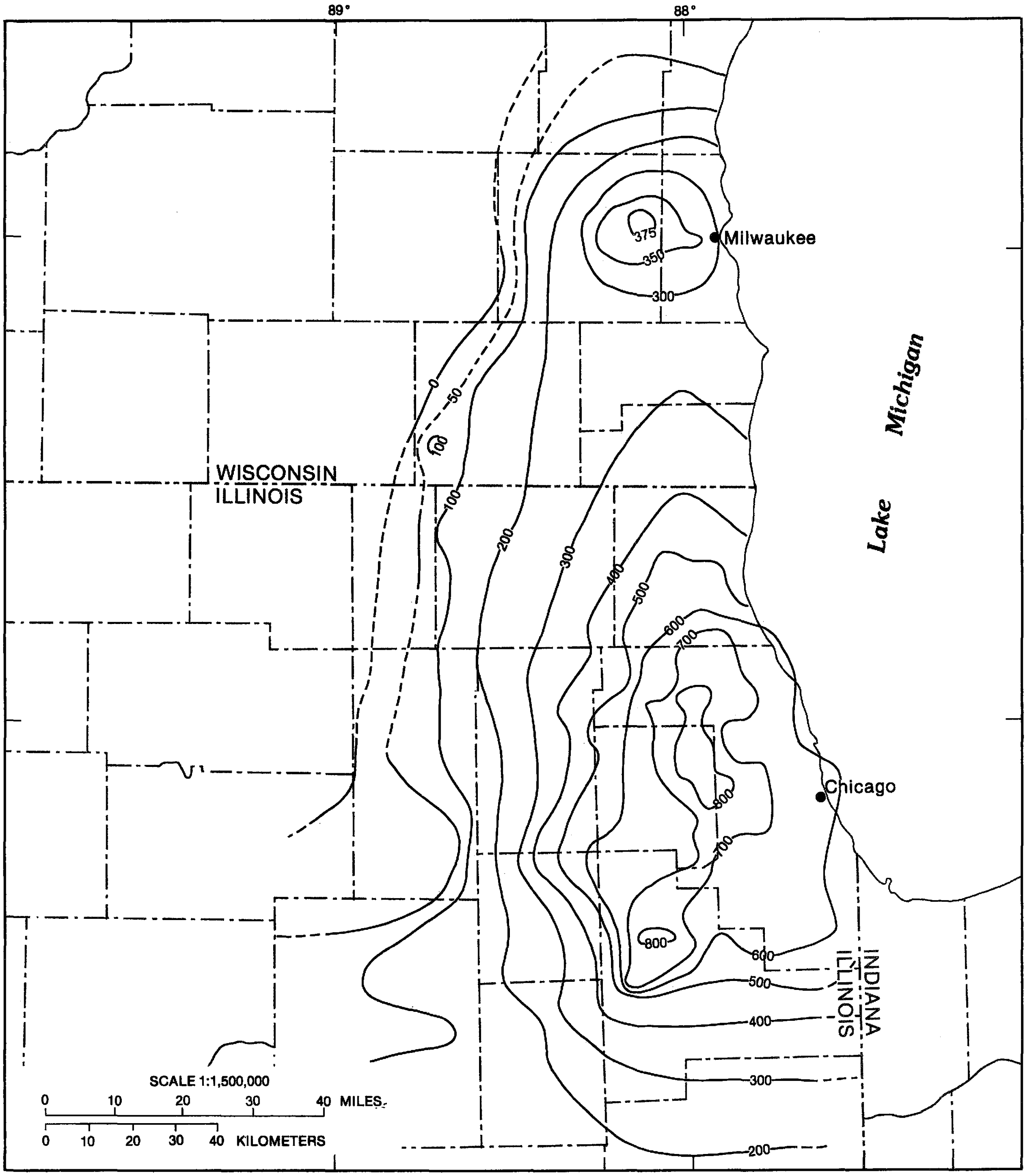


Figure 39.—Generalized predevelopment potentiometric surface of the St. Peter-Prairie du Chien-Jordan aquifer of the Cambrian-Ordovician aquifer system in the northern Midwest.

of the Minneapolis-St. Paul area (Guswa and others, 1982), and of Brown County, Wisconsin (Krohelski, in press).

The regional flow model was initially calibrated for steady-state predevelopment conditions, and then for transient conditions to simulate the effects of ground-water development from 1864 to 1980. Simulated drawdown is less than 50 feet throughout central Wisconsin, north-central Illinois, extreme northeastern Iowa, and



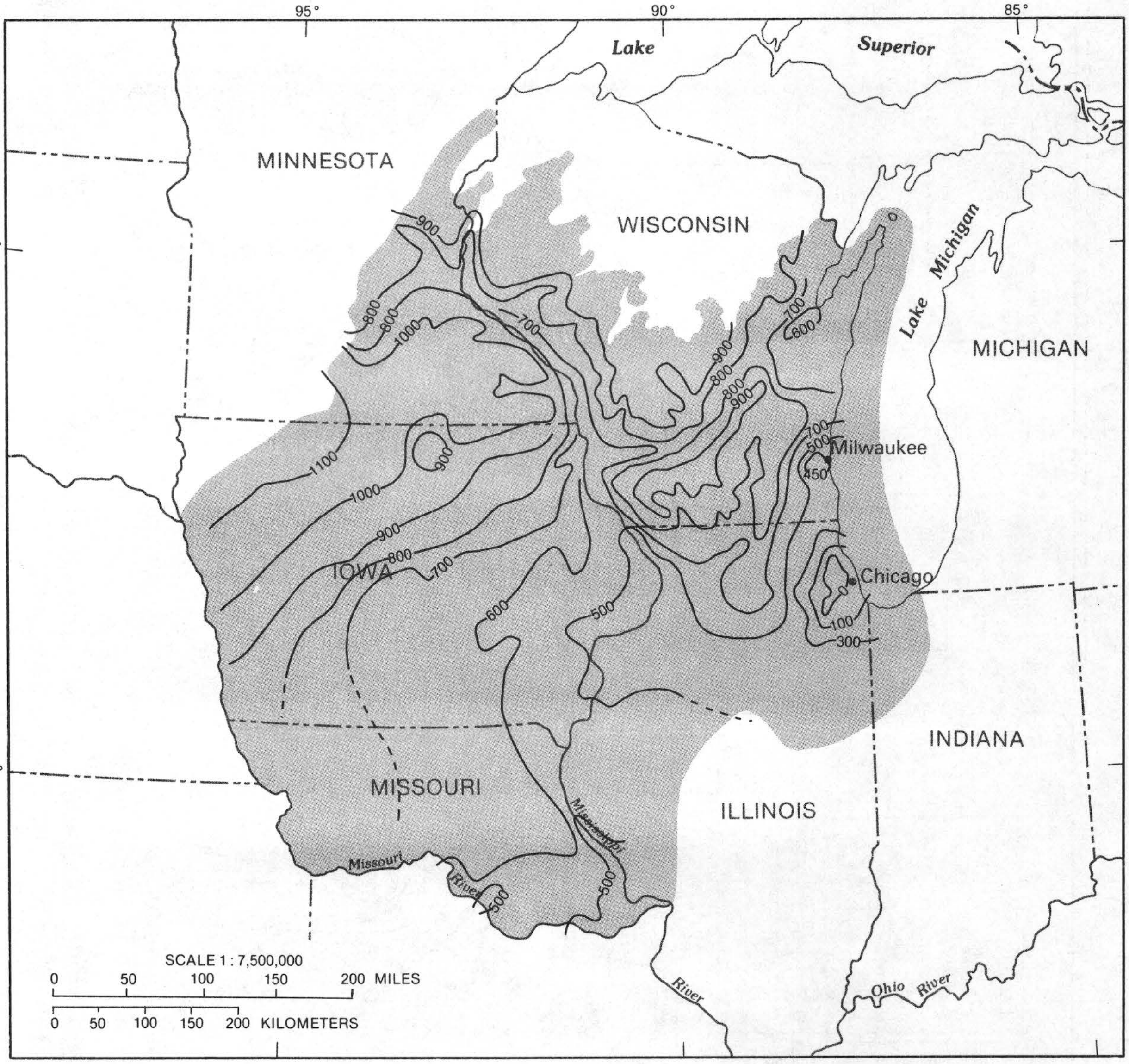
EXPLANATION

—200— — LINE OF EQUAL HEAD DECLINE -- Dashed where location is approximate. Interval is 100 feet with supplementary 25- and 50-foot lines

Figure 40.—Decline in the potentiometric surface of the Cambrian-Ordovician aquifer system in the Chicago-Milwaukee area, 1864-1980.

southern Minnesota (figs. 43-44), where the Maquoketa Shale is absent and the aquifers of the Cambrian-Ordovician aquifer system are mainly unconfined or loosely confined.

Large coalescing drawdown cones have developed in the Cambrian-Ordovician aquifer system in northeastern Illinois, northwestern Indiana, eastern Wisconsin, central and east-central Iowa, and central Missouri (figs. 43-44), where the aquifer system is tightly confined. Heads in the St. Peter-Prairie du Chien-Jordan aquifer most closely depict the composite head conditions in the Cambrian-Ordovician aquifer system. The large areal extent of the cone for the St. Peter-Prairie du Chien-Jordan aquifer shows that the aquifer system is tightly confined and that a major source of the water pumped is probably from storage in the aquifer. The pattern is similar for the Ironton-Galesville and the Mount Simon aquifers (fig. 44), except in Iowa and Missouri where these aquifers



EXPLANATION

—800— — POTENTIOMETRIC CONTOUR -- Shows altitude of potentiometric surface. Dashed where location is approximate. Interval, in feet, is variable. Datum is sea level

STUDY AREA

Figure 41.—Potentiometric surface of the St. Peter-Prairie du Chien-Jordan aquifer of the Cambrian-Ordovician aquifer system in the northern Midwest, 1980.

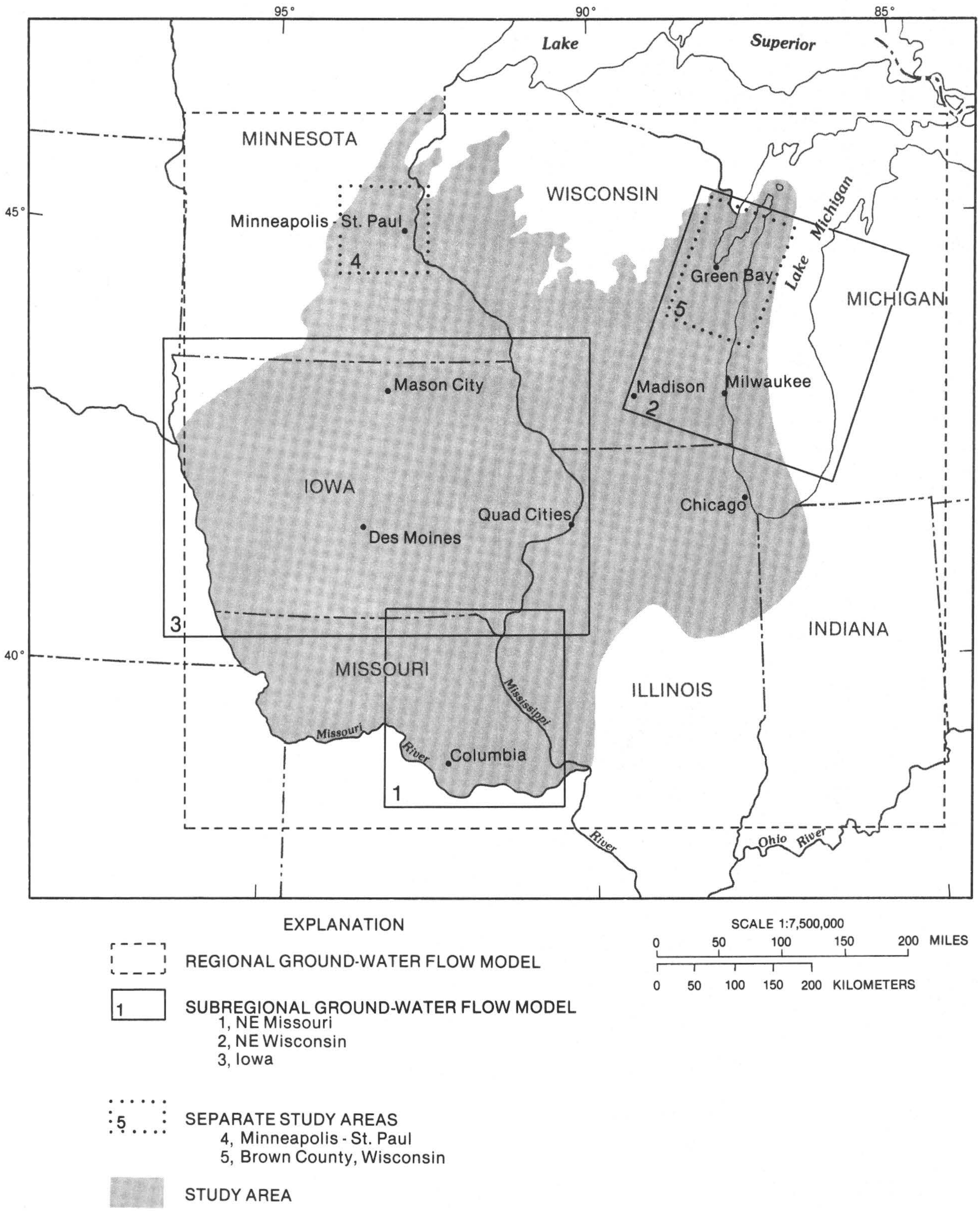


Figure 42.—Location and areal extent of and subregional ground-water flow models of the Cambrian-Ordovician regional aquifer system in the northern Midwest.

are little used. The extensive drawdown cone in eastern Wisconsin and northeastern Illinois and the drawdown cone in eastern Iowa and northwestern Illinois terminate near the edge of the Maquoketa Shale confining unit, showing the effect of recharge from the overlying unconfined aquifers. The drawdown cones spread farther in the confined area away from the outcrop area, because the primary source of pumped water there is probably from aquifer storage.

GEOCHEMICAL DATA

More than 3,000 chemical analyses of ground water were available in the study area. Major sources of the data were the WATSTORE data base of the U.S. Geological Survey, publications of State and Federal agencies, and scientific journals. In addition, about 200 water samples were collected during the study. These samples were

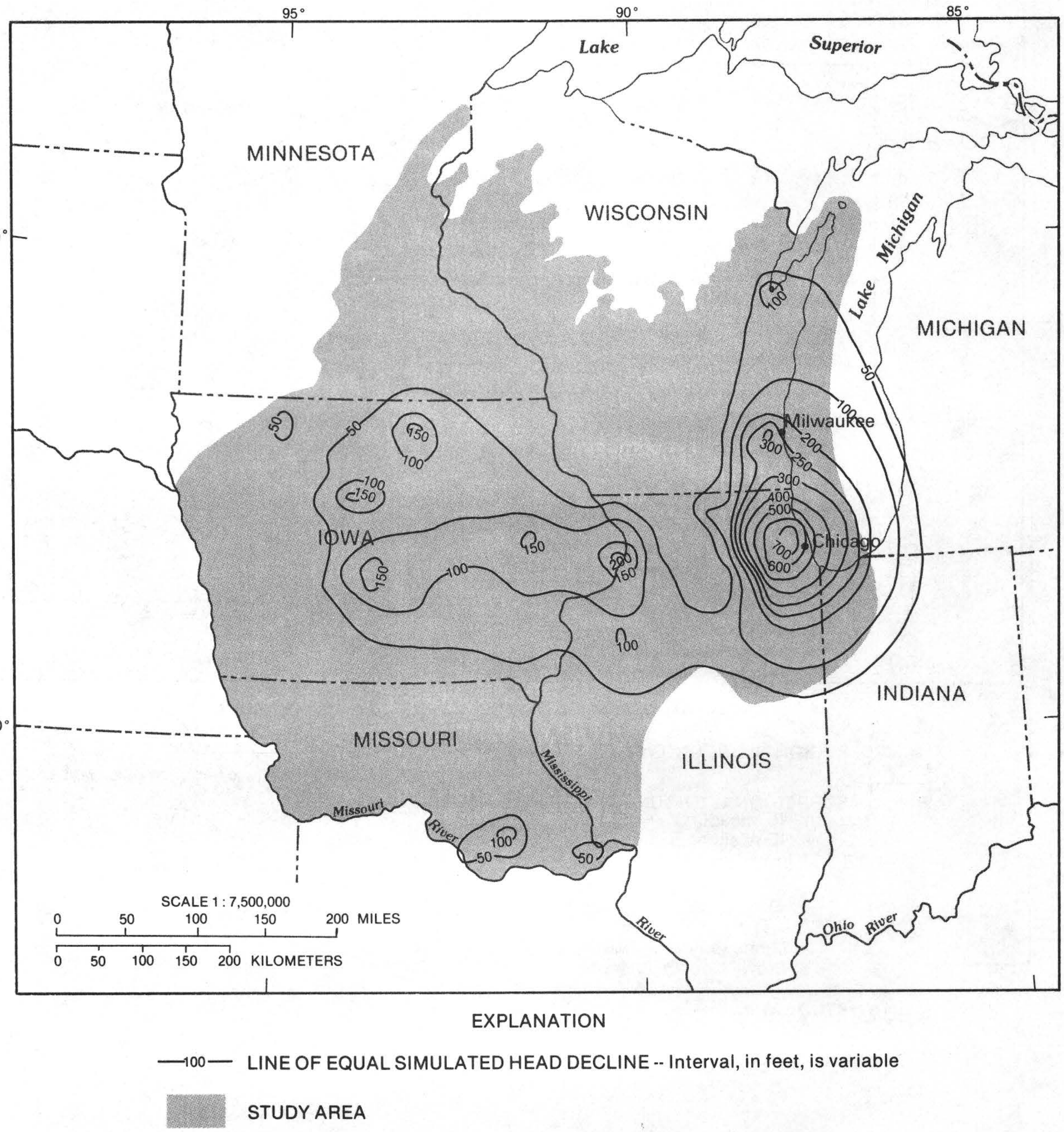


Figure 43.—Simulated head decline, 1864-1980, in the St. Peter-Prairie du Chien-Jordan aquifer of the Cambrian-Ordovician aquifer system in the northern Midwest.

analyzed for major ions, trace metals, and radiochemical species. In addition, 85 samples were analyzed for the isotopes of oxygen and hydrogen in water and carbon in bicarbonate, and 9 were analyzed for the isotopes of sulfur in sulfate.

Detailed evaluation of ground-water quality was restricted to the shallow drift and Cretaceous aquifer and the St. Peter-Prairie du Chien-Jordan aquifer, for which the areal distribution of data was considered sufficient to evaluate the trends and variation in concentrations of major dissolved constituents. In most of Wisconsin and north-central Illinois, the St. Peter-Prairie du Chien-Jordan, Ironton-Galesville, and Mount Simon aquifers are unconfined and are hydraulically connected, thus water quality in these aquifers is probably similar.

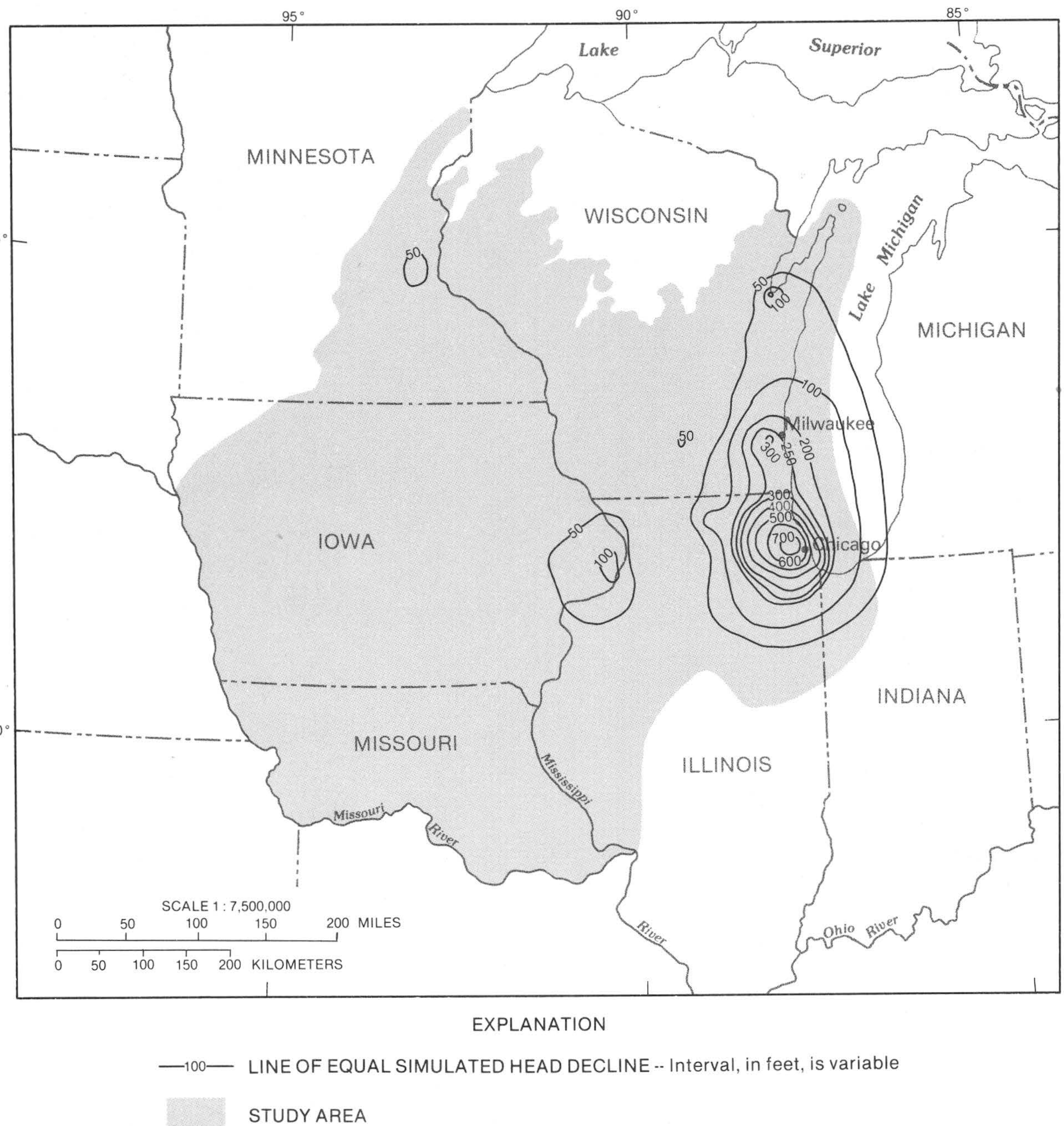


Figure 44.—Simulated head decline, 1864-1980, in the Mount Simon aquifer of the Cambrian-Ordovician aquifer system in the northern Midwest.

Ground water in outcrop areas and beneath the Maquoketa Shale confining unit to depths below land surface of about 2,000 feet generally has concentrations of dissolved solids of less than 1,000 milligrams per liter (mg/L). The water is of a calcium-magnesium-bicarbonate type. However, concentrations of dissolved solids tend to increase greatly where the aquifer is deeper and the formation is thicker. Ground water becomes a sodium-chloride brine in deeper parts of the structural basins. Commonly, the transition to higher concentrations of dissolved solids in the confined areas is accompanied by substantial increases in sulfate concentration and the occurrence of a calcium-sodium-sulfate type water (figs. 45 and 46).

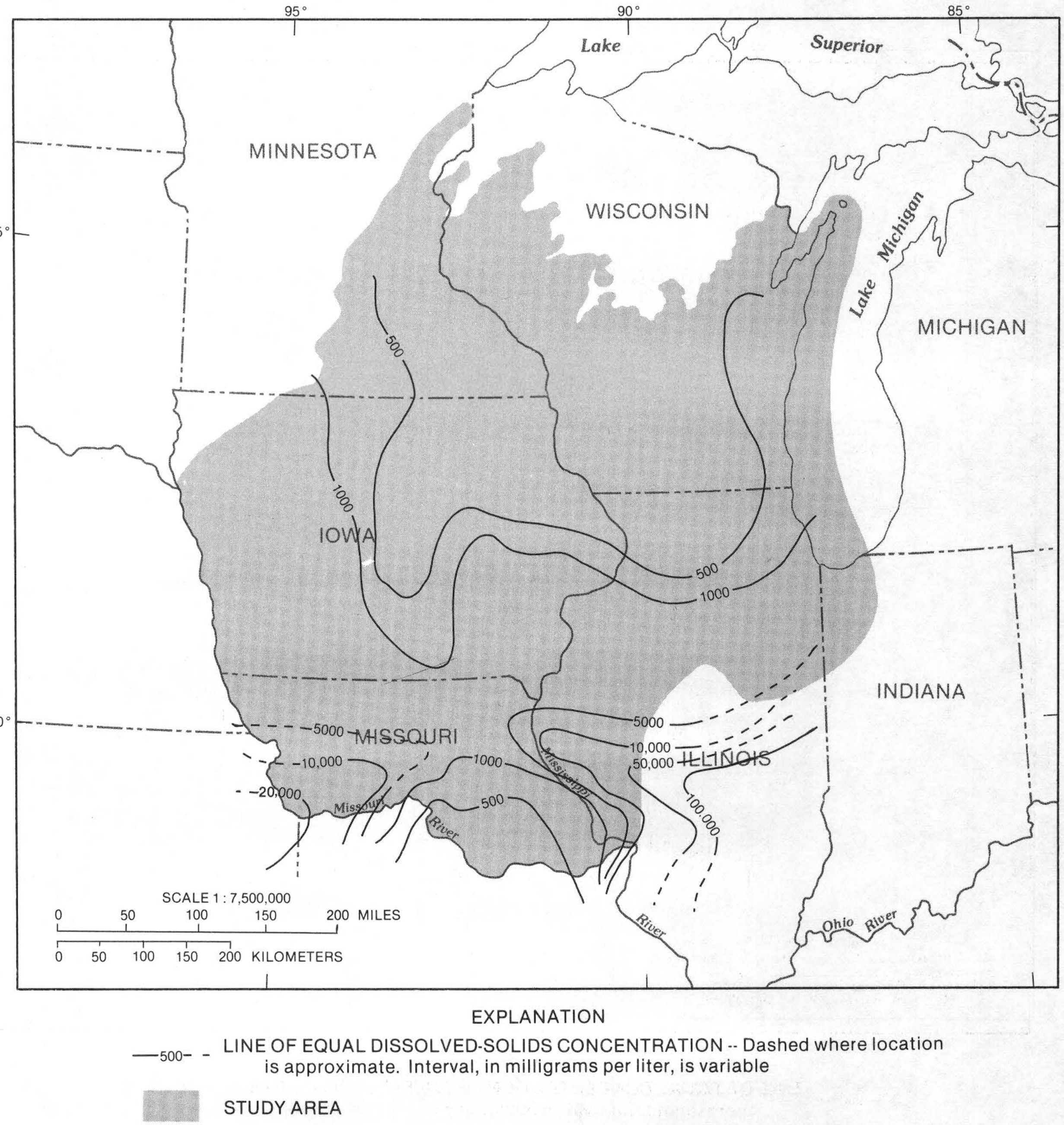


Figure 45.—Dissolved-solids concentrations of ground water in the St. Peter-Prairie du Chien-Jordan aquifer of the Cambrian-Ordovician aquifer system in the northern Midwest.

In recharge areas of Wisconsin, southern Minnesota, northeastern Iowa, and north-central Illinois, the calcium-magnesium-bicarbonate type water (Ca-Mg-HCO_3) (fig. 47) in the aquifer system is identical to that in the overlying glacial drift. It probably resulted from dissolution of carbonate minerals in the soil by carbon dioxide-rich recharge water. In northwestern Iowa and southwestern Minnesota, however, the water in both the glacial drift and the Cambrian-Ordovician aquifer system is of a calcium-sodium-sulfate-bicarbonate (Ca-Na-SO_4-HCO_3) type resulting from the oxidation of pyrite in the overlying Dakota Formation. Dissolved solids in this ground water can exceed 1,000 mg/L. The water is generally undersaturated with respect to gypsum; this fact, and its negative values of $\delta^{34}S$ (standard expression of the ratio of the less

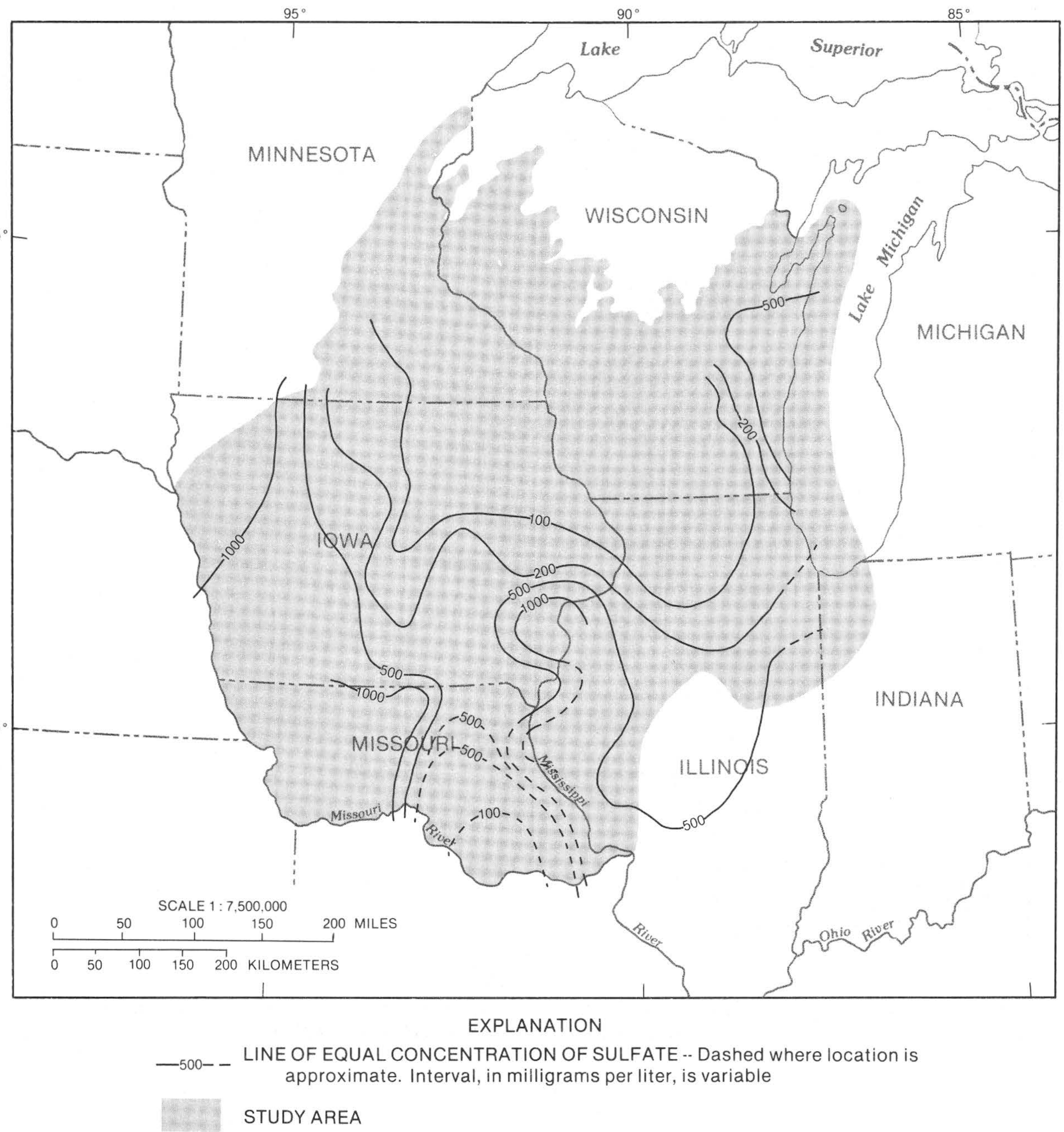
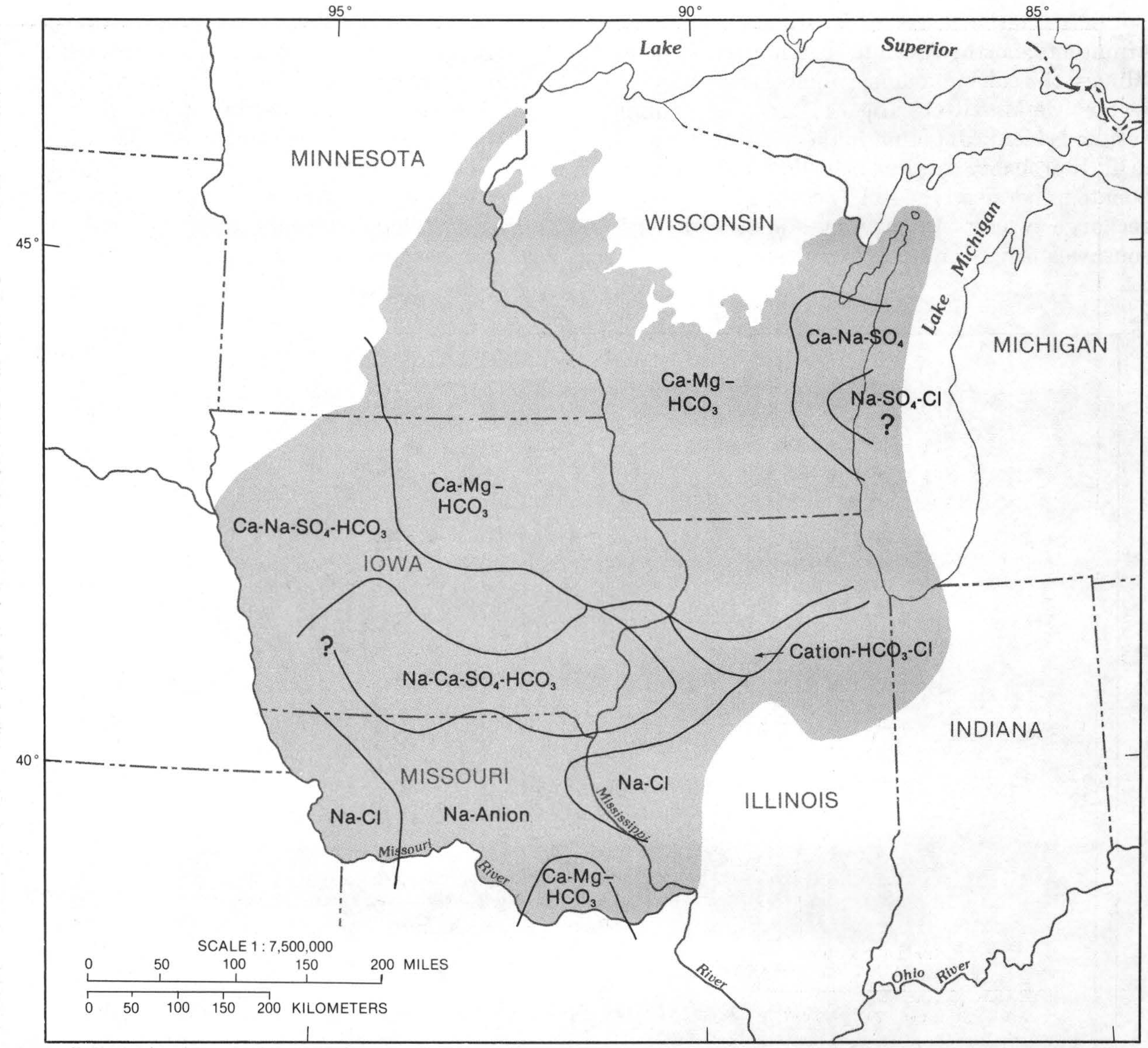


Figure 46.—Sulfate concentrations of ground water in the St. Peter-Prairie du Chien-Jordan aquifer of the Cambrian-Ordovician aquifer system in the northern Midwest.



EXPLANATION

FACIES ZONE BOUNDARY -- Queried where uncertain

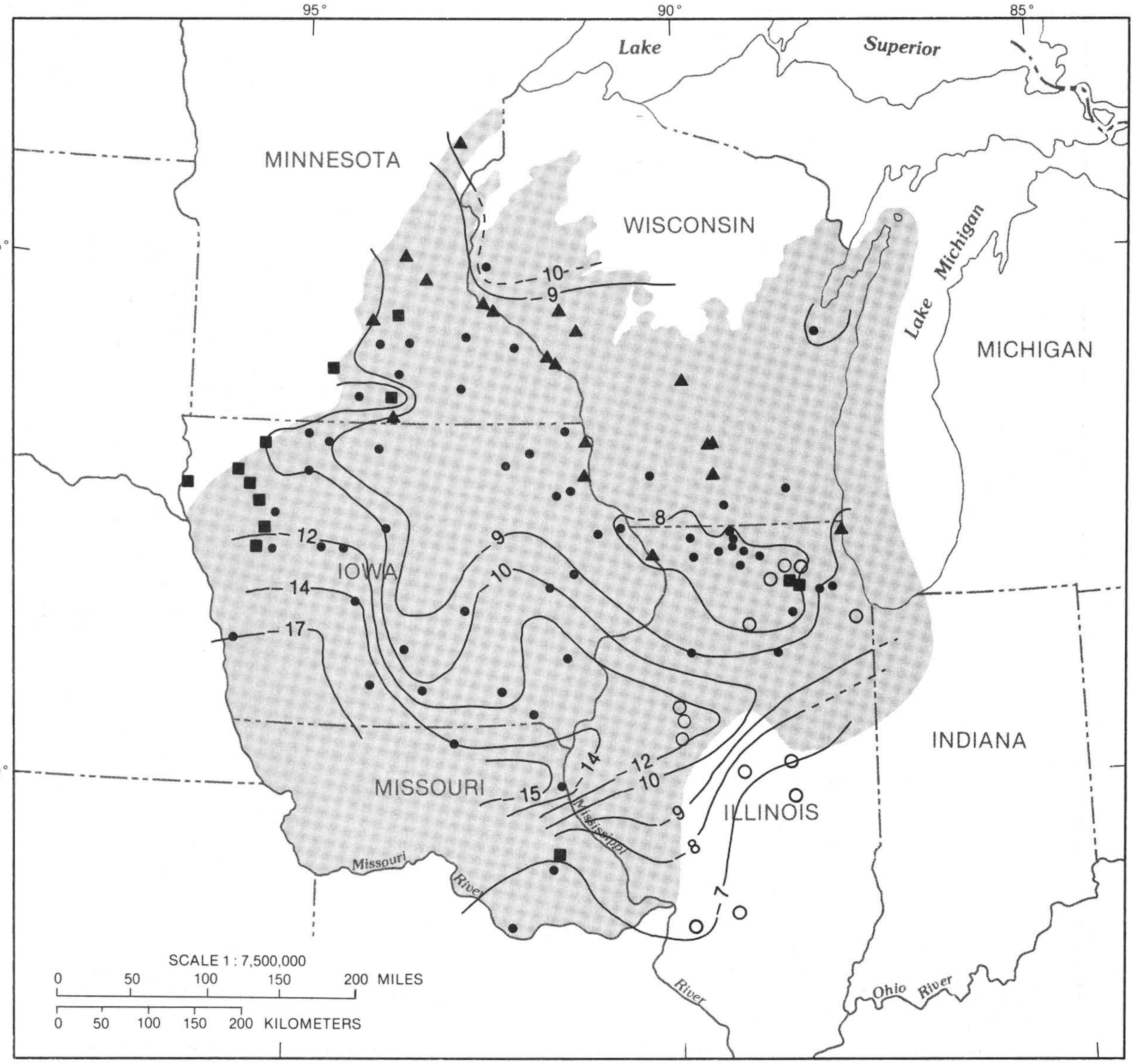
STUDY AREA

Figure 47.—Hydrochemical facies of ground water in the St. Peter-Prairie du Chien-Jordan aquifer of the Cambrian-Ordovician aquifer system in the northern Midwest.

abundant sulfur-34 ion with respect to the more common sulfur-32 ion) of sulfur in sulfate, indicates that pyrite is the primary source of the sulfur.

Ground water changes from a calcium-magnesium-bicarbonate type to a sodium-calcium-sulfate-bicarbonate (Na-Ca-SO_4-HCO_3) type and then to a sodium-chloride (Na-Cl) type along the major flow paths from northwestern Iowa to the Illinois Basin. The changes may be due to ion exchange of calcium and magnesium for sodium in clay or due to membrane filtration that could result in the precipitation of calcite coincident with increases in sodium and chloride.

The evolution of the water quality in the Cambrian-Ordovician aquifer system is com-



EXPLANATION

— — 10— — LINE OF EQUAL $\delta^{18}O$ -- Dashed where location is approximate. Interval, in parts permil, is variable

WELL SAMPLED FOR CHEMICAL ANALYSIS DURING THIS STUDY

▲ Open to Mount Simon aquifer

● Open to St. Peter - Prairie du Chien - Jordan aquifer

■ Open to glacial drift or Cretaceous aquifers

WELL WITH ANALYSIS FROM OTHER SOURCES

○ Open to St. Peter - Prairie du Chien - Jordan aquifer

STUDY AREA

Figure 48.—Distribution of $\delta^{18}O$ in ground water in the Cambrian-Ordovician aquifer system in the northern Midwest.

plicated because of many possible geochemical processes and the effects of Pleistocene glaciation on the aquifer system. The hydrochemical facies of ground water shown in figure 47 may, in part, reflect the water types before or during the Pleistocene. Analysis of the isotopic composition of the ground water strongly indicates that the source of much of the ground water in the confined part of the Cambrian-Ordovician aquifer system, other than in the deep basins, probably was recharge from glacial meltwater (Siegel and Mandle, 1984). The water is isotopically depleted in $\delta^{18}O$ (standard expression of the ratio of the less abundant oxygen-18 ion with respect to the more common oxygen-16 ion) with respect to modern precipitation (fig. 48). This indicates that the source of the ground water was precipitation in a climate similar to that at latitudes hundreds of miles north of the study area and suggests that the water probably was recharged during Pleistocene glaciations.

Pleistocene recharge probably also was a factor in the occurrence of a plume of water in central Iowa that has much lower concentrations of dissolved solids than in neighboring waters (fig. 45). The distribution of $\delta^{18}O$ and the concentration of lithium, radium-226, and bromide show a correlation with the plume. Concentrations of dissolved solids generally increase downgradient; however, in Iowa they increase from northeast to southwest, whereas the regional flow is southeastward across the State. This may suggest that the direction of paleoflow was perpendicular to the direction of flow today, possibly due to emplacement of subglacial meltwater by very high hydraulic gradients from ice loading.

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SNAKE RIVER PLAIN REGIONAL AQUIFER-SYSTEM STUDY

By Gerald F. Lindholm, Boise, Idaho

INTRODUCTION

The economy of southern Idaho is based largely on agriculture, and a key to successful agriculture is an adequate supply of good-quality water for irrigation. In the past, most water for irrigation was obtained from the Snake River. Today (1984), ground water is also a major source. As water use increases, so does competition for the right to use water. It follows that a better understanding of the hydrologic system in the Snake River Plain is needed.

The Snake River Plain is an area of about 15,600 mi^2 that extends across southern Idaho into eastern Oregon (fig. 49) and is included in the 69,200 - mi^2 drainage basin of the Snake River above Weiser, Idaho. Along its 502-mile course from Heise to Weiser, the Snake River descends 2,930 feet. The surface of the Plain decreases in altitude from about 6,000 feet above sea level in the northeast to 2,100 feet in the west. Surrounding mountains are as much as 12,000 feet above sea level.

Average annual precipitation on much of the Plain is less than 10 inches, one-third to one-half of which falls during the April through September growing season. Most water available to the Plain originates as snow on surrounding mountains where annual precipitation is as much as 60 inches.

The Snake River Plain regional aquifer-system study began in 1979 (Lindholm, 1981) and was completed in 1984. For study purposes, the Plain was divided into eastern and western parts on the basis of geology and hydrology (fig. 49).

Geophysical studies indicate that the 10,800-mi^2 eastern Plain is basically a downwarp. Most of the fill consists of Quaternary basalt. Near the margins, unconsolidated sedimentary rocks overlie and are intercalated with the basalt (fig. 50). Tops of basalt flows are typically broken and have high hydraulic conductivities. Consequently, thick sections of basalt, which include many flows, store and yield large quantities of water. In places, the basalt aquifer may be several thousand feet thick; however, the upper 200 to 500 feet are thought to be the most permeable. An estimated 200 to 300 million acre-ft of water are stored in the top 500 feet of the regional aquifer system. In much of the eastern Plain, the regional aquifer system is unconfined.

About two-thirds of the ground water discharged from the eastern Plain is through a series of springs that flow to the Snake River between Milner and King Hill. Included in that reach are 11 of the 65 springs in the United States that discharge an average of more than 100 ft^3/s (Meinzer, 1927, p. 42-51). The hydrologic boundary between the eastern and western parts of the Plain includes a 40-mile reach of the Snake River where springs are most concentrated. A second series of springs between Blackfoot and Neeley account for most of the remaining discharge.

The western Plain is bounded by well-defined high-angle faults and is a graben. Quaternary and Tertiary sedimentary rocks of variable thickness are the predominant fill material (fig. 50). In the Boise River Valley, most water is obtained from unconfined alluvial sand and gravel. Elsewhere, rocks are predominantly fine grained; included sand and gravel aquifers are largely confined. Ground-water discharge to the Snake River in the western Plain is small relative to that in the eastern Plain. A geothermal aquifer system underlies much of the western Plain.

GEOHYDROLOGIC FRAMEWORK

Investigators have used a variety of surface geophysical techniques to define the regional structural (crustal) features of the Snake River Plain (Whitehead, 1984a). However, little work has been done with the specific objective of defining the geohydrologic framework. The approach in this study was to use results of previous studies as a starting point and concentrate interpretive efforts on the geologic section that most likely contains significant amounts of water.

In about two-thirds of the Snake River Plain, Quaternary basalt of the Snake River Group contains and readily yields large quantities of water to wells. Most of the wells on the Plain penetrate only a few tens of feet of the basalt aquifer; accordingly, aquifer thickness is largely unknown.

Zohdy and Stanley (1973) demonstrated the usefulness of vertical electrical-resistivity soundings to help define the thickness of major rock types in the eastern Plain. After drill-hole data were assembled and analyzed during this study, electrical-resistivity soundings were made to help define the thickness of the Quaternary basalt. About 400 miles of resistivity profiles were generated. These new profiles supplemented about 300 miles of previous resistivity profiling. Other surface geophysical techniques were also investigated and gravity modeling was used in areas where no other geophysical data were available.

To help verify subsurface interpretations based on surface geophysics, a 1,123-foot deep test hole was drilled in Gooding County on the eastern Plain (fig. 49). About 75 percent of the hole was cored; the remainder of the hole was drilled by rotary methods. Five zones were isolated in the hole and piezometers installed to determine head changes with depth (Whitehead and Lindholm, 1984).

Quaternary basalt underlying the eastern Plain is estimated to be as much as 5,000 feet thick, of which a maximum of about 3,500 feet may be saturated (fig. 51). Thickness of the Quaternary basalt is greatest near the center of the eastern Plain between Arco and Lake Walcott. Individual basalt flows average 20 to 25 feet in thickness. The top of each flow, generally less than 6 feet thick, is typically fine grained, vesicular, highly broken, and has high hydraulic conductivity. The remainder of the flow is typically coarse grained, massive, and may be broken vertically. Hydraulic conductivity of the massive basalt is dependent on the degree of fracturing and is typically low. It follows that transmissivity of the total basalt section is largely dependent on the number of individual flows and the thickness of permeable flow tops.

Porosity and hydraulic conductivity of volcanic rocks in the Snake River Plain appear to decrease with depth. Most water probably moves through the highly conductive upper 200 to 500 feet of the Quaternary basalt. Miocene basaltic and rhyolitic rocks with lower hydraulic conductivity underlie the Quaternary basalt.

Ninety miles of the north wall of the Snake River canyon, from Milner to King Hill, were mapped during this study to determine geologic controls on the location and magnitude of springs issuing from the basalt. Results are being prepared to present as maps and sections in the U.S. Geological Survey Miscellaneous Geologic Investigations series.

Fine-grained Quaternary-Tertiary sediments, which predominate in the western Plain (fig. 50), are as much as 5,000 feet thick near the Idaho-Oregon State line. Alluvial sand and gravel in the Boise River Valley (fig. 49) are several hundred feet thick in places. Quaternary basalt in the central and eastern parts of the western Plain are as much as 2,000 feet thick. Locally, the basalt yields large quantities of water, although saturated thickness is generally less than 500 feet (fig. 51). Miocene basalt, stratigraphically equivalent to the Columbia River Basalt Group, underlies most of the western Plain.

Whitehead (1984b) summarized current understanding of the geohydrologic framework of the Snake River Plain and presented a series of illustrations, including maps, showing saturated and total thicknesses of the Quaternary basalt, geologic sections, resistivity profiles, gravity model profiles, and a table of physical and water-yielding characteristics of rocks underlying the Snake River Plain.

WATER BUDGETS AND GROUND-WATER AND SURFACE-WATER RELATIONS

Only a small part of the water available to the Snake River Plain regional aquifer system originates from precipitation falling on the Plain. Therefore, it was necessary to consider the entire

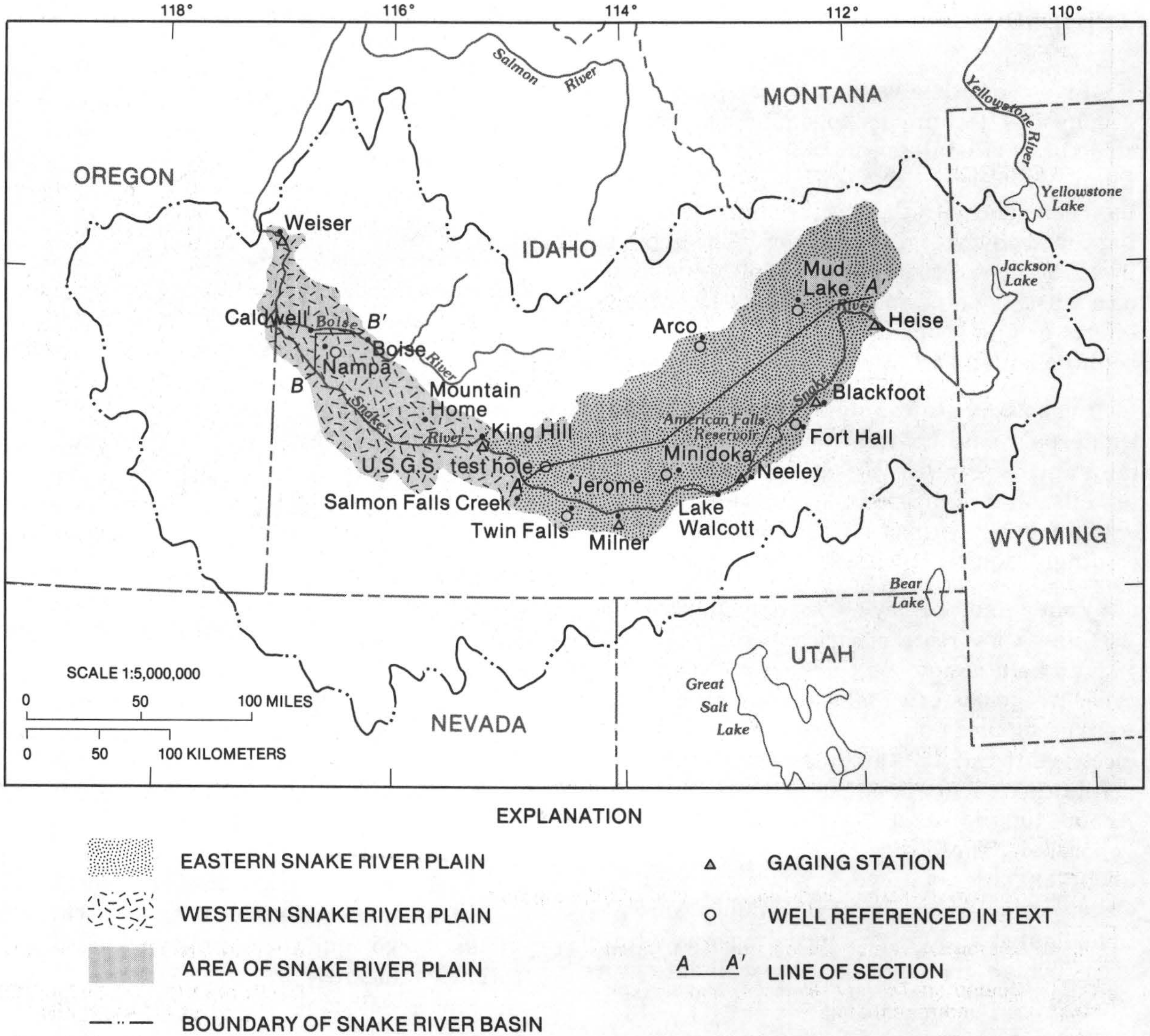


Figure 49.—Study area of the Snake River Plain regional aquifer system.

Snake River drainage basin above Weiser, Idaho, to study the water budget.

Existing hydrologic data were evaluated for adequacy, and additional data needs were identified. Supplemental data were collected on precipitation, streamflow, withdrawals, irrigation return flow, ground-water levels, and spring flows.

Data collection was concentrated in 1980. Water levels were measured in about 1,600 wells in spring of 1980 and repeated in about 800 wells in August 1980. Configuration of the regional water table in spring of 1980 is shown in figure 52. Annual water budgets from 1934 through 1980 for the entire Snake River drainage basin and annual ground-water budgets from 1912 through 1980 for the eastern Plain and from 1930 through 1980 for the western Plain were also estimated and reported by Kjelstrom (1984).

Millions of acre-feet of surface water are diverted annually for irrigation and have greatly increased the amount of water available for ground-water recharge. In one area south of the Snake River near Twin Falls, ground-water levels rose as much as 200 feet in 5 years after the start of irrigation (fig. 53). In much of the eastern Plain, water levels rose several tens of feet over tens of years as indicated by water levels observed in wells in Jerome and near Fort Hall (fig. 53). Water levels in the western Plain also rose in

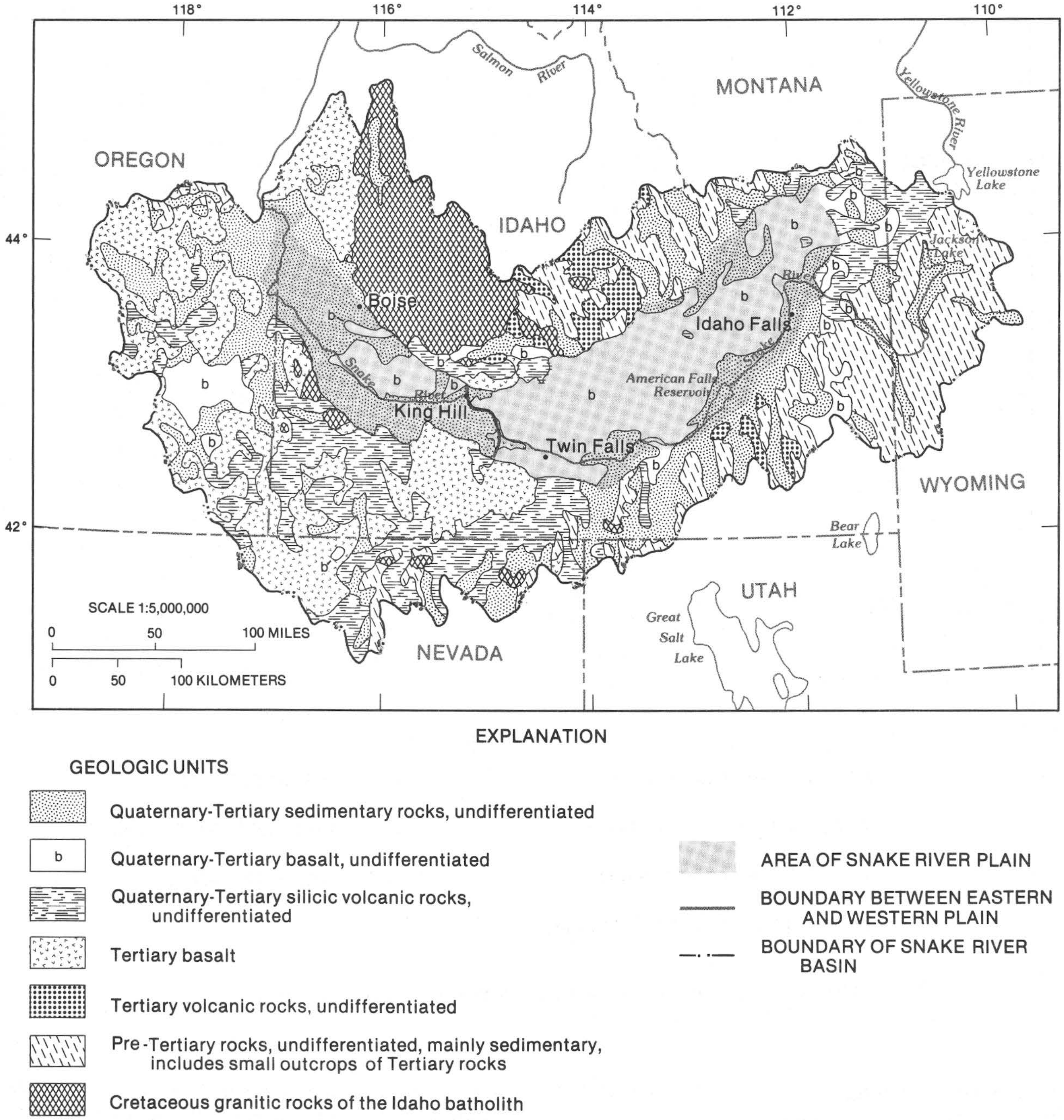


Figure 50.—Generalized geology of the Snake River drainage basin above Weiser, Idaho (modified from Whitehead, 1984b).

response to surface-water irrigation (well near Nampa, fig. 53).

As ground-water levels rose, discharge to rivers increased. Long-term data on gaining reaches of the Snake River that include the major springs are the key to understanding hydrologic changes in the eastern Plain. To supplement existing data, additional spring discharge measurements were made at various places and times during the study. Kjelstrom (1984) documented changes in ground-water discharge to reaches of the Snake River that include major groups of springs (fig. 54).

Discharge from the group of springs between Blackfoot and Neeley (fig. 54) has been relatively constant over the period of record (1912-80). Upgradient from the springs, considerable acreage was irrigated prior to the start of discharge-data collection and, despite an increase in surface-water diversions of about 1.5 million acre-ft from 1912 through 1952, ground-water discharge remained relatively stable. It seems likely that by 1912, ground-water recharge and discharge affecting the reach from Blackfoot to Neeley were approximately in balance.

In the reach from Milner to King Hill, some increase in ground-water discharge probably occurred between the start of irrigation (1880's) and the start of hydrologic-data collection (1902). A 40-year trend of increased ground-water discharge started in 1912 (fig. 54), the result of major increases in surface water irrigated acreage following completion of reservoirs and diversion structures on the Snake River. In the early 1950's, spring discharge stabilized, implying a temporary balance between recharge and discharge that lasted for several years. A period of overall decrease in ground-water discharge followed and continued to 1980. The decrease probably is due to a combination of factors, including increase in withdrawals of ground water, decrease in diversions of surface water, increase in efficiency of irrigation (largely the use of sprinklers), and climatological changes.

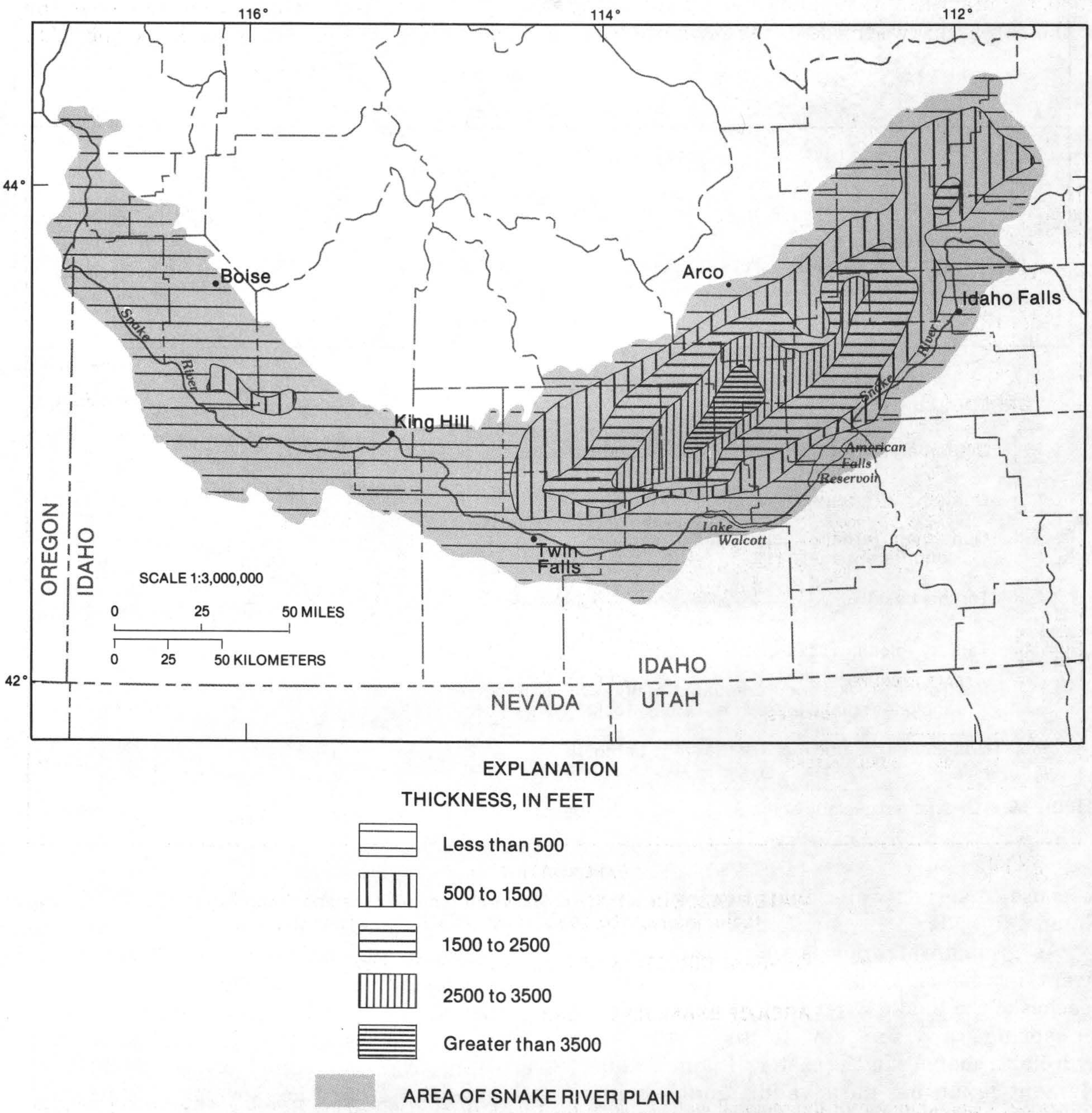


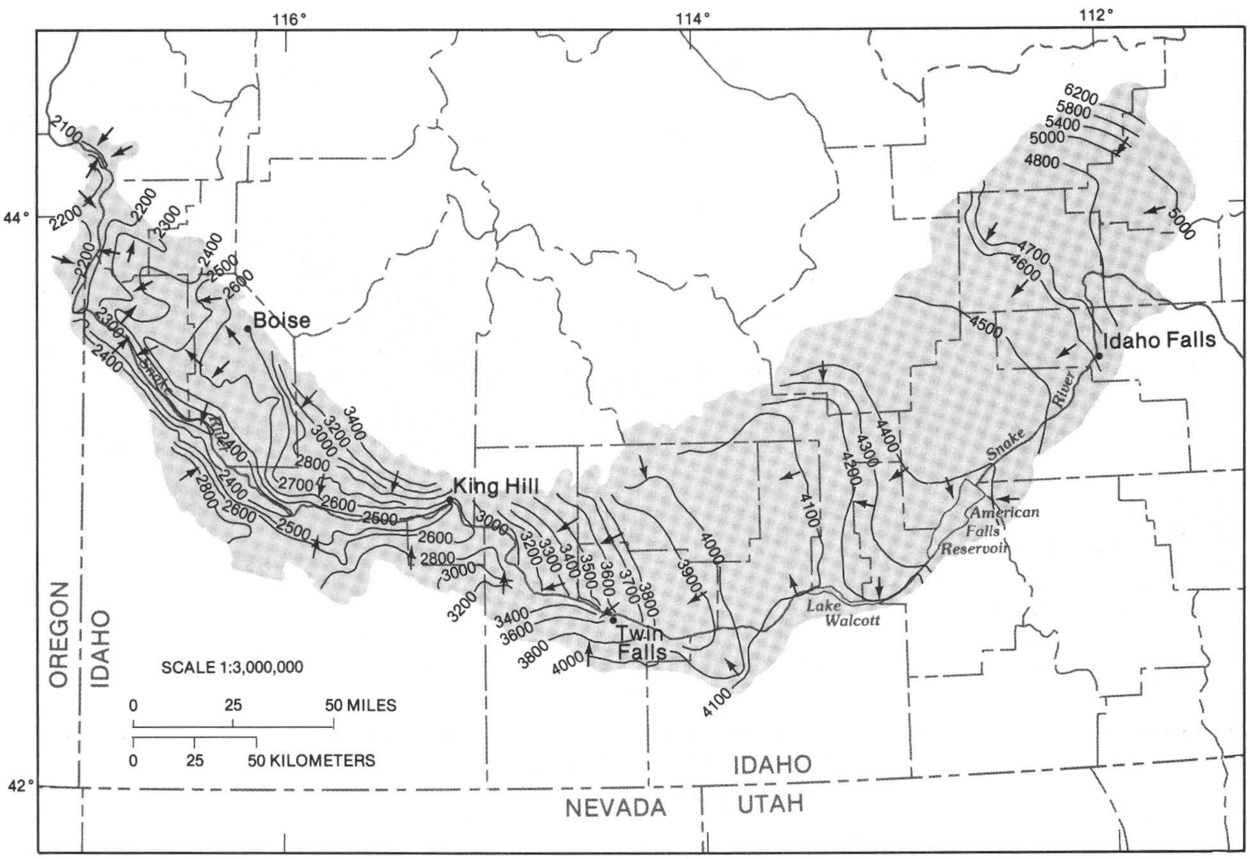
Figure 51.—Thickness of saturated Quaternary basalt underlying the Snake River Plain (modified from Whitehead, 1984b).

Water-level data for the eastern Plain are sparse prior to the 1950's. Therefore, correlation of pre-1950 water-level data with ground-water discharge is not possible. However, since the 1950's, there is a good correlation between ground-water levels and ground-water discharge to the reach of the Snake River from Milner to King Hill (fig. 55).

The net result of 100 successive years of irrigation on the eastern Plain is shown by the ground-water budgets for 1880 and 1980 (fig. 56). Before irrigation, about two-thirds of the total recharge was drainage from tributary basins. In 1980, tributary basins supplied about 20 percent of the recharge; two-thirds of the recharge was percolation of excess surface water diverted for irrigation. The remainder was from precipitation on the Plain and Snake River losses. The imbalance between recharge and discharge in 1980 (400,000 acre-ft) was water taken from aquifer storage. From 1880 to 1980, the amount of annual ground-water recharge increased from about 3.2 to 6.7 million acre-ft.

WATER QUALITY AND GEOCHEMISTRY

Several thousand surface- and ground-water samples have been collected on and near the Snake River Plain. Analyses from the U.S.



EXPLANATION

—4000— WATER-TABLE CONTOUR -- Shows altitude of the regional water table. Interval 100, 200, and 400 feet. Datum is sea level

GENERAL DIRECTION OF GROUND-WATER FLOW

AREA OF SNAKE RIVER PLAIN

Figure 52.—Configuration of the regional water table in the Snake River Plain, spring 1980 (modified from Lindholm and others, 1983).

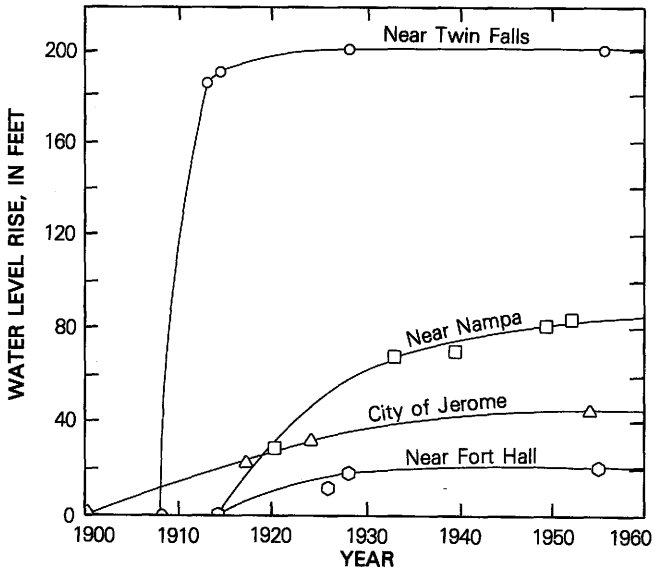


Figure 53.—Changes in Snake River Plain ground-water levels due to surface-water irrigation (locations shown in fig. 49).

Geological Survey files and published reports were assembled and used to determine the relative proportion of major ions in streams entering the Plain and the range of solute concentrations and areal distribution of selected constituents in ground water. Previous investigations emphasized local variations, both temporal and spatial, in water quality. The approach taken during this study was to describe regional variations of ion concentrations in water and determine how water reacts with the rocks through which it flows.

Ion concentrations in surface water flowing onto the Plain vary, depending on rock types in the tributary basin. Surface water from areas underlain by granitic rocks of the Idaho batholith is typically the least mineralized; surface water from areas underlain by pre-Cretaceous carbonate rocks is the most mineralized.

Concentrations of dissolved solids are less than 400 milligrams per liter (mg/L) at 78 percent of the 1,123 ground-water and spring sampling sites across the Snake River Plain. The areal distribution of concentrations of dissolved solids is shown in figure 57. Concentrations of dissolved solids are least in areas of exposed basalt and thin soil cover in the eastern Plain and greatest in areas of fine-grained sedimentary rocks and in the intensively irrigated areas. Concentrations of chloride are less than 20 mg/L at 60 percent of 1,649 sampling sites. Areas of high and low chloride concentrations generally coincide with areas of high and low dissolved solids (W. H. Low, U.S. Geological Survey, written commun., 1984).

Geochemistry of the aquifer system in the eastern Plain was defined by four indirect methods: (1) identification of minerals in the weathered rocks, (2) comparison of thermodynamic mineral indices with plausible ion reactions, (3) comparison of stable-isotope ratios in ground water with weathered mineral products from the rocks, and (4) calculation of chemical budgets.

Water and rock samples were collected in areas of recharge, from intermediate points along ground-water flowpaths, and in areas of discharge. Water samples were analyzed for common cations and anions, trace metals, and stable isotopes. Rock samples from outcrops and the subsurface were analyzed to determine mineralogic composition and weathering products.

Solutes in ground water in the eastern Plain are generally in equilibrium with minerals in the rock framework. Alteration minerals in the rock framework were identified by petrographic microscope, scanning-electron microscope, and energy-dispersive X-ray fluorescence methods. Chemical equations were written for weathering of the minerals identified and compared with water-mineral saturation indices. The validity of the proposed reactions was evaluated using isotopes of carbon and sulfur, which comprise 80 percent of the anions in most of the water samples.

The most plausible reactions controlling the ion concentrations appear to be the weathering of plagioclase to clay minerals; dissolution of olivine, anhydrite, and pyrite; and precipitation of calcite and quartz. About 25 percent of the solute load is from weathering of the rock framework or from man's activities; the remainder is in recharge water from tributary drainage basins.

The same geochemical processes may take place in the western Plain but were not defined. Concentrations of dissolved solids in thermal waters in the western Plain are similar to those in the overlying cold-water system. Reactions controlling solutes in thermal waters are also similar, but ion exchange processes produce higher concentrations of sodium and lower concentrations of calcium and magnesium; the pH is also higher. Preliminary results of carbon-14 age dating of water from selected geothermal wells suggest that residence time of some geothermal waters may be about 17,700 ±4,000 years.

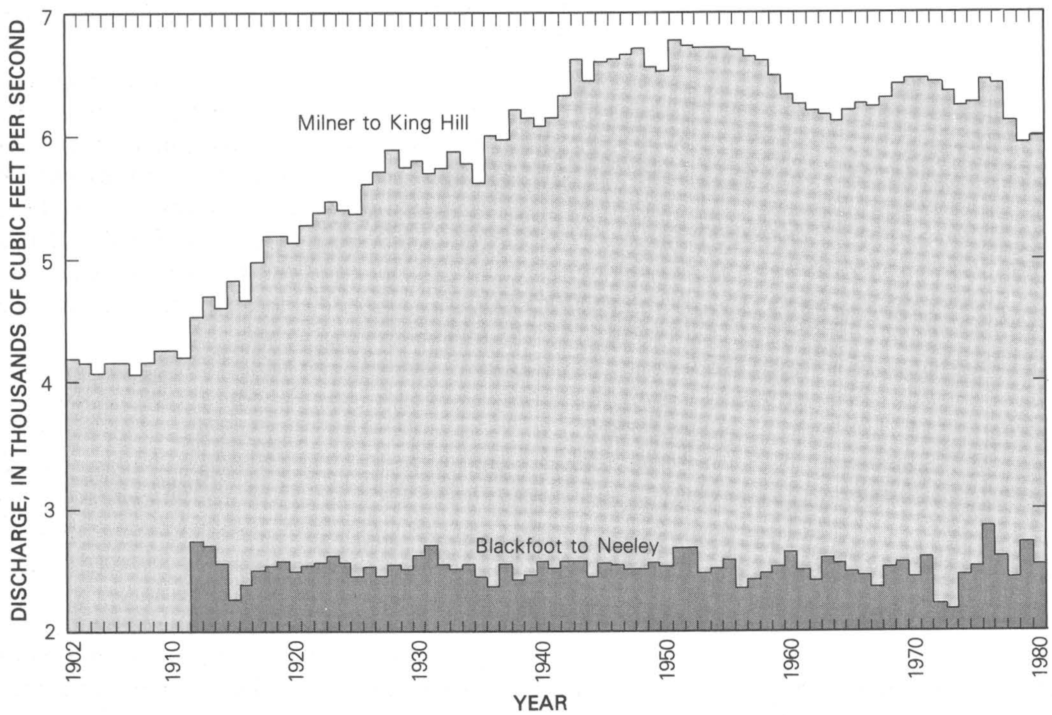


Figure 54.—Ground-water discharge to the Snake River from the eastern part of the Snake River Plain regional aquifer system (locations shown in fig. 49).

WATER USE

Irrigation is by far the largest consumptive use of water on the Snake River Plain. Therefore, the delineation of past and present irrigated areas was needed to determine stresses on the hydrologic system and to calculate recharge rates for ground-water flow simulation models.

Irrigation on the Snake River Plain started along the Boise River in about 1843. Development was slow until the 1880's, when Congress encouraged reclamation of desert lands and expansion of irrigated acreage in the arid West. The Desert Land Act of 1877, a survey of irrigable lands in 1889, and the Carey Act of 1894 provided incentive, and rapid irrigation development followed. By 1899, about 550,000 acres were irrigated with surface water. The Reclamation Act of 1902 stimulated further expansion by providing funds for construction of major instream storage facilities and canals. By 1929, about 2.2 million acres were irrigated with surface water. Increases in irrigated lands from 1929 through 1945 were moderate, owing to depressed economic conditions, World War II, and full appropriation of surface-water rights. In the 1940's, demand for irrigation water increased, and because surface-

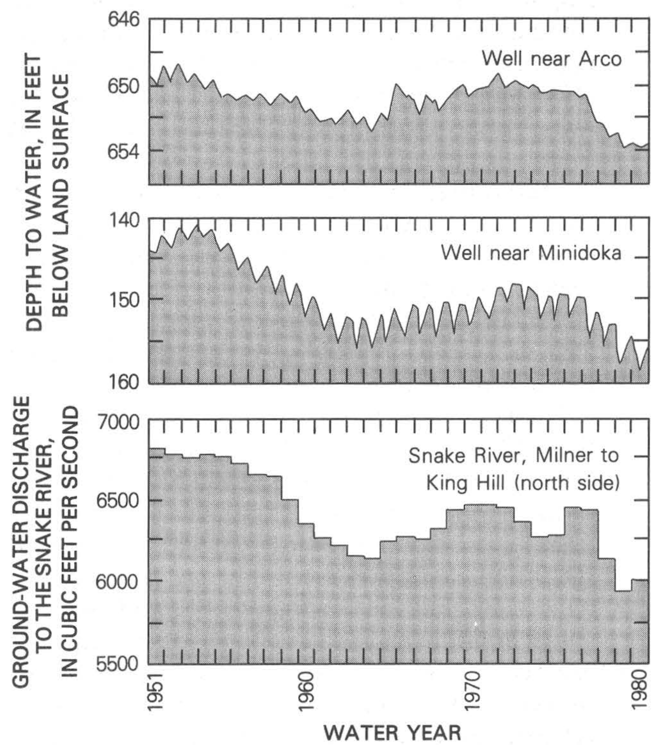


Figure 55.—Relation of ground-water levels to ground-water discharge from the eastern part of the Snake River Plain regional aquifer system (locations shown in fig. 49).

water rights were fully appropriated, ground water became a logical source of supply. By 1966, ground water was used to irrigate about 700,000 acres. In the 1960's, installation of large pumping stations on the Snake River began. Where the river is deeply entrenched, water is lifted several hundred feet to supply the adjacent irrigated uplands.

In 1980, about 3.1 million acres of land, or 32 percent of the Snake River Plain, were irrigated (Lindholm and Goodell, 1984). About 13 million acre-ft of surface water were withdrawn to irrigate 2.0 million acres. Another 2.3 million acre-ft were pumped from 5,300 wells to irrigate 1.0 million acres. The remaining 0.1 million acres were irrigated with combined surface and ground water.

Acreage irrigated in 1980 was determined from single-date digital multispectral scanner (Landsat) data by the Idaho Department of Water Resources Image Analysis Facility through a cooperative agreement with the U.S. Geological Survey. Source of water (surface, ground, or combined) was determined from published data, aerial photographs, and field checking. Historical acreages were estimated from published maps for 1899, 1929, 1945, and 1966. Although methods of defining irrigated areas varied, the location and relative numbers of acres determined are adequate for comparisons.

Ground-water and river pumpage in 1980 were estimated using electrical power-consumption data because nearly all pumps on the Plain are electrically powered. Field measurements were made on 79 wells and river pumping stations in 1980 to determine the power consumption per acre-foot of water pumped per foot of lift. On the eastern Plain, river pumping stations are typically 100 to 500 horsepower, pumping lifts from the Snake River are less than 200 feet, and the median drawdown in wells is 6 feet (Bigelow and others, 1984). Water commonly is distributed by sprinklers. On the western Plain, most irrigation wells are completed in sand and gravel aquifers that yield a moderate supply. River pumping stations are generally less than 100 horsepower on the western Plain, pumping lifts from the Snake River are similar to those on the eastern Plain but in places are as much as 500 feet, and median drawdown in wells is 56 feet. Gravity irrigation is more common on the Western Plain.

For comparative purposes, volumes of ground water pumped in 1980 were grouped into

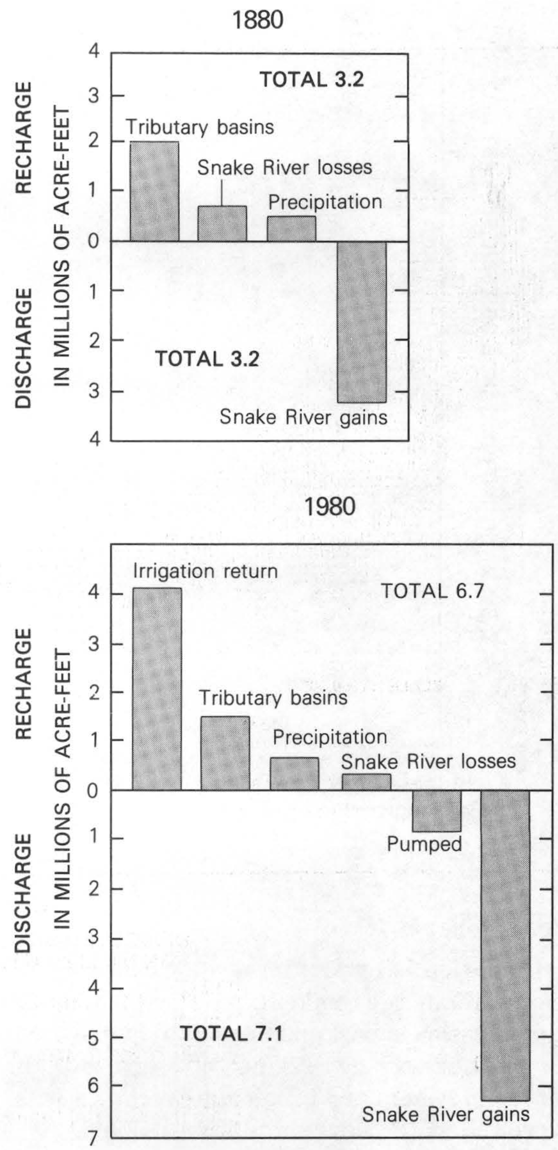


Figure 56.—Changes in ground-water budget, 1880 and 1980, the eastern part of the Snake River Plain regional aquifer system (area south of the Snake River from Neeley to Salmon Falls Creek excluded).

15-minute blocks (fig. 58). Ground-water pumpage is greatest on the eastern Plain in areas where surface water is not available and where topography limits irrigation to locally supplied sprinkler systems.

Other consumptive uses of water (public supply, industrial, rural, domestic, and stock) are small relative to irrigation. The quantity of water use was obtained from the State-Federal Water Use data base or was estimated. Hydroelectric power generation is the largest nonconsumptive use of water.

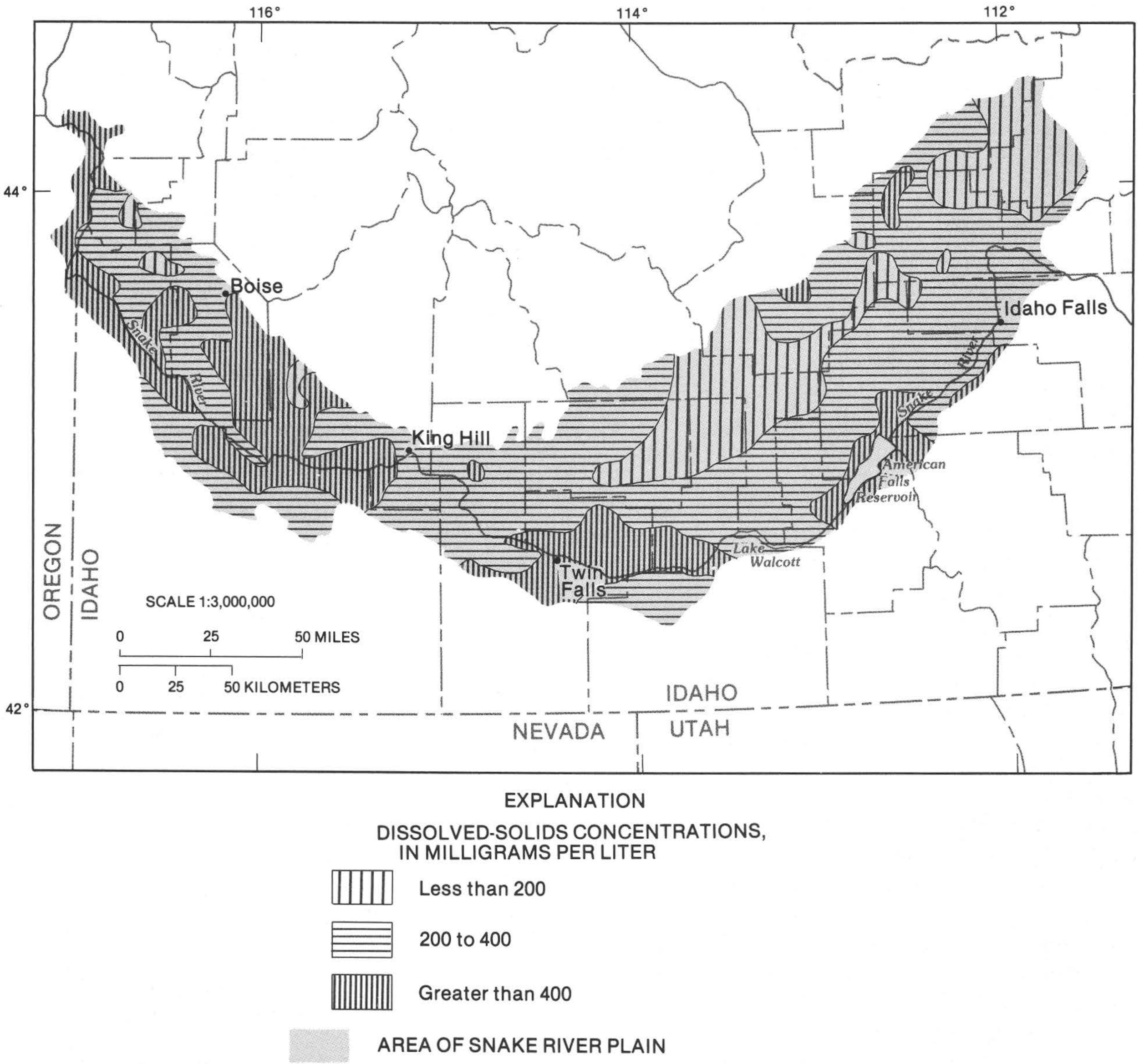


Figure 57.—Areal distribution of dissolved-solids concentrations in ground water in the Snake River Plain regional aquifer system.

SIMULATION OF GROUND-WATER FLOW IN THE EASTERN PART OF THE SNAKE RIVER PLAIN

A two-dimensional parameter estimation model (Garabedian, 1984) with 4 mile × 4 mile cells, which incorporated a nonlinear least-square regression technique, was used to estimate transmissivity, leakance, and boundary fluxes of the regional aquifer system in the eastern Plain. After the parameters were estimated with the parameter-estimation model, a four-layered, three-dimensional, finite-difference model (Trescott, 1976; Trescott and Larson, 1976) with the same cell size was used to simulate ground-water flow in the eastern part of the Snake River Plain (S. P. Garabedian, U.S. Geological Survey, written commun., 1984). Boundaries of the models were no-flow, constant-flux, and head dependent rivers and springs. It was assumed that there is no underflow between eastern and western parts of the Snake River Plain, owing to the apparent sink formed by the Snake River. Based on geologic and hydrologic conditions, the eastern Plain was then artificially subdivided into 40

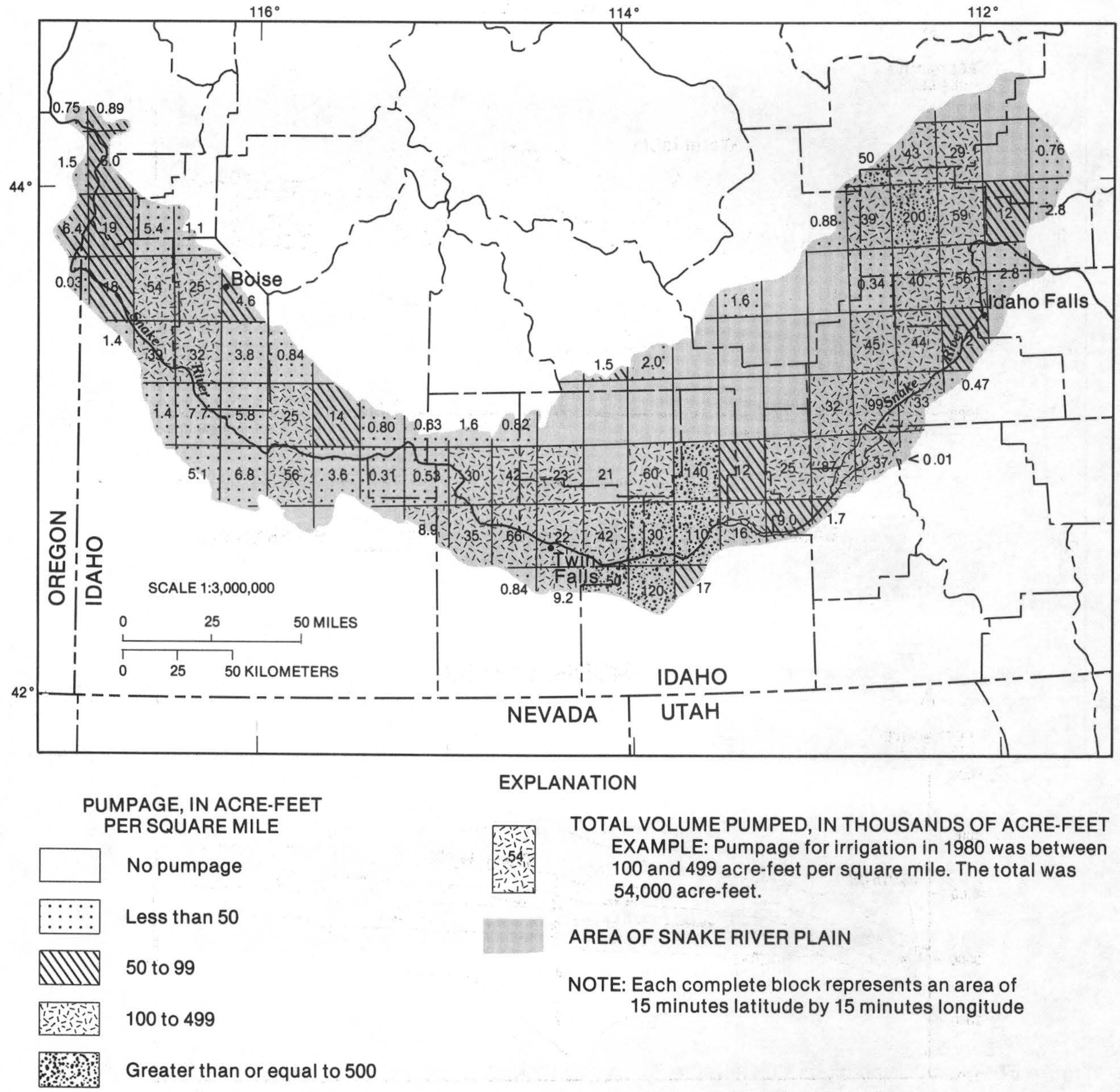


Figure 58.—Estimated 1980 ground-water pumpage from the Snake River Plain regional aquifer system (modified from Bigelow and others, 1984).

zones. Parameter values within each zone were assumed to be uniform. Throughout the modeling process, special efforts were made to keep values of geohydrologic parameters within reasonable ranges, consistent with what is known.

The two-dimensional parameter estimation model suggested that transmissivity ranges from 0.05 to 44 ft^2/s. The values are generally consistent with the ranges estimated by Mundorff and others (1964) using a flow-net analysis, by Norvitch and others (1969) using an analog model, and by Newton (1978) using a digital flow model. Along the margins of the Plain, and in the central part where better head control was available during this study, transmissivity values obtained by the parameter-estimation model were lower than those obtained previously.

Simulated and measured ground-water discharge to subreaches of the Snake River were similar. The coefficient of correlation between simulated and measured hydraulic head is 0.996 and standard error of estimates of head is 40 feet.

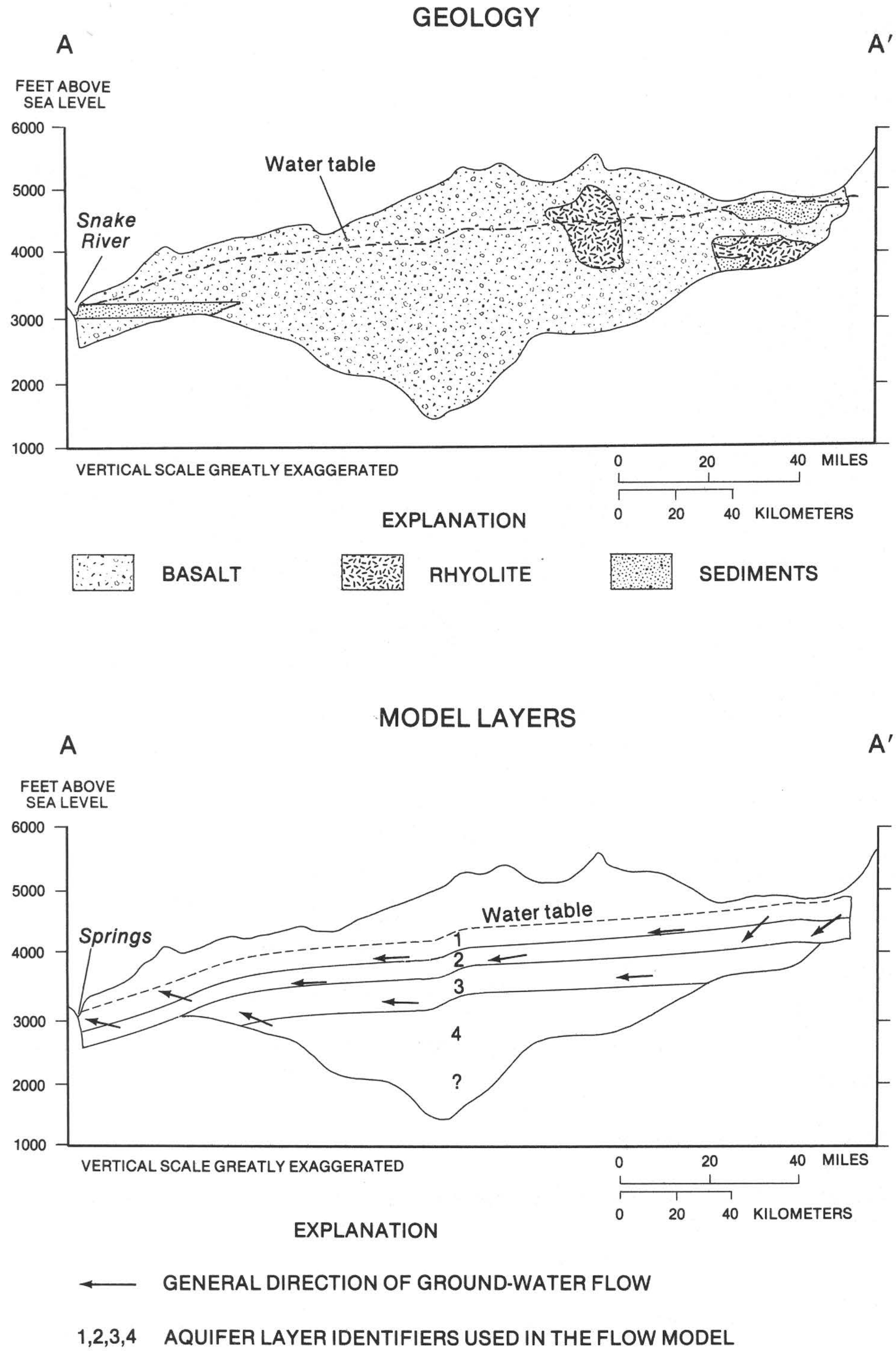


Figure 59.—Generalized geologic section and conceptual model, eastern part of the Snake River Plain regional aquifer system (location of section shown in fig. 49).

The flow model is more sensitive to changes in recharge than to changes in transmissivity.

Parameter values estimated from the calibrated two-dimensional parameter estimation model were the initial input data for the four-layered, three-dimensional steady-state and transient flow models. The basalt aquifer was artificially subdivided into four aquifer layers in order to simulate head changes with depth (fig. 59). Modeled aquifer layers 1 and 2 are 200 and

300 feet thick and represented the basalts having the highest hydraulic conductivity. Most wells are completed in the upper 200 feet; consequently, less is known about hydraulic conditions in layer 2 and little is known about conditions in layers 3 and 4. On the basis of scant evidence, simulated hydraulic conductivities in the modeled aquifer layers 3 and 4 were reduced by one-third and two-thirds of the observed values.

Simulated transmissivities for modeled aquifer layer 1 are shown in figure 60. Transmissivity is greatest in the central part of the eastern Plain. The area of relatively low transmissivity comprising model zones 12 and 22 (fig. 60) probably is related to a major rift zone; the low transmissivity area comprising zone 15 is composed of sediments interlayered with basalt.

The model was tested for anisotropy of the basalt aquifer. Although inconclusive, test results suggest that transmissivity is slightly greater in the direction perpendicular to the longitudinal

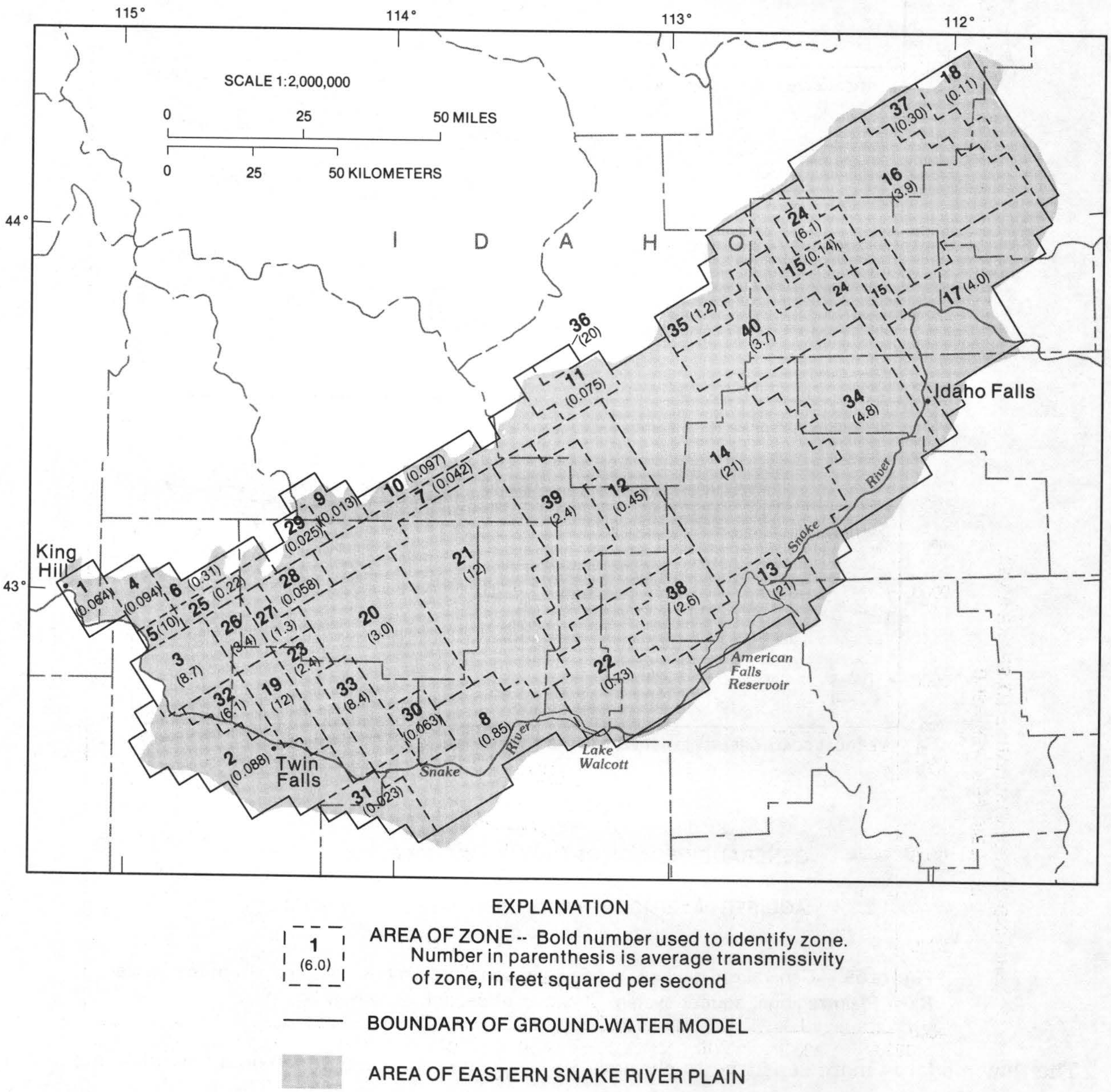


Figure 60.—Simulated transmissivities for modeled aquifer layer 1, eastern part of the Snake River Plain regional aquifer system.

axis of the eastern Plain. The results are consistent with geologic evidence; fractures and faults in southeast-trending rift zones may increase hydraulic conductivity in that direction.

The storage coefficient of aquifer layer 1 varies from 0.05 to 0.20 according to the rock type. Aquifer layers 2, 3, and 4 were assigned a uniform storage coefficient of 0.0001.

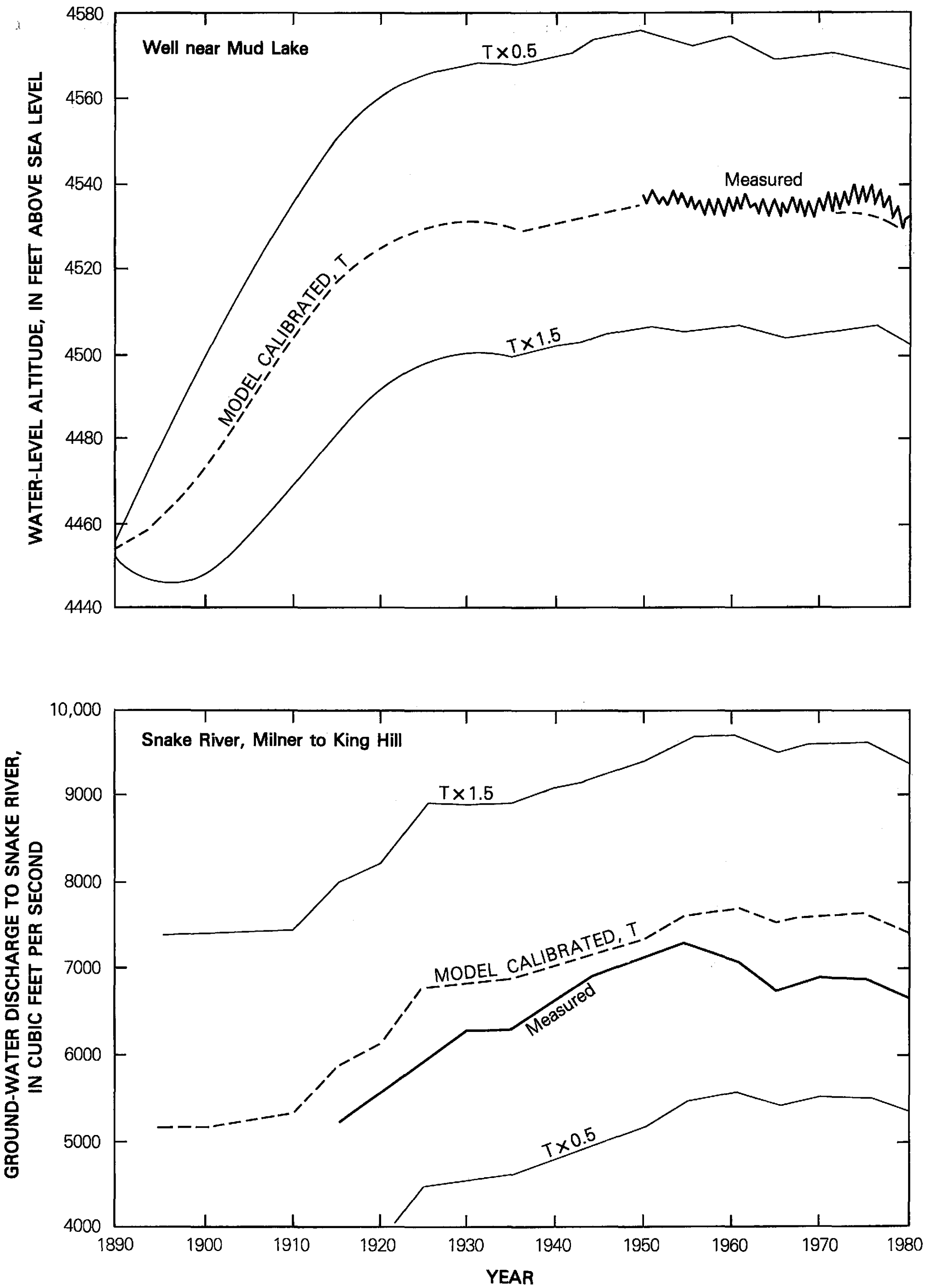


Figure 61.—Simulated and measured water levels in a well near Mud Lake, ground-water discharge to the Snake River from Milner to King Hill, and model sensitivity to changes in transmissivities, eastern part of the Snake River Plain regional aquifer system.

Transient simulations, using 5-year time steps from 1890 to 1980, were made to determine hydrologic changes in the eastern Plain resulting from continued use of surface water for irrigation. Simulated and measured water levels and discharge to subreaches of the Snake River were comparable (fig. 61). Model response to changes in transmissivity is also shown in figure 61.

The flow model was used to simulate pre-irrigation hydrologic conditions and to determine probable aquifer response to hypothetical future water-resource development in the eastern Plain.

If 1980 conditions of recharge and discharge are extended to the year 2010, the model indicates that water levels may decline 2 to 8 feet in most of the eastern Plain with greater declines along the boundaries. Spring flow may decrease about 5 percent and river leakage into the aquifer may increase 9 percent. As an extreme example, if ground-water withdrawals were increased by 2,400 ft^3/s in order to irrigate another 1 million acres in the eastern Plain, head declines of 10 to 50 feet and decreases in spring flow of 20 percent might be expected within 30 years.

SIMULATION OF GROUND-WATER FLOW IN THE WESTERN PART OF THE SNAKE RIVER PLAIN

Geologic complexities and general lack of data for model input dictated that major emphasis on the western Plain be given to development of a conceptual flow model.

Rather than a single regional aquifer, a number of local aquifers collectively operate as a regional aquifer system. Basalt of the Snake River Group, particularly in the vicinity of Mountain Home, yields large quantities of water, but areal extent is small. Alluvial sand and gravel aquifers in the Boise River Valley yield moderate to large amounts of water to wells. Elsewhere, sand lenses in the predominantly fine-grained Quaternary-Tertiary Bruneau Formation and Idaho Group are the best aquifers. Thermal water in the Tertiary Banbury Basalt, Idavada Volcanics, and Columbia River Basalt Group is confined. The thermal water is used mainly for space heating in the Boise area and for irrigation south of the Snake River.

Natural recharge to shallow aquifers is from tributary basin underflow, leakage of streams that drain the tributary basins, and direct precipitation. Deeper aquifers are recharged laterally from faulted margins of the Plain and by vertical leakage. As in the eastern Plain, use of surface water for irrigation for more than 100 years has effected major changes in the hydrologic system. In 1980, infiltration of surface water applied for irrigation provided about two-thirds of the recharge. Since the start of irrigation, ground-water levels in the Boise River Valley have risen several tens of feet (well near Nampa, fig. 53). In the lower Boise River Valley, high water levels necessitate drainage ditches to relieve waterlogging. Ground water in the western Plain also discharges to the Snake River and its tributaries. Owing to the magnitude of flow in the Snake River, estimation of river gains is difficult at best.

A three-layer finite-difference flow model (Trescott, 1976; Trescott and Larson, 1976) with 2 mile × 2 mile grid cells was developed by G. D. Newton (U.S. Geological Survey, written commun., 1984) to simulate the conceptualized system in the western Plain (fig. 62). The steady-state model was calibrated to 1980 hydrologic conditions. The transient model simulated 1890-1980 conditions in 10-year time increments. Transient calibration helped refine initial estimates of transmissivity, recharge and discharge rates, specific yield, and vertical hydraulic conductivity. Calibration in 10-year increments was uncertain because long-term hydrologic data were generally lacking. Therefore, the model also was calibrated to several successive 1-year periods using monthly time steps. All rivers were modeled as leaky connections to the aquifer system; canals and drains in the lower Boise River Valley were similarly modeled.

The top aquifer layer, about 500 feet thick, is largely alluvial sand and gravel west of Boise. East of Boise, the top layer includes basalt of the Snake River Group and Bruneau Formation, and sediments of the Bruneau Formation and Idaho Group. Simulated transmissivities for modeled aquifer layer 1 are highly variable and range from 0.01 to 0.22 ft^2/s (fig. 63). Transmissivity is highest in the Boise River Valley and lowest along the margins of the western Plain where fine-grained sediments predominate.

Modeled aquifer layer 2 represents several thousand feet of predominantly fine-grained sediments. The estimated transmissivity is about 0.005 ft^2/s.

Modeled aquifer layer 3 represents Tertiary volcanic rocks, some of which are exposed along the boundaries of the western Plain. Thickness of the volcanic rocks is largely unknown but is assumed to be several thousand feet. Estimated transmissivity ranges from about 0.01 to 0.2 ft^2/s. Water in modeled aquifer layer 3 is confined by the fine-grained sediments in modeled layer 2,

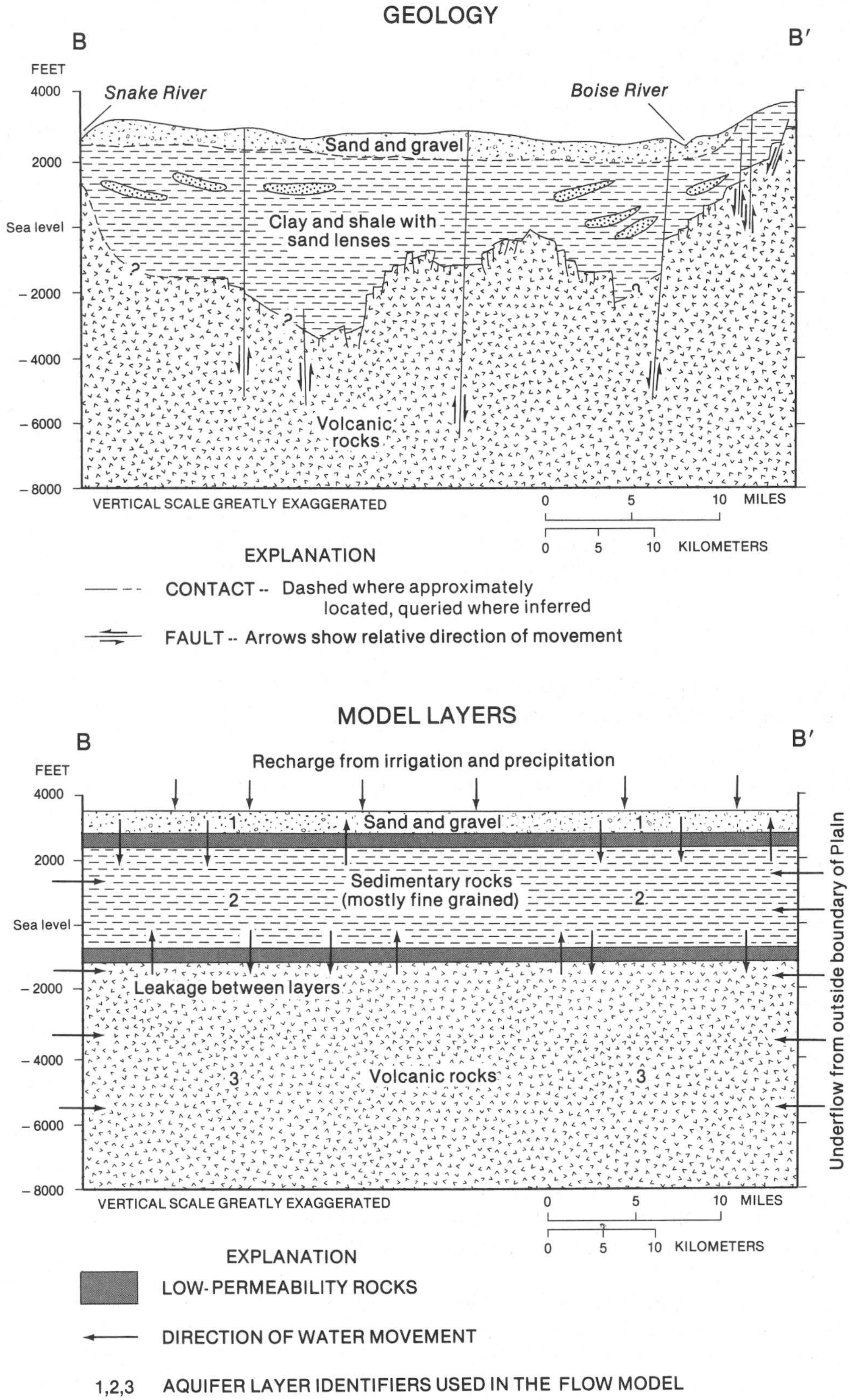


Figure 62.—Generalized geologic section and conceptual model, western part of the Snake River Plain regional aquifer system (location of section shown in fig. 49).

which have an average vertical hydraulic conductivity of about 3×10^{-9} ft/s. Although hydraulic heads are high in layer 3, the volume of recharge to upper layers is probably small and is masked by other sources of recharge.

Storage coefficients of 0.10, 0.002, and 0.001 were assigned for layers 1, 2, and 3 to yield the best simulation results. By removing all pumpage and recharge from irrigation return flow resulting from surface-water irrigation, preirrigation hydrologic conditions were simulated. The transient model was used to simulate possible aquifer conditions to the year 2010, using 1980 recharge rates. Model response suggests that future water-level declines can be expected if ground-water withdrawals increase.

SUMMARY

Large quantities of good-quality ground and surface water are available on the Snake River Plain. For study purposes, the Plain was divided into eastern and western parts. As much as 3,500

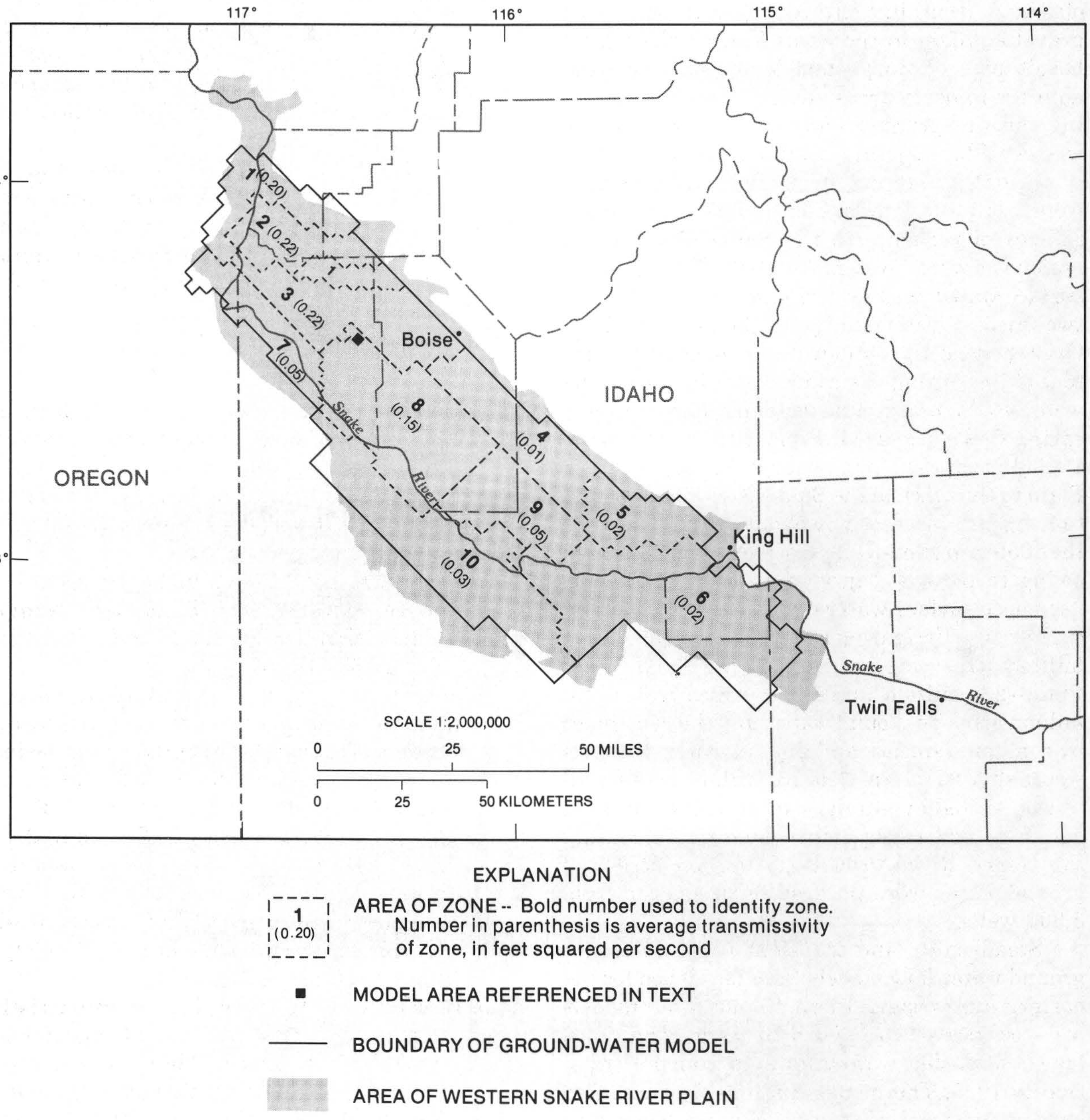


Figure 63.—Simulated transmissivities for modeled aquifer layer 1, western part of the Snake River Plain regional aquifer system.

feet of saturated Quaternary basalt underlie the eastern Plain. The upper 200 feet have the highest hydraulic conductivity, and estimated transmissivity ranges from 0.05 to 44 ft^2/s. An estimated 200 to 300 million acre-ft of water are stored in the upper 500 feet. The thickness of the basalt aquifer was estimated largely from electrical-resistivity soundings. Interpretations of surface geophysical data were checked by drilling a 1,123-foot test hole. In the western Plain, generally fine-grained Tertiary sedimentary rocks predominate; water in the western Plain is obtained from unconfined alluvial sand and gravel aquifers in the Boise River Valley, from basalt east of Boise, and from confined sand aquifers in other areas. Volcanic rocks underlying the fine-grained sedimentary rocks in the western Plain contain confined thermal water.

Prior to irrigation, streamflow and underflow from tributary drainage basins were the major sources of recharge to the Snake River Plain regional aquifer system. In 1980, infiltration of surface water used for irrigation supplied about two-thirds of the recharge in the eastern Plain. Over the years, ground-water levels rose several tens of feet, owing to surface-water irrigation. As water levels rose, ground-water discharge, largely spring flow, increased. From 1912 to the early 1950's, ground-water discharge from the eastern Plain to the reach of the Snake River from Milner to King Hill increased by about 2,600 ft^3/s. Since the 1950's, ground-water discharge has decreased owing to increased ground-water withdrawals, decreased surface-water diversions, and more efficient use of irrigation water. In 1980, about 3.1 million acres were irrigated on the Snake River Plain—2.0 million acres by surface water, 1.0 million acres by ground water, and the remainder by combined surface and ground water. Surface-water diversions totaled 13 million acre-ft; all were diverted by gravity except about 1 million acre-ft, which were pumped from rivers, mainly the Snake River. About 2.3 million acre-ft of ground water were pumped for irrigation from 5,300 wells.

Steady-state and transient finite-difference ground-water flow models were developed for the eastern and western Plain. Steady-state models were calibrated to 1980 hydrologic conditions; transient models were calibrated from preirrigation to 1980. The models reasonably simulated current and past hydrologic conditions. They were used to evaluate aquifer response to possible future ground-water development. Should ground-water withdrawals increase, additional water-level declines and decreases in ground-water discharge to the Snake River are expected.

Water quality is generally good. Most solutes originate in tributary basins, and concentrations of ions change little as water flows from areas of recharge to areas of discharge.

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SOUTHWEST ALLUVIAL-BASIN REGIONAL AQUIFER-SYSTEMS STUDY

Hundreds of isolated alluvial basins are in the southwestern United States. Some of them have surface-water outflow, others do not. It is impossible to study all alluvial-basin aquifer systems at once. Therefore, basin aquifers were selected to be studied under the RASA Program to understand the occurrence, movement, and water quality of ground water in these alluvial-basin aquifer systems. It is hoped that the results of the study can be transferred to other basin aquifer systems which need to be evaluated in the future. Geographically, the study was divided into two parts. Alluvial basins in parts of Colorado, New Mexico, and Texas were investigated by the U.S. Geological Survey in Albuquerque, NM, and alluvial basins in parts of southern and central Arizona and parts of adjacent States were investigated by the Geological Survey in Tucson, AZ. Following are discussions of these studies.

STUDY IN PARTS OF COLORADO, NEW MEXICO, AND TEXAS

By David W. Wilkins, Albuquerque, New Mexico

INTRODUCTION

The study of aquifer systems underlying the southwest alluvial basins in parts of Colorado, New Mexico, and Texas was started in 1978 and completed in 1984, except for report writing. The study covers a total area of 70,000 mi^2 within or adjacent to three physiographic provinces (fig. 64). The northern part of the study area is in the Southern Rocky Mountains Province, the central part is in the Basin and Range Province, and the west-central part is in the Colorado Plateau Province. The Great Plains Province is east of the study area. Except for a small area in its southwest corner, the study area is bounded by the Continental Divide on the west.

GENERAL GEOLOGY AND HYDROLOGY

Two types of basins occur in the study area: (1) open basins are within the Rio Grande rift, (2) closed basins are predominantly in southwest New Mexico and west Texas having no surface-water outflow (fig. 65). The Rio Grande rift is a fault-bounded structural feature with uplifted blocks on the east and on the west. Uplifted blocks to the east of the basins generally rise several thousand feet above the valley floor of the basins. The basins are bounded on the north, east, and west mainly by Quaternary and Tertiary volcanics, and Mesozoic and Paleozoic rocks (fig. 66). The southern boundary of the study area is not defined by structural or hydrologic boundaries but is arbitrarily placed at the Mexico-United States International Boundary. Alluvial sediments in the basins were derived from the surrounding highlands and mountains and are of Cenozoic age. These are composed of flood-plain deposits and sediments of the Santa Fe Group. The Santa Fe Group is a rock-stratigraphic unit that consists of unconsolidated to moderately consolidated lenticular deposits of gravel, sand, and clay interbedded in some places with volcanics. Most municipal and industrial wells in the study area are completed in the Santa Fe Group. Wells in these sediments may produce up to 3,000 gal/min.

Precipitation in the uplifted mountainous blocks east and west of the basins is high and is the source of the surface water which eventually recharges the aquifers near the base of the mountains. Very little of the surface runoff is debouching onto the alluvial fans because of high rates of infiltration through arroyos. Most recharge to aquifers underlying the basins is near mountain fronts. The other major recharge in the open basins is the seepage of the Rio Grande river and irrigation return flows. The regional potentiometric surface of the alluvial basins (fig. 67) indicates that ground-water flow in most basins is from the east and west margins of the basins toward the basin axis and then southward. Exceptions to this flow pattern are the Animas Basin and lower part of the Salt Basin. Within these basins, flow is northward (figs. 64 and 67). Wilkins (1984) has described the ground-water flow system and the aquifer characteristics of each basin.

Quality of ground water changes areally and vertically. Kelly (1974) reported that ground water underlying the Rio Grande river from near Espanola to east of Socorro, NM, has concentrations of dissolved solids less than 1,000 milligrams per liter (mg/L) from land surface to a depth of about 2,000 feet. Below this depth, concentrations of dissolved solids are as much as 3,000 mg/L. Concentrations of dissolved solids increase to a range of 3,000 to 10,000 mg/L below about 1,900 feet in the Jornada del Muerto Basin to the south (location of the Basin is shown in fig. 64). A vertical profile along the Rio Grande in the Mesilla Basin, further to the south, shows water with concentrations of dissolved solids from 1,000 to 3,000 mg/L in the flood-plain sediments. Within 100 feet below the water table, concentrations of dissolved solids normally are less than 1,000 mg/L; however, concentrations increase with depth. At a depth of about 4,000 feet below the water table, concentrations of dissolved solids may range from 3,000 to 10,000 mg/L (Wilson and others, 1981). Recharge waters are characterized by a calcium bicarbonate type such as that found in the San Luis and Espanola Basins (figs. 64 and 68). A sodium bicarbonate type water is found in the central part of many closed basins and at the lower end of some open basins (figs. 64, 65, and 68). The predominant chemical process responsible for sodium bicarbonate type water is probably ion exchange on clay minerals. In many sodium bicarbonate type water areas a fine-grained lithology is present. Calcium or magnesium sulfate or chloride type waters probably result from the dissolution of gypsum or salts. These latter three types of water are found in many closed basins. Sodium sulfate or sodium chloride type waters are also found in many closed basins.

GEOLOGIC STUDIES

Geologic studies were initiated early during the study with the objective of selecting representative basins and of characterizing these selected basins. Basin boundaries were delineated on the basis of bedrocks or faults that separated the basins into distinct hydrologic areas. Topographic and surface-water divides were also considered. This process resulted in dividing the alluvial basins in the study area into 22 basins, shown in figure 64.

The characterization studies were based upon driller's logs, geophysical logs, and published reports. Estimates of percent sand and clay were derived from the logs and were used to estimate the relative permeability of the sediments which generally are correlated with the permeability values resulting from calibrated ground-water flow models.

Subsurface configuration of the basins could not be defined with available data. Surface geophysical surveys were conducted in 16 basins. Birch (1980a, 1980b) interpreted existing gravity data to define volume of sediments and bottom of alluvial material. Birch (1980a and b) estimated the average thickness of sediments in the Albuquerque-Belen Basin probably is about 4,920 feet; 2,400 feet in Mesilla Basin; 1,600 feet in Mimbres Basin; 2,600 feet in Tularosa-Hueco Basin; 2,400 feet in Jornada del Muerto Basin; 1,600 feet in Palomas Basin; and 2,000 feet in Engle Basin. Seismic refraction data were used to define the size and shape of the basins and to estimate the depth of alluvial sediments.

GEOCHEMICAL STUDIES

Prior to this study, few geochemical studies had been done. Data stored in the WATSTORE file of the U.S. Geological Survey, having identifiable sources of water, were evaluated for cation and anion balance. Four basins, San Luis, Albuquerque-Belen, Socorro, and Mesilla, were selected for geochemical studies and are discussed below.

San Luis Basin.—The geochemical study in the San Luis Basin suggests that chemical data of ground water are insufficient. Data were collected from 15,000 wells in the basin, but only data from 99 wells were useful. Primary conclusions are that ground water in the aquifer north of the Rio Grande changes in water chemistry from calcium bicarbonate to sodium bicarbonate due to cation exchange on clays.

Albuquerque-Belen Basin.—The study in the Albuquerque-Belen Basin verified the existence of a layer of silt and clay at depth on the mesa west of Albuquerque. It is suggested that ground-water flow enters the Albuquerque-Belen Basin from the San Juan Basin to the west. The ground-water flow direction in the southern end of the basin is upward over a bedrock layer, which probably is the basin boundary.

Socorro Basin.—The geochemical study in the Socorro Basin was useful in verifying the suspected mixing of water upwelling along faults with ground water at shallower depths. These upward-flowing waters are geothermal.

Mesilla Basin.—The major geochemical process in the Rio Grande flood plain of the Mesilla Basin is mixing of water flowing from the uplifted recharge areas with water in the flood-plain sediments. The study also confirmed the location and temperature of the geothermal water.

RECHARGE MODELS

Prior to this study, data on aquifer recharge in the basins were not available. Three models were developed to estimate recharge. The "mountain-front recharge model" provides

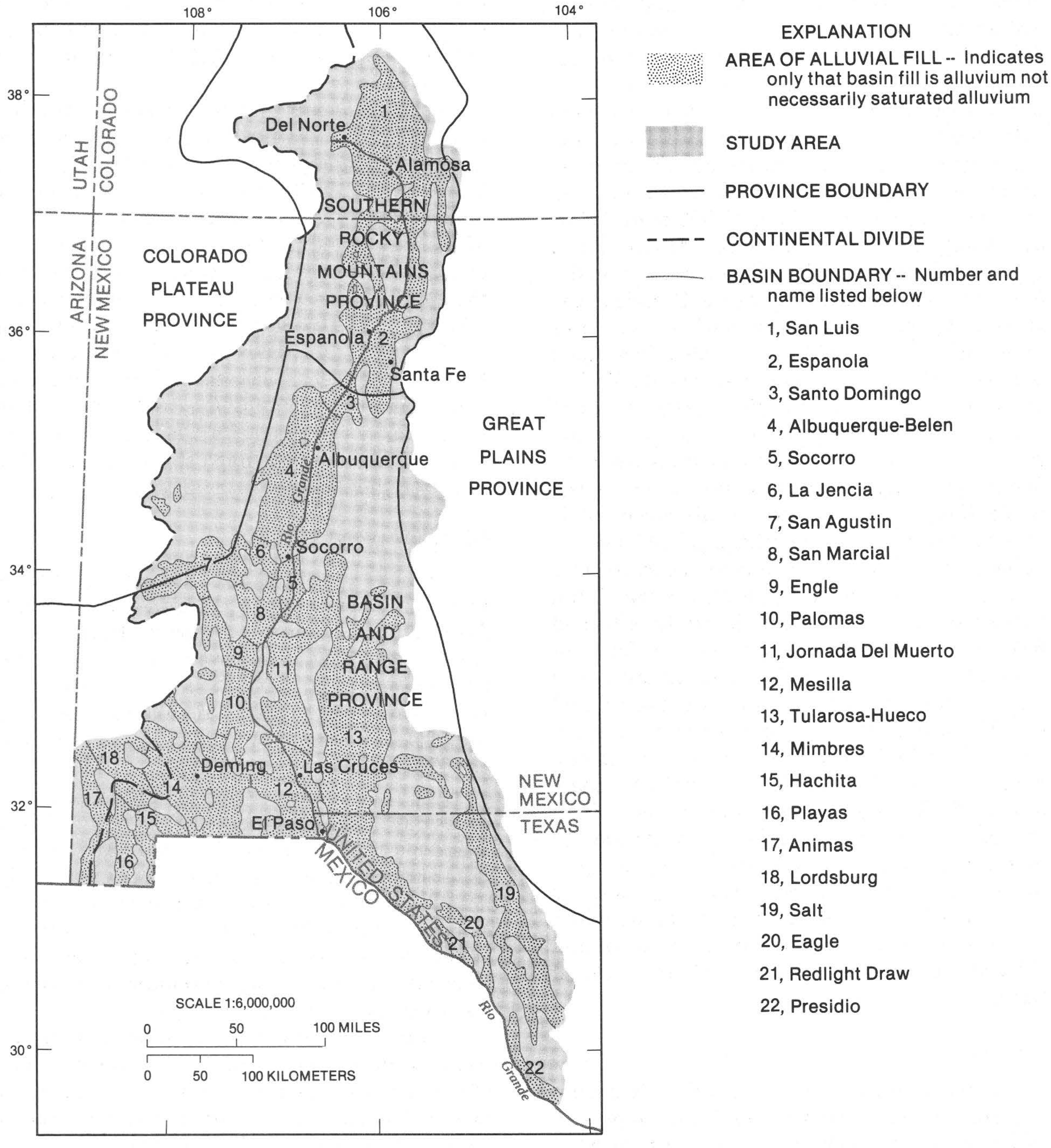

Figure 64.—Study area and basin boundaries of alluvial basins in parts of Colorado, New Mexico, and Texas.

estimates of volumes and locations of recharge supplied from streams with drainage areas on relativly impervious bedrock masses. Recharge occurs as flow crosses permeable alluvial fan deposits at the toe of the mountains. These results are applied to simulation of specific basins. Simulation suggests that the estimated recharge is reasonable. During development of the "mountain-front recharge model," the following runoff equations were established empirically. If precipitation is less than 7.36 inches, runoff may be calculated as:

$$Q = 1.074 \times 10^{-5} \times A^{1.216} \times P^{2.749} \times S^{0.535}$$

If precipitation is greater than 7.36, the runoff may be calculated as:

$$Q = 7.621 \times 10^{-5} \times A^{0.977} \times P^{3.596}$$

where:

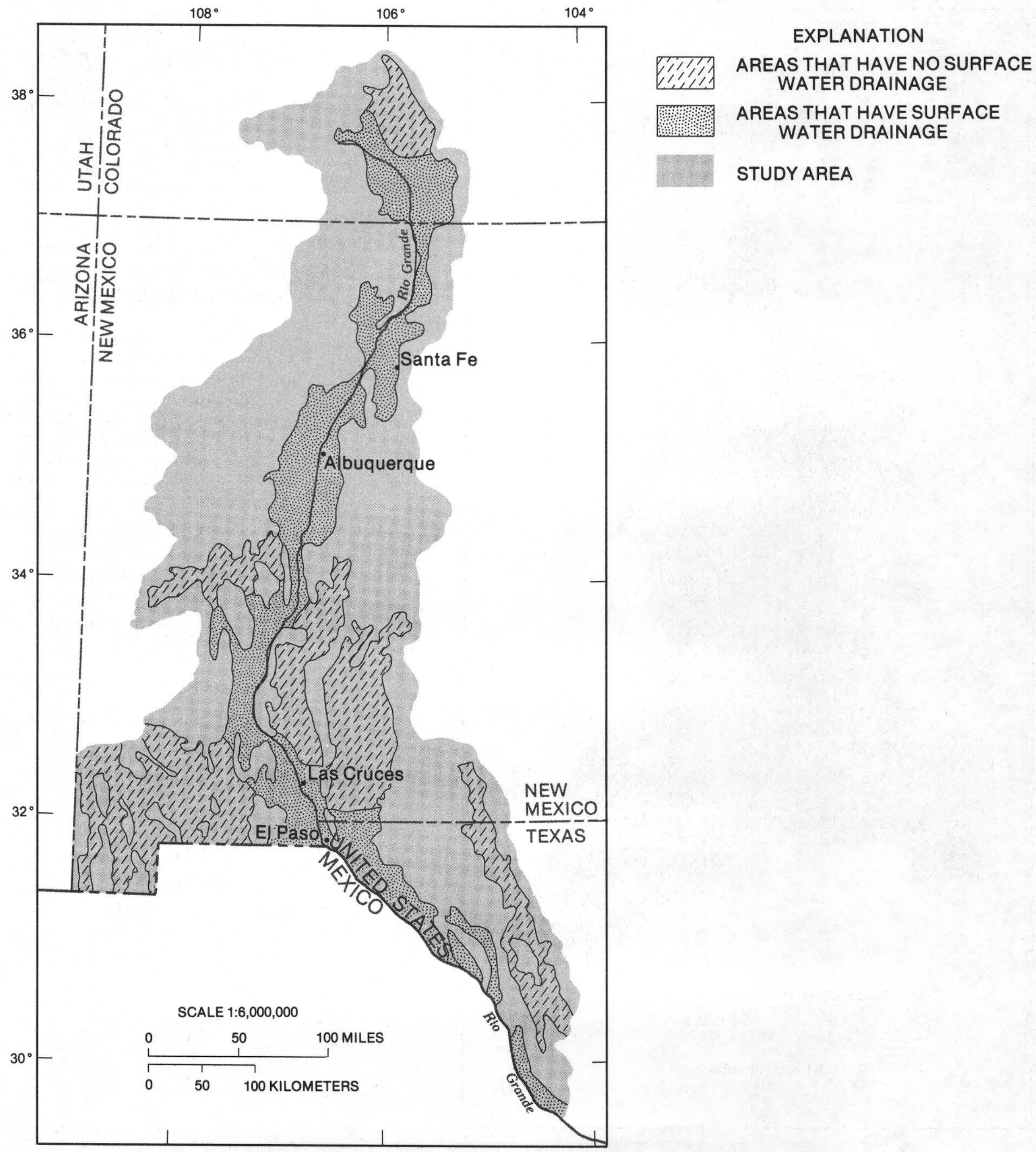

Figure 65.—Alluvial basins with open and closed surface water drainages, in parts of Colorado, New Mexico, and Texas.

Q is runoff, in cubic feet per second,
A is the drainage area, in square miles,
P is precipitation, in inches, and
S is slope of the drainage channel, in feet per mile.

The second recharge model (the mountain water-balance model) was developed to estimate direct recharge within uplifted mountainous areas composed of relatively impervious rocks, such as Quaternary or Tertiary volcanics. The "mountain water-balance model" was especially useful for estimating recharge entering the San Luis Basin from the San Juan Mountains on the west. The model uses precipitation, types of vegetation, and ambient temperature to estimate recharge within a particular drainage basin. Some water drains to streams through local flow

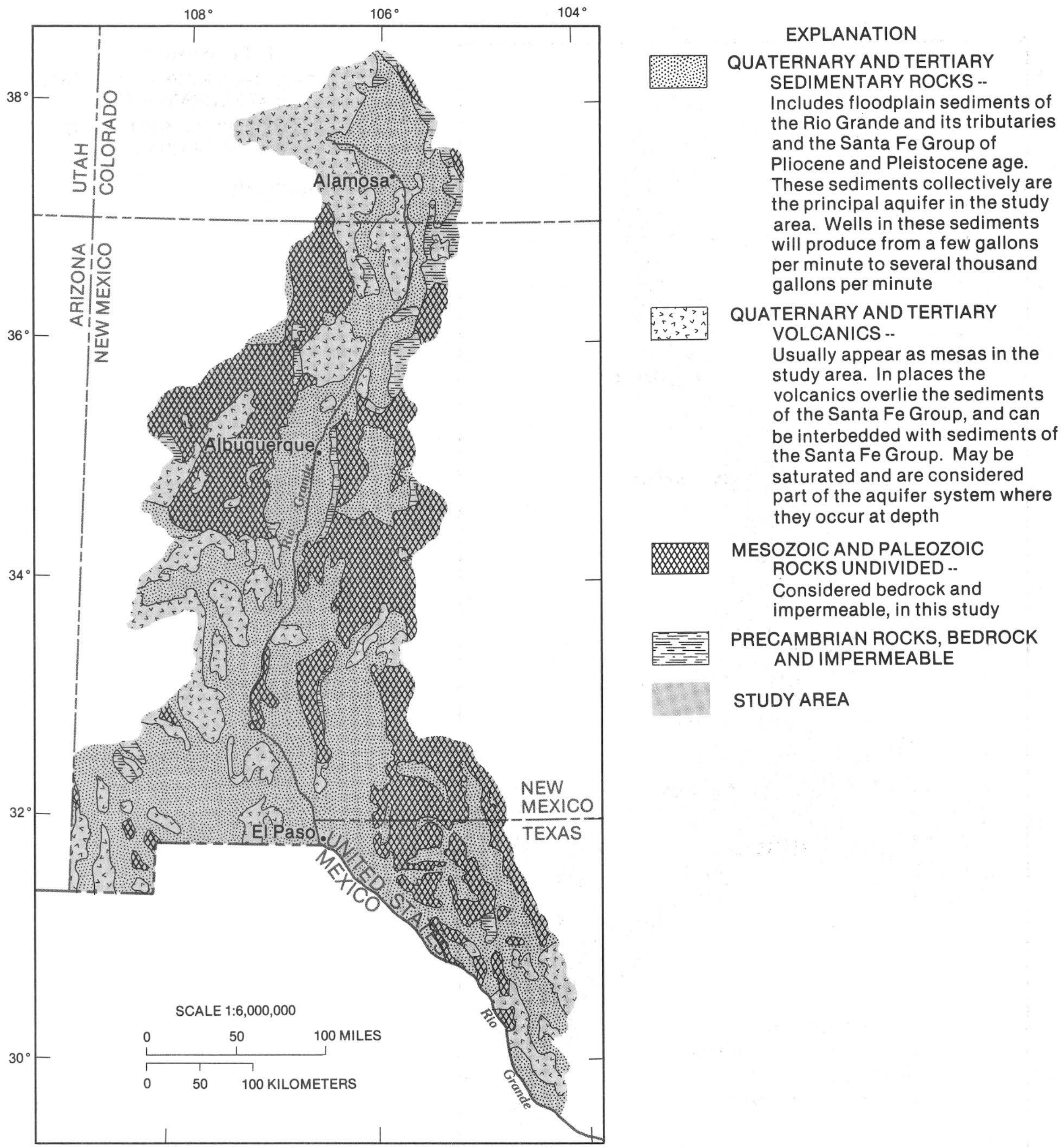

Figure 66.—Generalized geologic map of the alluvial basins in parts of Colorado, New Mexico, and Texas.

systems. The remaining water is assumed to flow through permeable rock and recharge the alluvial sediments at the toe of the mountains. Results from this recharge model and water budgets in the San Luis Basin suggest that there is some lateral ground-water movement from areas in the San Juan Mountains to adjacent downgradient areas.

The third recharge model is a "valley evapotranspiration (ET) model." The model takes into account precipitation, depth to water, types of natural vegetation and crops, irrigated acreage, and consumptive use of vegetation and crops. In irrigated areas of the San Luis Basin, it is estimated that about 90 percent of the precipitation is returned to the atmosphere by

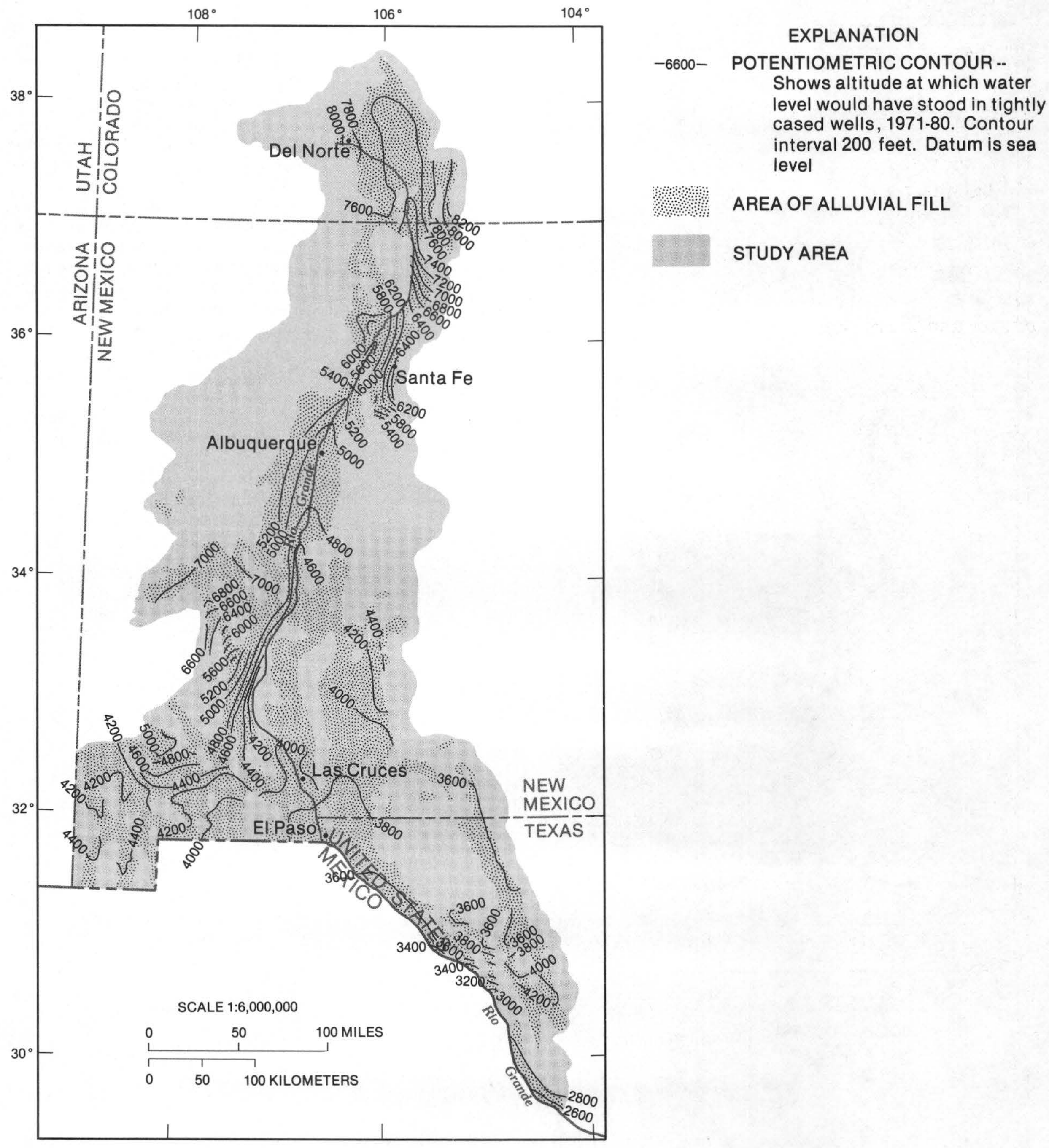

Figure 67.—Regional potentiometric surface of the alluvial basin aquifer systems in parts of Colorado, New Mexico, and Texas, 1971-80.

evapotranspiration during the growing season and about 27 percent is returned during the nongrowing season.

SIMULATIONS

For basins where sufficient data were available, the ground-water flow systems were simulated by computer-based models. Basins selected for simulation were the San Luis, Albuquerque-Belen, Mesilla, and Animas Basins.

San Luis Basin.—Cross-sectional simulations were made to evaluate the conceptualized flow in the vertical direction. Several sections, located where the most data were available, were simulated. The simulated depth of the basin

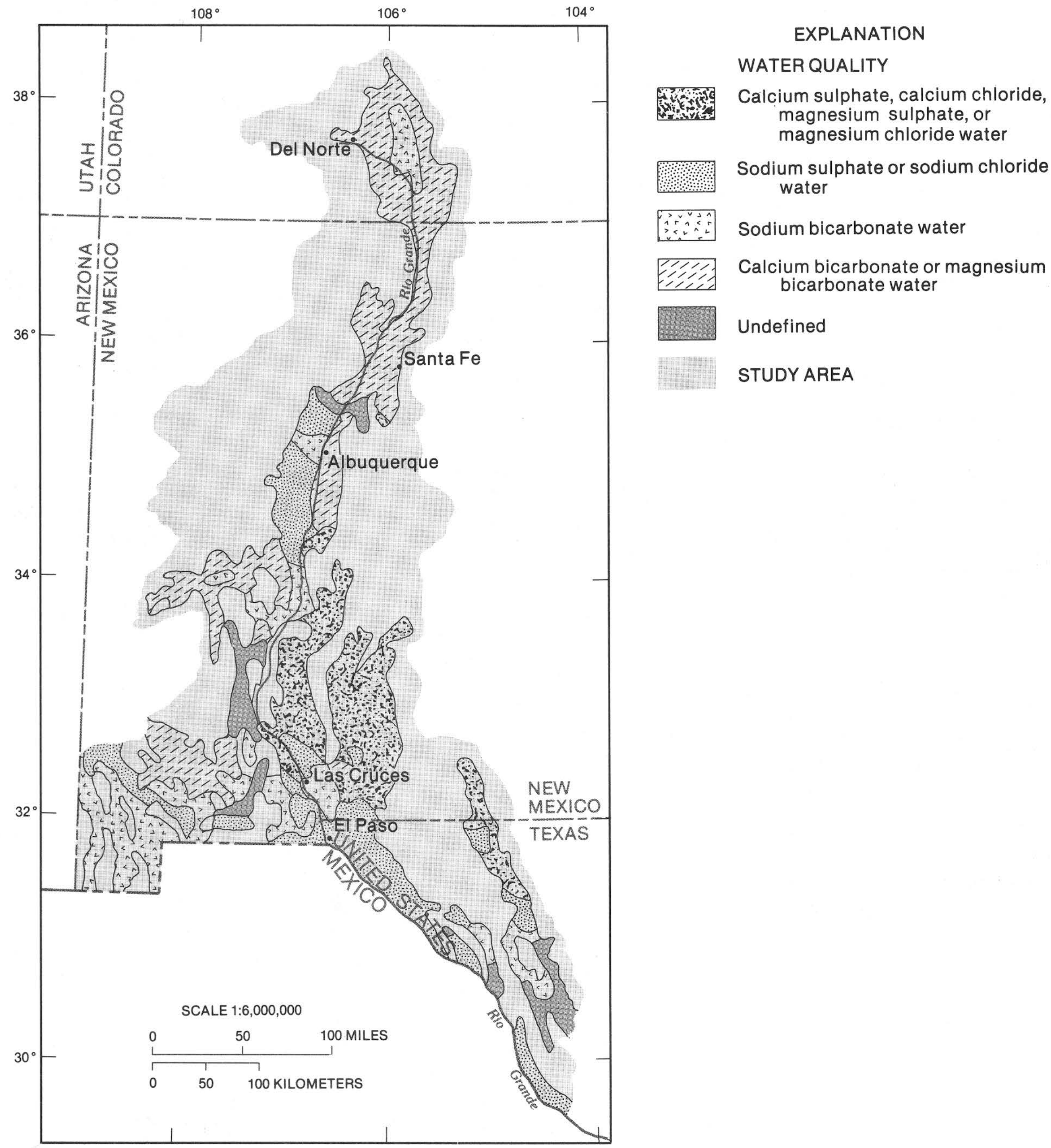

Figure 68.—Regionalized water quality of the alluvial basin aquifer systems in parts of Colorado, New Mexico, and Texas.

aquifer was stopped at 3,200 feet during the cross-sectional simulation. Extending the depth beyond 3,200 feet had no significant improvement on results. Sensitivity of dividing the aquifer into several layers having different thicknesses was tested by the cross-sectional simulation. The tests indicated that better results can be obtained where a given layer is not more than 1.5 times the thickness of the layer above.

A three-dimensional flow model (Trescott, 1975; Trescott and Larson, 1976) was used to simulate the aquifer underlying the San Luis Basin north of the Colorado-New Mexico State line. Because ground-water development in the basin was begun in the late 1800's, there were few data available to define the predevelopment steady-state conditions. Information on hydrologic changes since 1950 was used for the transient calibration. Simulation indicates that in 1980 about 80 percent of ground-water withdrawals were from salvage of evapotranspiration, 14 percent from aquifer storage, and 6 percent from reduction in discharge as stream baseflow.

Albuquerque-Belen Basin.—The ground-water flow model of the Albuquerque-Belen Basin is also a three-dimensional model representing the aquifer to a depth of 6,180 feet with 6 active layers. Both steady-state and transient simulations were made. The flood plain was represented by a constant-head boundary. This assumption is made on the basis that no long-term water-level fluctutations in the flood-plain aquifer existed prior to 1975. Simulation confirms the results of the geochemical study in the basin that upwelling of flow exists at the southern end of the basin. Simulation also suggests that changes in recharge will not totally disappear for about 800 years; that indicates that the aquifer system underlying the basin does not reach equilibrium easily. Simulation also suggests that about 25 percent of the water withdrawn from 1907 to 1979 was from aquifer storage outside the flood plain. About 75 percent was from the Rio Grande and flood plain aquifer, assuming that evapotranspiration remained constant.

Mesilla Basin.—The Mesilla Basin was simulated with a 5-layer model. The flood plain of the Rio Grande was represented in the model with as much detail as possible. Data for irrigation pumpage, evapotranspiration, domestic and industrial use, and irrigation return flows were compiled or estimated from the late 1800's to 1975. Flow in rivers and drains was routed. Simulation indicates that most hydrologic details observed in the flood plain of the Rio Grande can be simulated. Simulated average volume of ground-water discharge to the river or drains from 1940 to 1975 was within about 5 percent of the observed values. In 1960 about 75 percent of the pumped ground water was induced from surface-water bodies and the remainder was from aquifer storage.

Animas Basin.—A two-dimensional flow model was used to simulate the aquifer system underlying the Animas Basin in southwestern New Mexico. This model was done under contract by the New Mexico Bureau of Mines and Mineral Resources (O'Brien and Stone, 1983). The model serves as a type model that could be used to simulate a closed-basin aquifer system in which vertical flow is insignificant. The two-dimensional flow model produced acceptable results. During the transient calibrations, change in head from 1948 to 1955 were simulated to within 10 feet of the observed head changes. From 1948 to 1981, the simulated head changes were within 18 feet of the observed drawdowns. Sensitivity tests showed that the aquifer system is most sensitive to increases in transmissivity and withdrawal and to decreases in storage coefficient and transmissivity (O'Brien and Stone, 1983, p. 50).

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STUDY IN SOUTHERN AND CENTRAL ARIZONA AND PARTS OF ADJACENT STATES

By **Thomas W. Anderson, Tucson, Arizona**

INTRODUCTION

The study of the alluvial-basin regional aquifer systems in southern and central Arizona and parts of California, Nevada, and New Mexico covering an area of about 82,000 mi^2 (fig. 69) was started in 1978. All activities were completed in 1984, except for report writing. The area contains 72 basins that are virtually separate hydrologic entities. The boundaries between basins mainly correspond to surface-water drainage divides, ground-water divides, or areas of minimal interbasin connection. The study area is characterized by sharply rising mountains that separate wide, flat basins filled with varying amounts of alluvial deposits. These deposits form the major aquifers and store large amounts of water. Altitude of the land surface generally increases from the southwest toward the north and northeast. Climatic factors have a wide range—precipitation increases and evaporation and temperature decrease from southwest to northeast.

GEOHYDROLOGIC SETTING

The basins of the study area are the result of formation of the Basin and Range topography about 10 to 15 million years ago (Scarborough and Peirce, 1978, Shafiqullah and others, 1980). Movement along high-angle normal faults down-dropped the basins in relation to the mountain masses. The result was a series of generally northwest-trending basins. Basin subsidence was a gradual process that was accompanied by deposition of locally derived sediments within the internally drained basins. The bedrock of the mountains consists of igneous, metamorphic, and sedimentary rocks that are virtually impermeable. These rocks underlie the basins and form the side and bottom boundaries of the alluvial basin aquifers.

The basins are filled with alluvial deposits that range from a few thousand feet to more than 10,000 ft in thickness. Basin subsidence and sediment deposition occurred at varying rates throughout the study area; as a result, the thickness, areal extent, and grain size of the alluvial deposits are highly variable. The sediments range in grain size from clay to gravel and may vary from unconsolidated to highly consolidated in a single basin. The general sequence of sedimentary units, in ascending order, consists of: (1) sedimentary rocks deposited prior to development of the Basin and Range topography, (2) lower and upper basin fill, and (3) stream alluvium (fig. 70). The sedimentary units are distinguishable on the basis of structural relation, degree of consolidation and deformation, sources of clasts, age, and water-bearing characteristics. The sedimentary units reflect different conditions of deposition. Although each basin is unique, general spatial patterns in character, thickness, and extent of the sedimentary units can be documented among groups of basins.

The sedimentary rocks deposited prior to development of the Basin and Range topography consist of moderately to highly consolidated continental deposits that range from clay to gravel and also include some volcanic rocks. The sedimentary rocks were faulted and tilted along with the underlying bedrock. These sedimentary rocks are deeply buried in the basins. Although a few wells penetrate these sedimentary rocks near the basin edges, little is known about their hydrologic characters.

The basin-fill sediments, which constitute the principal aquifers in most basins, were deposited during and after the formation of the Basin and Range topography. The lower basin fill is moderately to highly consolidated, has extensive fine-grained material, and contains evaporites that range from disseminated gypsum to massive halite deposits. The lower basin fill was deposited in topographically closed basins. The upper basin fill was deposited during a transition period in which the basins changed from a closed basin to an integrated drainage system. The upper basin fill is less consolidated, generally is thinner, and contains fewer fine-grained

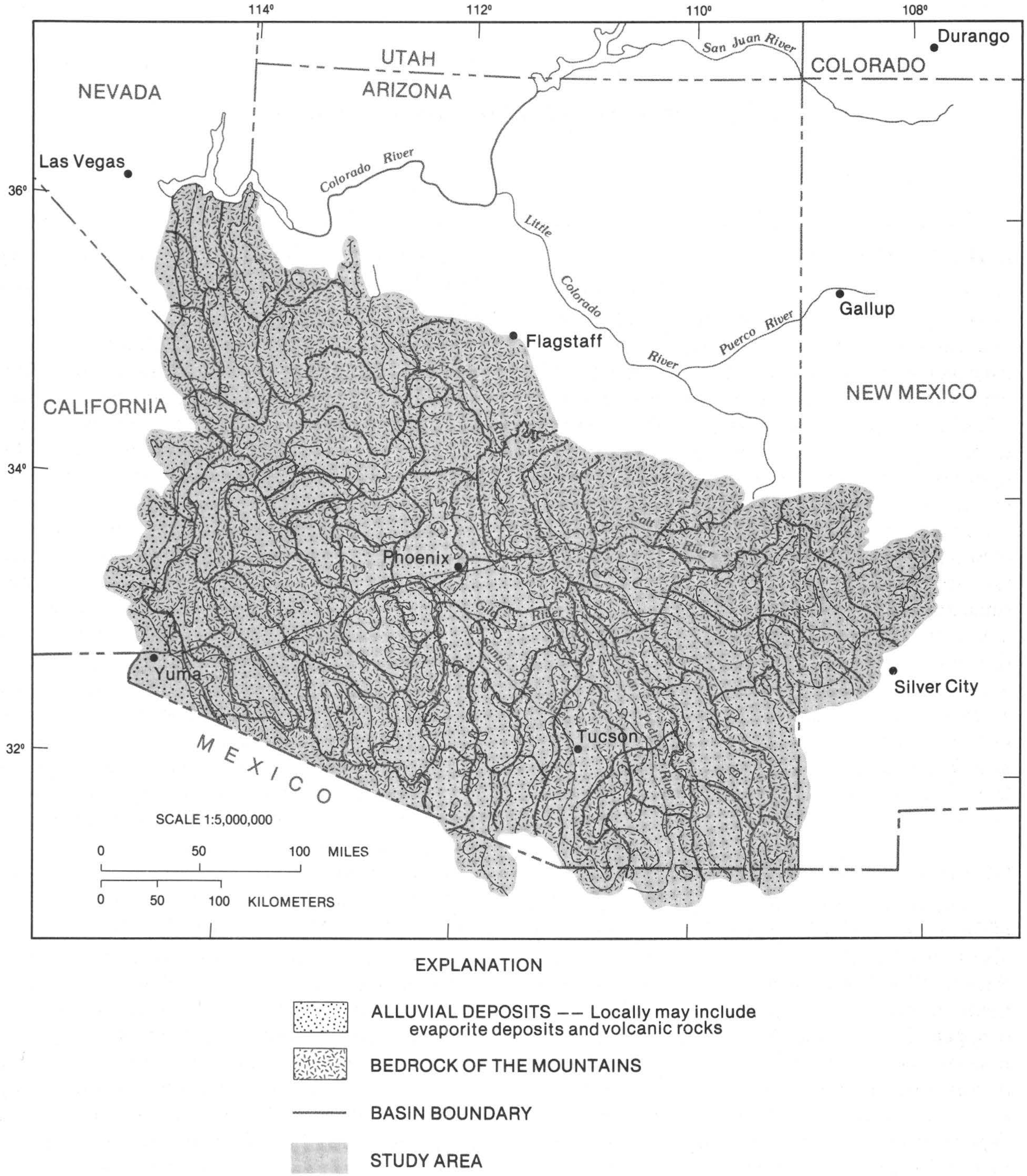


Figure 69.—Study area and basin boundaries of the alluvial basin aquifer systems in southern and central Arizona and parts of adjacent States.

sediments than the lower basin fill. Both lower and upper basin fills grade laterally from coarse-grained sediments at the basin margins to fine-grained sediments near the basin center.

The stream alluvium was deposited contemporaneously with the establishment of the present surface-drainage system. The alluvium ranges from clay to boulders and is unconsolidated. Where saturated, the stream alluvium forms the most productive part of the aquifers.

The sedimentary units are hydraulically interconnected and form a single aquifer system within most basins. The basins are interconnected in a dendritic pattern—much like the surface drainage—and form an integrated regional ground-water flow system. The individual basin-aquifer system serves mainly as a subsurface reservoir with only small areas of interconnection between the basin aquifers. A small quantity of ground water flows from basins of higher altitude to those of lower altitude.

The chemistry of ground water generally is good and is similar from basin to basin. Concentrations of dissolved solids generally are less than 1,000 milligrams per liter (mg/L). Where appreciable gypsum or halite is present, such as in the fine-grained facies of the lower basin fill, concentrations of dissolved solids may increase severalfold. Trace elements, such as mercury, lead, arsenic, and nitrate associated with the rock matrix in which the ground water occurs, locally restrict the use of the water.

Ground water generally is unconfined. Confined conditions occur in a few basins where extensive fine-grained deposits overlie the principal water-bearing formation. Depth to water ranges from land surface near perennial streams to more than 500 feet below land surface near the mountain fronts in some basins. An estimated 900 million acre-ft of recoverable water was stored in the upper 1,200 feet of the alluvial deposits of the basins prior to development (Freethey and Anderson, 1985). This amount is much greater than the amounts of water that move into and out of the aquifers annually.

Sources of recharge are: (1) infiltration of streamflow along major streams, (2) infiltration of streamflow along mountain fronts, and (3) subsurface inflow from adjacent basins (fig. 71). The average annual recharge to all basins in the study area was estimated to be 2.5 million acre-ft before development. In general, ground water discharges to the atmosphere as evapotranspiration, through springs, to the downgradient adjacent basin as subsurface outflow, or to streams where head in the aquifer is higher than the stage in the stream. The specific flow characteristics in a basin or for a group of basins are dependent on the hydraulic

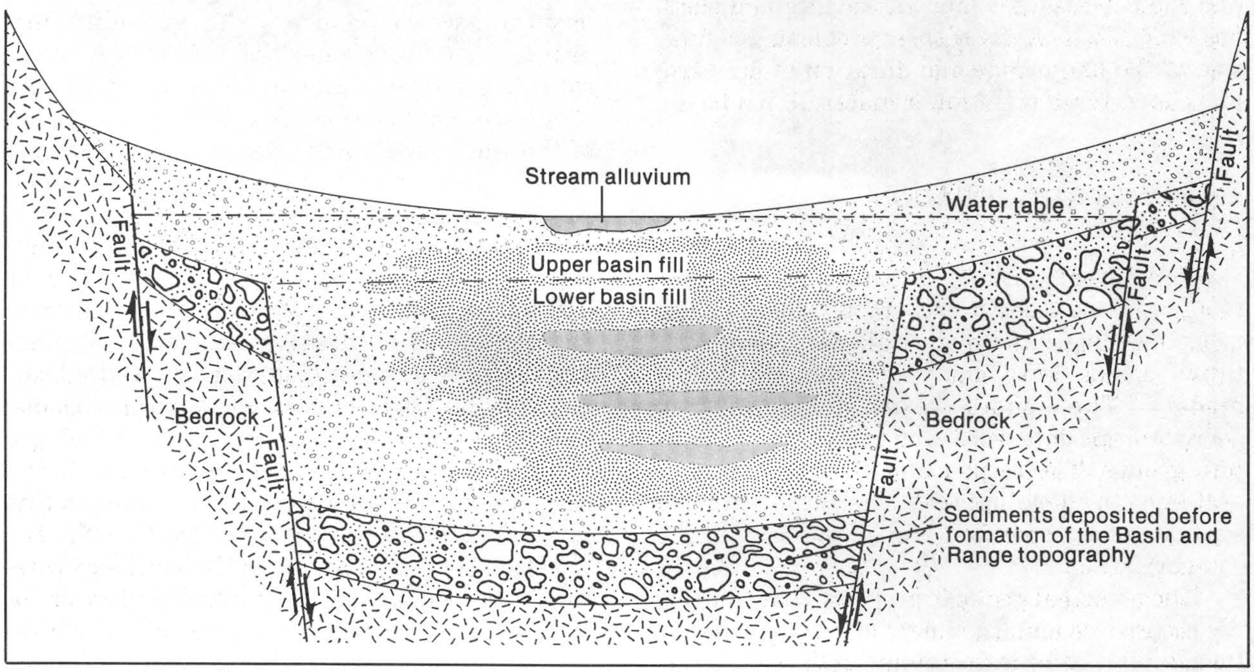


Figure 70.—Typical sequence of sedimentary units that fill a basin in southern and central Arizona and parts of adjacent States.

properties of the aquifer material and on the total flow through the aquifer.

Ground-water development in the study area has been principally for irrigation and has resulted in the depletion of surface-water flow in several basins and water-level declines in developed basins. The greatest development began after World War II. Ground-water pumpage increased from 1.7 million acre-ft in 1942 to 3.8 million acre-ft in 1952 and to 4.8 million acre-ft in 1962 (fig. 72). During 1950-80, an average of 4.8 million acre-ft per year was pumped. Pumpage in individual basins was 2 to more than 100 times greater than the estimated annual recharge. Through 1980, an estimated 184 million acre-ft of water was pumped from the basins (U.S. Geological Survey, 1982). Of the estimated total pumpage, not all was derived from aquifer storage; some pumpage was balanced by irrigation return flow, and leakage from canals and streams.

The major developed basins have experienced water-level declines that range from less than 50 feet to more than 400 feet (fig. 73). The magnitude of water-level decline varies from basin to basin and reflects the characteristics of the geohydrologic environment and the magnitude and duration of pumping. Associated with water-level declines are the problems of decreases of well yield, changes in water quality, and the potential for land subsidence and earth fissuring. The severity of these problems is a function of the magnitude and duration of pumping and character of the aquifer material in a basin.

REGIONAL CATEGORIZATION OF BASINS

Conceptual models of the basin aquifer systems in the study area were developed on the basis of geologic and hydrologic data. Representative basins were analyzed using numerical models. The model results and related geohydrologic data were used to categorize basins into groups. The response of the basin aquifer systems to development therefore can be evaluated for a specific basin on the basis of the categorization.

The principal geologic property that controls the occurrence and movement of ground water in the alluvial basins is the texture of the sediments. The important hydrologic factor is the total annual flow through the basin, which represents the annual renewable source of water. On the basis of the similarity of these geohydrologic factors, the alluvial basins in the study area are grouped into five categories and are named geographically as: (1) southeast, (2) central, (3) west, (4) Colorado River and (5) highland (fig. 74).

Flow simulation of basin aquifer systems of each category requires different input data. Aquifers underlying 12 alluvial basins (fig. 74) were simulated to test the concepts of the categorization. Simulation suggests that hydraulic conductivity, recharge rate, and extent and integrity of a confining unit constitute the most important information for steady-state conditions of the southeast category. Specific yield, pumpage, and pumping location constitute the most important information for transient conditions of the central and west categories. Stream-aquifer interaction, evapotranspiration, and return of excess irrigation water constitute the most important information for the Colorado River and highland categories, because infiltration of surface water is a major inflow component in these basins.

SUMMARY OF BASIN CHARACTERISTICS

Southeast Category

The aquifer systems underlying the basins in the southeast category consist of at least two saturated units, the upper basin fill and the lower basin fill, separated by a low-permeability fine-grained confining unit (fig. 75a). The lower saturated unit may include at least the upper part of the sediments deposited before the formation of the Basin and Range topography. Mudstone and evaporites may be found in the lower basin fill.

The major ground-water components are streamflow infiltration along the mountain front and along the major streams. Ground-water discharge before development was through evapotranspiration, stream baseflow, and subsurface outflow to downgradient adjacent basins. Because of similarities in geohydrologic characteristics, the Big Sandy River basin in the northwestern part of the area is included in this category despite its geographical location (fig. 74).

Development of basins in the southeast category has resulted in a gradual reduction of natural discharge, increase in streamflow infiltration, and a small change in the volume of water stored in the aquifer. Models of the basin aquifers of the southeast category need to be multilayered and allow for flow through leaky confining units.

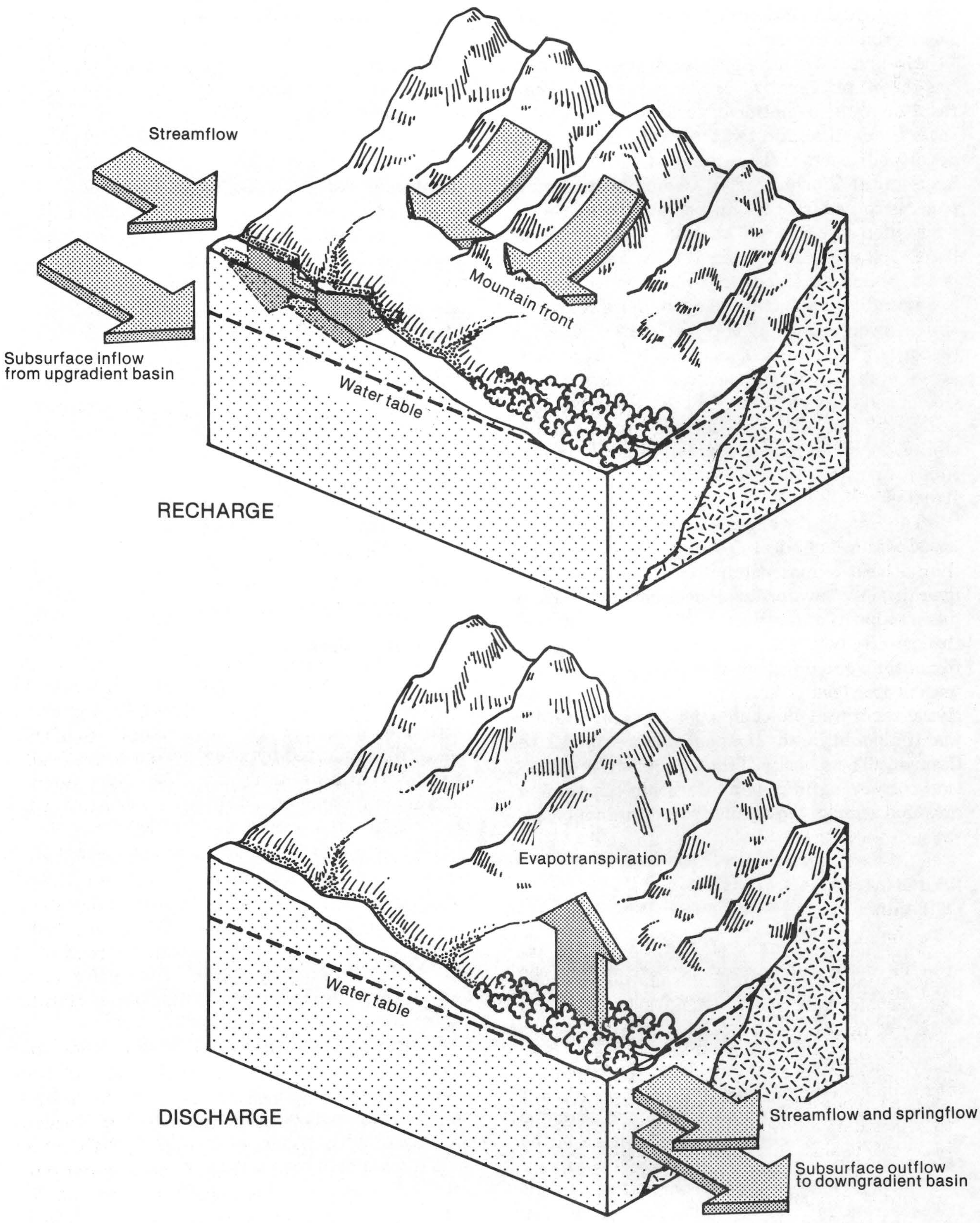


Figure 71.—Recharge and discharge components in alluvial basins in southern and central Arizona and parts of adjacent States.

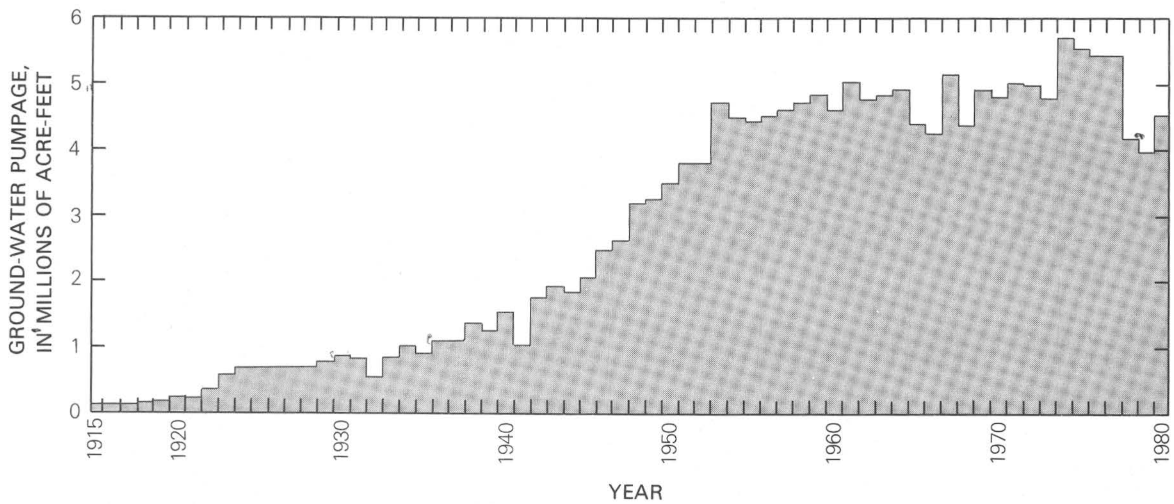


Figure 72.—Annual ground-water withdrawal, 1915-80, from the alluvial basin aquifer systems in southern and central Arizona and parts of adjacent States.

Central Category

Sediments of the aquifer systems underlying the basins of the central category are the thickest in the study area. Few wells have penetrated the sediments that were deposited before the formation of the Basin and Range topography; therefore, little is known about their extent and character. Typically, aquifers underlying the basins of this category contain as much as 5,000 feet of lower basin fill and less than 1,000 feet of upper basin fill (fig. 76A). Evaporites typically are found in the lower basin fill. Fine grained sediments commonly comprise more than 60 percent of the sediments near the center of the basins. The upper basin fill grades laterally to coarse material near the mountain fronts. Stream alluvium along major streams is as much as 300 feet thick.

Aquifer recharge consists of small to moderate amounts of mountain-front recharge and streamflow infiltration. Subsurface inflow from adjacent upgradient basins is also a significant recharge source in several basins (fig. 76B). Before development, ground water discharged principally as evapotranspiration. Subsurface outflow to adjacent downgradient basins was significant in several basins. Discharge to streams was small.

Development has greatly decreased the amount of discharge as evapotranspiration and stream baseflow. Most of the pumpage is from aquifer storage. Further increases in pumpage will cause further decline in water levels. Models used to simulate the alluvial aquifers underlying the basins of the central category need to be multilayered for simulating vertical-flow components and variable hydraulic characteristics with depth in the basins.

West Category

The aquifer systems underlying the basins of the west category contain sediments deposited before the formation of the Basin and Range topography, which are overlain by moderate- to coarse-grained lower basin fill. The lower basin fill is the principal aquifer in the basins of this category. Upper basin fill is thin and generally above the water table (fig. 77A). Stream alluvium is not typical of this category and is found only along the lower Gila River.

Recharge in this category is small and consists of minor amounts of mountain-front recharge, subsurface inflow from upgradient basins, and streamflow infiltration (fig. 77B). Discharge is also small and consists of evapotranspiration and subsurface outflow to adjacent downgradient basins. Discharge to streams is insignificant.

Because natural recharge, discharge, and streamflow are small in the basins of this category, no significant capture of natural discharge or increase in streamflow infiltration has occurred following pumping (fig. 77C), except locally along the Gila River. Ground-water pumpage is derived almost entirely from aquifer storage (fig. 77D). Multilayered models need to be used to simulate the aquifer systems underlying the basins of this category. The Big Sandy River basin southeast of Kingman is excluded from this category, despite its western geographic location.

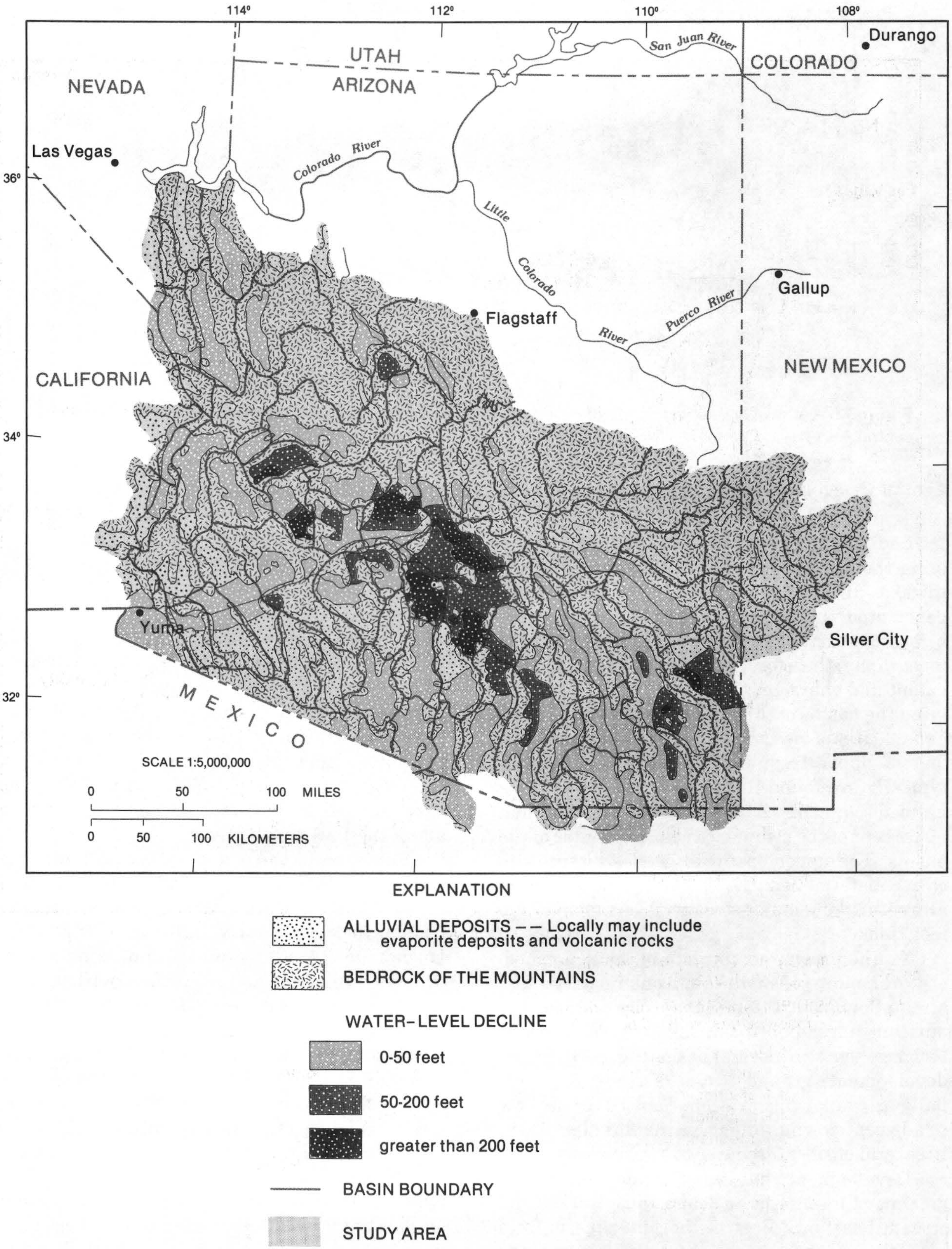
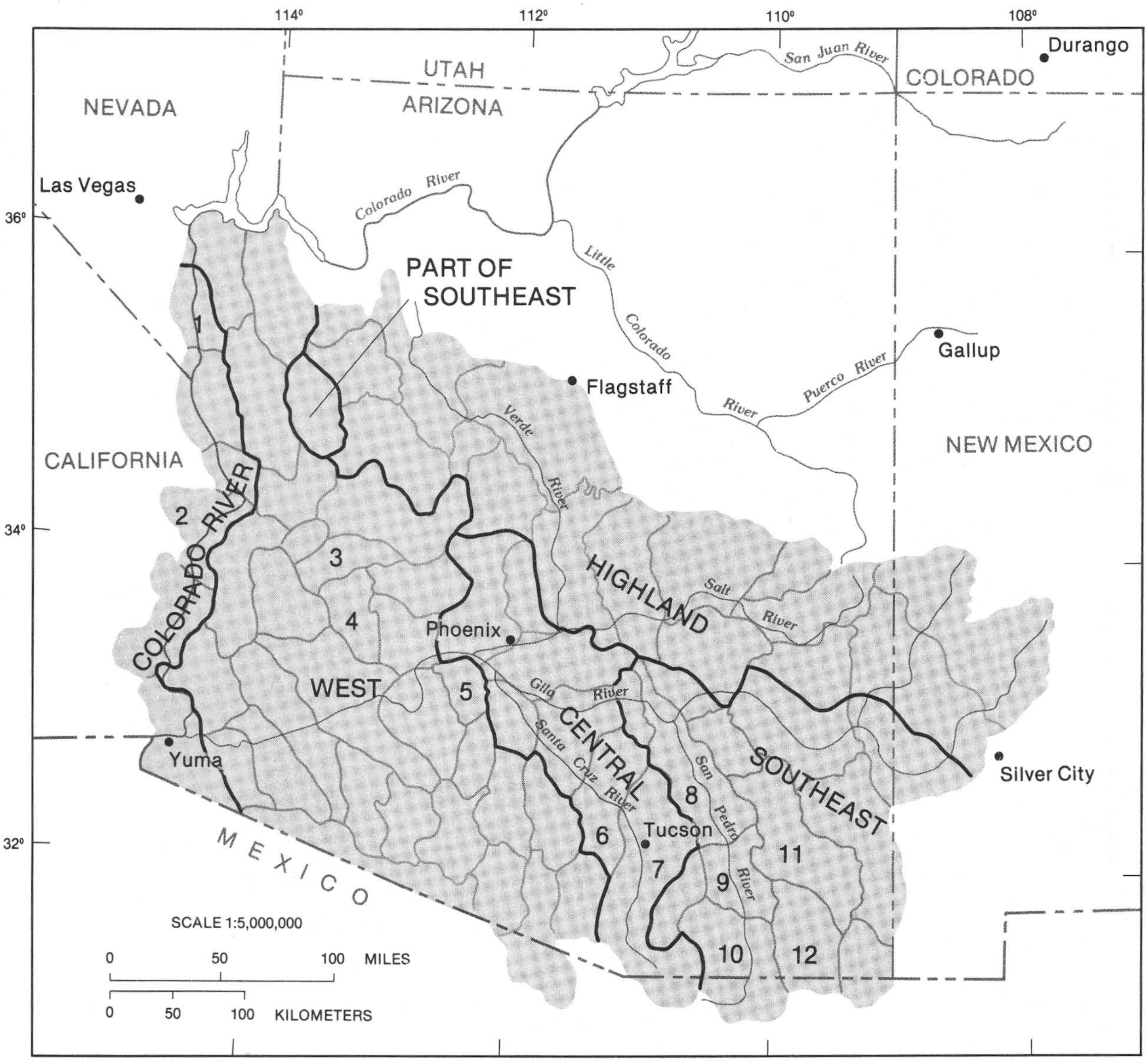


Figure 73.—Areas of water-level declines in the alluvial basin aquifer systems in southern and central Arizona and parts of adjacent States.



EXPLANATION

BOUNDARY OF GENERAL CATEGORIC ALLUVIAL BASINS

5 BASIN BOUNDARY -- Number indicates modeled basins listed below

1, Mohave
2, Parker
3, McMullen
4, Harquahala
5, Waterman
6, Avra Basin
7, Tucson
8, Lower San Pedro
9, Benson
10, Upper San Pedro
11, Willcox
12, Douglas

STUDY AREA

Figure 74.—Generalized categories of basins and modeled basins of the alluvial basin aquifer systems in southern and central Arizona and parts of adjacent States.

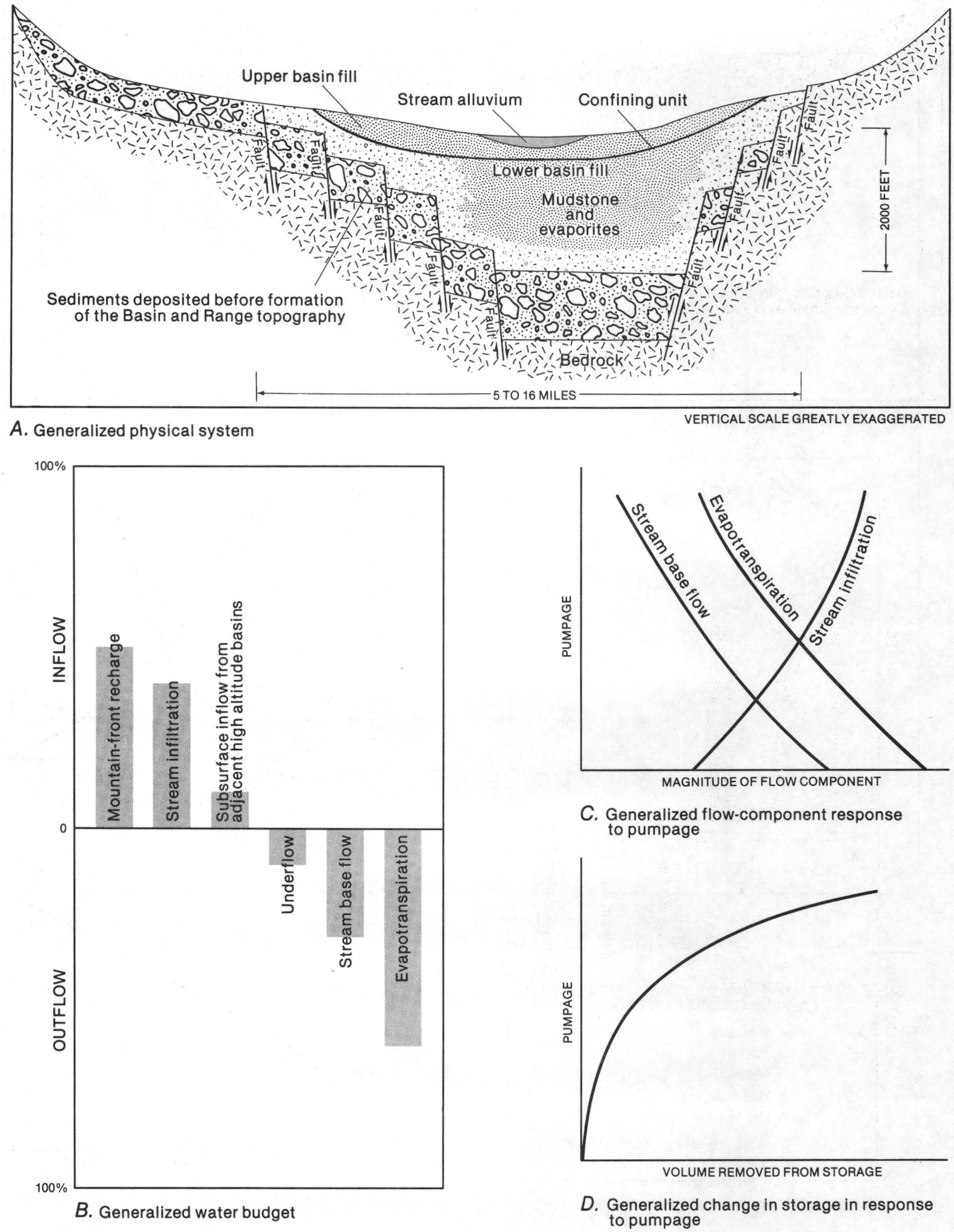


Figure 75. — Generalized characteristics of the aquifer system underlying the basins of the southeast category in southern and central Arizona and parts of adjacent States.

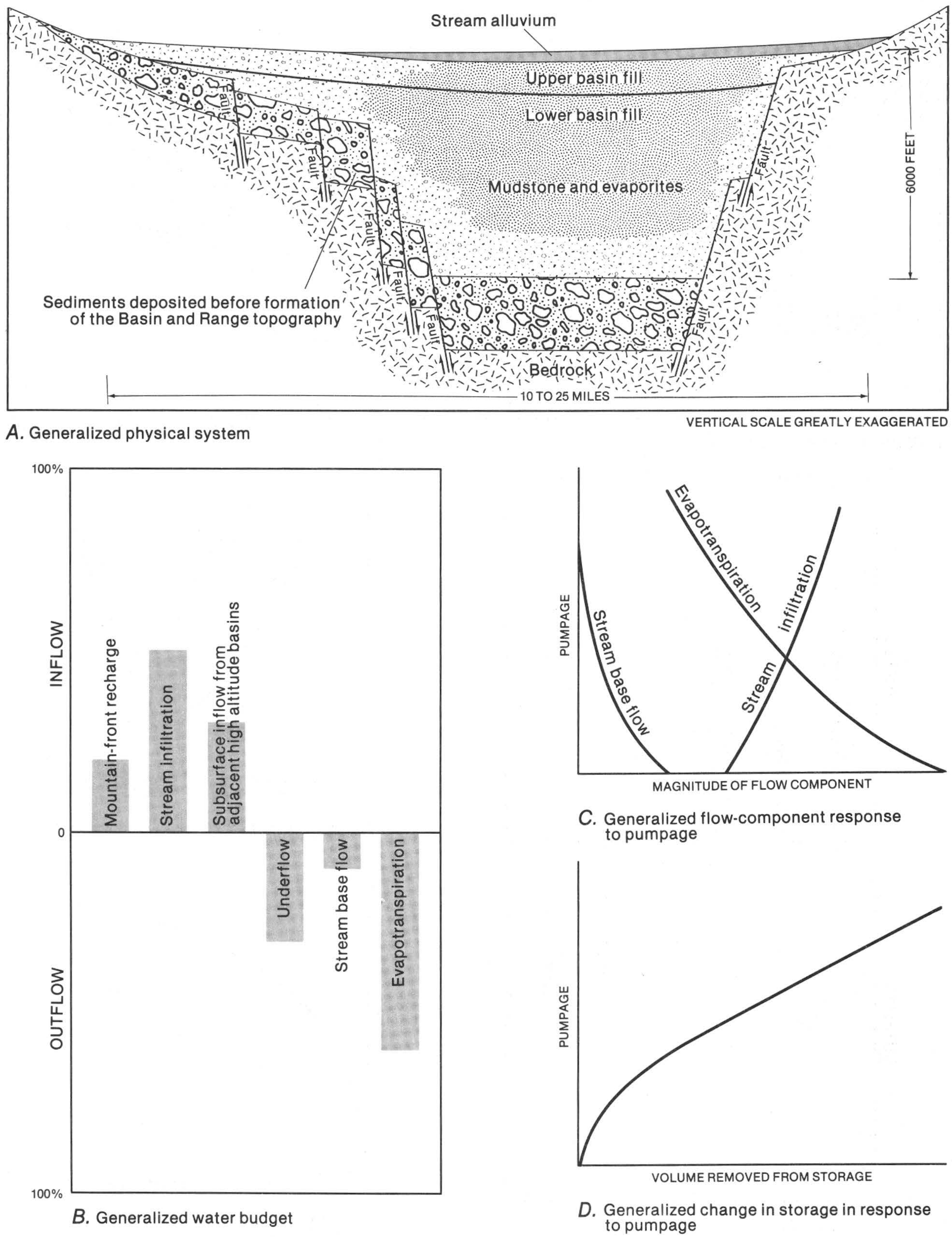


Figure 76.—Generalized characteristics of the aquifer system underlying the basins of the central category in southern and central Arizona and parts of adjacent States.

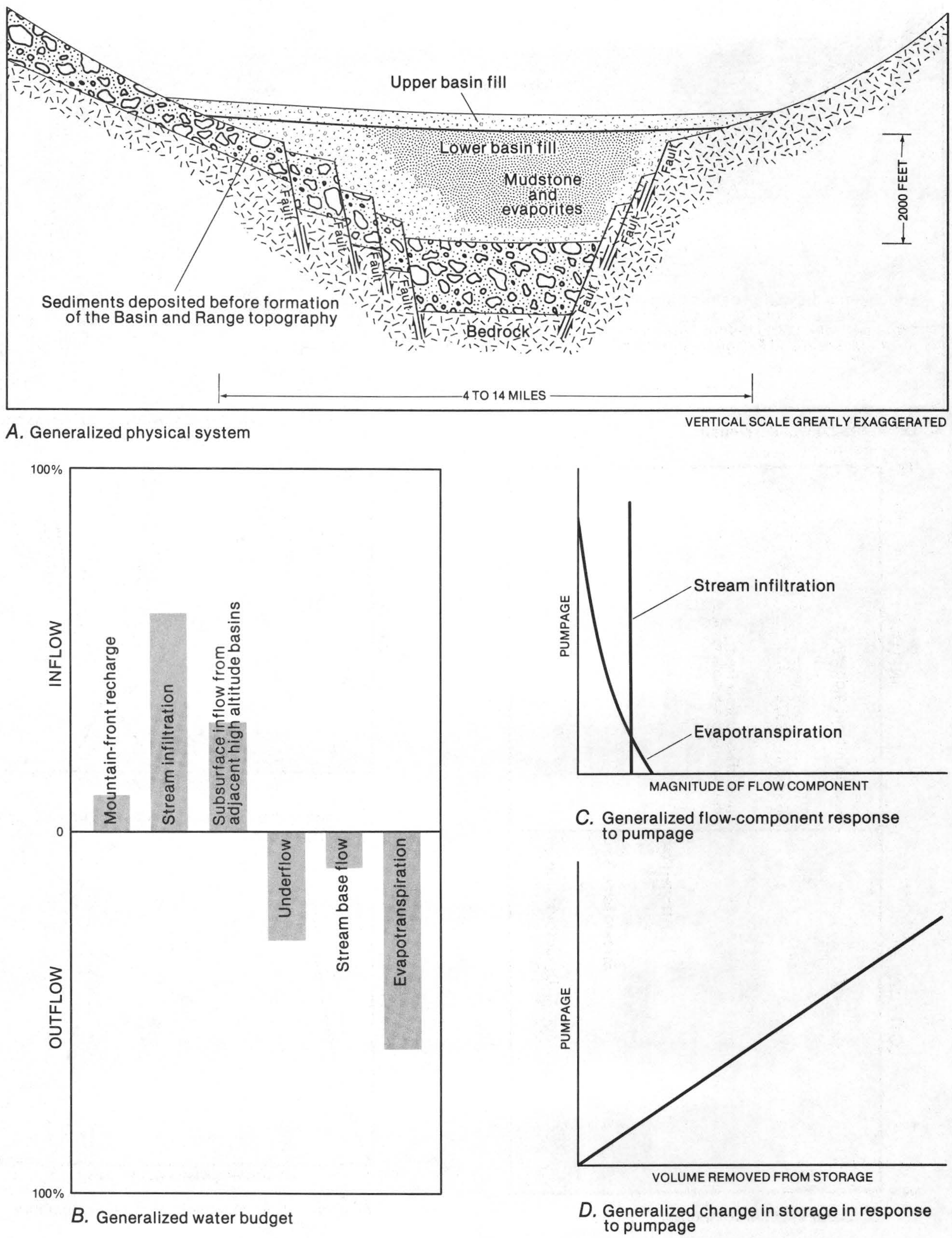


Figure 77.—Generalized characteristics of the aquifer system underlying the basins of the west category in southern and central Arizona and parts of adjacent States.

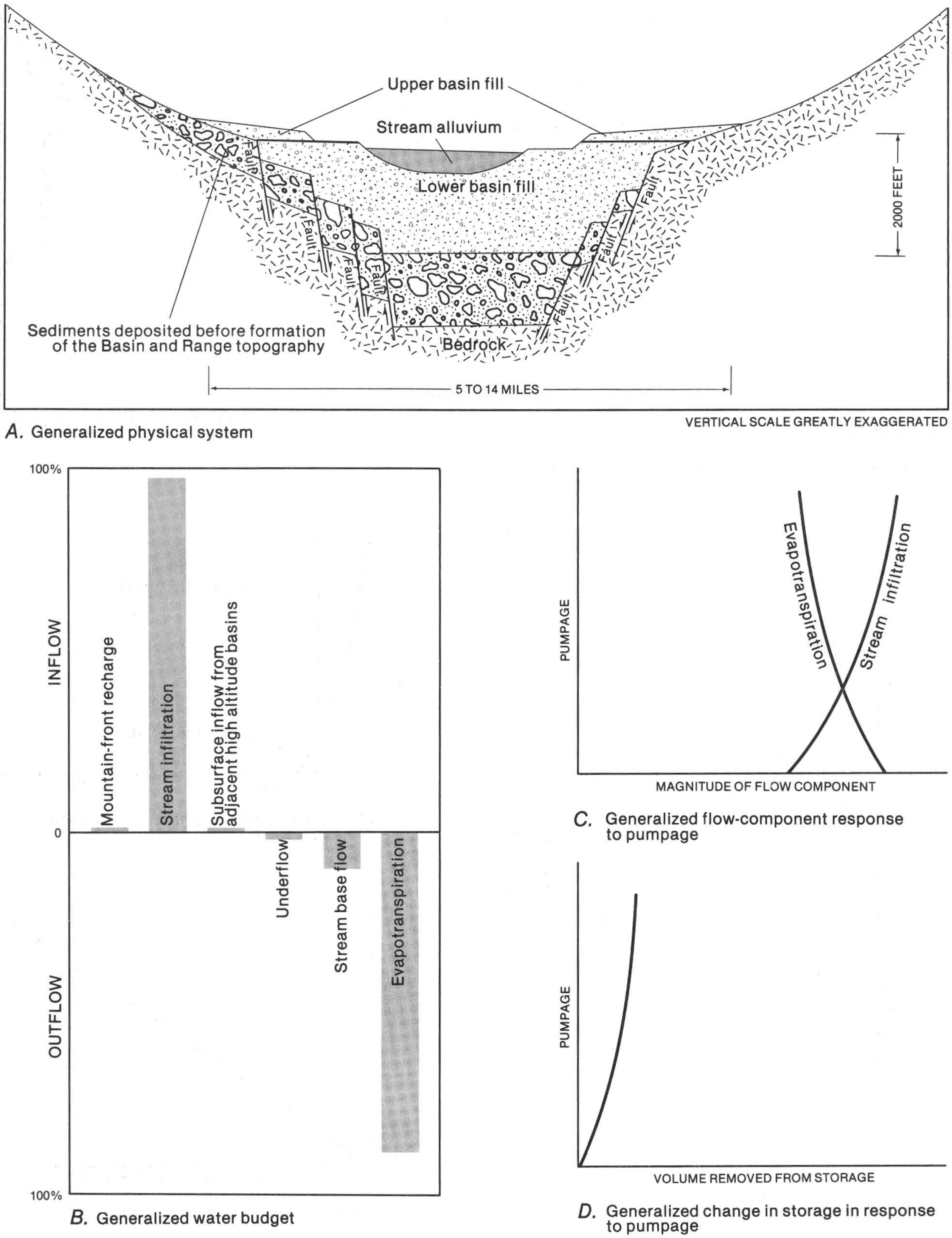


Figure 78.—Generalized characteristics of the aquifer system underlying the basins of the Colorado River category in southern and central Arizona and parts of adjacent States.

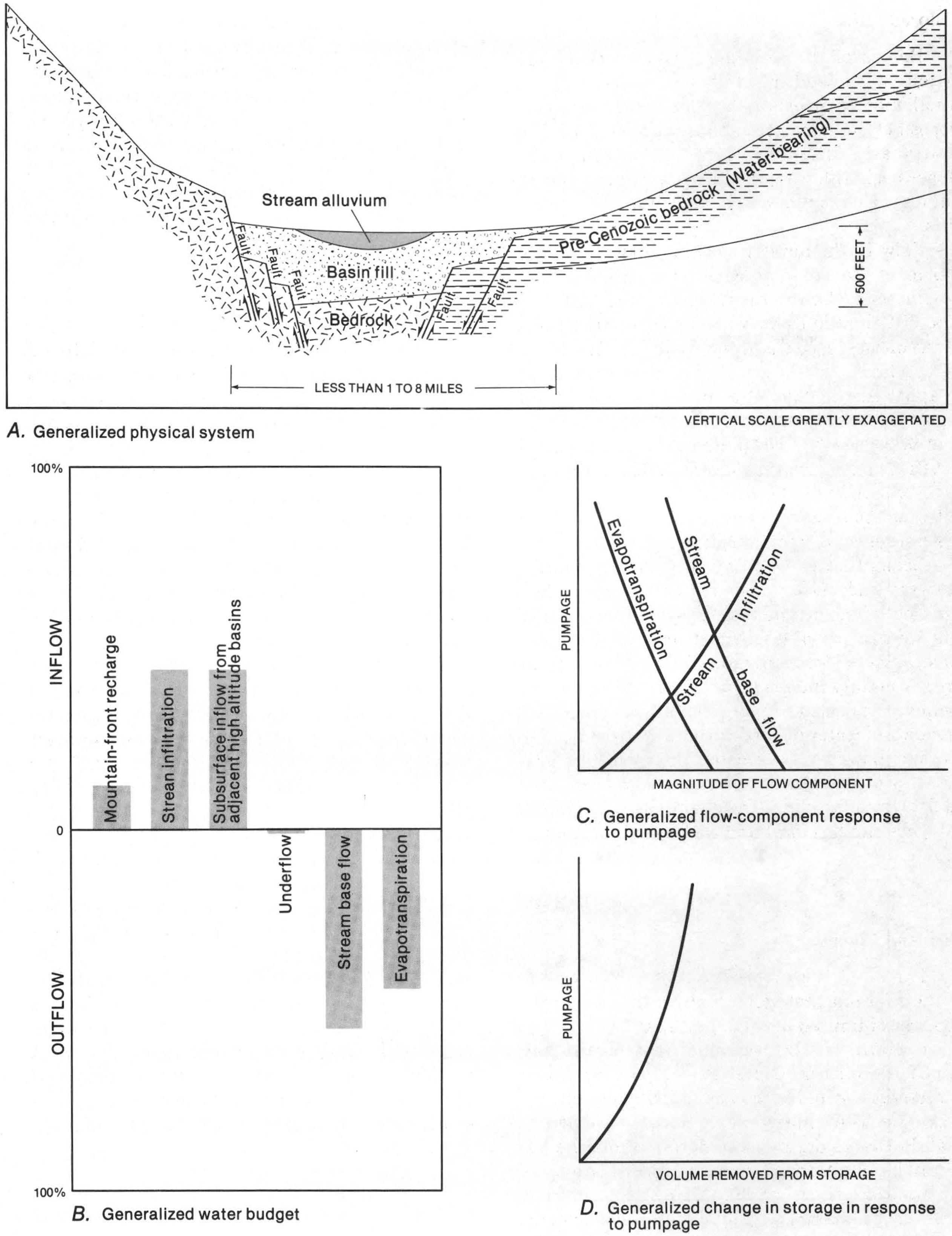


Figure 79.—Generalized characteristics of the aquifer system underlying the basins of the highland category in southern and central Arizona and parts of adjacent States.

Colorado River Category

Basins of the Colorado River category are filled with sediments that consist of a fanglomerate overlain by a marine estuarine deposit of the same age as the lower basin fill. The lower basin fill unit, in turn, is overlain by the upper basin fill. Stream alluvium overlies and occupies the channels which had cut into the basin fills.

Flow in the aquifer systems underlying the basins of the Colorado River category is totally dominated by flow in the Colorado River. Infiltration of Colorado River water is the main source of recharge to the aquifer systems. Discharge before development was almost entirely through evapotranspiration by phreatophytes and riparian vegetation (fig. 78B). Stream alluvium that underlies the flood-plain area is the principal aquifer. Ground-water flow occurs laterally and longitudinally in the basins. Surface water infiltrates and flows away from the river and discharges as evapotranspiration.

Ground-water withdrawal from the aquifer systems underlying basins of this category has minimal effect on the aquifer system because of the immensity of seepage from the Colorado River. Ground water pumped from the flood-plain area is mainly induced from the river. In places, evapotranspiration may decrease and minor amounts of water may be withdrawn from aquifer storage; however, increasing the hydraulic gradient by pumping will also result in more induced flow from the river. The aquifer systems in this category can be simulated with two-dimensional models.

Highland Category

The aquifer systems underlying the basins of the highland category contain thin basin-fill deposits of limited areal extent that occur in the flood plain of the streams that drain the area.These alluvial deposits are superimposed on a sequence of older, consolidated sedimentary rocks (fig. 79A). Many of the streams are perennial. Recharge occurs as underflow from the adjacent consolidated-rock aquifers and as streamflow infiltration (fig. 79B). Mountain-front recharge only occurs locally. Before development, ground water discharged as evapotranspiration and as stream baseflow.

The response of aquifers to development is dependent upon the magnitude of the development and its proximity to the perennial stream, if one is present. Typically, the result will be more induced streamflow infiltration, less discharge as stream baseflow, and reduced evapotranspiration. Ground-water pumpage from the aquifer systems in the basins of this category could result in some amount of water being withdrawn from aquifer storage (fig. 79D). Two-dimensional models are adequate to simulate the aquifer systems in this category.

SUMMARY

The study area is composed of 72 alluvial basins. The basins are filled with alluvial deposits that range from a few thousand feet to more than 10,000 feet in thickness. In almost all basins, the general vertical sequence of sedimentary units is, in ascending order, sediments deposited before the formation of the Basin and Range topography, lower and upper basin fill, and stream alluvium. Each of the hydrogeologic units has different physical, geologic, and hydrologic properties largely because of differences in the depositional environment and source area of the sedimentary material.

An estimated 900 million acre-ft of recoverable water was stored in the upper 1,200 feet of the sediments before development. The amount of water entering and leaving the basin aquifers is estimated to be about 2.5 million acre-ft per year. From the beginning of development through 1980, an estimated 184 million acre-ft of water has been pumped. Although a part of this volume has been balanced by recharge, water levels have declined more than 400 feet in some basins.

The basins of the study area are grouped into five categories on the basis of geologic and hydrologic properties. The groups are: (1) southeast, (2) central, (3) west, (4) Colorado River, and (5) highland. The character of the sediments filling the basins and the important flow components are similar within a category. The effect of development on the aquifer system and changes in flow components generally can be evaluated for each category by model simulation.

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ACTIVE PHASE I REGIONAL AQUIFER-SYSTEM ANALYSIS PROJECTS

CENTRAL MIDWEST REGIONAL AQUIFER-SYSTEM STUDY

By Donald G. Jorgensen, Robert B. Leonard, Donald C. Signor, and John O. Helgesen, Lawrence, Kansas

INTRODUCTION

The Central Midwest regional aquifer system study was started in 1980 and is scheduled for completion in 1986. The study area extends eastward from the foothills of the Rocky Mountains in Colorado to the valleys of the Missouri and Mississippi Rivers, and extends southward from northern Nebraska to south-central Arkansas. The area includes the Ozark Plateau and a large part of the Great Plains (fig. 80).

The sedimentary rocks underlying the study area, except in the St. Francois Mountains, are generally water-yielding formations and range in thickness from a featheredge where they pinch out against the St. Francois Mountains to more than 40,000 feet in the Anadarko Basin in central Oklahoma. The igneous and metamorphic basement rocks that underlie the water-yielding formations generally do not yield significant quantities of water to wells. Therefore, the surface of the basement rock effectively forms the base of the ground-water flow system in the study area.

Hydraulic properties of the various rocks in the study area differ greatly. These rocks include sandstone, shale, and evaporites of Cretaceous, Jurassic, and Permian age; limestone and shale of Pennsylvanian and Mississippian age and Dolomite and sandstone of Silurian, Ordovician, and Cambrian age. Except in the Ozark Plateau, little is known about the ground-water flow, and it is probable that not all aquifers have been identified. In much of the study area, the water-yielding rocks are deeply buried, and ground-water related data are scarce except for data collected incidentally by the petroleum industry. Because the cost of collecting additional hydrologic data in the deep subsurface is prohibitive, special efforts and techniques are needed to evaluate and analyze existing data.

PROGRESS OF THE STUDY

The aquifers studied consist generally of Early Cretaceous and older rocks. These aquifers constitute the major components of the regional-flow system within the study area. Local aquifers are not studied unless they are hydrologically important components of the regional-flow system.

Available maps and reports of previous investigations, along with readily accessible data describing hydrologic and geologic characterstics, were examined to define the regional hydrogeologic units and to form a conceptual model of the ground-water flow system.

Data have been collected on water quality, water use, water levels, hydraulic properties, geohydrologic-unit thicknesses, and altitude. The data bases in which the collated data are stored are nearly complete. Maps and tables are now being prepared from the information stored in these data bases.

A ground-water flow model for the entire study area incorporating variable-density effects is being constructed and used. Subregional flow models are being used to simulate the freshwater flow system of the Ozark area, and to simulate flow conditions in the Great Plains aquifer system. A model simulating flow into and out of the water-table aquifer has also been developed and models of water chemistry are being tested.

Several special studies have been undertaken to evaluate specific items of particular importance in understanding the hydrology in the study area. The special studies, some of which have been completed, are

- Geohydrology of the aquifers in the Ozark area.
- Geohydrology of the aquifers consisting of Lower Cretaceous rocks in the Great Plains aquifer system.
- Hydrology and geochemistry of the transition zone between the freshwater flow system in the Ozark area and the saline water flow system in the western part of the Interior Plains.
- Ground-water and surface-water relationships in the Ozark area.

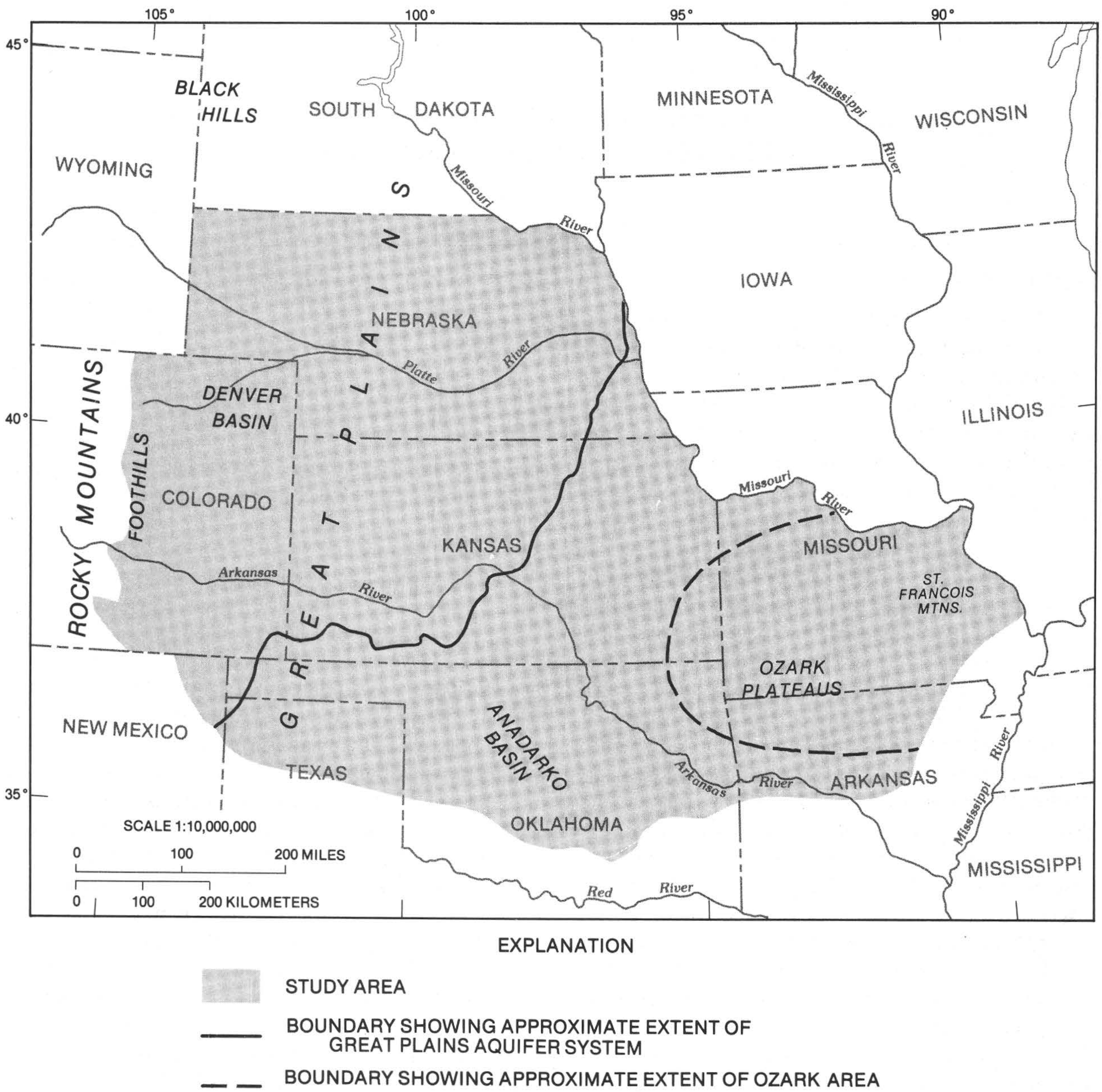


Figure 80.—Study area of the Central Midwest regional aquifer system.

•Recharge to and discharge from the shallow water-table.

•Hydrogeologic controls of recharge to the Great Plains aquifer system along the foothills of the Rocky Mountains.

An evaluation of past and future impacts on the regional flow system resulting from development of ground water is being conducted.

RESULTS

Results from several of the special studies listed above have been important in developing the overall description of the Central Midwest regional aquifer system and in defining the following major regional geohydrologic units:

•Confining units consisting of Late Cretaceous and younger rocks.

•The Great Plains aquifer system in the Great Plains, consisting of "Early" Cretaceous rocks.

•Confining units of Jurassic, Triassic, Permian, and Pennsylvanian age rocks.

•An aquifer system consisting of rocks of Early Paleozoic age outside the Ozark area (western part of Interior Plains).

•The Springfield Plateau aquifer consisting of limestones of Mississippian age in the Ozark area.

•The confining unit that separates the Springfield Plateau aquifer from the underlying Ozark aquifer in the Ozark Area.

•The Ozark aquifer consisting mostly of dolomites of Lower Ordovician and Cambrian age in the Ozark area.

•A confining unit of shale and dolomite that separates the Ozark aquifer and the underlying St. Francois aquifer.

•The St. Francois aquifer of sandstone of Cambrian age.

•The crystalline basement.

The confining units of Late Cretaceous and younger rocks overlying the Great Plains aquifer system in the Great Plains consist mainly of shale and range in thickness from featheredge at the outcrop to about 8,000 feet in the Denver basin. The shale has extremely low permeability except where microfractures that exist to depths of about 800 feet, markedly increase the vertical permeability.

The Great Plains aquifer system in the Great Plains, generally referred to as the "Dakota aquifer" in the hydrologic literature is one of the most extensive aquifer systems in North America. The aquifer system commonly consists of an upper unit (Maha aquifer) and a lower unit (Apishapa aquifer) separated by a confining unit (fig. 81). Both the Maha and Apishapa aquifers consist primarily of loosely cemented fine sand. In the study area, regional ground-water flow through the Great Plains aquifer system is east-northeastward towards discharge areas (fig. 82).

Thick shale, salt, and anhydrite of Jurassic to Pennsylvanian age restrict vertical flow between the Great Plains aquifer system and the underlying aquifers. The shale has very low permeability, and the anhydrite and salt are virtually impermeable. Thus, these sediments form a very effective confining unit.

Lower Paleozoic rocks, exclusive of the Ozark area, comprise a separate regional aquifer system in the study area. This aquifer system contains a large quantity of saline water. Permeable zones, resulting from paleokarst development in dolomite, exist in this aquifer system.

Within the Ozark area, the Springfield Plateau and Ozark aquifers are important sources of water. The Springfield Plateau aquifer, the upper aquifer, is composed mainly of fractured limestone of Mississippian age. The Ozark aquifer, except in northwestern Arkansas, usually yields more water than the Springfield Plateau aquifer and is composed mostly of fractured dolomite of Ordovician and Cambrian age. At most locations, the two aquifers are separated by a confining unit of Chattanooga Shale. The basal water-yielding unit in the Ozark area is the St. Francois aquifer consisting of sandstone of Cambrian age.

Throughout the study area, the basement is composed mostly of fractured igneous and metamorphic rocks of Cambrian and Precambrian age. The rocks generally have moderate to low permeability, depending on the degree of fracturing. Because there is no information indicating the magnitude or direction of flow through the basement rock, it is being treated regionally in this study as a no-flow boundary of the delineated aquifer system.

Rapid interchange of surface water and ground water characterizes the Ozark Plateaus, especially at locations where the fractured limestone and dolomite of the Springfield Plateau and Ozark aquifers crop out or are near land surface. Many springs are present and karst topography is well developed locally. Large gains and losses of streamflow are evident in much of the Ozark area, thus determination of the relations between the aquifers and streams is essential in evaluating the ground-water flow system

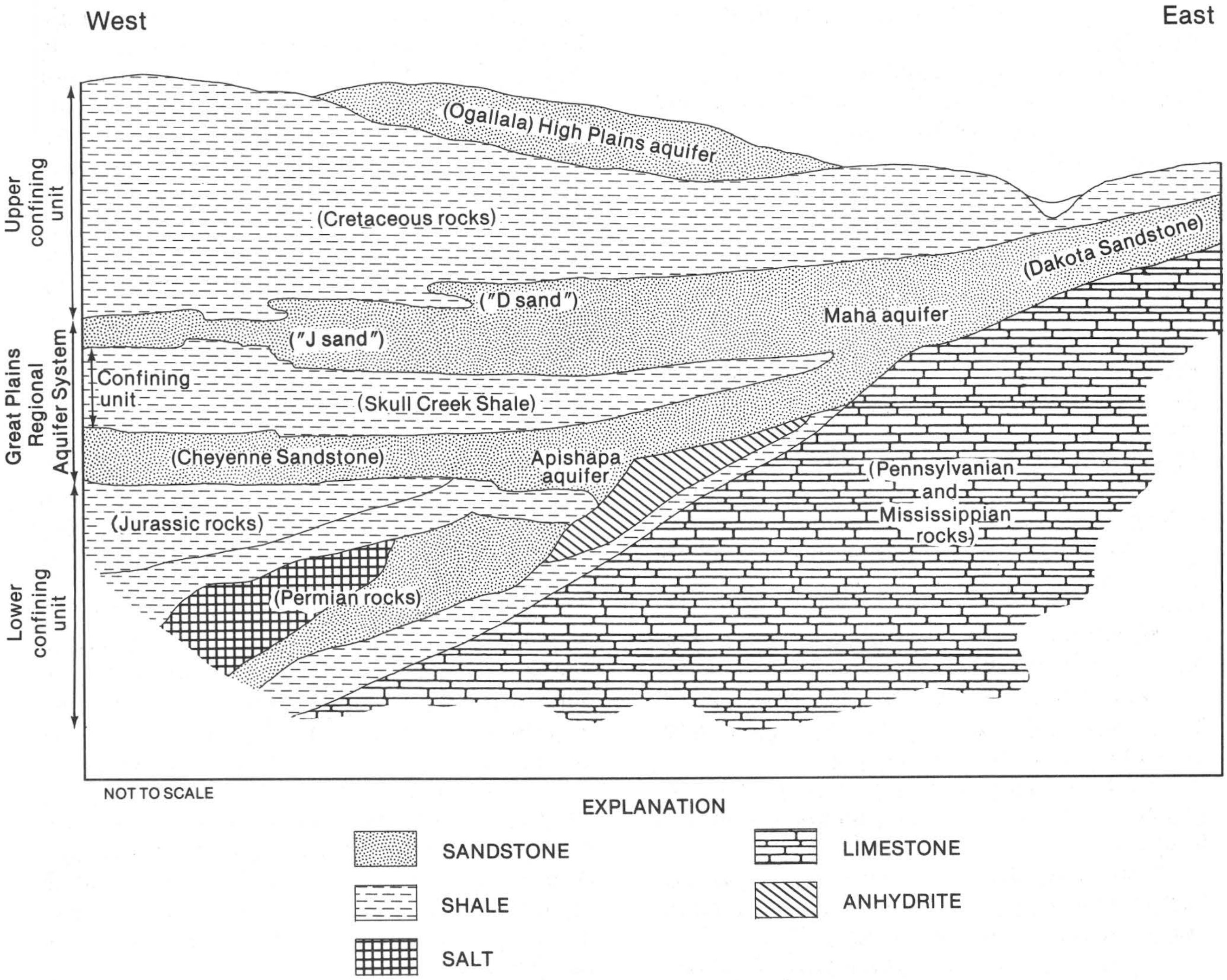
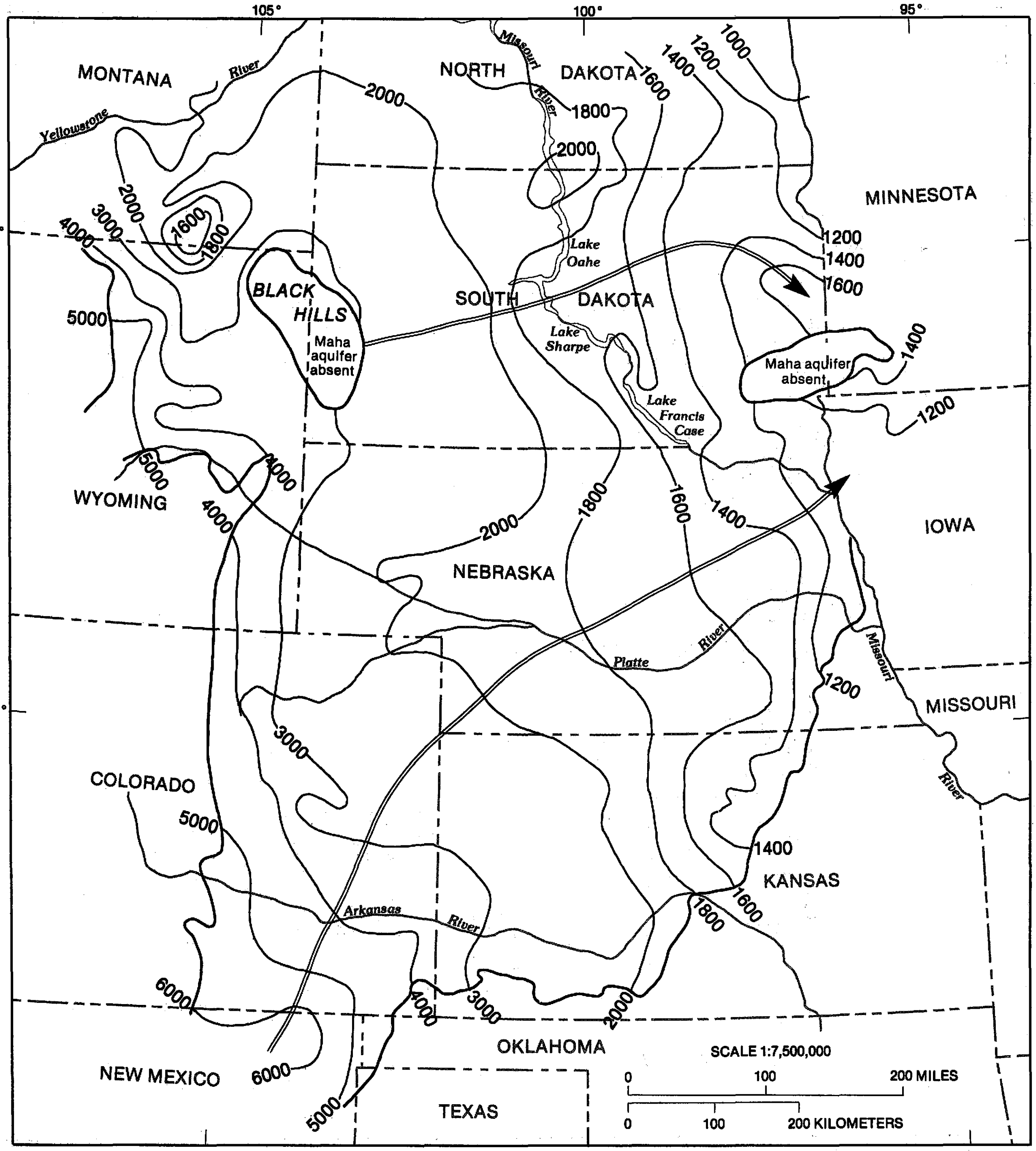


Figure 81.—Generalized geologic section of the Maha and Apishapa aquifers of the Great Plains regional aquifer system in central Midwest.

in the Ozark area. Accordingly, a study was made of the Ozark area in parts of Arkansas, Kansas, Missouri, and Oklahoma to determine the flow characteristics of selected springs and streams. Hedman and others (1984) summarized flow-duration data for continuous-record spring- and streamflow-gaging stations. They also included low-flow frequency data for continuous-record spring- and streamflow-gaging stations and for miscellaneous-measurement sites as well as values of mean annual runoff. A finding of particular interest is that the difference between annual precipitation and runoff (which represents the sum of consumptive use and evapotranspiration) has a relatively uniform value of about 29 inches across the Ozark area; although the annual precipitation ranges from 34 to 56 inches.

Information describing flow to and from the water-table aquifer is needed as data-input for simulating the regional ground-water flow system. As a part of this study, Dugan (1984) developed a procedure to determine such flow. In general, the amount of water moving through the unsaturated zone above the water table is a function of many factors, such as soil type, slope of land surface, vegetation, soil permeability, water-retention properties, depth to the water table, and climatic factors such as precipitation, solar radiation, and temperature.

Determining the direction of flow in an aquifer containing water with different densities at different locations is a complex problem. Analysis usually is accomplished by modeling the flow of the variable density fluid in three dimen-



EXPLANATION

–1800– POTENTIOMETRIC CONTOUR -- Shows altitude of potentiometric surface. Contour intervals 200 and 1000 feet. Datum is sea level

⟹ GENERALIZED GROUND-WATER FLOW DIRECTION

—— LIMIT OF MAHA AQUIFER

Figure 82.—Preliminary potentiometric surface of the Maha aquifer of the Great Plains regional aquifer system in the central Midwest (modified from Helgesen and others, 1982).

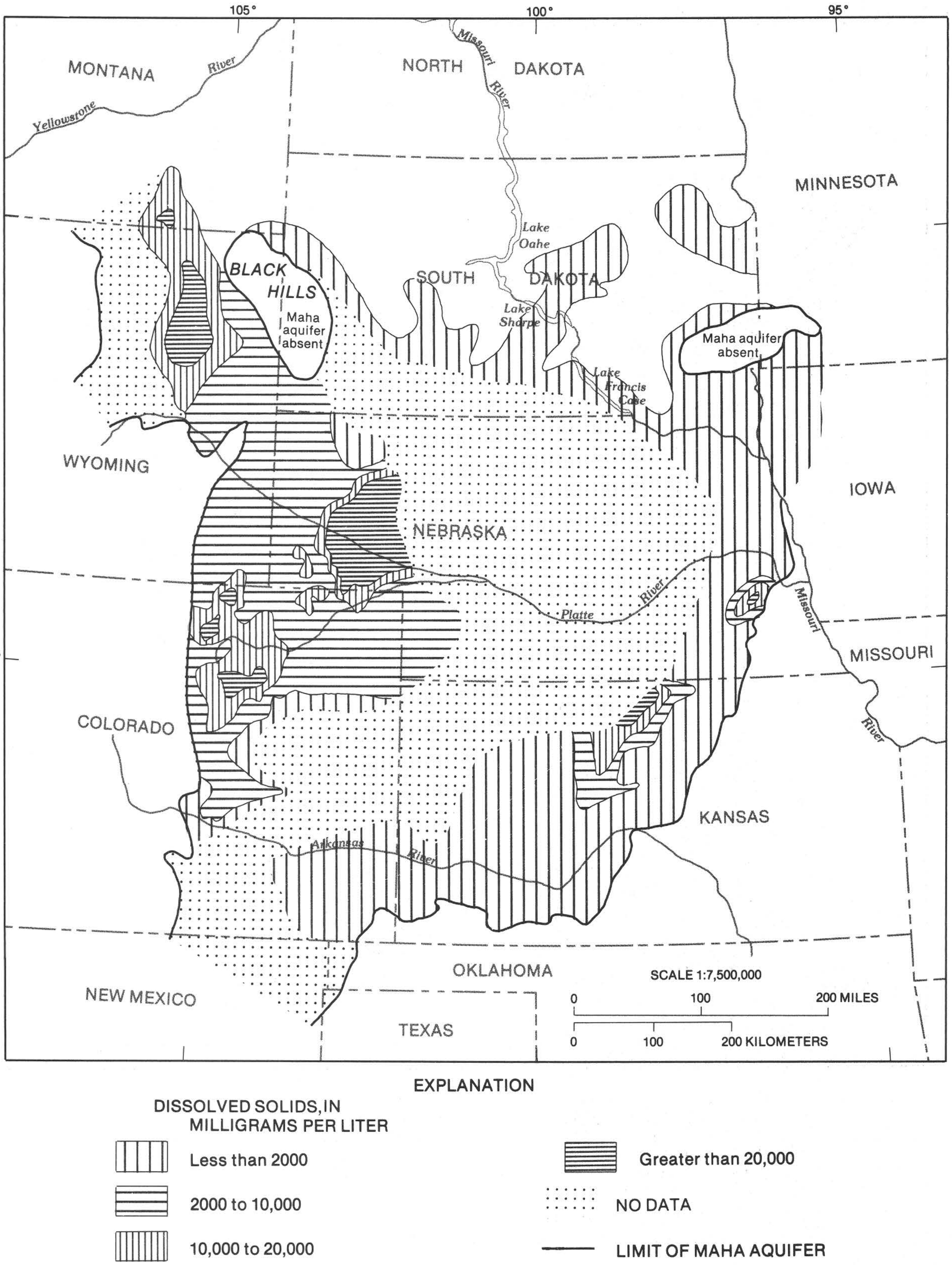


Figure 83.—Dissolved-solids concentrations of ground water in Maha aquifer of the Great Plains regional aquifer system in the central Midwest (modified from Leonard and others, 1983).

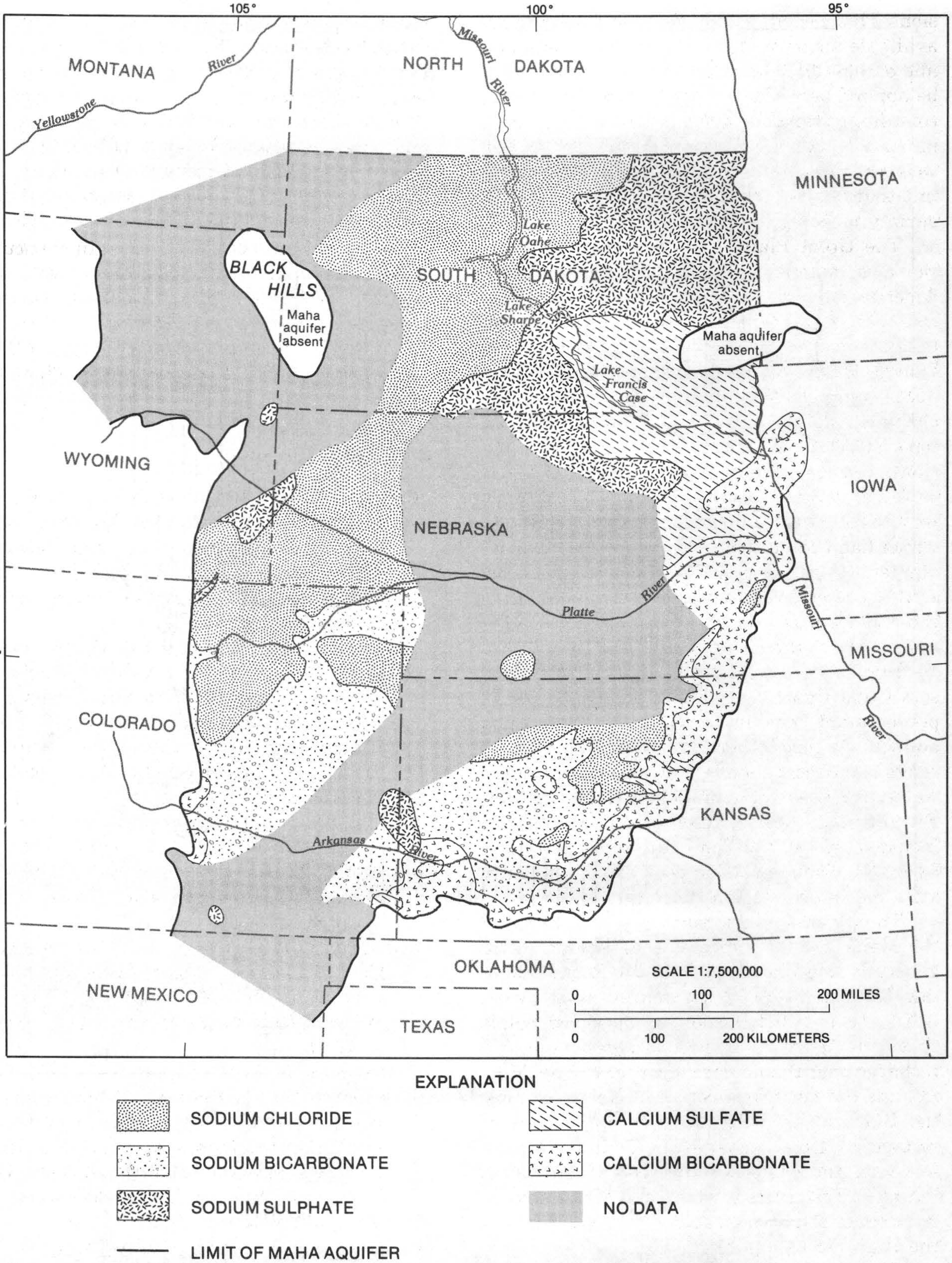


Figure 84.—Dominant ions of ground water in the Maha aquifer of the Great Plains regional aquifer system in the central Midwest (modified from Leonard and others, 1983).

sions. This modeling requires more data than are available for most of the study area. Jorgensen and others (1982) have developed a technique to be applied between two points in a flow system containing variable density water that determines whether a component of flow exists between the two points. This technique requires less information than the three-dimensional variable density modeling technique.

The Great Plains aquifer system is one of the most extensive aquifer systems in North America. No regional data synthesis and evaluation of the aquifer characteristics had been undertaken since the classic work by Darton (1905). During the RASA study, Helgesen and others (1982) summarized the important work and the changes in concepts that have occurred since Darton's 1905 study and concluded that conceptualization of the geohydrology was handicapped by the confusing nomenclature used in describing the various sandstone units of Lower Cretaceous age, which had been collectively called the "Dakota aquifer." Also, Leonard and others (1983), described the geohydrology and hydrochemistry of the Maha aquifer which is considered to be the upper part of the Great Plains aquifer system.

Figure 83 shows the distribution of dissolved solids and figure 84 shows the distribution of predominant ions in the water of the Maha aquifer. As used therein, "predominant ion" refers to the most abundant cation or anion in the water, as measured, in milliequivalents per liter. For simplicity, mixed anion or cation types are included within the dominant types shown in figure 84. Their inclusion does not significantly alter the interpretation.

The potentiometric surface of the water in the Maha aquifer indicates that ground water generally flows laterally from the Rocky Mountain foothills toward outcrop areas to the east. However, the distribution of dissolved solids shown in figure 83 strongly suggests zones of recharge near the northeast corner of New Mexico, and the northwest corner of Nebraska near the Black Hills, with an intervening zone of stagnation. Local recharge also occurs along the southern and eastern boundaries of the Great Plains aquifer system where the Maha aquifer crops out or is truncated at shallow depth beneath the High Plains aquifer.

Concentrations of dissolved solids exceeding 20,000 milligrams per liter (mg/L) in the Denver basin and ranging from about 2,000 to 10,000 mg/L along the Rocky Mountain front, west of the basin, suggests minimal recharge of freshwater in the foothill areas. Recharge from west of the Denver basin probably is restricted by low permeability resulting from facies change, including pinchout of sandstone to the west, and faults that constitute barriers to lateral flow.

These interpretations are preliminary and subject to revision; however, the chemical data shown in figures 83 and 84 reveal the complexity of the Great Plains aquifer system. In particular, when combined with the hydrologic data, they suggest extensive vertical leakage through presumed confining units, localized recharge at the truncated eastern boundary, and a partial barrier to recharge along the Rocky Mountain front.

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COLUMBIA PLATEAU BASALT REGIONAL AQUIFER-SYSTEM STUDY

By John Vaccaro, Tacoma, Washington

INTRODUCTION

The basaltic rocks that comprise the regional aquifer system underlying the Columbia Plateau are located in central and eastern Washington, northern Oregon, and a small part of northwestern Idaho (fig. 85). The Plateau covers about 70,000 mi^2 entirely within the drainage of the Columbia River and is bordered on the west by the Cascade Range, on the north and east by the Rocky Mountains, and on the south by the Blue Mountains. Major tributaries to the Columbia River on the Plateau are the Snake, Spokane, John Day, Yakima, Palouse, and Deschutes Rivers. The topography of the Plateau is varied and includes: (1) major mountains consisting of a geologically young folded region of large anticlines and synclines, and (2) low relief features.

Rocks of the Columbia River Basalt Group underlie the Columbia Plateau and compose a multilayered aquifer system which, for study purposes, has been conceptualized as a three-layered aquifer system. The three aquifer layers correspond to three basalt formations—the Saddle Mountains, Wanapum, and Grande Ronde Basalts (Swanson and others, 1979) (fig. 86). The Saddle Mountains Basalt, less extensive than the other basalts, is the youngest formation, and is unconfined where it is not overlain by saturated unconsolidated deposits. The Wanapum Basalt underlies most of the study area and is loosely confined where it is overlain either by the Saddle Mountains Basalt or unconsolidated sediments. The Grande Ronde Basalt underlies the entire study area and is loosely confined where it is overlain by younger basalts or by sediments of Holocene and Pliocene age. Both the Wanapum and Grande Ronde Basalts are unconfined near their respective margins. All basalts are connected hydraulically either directly or through sedimentary interbeds.

These three major aquifers form the regional ground-water flow system that provides water for most municipal, industrial, and domestic needs and for most of the irrigated lands outside of the Columbia Basin Irrigation Project and the Yakima River Basin.

The predominant economic activity in the study area is agriculture, and the importance of this activity to the economy of Washington and Oregon underlines the need to better understand the ground-water flow system.

The Columbia Plateau Basalt regional aquifer system study was started in 1982 and is scheduled for completion in 1986. The study was designed to address some of the hydrologic problems currently being encountered on the plateau. These problems include: (1) declining water levels of as much as 20 ft/yr (Cline, 1984), (2) the occurrence of sodium-enriched water, (3) the need for additional ground water for expanding irrigated land, (4) the lack of knowledge of the effects of increased development of the aquifer system, (5) the lack of knowledge of the interaction between ground water and surface water, and (6) the potentiality of using the low-permeability zones of the deep basalts as a national repository site for solidified high-level nuclear wastes near Richland, WA.

HYDROGEOLOGIC FRAMEWORK

Rocks that comprise the Columbia River Basalt Group were emplaced through extrusion of numerous individual lava flows through fissure eruptions. These individual lava flows range in thickness from a few inches to more than 300 feet, and average about 110 feet (Swanson and Wright,

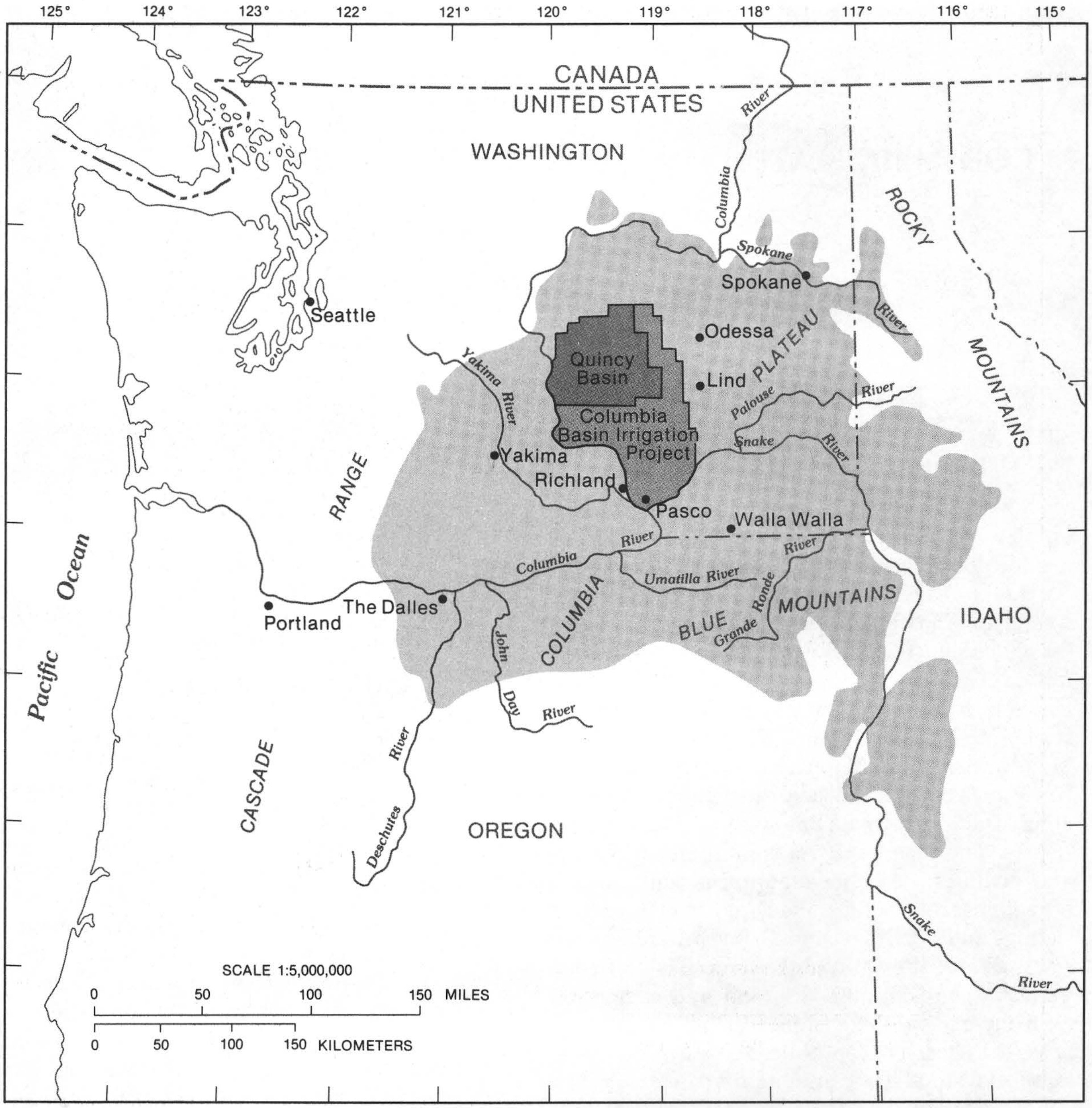

Figure 85.—Study area (shaded) of the Columbia Plateau basalt regional aquifer system.

1978). The structure of a lava flow generally consists of three units (fig. 87): the flow top, the entablature, and the colonnade. The flow tops may also include the basal part of the overlying flow that consists of pillow-palagonite complexes (Swanson and Wright, 1978). Ground-water flow is predominantly horizontal and occurs in the flow tops, which average about 5 to 30 percent of the total flow thickness. In general, the entablature and colonnade are denser than the flow tops. Because fractures are generally oriented vertically, flow through fractures in the entablature and colonnade is predominantly in the vertical direction.

The basalt flows are locally interbedded with sediments that function either as confining units or aquifers. The lithology of the interbeds varies from shalelike material to sand and gravel with fine-grained materials predominating. There are generally more interbeds in the younger basalt formation than in the older basalt. Basalt flows underlying the Columbia Plateau have been grouped into three major aquifers in this study. Movement of ground water in each aquifer is con-

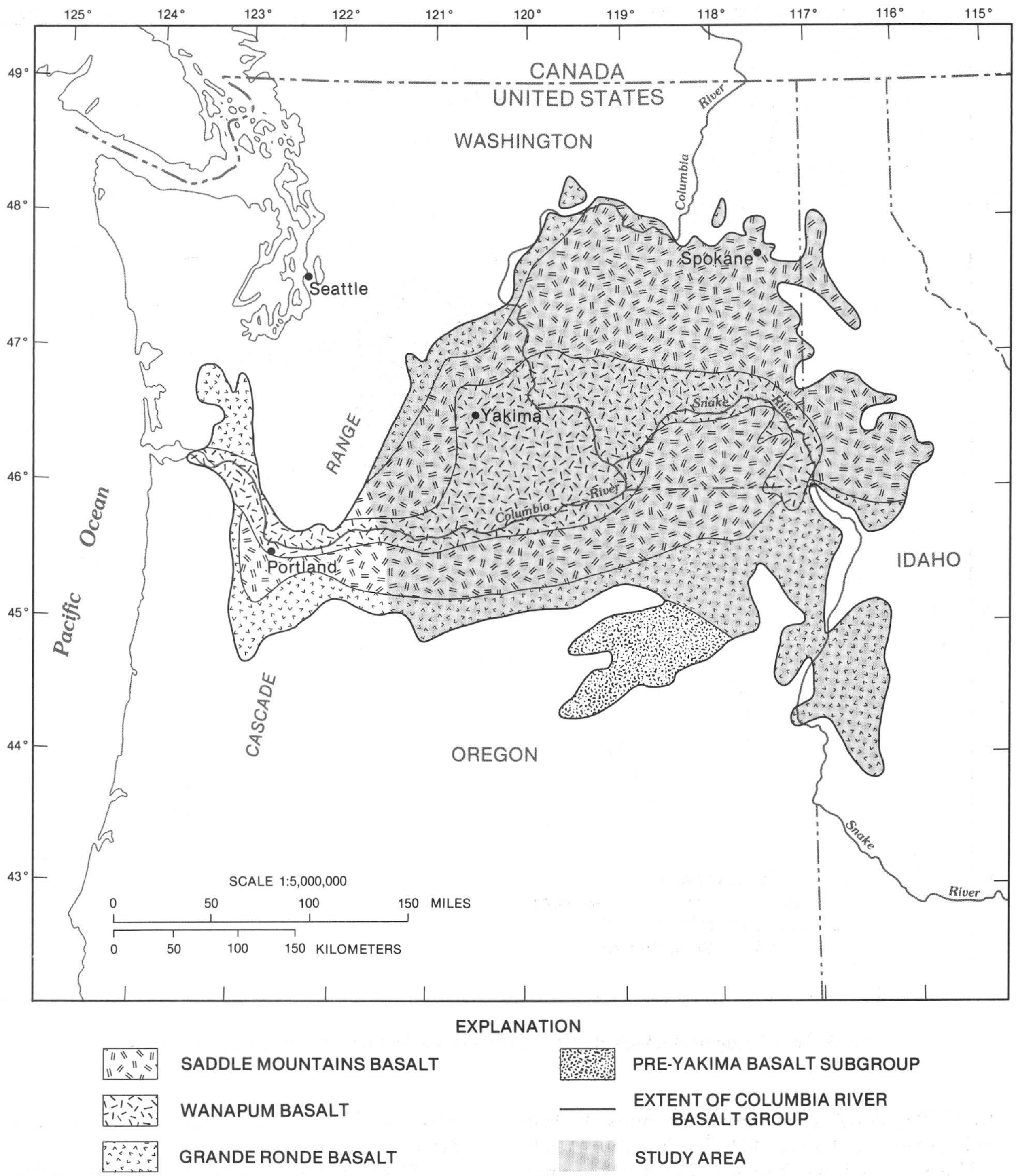

Figure 86.—Extent of the Columbia River Basalt Group.

sidered to be lateral with some vertical flow through sedimentary interbeds and fractures.

The hydrogeologic framework and generalized pattern of regional ground-water flow is depicted in figure 87. On a regional scale, water levels measured in the spring of 1983 indicate that the ground-water flow reasonably agrees with the conceptualized flow pattern shown in figure 87. Because of pumping, local geologic structures, and the inhomogeneity and anisotropy of the aquifer system, local variations in ground-water flow are common. For example, the regional water-level data show that in some areas ground-water flow is downward and that gradual,

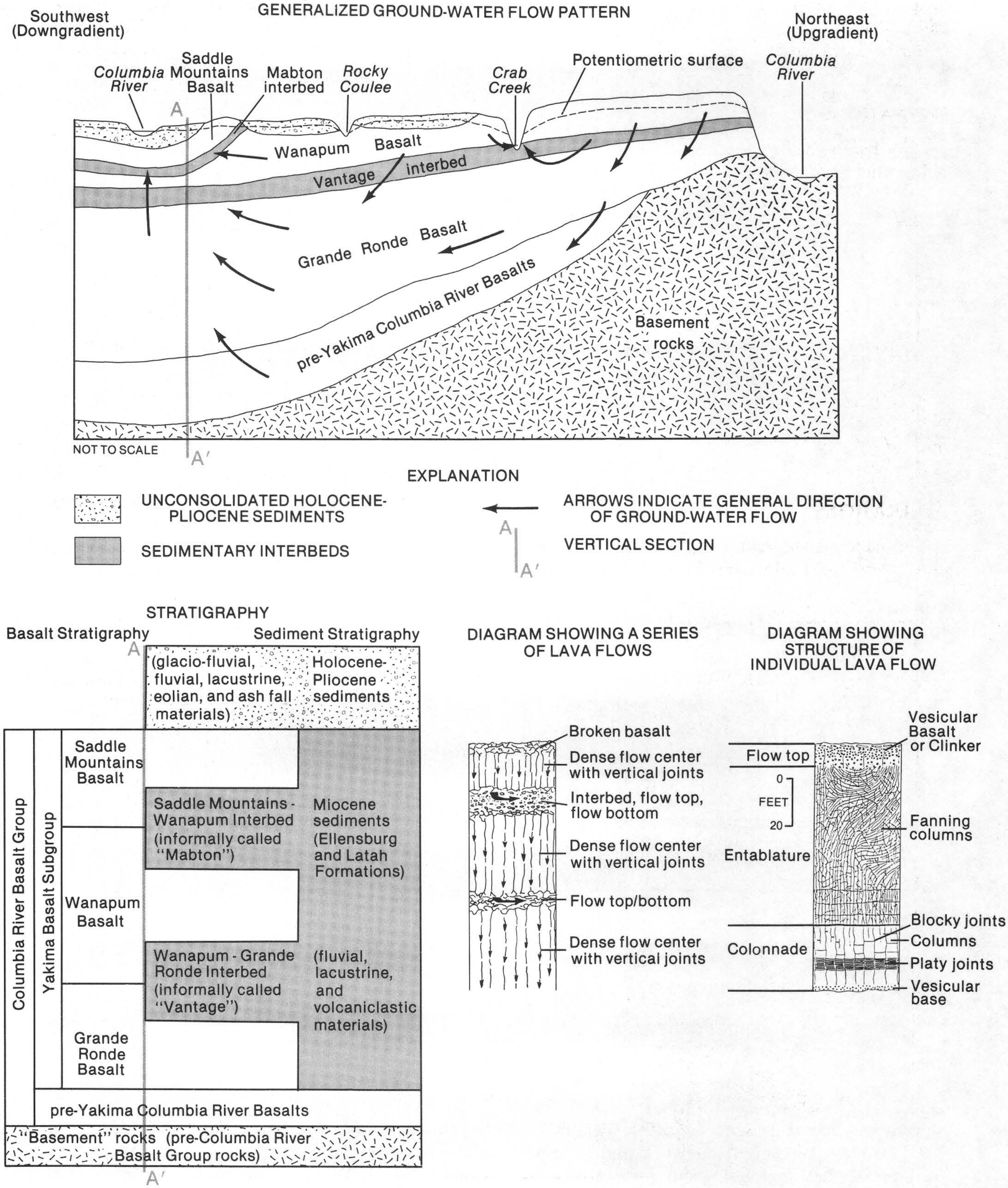

Figure 87.—Generalized ground-water flow pattern in the Columbia River basalt regional aquifer system and idealized vertical sections showing basalt interflow structures.

almost linear, changes of head occur with depth. Locally, however, changes of head with depth may be small within the basalt unit but quite large, up to 200 feet, at or near the boundaries between two basalt units.

RESULTS AND SCHEDULED WORK

As of 1984, the following work had been completed:

•The hydrogeologic framework for Washington State has been defined. This includes areal extent, thickness, tops and bottoms, and structural features for the three aquifers and the two major sedimentary interbeds.

•An observation well network was set up and two mass water-level measurements were made in the spring of 1983 and 1984.

•Potentiometric surface maps were prepared for the three basalt aquifers and the overlying sedimentary deposits for the spring of 1983.

•Conceptualization and development of a computer program of a recharge model was made.

•The first draft of a report documenting the recharge model was completed.

•The meteorological data base such as temperature, wind velocity, radiation, and others for input to the recharge model was established.

•Average annual runoff and precipitation maps were constructed.

•Computer programs for plotting, manipulating, and retrieving the hydrogeologic data were completed.

•Files of more than 4,000 wells in the U.S. Geological Survey's GWSI data base were revised.

The following work is in progress:

•Estimating ground-water pumpage and recharge rates.

•Developing a regional ground-water flow model.

•Preparing soil maps.

•Developing computer programs to estimate horizontal hydraulic conductivity values on the basis of specific-capacity test data.

•Determining boundary conditions for model simulations.

•Determining the location and altitude of surface-water bodies and identifying underlying geohydrologic units.

•Establishing rainfall-runoff relations.

•Generating land-use data for 1979 and 1983 on half-mile blocks through Landsat imagery.

•Estimating irrigation acreage associated with crop types.

•Using a model to estimate evapotranspiration.

•Developing the hydrogeologic framework for the Oregon and Idaho part of the study area.

•Completion of the 1984 potentiometric map.

Scheduled future work includes the following items:

•Complete maps for water levels, average annual precipitation, basalt-unit thickness, hydraulic conductivity, irrigation rates, and average annual runoff.

•Complete a computerized data base which will contain land-use data, model-node locations, locations and altitudes of water levels of the surface-water bodies, thickness of hydrogeologic units, and recharge and discharge rates.

•Construct pumpage maps, which will show location and average annual pumping rates.

•Calibrate the regional ground-water flow model, which will be used to evaluate current, predevelopment, and projected future conditions.

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GREAT BASIN REGIONAL AQUIFER-SYSTEM STUDY

By James R. Harrill, Carson City, Nevada

INTRODUCTION

The Great Basin regional aquifer system study was started in 1980 and is scheduled for completion in 1985. The study area encompasses about 140,000 mi^2 in parts of Nevada, Utah, and adjacent States (fig. 88). The area is characterized by generally north-trending mountain ranges which have a width ranging from 5 to 15 miles. The mountain ranges rise 1,000 to 5,000 feet above adjoining valleys. The widths of the valleys are about the same as widths of the adjacent mountain ranges. The valleys are typically elongated, and many extend more than 50 miles in a north or northeast direction. The area has a complex geologic history that includes major episodes of sedimentation, igneous activity, orogenic deformation, and continental rifting. A major tectonic change occurred about 17 million years ago with the onset of extensional faulting, which has formed the major basins and ranges that characterize the topography. The Great Basin contains a regional aquifer system in that most of its separate valley basins share common geologic and hydrologic characteristics. Currently, some 242 hydrographic areas are recognized within the study area. Most include one or more structural basins and associated basin-fill aquifers. A special situation exists in eastern Nevada and western Utah, where permeable carbonate rocks underlying the basin-fill deposits form a complex ground-water flow system that reflects the characteristics of both basin-fill and carbonate aquifers.

In recent years, much of the study area has been considered for use by the MX missile system; large coal-fired powerplants are being constructed at several locations, and the potential for disposal of solidified high-level radioactive waste at the Nevada Test Site is being studied. These activities will greatly affect the ground-water resources in much of the study area within the next several decades.

Impacts from existing and anticipated developments would have both regional and local effects. However, most of the known water resources, which include the surface water and much of the ground water in basin-filled deposits, are either used or appropriated to the extent of current estimates of their availability. Therefore, continued development in some of the more intensively developed basins requires new sources of water or new methods of managing the existing resources to make more water available. This regional aquifer-system study will provide an areal perspective so that effects of intensive development can be evaluated on both a local and regional basis.

The objective of this study is to describe the aquifer systems in the Great Basin and, to the extent possible, develop techniques that can be used for quantitative evaluation of the aquifer systems.

The approach has been strongly influenced by the diverse nature of the ground-water flow systems and the large number of basins. Some hydrologic characteristics can be described on a regional basis. However, a detailed appraisal of 242 individual areas is not feasible. Consequently, the approach taken was to study selected typical areas, identify key hydrologic processes, and then attempt to transfer the knowledge developed to areas of similar hydrology. Initial efforts have been directed toward describing the regional hydrology, conducting detailed studies of representative areas, and studying selected hydrologic processes. Efforts during the final part of the study would be directed toward developing a regional approach of evaluating the aquifers in the area of the Great Basin.

PROGRESS OF THE STUDY

As of 1984, efforts to describe the regional system were mostly complete. Findings are being presented as maps or reports which are in various stages of preparation. These are:

- U.S. Geological Survey Hydrologic Atlases showing:
  - Generalized water-level contours in all basin-fill and consolidated rock aquifers within the carbonate rock province.
  - Regional hydrogeology.
  - Regional flow systems.
- Data reports such as:
  - A series of 1:250,000 quadrangle maps showing quality of ground water in the region.
  - A summary of selected hydrologic data collected by the MX missile siting investigation.
  - A field canvas of wells drilled in Utah as part of the MX missile program.

Study of important hydrologic processes is also in progress, as described below.

Recharge.—The Desert Research Institute of the University of Nevada was contracted to investigate the mechanics of recharge processes at selected sites in both northern and southern Nevada. Results of the work were being compiled into a report. In addition, a technique to estimate recharge based on the mass balance of chloride between precipitation and ground water has been developed and applied to selected areas.

Evapotranspiration.—Efforts to better understand the process of evapotranspiration have been directed toward measuring evapotranspiration rates under field conditions. Two field sites have been instrumented—one near Soda Lakes, about 50 miles east of Carson City, NV, and the other is in Smith Creek Valley, about 40 miles west of Austin, NV. Weather stations were established at each site to obtain continuous meteorologic data. The Bowan-ratio technique (which measures evapotranspiration on the basis of humidity gradients) was also used to provide a comparison with the eddy correlation measurements. Data were collected during part of the 1982 growing season and through the 1983 growing season, which were wetter than normal. Reports of this study are interesting and useful for estimating ground-water discharge.

Basin flow models.—Eight basins were selected for ground-water flow modeling, which collectively represent most conditions present in the study area (fig. 88). Two basins were modeled as parts of cooperative studies with State and local agencies, and the results are available for use in this study. Of the remaining six models, all were calibrated for steady-state conditions, two were calibrated for transient conditions, and the remaining are in final stages of transient calibrations. Grid spacing differs in each model depending on the basin characteristics and ranges from 2,000 feet to 2 miles. The model for Smith Creek Valley was completed and stressed for several proposed development alternatives.

Regional flow models.—Two regional flow models were being constructed, one is the carbonate rock province between the Great Salt Lake Desert and Death Valley, and the other is a smaller scale model of the Fish Springs flow system that drains northward toward the Great Salt Lake Desert (fig. 88). The primary purpose of these models is to test hypotheses concerning the regional flow system and to ensure that information used to describe the regional flow system is internally consistent. Both models were run for steady-state conditions. The large regional carbonate-rock flow model is being restructured with a grid spacing half the initial size of 10 by 15 miles to allow for more detailed analyses.

Regional-scale analysis.—Intermediate and final results of this study were being evaluated in terms of their significance to the overall regional flow system. Constraints posed by geochemical, geologic, geophysical, and hydrologic evidence was generally in good agreement among all elements, and overall analysis based on various techniques of investigation appeared to be converging.

SIGNIFICANT RESULTS

Regional flow is driven by hydraulic gradients that extend over long distances. Figure 89 shows the regional flow of the Great Basin aquifer system based primarily on the lowest water-level altitudes in each basin. The regional flow is apparently toward either the Colorado River or major regional discharge areas. The 242 identified basin areas had been grouped into 39 major flow systems (fig. 90). Of these, 14 are single-basin systems, the rest are multibasin systems. Large multibasin systems outside the carbonate-rock province are generally coincident with major drainage systems. Large multibasin systems within the carbonate-rock province typically have little surface-water flow; instead they may contain ground-water flow paths that are more than

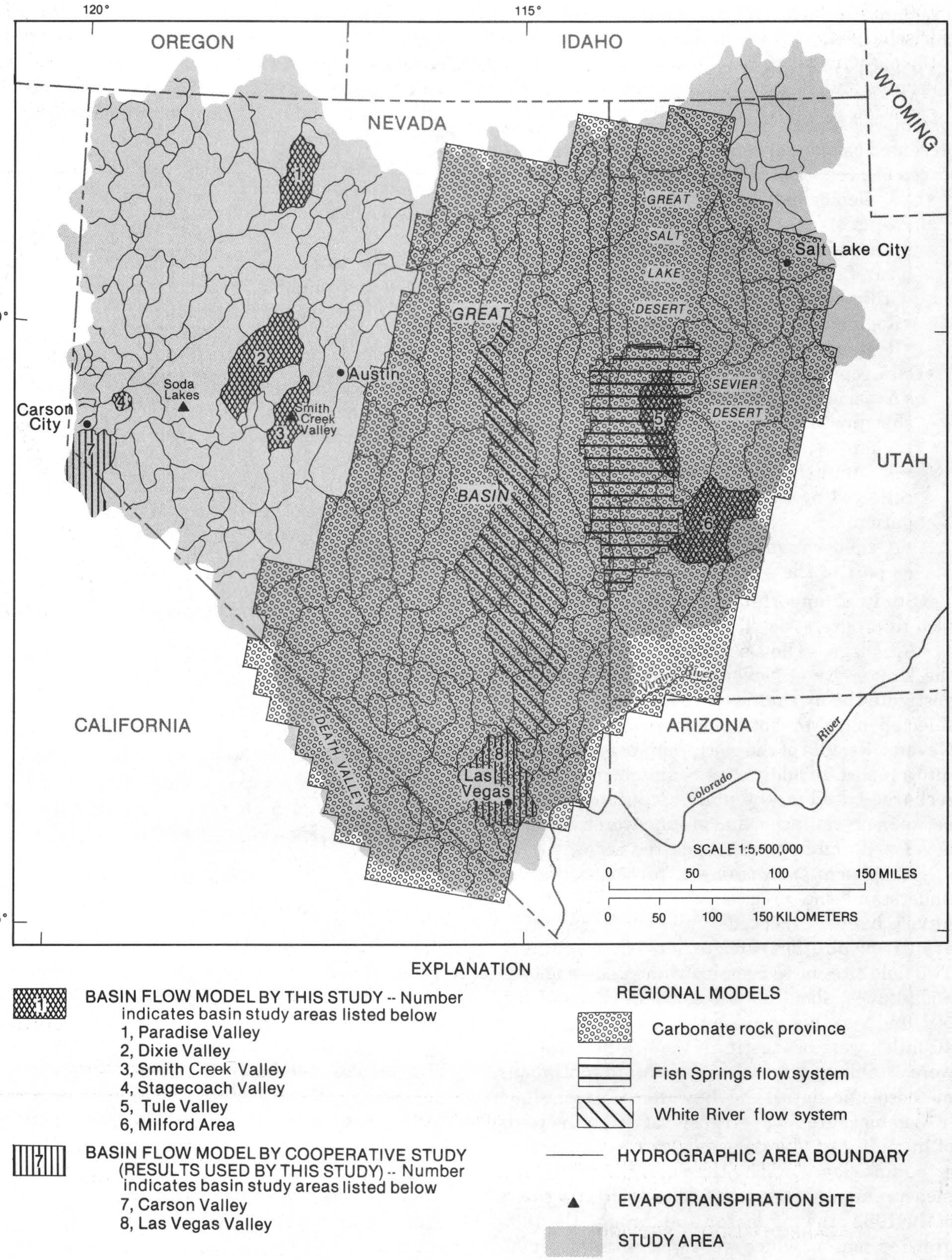

Figure 88.—Study area and types of ground-water flow models, Great Basin regional aquifer system.

100 miles long and traverse several basins. Discharge from these multibasin systems within the carbonate-rock province is typically to large springs and is usually consumed by evapotranspiration in the vicinity of the springs.

Geochemical studies of the White River ground-water flow system suggest that the system can be further subdivided into several subsystems, with relatively small quantities of flow between the subsystems. Analysis of

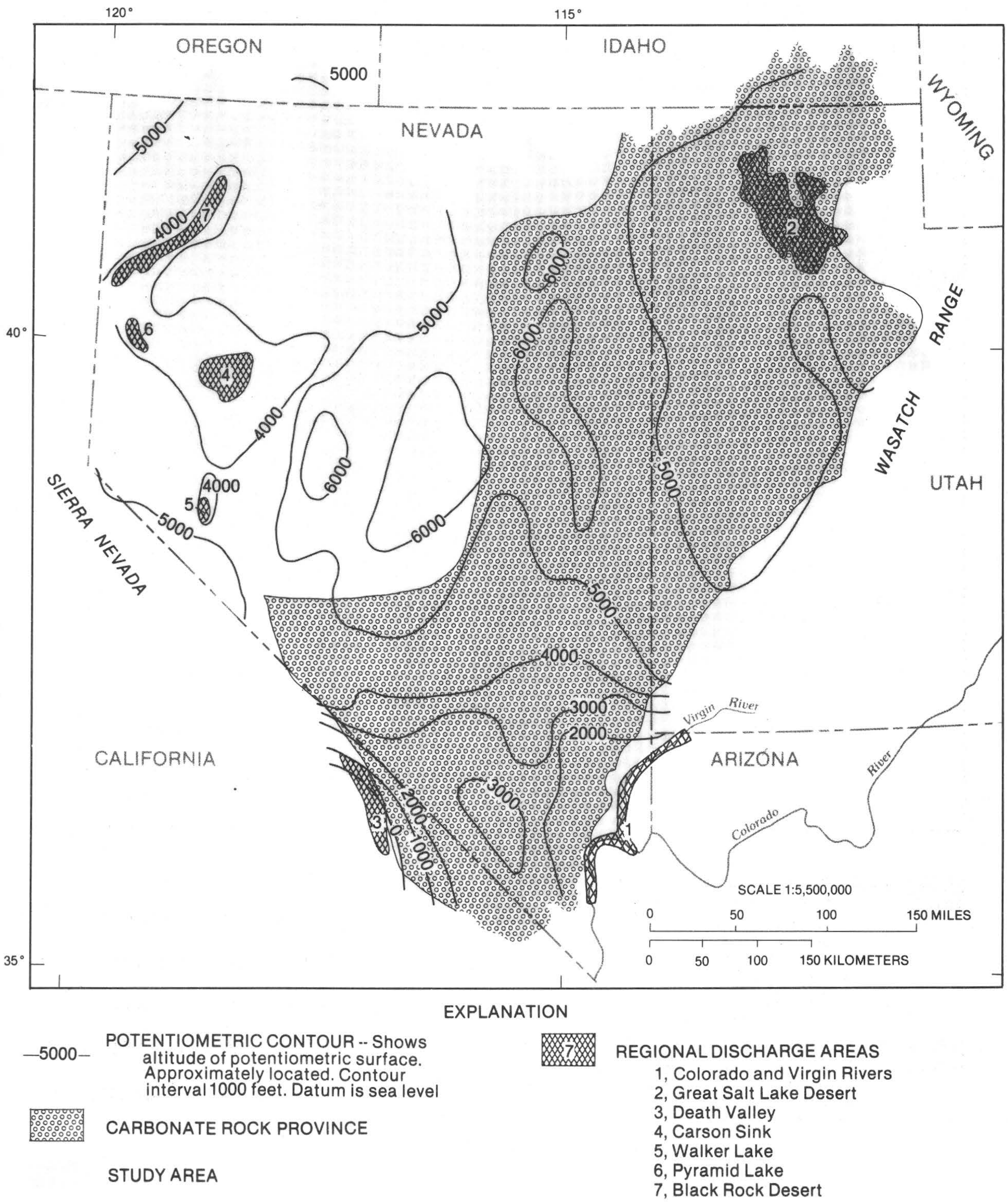

Figure 89.—Regional ground-water flow of the Great Basin regional aquifer system.

aeromagnetic data suggests that subsurface masses of igneous rock probably act as barriers and provide significant controls on regional flow in parts of east-central Nevada and western Utah. Regional analysis of hydrologic conditions in southwestern Utah suggests that the transmissivity of the carbonate rocks is higher than originally anticipated, and that some degree of hydraulic continuity exists between basins throughout that part of the area. Much of this con-

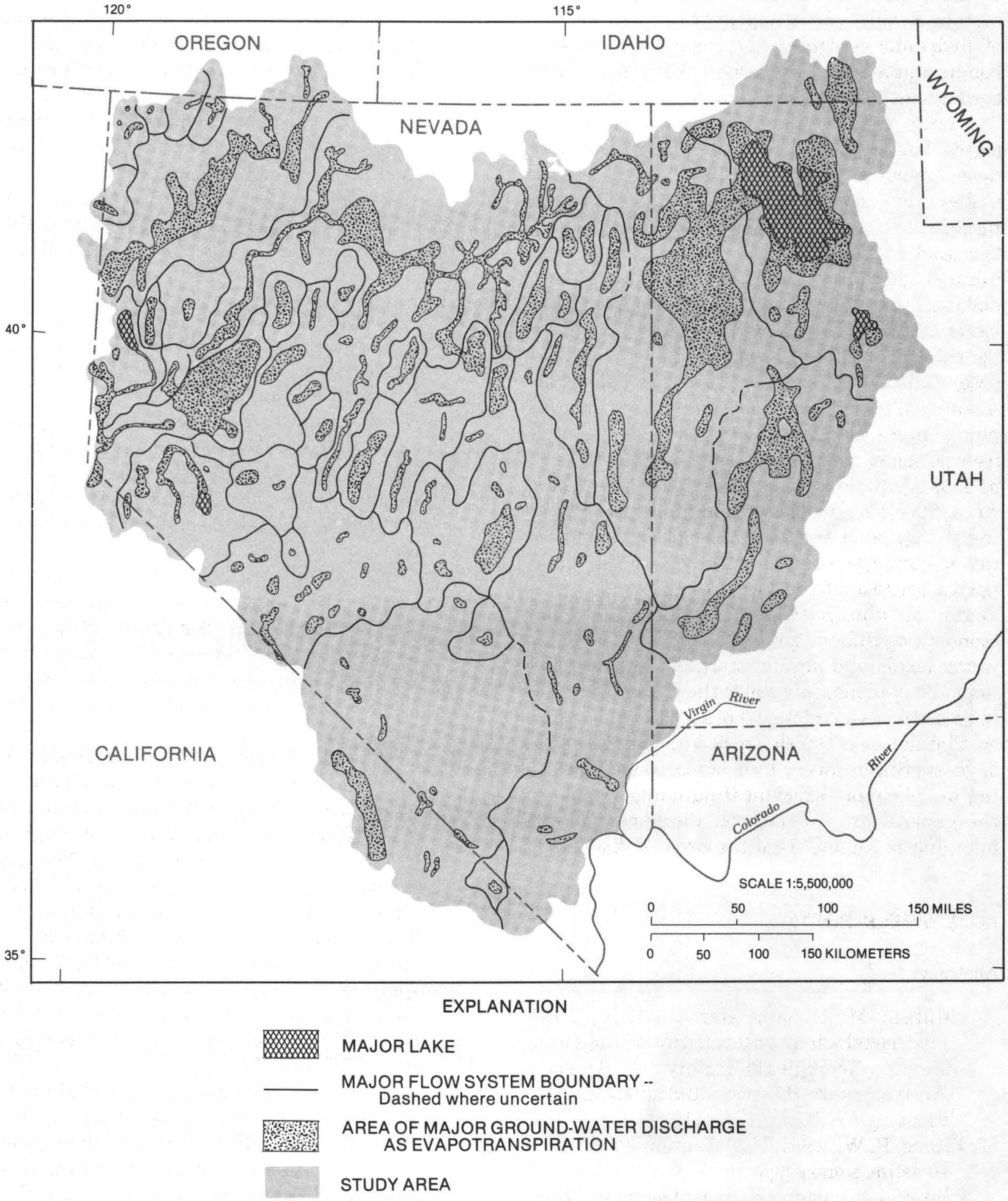

Figure 90.—Identified major ground-water flow systems and areas of major ground-water discharge of the Great Basin regional aquifer system.

tinuity is inferred from information obtained at shallow depth. For example, shallow test drilling in the Sevier Desert of Utah (fig. 88) suggests that Sevier Lake is not the final terminus of the Sevier flow system but that there may be some leakage west and north to the Great Salt Lake Desert flow system. Finally, there appears to be some degree of hydraulic continuity between the deep carbonate aquifer and the overlying basin-fill aquifer throughout most of the carbonate-rock province.

Smith Creek Valley is generally typical of closed basins. Simulation of several proposed development alternatives using the Smith Creek Valley flow model illustrates that long-term response to pumping is a function of both the location and rate of pumping. Efficient capture of natural discharge with minimal water-level declines is hydrologically a viable alternative. However, less strategic distributions of pumping, which still provide for reasonably efficient capture of discharge, may be preferable because of the additional freedom allowed in locating pumping centers. Locating pumping centers remote from areas of natural discharge results in long periods of aquifer-storage reduction and declining water levels regardless of pumping rate. Pumping in excess of the estimated predevelopment rate of recharge is hydrologically a feasible alternative because of the large quantities of water stored in the basin-fill aquifers. However, economic considerations associated with declining water levels and aquifer-storage reduction will most likely ultimately cause the pumping rate to be reduced as the thickness of saturation is reduced. Simulation of Smith Creek Valley flow system suggests that recovery from sustained overpumping may be more efficient if pumping is close to the general area of natural discharge than if pumping is remote from the area of discharge.

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GULF COASTAL PLAIN REGIONAL AQUIFER-SYSTEM STUDY

By Hayes F. Grubb, Austin, Texas

INTRODUCTION

The Gulf Coastal Plain regional aquifer system study was started in 1980 and is scheduled for completion in 1988. The study area includes about 225,000 mi^2 of the Gulf Coastal Plain in parts of Alabama, Arkansas, Florida, Illinois, Kentucky, Mississippi, Missouri, Tennessee, and Texas and all of Louisiana (fig. 91). The thick wedge of sediments of Tertiary and younger age, yields large quantities of water for municipal, industrial, and agricultural use.

In addition to the objectives of all RASA studies, specific objectives or approaches of this study include: (1) evaluation of effects of highly saline water on the regional flow system, and (2) evaluation of potential for compaction of confining units as a result of changes in fluid pressures.

The principal findings of the study, as of 1984, consist largely of the development of a conceptual framework for studying the regional aquifers, identification of data sources, compilation of the data into computer files, and preliminary simulations of the ground-water flow system.

CONCEPTUAL GEOHYDROLOGIC FRAMEWORK

A regional geohydrologic framework was developed based on the concept of the aquifer system. This concept recognizes the interconnection of aquifers and associated confining units and provides guidelines for simulation of the ground-water flow on a regional scale where geologic units are aquifers in one part and confining units in another part of the study area.

Three aquifer systems are delineated in the study area: the Mississippi Embayment aquifer system, the Texas Coastal Uplands aquifer system, and the Coastal Lowlands aquifer system (fig. 92). The designation of these aquifer systems is based upon differences in geohydrology, regional flow patterns, and the presence of two or more significant aquifers of regional extent.

The differences in flow patterns and the distribution of sediment characteristics provided the basis for delineating the three aquifer systems. Ground-water flow in the Mississippi Embayment aquifer system is generally from the outcrop areas around the margin of the Mississippi embayment toward the Mississippi River. In the Texas Coastal Uplands aquifer system the flow is generally from the outcrop areas toward the Gulf of Mexico. The bulk of fine-grained sediments that comprise the confining units in the Mississippi Embayment aquifer system and in the Texas Coastal Uplands aquifer system are generally massive clay layers that can be correlated over large areas, both at the surface and in the subsurface. By contrast, the fine-grained sediments that comprise the confining units within the Coastal Lowlands aquifer system typically occur in thinner layers, cannot be correlated over large areas, and occur as isolated bodies throughout the aquifer system.

Mississippi Embayment Aquifer System

The Mississippi Embayment aquifer system consists of sediments of Tertiary age above the uppermost massive marine clay of the Midway Group and below the lowermost massive marine clay of the Jackson Group, or the Jackson and Vicksburg Groups where they are undifferentiated (table 2). The Upper Cretaceous sediments that underlie the Paleocene Midway Group in an area of about 25,000 mi² in the northernmost part of the Mississippi embayment are also included in the Mississippi Embayment aquifer system. The Upper Cretaceous sands are laterally isolated from the extensive aquifers of Cretaceous age to the south and southeast by calcareous and clayey strata of low permeability. The sediments which make up the Mississippi Embayment aquifer system are exposed at land surface from the southern tip of Illinois to the Jackson-Vicksburg outcrop belt in central Louisiana, central Mississippi, and southwestern Alabama. The eastern limit of this aquifer system is the contact between sediments of Cretaceous and Paleozoic age in western Kentucky and the outcrop belt of the Ripley Formation of Cretaceous age in northeastern Mississippi and western Tennessee. The western limit of this aquifer system is the Texas-Louisiana State line, which runs in a north-south direction approximately through the center of the surface expression of the Sabine uplift, which disrupts the general outcrop pattern of the Tertiary sediments.

Eleven aquifers of regional significance have been mapped by previous investigations in the Mississippi Embayment aquifer system (table 2), the oldest is the Ripley Formation of Cretaceous age (Boswell and others, 1965) and the youngest is the Mississippi River Valley alluvial aquifer of Holocene and Pleistocene age (Boswell and others, 1968). Identified aquifers of Eocene age, in ascending order, are: (1) lower Wilcox aquifer (Hosman and others, 1968), (2) Wilcox Group (Hosman and others, 1968), (3) Carrizo Sand and Meridian-Upper Wilcox aquifer (Hosman and others, 1968), (4) Carrizo and Meridian Sand aquifer (Payne, 1975), (5) Sparta hydraulic system (Payne, 1968), (6) Sparta Sand (Hosman and others, 1968), (7) Memphis aquifer (Hosman and others, 1968), (8) Cockfield Formation (Hosman and others, 1968), and (9) Cockfield aquifer system (Payne, 1970). The number of aquifers to be delineated in the Mississippi Embayment aquifer system during this study has not been decided. Similarly, the delineation of aquifers in the other two aquifer systems—Texas Coastal Uplands aquifer system and the Coastal Lowlands aquifer system—is also pending.

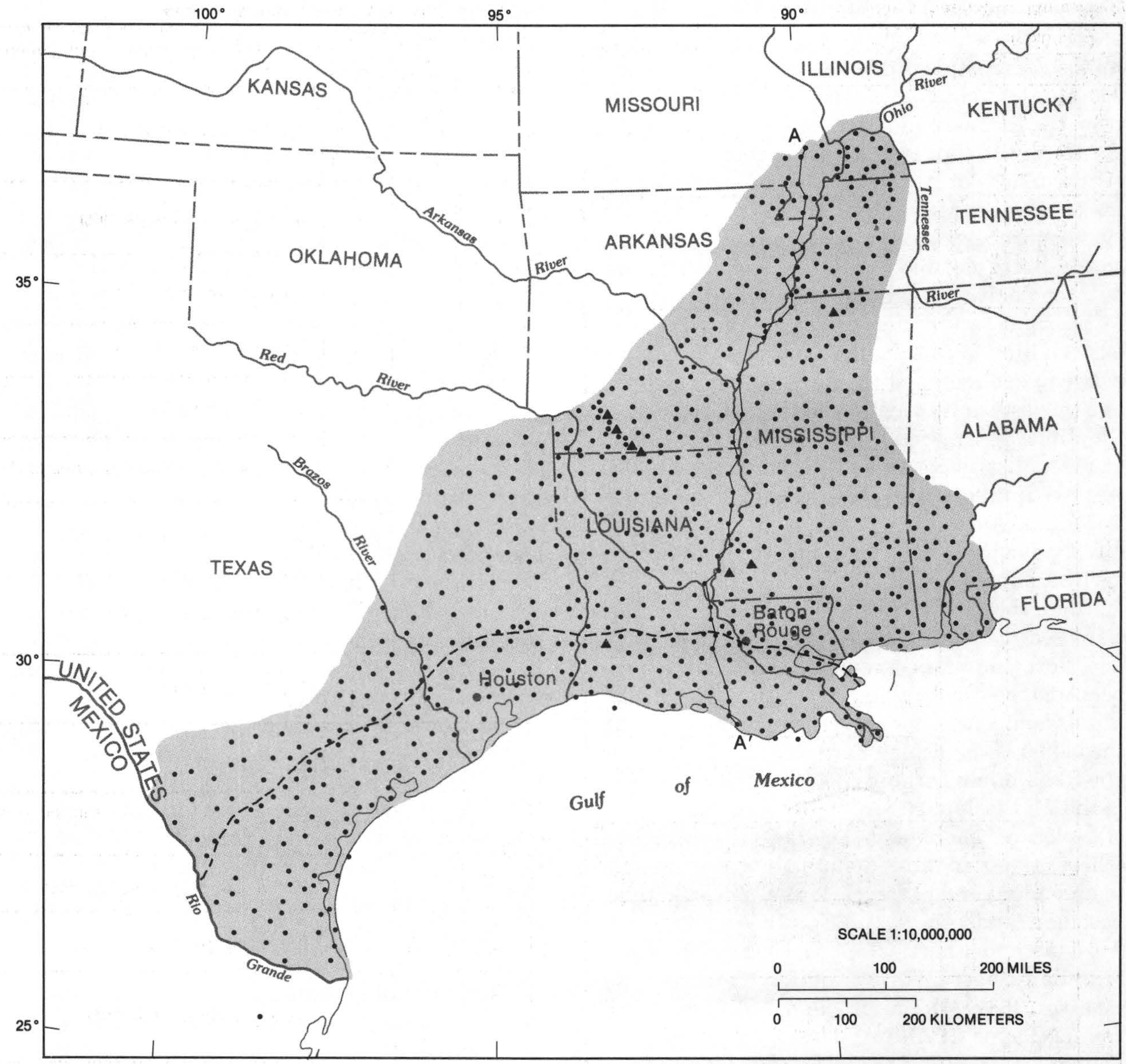
Texas Coastal Uplands Aquifer System

The Texas Coastal Uplands aquifer system consists of sediments of Tertiary age above the uppermost massive clay of the Midway Group and below the lowermost massive clay of the Jackson Group, or the Jackson and Vicksburg Groups where they are undifferentiated. The sediments which make up the Texas Coastal Uplands aquifer system are exposed at land surface gulfward of a narrow band of the Midway outcrop, and extend to a narrow outcrop band of the Jackson and Vicksburg Groups which is located about 100 miles inland from the Gulf of Mexico. The southwestern limit of this aquifer system extends beyond the Rio Grande into Mexico but this area is not included in this study. The eastern limit of this aquifer system is the Texas-Louisiana State line. Many of the aquifers and confining units of the Texas Coastal Uplands aquifer system extend laterally into equivalent hydrologic units that are part of the Mississippi Embayment aquifer system.

Seven aquifers of regional significance have been identified by different investigators in the Texas Coastal Uplands aquifer system (table 2). All are of Eocene age and range from the oldest in sediments of the Wilcox Group to the youngest in sediments of the Yegua Formation of the Claiborne Group. The seven identified aquifers are: (1) Lower Wilcox (Jones and others, 1976), (2) Upper Wilcox (Jones and others, 1976), (3) Carrizo-Wilcox aquifer (Klemt and others, 1976), (4) Carrizo Sand (Payne, 1975), (5) Sparta Sand (Payne, 1968), (6) Queen City Sand (Payne, 1972), and (7) Yegua aquifer system (Payne, 1970).

Coastal Lowlands Aquifer System

The Coastal Lowlands aquifer system consists of sediments of Miocene and younger age above the uppermost massive clay of the Vicksburg Group or the Jackson and Vicksburg Groups where they are undifferentiated. The sediments which make up the Coastal Lowlands aquifer system are exposed at land surface in Alabama, Florida, Louisiana, Mississippi and Texas, and extend gulfward from the Jackson-Vicksburg outcrop belt to the Gulf of Mexico. The eastern limit of this aquifer system is the Escambia River in Florida and the Alabama River and Big Escambia Creek in southwestern Alabama. Sediments of Miocene age extend beyond the eastern limit of the aquifer system but they are thin and are significant only as a source of recharge to the Floridan aquifer system. The Coastal Lowlands aquifer system extends to the west beyond the Rio Grande into Mexico but this area is not included in this study.



EXPLANATION

• LOCATION OF WELL WITH LOGS

▲ LOCATION OF TWO OR MORE WELLS

A—A′ LINE OF SECTION ON FIGURE 93

- - - - LINE DIVIDING BASE OF THE FLOW SYSTEM -- Coastal Uplands confining unit is the base of the flow system to the north of line. The top of the geopressured zone is the base of the flow system to the south of line

STUDY AREA

Figure 91.—Study area, location of geophysical logs used to define the hydrogeologic framework of the Gulf Coastal Plain regional aquifer system.

Mississippi Embayment aquifer system

Geologic unit: System	Series	Group	Aquifers, (references), and confining units
Quaternary	Pleistocene and Holocene		Mississippi River Valley Alluvial aquifer (Boswell, and others, 1968)
Tertiary	Eocene and Oligocene	Jackson and Vicksburg	Coastal Lowlands confining unit [1]
	Eocene	Claiborne	Cockfield aquifer system (Payne, 1970)
			Cockfield Formation (Hosman and others, 1968)
			Memphis aquifer (Hosman and others, 1968)
			Sparta Sand (Hosman and others, 1968)
			Sparta hydraulic system (Payne, 1968)
			Carrizo and Meridian Sand aquifer (Payne, 1975)
			Carrizo Sand and Meridian-Upper Wilcox aquifer (Hosman and others, 1968)
		Wilcox	Wilcox Group (Hosman and others, 1968)
	Paleocene		Lower Wilcox aquifer (Hosman and others, 1968)
		Midway	Coastal Uplands confining unit [1]
Cretaceous	Upper Cretaceous		Ripley Formation (Boswell and others, 1965)

[1] The confining units are defined as the massive clay section (with interbedded sands) of the Midway Group and the undifferentiated Jackson and Vicksburg Groups that are recognizable on geophysical logs. The recognizable lithologic unit may not be equivalent to the geologic unit as determined by fossils or other means of correlation and dating, because the upper or lower part of either the Midway Group or the undifferentiated Jackson and Vicksburg Groups may be sandy and therefore included in the adjacent aquifer.

Texas Coastal Uplands aquifer system

Geologic unit: System	Series	Group	Aquifers, (references), and confining units
Tertiary	Eocene and Oligocene	Jackson and Vicksburg	Coastal Lowlands confining unit [1]
	Eocene	Claiborne	Yegua aquifer system (Payne, 1970)
			Queen City Sand (Payne, 1972)
			Sparta Sand (Payne, 1968)
			Carrizo Sand (Payne, 1975)
			Carrizo-Wilcox aquifer (Klemt and others, 1976)
		Wilcox	Upper Wilcox (Jones and others, 1976)
	Paleocene		Lower Wilcox (Jones and others, 1976)
		Midway	Coastal Uplands confining unit [1]

Coastal Lowlands aquifer system

Geologic unit: System	Series	Group	Aquifers, (references), and confining units	
Quaternary	Pleistocene and Holocene		Chicot Reservoir (Jones and others, 1956)	
			Chicot aquifer (Meyer and Carr, 1979)	
Tertiary	Pliocene		Evangeline aquifer (Whitfield, 1975)	
			Evangeline aquifer (Meyer and Carr, 1979)	
	Miocene		Jasper aquifer (Whitfield, 1975)	Miocene aquifer system (Gandl, 1982)
			"2000-foot" sand of the Baton Rouge area (Torak and Whiteman, 1982)	
	Eocene and Oligocene	Jackson and Vicksburg	Coastal Lowlands confining unit [1]	

Table 2.—Relation among previously mapped aquifers to geology and regional confining units in the Gulf Coastal Plain regional aquifer systems study area.

At least three aquifers of regional significance in the Coastal Lowlands aquifer system have been identified by other investigators. The oldest is in sediments of Miocene age and the youngest in sediments of Pleistocene-Holocene age. Subdivision into aquifers within this aquifer system is difficult due to the discontinuous nature of the sediments, which were deposited in fluvial and deltaic environments. This problem is illustrated by the aquifers mapped by different investigators in Mississippi and Louisiana. The Miocene aquifer system mapped by Gandl (1982) includes equivalent units that were mapped as the Jasper aquifer by Whitfield (1975) and the "2000-foot" sand of the Baton Rouge area by Torak and Whiteman (1982) as shown in table 2. Investigation of vertical head gradients in areas having heavy pumping suggests that this regional aquifer system probably can be differentiated into five aquifers regionally.

Regional Confining Units

Two regional confining units, the Coastal Uplands confining unit and the Coastal Lowlands confining unit, are the major restrictions to vertical flow in the study area. The Coastal Uplands confining unit is comprised of the predominantly fine-grained marine sediments of the Midway Group. The Coastal Uplands confining unit is assumed to be the base of the ground-water flow system in the overlying Texas Coastal Uplands aquifer system. The Coastal Lowlands confining unit is comprised predominantly of thick marine clays of the undifferentiated Jackson and Vicksburg Groups which separate the Texas Coastal Uplands and the Mississippi Embayment aquifer systems from the Coastal Lowlands aquifer system down dip of the Jackson-Vicksburg outcrop belt. Sand beds near the top of the Midway Group or near the top or base of the Jackson and Vicksburg Groups are considered to be part of the adjacent aquifers. The top and base of these confining units are determined by lithology (clay being indicative of low permeability sediments) rather than by the geologic age of the sediments.

The data available for mapping these confining units are mostly derived from geophysical logs obtained in oil and gas wells. Therefore, the confining units are defined as the massive clay sections with interbedded sands of the Midway Group (Coastal Uplands confining unit) and the undifferentiated Jackson and Vicksburg Groups (Coastal Lowlands confining unit) that can be recognized on geophysical logs.

Two confining units within the Texas Coastal Uplands and the Mississippi Embayment aquifer systems can be correlated across the study area by use of geophysical logs. They correspond closely to the Cook Mountain Formation and part of the Cane River Formation of the Claiborne Group. These confining units are defined in terms of lithology rather than being restricted to a geologic unit.

Local vertical resistance to flow within the Coastal Lowlands aquifer system is typically due to numerous clay layers that are discontinuous and generally cannot be correlated over large areas by use of geophysical logs. In the Houston, TX, area, Meyer and Carr (1979, p. 5) report that about one-half of the total thickness of the Chicot aquifer consists of discontinuous clay layers.

DATA SOURCES

Data that are useful in defining the regional geohydrologic framework are abundant in the study area. Hundreds of thousands of wells have been drilled for oil and gas exploration and production, and geophysical logs are available from commercial sources for many of these wells. The Petroleum Information Corporation[1] has compiled additional data for many oil and gas wells in its Well History Control System.

Major sources of hydraulic head and chemical data other than the U.S. Geological Survey's WATSTORE file are the files of the Texas Department of Water Resources and brine data from the Petroleum Data System at the University of Oklahoma.

Estimates of pumpage at state level have been made throughout the study area by the U.S. Geological Survey in cooperation with State and other Federal agencies at 5-year intervals since about 1950 (Solley and others, 1983). Additional detail is needed for use of pumpage in ground-water flow models; therefore, the pumpage data at the county level will be compiled and used to estimate the amount of pumpage for each grid block of the flow models.

Pumpage data for 1980 are summarized by state and aquifer systems in table 3 and show that more than one-half of the pumpage is from the Mississippi River Valley alluvial aquifer. Pumpage from aquifers in Arkansas is about twice that of any other state within the study area.

[1] The use of company names or brand names in this report are for identification purposes only and do not imply endorsement by the U.S. Geological Survey.

Geophysical logs were selected for use in defining the regional geohydrologic framework. Location and general information about the wells and conditions at the time of logging were entered into a computer-based file along with the thickness, resistivity, spontaneous potential, and depth to each sand bed with a thickness exceeding 20 feet. Geologic units were correlated across the study area and the depth to the top of each unit was entered into the computer file. Of 1,011 selected logs, 887 have been interpreted and entered into the computer file. An average of 22 sand beds per well have been identified for almost 7 million feet of logged section. The vertical relation of the aquifers and confining units as defined from the geophysical logs is shown in figure 93.

A computer-based file of selected chemical analysis of ground water has also been constructed. About half of the nearly 52,000 analyses have data on well depths (or depth of sample collection), and about 75 percent of the analyses have data for dissolved solids.

SIMULATION OF GROUND-WATER FLOW

Both regional and subregional finite-difference ground-water flow models are being used for simulation of the ground-water flow within the Gulf Coastal Plain regional aquifer system. The regional model will encompass the entire study area including those parts of the aquifers that contain saline water. The subregional models will be limited to smaller areas as shown in figure 94, and to those parts of the aquifers containing fresh to moderately saline water.

The regional model consists of a rectangular grid that is divided into blocks which are 10 miles on each side. The model probably will consist of 11 aquifers; however, the details of how each aquifer will be delineated has not been finalized. Flow of ground water in all of the aquifers, with the exception of the Upper Cretaceous aquifer, will be simulated using a variable-density flow model developed by Kuiper (1983). Spatial variation in water density and water temperature is calculated from data obtained from geophysical logs.

The subregional models consist of rectangular grids with blocks that are 5 miles on each side. Four of these blocks cover the same area as a single block in the regional model. The number of aquifers in the subregional models probably will range from two for the Mississippi River Valley alluvial aquifer, a subregional flow system of the Mississippi Embayment aquifer system, to six for the entire Mississippi Embayment aquifer system.

Preliminary model runs have been made using aquifer thicknesses calculated from data obtained from geophysical logs and with uniform values of hydraulic conductivity assigned to aquifers and confining units. Initial results have demonstrated that: (1) there is at least a small volume of water that moves vertically upward from the Mississippi Embayment aquifer system through the Coastal Lowlands confining unit (marine clays of the undifferentiated Jackson and Vicksburgs Group) and discharges to areas at or near land surface, and (2) the vertical hydraulic conductivity of the thick sequence of interbedded sand and clay within the Coastal Lowlands aquifer system is almost as small as the vertical hydraulic conductivity of the thick clay units that are areally persistent over part of the study area and mapped as the regional confining units.

Table 3.—Ground-water pumpage, in million gallons per day, 1980, Gulf Coastal Plain regional aquifer system

State	Mississippi Embayment aquifer system			Coastal Lowlands aquifer system	Texas Coastal Uplands aquifer system	Total
	Mississippi River Valley alluvial aquifer	Aquifers in sediments of Tertiary age	Upper Cretaceous aquifer			
Arkansas	3,745	199	6	—	—	3,950
Louisiana	234	82	—	1,148	—	1,754
Mississippi	1,140	154	—	146	—	1,440
Missouri	102	5	4	—	—	111
Tennessee	2	251	7	—	—	260
Texas	—	—	—	1,269	458	1,727
Total	5,223	691	17	2,853	458	9,242

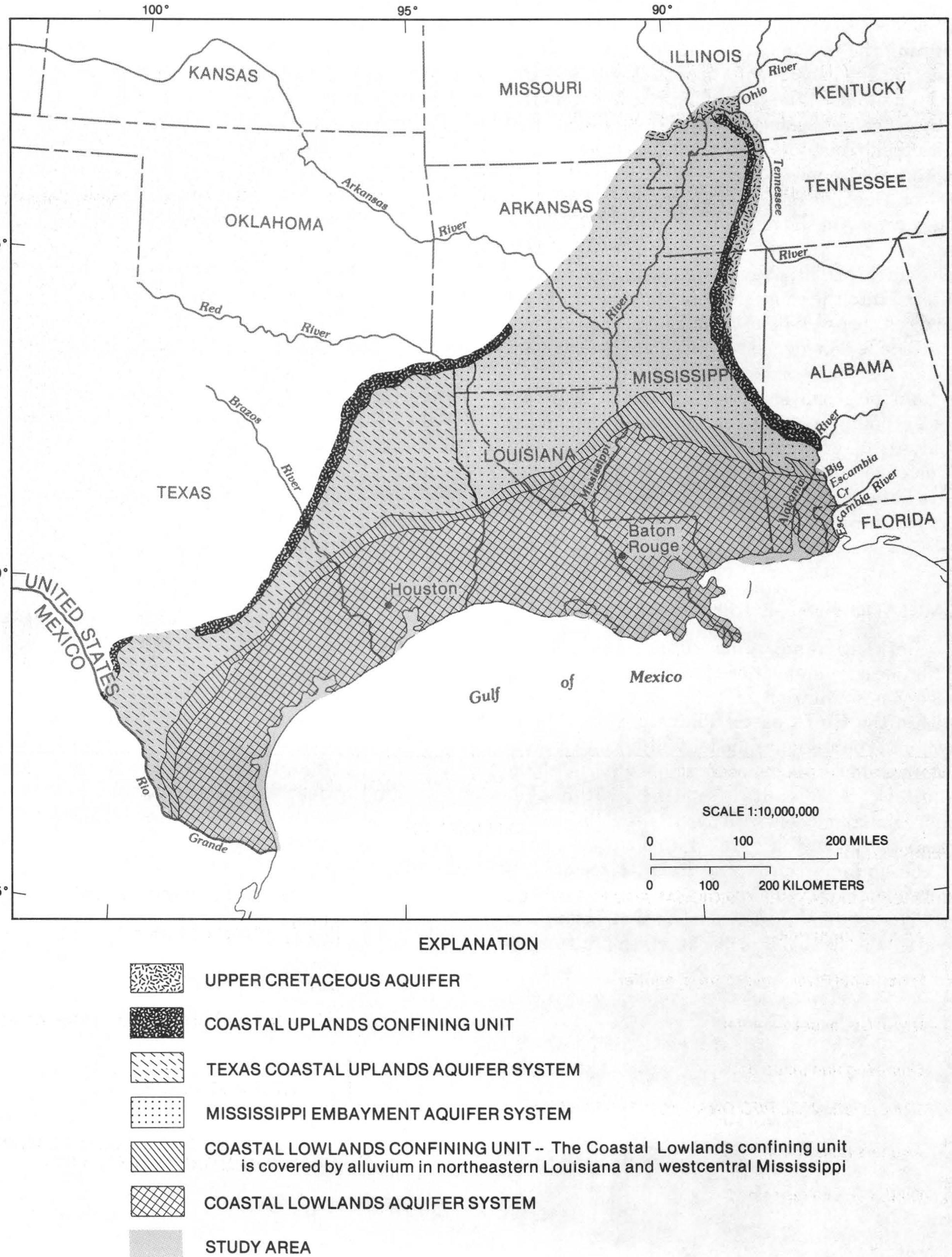
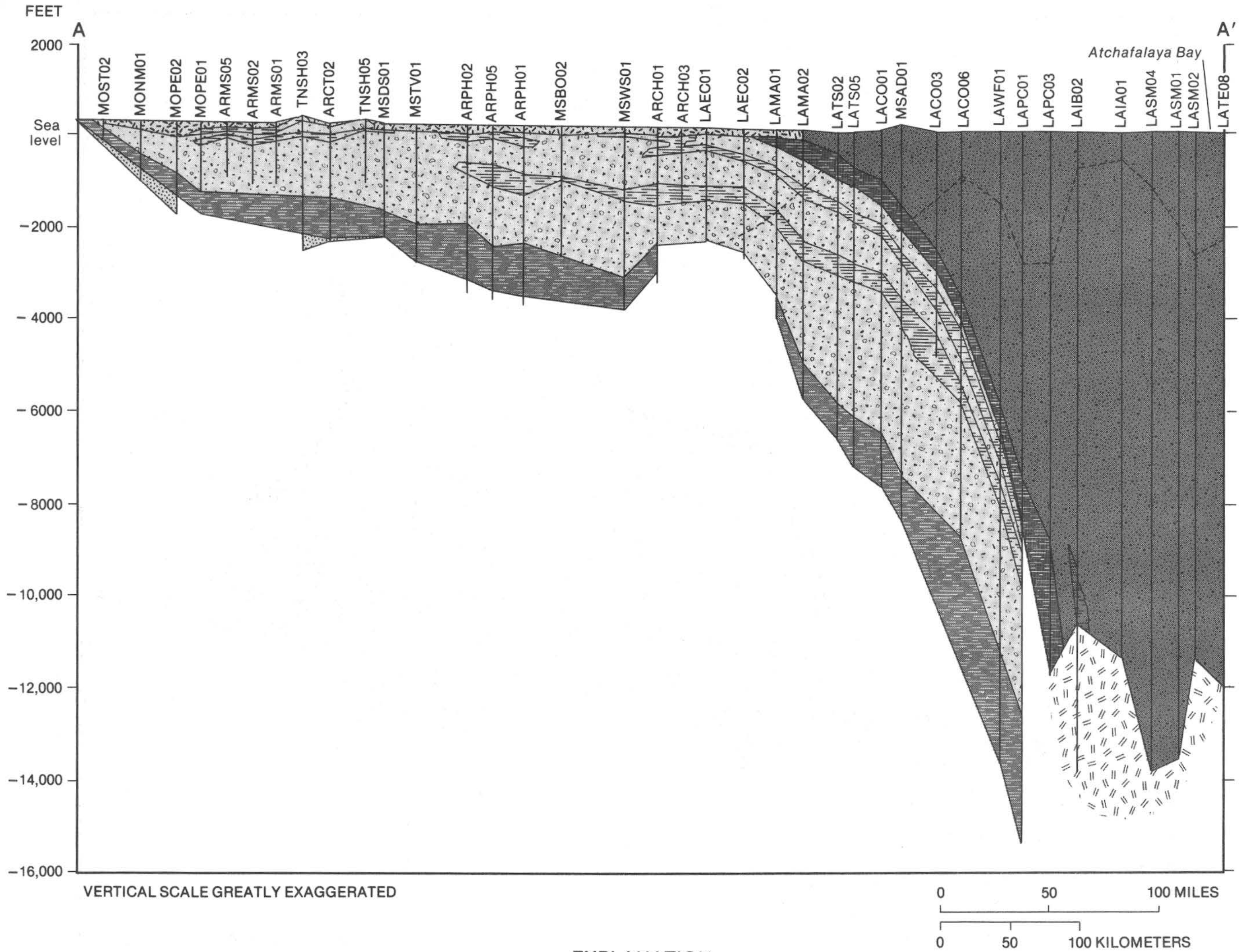


Figure 92.—Generalized outcrop of major aquifers and confining units in the Gulf Coastal Plain regional aquifer system study area. The Coastal Lowlands confining unit is covered by alluvium in northeastern Louisiana and west-central Mississippi.



EXPLANATION

AQUIFER SYSTEMS

MISSISSIPPI EMBAYMENT REGIONAL AQUIFER SYSTEM

- Aquifers (unnamed)
- Mississippi River Valley alluvial aquifer
- Upper Cretaceous aquifer
- Confining unit (unnamed)

COASTAL LOWLANDS REGIONAL AQUIFER SYSTEM

- Aquifers (unnamed)
- Confining unit (unnamed)

REGIONAL CONFINING UNIT

- Coastal uplands
- Coastal lowlands

ZONE OF ABNORMALLY HIGH PRESSURE

ARMS02 — WELL AND NUMBER

------- DISSOLVED SOLIDS EQUAL TO 10,000 MILLIGRAMS PER LITER

Figure 93.—Generalized section showing relation of aquifers and confining units from southeastern Missouri to Atchafalaya Bay, Louisiana, (location of section shown in fig. 91).

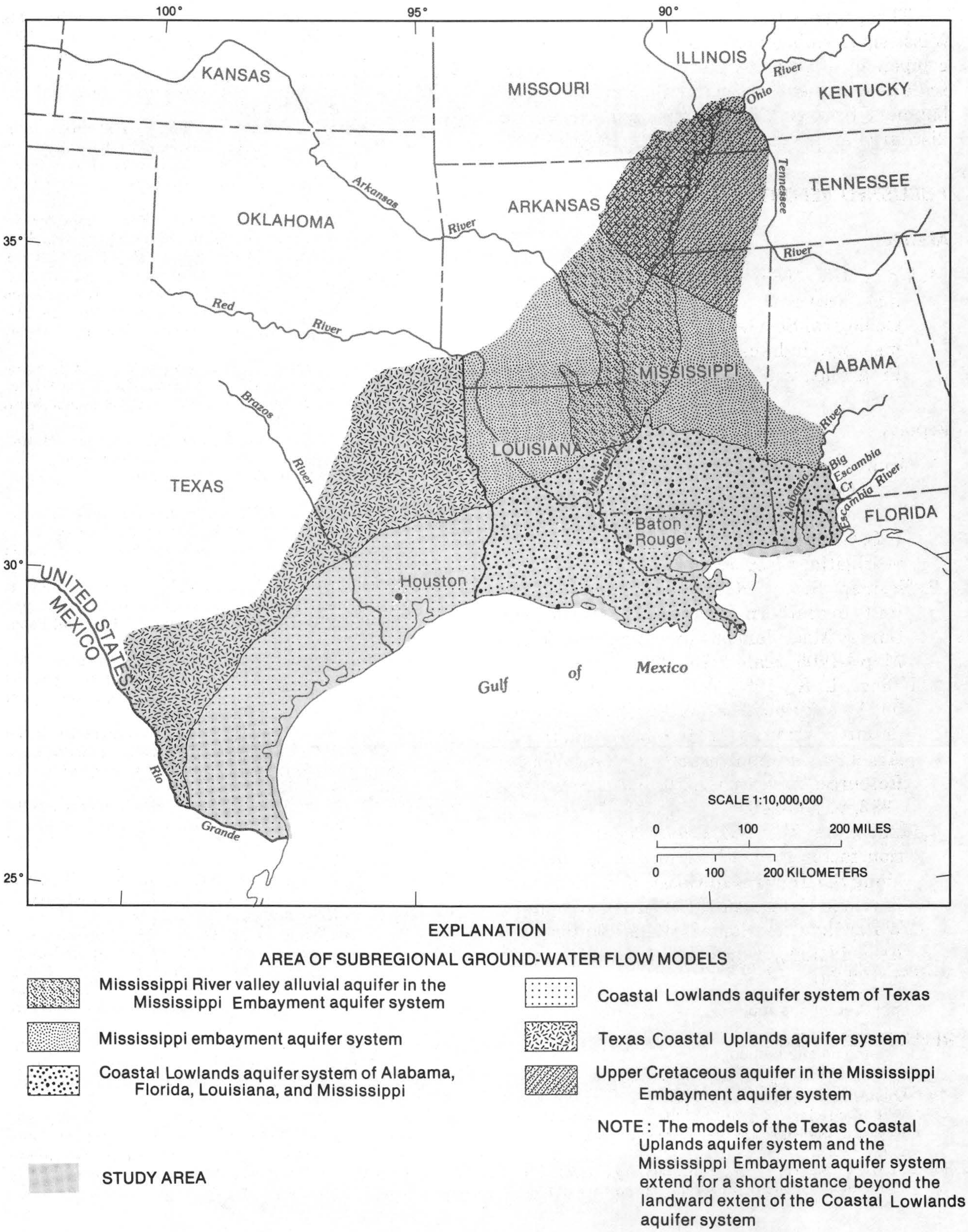


Figure 94.—Areas of subregional ground-water flow models, Gulf Coastal Plain regional aquifer system.

Three cross-sectional models across the Mississippi embayment indicate that a major component of ground-water flow prior to large-scale development was across the axis of the embayment from east of the Mississippi River to discharge areas west of the river (Weiss, 1983).

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NORTHEAST GLACIAL REGIONAL AQUIFER-SYSTEM STUDY

By Forest P. Lyford, Albany, New York

INTRODUCTION

The regional assessment of the Northeast glacial aquifers was started in 1981 and is scheduled for completion in 1986 (Lyford and others, 1984). The purpose of the study is to investigate the sand and gravel aquifers that were formed during advances and retreats of the continental glaciers in the northeastern United States. Glacial sand and gravel aquifers, though consisting of many geographically separated independent systems (fig. 95), share many geologic, hydrologic, and geochemical characteristics because of their common depositional origins and physiographic settings.

This study will document the hydrologic characteristics of the glacial aquifers in the northeastern United States through study of the variations in magnitude and areal distribution of key components of the aquifers and through evaluation of the response of the aquifers to pumping and to climatic stresses.

The study area includes most of the glaciated parts of the northeastern United States and extends approximately as far west as the edge of the glaciated Appalachian Plateau in Ohio. The areas of Long Island, NY, and Cape Cod, MA, are excluded from the study because the ground-water hydrology of these systems has been extensively studied. The study area includes several physiographic provinces, which range from mountainous areas such as the White Mountains of New Hampshire and Maine, the Green Mountains of Vermont, and the Adirondack and Catskill Mountains of New York; to low-lying areas along the Great Lakes, the St. Lawrence River valley, the Hudson and Mohawk River valleys; and seaboard lowland areas along the Atlantic Coast (fig. 96).

PROGRESS AND FINDINGS

Aquifer Geometry

The study area has been subdivided into three major geohydrologic areas to group and describe aquifer systems that have similar characteristics. These subdivisions, shown in figure 97, are based largely on the geology of glacial deposits and physiography.

Much of the study area and most of the productive glacial aquifers are in type A areas (fig. 97), where aquifers were formed largely in valleys that generally drained away from glacial ice. In type A areas, stratified drift commonly fills valleys or narrow lowlands, bordered by uplands or ridges. The stratified drift generally exhibits a succession of overlapping, shingled depositional profiles classified as morphosequences (Koteff, 1974). Each morphosequence ideally grades down valley from coarse-grained heterogeneous ice-contact deposits to a valley train capped by pebbly coarse sand that may overlie lacustrine deposits. Generalized transverse and longitudinal sections of valleys draining away from glacial ice are shown in figure 98.

Glacial aquifers in type B areas (fig. 97) were formed largely in lake- or marine-dominated terranes where fine-grained stratified drift predominates. Commonly, a surfical aquifer, consisting of medium to very fine sand overlies silt and clay as shown in figure 99. Ice-contact deposits, which are commonly scattered beneath the lacustrine clay, probably reflect local ice-front positions and meltwater channels in the decaying ice. In southern Maine, northwestern Vermont, and south of Lake Ontario, marine or lake-bottom silt and clay immediately underlie the land sur-

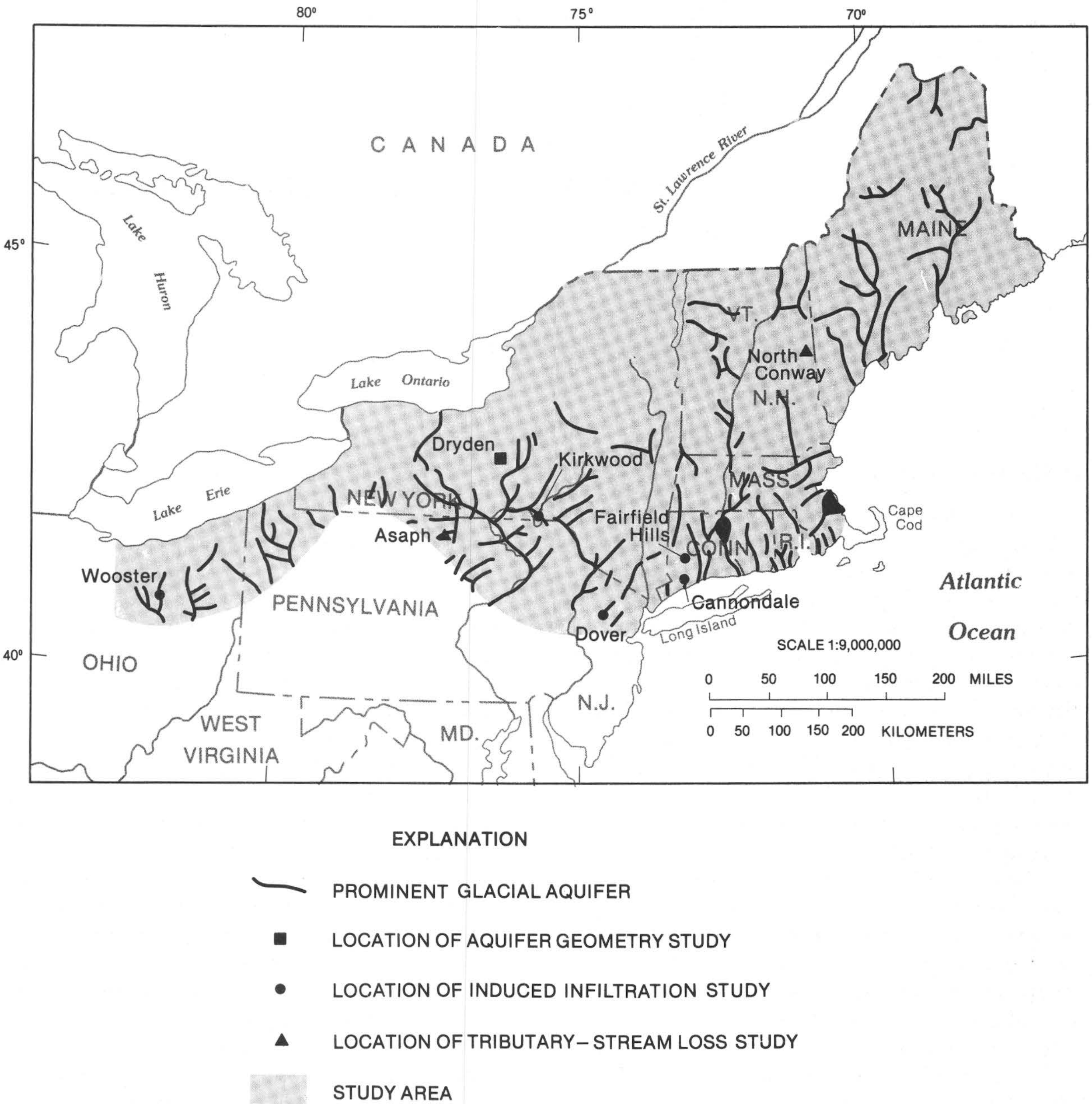

Figure 95.—Study areas, specific site study area, and prominent glacial aquifers in the northeastern United States (modified from McGuinness, 1964).

face over large areas. Coarse-grained stratified drift has limited distribution in type B areas.

Glacial aquifers in type C areas (fig. 97) are characterized by valleys that drained toward glacial ice. Glacial drift is commonly several hundred feet thick in type C areas and consists largely of lake-bottom silt and clay that are interlayered in places with till composed of reworked lacustrine sediments. Coarse sediments, a minor fraction of the valley fill, may have originated in two ways: (1) as alluvium or possibly deltaic deposits of north-draining streams deposited during intervals between ice readvances and, (2) as local outwash deposited by meltwater at the base of an ice tongue during advance or retreat of the ice. The origin of buried coarse-grained sediments in a type C area is being studied near Dryden in central New York (fig. 95).

In support of aquifer geometry studies and to determine the application of surface geophysics, several methods, including electromagnetic surveys and marine seismic reflection, are being tested in several depositional environments to aid

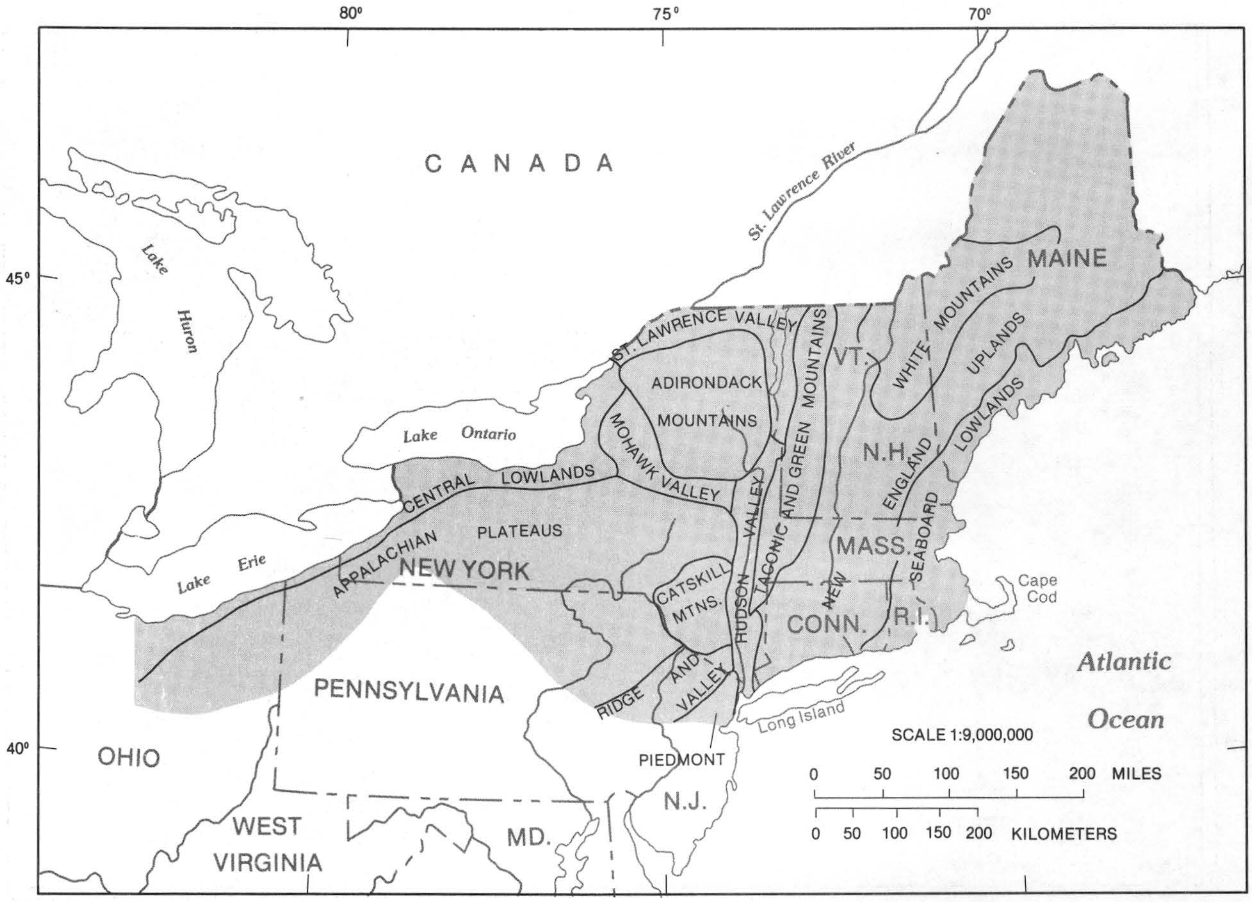

Figure 96.—Physiographic divisions of the northeastern United States in the study area (shaded) of the Northeast glacial regional aquifer system (from Fenneman, 1938).

in determining the subsurface distribution of glacial materials. Preliminary results indicate that these methods are useful for interpolation of the subsurface geology between points where geologic data are available, and for mapping the distribution of buried coarse-grained material.

Studies of low-flow characteristics of streams will help identify the elements of aquifer geometry that strongly influence the hydrology of glacial aquifers as well as interaction between aquifers and streams. Earlier studies in Connecticut and New York indicate that low flow is chiefly a function of the size of the area covered by coarse-grained stratified drift. Low-flow studies, as part of this study conducted in Connecticut, have shown that the percent of area covered by wetlands is also a significant factor affecting low flow (Robert Melvin, U.S. Geological Survey, written commun., 1983). Studies of geologic and physiographic controls on low flow in streams are underway in other parts of the study area.

Regional Analysis of Water Chemistry

The water-quality data in the U.S. Geological Survey's WATSTORE file are being analyzed to define water chemistry in glacial aquifers and to determine relations between rock minerals and water chemistry. Statistical analysis of concentrations of dissolved solids from water samples taken from stratified drift and shallow bedrock wells (generally less than 200 feet deep) indicates that the chemistry of water in drift is similar to that of the adjoining bedrock. Studies of the areal distribution of various constituents and of the geologic and hydrologic factors that affect water chemistry are in process.

Hydrologic Budget of Glacial Aquifers

Streamflow and water-level data are being analyzed to determine the hydrologic budget of selected glacial aquifers. Recharge and discharge of an entire basin have been studied by many investigators; however, information on recharge

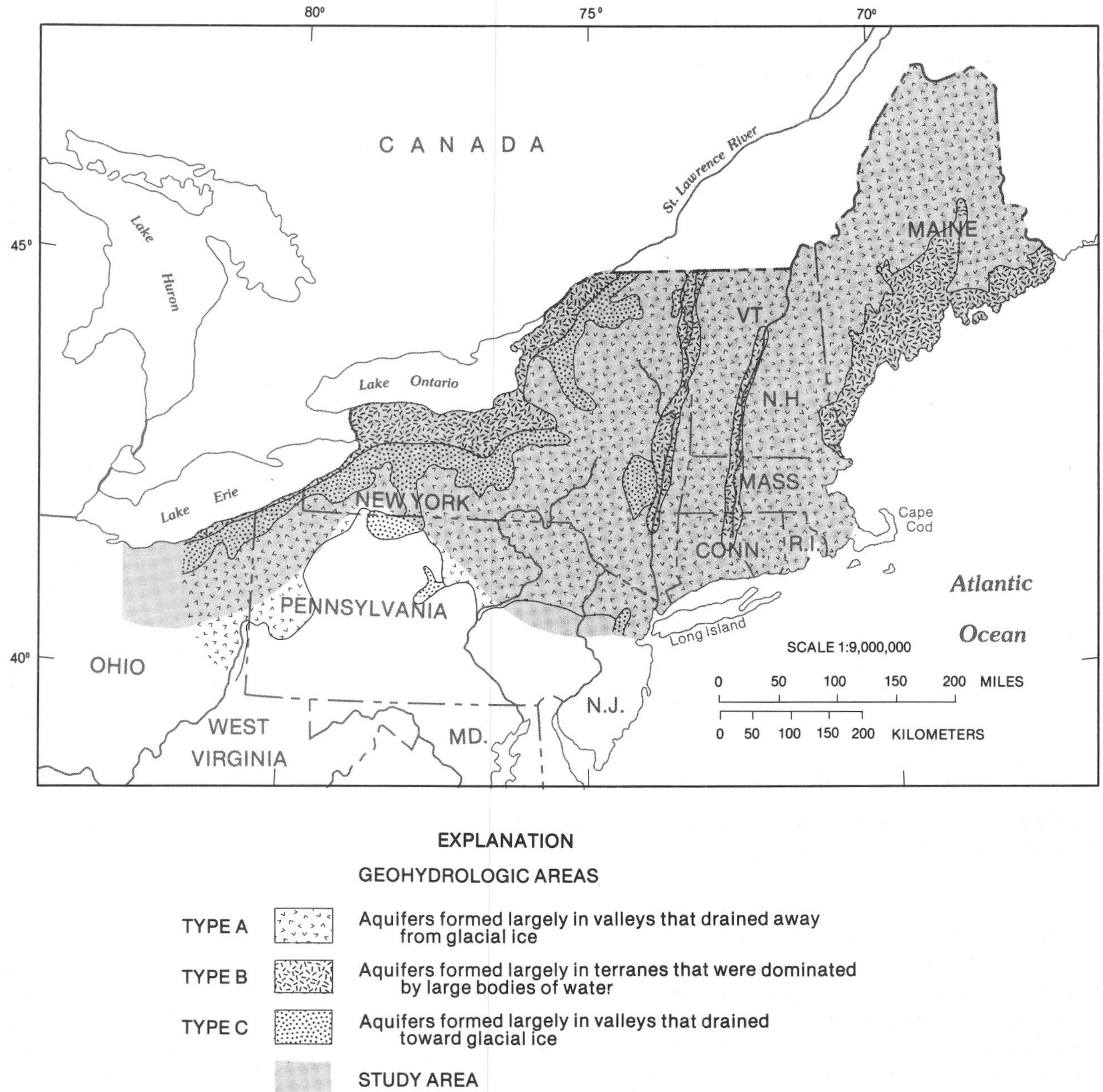

Figure 97.—Three major types of geohydrologic areas to group and describe the glacial aquifers in the northeastern United States.

and discharge of sand and gravel aquifers within a basin is lacking. Analytical and modeling techniques are being used to evaluate ground-water runoff in aquifers of different geometry.

Recharge to glacial aquifers from tributary streams that flow across the valleys is an important component of the hydrologic budget for certain valley-fill aquifers. Field studies to quantify the recharge process from tributary streams are underway in the Asaph area of Pennsylvania and the North Conway area of New Hampshire (fig. 95).

Hydraulic Properties of Aquifers Near Streams

A well field near a stream in a glacial aquifer may strongly affect the flow in the stream. The controlling factors are the hydraulic conductivities of streambed and aquifer materials, and the head difference between the stage in the stream and head in the aquifer near the stream. Five sites have been selected for studies of these controlling factors. These sites are: (1) Wooster well field on Killbuck Creek, OH; (2) Dover well field on the Rockaway River, NJ; (3) Cannondale well

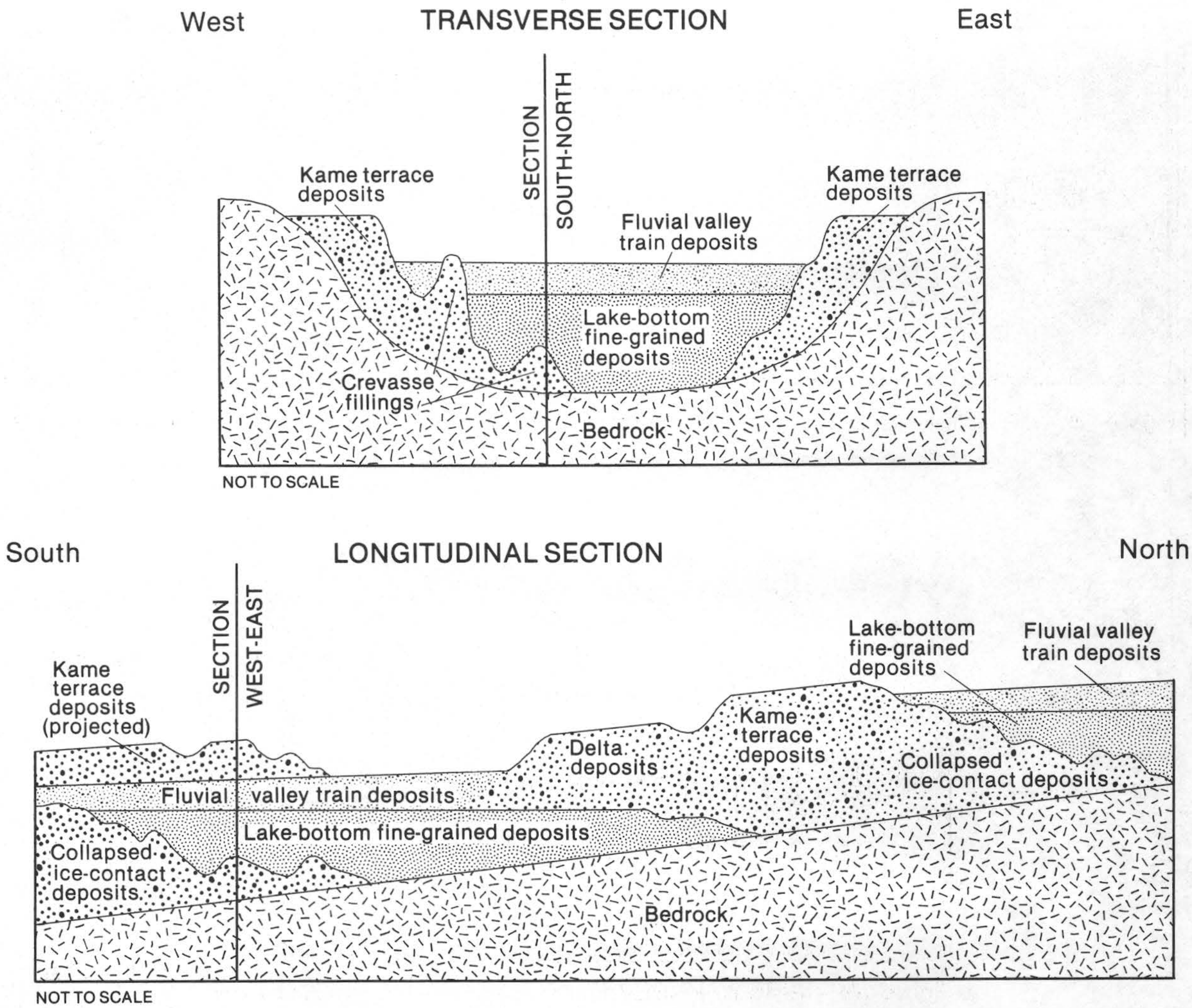

Figure 98.—Generalized transverse and longitudinal geologic sections of valleys draining away from glacial ice (type A).

field on the Norwalk River, CT; (4) Fairfield Hills well field on the Pootatuck River, CT; and (5) Kirkwood well field on the Susquehanna River, NY (fig. 95). Techniques that are used to determine hydraulic properties of aquifer materials near streams and of streambeds include: (1) simultaneous measurement of seepage loss or gain and head in aquifers near the stream, (2) chemical analyses of water from the stream and nearby wells to estimate quantity of water diverted from the streams, (3) temperature surveys in wells to determine the ground-water flow rate, and (4) flow modeling.

Simulation of Glacial Aquifers

Flow models are being used to evaluate and synthesize concepts of ground-water flow in glacial aquifers and to analyze sensitivity of flow to geohydrologic factors. A three-dimensional conceptual model consisting of three layers representing an idealized valley-aquifer system is depicted in figure 100. In the valley part, layers 1 and 2 represent coarse-grained material separated by a confining layer. In the upland area, layer 1 represents a layer of soil or weathered till at land surface; layer 2 represents a layer of till or fractured bedrock; and layer 3 represents bedrock throughout the modeled area. These features of a glacial aquifer are commonly observed in the northeastern United States. The flow model is being used to evaluate variations in ground-water runoff with time, the relation between aquifer head and ground-water runoff, and the response of streamflow to pumping stresses.

Bibliography of Previous Investigations

A bibliography that summarizes published information on ground-water studies related to

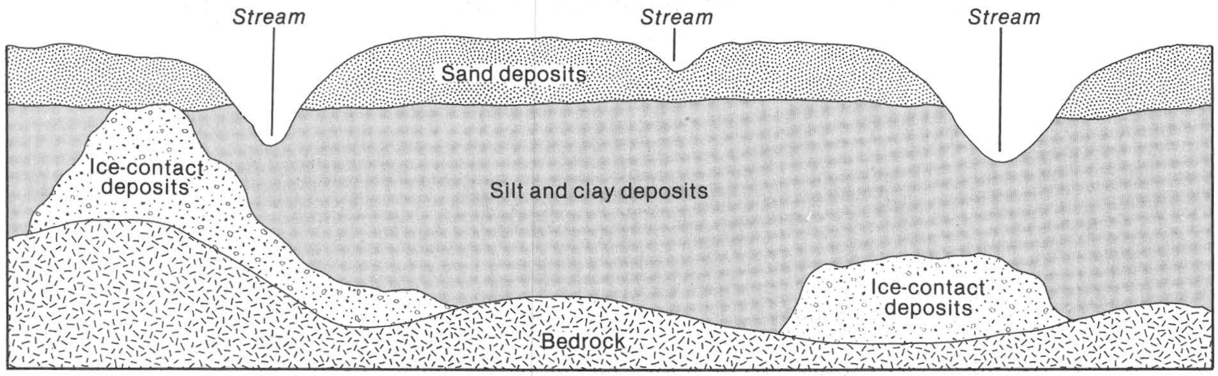

Figure 99.—Generalized geologic section of deposits in an area having extensive proglacial lakes (type B).

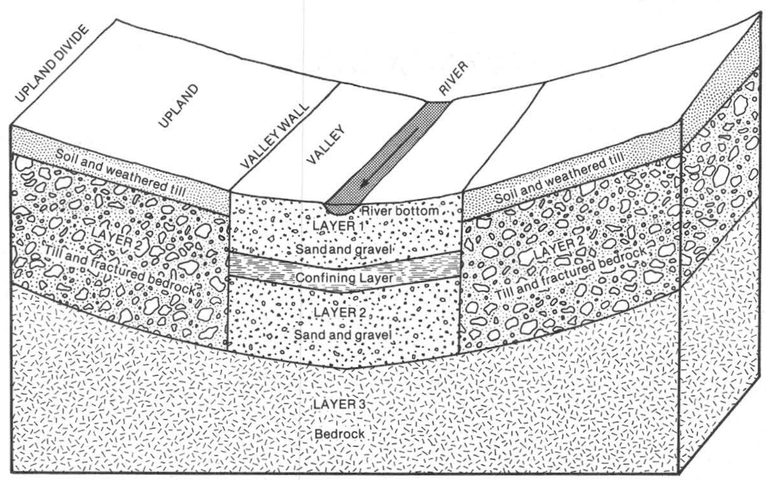

Figure 100.—Generalized valley-aquifer system used in a ground-water flow model for the study of the Northeast glacial regional aquifer system.

glacial aquifers in the study area includes about 750 references. Types of information contained in each reference such as well data, modeling results, and description of water quality have been identified.

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NORTHERN ATLANTIC COASTAL PLAIN REGIONAL AQUIFER-SYSTEM STUDY

By **Harold Meisler, Trenton, New Jersey**

INTRODUCTION

The northern Atlantic Coastal Plain is a gently rolling to flat region of about 50,000 mi^2. The study area extends along the Atlantic coast from Long Island, NY, to North Carolina (fig. 101). It is underlain by a wedge of predominantly unconsolidated sediments that thickens from a feather edge at the Fall Line to 8,000 feet along the coast of Maryland and 10,000 feet at Cape Hatteras, NC, (fig. 102). The sediments consist mostly of sand, silt, clay, and gravel of Jurassic to Holocene age. Limestone occurs in North Carolina. A regional aquifer system study of the Northern Atlantic Coastal Plain was begun in 1979 and is scheduled for completion in 1986 (Meisler, 1980).

This sedimentary wedge forms a complex aquifer system in which the sand, gravel, and limestone function as aquifers, whereas the clay and silt act as confining units. Withdrawal of water from this system, principally for municipal and industrial use, has grown from about 100 Mgal/d in 1900 to about 1,200 Mgal/d in 1980.

Recharge to the northern Atlantic Coastal Plain aquifer system is derived from precipitation and occurs chiefly in upland and interfluvial areas. It ranges from 10 to 25 in./yr, but most of this water flows only through the shallow unconfined parts of the system and discharges to local streams that dissect the Coastal Plain. A small amount of the precipitation, generally less than 1 in/yr, recharges the deeper confined aquifers. Under natural conditions, discharge from the deeper aquifers is primarily upward across the confining units into shallower aquifers and ultimately into the sea or coastal estuaries, sounds, and bays.

HYDROGEOLOGIC FRAMEWORK

By **Henry Trapp, Jr., Trenton, New Jersey**

Northern Atlantic Coastal Plain sediments range in age from Jurassic to Holocene. Late Jurassic sediments have been identified in a few wells near the coast, but for the most part, the Jurassic sediments are offshore (fig. 103), and do not contain freshwater. In general, the lowermost Coastal Plain deposits are fluvial or fluviodeltaic in origin, and contain discontinuous lenses of sand, silt, clay, gravel, and lignite. Younger deposits successively overlap older ones (figs. 103-105). In the Cretaceous section, there is a general upward transition from fluvial and fluviodeltaic to marginal-marine to marine deposits. The marine parts of the section consist primarily of glauconitic sand, silt, clay, and some limestone beds, which are traceable over longer distances than the more lenticular nonmarine beds. The Tertiary sediments are predominantly marine except for the Upper Miocene and Pliocene beds, which are, in part, nonmarine. The Pleistocene section includes glacial drift on Long Island, and marine, dune, alluvial and terrace deposits elsewhere. The Holocene section includes alluvial, terrace, and marine deposits, beaches, and dunes.

For this study, the Coastal Plain sediments have been grouped into ten regional aquifers (fig. 106), consisting principally of sand, gravel, or limestone, separated by nine confining units, consisting principally of clay and silt. A regional aquifer may coincide with a recognized local aquifer in one area and comprise several local aquifers in another, or it may constitute only part of an aquifer. None of the regional aquifers extends over the entire study area. Because they are delineated on the basis of permeability contrast, and the distribution of permeable zones changes from place to place, the regional aquifers represent sediments of different ages in different areas.

SIMULATION OF GROUND-WATER FLOW

By **P. Patrick Leahy and Mary Martin, Trenton, New Jersey**

Ground-water flow in the northern Atlantic Coastal Plain regional aquifer system was simulated using a multilayered finite-difference flow model (Leahy, 1982) that includes ten aquifers and nine confining units. The grid used has 85 rows and 32 columns. The grid sizes are variable; however, most of them are 7 miles × 7 miles or 49 mi².

The aquifer system is bounded laterally and below by no-flow boundaries. The bottom boundary corresponds to the sloping contact between Coastal Plain sediments and the underlying, nearly impermeable, crystalline rocks or consolidated rocks of Mesozoic age. The Fall Line represents an irregular no-flow boundary at the updip limit of the aquifer system to the north and west. The downdip offshore no-flow boundary to the east and south represents an assumed saltwater-freshwater interface located where the ground water contains concentrations of chloride of 10,000 milligrams per liter (mg/L) as delineated by Meisler (1981). The southwestern boundary is coincident with a major ground-water divide located at the Pee Dee River in South Carolina. However, in the deepest part of the aquifer system, ground water appears to flow northeast in South Carolina toward North Carolina beneath the shallower ground-water divide. Therefore, the southwestern boundary of the lowest aquifer was defined as a constant-head boundary instead of a no-flow boundary. The northeastern no-flow boundary of the aquifer system occurs at the intersection of the Fall Line and an estimated location where ground water contains concentrations of chloride of 10,000 mg/L northeast of Long Island.

During steady-state simulations, two top boundary conditions were used to represent the shallow aquifers. One of the top boundaries has water-table altitudes specified as constant heads (fig. 107). In this case, the water-table aquifer was not simulated; however, the flow recharged from the water-table aquifer to the deeper aquifers was simulated. The other top boundary was used to simulate the water-table aquifer and streams as part of the aquifer system (fig. 108). In this case, the base flow discharged from the water-table aquifer and recharge rate to the deeper aquifer system were simulated.

Simulation of the steady-state flow system using water-table altitudes as a constant-head boundary yields heads in every cell and a value of flow into or out of the constant-head cells. The flow out of the constant-head cells represents the component of flow recharging to the deeper aquifers and is herein referred to as deep percolation. Deep percolation is treated as a known flow for the second conceptualization of the top boundary condition.

The second conceptualization includes the water-table aquifer and streams as part of the aquifer system and is necessary in order to simulate transient water levels in aquifer-outcrop areas where pumping centers are located and water levels are declining. These water-level declines cannot be simulated using the water-table constant-head boundary.

Simulation of the flow system using the second conceptualization includes ground-water recharge of 10 to 25 in./yr from precipitation to the water-table aquifer, and streams acting as local drains. The streambed leakance (hydraulic conductivity divided by streambed thickness) controls the amount of ground water discharged to or recharged from streams as base flow or seepage depending on head difference between the aquifer and the stage in the stream. A water budget for each model cell is used to compute an effective streambed leakance. Inasmuch as recharge (QRE) is given and deep percolation (DP) is computed by the steady-state simulation, the discharge (BF) to a stream is then calculated. Using the calculated discharge and the known altitudes for the water table (h_{wt}) and stage of streams (h_s), streambed leakances (TK_s) can be calculated. Because of the large grid size used in the model, numerous streams exist in every cell and one average representative stream elevation must be estimated for each model cell. Also, regional lateral flow in the water-table aquifer is assumed to be negligible.

Simulating ground-water flow under prepumping conditions results in virtually equivalent values of head and of deep percolation using either a constant-head boundary to represent the water-table aquifer, or simulating the water-table aquifer and streams as part of the aquifer system. However, using the latter ap-

Figure 101.—Location of the northern Atlantic Coastal Plain.

proach, the water-table altitude, deep percolation, and discharge to or recharge from streams are able to change in response to pumping stresses during transient simulations.

Figures 109 and 110 show the simulated prepumping potentiometric surfaces for two selected aquifers, a deep aquifer consisting of sediments of Late Cretaceous age (model aquifer layer 2) and a shallow aquifer consisting of sediments of Eocene age (model aquifer layer 7). From the potentiometric surface of the shallower aquifer, the discharge to major surface-water bodies may be deduced. The influence of major rivers, estuaries, and embayments on the flow system is apparent. In contrast, the potentiometric surface of the deeper aquifer shows smoother contours reflecting a deeper regional ground-water flow. Although the streams and rivers seem to affect the flow pattern in the updip part of the deeper aquifer, most of the ground-water flow is not influenced by the presence of overlying surface-water bodies. Flow lines in the deeper aquifer are several tens of miles long and cross State boundaries. The flat hydraulic gradient suggests that the ground-water flow is sluggish in areas away from outcrops.

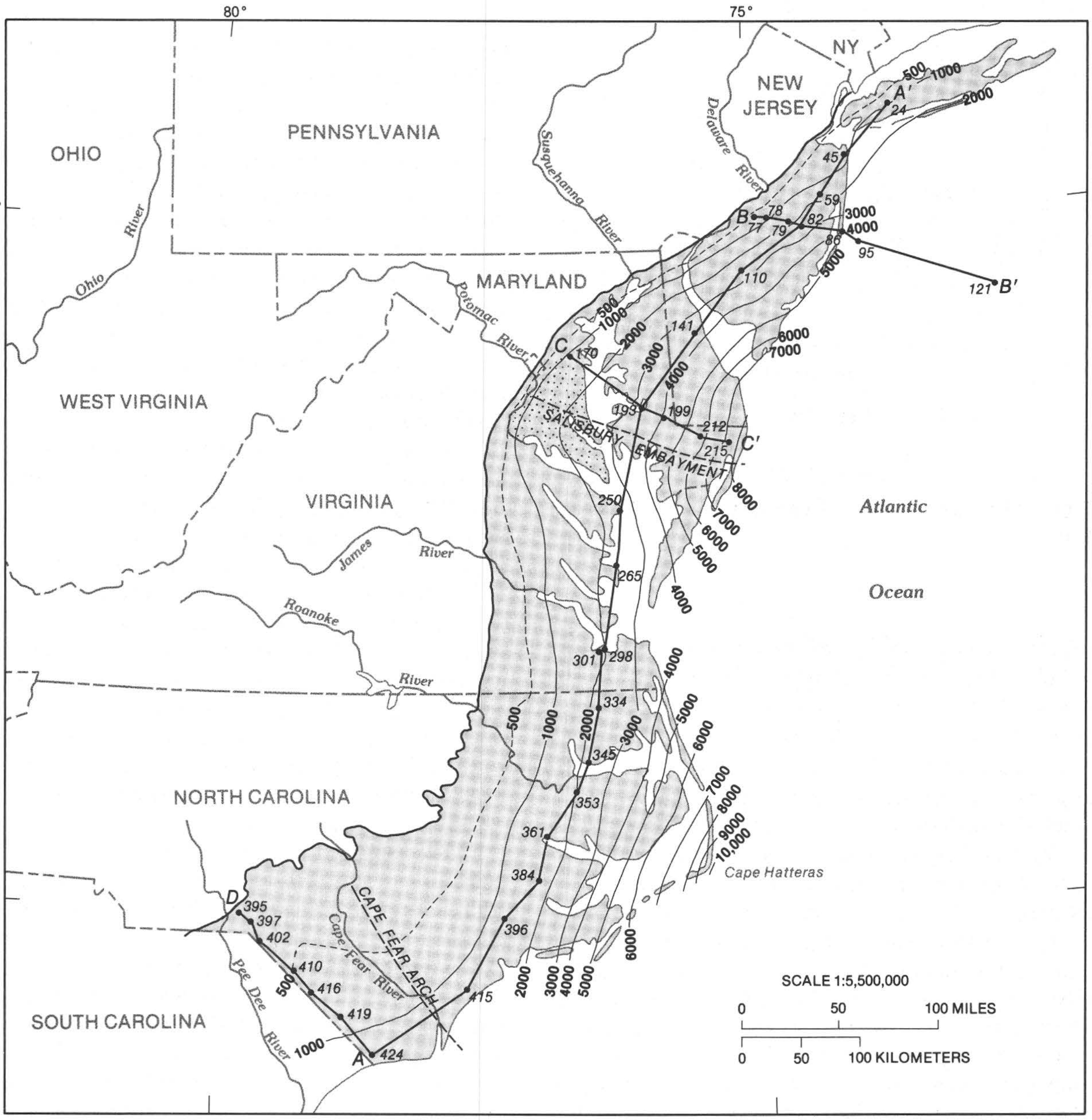
The simulated prepumping potentiometric surfaces show that: (1) recharge to the confined aquifers generally is downward along or near the Fall Line, and (2) discharge from the deeper confined aquifers usually is upward leakage through confining units into the ocean or coastal estuaries and bays.

Figure 111 shows the simulated flow to and out of the overlying water-table aquifer under steady-state conditions. The flow ranges up to 20 in./yr of recharge into the confined aquifers and up to 17 in./yr of discharge from the confined aquifers. In general, however, recharge is 1 in./yr or less, indicating that only a small part of the water in the water-table aquifer infiltrates to the deeper confined aquifers. High infiltration rates are located in areas where the water-table aquifer is directly connected with the underlying aquifers and there is no intervening confining unit. Examples of this relation occur between the glacial aquifer and the Magothy aquifer in Long Island, and between the surficial and Castle Hayne aquifers in North Carolina.

Generally, development of ground water in the northern Atlantic Coastal Plain began in the late 1800's. Pumpage for 1900 was estimated to be 100 Mgal/d; most of it was for water supply of Brooklyn and Queens on Long Island. Withdrawals have steadily increased and total pumpage in 1980 was estimated to be about 1,200 Mgal/d (fig. 112). These estimates do not include domestic and irrigation usage obtained from shallow water-table aquifers, because the impact of these water-table pumpages is negligible regionally. Water-level declines resulting from deep-aquifer withdrawals are observed in every State in the study area.

Estimates of pumpage from 1900 to 1980 were used in determining the average pumpage for ten pumping periods for the simulation. Figures 113 and 114 show the simulated 1980 potentiometric surfaces for a deeper Late Cretaceous aquifer and a shallow Eocene aquifer, respectively. Pumping in the deeper aquifer has reduced water levels in Virginia to more than 100 feet below sea level and to more than 50 feet below sea level in New Jersey. Pumping in the shallower aquifer reduced water levels to more than 50 feet below sea level in North Carolina and Delaware.

Sources of pumped water in the Coastal Plain are: (1) an increase in deep percolation from the overlying water-table aquifer to deeper confined aquifers corresponding to a decrease in discharge to streams; (2) water withdrawn from aquifer storage; and (3) a reduction in upward discharge from the confined aquifers through overlying confining units to surface-water bodies, including the ocean. Preliminary simulation indicates that the aquifer system approaches steady-state conditions within a 5-year pumping period. Therefore, water is released from aquifer storage with an accompanying water-level decline for a relatively short period of time after increases in pumpage. Pumping has caused a long-term decrease in ground-water discharge to streams in areas of increased infiltration to the deeper aquifers. In other areas, pumping has resulted in a decrease in discharge from the deeper aquifers upward to the overlying aquifers and lower water levels in the deeper



EXPLANATION

—2000— LINE OF EQUAL THICKNESS OF COASTAL-PLAIN SEDIMENTS -- Interval in feet, is variable

A A' (424 24) TRACE OF HYDROGEOLOGIC SECTION -- Numbers indicate the control wells

---- AXIS OF STRUCTURAL FEATURE

AREA OF INTENSIVE SAMPLING OF AQUIA-AQUIFER WATER

FALL LINE

STUDY AREA

Figure 102.—Thickness of Coastal-Plain sediments and location of hydrogeologic sections in the northern Atlantic Coastal Plain.

aquifers, possibly causing migration of the freshwater-saltwater transition zone, toward pumping centers.

Simulation has indicated: (1) the need for rigorous definition of confining-unit leakance, (2) the importance of discretization scale in the simulation of regional flow as related to the purpose of the simulation, and (3) the importance of the boundary condition representing the freshwater-saltwater transition zone.

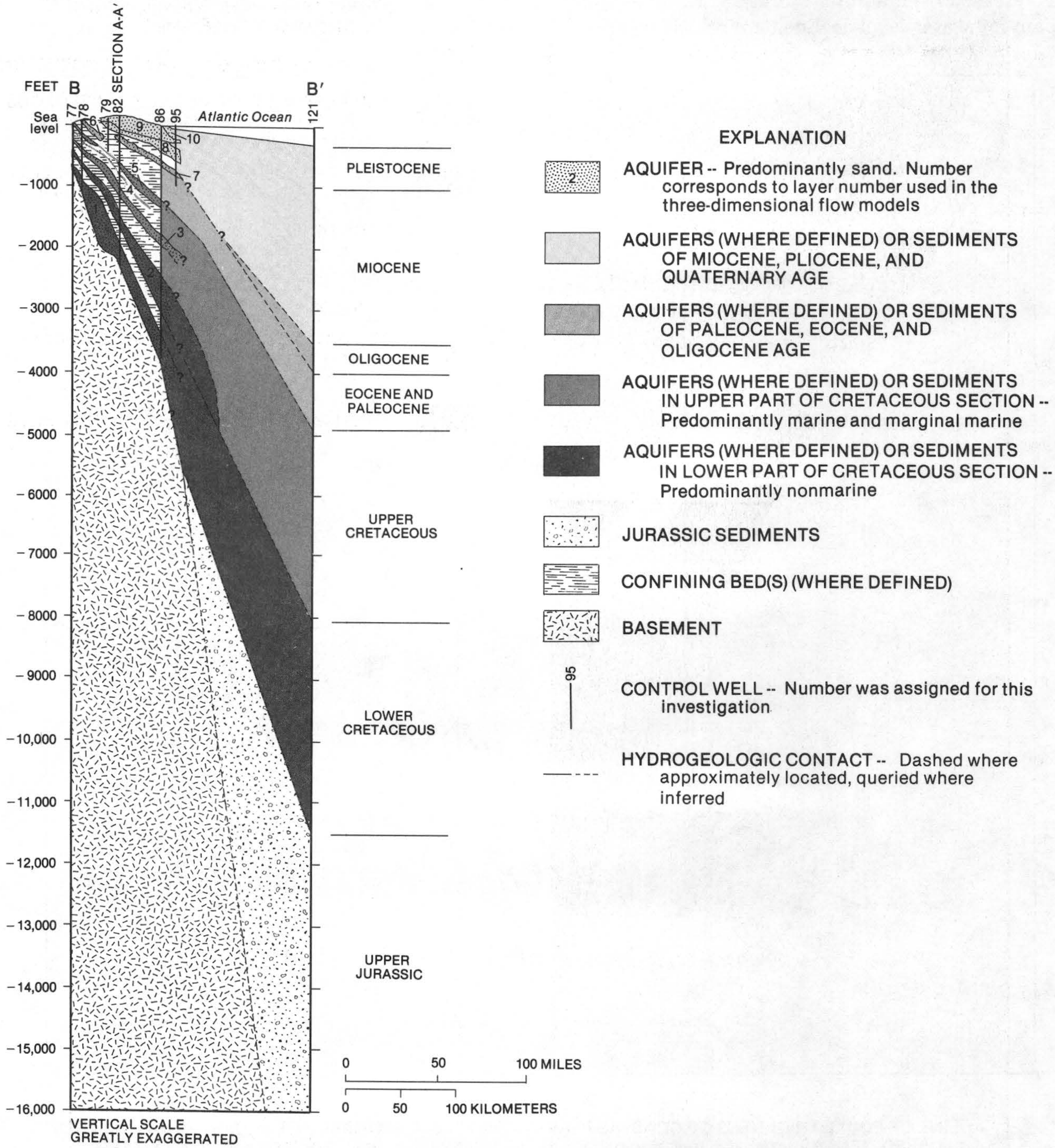


Figure 103.—Hydrogeologic section B-B′, from New Jersey to the offshore COST B-2 well, in the northern Atlantic Coastal Plain (location of section shown in fig. 102).

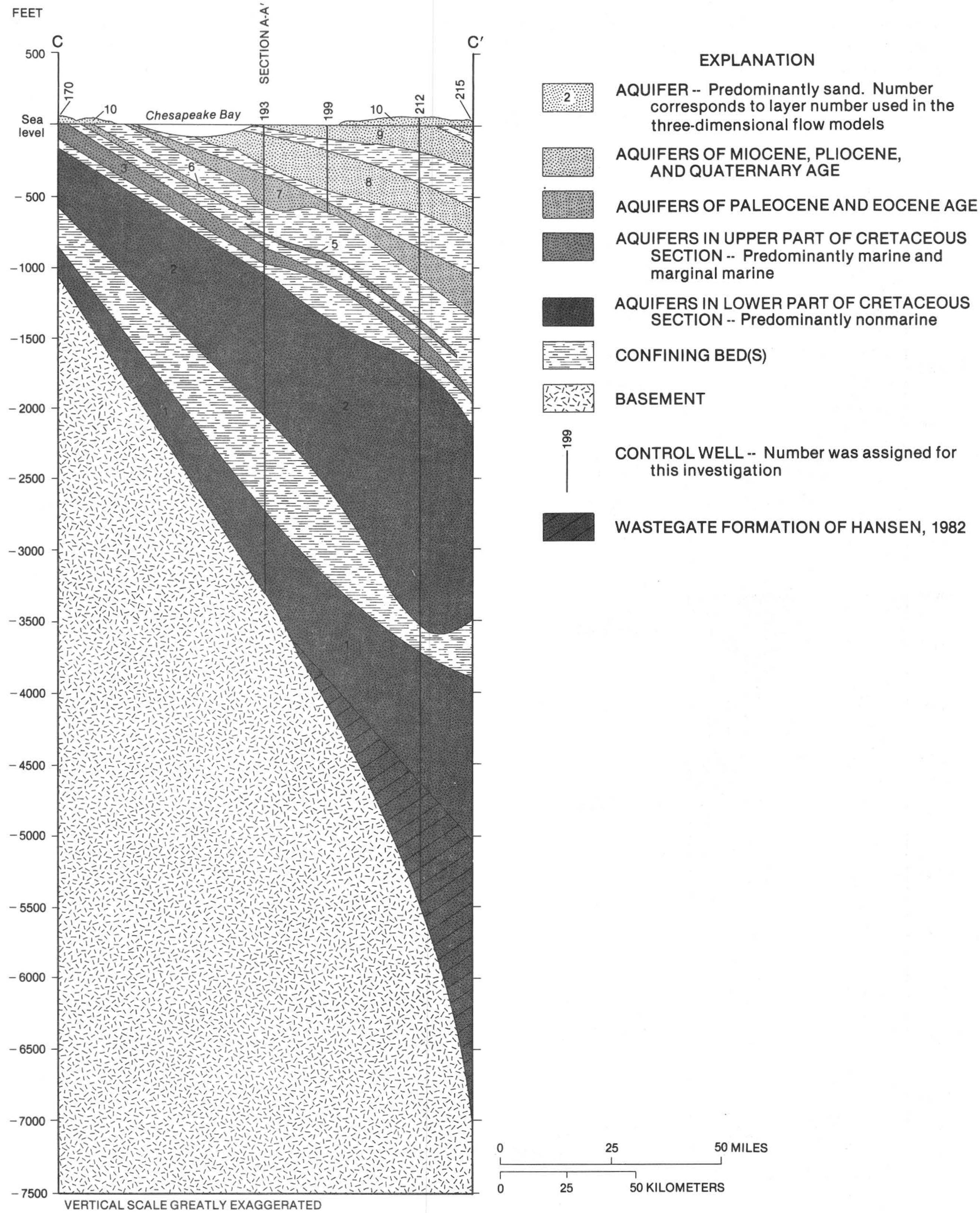


Figure 104.—Hydrogeologic section C-C′, Maryland, in the northern Atlantic Coastal Plain (location of section shown in fig. 102).

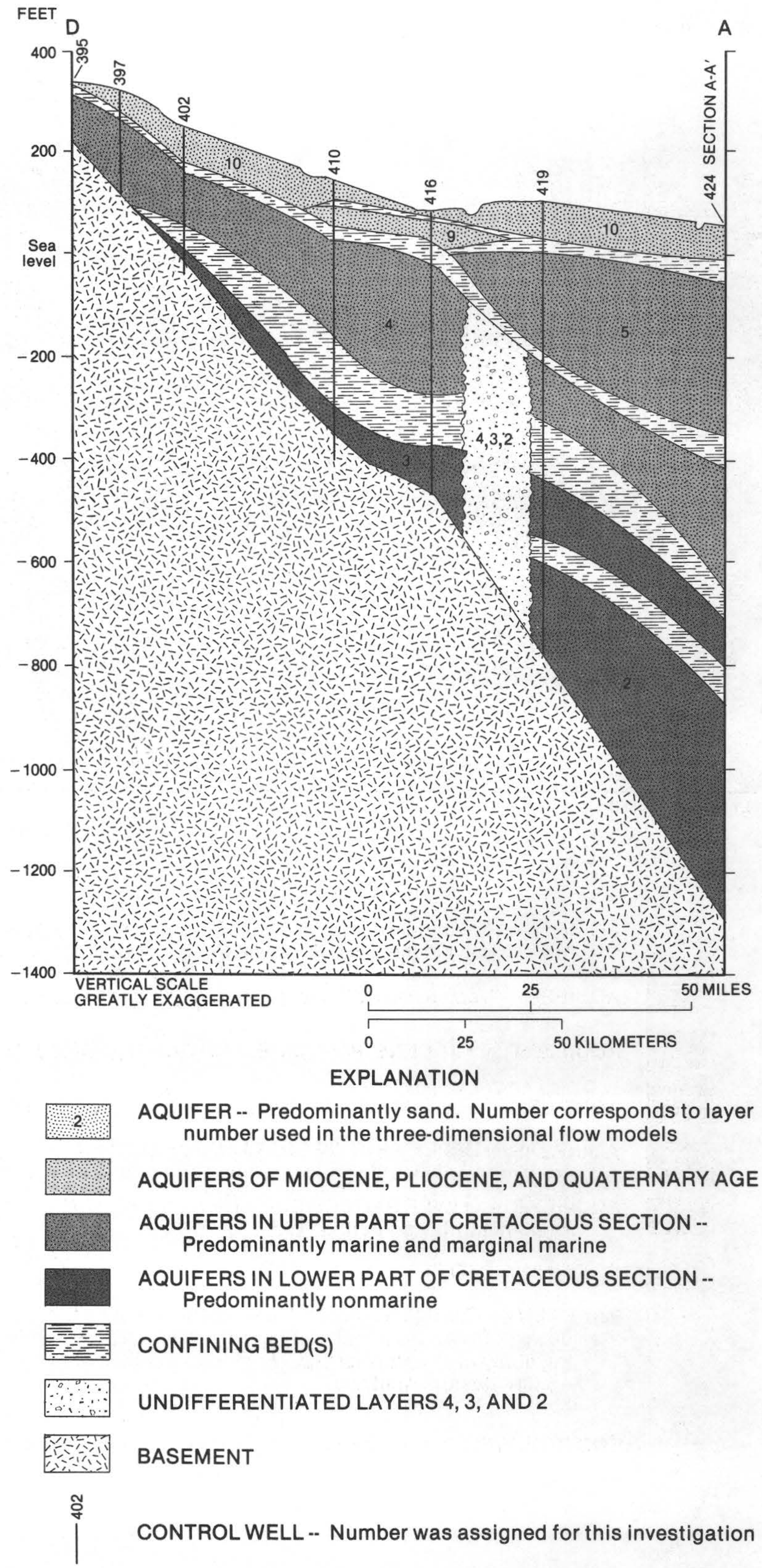
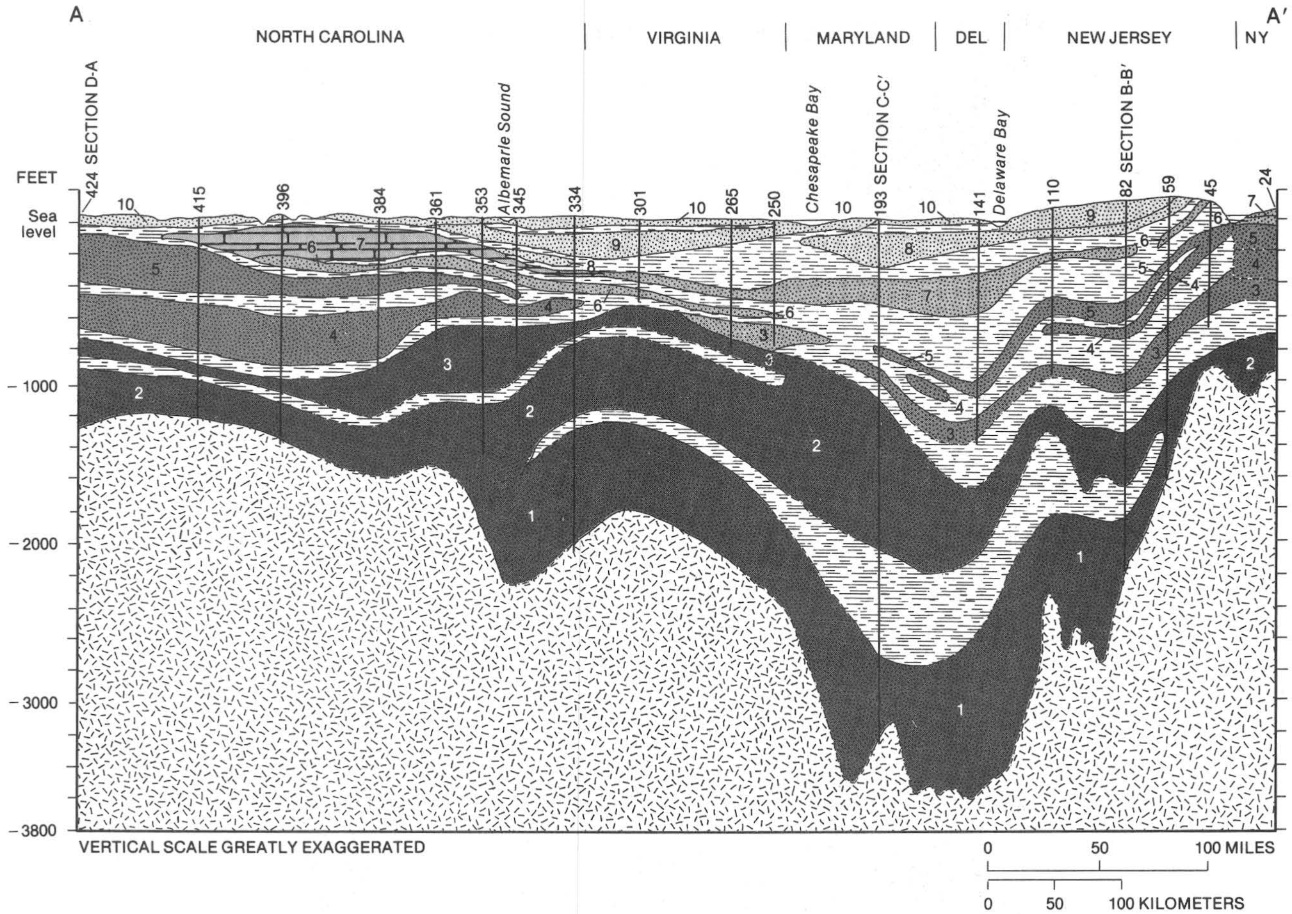


Figure 105.—Hydrogeologic section D-A, southern North Carolina, in the northern Atlantic Coastal Plain (location of section shown in fig. 102).



EXPLANATION

AQUIFER -- Predominantly sand. Number corresponds to layer number used in the three-dimensional flow models

AQUIFER -- Predominantly limestone and lime sand, layer 7

AQUIFERS OF MIOCENE, PLIOCENE, AND QUATERNARY AGE

AQUIFERS OF PALEOCENE, EOCENE, AND OLIGOCENE AGE

AQUIFERS IN UPPER PART OF CRETACEOUS SECTION -- Predominantly marine and marginal marine

AQUIFERS IN LOWER PART OF CRETACEOUS SECTION -- Predominantly nonmarine

CONFINING BED(S)

BASEMENT -- Configuration of the basement drawn on the basis of the contour maps of the basement prepared under the study of the Northern Atlantic Coastal Plain Regional Aquifer System Analysis

301 CONTROL WELL -- Number was assigned for this investigation

Figure 106.—Hydrogeologic section A-A′, from North Carolina to Long Island, New York, in the northern Atlantic Coastal Plain (location of section shown in fig. 102).

SALTWATER-FRESHWATER RELATIONS

By Harold Meisler

Salty ground water underlies freshwater in the eastern part of the northern Atlantic Coastal Plain. Chloride concentrations generally increase with depth within a transition zone between the deepest freshwater and the underlying saltwater. The zone containing concentrations of chloride ranging from 1,000 to 18,000 milligrams per liter (mg/L) (the approximate chloride concentration in seawater) is depicted in a section across southern New Jersey and the adjacent Continental Shelf

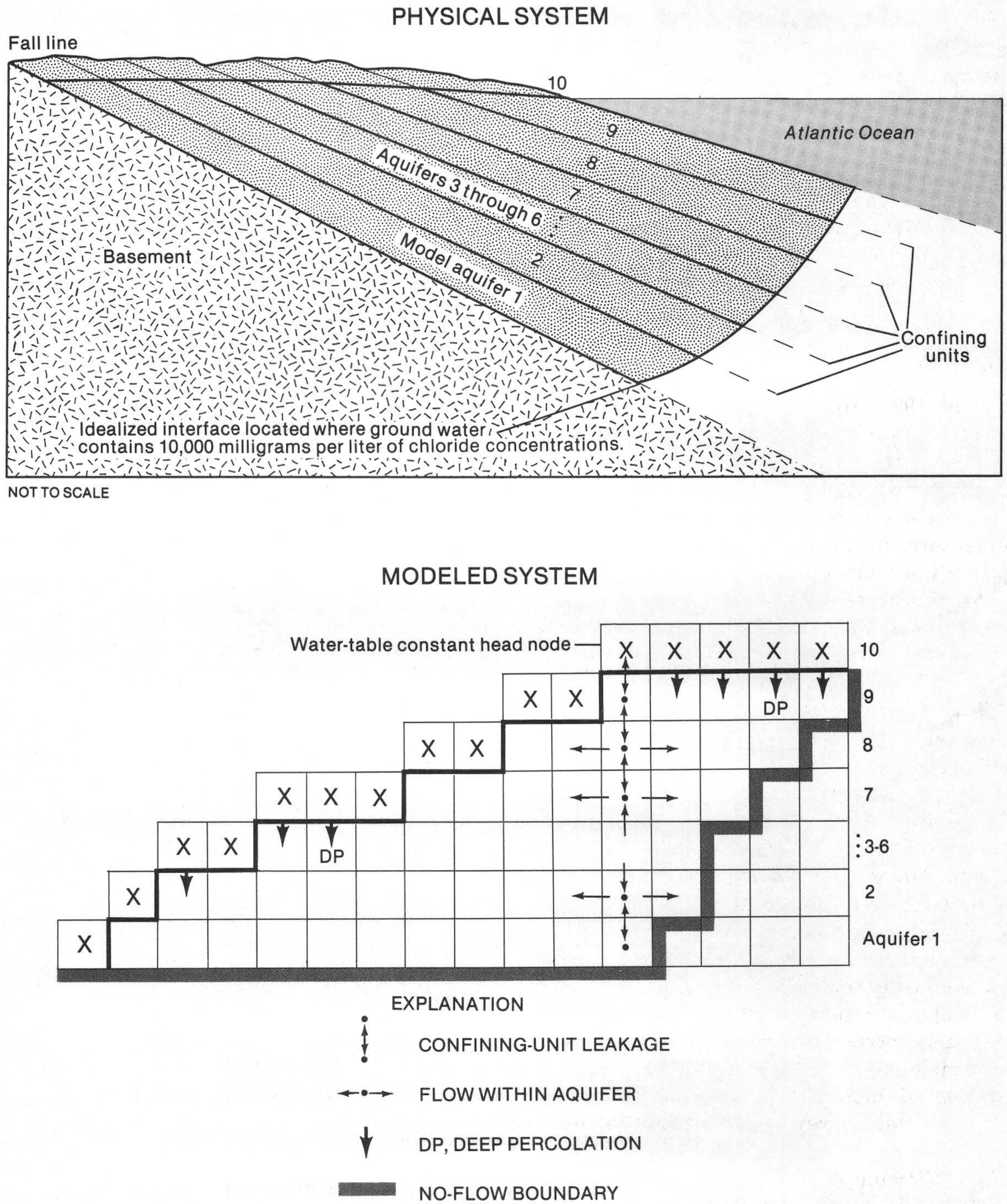


Figure 107.—Comparison of physical and model-simulated systems used for steady-state simulations of the Northern Atlantic Coastal Plain regional aquifer system.

(fig. 115). Altitudes of the 250 and 10,000 mg/L concentrations of chloride are shown in figures 116 and 117, respectively.

The saltwater-freshwater transition zone generally deepens inland from the coast except in New Jersey, where it is deepest along the coast, and locally in North Carolina (fig. 117). The transition zone is shallowest in North Carolina, particularly in the vicinities of the Cape Fear River and Albemarle Sound, and deepens northward to Maryland and the coast of New Jersey. It is relatively shallow, however, in the vicinity of Delaware Bay. Concentrations of chloride of less than 5,000 mg/L extend about 55 miles off the New Jersey coast (fig. 115). The offshore extension of the transition zone decreases southward toward southern Virginia.

Depth to the saltwater-freshwater transition zone is partly controlled by the natural flow pattern of the fresh ground water in the aquifer system. Areas where the transition zone is relatively shallow generally coincide with areas of natural ground-water discharge—the most prominent of these are Delaware Bay, lower Chesapeake Bay, Albemarle Sound, and the Cape Fear River.

The effect of eustatic sea-level fluctuations on the location and development of the transition zone in New Jersey was analyzed by means of a finite-difference cross-sectional flow model (Meisler, and others, 1984). The model simulates the sedimentary wedge from the Delaware River to the Continental Slope (fig. 115) as a highly anisotropic porous medium (ratio of simulated lateral to vertical hydraulic conductivity ranges from 1,000:1 to 1 million:1). Simulated steady-state freshwater flow is separated from static saltwater (18,000 mg/L of chloride) by a sharp interface. Freshwater pressure at the simulated interface equals the pressure exerted by a column of static seawater.

Positions of the interface in a steady-state condition of the aquifer system were simulated for several hypothetical sea levels and for dif-

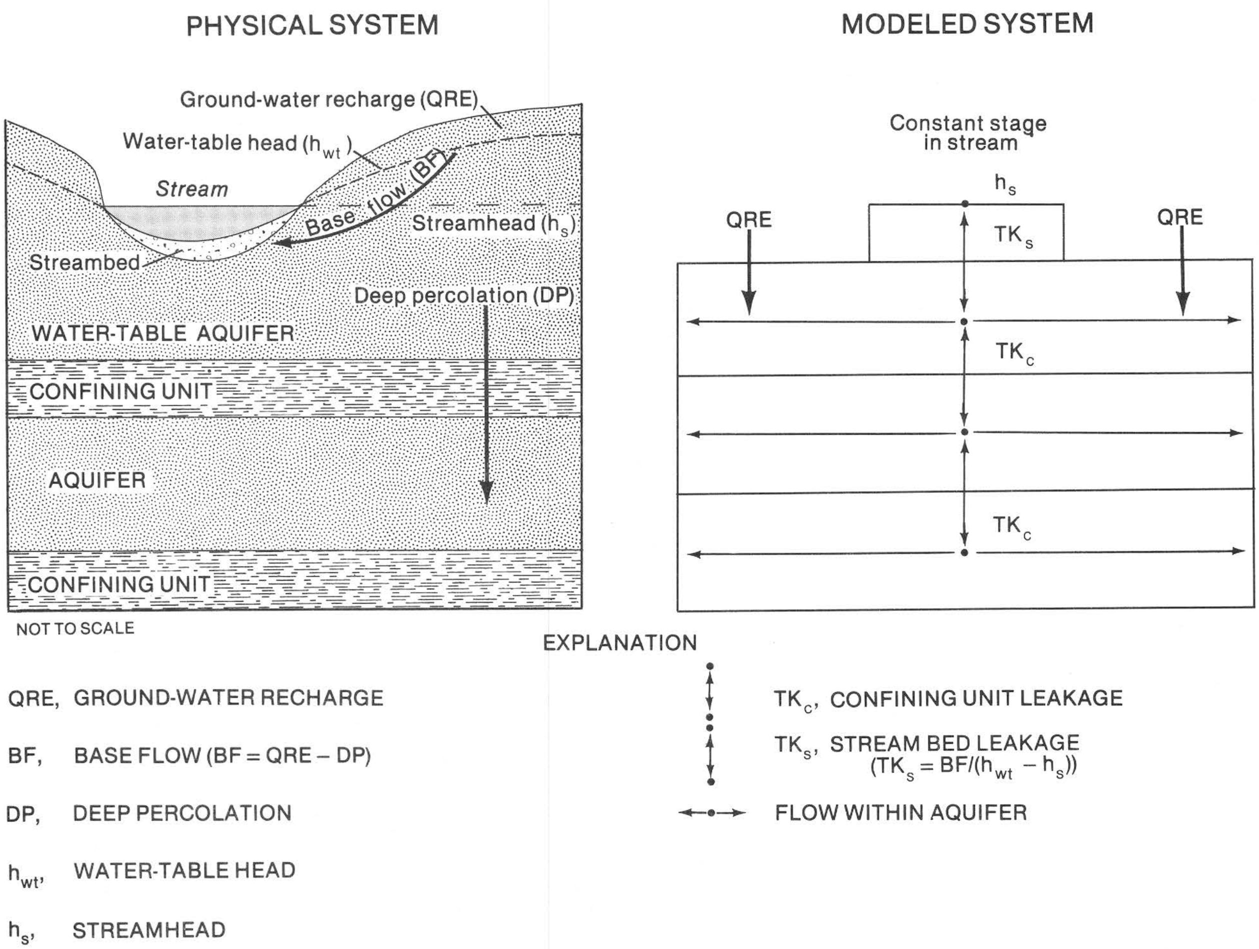


Figure 108.—Comparison of physical and model-simulated systems used for transient simulations of the Northern Atlantic Coastal Plain regional aquifer system.

ferent distributions of vertical and lateral hydraulic conductivity. The simulated sea levels range from present sea level to 150 feet below present sea level. Simulated hydraulic conductivities range from 1 to 20 ft/d lateral conductivity, and 1×10^{-6} to 3.3×10^{-4} ft/d for vertical conductivity. The simulated interface position is sensitive to anisotropy (ratio of lateral conductivity to vertical conductivity) and to sea level. Increasing anisotropy causes the interface to be

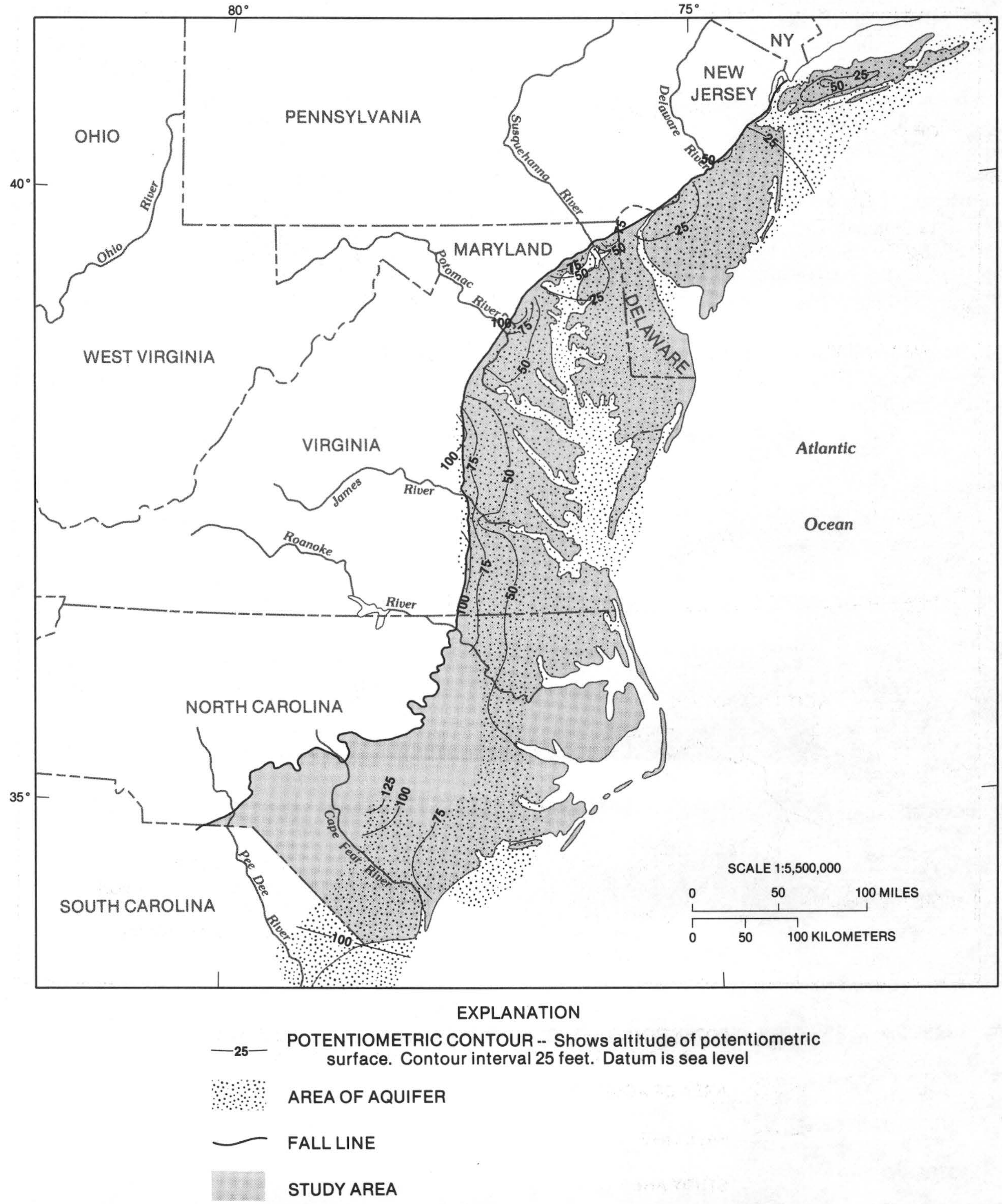


Figure 109.—Simulated prepumping potentiometric surface of an aquifer consisting of sediments of Upper Cretaceous age (model aquifer layer 2), the Northern Atlantic Coastal Plain regional aquifer system.

shallower and to extend farther offshore. Lowering sea level causes the interface to be deeper and to extend farther offshore.

A comparison of several interfaces, computed for different sea levels, with observed isochlors is shown in figure 115. The most probable distribution of hydraulic conductivities (2.5 to 10 ft/d for lateral hydraulic conductivities and 8.3×10^{-5} to 3.3×10^{-4} ft/d for vertical conductivity) and a constant anisotropic ratio of 30,000:1 were used

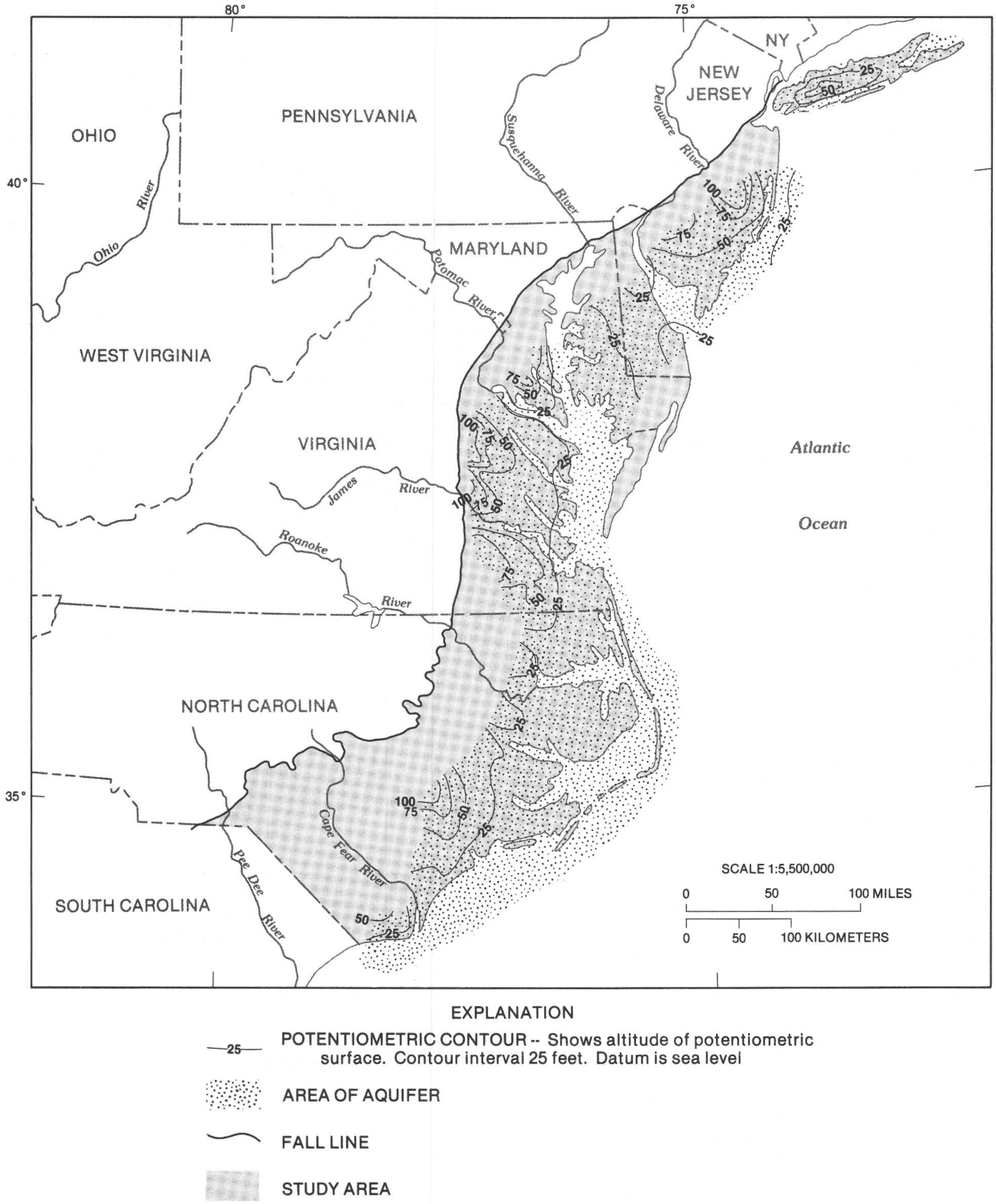


Figure 110.—Simulated prepumping potentiometric surface of an aquifer consisting of sediments of Eocene age (model aquifer layer 7), the Northern Atlantic Coastal Plain regional aquifer system.

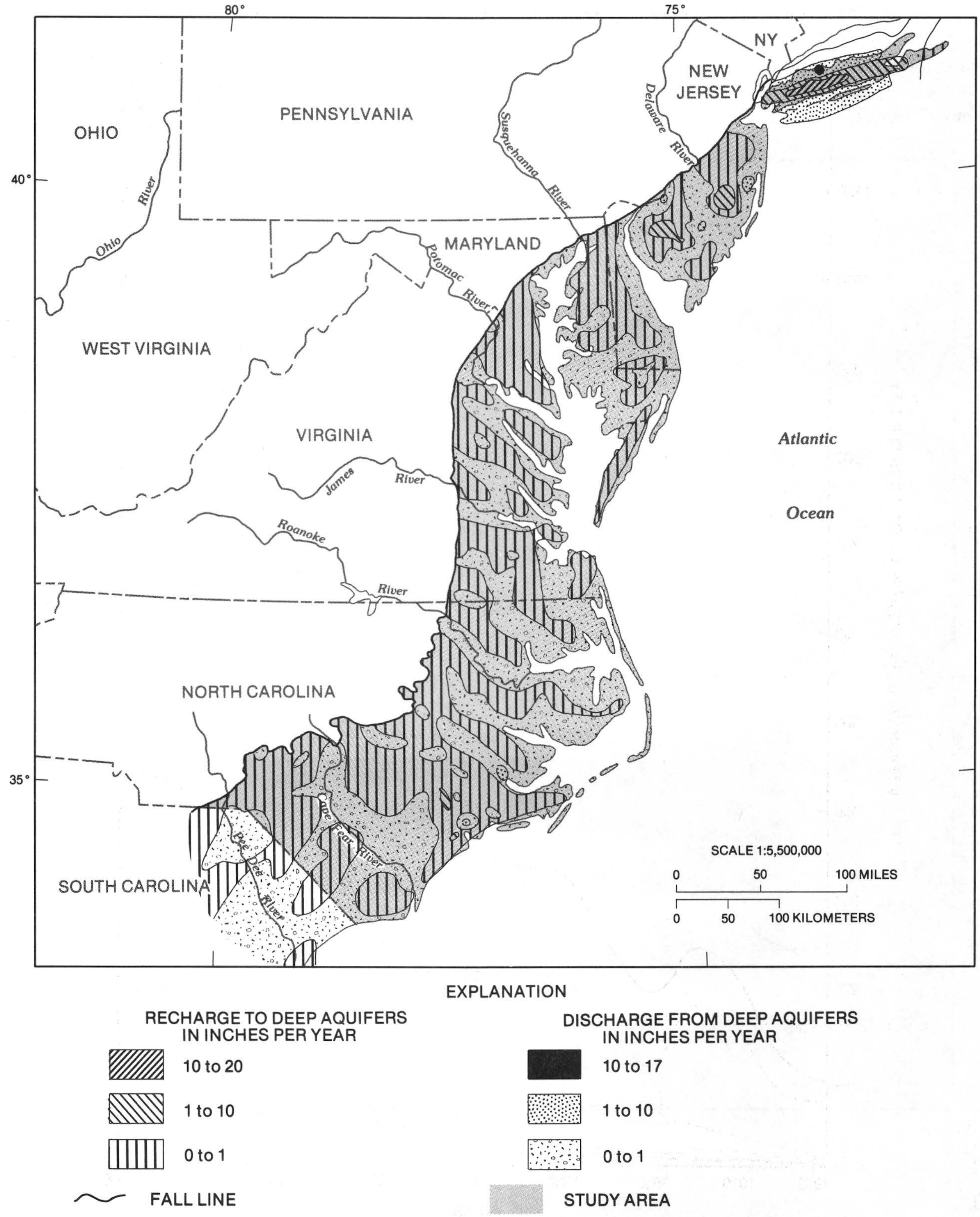


Figure 111.—Simulated deep percolation to and discharge from the confined aquifers of the Northern Atlantic Coastal Plain regional aquifer system.

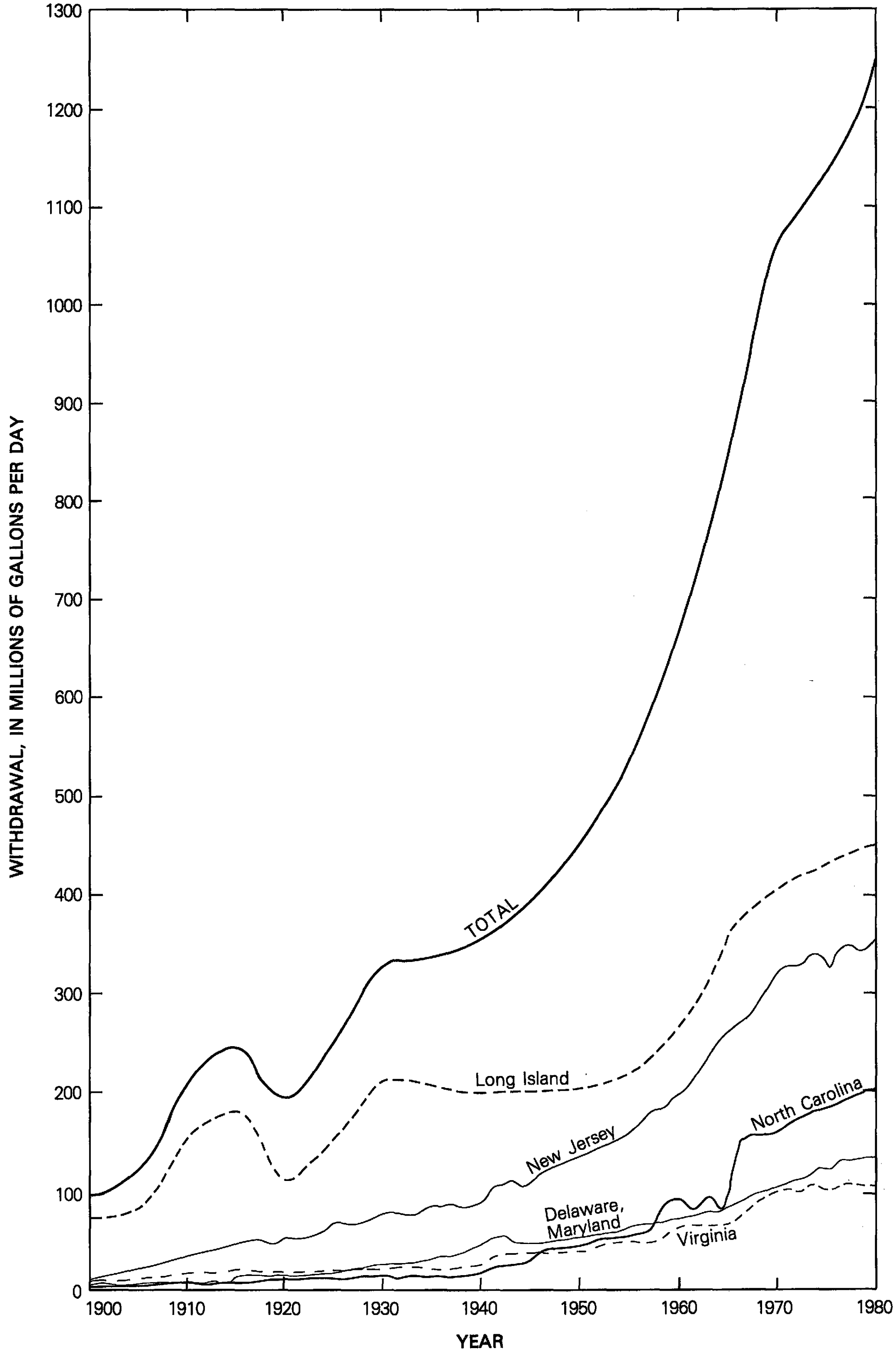
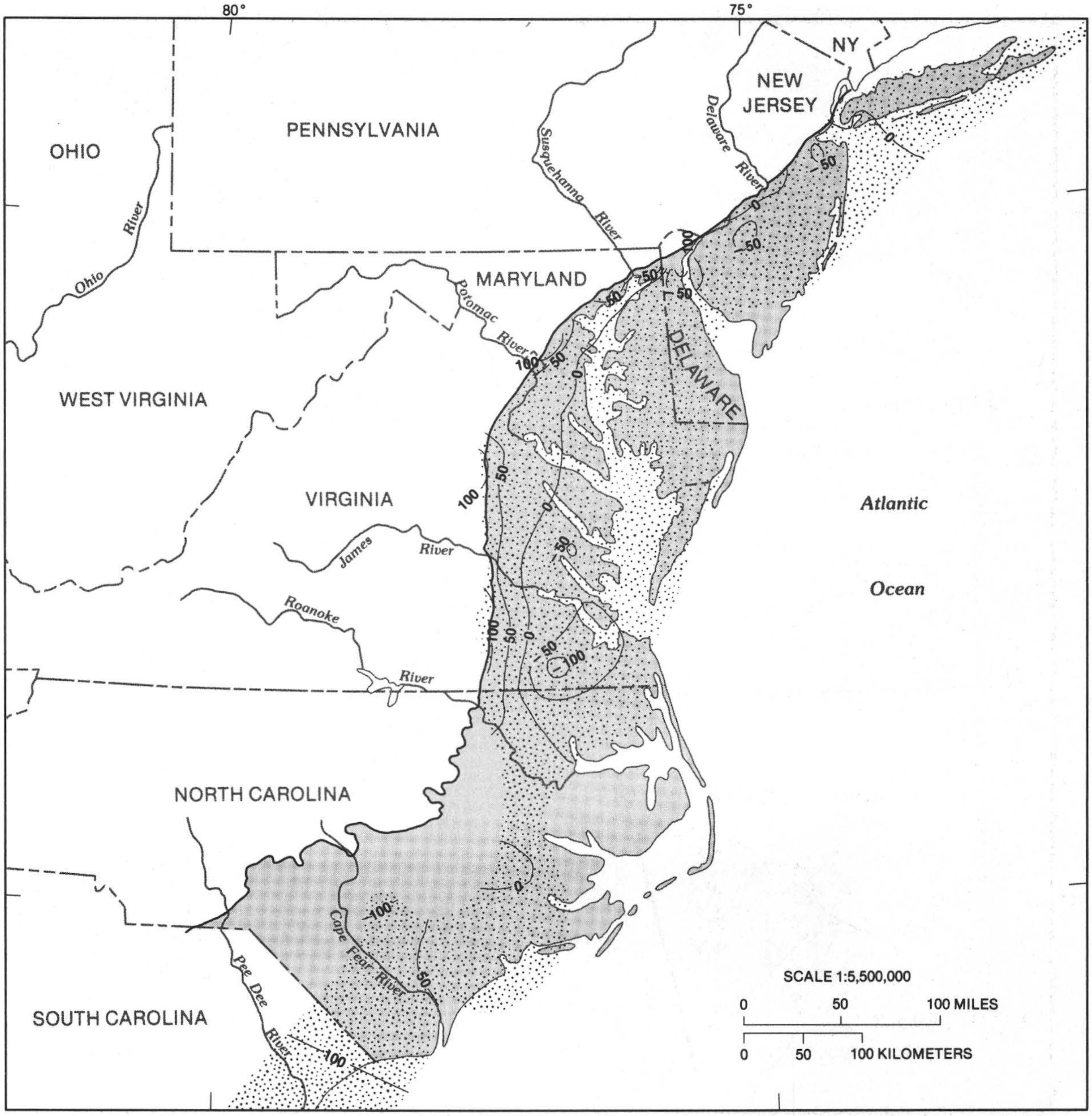


Figure 112.—Ground-water withdrawals from the Northern Atlantic Coastal Plain regional aquifer system, 1900-80.



EXPLANATION

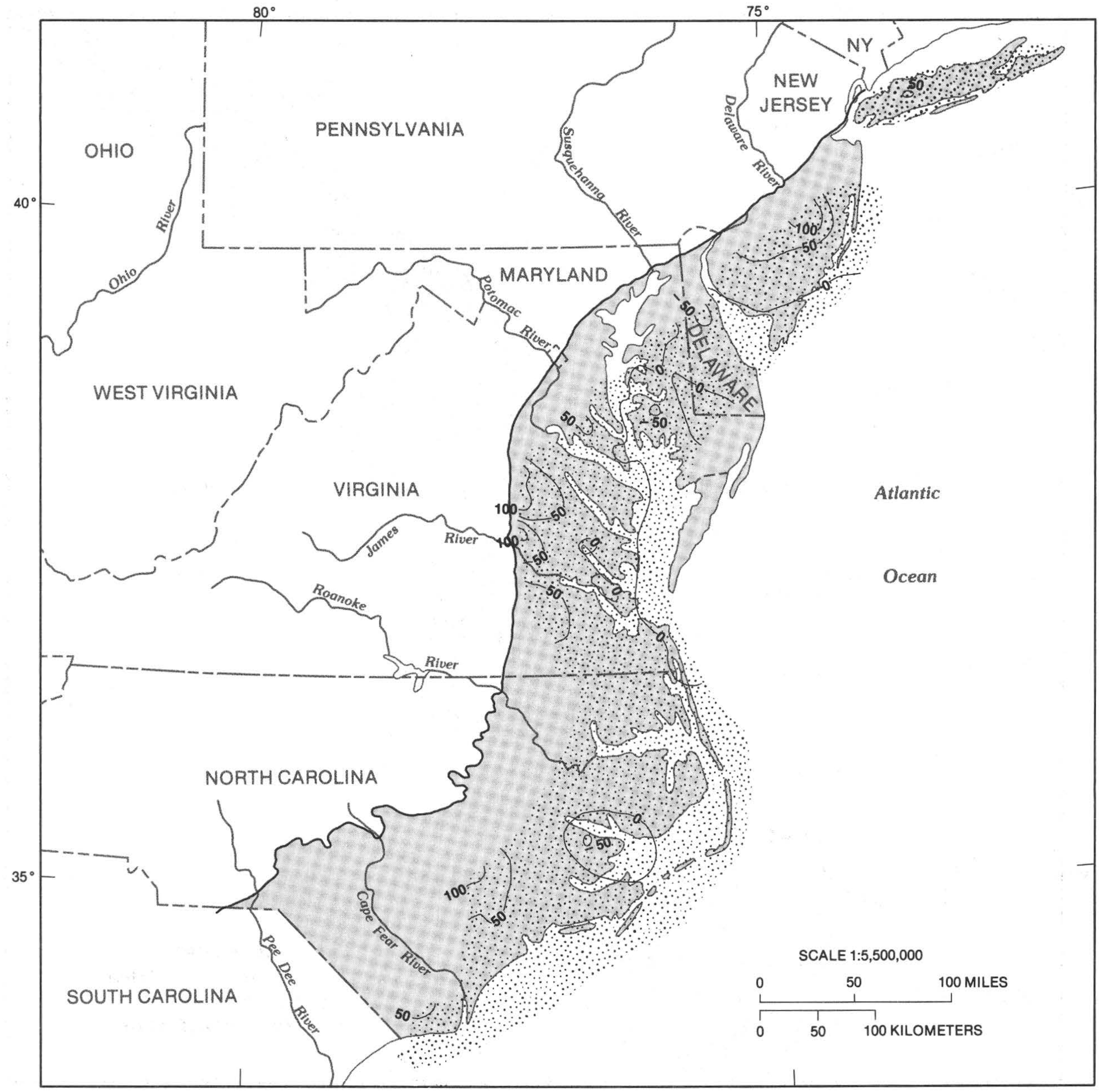
—50— POTENTIOMETRIC CONTOUR -- Shows altitude of potentiometric surface. Contour interval 50 feet. Datum is sea level

AREA OF AQUIFER

FALL LINE

STUDY AREA

Figure 113.—Simulated 1980 potentiometric surface of an aquifer consisting of sediments of Upper Cretaceous age (model aquifer layer 2), the Northern Atlantic Coastal Plain regional aquifer system.



EXPLANATION

—50— POTENTIOMETRIC CONTOUR -- Shows altitude of potentiometric surface. Contour interval 50 feet. Datum is sea level

AREA OF AQUIFER

FALL LINE

STUDY AREA

Figure 114.— Simulated 1980 potentiometric surface of an aquifer consisting of sediments of Eocene age (model aquifer layer 7), the Northern Atlantic Coast regional aquifer system.

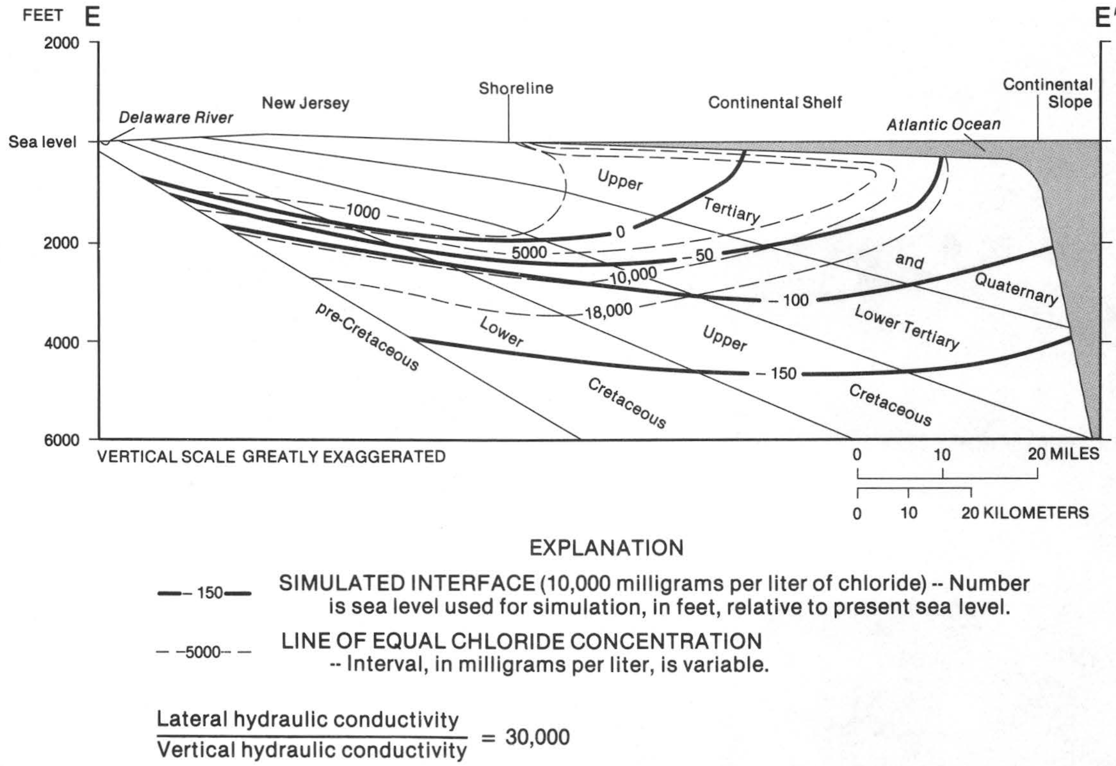
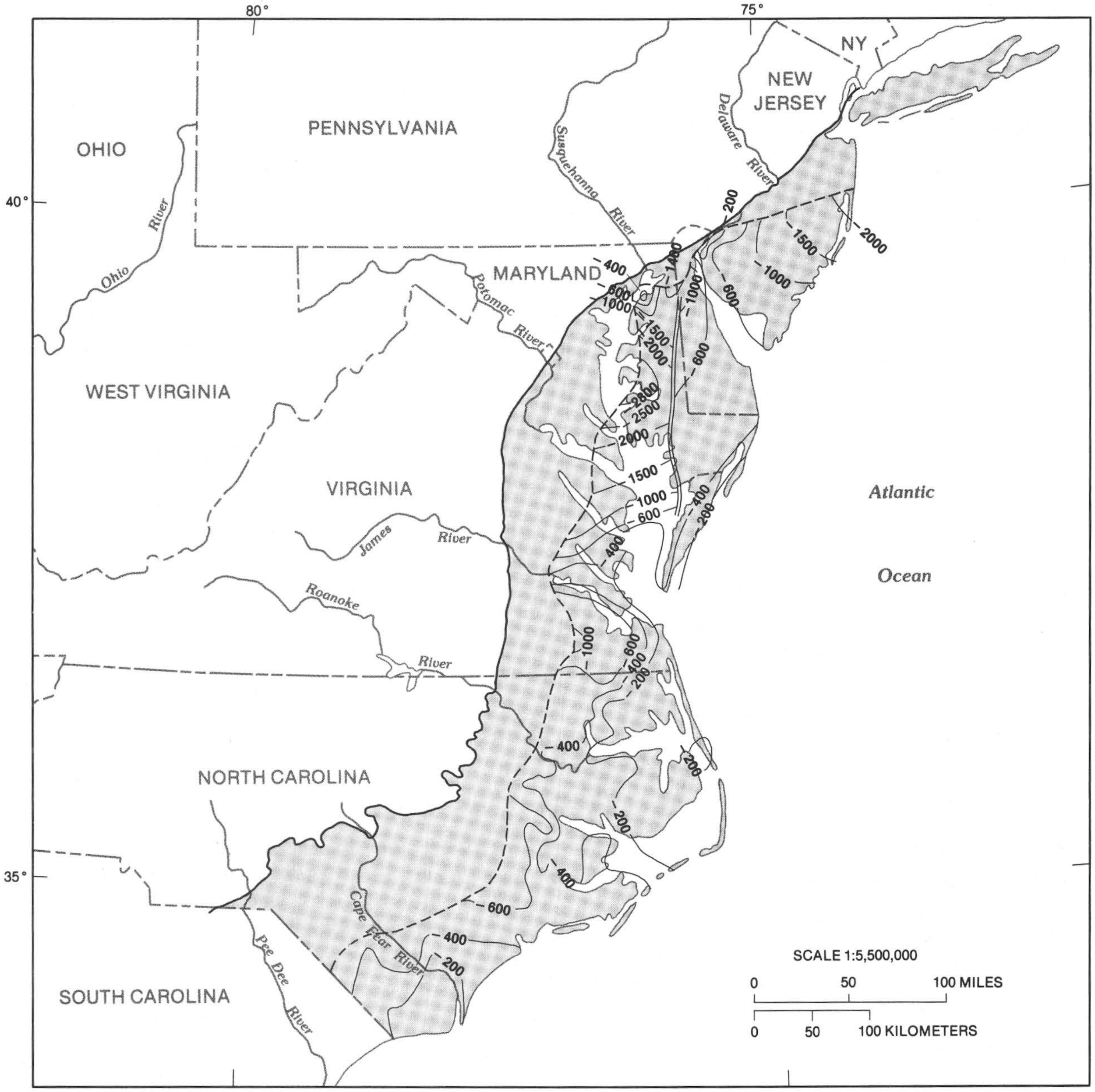


Figure 115.—Hydrogeologic section from southern New Jersey to the Continental Slope, showing isochlors and simulated saltwater-freshwater interface in the northern Atlantic Coastal Plain (location of section shown in fig. 117).

to compute the interfaces shown in figure 115. The offshore position of the center of the transition zone (10,000 mg/L isochlor) estimated from field data coincides approximately with the interface computed for a sea level of 50 feet below present sea level. Onshore the estimated 10,000 mg/L isocholor coincides with the interface computed for a sea level of 100 feet below present sea level (fig. 115).

These simulations suggest that the position of the freshwater-saltwater transition zone estimated from field data in southern New Jersey, and the occurrence of relatively fresh ground water (compared to seawater) in the offshore area reflect sea levels that were probably between 50 to 100 feet below present sea level. Although the average sea level during the Quaternary period probably was in this range, it is likely that the observed and estimated transition zone reflects sea-level conditions for a longer period of time that includes the Pliocene and part of the Miocene. The saltwater-freshwater transition zone off the coast of southern New Jersey is presently moving slowly landward in response to the most recent sea-level rise which started about 16,000 years ago. Estimates of lateral ground-water velocity suggest that this movement probably is at a rate of about 0.2 mile per 10,000 years (Meisler and others, 1984). Ground-water development in southern New Jersey, however, has probably increased the lateral velocity, but the magnitude of the influence of pumping has not yet been investigated.

Farther south, in southeastern Virginia and North Carolina, the transition zone is much shallower, and relatively fresh ground water (compared to seawater) probably does not extend very far offshore (fig. 117). Here, the position of the transition zone probably reflects higher average sea levels than those suggested for New Jersey. Indeed, depositional environments inferred from sediments of Middle Miocene age and younger indicate that marine submergence was generally more prevalent to the south. The broad saltwater-freshwater transition zone probably developed as a result of saltwater circulation caused by large-scale sea-level fluctuations during the later Tertiary and Quaternary. Repeated advance and retreat of the salty ground water caused the saltwater and freshwater to mix.



EXPLANATION

--800-- WATER-QUALITY-ZONE CONTOUR -- Shows altitude of 250 milligrams per liter chloride concentration. Contour interval, in feet, is variable. Datum is sea level

EASTERN LIMIT OF DEEP FRESHWATER WEDGE

LIMIT OF 250 MILLIGRAMS PER LITER CHLORIDE CONCENTRATION IN COASTAL PLAIN SEDIMENTS

FALL LINE

STUDY AREA

Figure 116.—Altitudes of the 250 milligrams per liter chloride concentration of ground water in the northern Atlantic Coastal Plain.

GEOCHEMISTRY

By LeRoy L. Knobel, Trenton, New Jersey
and Francis H. Chapelle, Towson, Maryland

The chemical nature of ground water in the study area is highly variable (Back, 1966). This variability results from both the mineralogical complexity of the sediments and the diverse nature of flow patterns.

The environment of deposition causes the most striking differences in the mineralogy of the sediments, which were deposited under marine, marginal marine, and nonmarine conditions. The resulting differences in mineralogical composition and their effect on the chemical character of the ground water are discussed below.

Nonmarine Sediments

Nonmarine sediments, which occur primarily in the Lower Cretaceous and the lower part of the Upper Cretaceous, are commonly fluviodeltaic deposits—they generally consist of quartz sand with minor amounts of silicate minerals and interbedded silt and clay lenses. The clay lenses are usually light colored, consist mostly of illite, kaolinite, and montmorillonite, and contain significant quantities of pyrite and lignite. Hard beds rich in iron minerals occasionally occur at or near the contacts of sandy and clayey layers. These sediments commonly grade laterally into marginal marine deposits.

Precipitation entering the nonmarine sediments in the outcrop areas is saturated with dissolved oxygen and has a low pH. The dissolved oxygen reacts with lignite and pyrite to produce ferric hydroxide, dissolved carbon dioxide, and hydrogen and sulfate ions (Chapelle, 1984). The dissolved carbon dioxide dissociates into additional hydrogen and bicarbonate ions. The hydrogen ions are consumed by hydrolysis of silicate minerals such as the feldspars and micas to produce dissolved silica (SiO_2) and the following dissolved ions: calcium (Ca^{2+}), magnesium (Mg^{2+}), sodium (Na^+), potassium (K^+), and bicarbonate (HCO_3^-). The composition of silicate minerals is variable and controls which ions are actually released into the ground water (Faust and Aly, 1981).

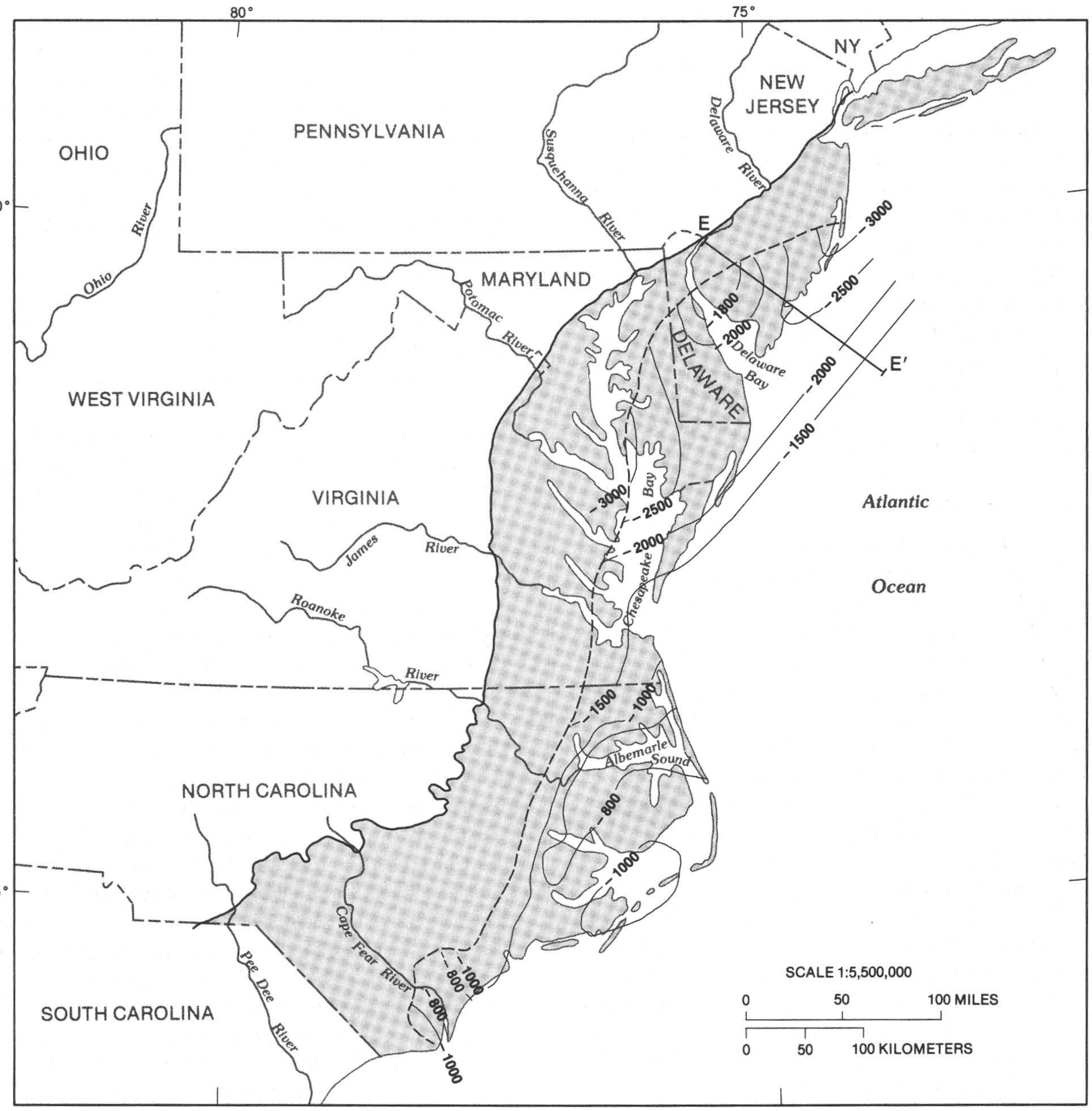
Ca^{2+} and Mg^{2+} released by weathering of the silicate minerals are removed from solution by cation exchange processes and are replaced by Na^+. The source of sodium on cation exchange minerals is probably either from loading of the exchange sites during episodes of saltwater intrusion or from formation of sodium-rich clay by weathering processes prior to deposition of the clay. The sulfate ion (SO_4^{-2}) produced by the dissolution of pyrite is reduced to sulfide or remains in the ground water as SO_4^{-2} depending on local conditions. Precipitated ferric hydroxide ($Fe(OH)_3$) is commonly observed as secondary mineralization in the nonmarine sediments.

Marine Sediments

Glauconitic Sediments.—Glauconite-rich sediments constitute a significant part of the marine deposits of the northern Atlantic Coastal Plain, primarily in the aquifers composed of sediments of Late Cretaceous and early Tertiary age. The Aquia aquifer in southern Maryland (fig. 102) is typical of these sediments and will be used to summarize the chemical processes. The Aquia aquifer is composed primarily of quartz sand, glauconite, and shell debris. The shell material is composed of magnesium calcite and aragonite and is commonly associated with secondary calcite cementation. Detrital lignite is found throughout these sediments.

Glauconite has a high capacity to exchange ions adsorbed onto its surface for ions in the ground water. Because glauconite is formed in a marine environment, the exchange sites are initially occupied by sodium; changes in water chemistry due to ion exchange reflect changes in the exchangeable cation composition of glauconite. Analyses of paired samples of ground water and glauconite from the Aquia aquifer in Maryland indicate that the exchangeable cation composition of glauconite changes systematically along flow paths in a manner related to changes in the concentrations of cations in the ground water. Cation exchange reactions thus appear to be processes that simultaneously alter water and glauconite composition along a flow path (Chapelle and Knobel, 1983). Because the exchange reaction is rapid, any Ca^{2+} in solution would exchange for Na^+ on glauconite, a process that continues until sodium is no longer available and Ca^{2+} is the dominant ion both in ground water and on glauconite. The concentrations of Na^+ on Aquia glauconite are plotted in figure 118 as a function of distance along a flow path. Examination of figure 118 indicates that the exchange front in the Aquia aquifer is currently more than 30 miles along the hydraulic gradient.

Ground water in the outcrop area of the Aquia aquifer in southern Maryland is calcium-magnesium bicarbonate in character, and



EXPLANATION

--800-- WATER-QUALITY-ZONE CONTOUR -- Shows altitude of 10,000 milligrams per liter chloride concentration. Contour interval, in feet, is variable. Datum is sea level

E E′ TRACE OF HYDROGEOLOGIC SECTION

LIMIT OF 10,000 MILLIGRAMS PER LITER CHLORIDE CONCENTRATION IN COASTAL PLAIN SEDIMENTS

FALL LINE

STUDY AREA

Figure 117.—Altitudes of the 10,000 milligrams per liter chloride concentration of ground water in the northern Atlantic Coastal Plain.

changes to sodium bicarbonate downgradient. To explain the chemical processes further, the concentrations of $Ca^{2+} + Mg^{2+}$, Na^+ and HCO_3^- in Aquia water were plotted as a function of distance along a flow path as shown in figure 119. The aquifer can be divided into three regions based on changes in water chemistry.

Region I.—This region includes the outcrop area. $Ca^{2+} + Mg^{2+}$ and HCO_3^- concentrations increase to a peak and begin to decrease as water flows downgradient, due to the rapid dissolution of high-magnesium calcite shell material in the presence of CO_2. Because high-magnesium calcite has a higher solubility than low-magnesium calcite, the water becomes saturated with respect to the latter and precipitation of calcite occurs simultaneously with dissolution of the shell material. These processes account for the distribution of ions in the ground water occurring in Region I as well as for the observed distribution of secondary calcite cementation in this area (Chapelle, 1983). Ion exchange is not a significant process in Region I because the exchange capacity of the glauconite has been used up.

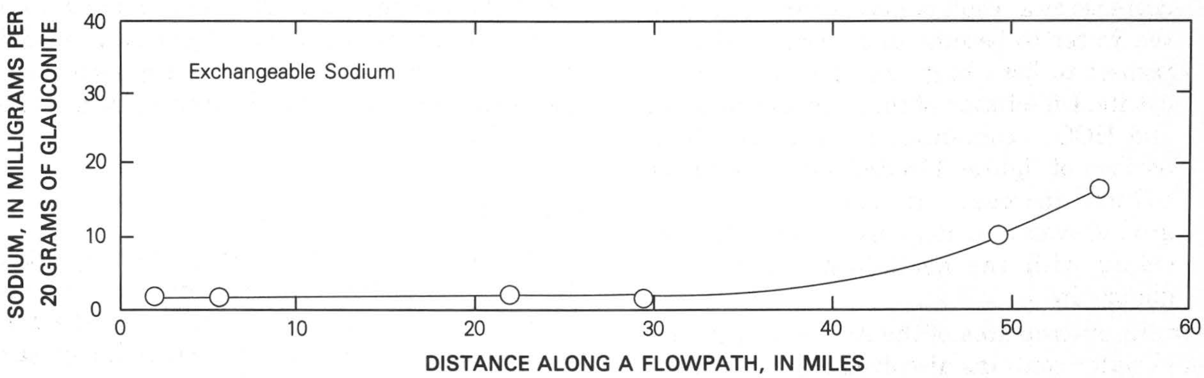


Figure 118.—Concentrations of exchangeable sodium on Aquia glauconite samples plotted against distance along a flow path in southern Maryland.

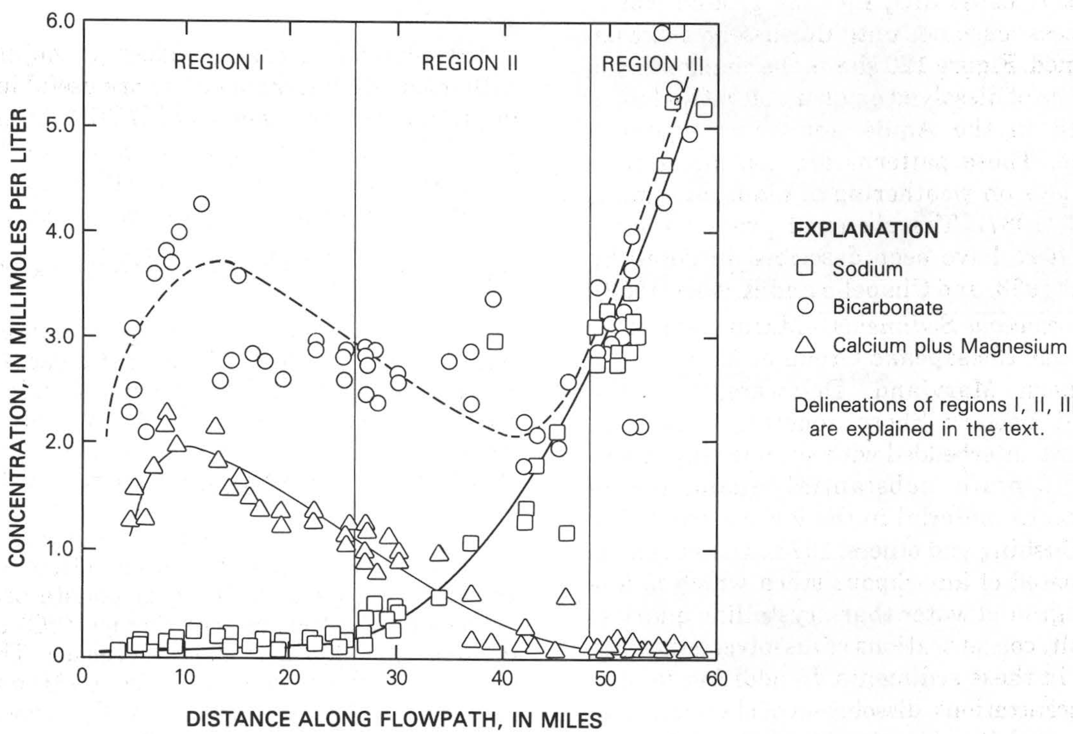


Figure 119.—Concentrations of dissolved sodium, calcium plus magnesium, and bicarbonate of the Aquia aquifer water plotted against distance along a flow path in southern Maryland (modified from Chapelle, 1983).

Region II.—This region is downgradient from Region I. The aquifer is confined and therefore is closed off from atmospheric CO_2. The water chemistry exhibits rising Na^+ concentrations and declining $Ca^{2+} + Mg^{2+}$ concentrations. These changes in the water chemistry are primarily due to ion exchange.

Region III.—This region is downgradient from Region II and extends to the downdip boundary of the aquifer. The water is characterized by low $Ca^{2+} + Mg^{2+}$ concentrations and increasing Na^+ and HCO_3^- concentrations. The concentrations of $Ca^{2+} + Mg^{2+}$ decrease as a result of ion-exchange, causing the water to become undersaturated with respect to both high- and low-magnesium calcite. Dissolution of these minerals occurs and HCO_3^- concentration increase. Comparison of figures 118 and 119 shows that sodium increases in concentrations in ground water in Regions II and III, coinciding with the availability of Na^+ on glauconite.

In the outcrop area of the Aquia aquifer, the ground water contains dissolved oxygen and incongruent dissolution of glauconite is an important chemical process that consumes dissolved oxygen and releases SiO_2 into the ground water. This process continues until the dissolved oxygen is consumed. Figure 120 shows the change of concentrations of dissolved oxygen and SiO_2 along a flow path in the Aquia aquifer in southern Maryland. These patterns are consistent with observations on weathering of glauconite made by Wolff (1967). The chemical processes summarized here have been described in detail by Chapelle (1983) and Chapelle and Knobel (1983).

Diatomaceous Sediments.—Marine sediments of the lower Chesapeake Group of Miocene age in southern Maryland, Delaware, and the Maryland Eastern Shore generally consist of quartz sand interbedded with silt and clay. These deposits contain substantial quantities of diatomaceous material in the lower parts of the section (Cushing and others, 1973). Diatom shells are composed of amorhpous silica which is less stable in ground water than crystalline quartz—as a result, concentrations of dissolved silica are elevated in these sediments. In addition to high silica concentrations, dissolution of shell material generates calcium-bicarbonate water which is modified by ion exchange to form a sodium bicarbonate water.

Limestone Deposits.—The Coastal Plain of North Carolina contains extensive deposits of molluscan limestone with substantial secondary recrystallization, which grades laterally into bryozoan limestone and downward into calcareous sand (Brown and others, 1972). Chapelle (1983) developed a model to predict the chemical composition of ground water resulting from carbonate dissolution reactions in the presence of carbon dioxide. The model indicates that two moles of bicarbonate will be produced for each mole of calcium plus magnesium. $Ca^{2+} + Mg^{2+}$ values plotted against HCO_3^- values for all available analyses of ground water from the Castle Hayne aquifer in North Carolina show a good match between observed data and model-derived values, suggesting that the model is a good representation of the chemical system in this area (fig. 121).

Origin of Carbon

Foster (1950) noted that ground water in the Atlantic and Gulf Coastal Plains has anomalously high bicarbonate concentrations; several times higher than those that could be generated by reactions involving dissolved CO_2 gas in recharge water in outcrop areas. Foster (1950) concluded that CO_2 was being generated within the aquifer system.

Stable carbon isotopes used in conjunction with major-ion water chemistry are useful in helping to identify the sources of HCO_3^- in ground water. It is a common practice to relate $^{13}C/^{12}C$ ratios to the Peedee Belemnite (PDB) carbonate standard according to the following expression.

$$\delta^{13}C = \frac{(^{13}C/^{12}C)\text{sample} - (^{13}C/^{12}C)\text{PDB}}{(^{13}C/^{12}C)\text{PDB}} \times 1000$$

If the carbon-containing substance is isotopically lighter than the PDB standard (for example, has less ^{13}C relative to ^{12}C), then it would have a negative value of $\delta^{13}C$. Conversely, if the substance is isotopically heavier than the PDB standard, then it would have positive values of $\delta^{13}C$.

Stable carbon isotope data for Aquia water are plotted versus distance along a flow path as shown by figure 122. The bicarbonate becomes enriched with the heavier isotope (^{13}C) as the distance along a flow path increases. This increase is more pronounced in Region III and corresponds to the rapid rise of HCO_3^- concentrations (see figs. 119 and 122). The authors, using mass balance techniques, concluded that the increases in both the heavier carbon isotope in bicarbonate and the bicarbonate concentrations in the ground water require an aquifer-generated source of CO_2 that is enriched in the heavy isotope. They also noted that in Region III where CO_2 is being generated, anoxic conditions exist (fig. 120).

Winograd and Farlekas (1974) noted similar trends along three hydrochemical sections in New Jersey and also concluded that an aquifer-generated source of CO_2 was required. It is significant to note that the observations of Winograd and Farlekas were based on a study of sediments deposited in a nonmarine to marginal marine environment whereas the observations of the authors were based on a study of marine sediments. This suggests that similar processes may generate CO_2 in aquifers of widely varying characteristics. The possible sources of aquifer-generated CO_2 include: (1) bacterially-mediated fermentation of lignitic material, and (2) bacterially-mediated methanogenesis.

SALTWATER

***By* Harold Meisler**

The saltwater-freshwater transition zone is characterized by increases in concentrations of chloride with increasing depth. Also, concentrations of calcium, magnesium, sodium, potassium,

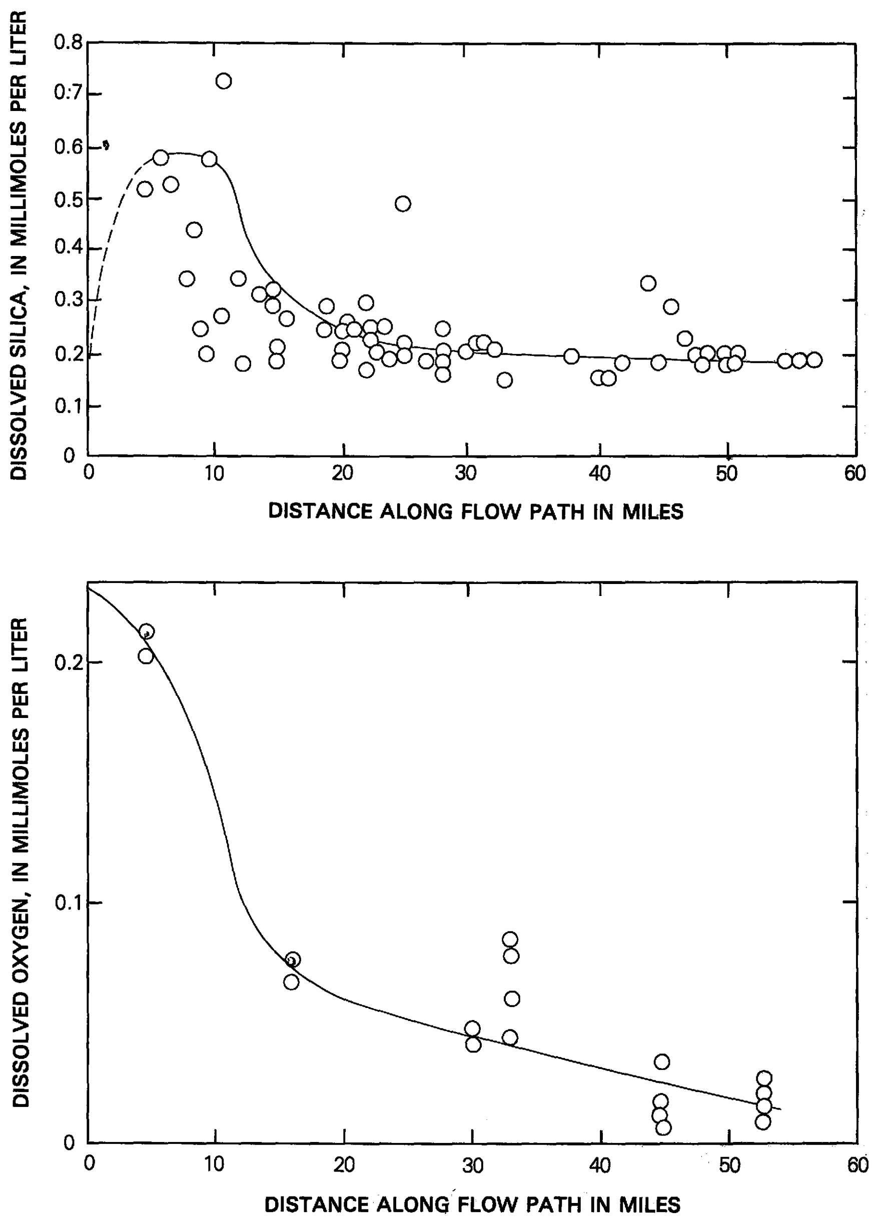


Figure 120.—Concentrations of dissolved silica and dissolved oxygen of the Aquia aquifer water plotted against distance along a flow path in southern Maryland.

and sulfate generally increase as concentrations of chloride increase. Bicarbonate, on the other hand, tends to decrease as concentrations of chloride increase (Meisler, and others, 1984).

The relation of concentrations of these ions to concentrations of chloride indicates that the saltwater-freshwater transition zone in North Carolina is primarily a mixture of seawater and fresh sodium bicarbonate water. In the area from Virginia to New Jersey, the transition zone appears to be largely a mixture of sodium-calcium chloride brine and fresh sodium bicarbonate water. The brine has concentrations of chloride several times higher than that of seawater. It differs significantly from seawater, also, in having higher concentrations of calcium and sodium and lower concentrations of potassium, sulfate, and bicarbonate. A comparison of the equivalent fractions of the principal cations of the most concentrated ground-water samples (Meisler, and others, 1984, table 2) with seawater indicates that, proportionately, the brine has significantly more calcium and less magnesium and potassium than does seawater. The occurrence of a similar brine beneath the transition zone in North Carolina is indicated by interpretation of geophysical logs (Meisler, 1981) and by chemical analyses of water samples taken from depths of greater than 6,000 feet from a well at Cape Hatteras (Meisler, and others, 1984).

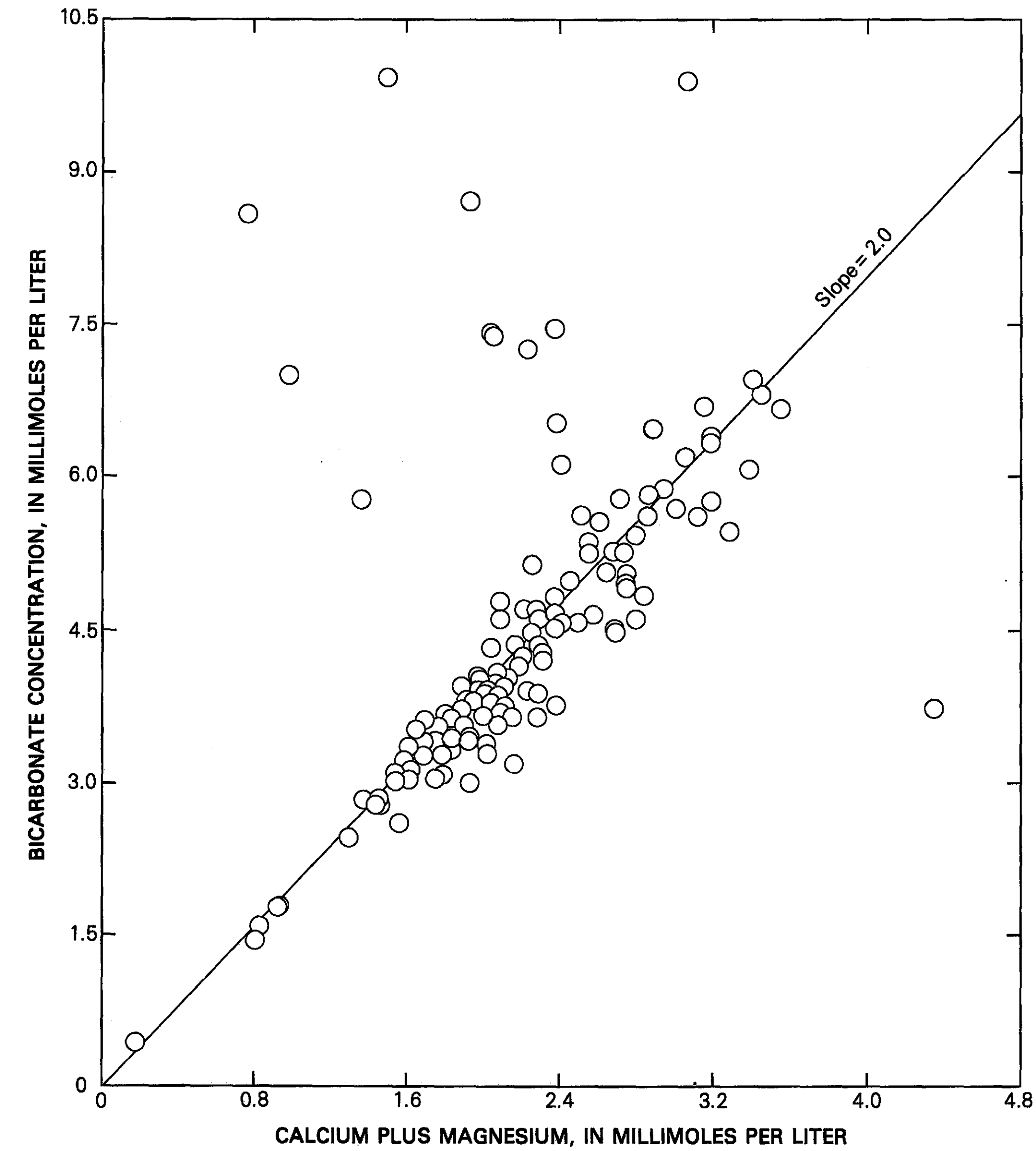


Figure 121.—Calcium plus magnesium concentrations plotted against bicarbonate concentrations for analyses of water taken from the Castle Hayne aquifer in North Carolina.

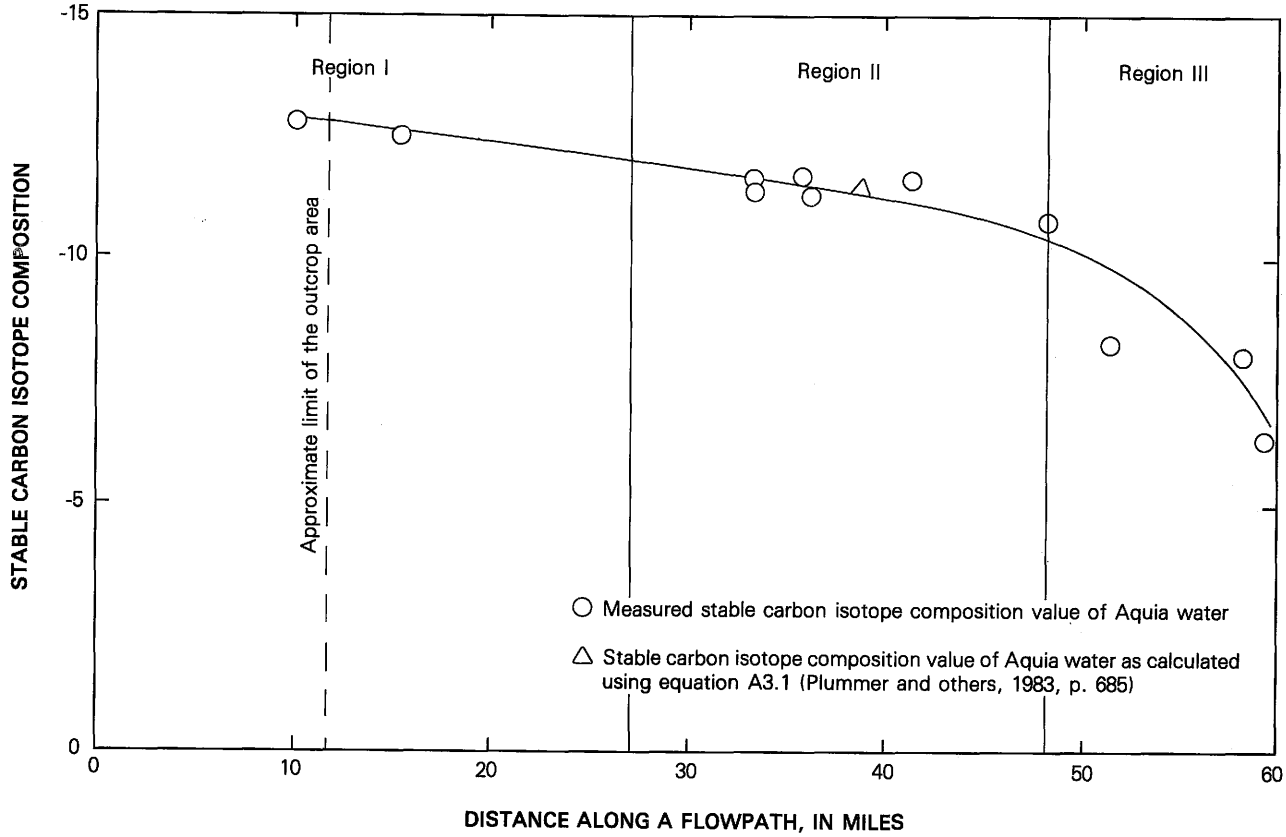


Figure 122.—Measured and calculated values of stable carbon isotope composition plotted against distance along a flow path for water taken from the Aquia aquifer in southern Maryland.

The origin of brines in the Atlantic Coastal Plain is not well understood. Two possible sources are: (1) leaching of evaporitic strata, and (2) concentration of dissolved solids through membrane filtration. Graf (1982) discusses three of the most important mechanisms in the development of the overpressured zones in aquifers, that are needed to drive membrane filtration. One of these, tectonic compression, is not applicable to the northern Atlantic Coastal Plain. A second mechanism, abnormally high geothermal gradients, appears to be ruled out by the geothermal data available. The third mechanism, rapid accumulation of fine-grained sediments, has been evaluated by Bredehoeft and Hanshaw (1968, p. 1097); they concluded that "a sedimentation rate of 500 m [1640 ft] per 1 million years (reasonable for the Gulf Coast) will create fluid pressures approaching lithostatic***." In the northern Atlantic Coastal Plain, however, sedimentation rates have generally been 80 to 160 feet per 1 million years in the outer Continental Shelf and considerably lower on the emerged Coastal Plain. Head measurements in the Atlantic Coastal Plain wells and reported pressures in several offshore oil-test wells do not indicate excess fluid pressures. Any excess pressure generated has probably been dissipated laterally.

Manheim and Horn (1968, p. 229-233), in their discussion of the origin of the Atlantic Coastal Plain brines conclude that leaching of evaporitic strata along with updip movement of the brine accounts for the present distribution of highly saline water. Indeed, drilling and geophysical data (Hathaway and others, 1979, p. 529) have indicated the presence of evaporitic strata beneath the Continental Shelf and Slope.

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OAHU ISLAND REGIONAL AQUIFER-SYSTEM STUDY, HAWAII

By Charles J. Ewart, Honolulu, Hawaii

INTRODUCTION

Several recent studies have concluded that the ground-water resource of the Island of Oahu will be near maximum development by the year 2000. Estimates of the long-term potential of ground-water development of the Oahu regional aquifer system range between 480 and 635 Mgal/d. In 1980, the ground-water withdrawal rate was about 400 Mgal/d, which is 85 percent of the island's total water use. Development of this magnitude unquestionably imposes substantial stresses on the aquifer system. To establish background information and to evaluate the impact of the potential development, a study of the Oahu regional aquifer system was started in 1982 and is scheduled for completion in 1986.

The Island of Oahu has a land area of 604 mi^2 and was formed through the building and subsequent coalescence of two shield volcanoes, the Waianae and Koolau Volcanoes (fig. 123). The Waianae Volcano forms the western part of Oahu, with the Koolau Volcano making up the eastern part. A long period of quiescence followed the initial mountain-building phase. During the quiescence, both volcanoes were deeply eroded. The Waianae Volcano, which is older, became dormant first, thus allowing the westward dipping flows of the Koolau Volcano to overlap the eroded surface of the Waianae Volcano in the central part of the island.

Volcanic activity resumed at the southeastern end of the Koolau Range after a long erosional period. This renewed activity produced the rocks of the Honolulu Volcanic Series. These rocks are confined to relatively small scattered areas. Because the lava flows tended to be ponded in deep valleys, and in general are very dense and of low permeability, they do not constitute an important aquifer. The bulk of the Koolau Volcano consists of thin-bedded, highly permeable basaltic lava flows. An andesitic veneer is common to most Hawaiian volcanoes, but is absent in the Koolau rock mass. In contrast, much of the Waianae range has the andesite veneer, making the surficial lavas somewhat less permeable than those of the Koolau rock mass.

Isostatic adjustment resulted in the subsidence of Oahu to a significant depth, thus submerging permeable lava flows originally extruded subaerially, and placing them in hydraulic contact with the surrounding ocean water. Shifts in sea level and erosion allowed deposits of marine and terrestrial sediments to accumulate behind barrier reefs, forming coastal plains in some areas (fig. 124). The most prominent example of this on Oahu is the southern Oahu coastal plain (fig. 123).

Due to the complexity of the volcanic rocks, a local terminology has developed in Hawaii to describe the occurrence of ground water. A body of ground water floating on and in hydrodynamic equilibrium with saltwater is termed "basal ground water" or "basal water" in Hawaii. Ground water stored in lavas between low permeability intrusion dikes is termed "dike-impounded ground water." In areas where water levels are much higher than the expected basal water table but the geological reason for this occurrence is unknown (ground water is impeded by low permeability dikes, ash beds, or other low permeability rocks), the ground water is referred to as "high-level ground water." The dike-impounded or high-level ground water eventually seeps through or over the low permeability dikes or barriers and discharges into a basal ground-water body, as shown in figure 125, or leaks to streams in valleys that have cut deep into the aquifer. In places, perched ground water also occurs. The perched water is an unconfined ground water separated from an underlying basal water body by an unsaturated zone due to the presence of ash beds or other low permeability materials overlying the basal water body (fig. 125). The Oahu regional aquifer system is divided into several aquifers using the technical terms discussed above.

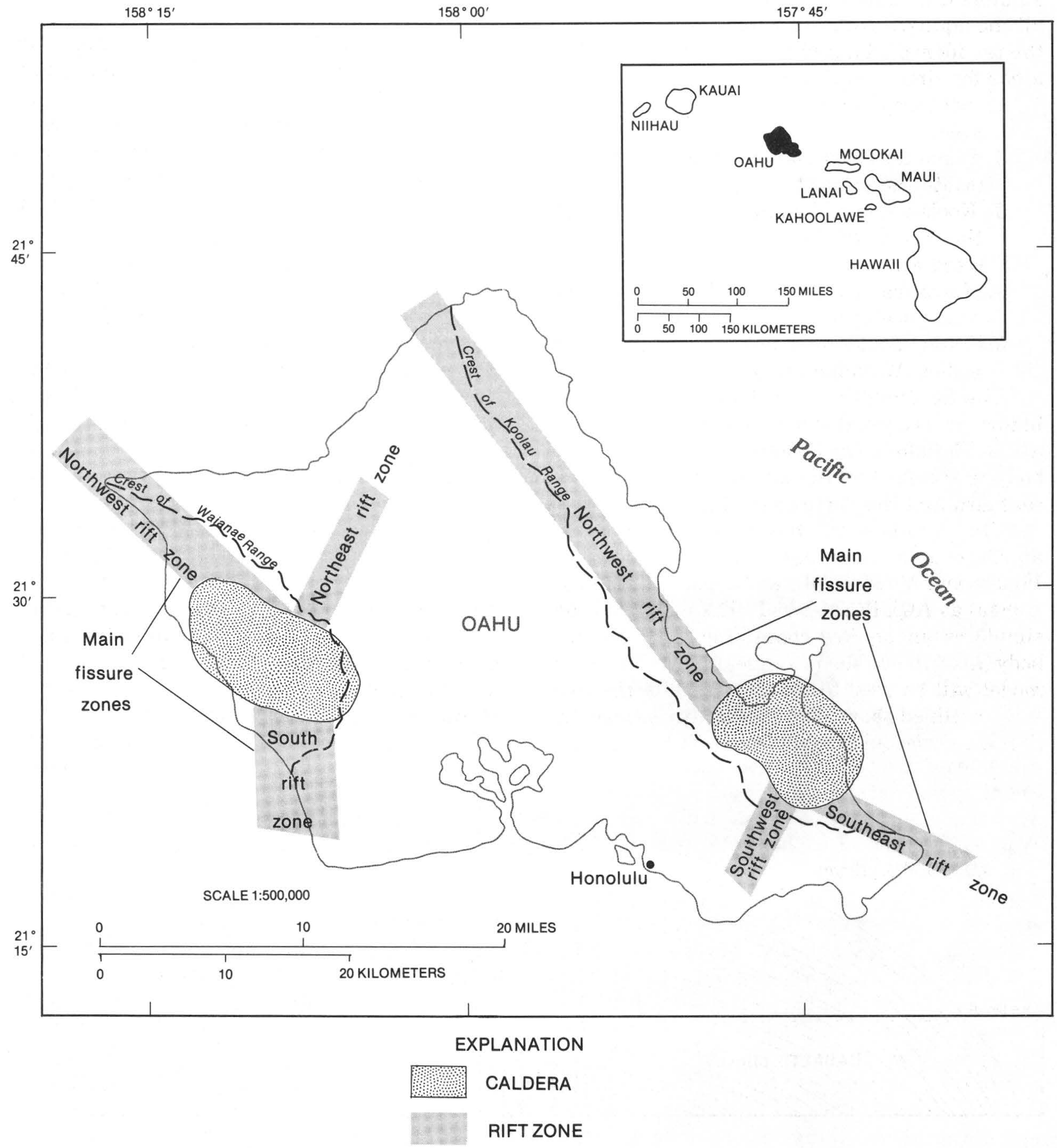

Figure 123.—Rift zones and calderas of Koolau and Waianae Volcanoes on the Island of Oahu, Hawaii (modified from Macdonald, 1972).

PROGRESS AND RESULTS

An annotated bibliography has been comiled. Compilation of a hydrologic data base for all aquifers is virtually completed. From available data, interpretation of information on hydrology and hydraulics of three of the ten identified aquifers (fig. 126) has been completed. Because all the aquifers are interrelated to some extent, the ten identified aquifers are grouped into five areas for simulation purposes. They are:

1. Southeast basal water body (southeast area).
2. Honolulu-Pearl Harbor basal water body (southern area).
3. Koolau dike-impounded water body and the northeast basal water body (windward area).
4. Kawailoa, Waialua, and Mokuleia basal water bodies (north-central area).
5. Waianae dike-impounded and basal water bodies (Waianae area).

The Schofield high-level water body is tributary to the southern and the north-central areas. Therefore, the Schofield high-level water body is included in the simulation of both the southern and the north-central areas.

The ground-water flow model selected for aquifer evaluation is AQUIFEM as described by Pinder and Voss (1979), and modified by Voss (1984a) as AQUIFEM-SALT. The modified code simulates an aquifer containing a freshwater body that freely floats on seawater. The flow model will be used to simulate each of the five areas outlined above. The southeast area was the first area selected for simulation. This area comprises the leeward (south) side of Oahu's eastern end, and the aquifer consists principally of thin-bedded, gently dipping and highly permeable lava flows from the Koolau Volcano. The aquifer is bounded on the north and west by near-vertical volcanic dikes of low permeability, and on the south and east by the Pacific Ocean and Koko volcanics (fig. 127).

Recharge to the southeast basal water body is greatest near the crest of the Koolau Range, and water flows principally from north to south, generally along the direction of the lava flows. Natural discharge from the southeast basal water body occurs as diffuse leakage upward through the coastal sediments along the southern coast.

A northeast-trending dike zone internally divides the southeast basal water body into the Waialae aquifer on the west and the Wailupe-Hawaii-Kai aquifer on the east (fig. 127). Recharge to these aquifers of 6 and 9 Mgal/d, respectively, has created 2 freshwater lenses that "float" on underlying seawater. In the Waialae aquifer area, rainfall is greater than in the neighboring Wailupe-Hawaii-Kai aquifer area, and the poorly permeable coastal sediments are thicker, thus impeding the coastal discharge of ground water. The greater rainfall and thick low permeability coastal sediments cause freshwater heads in the Waialae aquifer to be 9 to 15 feet above sea level. Lesser rainfall and thinner low permeability coastal sediments allowing easier discharge to the sea result in freshwater heads of 1 to 5 feet above sea level in the Wailupe-Hawaii-Kai aquifer. The Koko volcanics (fig. 127) are posterosional volcanics that lie seaward of the

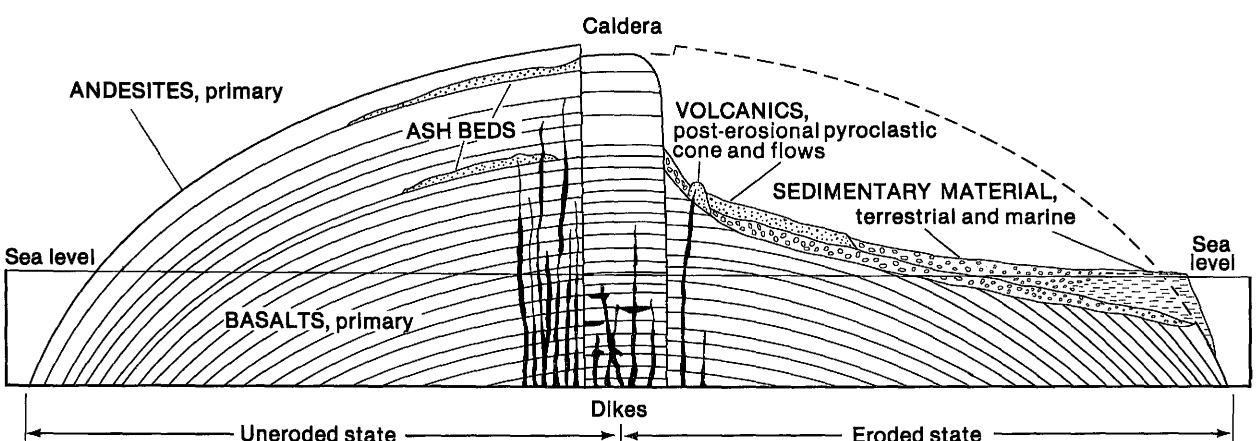

Figure 124.—Geologic structure of an idealized Hawaiian volcano dome (from Cox, 1954).

former shoreline, and consist of tuff and recent lavas underlain by coralline material. Ground water in the Koko volcanics is probably not hydraulically connected with the southeast basal water body and the water in the volcanics is not potable (Takasaki and Mink, 1982). Therefore, the Koko volcanics are not included in the simulation of the southeast area.

Ground-water flow was simulated by using a finite-element mesh constructed for the discretization of the aquifer. Average annual recharge was determined for each element by use of a water balance approach (Thornthwaite and Mather, 1955). The model simulates head changes only in the freshwater flow field. The base of the freshwater aquifer is either a confining unit underlying the aquifer or the freshwater-saltwater interface, depending on the position of the interface during the time of interest. If the simulated freshwater-saltwater interface is below the confining unit underlying the aquifer, then the aquifer base is the confining unit; otherwise, the aquifer-base boundary is the freshwater-saltwater interface. The position of the freshwater-saltwater interface is determined by assuming static equilibrium between the freshwater and saltwater. Data from several wells provided a general distribution of heads. Because most of the historic head data were obtained after development began in 1899, the predevelopment heads were estimated by calibrating the model by matching the simulated head to observed head data after development and then "turning off" the simulated pumpage and allowing the simulation to continue until the heads recovered to an equilibrium condition. The simulated and observed head data match closely, as shown in table 4. The simulated predevelopment potentiometric surface for the southeast basal water body is shown in figure 128.

Because of the assumptions used in the model—namely, a sharp interface between the freshwater and saltwater, no vertical movement, and a constant saltwater head in the aquifer—this model cannot be used to analyze the movement of a freshwater lens. However, it is appropriate for long-term analysis of the aquifer system regionally.

The model selected to evaluate the movement of a freshwater lens is SUTRA (Voss, 1984b). SUTRA is a finite-element simulation model for saturated-unsaturated fluid density-dependent

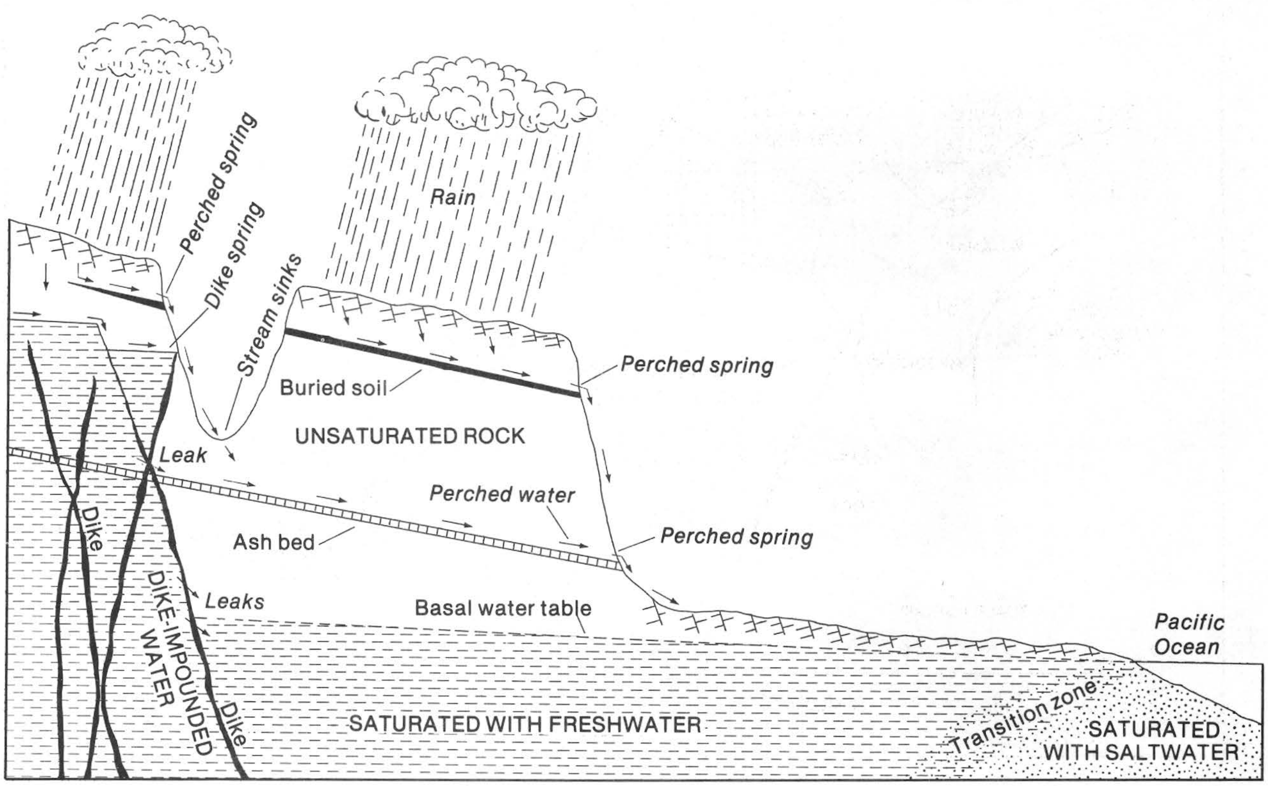

Figure 125.—Occurrence of ground water in the Oahu regional aquifer system, Hawaii (modified from Stearns and Macdonald, 1946).

ground-water flow with energy or solute transport. The principal objective of using SUTRA is to gain an understanding of the dynamics of the freshwater and saltwater movement. Of particular interest is the size and movement of the transition zone between freshwater and saltwater in response to natural and man-induced stresses on the aquifer system.

The most heavily developed aquifer in Hawaii is the Honolulu-Pearl Harbor basal water body. This aquifer consists of thin-bedded overlapping Koolau lava flows, with generally higher horizontal permeability. A thick freshwater lens has been formed because ground-water discharge to the ocean is inhibited by a thick wedge of poorly permeable coastal sediments, locally termed "caprock." Along the base of the lens, the freshwater and saltwater are separated by a transition zone of variable thickness (fig. 129).

Natural discharge from the Honolulu-Pearl Harbor basal water body is by upward leakage through the caprock and through springs that

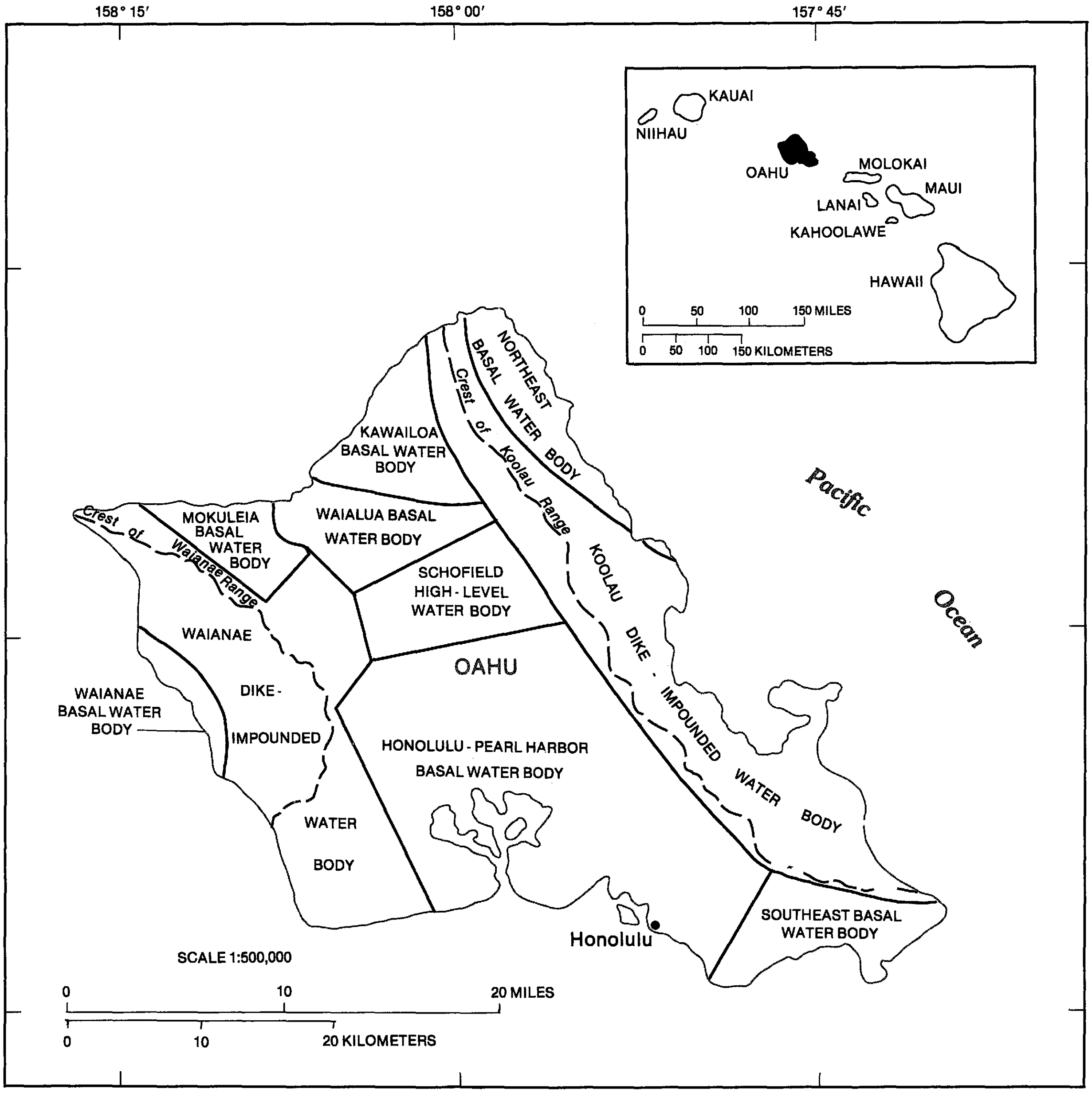

Figure 126.—Ten identified aquifers underlying the Island of Oahu, Hawaii (modified from Dale and Takasaki, 1976).

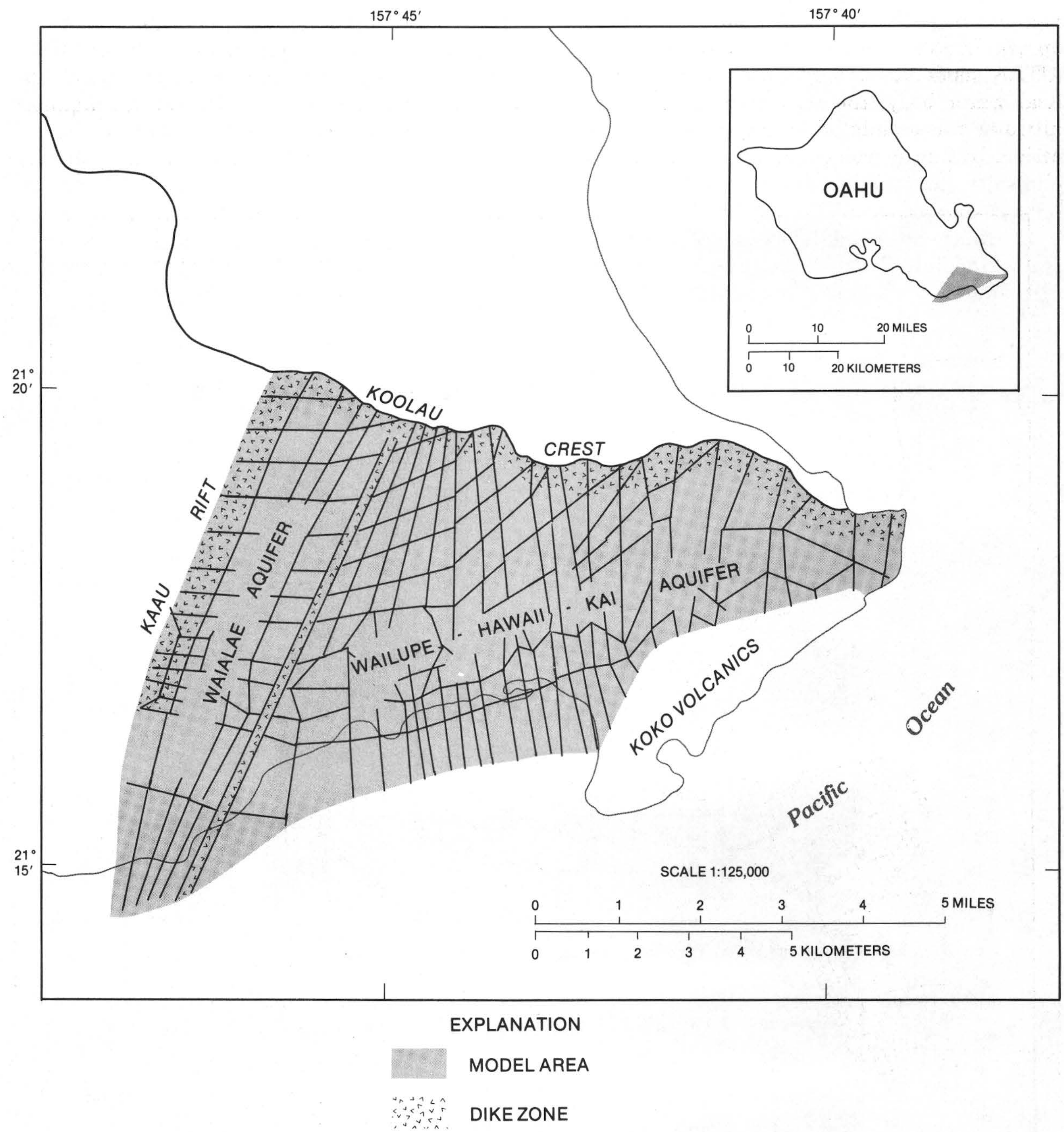

Figure 127.—Finite-element mesh for the southeast basal water body, Oahu, Hawaii.

Table 4.—Simulated and observed heads in southeast basal water body, Oahu, Hawaii, in feet above sea level

Node number	41	45	62	76	78	144	163	206	237	277	316	317
Observed head	9.6	11.5	10.6	8.5	9.5	2.7	3.7	2.2	3.7	0.81	1.3	2.2
Simulated head	9.7	11.2	10.4	8.2	10.2	2.8	3.8	2.4	3.8	0.40	1.4	2.4

originate near the contact between basalt and caprock. In order to assess the applicability of the SUTRA model to the Honolulu-Pearl Harbor basal water body, the flow of freshwater and saltwater was simulated for a generalized cross section. Initial hydrologic parameter values for the Honolulu-Pearl Harbor basal water body were those determined by other investigators. A finite-element mesh of rectangular elements of variable size was constructed to represent the aquifer system (fig. 130). Constant freshwater recharge was put into nodes at the top part of the upstream

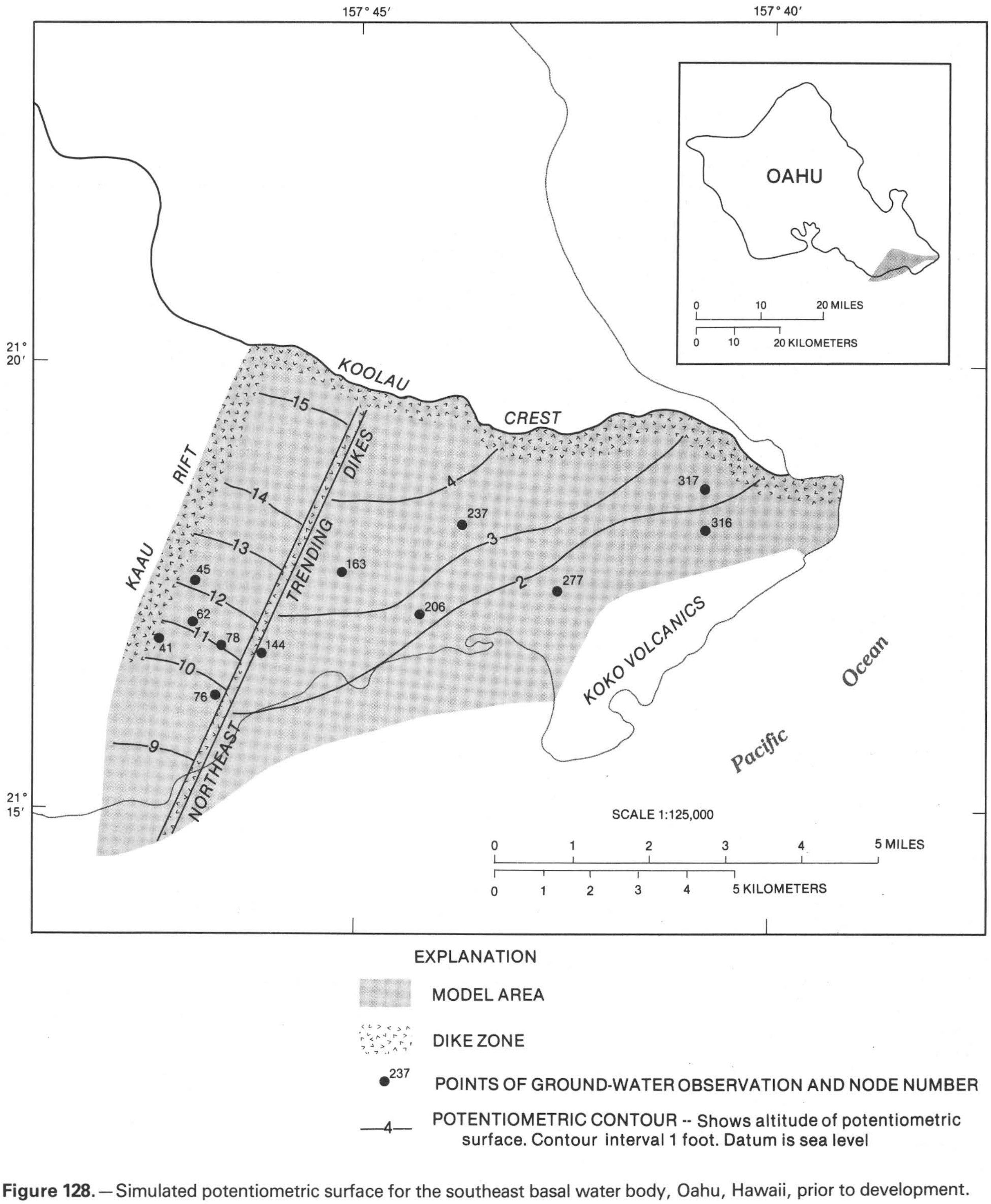

Figure 128.—Simulated potentiometric surface for the southeast basal water body, Oahu, Hawaii, prior to development.

boundary (constant flux) while the bottom part of the same upstream boundary was assumed to be a no-flow boundary. The downstream seawater boundary is a specified-pressure boundary which is equivalent to the depth of the sea at the respective position. A regional ground-water discharge of 20 Mgal/d per mile of coastline was used for the steady-state simulation. The simulated contours of seawater percentage for the steady-state, and transient conditions, are shown in figure 131 which seem to be consistent with the observed data.

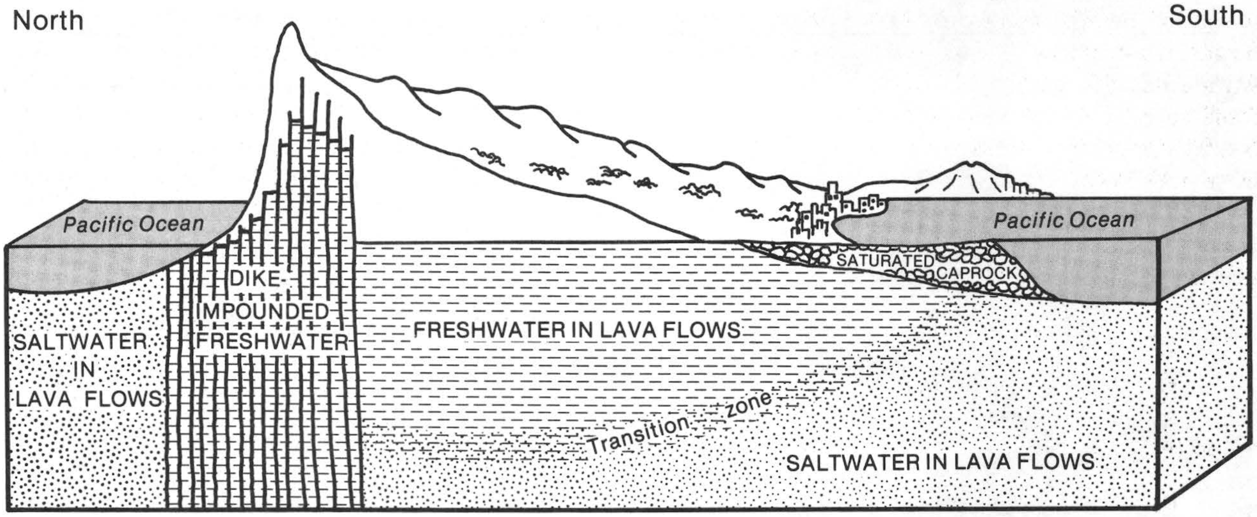

Figure 129.—Generalized section of Honolulu-Pearl Harbor basal water body, Oahu, Hawaii.

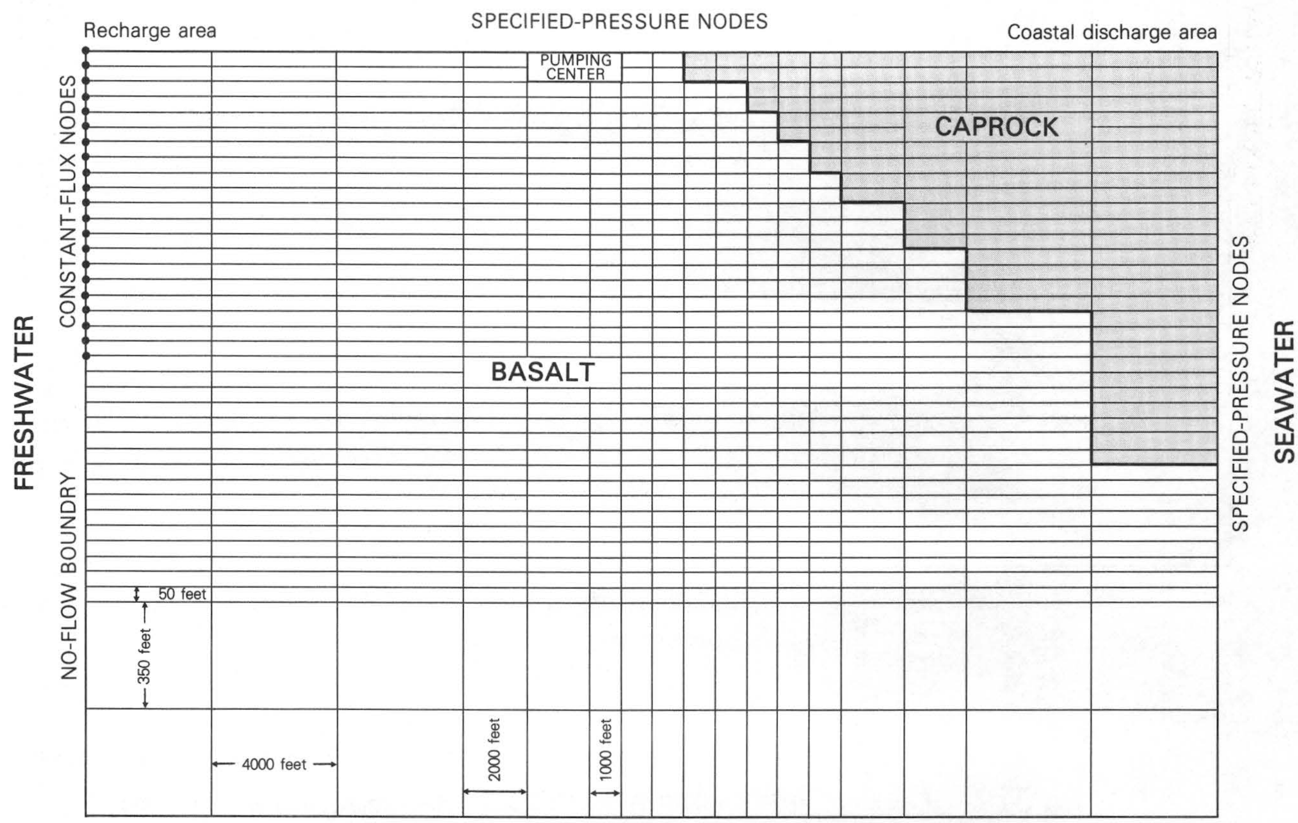

Figure 130.—Finite-element mesh for cross sectional simulation of the Honolulu-Pearl Harbor basal water body, Oahu, Hawaii.

During the transient simulation, a cyclic pumping stress was applied to the steady-state condition. Pumping was assumed to withdraw 150 percent of the estimated regional ground-water discharge (30 Mgal/d per mile of coastline) for 6 months, followed by 6 months of recovery. This cyclic pumping schedule roughly corresponds to that of the agricultural industry water use (the largest water users) in southern Oahu. During the simulated 65 years of cyclic pumping, the freshwater lens shrank considerably, and the freshwater-saltwater transition zone moved landward and upward to a significant degree (fig. 131).

The SUTRA finite-element model was successfully used to simulate the flow of fresh and saline ground water. The initial phase of testing for suitability of the model is completed. Extensive use of the model is underway to better understand the interrelation of the aquifer parameters.

The first phase of using SUTRA model did not fully utilize the transport aspect of the model, except to demonstrate that the general nature of a freshwater lens and the freshwater-saltwater transition zone could be reproduced. The next phase of using SUTRA will be to investigate the nature of the freshwater-saltwater transition zone and its movement in relation to: (1) boundary conditions, (2) long- and short-term stresses, (3) values of dispersivity, (4) sea-level changes, (5) movement of sea-water, and (6) other aquifer characteristics.

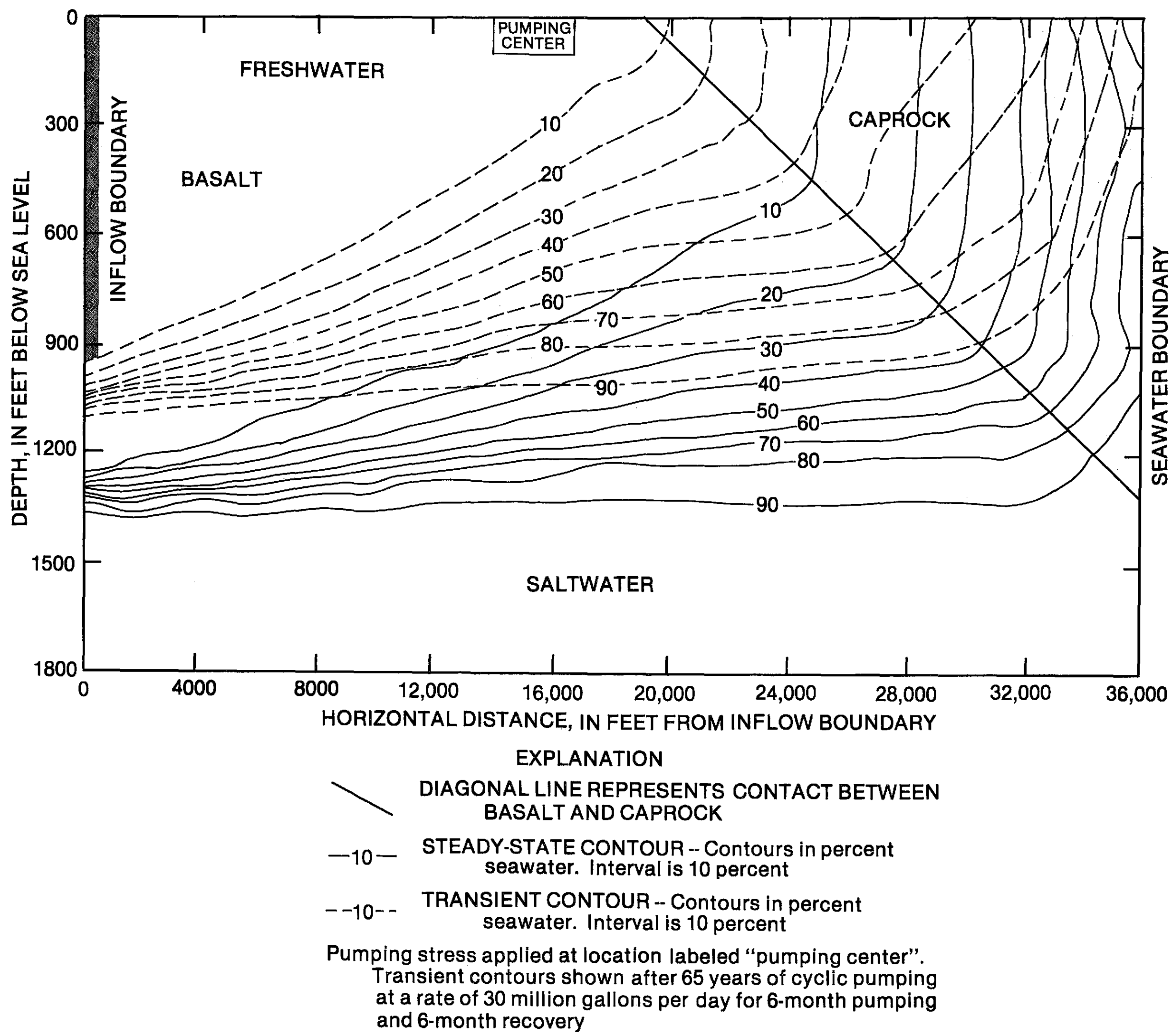

Figure 131.—Simulated contours of seawater percentage for steady-state and transient conditions, the Honolulu-Pearl Harbor basal water body, Oahu, Hawaii.

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SOUTHEASTERN COASTAL PLAIN REGIONAL AQUIFER-SYSTEM STUDY

By Robert L. Wait, Robert A. Renken, René A. Barker, Roger W. Lee and Virginia Stricker, Atlanta, Georgia

INTRODUCTION

Clastic sediments of Cretaceous and Tertiary age in South Carolina, Georgia, Alabama, Mississippi, and adjacent areas of northern Florida and southwestern North Carolina comprise a major aquifer system that underlies an area of about 130,000 mi^2 and is informally called the Southeastern Coastal Plain aquifer system. This aquifer system extends from the southwestern flank of the Cape Fear arch in North Carolina into the Mississippi embayment in northern Mississippi (fig. 132). The Southeastern Coastal Plain aquifer system is located between three adjacent regional aquifer systems: the Northern Atlantic Coastal Plain to the northeast; the Floridan aquifer system to the south and southeast; and the Gulf Coastal Plain to the west.

No previous hydrologic studies have considered the Southeastern Coastal Plain aquifer system as a single system. The solutions to problems such as multicounty declines in water levels or the degradation of ground-water supplies due to saltwater encroachment need to be examined regionally. One of the merits of the regional investigation is that a comprehensive view is obtained of the whole system, rather than only a small part of it. The Southeastern Coastal Plain regional aquifer system study was started in 1979 and is scheduled for completion in 1986.

For the most part, this study is based on information from previous studies. However, some additional data were collected to fill major gaps in information. For example, three test wells were drilled between 1980 and 1983 (one in western Alabama; one each in central and eastern South Carolina) to fill voids in the data. In addition, a four-State mass measurement of water levels was made in 1982 to evaluate the decline of water levels in different hydrogeologic units. Water samples were collected from 105 wells over the four-State area and were filtered and measured for pH, conductivity, temperature, trace metals, stable and radioactive isotopes, nutrients, and dissolved gases.

The clastic sediments that comprise the Southeastern Coastal Plain regional aquifer system have been grouped into seven major hydrogeologic units; some of these are highly interconnected with the interfingering and partially overlying carbonate rocks of the Floridan aquifer system. The composition, texture, and bedding character of the major hydrogeologic units vary considerably from place to place. Sand aquifers of this system are massive to thinly bedded, fine to coarse grained, quartzose, and feldspathic, with minor amounts of glauconitic sand, sandstone, gravel, and occasional limestone beds. Chalk, clay, shale, and mudstone form the confining units that separate the major aquifers.

Rainfall ranges from 44 to 64 inches in the study area. Most of the rainfall that enters this clastic system is discharged to nearby streams and rivers. In outcrop areas, the predominant flow is downward or upward near streams. However, in middip and downdip areas where the major aquifers are confined by thick clay, chalk, and shale, the predominant flow direction is horizontal.

The average hydrologic conditions in the study area can be summarized as follows:

- precipitation is approximately 50 in./yr,
- overland runoff is approximately 7 in./yr,
- evapotranspiration is approximately 35 in./yr, and
- recharge to the aquifer system is approximately 8 in./yr.

Most of this recharge eventually discharges into streams or rivers as base flow through shallow, local scale aquifers; however, a small amount, about 1 in./yr, recharges downward into the deeper aquifers. This deep recharge eventually discharges either into streams or rivers as base flow or discharges upward through confining units into overlying aquifers. This conceptualized recharge pattern has been demonstrated to be reasonable by preliminary simulations as discussed later.

Five geochemical zones have been identified for ground water flowing from recharge areas into the deeper parts of the system. The quality of the ground water has been adversely affected locally by heavy pumping which has caused an increase in concentrations of dissolved solids in several areas. Similarly, high concentrations of dissolved solids near major streams in central Alabama may be associated with potentiometric lows caused by natural ground-water discharges there.

REGIONAL GEOLOGIC SETTING

Coastal Plain sediments in the southeastern United States form a thick wedge of unconsolidated to poorly consolidated clastic and carbonate rocks of Jurassic to Holocene age. They are underlain in most places by relatively impervious metamorphic, sedimentary, and igneous rocks of Paleozoic and early Mesozoic age that are, in part, a southeastern extension of the Piedmont Province. In other places, they are underlain by sedimentary rocks of Paleozoic age which are a southwestern extension of the Appalachian Mountains. These rocks form the base of the Atlantic Coastal Plain.

Except where they are covered by younger strata, the aquifers and confining units that comprise the Southeastern Coastal Plain regional aquifer system crop out in adjacent bands from Mississippi to South Carolina (fig. 133). These units extend seaward from the inner Coastal Plain margin to the Atlantic Ocean, Gulf of Mexico, or Peninsular Florida.

The Coastal Plain sediments of the southeastern United States are the product of cyclical advance and retreat of ancient seas over the Paleozoic-Mesozoic basement complex during Jurassic to Holocene time and were deposited under marine, marginal marine, and nonmarine conditions. The fluctuating depositional conditions, resulting from an ever-shifting shoreline, caused the lithology—and, therefore, the hydraulic conductivity—of these sediments to differ greatly from place to place. The variation of hydraulic conductivity significantly affects the occurrence and flow of ground water within the Southeastern Coastal Plain. Consequently, the texture, bedding character, and composition of the aquifers and confining units greatly control the regional ground-water movement.

Clastic Tertiary and Cretaceous aquifers are the focus of this investigation; however, in Florida, southern Georgia, southern Alabama, and southwest South Carolina, the Southeastern Coastal Plain aquifer system is partly overlain by, and is hydraulically connected to, highly permeable carbonate rocks of the Floridan aquifer system. The limestone units that comprise the Floridan aquifer system generally grade into or interfinger with the clastic rocks that compose the Southeastern Coastal Plain regional aquifer system. Low-permeability rocks of Oligocene to Pliocene age make up the upper confining unit of the Floridan aquifer system (fig. 133) and are, in turn, overlain by a surficial aquifer consisting of unconsolidated sand and gravel of Pliocene to Holocene age. Aquifers of the Southeastern Coastal Plain regional aquifer system consist of quartzose, feldspathic, locally glauconitic, coarse to fine sand, with minor amounts of gravel and limestone; confining units that bound and separate these aquifers are composed of clay, mudstone, shale, marl, and chalk.

HYDROGEOLOGIC FRAMEWORK

The Southeastern Coastal Plain regional aquifer system consists of interbedded strata having wide areal distribution and containing numerous aquifers and confining units. The aquifers are hydraulically connected in varying degrees and can be treated as a single system. The arrangement, distribution, and physical attributes of individual aquifers and confining units that serve as conduits for the regional ground-water flow constitute the hydrogeologic framework.

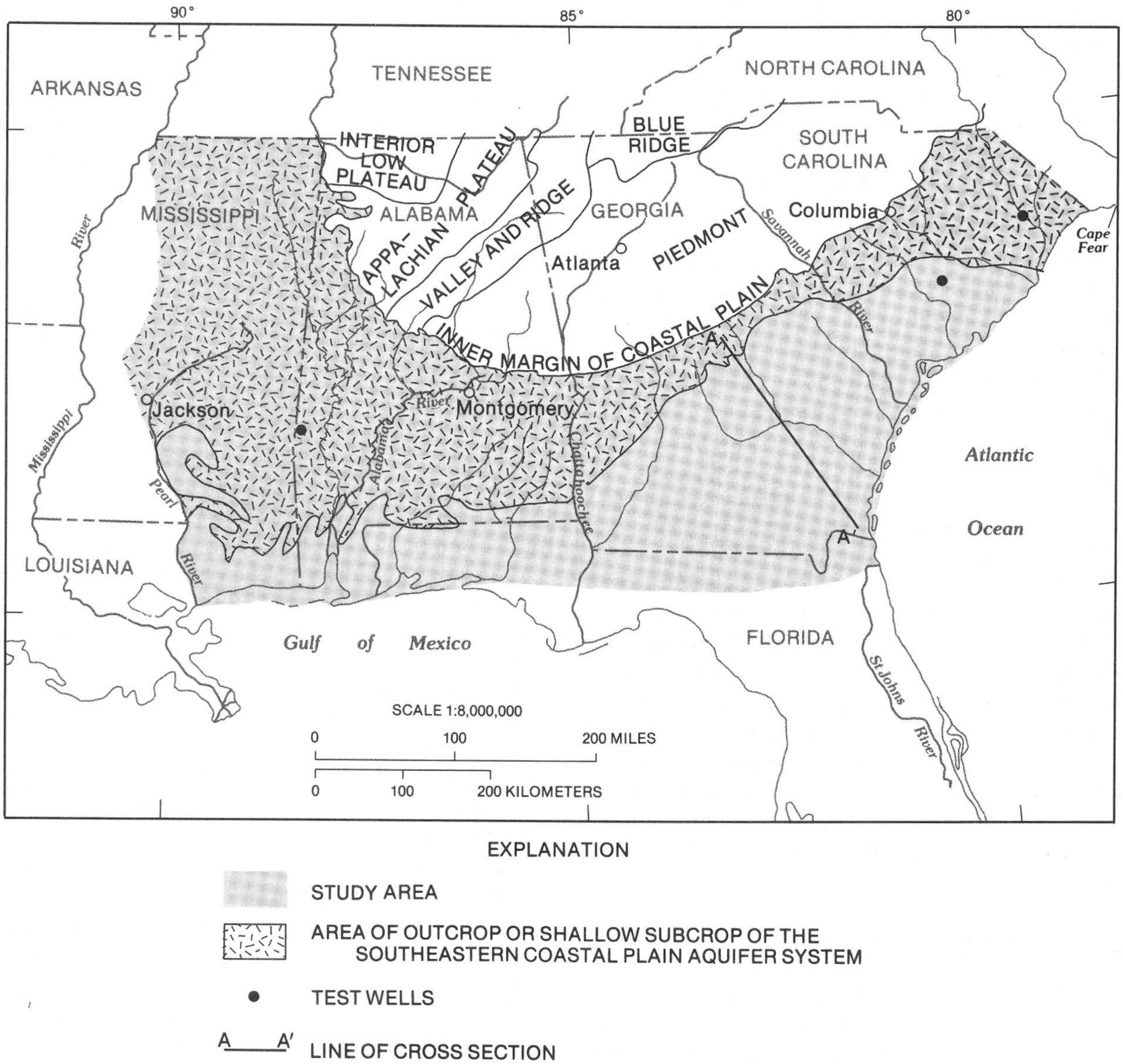


Figure 132.—Study area of the Southeastern Coastal Plain regional aquifer system.

The hydrogeologic framework established for this study is designed to evaluate permeability, and is therefore unlike sedimentary basin studies that usually center on the study of the stratigraphic characteristics of the rocks that have paleoenvironmental significance. The physical limits of an aquifer may locally parallel boundaries of a stratigraphic unit defined by lithology, time, or mode of deposition; however, the limits of a regionally-extensive body of hydraulically interconnected permeable rock do not always fall within such boundaries. Rocks that are stratigraphically equivalent may be an aquifer in one location and a confining unit in another location.

Consequently, the hydrogeologic units delineated herein cannot be described adequately by existing nomenclatural schemes, in part due to their regional extent, and in part due to their poor correspondence with physical boundaries of rock- and time-stratigraphic units. Therefore, the hydrogeologic units have been given a provisional alphanumeric designation; regional aquifers are designated by the letter A, and regional confining units by the letter C. A sequential number is assigned to each hydrogeologic unit to indicate its relative position in the hydrogeologic section. Figure 134 shows the relation between rock-stratigraphic units and the regional hydrogeologic units.

Strata that together comprise a regional hydrogeologic unit were combined according to: (1) their degree of interconnection, (2) the uniformity and continuity of the strata, (3) their overall distribution, (4) their ability to be mapped over a wide area with definable physical boundaries, and (5) hydraulic heads in the strata. Many of the local water-yielding units within the regional aquifer system have been largely treated as discrete aquifers or confining units at the State

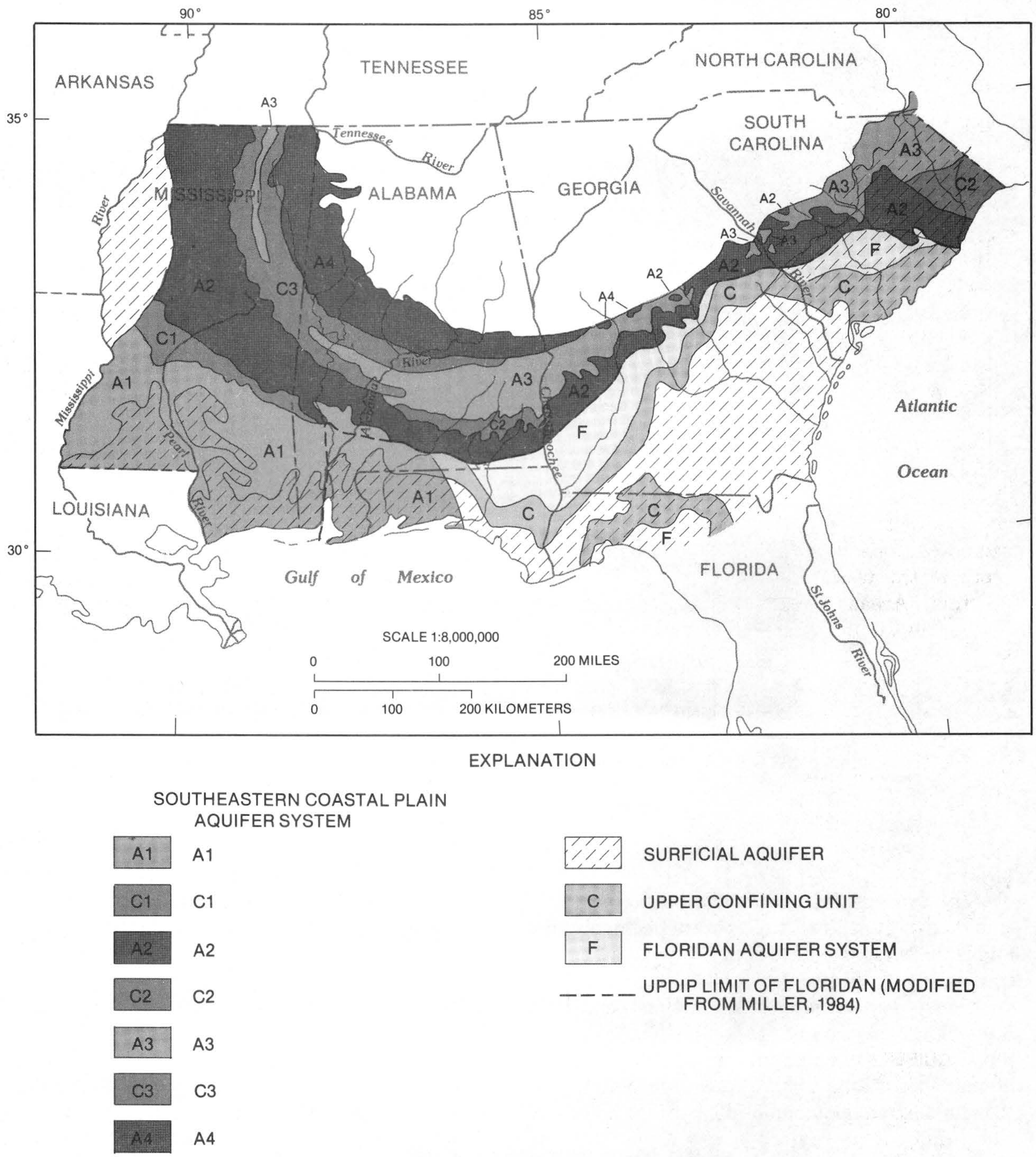


Figure 133.—Location and outcrop areas of major hydrogeologic units of the Southeastern Coastal Plain regional aquifer system.

or county level. However, several of these discrete aquifers and confining units are grouped into regional aquifers or confining units if they can be shown to behave regionally as a single hydrologic unit.

The Southeastern Coastal Plain sediments are divided into 7 hydrogeologic units (Renken, 1984); 4 regional aquifer units, A1 to A4, numbered downward, separated by 3 regional confining units, C1 to C3 (fig. 134). The uppermost

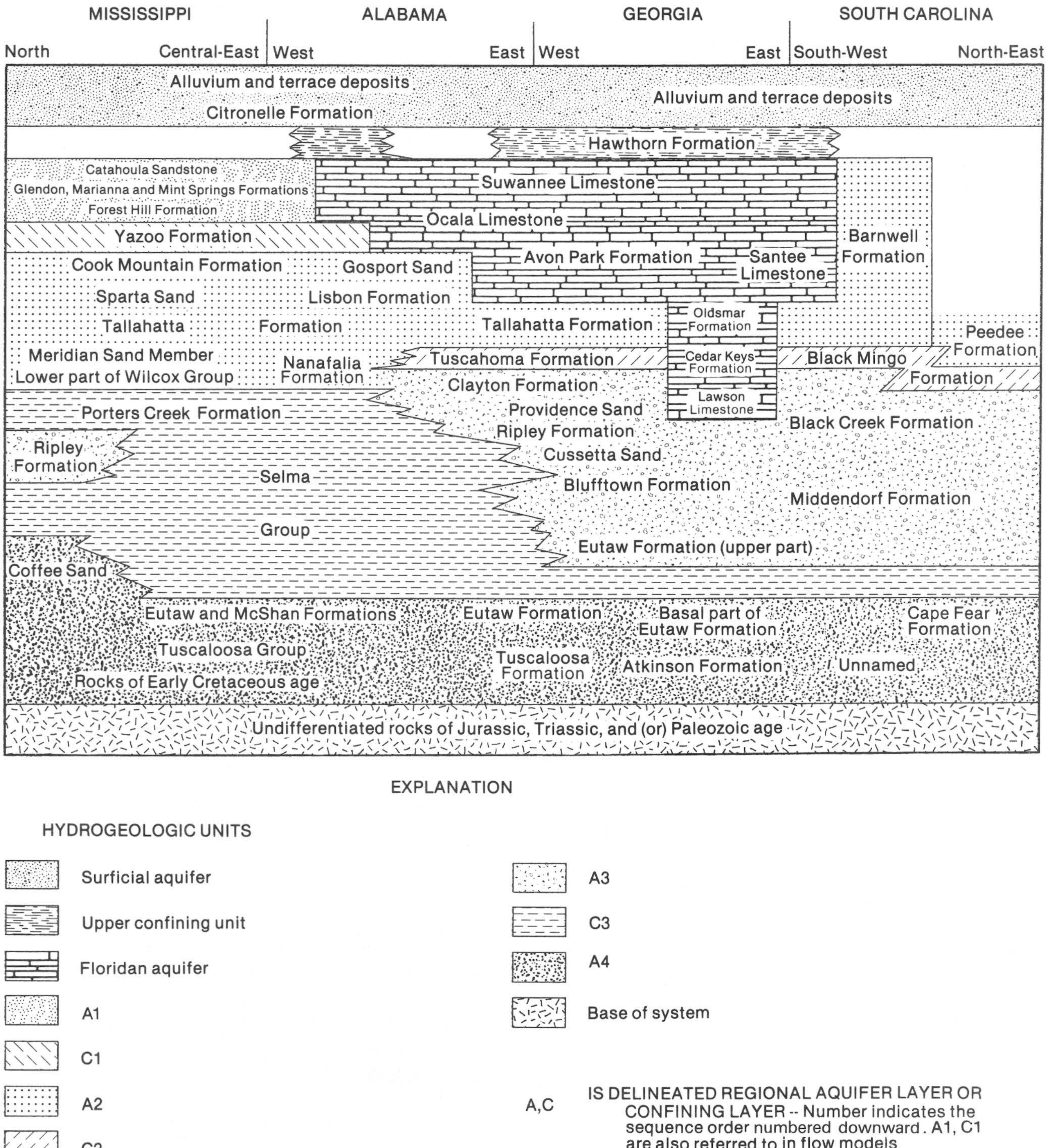


Figure 134.—Schematic classification of rock-stratigraphic and regional hydrogeologic units of the Southeastern Coastal Plain regional aquifer system.

unit, A1, extends across southwest Alabama and southern Mississippi into Louisiana, where it becomes part of the Gulf Coastal Plain regional aquifer system. Comprised largely of Oligiocene and Miocene clastics and minor limestone beds, it is underlain in southern Mississippi by calcareous marine clay (Yazoo Formation, regional confining unit C1) that separates the A1 aquifer from underlying permeable strata. In southwest Alabama, the A1 unit is interbedded and interconnected with the upper part of the Floridan aquifer system. Although aquifer unit A1 has been mapped regionally, it is not simulated as an active layer in the regional ground-water flow model; rather, it is referred to as a source-sink layer.

The A2 regional aquifer consists of a thick sequence of sand, sandstone, gravel, and minor limestone beds, all of Paleocene to late Eocene age. Locally, sands of Late Cretaceous age in South Carolina are included as part of the A2 aquifer. The sediments that comprise the A2 aquifer were deposited largely under marine conditions except in Mississippi where they grade into a thick fluvial sequence. The A2 aquifer is largely quartzose sand (but contains glauconitic and feldspathic sand in places), fine to coarse grained, massive to thinly bedded, and is locally stratified by interbedded to inter-laminated clay, shale, marl, and mudstone.

The outcrop area of the A2 aquifer is shown in figure 133. This aquifer connects to the west with the Gulf Coastal Plain regional aquifer system that can be mapped as far west as Texas. The A2 aquifer is the uppermost regionally-extensive clastic aquifer except in southwestern Alabama and southern Mississippi where younger sand aquifers are present. The highly permeable Floridan aquifer system partly overlies, and is hydraulically interconnected with, the less permeable A2 aquifer in Georgia, southwest South Carolina, and southern Alabama. In most places, the Floridan aquifer system and the underlying A2 aquifer are not separated by an intervening confining unit. The boundary between the two aquifer systems represents a facies transition separating predominantly carbonate rocks from underlying predominantly clastic rocks. The Floridan aquifer system has permeability values that are generally one or more orders of magnitude higher than those of the A2 aquifer. The top of the A2 aquifer is coincident with the base of the Floridan aquifer system (Miller, 1982), where the two systems are juxtaposed (fig. 135). The A2 aquifer in most places grades seaward from sand, sandstone, gravel, and minor limestone beds into low-permeability rocks that mark the downdip limit of the aquifer. In southeastern Georgia, however, the A2 aquifer grades downdip into permeable limestone that is part of the Floridan aquifer system (fig. 135).

The permeable clastic rocks of the A2 aquifer are separated from underlying aquifers by two extensive, separate confining units (C2 and C3) that range from Paleocene to Late Cretaceous in age (fig. 134). Locally, in updip areas of the Coastal Plain in northeast Georgia and northwest South Carolina, no underlying confining unit is present, and the A2 aquifer is hydraulically connected with underlying massively-bedded, quartzose, nonmarine Cretaceous sand that is part of the A3 aquifer. The C2 confining unit is largely covered by overlapping rocks of the A2 aquifer (fig. 133) but crops out in eastern Alabama (Tuscahoma Formation and equivalents) and in eastern South Carolina (parts of the Black Mingo and Peedee Formations). Although the lithology of this confining unit varies, the C2 confining unit consists everywhere of low-permeability rocks that effectively separate the overlying A2 aquifer and the underlying A3 aquifer.

To the west, in central Alabama, the C2 confining bed grades laterally into fine-grained glauconitic sand with minor interbedded silty clay that is part of the A2 aquifer (fig. 134). Consequently, a deeper, regionally extensive confining unit (C3) in these places separates the A2 aquifer from the underlying permeable strata, the A3 aquifer. Where the C3 confining unit underlies the A2 aquifer, this confining unit is largely equivalent to the outcropping massive clay of the Porters Creek Formation of Paleocene age and to chalk and shale of the Late Cretaceous Selma Group.

The area of outcrop of the A3 aquifer is shown in figure 133. In western South Carolina and eastern Georgia, this aquifer crops out in a discontinuous pattern where it is covered by younger, overlapping rock. In the subsurface, the A3 aquifer extends as a continuous unit from eastern Alabama into North Carolina where it is continuous with aquifers that are part of the Northern Atlantic Coastal Plain regional aquifer system. To the west, the A3 aquifer grades by facies change laterally into calcareous shale and chalk that are part of the C3 confining unit (Selma Group, fig. 134). Locally, glauconitic quartz sand of Late Cretaceous age in northern Mississippi (Ripley Formation) is considered part of the A3 aquifer. This aquifer extends north and northwest into the northernmost reaches of the Mississippi embayment where it is part of the Gulf Coastal Plain regional aquifer system.

The A3 aquifer is mostly massive to thinly bedded, largely quartzose, fine to coarse-grained quartz sand that is glauconitic to feldspathic in places; highly permeable, sandy, glauconitic limestone of local extent (Clayton Formation) is also part of the aquifer. In downdip areas, the A3

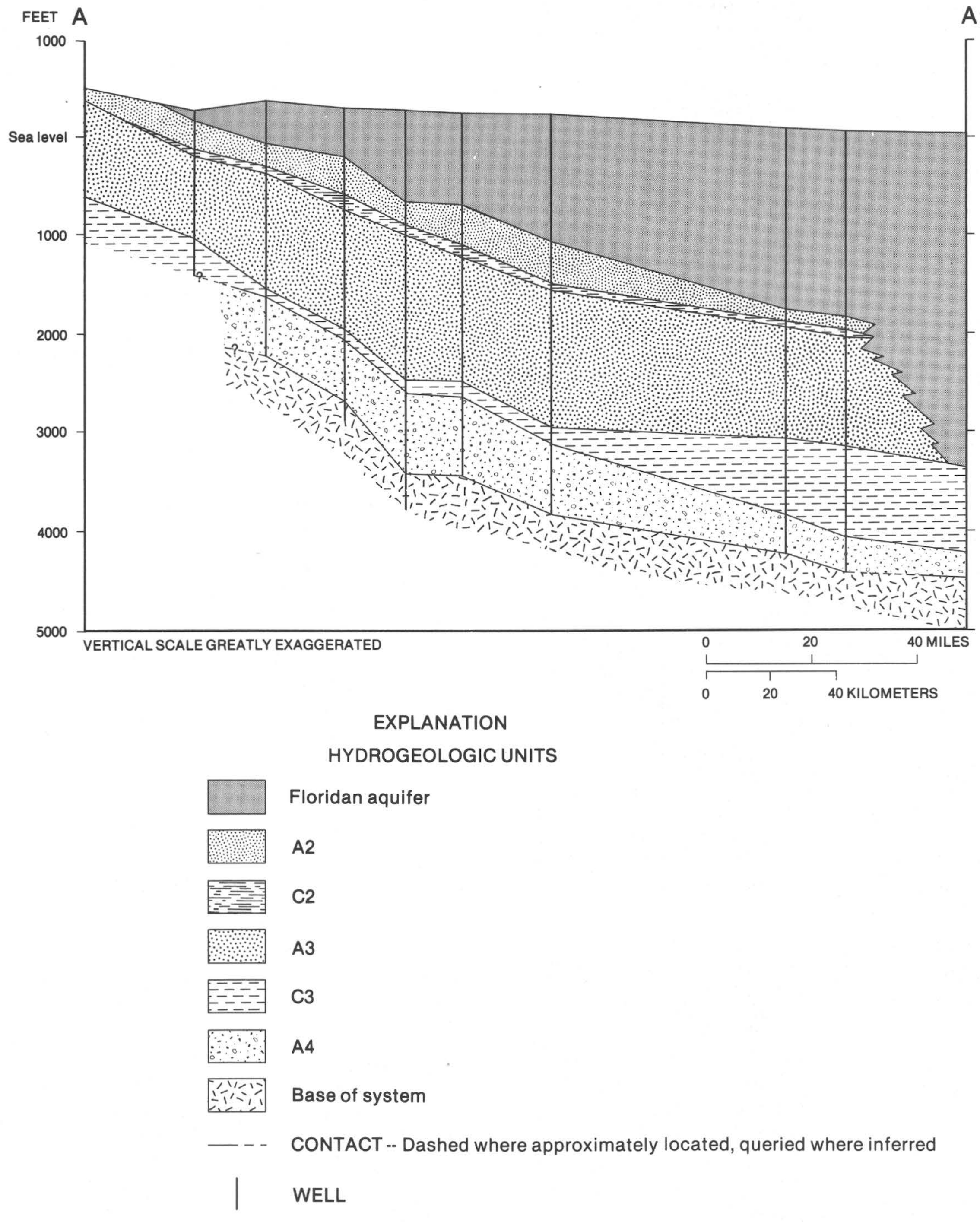


Figure 135.—Generalized hydrogeologic section in east-central Georgia (location of section shown in fig. 132).

aquifer grades into calcareous shale and chalk; locally in southeast Georgia, however, the aquifer grades into permeable limestone that is part of the Floridan aquifer system (fig. 135).

Fissile, carbonaceous to varicolored clay and shale found in the sub-surface of South Carolina and Georgia combine with outcropping and subsurface clay and chalk of the Midway (Paleocene) and Selma (Late Cretaceous) Groups of Alabama and Mississippi to form confining unit C3. This confining unit is everywhere thick and effective, separating aquifers A3 and A4.

The A4 aquifer is the most regionally extensive clastic aquifer underlying the Southeastern Coastal Plain. Correlative sands of this aquifer extend in the subsurface or outcrop from Tennessee eastward into central North Carolina. This aquifer crops out in a wide band in Mississippi (fig. 133) but the outcrop width gradually diminishes in central Georgia where the aquifer is completely covered by onlapping Late Cretaceous and Tertiary rocks. Aquifer A4 is not known to crop out in eastern Georgia or in South Carolina.

Aquifer A4 includes sediments of Early and Late Cretaceous age and is composed of medium to coarse, massively to thinly bedded, sand and sandstone mostly of fluviodeltaic origin. Locally, this aquifer contains various trace minerals ranging from limonite to glauconite; very locally, sediments of this aquifer are of marginal marine origin. Interbedded varicolored and mottled clay, shale, and siltstone form local confining units within the aquifer. A major subsurface part of this aquifer consists of sand that is part of the Tuscaloosa Group. However, four other rock-stratigraphic units are included in the upper part of this aquifer (fig. 136), thus illustrating the poor regional correspondence of a regionally extensive aquifer with previously defined rock-stratigraphic units

Deeply buried sedimentary rocks of Jurassic age that underlie Early Cretaceous rocks in parts of Alabama, Mississippi, Georgia, and Florida are not considered to be part of the regional hydrogeologic framework of the Southeastern Coastal Plain regional aquifer system because they contain water everywhere with concentrations of dissolved solids greater than 10,000 milligrams per liter (mg/L). The southernmost extent of ground water having concentrations of dissolved solids less than 10,000 mg/L in aquifer A4 is shown in figure 136. Downdip of the line representing the limit of 10,000 mg/L concentration of dissolved solids, there is considered to be little or no ground-water movement in the aquifer.

HYDROLOGY

Ground water originates as precipitation that enters the aquifer system in outcrop areas. Annual precipitation ranges between 44 to 64 in./yr over the entire study area (Cederstrom and others, 1979); however, outcropping rocks of the aquifer system that are adjacent to the inner margin of the Coastal Plain receive lesser amounts of precipitation. Only a small part of the precipitation enters the regional ground-water flow system; the majority either discharges to streams as (1) direct runoff, and (2) through shallow and intermediate aquifer systems as base flow, or is transpired by plants, or evaporates. Runoff in outcrop areas of the regional aquifer system is generally 10 to 20 in/yr, except in western Alabama where it ranges between 20 to 25 in/yr (Busby, 1966). Major rivers that are considered to have a significant impact on the regional ground-water flow system are shown in figure 137.

Aquifers extend downdip under younger sediments, and generally contain freshwater for a distance of about 100 miles or more downdip of the outcrop. Where the regional aquifer system is overlain by rocks of the Floridan aquifer system, the clastic aquifers are little used because of the abundance of water available from the Floridan. Ground-water development in the Southeastern Coastal Plain regional aquifer system is largely limited to rural use. However, in places, municipal and agricultural use has resulted in local water-level declines of as much as 100 feet from predevelopment water levels.

The general pattern of ground-water flow within the Southeastern Coastal Plain regional aquifer system is controlled by variations in aquifer permeability and the distribution of recharge and discharge. Water enters the aquifer system in upland outcrop areas where the ground-water movement is through relatively short distances and predominantly in the vertical direction; whereas ground-water flow in middip and downdip areas is relatively slow and predominantly in the horizontal plane. In places, ground water may flow parallel to the strike of the aquifers (fig. 137). The shift from predominantly vertical flow in updip areas to largely horizontal flow in middip areas (fig. 138) can be attributed to: (1) a diminishing downdip interconnection between surface drainage and the aquifers, and (2) an overall decrease in vertical hydraulic conductivity combined with an increase in thickness of the confining units in a downdip direction. Distinctions between regional and local ground-water flows and the determination of the relative importance of vertical and horizontal flows are made possible through use of simulation.

GROUND-WATER FLOW SIMULATIONS

Computer models of the Southeastern Coastal Plain regional aquifer system are being used to

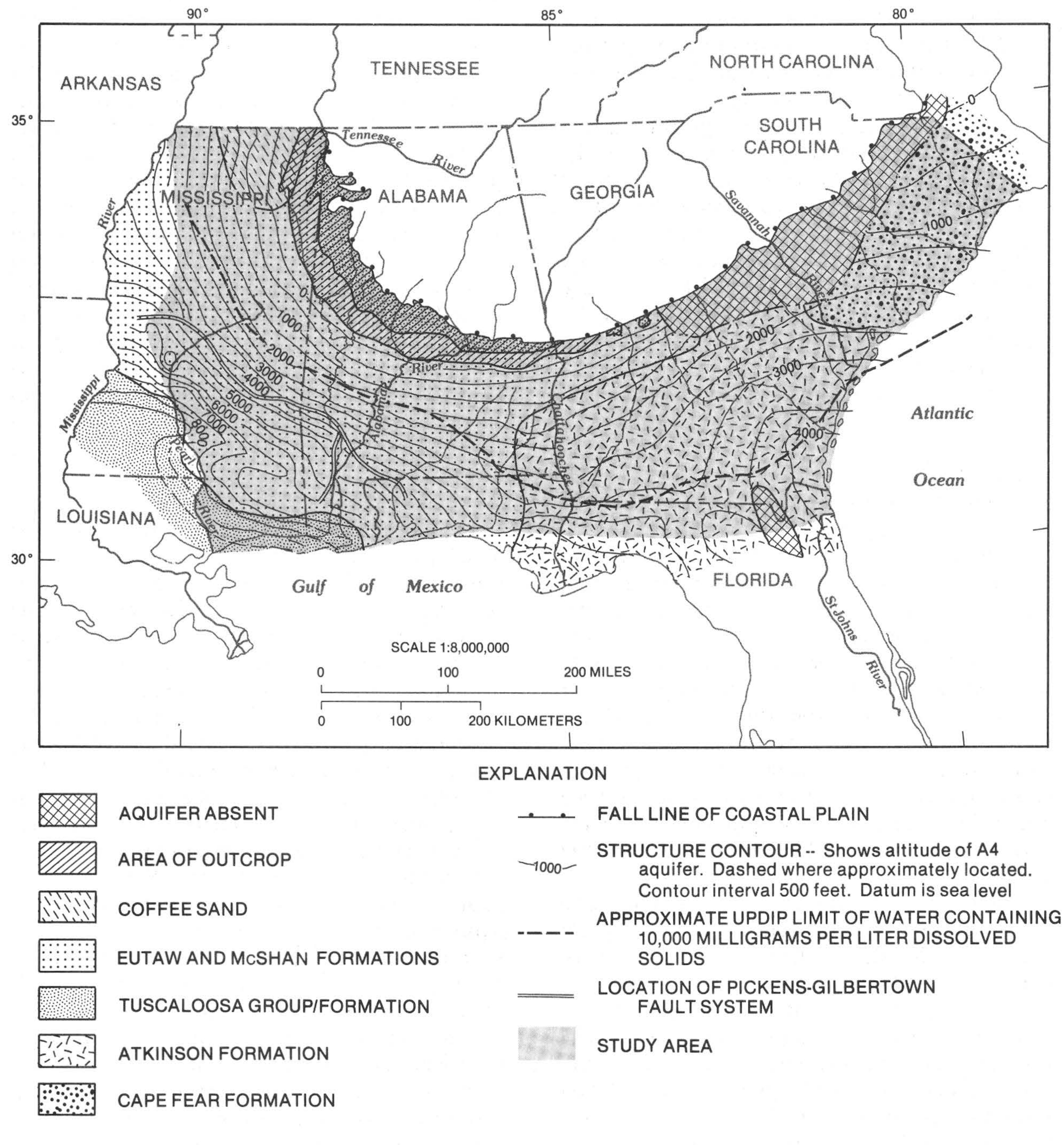
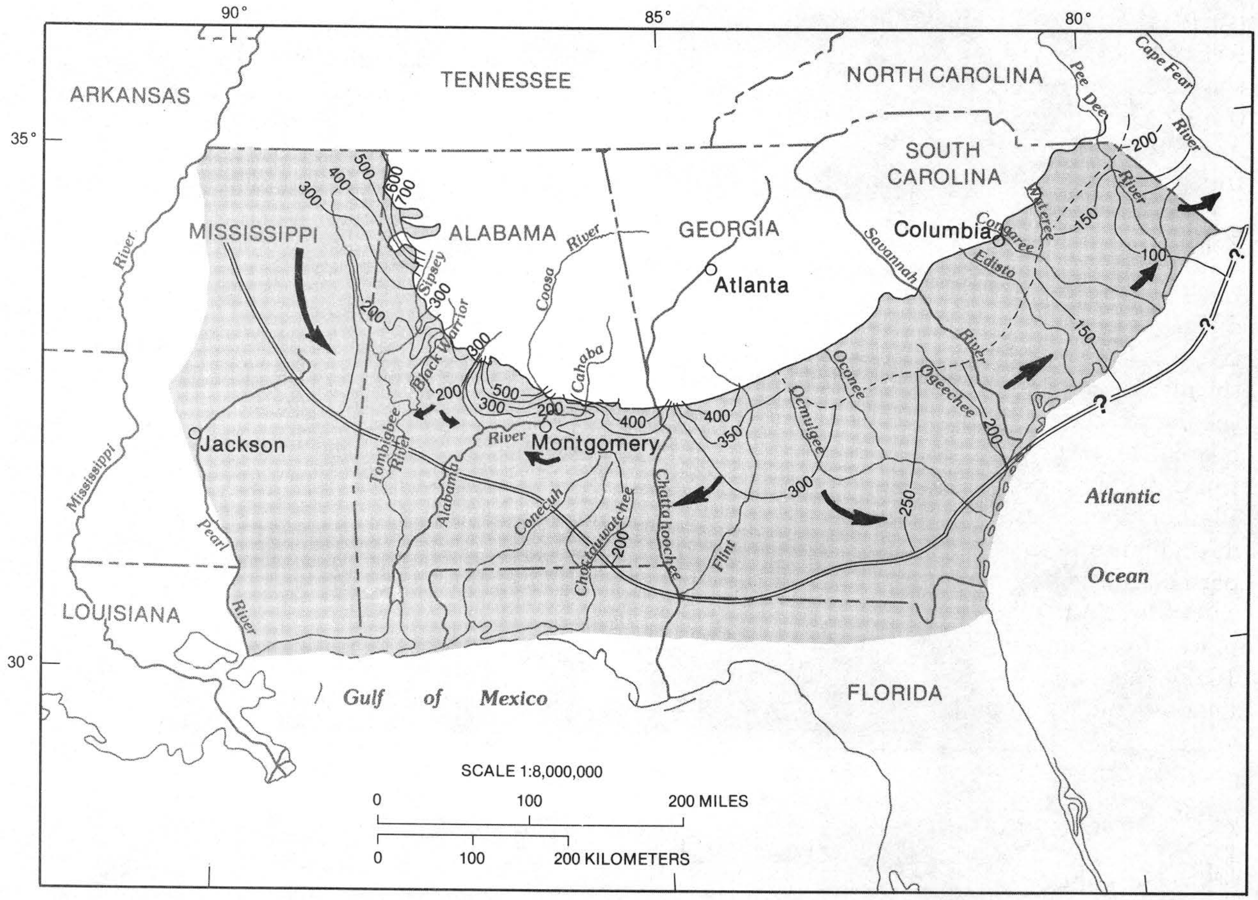


Figure 136.—Geology and configuration of the surface of the lowest aquifer (A4) and approximate location of updip limit of water containing concentrations of 10,000 milligrams per liter of dissolved solids, Southeastern Coastal Plain regional aquifer system.

help estimate predevelopment potentiometric surfaces and hydrologic properties of the aquifers and confining units, and to simulate ground-water movement. The models additionally are being used to help test the validity of the delineated regional hydrogeologic framework.

Because hydrologic conditions within the regional aquifer system were not known initially for large parts of the study area, the strategy for calibrating the regional flow model was to: (1) accept a plausible set of boundary conditions; (2) make preliminary estimates of recharge, transmissivity, leakance of confining units, and conductance of riverbeds; and (3) refine these preliminary estimates by way of trial- and-error simulations. The objective of model calibration is to simulate potentiometric surfaces and regional ground-water flow conditions consistent with the observed data and the delineated geohydrologic framework.

The models utilize a three-dimensional finite-difference code described by McDonald and Harbaugh (1984). Four subregional models that overlap are directly linked with the regional



EXPLANATION

FALL LINE OF COASTAL PLAIN

UPDIP LIMIT OF AQUIFER A4

—200— ALTITUDE OF POTENTIOMETRIC SURFACE -- Interval, in feet, is variable. Datum is sea level

GENERAL DIRECTION OF REGIONAL GROUND-WATER FLOW

SOUTHERN EXTENT OF GROUND WATER WITH LESS THAN 10,000 MILLIGRAMS PER LITER DISSOLVED SOLIDS

STUDY AREA

Figure 137.—Location of major rivers that have significant impact on the regional ground-water flow system and potentiometric surface of the lowest aquifer (A4), Southeastern Coastal Plain regional aquifer system.

model (fig. 139) and share a common data base. Although the regional model is less detailed than the subregional models, it simulates flow across the boundaries of the subregional models and calculates a water budget for the entire aquifer system.

As shown in figure 140, the regional model is comprised of three active aquifers (A2, A3, and A4) which are overlain by an aquifer assumed to have constant water levels and to act as a source or sink. In addition to the Upper Floridan aquifer (Miller, 1984), the source-sink aquifer includes the 'surficial aquifer' as shown in figure 134. To effectively simulate the interconnection between the A2 aquifer and the Lower Floridan aquifer (Miller, 1984), these aquifers are combined into a single aquifer in the simulation model as shown in figure 140.

Boundary Conditions

To properly simulate observed hydraulic conditions near the boundaries of the modeled area, the appropriate conditions must be specified for the model boundaries. Zero ground-water flow in the horizontal direction is maintained by no-flow boundaries to represent conditions along the Fall Line (updip limit of the Southeastern Coastal Plain regional aquifer system) and along the downdip limits these represent either the loss of permeability by facies change or the freshwater-saltwater interface which is defined as water in place that contains concentrations of at least 10,000 mg/L of dissolved solids. The constant-head boundary conditions atop parts of the Southeastern Coastal Plain regional aquifer system are derived from potentiometric surfaces simulated by flow models of the Floridan aquifer system and the Gulf Coastal Plain regional aquifer system. Flow across all model boundaries is monitored during simulations to ensure that realistic hydrologic conditions are being simulated.

Model Sensitivity

The steady-state model is most sensitive to changes in rates of recharge. The potentiometric surface simulated directly results from recharge into outcrop areas. Different recharge rates affect the simulated rates of aquifer-to-stream leakage by roughly the same percent that the recharge rate differs. When recharge is held constant, the model is more sensitive to variation in leakance of confining units and conductance of streambeds than it is to comparable changes in transmissivity. The model is more sensitive to changes in transmissivity in updip areas where hydraulic gradients are steep than it is to changes in transmissivity in downdip areas where hydraulic gradients are relatively flat.

The model's sensitivity was tested also to changes in the location of no-flow boundaries used to simulate the presumed effect of the downdip limits of the freshwater-saltwater interface. Changing the location of the southern no-flow boundary through a distance of about 40 miles in a southeast-to-northwest (inland) direction, produces simulated head differences that are everywhere less than 10 feet, and average less

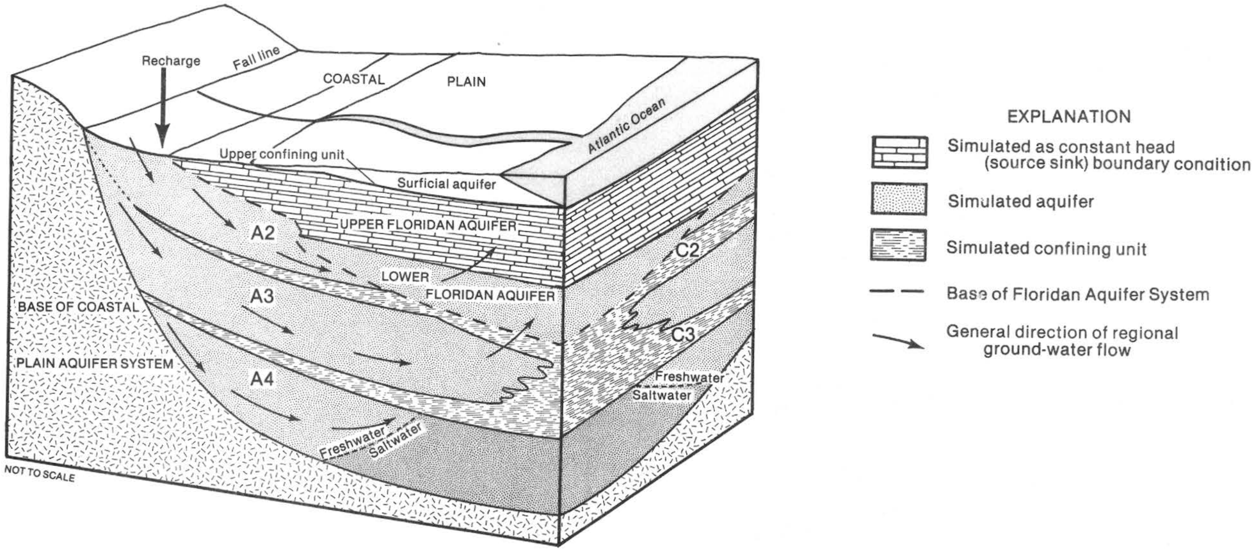


Figure 138. — Relation between hydrogeologic framework and simulated flow direction along a hypothetical dip section through Georgia.

than 5 feet, over the affected parts of the simulated aquifer system. Differences of such magnitude and distribution are considered inconsequential to the model's capability of satisfying the objectives of the study.

Preliminary Results of Simulation

Limited by the model's grid block size of 8 miles to a side, the regional flow model can simulate discharge as base flow adequately only to major streams in the Southeastern Coastal Plain. All major rivers simulated in the regional model have drainage basins exceeding 1,000 mi². To simulate recharge and discharge rates that are consistent with the model's grid scale, stream reaches within drainage basins of less than 1,000 mi² are considered to be of local or intermediate scale and are not modeled (fig. 141). Consequently, the regional model cannot simulate recharge that is discharged directly to these streams which drain the shallower aquifers.

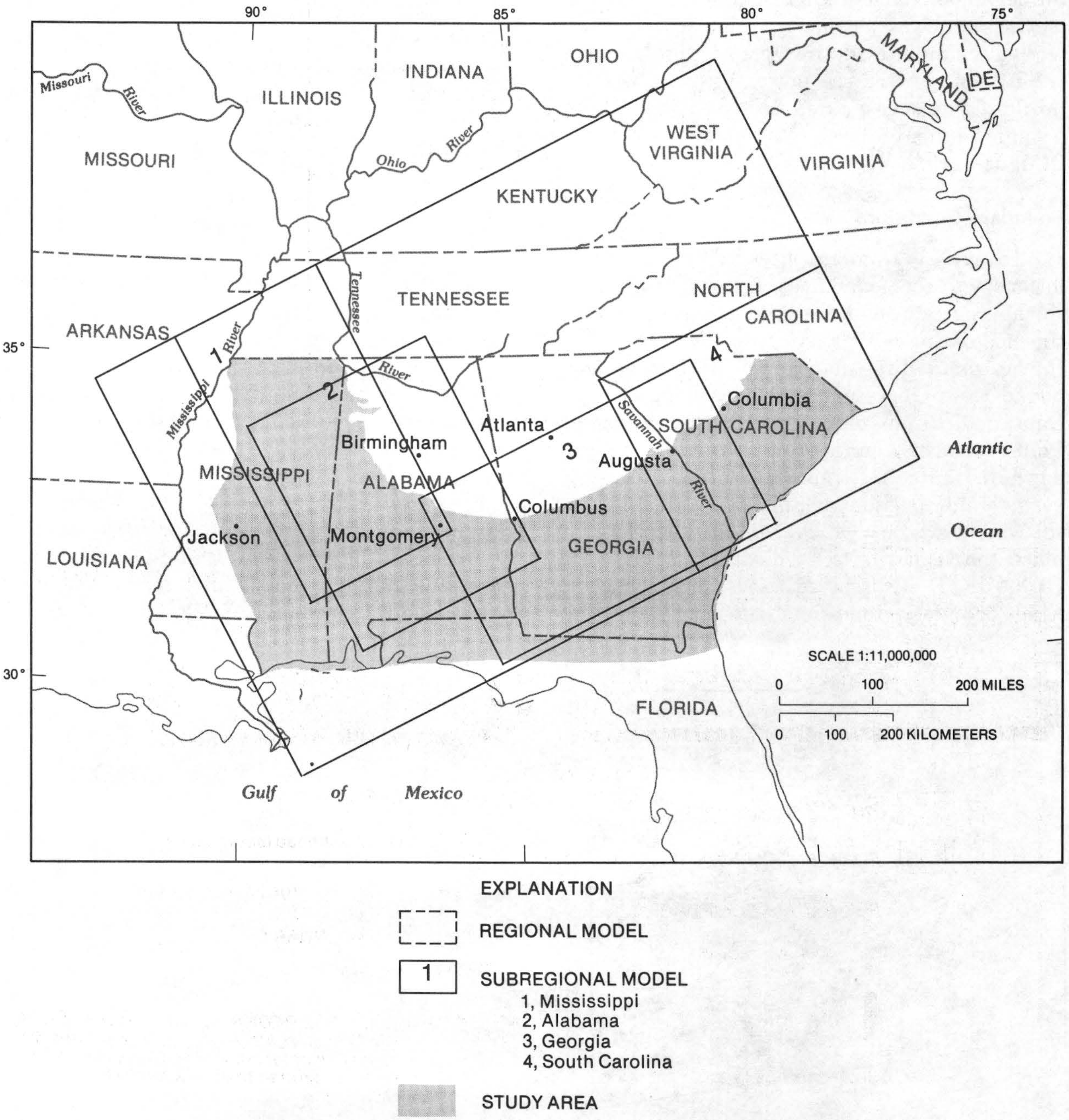


Figure 139.—Areas modeled by regional and subregional models, Southeastern Coastal Plain regional aquifer system.

Therefore, the amount of simulated recharge associated with the regional scale ground-water flow system is less than the total recharge to the entire aquifer system.

Recharge in the outcrop area can be calculated by the following equation:

$$R = P - DRO - ET \quad (1)$$

Where,

R = total recharge, in./yr,
P = precipitation, in./yr,
DRO = direct runoff (does not include base flow), in./yr,
ET = evapotranspiration, in./yr.

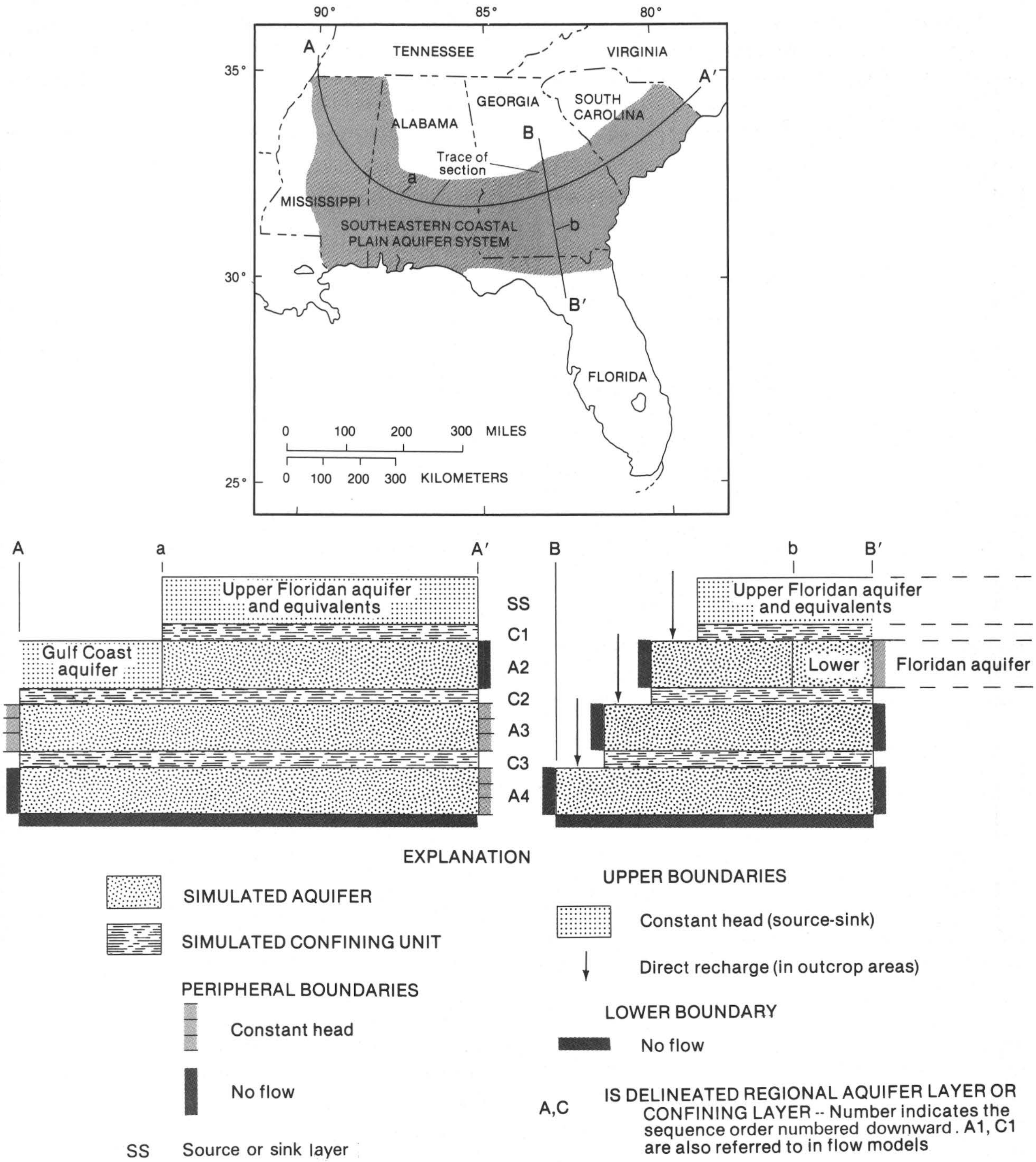


Figure 140.—Schematic section A-A′ and B-B′, showing simulated aquifers, confining units, and boundary conditions, Southeastern Coastal Plain regional aquifer system.

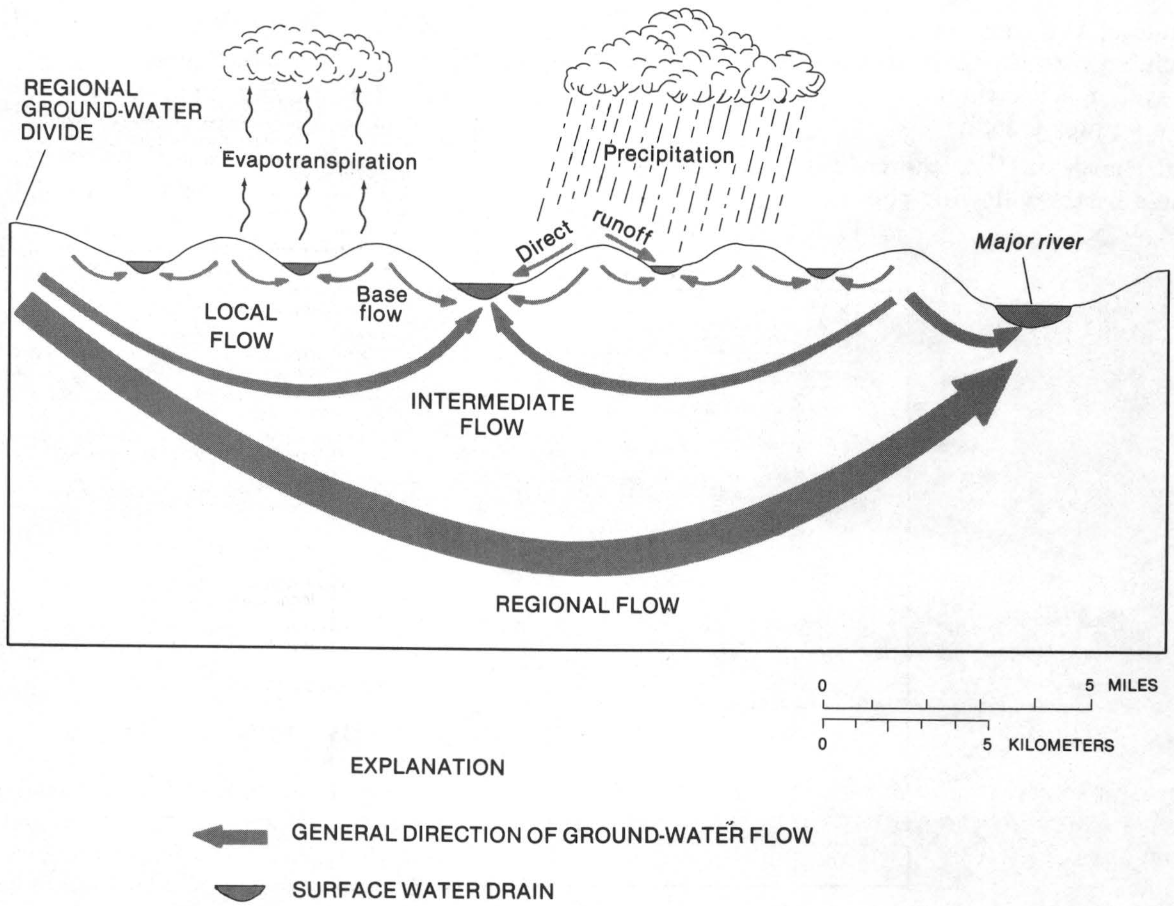


Figure 141.—Relation between surface drainage system and scale of aquifers which discharge into the surface drainage system, Southeastern Coastal Plain regional aquifer system.

Equation 1 can be modified to show the difference between simulated recharge (r) associated with the regional flow system and the total recharge (R) to the entire aquifer system; thus:

$$r = P - DRO - ET - BF, \text{ or}$$
$$r = R - BF \qquad (2)$$

Where, r = simulated regional recharge, in./yr,
BF = base flow to rivers discharged by local and intermediate flow systems which are not simulated, in./yr.

The relatively small component of simulated regional recharge (r) that does not discharge to major rivers can be expressed by:

$$dp = r - bf \qquad (3)$$

Where, dp = simulated deep infiltration of ground water, in./yr,
bf = simulated base flow discharge to major rivers from the regional scale flow system, in./yr.

The schematic diagram of the water budget (inset, fig. 142) illustrates the results of combining the simulated hydrologic components with estimates of precipitation, total runoff, and base flow which are assumed to be known. The estimated evapotranspiration (ET) rate of about 35 in./yr results from the difference between the estimated average precipitation (about 50 in./yr) and the estimated average total runoff (about 15 in./yr) which is the sum of direct runoff and base flow. The base flow of approximately 7 in./yr through local and intermediate scale ground-water flow systems (BF) is the area-weighted average of base flow provided by Stricker (1983, table 1, p. 8) for 26 nonsimulated stream reaches in the outcrop area of the regional aquifer system, all of which have drainage areas of less than 1,000 mi^2. The estimated 7 in./yr of direct runoff to streams (DRO) is the rounded-off result of subtracting the sum of the nonsimulated base flow (BF) and the simulated base flow discharged from the regional scale aquifer systems (bf, about 0.75 in./yr) from total runoff (about 15 in./yr). The total recharge (R) of about 8 in./yr is the difference between precipitation (P) and the sum of

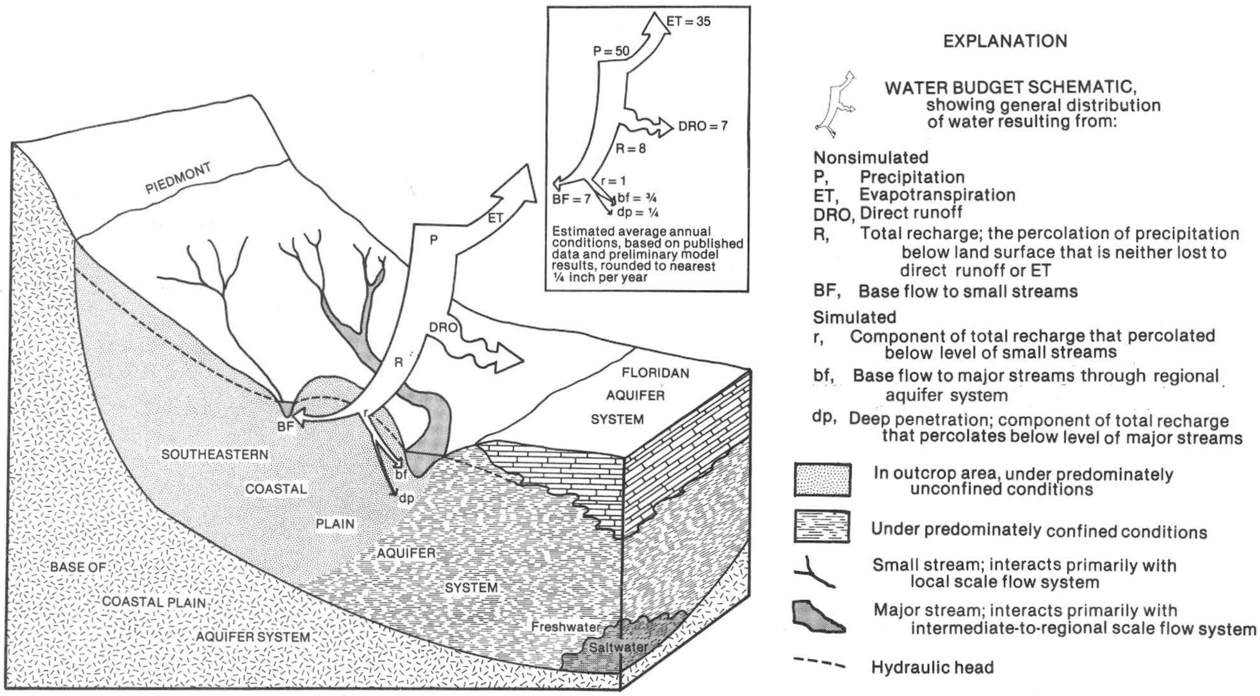


Figure 142.—Simulated hydrologic components, Southeastern Coastal Plain regional aquifer system.

evapotranspiration (ET) and direct runoff (DRO); total recharge also equals the sum of base flow to the unmodeled streams (BF) and the simulated regional recharge (r).

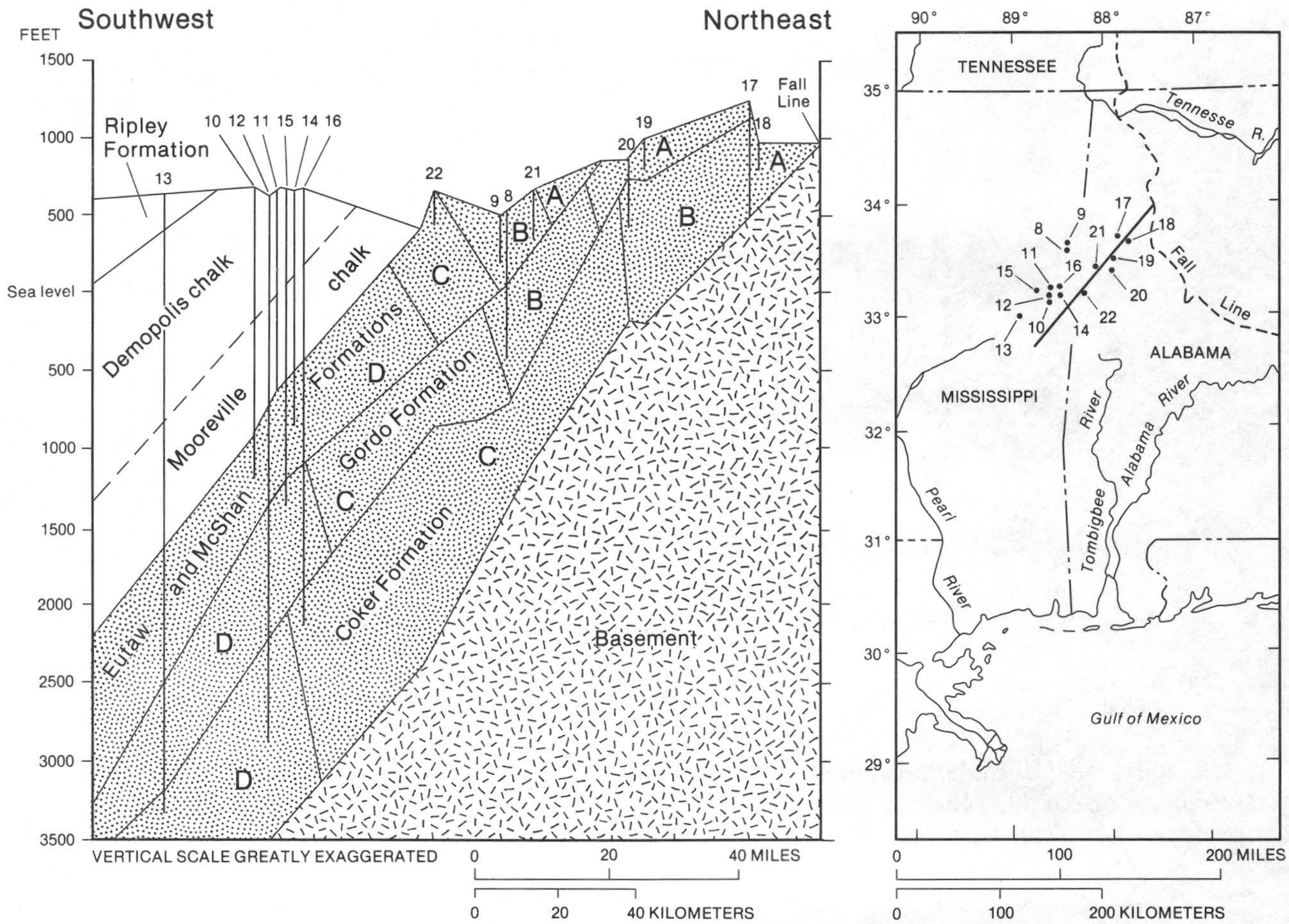
The model-simulated recharge (r) to the regional aquifer systems is about 1 in./yr. The regional aquifer system discharge to major rivers as base flow (bf) is about 0.75 in./yr; the remainder of about 0.25 in./yr (dp) discharges diffusively upward through confining units into the overlying Floridan aquifer system and its equivalents.The simulated values of r, bf, and dp are rounded to the nearest 0.25 in./yr. The sum of simulated boundary flow among the Southeastern Coastal Plain aquifer system and the adjacent Gulf Coastal Plain and Northern Atlantic Coastal Plain regional aquifer systems is less than 10 ft^3/s. The ratio of total recharge (R) to simulated recharge (r) is 10 to 1 and the boundary outflow is 250 to 1.

GEOCHEMISTRY

Water from all aquifers in the Southeastern Coastal Plain regional aquifer system have a similar chemical evolution as the water moves down the hydraulic gradient from outcrop recharge areas to deeper downgradient parts of the aquifers. In the recharge area and at shallow depth (Zone A, fig. 143), the concentrations of dissolved solids are less than 50 mg/L, the water is calcium or sodium-bicarbonate dominated, pH is about 4.5 to 6.0, dissolved oxygen concentrations are 2 to 10 mg/L. At greater depth (Zone B) the concentrations of dissolved solids range from 100 to 200 mg/L, the water is sodium and bicarbonate dominated, pH is about 6.0 to 7.0, with negligible dissolved oxygen, and concentrations of dissolved iron range from 500 to 20,000 mg/L. Downgradient from the recharge area (Zones C and D) the concentrations of dissolved solids range from 300 to 500 mg/L, the water is sodium bicarbonate dominated with increased chloride, pH is about 7.0 to 9.0, and the concentrations of dissolved iron are less than 300 mg/L. Further downgradient (Zone E, not shown in fig. 143) the concentrations of dissolved solids range from 1,000 to 30,000 mg/L, the water is sodium and chloride dominated, pH is about 8.0 to 9.0, and the concentrations of dissolved iron range up to 5,000 mg/L.

Zones A through D were examined in greater detail in Mississippi and Alabama (fig. 143) to gain further insight concerning the types of geochemical reactions that occur after water recharges the aquifers. The various geochemical reactions associated with each zone are summarized in table 5. A fifth zone (zone E, not included in table 5) is characterized as a zone where freshwater is mixed with sodium-chloride brine.

Evidence of water-quality degradation is observed in several locations; large ground-water withdrawals at major pumping centers in



EXPLANATION

A Geochemical zone, explained in text

EXPLANATION

• Sampling well, number assigned during this study

—— Line of section

Figure 143.—Generalized geologic section along a flow path from Alabama to Mississippi showing approximate location of geochemical zones, A, B, C, and D, in the southeastern Coastal Plain.

Mississippi, Alabama, and coastal South Carolina have resulted in higher concentrations of dissolved solids and chloride. An anomalous low in the potentiometric surface of the A4 aquifer in central Alabama (fig. 137) may have contributed to the observed increase in concentrations of dissolved solids and chloride there.

Detailed study of zones A and B indicates that the geochemical character of ground water associated with calcareous clastic aquifers differs from that of sand aquifers that are noncalcareous. Sand aquifers that are calcareous contain water of calcium and bicarbonate dominance in zones A and B, but do not attain high concentrations of dissolved iron in zone B owing to precipitation of siderite. In noncalcareous sand aquifers, water develops high dissolved iron concentrations in zone B and the iron is precipitated in zone C.

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Table 5.—Zonation of chemical reactions of water along a flow path from Alabama to Mississippi in the Southeastern Coastal Plain regional aquifer system [Location of the flow path is shown in fig. 143]

Recharge zone:	Rainfall-input chemistry, atmospheric carbon dioxide + soil carbon dioxide + dissolved oxygen
Zone A:	Calcite dissolution Sodium feldspar hydrolysis to kaolinite Oxidation of carbonaceous matter by dissolved oxygen
Zone B:	Calcite dissolution Sodium feldspar hydrolysis to kaolinite Ferric iron reduction to siderite to saturation of siderite Oxidation of carbonaceous matter Silica precipitation
Zone C:	Calcite dissolution Sodium feldspar hydrolysis to kaolinite Siderite precipitation Ferric iron reduction Oxidation of carbonaceous matter Silica precipitation
Zone D:	Smectite precipitation Calcite dissolution Na-Ca cation exchange Gypsum dissolution Sulfate reduction FeS_2 precipitation Ferric iron reduction Oxidation of carbonaceous matter Silica dissolution

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UPPER COLORADO RIVER BASIN REGIONAL AQUIFER-SYSTEM STUDY

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INTRODUCTION

Water shortage is common in the Colorado River Basin and increasing water demand is expected. The Colorado River Compact of 1922 divided the Colorado River Basin into an upper and a lower basin in order to allocate water supplies. The Upper Colorado River Basin has a drainage area of about 113,500 mi^2 in western Colorado, eastern Utah, southwestern Wyoming, northeastern Arizona, and northwestern New Mexico.

Local studies of the ground-water resources in the Upper Colorado River Basin are numerous; however, regional studies are few. Some regional studies have been described by Iorns and others (1965), the Upper Colorado Region State-Federal Interagency Work Group (1971), and Price and Arnow (1974). To obtain additional regional information systematically on hydrology, geology, and water chemistry of the Upper Colorado River Basin aquifer system, in 1981, the U.S. Geological Survey started the Upper Colorado River Basin regional aquifer study, scheduled for completion in 1986. Excluded from this study is the upper part of the San Juan River Basin, which will be investigated in a separate study. The remaining study area is about 100,000 mi^2 (fig. 144).

The area covered by this study contains a variety of landforms: rugged mountains, broad plains, deeply dissected canyons, relatively flat flood plains, and many erosional features. Consolidated sedimentary formations of Paleozoic, Mesozoic, and Cenozoic age attain a maximum thickness of tens of thousands of feet. These formations include aquifers within beds of fractured limestone, dolomite, sandstone, and shale. Low-permeability limestone, dolomite, shale, and evaporite deposits act as confining units. Igneous rocks, especially volcanic rocks, are also present in part of the study area, but they are not regional aquifers.

The study area has been subjected to repeated tectonism. The predominant tectonic features are numerous basins and uplifts (fig. 145). The resulting structural relief is nearly 30,000 feet above the basin floors in places. Because of this relief, several aquifers that are deeply buried in basins are exposed on the margins of uplifts, where precipitation partly recharges the aquifers. Aquifers within stratigraphically younger formations tend to be exposed and recharged over extensive areas.

Annual precipitation ranges from approximately 6 inches on the plains of Utah to about 40 inches in mountainous areas. Precipitation, in form of snowmelt and rainfall, is the only source of recharge to the aquifers.

PROGRESS AND RESULTS

Ground-water flow systems have been classified into three major groups. In descending order, they are: (1) Cenozoic-rock aquifers, (2) Mesozoic-rock aquifers, and (3) Paleozoic-rock aquifers. These flow systems are being studied individually in U.S. Geological Survey offices in Colorado, Utah, and Wyoming. Studies of the flow systems are being integrated to ensure that interactions among the three identified flow systems are investigated. Within each flow system, rocks are grouped into aquifers and confining units on the basis of lithology, depositional environment, and hydrologic characteristics. A total of 11 hydrogeologic units has been identified for the Upper Colorado River Basin regional aquifer system.

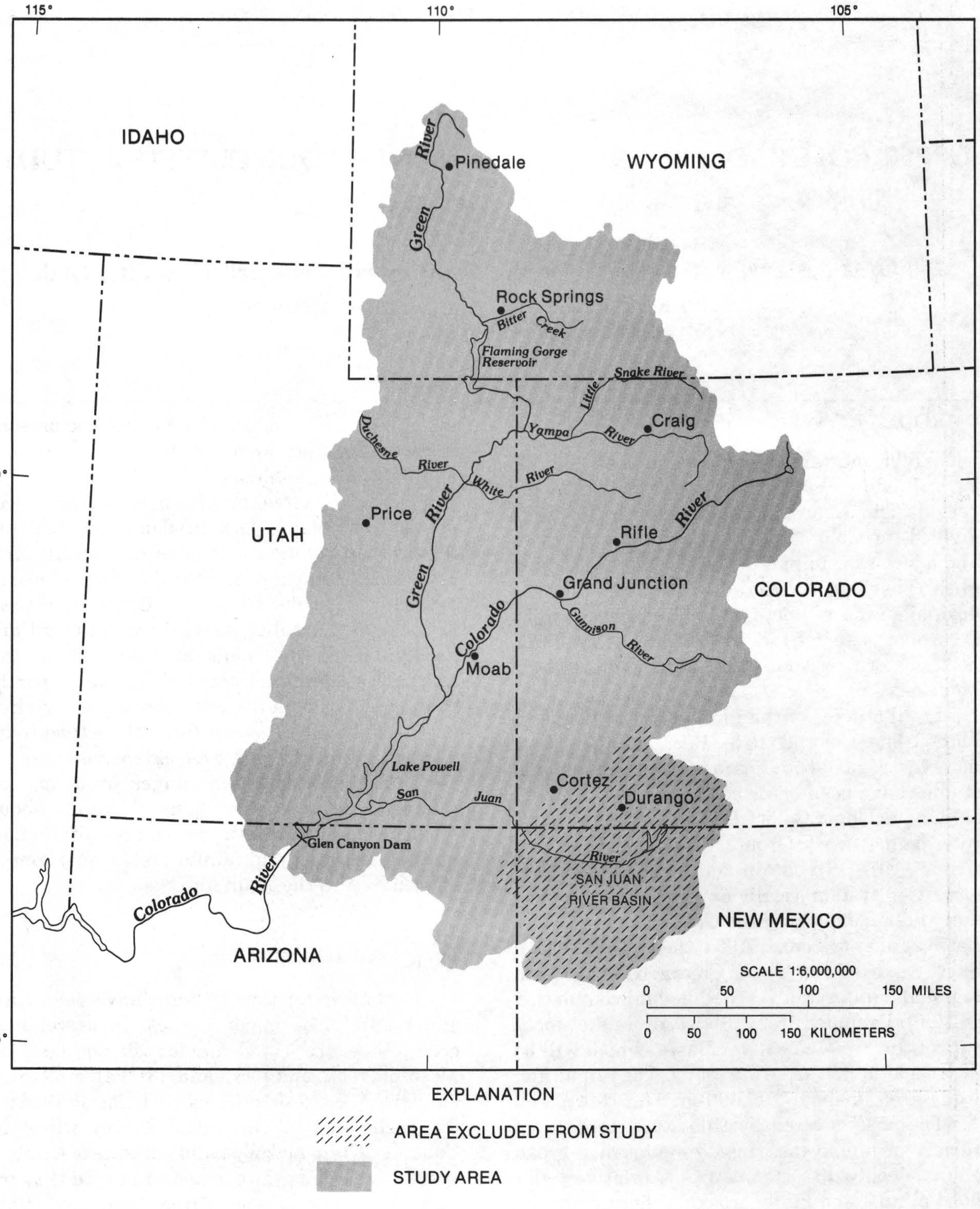


Figure 144.—Study area of the Upper Colorado River Basin regional aquifer system.

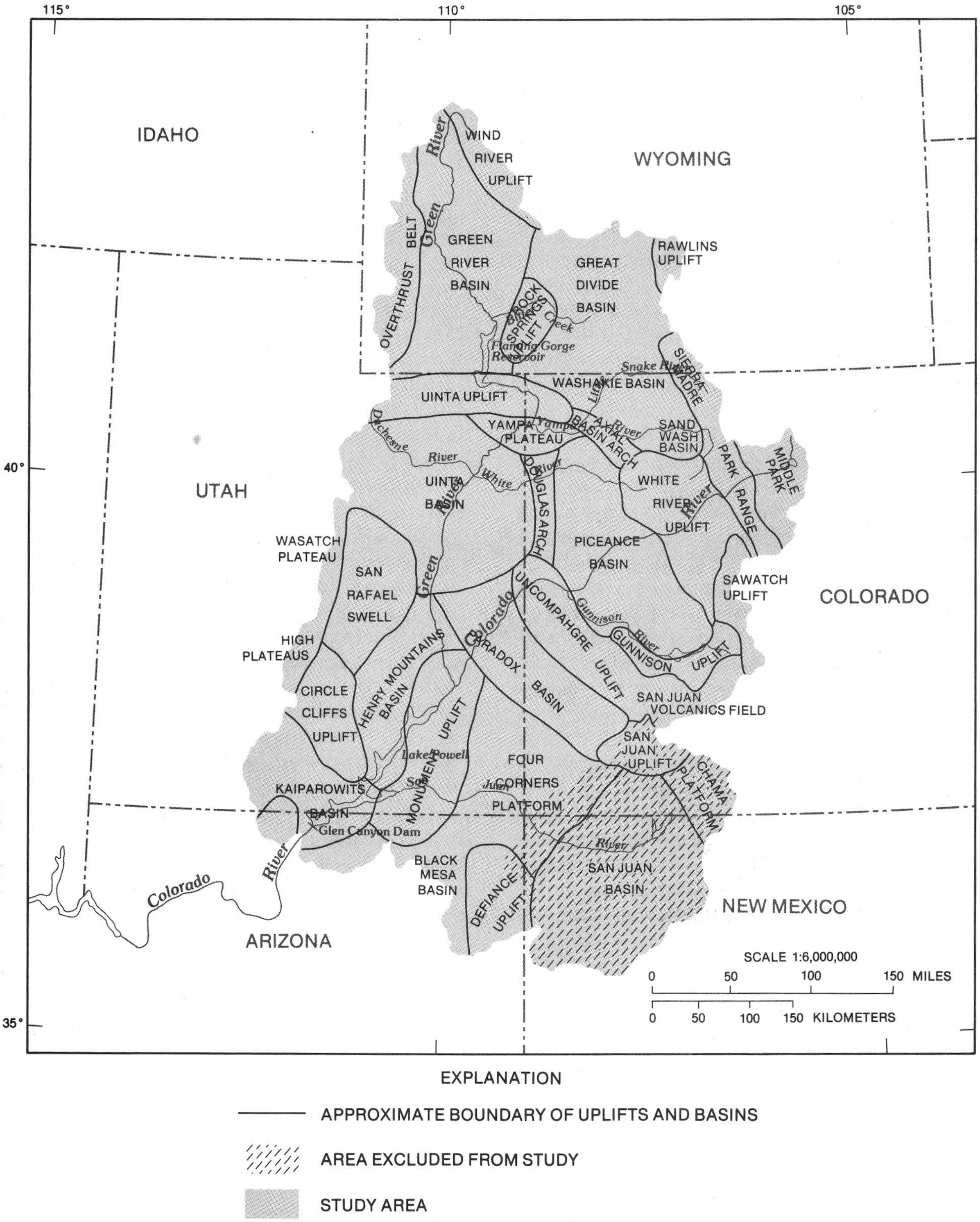


Figure 145.—Principal tectonic features in the Upper Colorado River Basin (modified from Grose, 1972).

Paleozoic Rocks

Paleozoic rocks extend over most of the study area (fig. 146). These rocks were subdivided into four hydrogeologic units. In descending order, they are: (1) an upper Palezoic aquifer and confining unit which contains sandstone, red beds, and evaporites of Mississippian, Pennsylvanian, and Permian age; (2) a middle Paleozoic aquifer which contains limestone and dolomite of Devonian and Mississippian age; (3) a lower Paleozoic aquifer and confining unit which contains limestone and dolomite of Cambrian and Ordovician age; and (4) a basal Paleozoic aquifer which contains sandstone and quartzite of Cambrian age. Due to the great depth of the lower two units, very few drillers' logs are available. Therefore, the analysis of Paleozoic rocks will be limited to the upper two hydrogeologic units.

Structure contours drawn to represent the top of each hydrogeologic unit indicate considerable relief because of the numerous basins and uplifts. Isopach maps indicate a large range in thickness, especially for the uppermost hydrogeologic unit. Potentiometric surface maps were prepared from shut-in pressures obtained during drill-stem tests. The potentiometric surface maps indicate that the upper two hydrogeologic units are recharged by precipitation in uplifted areas. Water moves through the formations toward the valleys and discharges to major streams and tributaries. Maps of concentrations of dissolved solids generally substantiate the identified flow pattern. In areas of recharge, the concentrations of dissolved solids are relatively low. Water quality deteriorates downgradient and with depth. In some basins, concentrations of dissolved solids exceed 400,000 milligrams per liter (mg/L) at depth.

A detailed analysis of the major flow systems is underway. Hydraulic conductivity data have been obtained by interpreting drill-stem tests (Teller and Chafin, 1984) and from literature. Reported values of hydraulic conductivities for Paleozoic sandstones range from 3×10^{-5} to 100 ft/d; reported values for Paleozoic carbonate rocks range from 6×10^{-5} to 600 ft/d. The discharge of major aquifers to valley streams is being estimated by determining base flow from analysis of historical streamflow data at major gaging stations. Quality data of streamwater collected at gaging stations during base-flow periods are being analyzed to identify which aquifers are discharging to streams. Hydrogeologic sections in selected areas are in preparation, and will be used to interpret and simulate the vertical flow components between aquifers.

Model simulation is being used to evaluate the flow system of the Paleozoic-rock aquifers in parts of south-central Utah and northwestern Colorado. The simulation includes the middle and upper Paleozoic aquifers and their associated confining units. The simulation is used to evaluate the hydrologic effects and potential of ground-water development.

Mesozoic Rocks

Based on hydrogeologic characteristics, the Mesozoic-age rocks were divided into three units. In descending order, they are: (1) an upper Mesozoic aquifer and confining unit which contains shale and sandstone of Jurassic and Cretaceous age; (2) a middle Mesozoic aquifer which contains sandstone and shale of Triassic and Jurassic age; and (3) a lower Mesozoic confining unit which contains siltstone, and mudstone of Triassic age. The three hydrogeologic units have been described in detail by Freethey and others (1984). Figure 147 shows the areal extent of the Mesozoic rocks.

Preliminary analysis of stratigraphic data indicates that Jurassic and Triassic sandstones constitute a regional aquifer (the middle Mesozoic aquifer), that extends over about 75 percent of the Upper Colorado River Basin. Aggregate thickness of this hydrogeologic unit increases westward to more than 2,000 feet, but saturated thickness is less than aggregate thickness where the hydrogeologic unit crops out. Cretaceous sandstone underlies about 50 percent of the Upper Colorado River Basin. Although the Cretaceous sandstone yields water to wells, the water quality generally is poor. Most of the Mesozoic rocks are laterally discontinuous and are only locally significant as sources of water.

Water-level data are being compiled and potentiometric surface maps were constructed for the significant formations of the Mesozoic-rock aquifers. They are the Glen Canyon Group, Entrada Sandstone, Morrison Formation, lower Cretaceous Sandstone, and Mesaverde Group. The Glen Canyon Group is the most laterally continuous system.

Ground-water occurrence and direction of movement can be determined in areas where the Mesozoic rocks are not deeply buried. Potentiometric surface maps of the Mesozoic-rock aquifers indicate that the regional movement of ground water is toward the large river valleys,

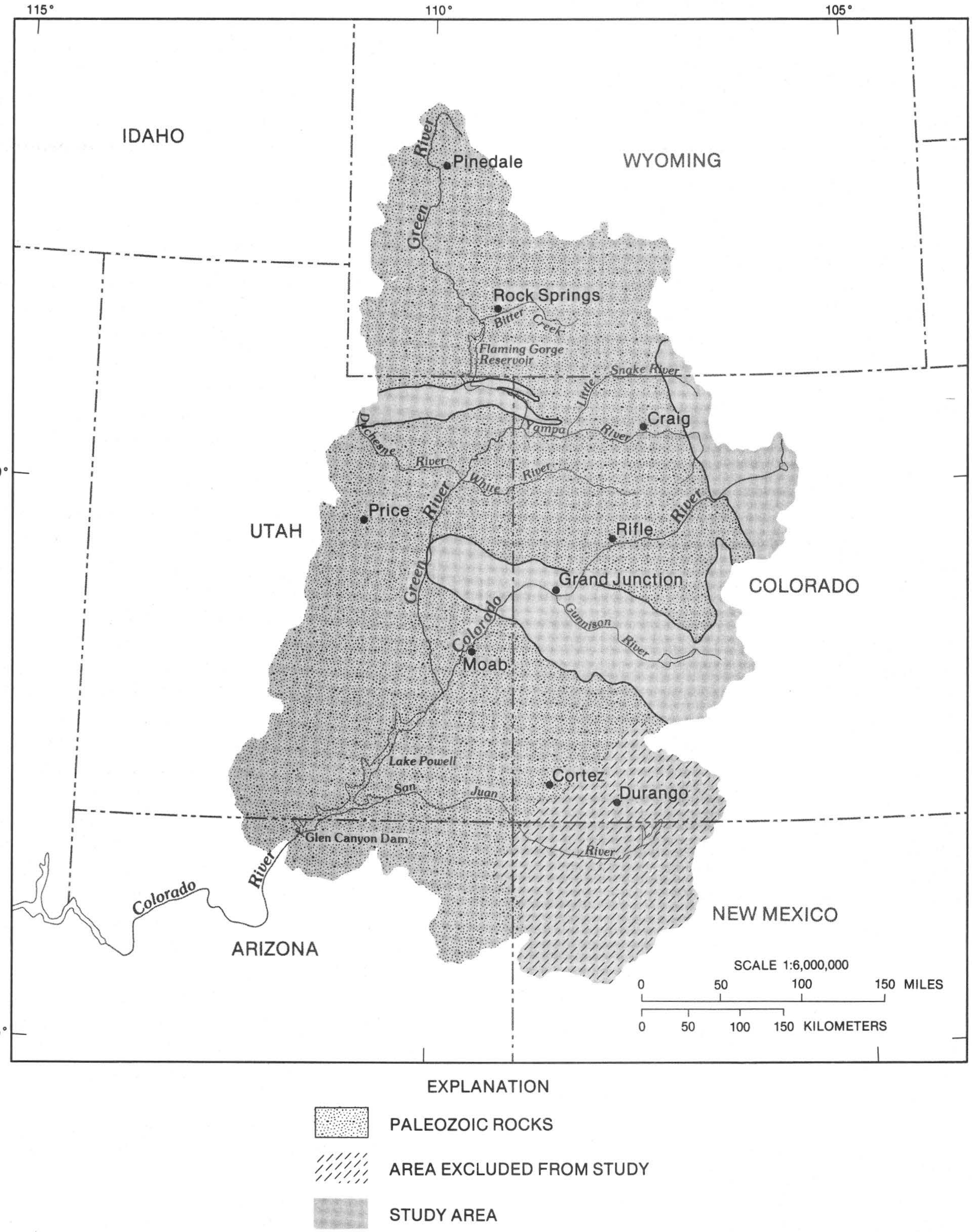


Figure 146.—Areal extent of Paleozoic rocks in the Upper Colorado River Basin (modified from King and Beikman, 1974, and Rocky Mountain Association of Geologists, 1972).

mainly those of the Colorado, Green, Yampa, and San Juan Rivers (fig. 144). Small river valleys control ground-water discharge locally. Potentiometric data indicate vertical movement of water also takes place from one formation to another, but the direction of movement is governed largely by the relative locations of recharge and discharge.

Quality of ground water is related to: (1) distance and time of travel from the recharge areas to the discharge areas, and (2) solubility of minerals in the sediments. Ground water in the upper Mesozoic aquifer and in the middle Mesozoic aquifer generally is fresh with calcium-bicarbonate dominance where the formations are near land surface or close to recharge areas. Ground water is briny with sodium-chloride dominance in deep basins. Large concentrations of dissolved solids are present where ground water flows through formations containing evaporites.

Available data indicate that hydraulic conductivity in unfractured Mesozoic sandstone ranges from about 1×10^{-4} to 10 ft/d, and ranges from 1×10^{-2} to 200 ft/d for fractured sandstones.

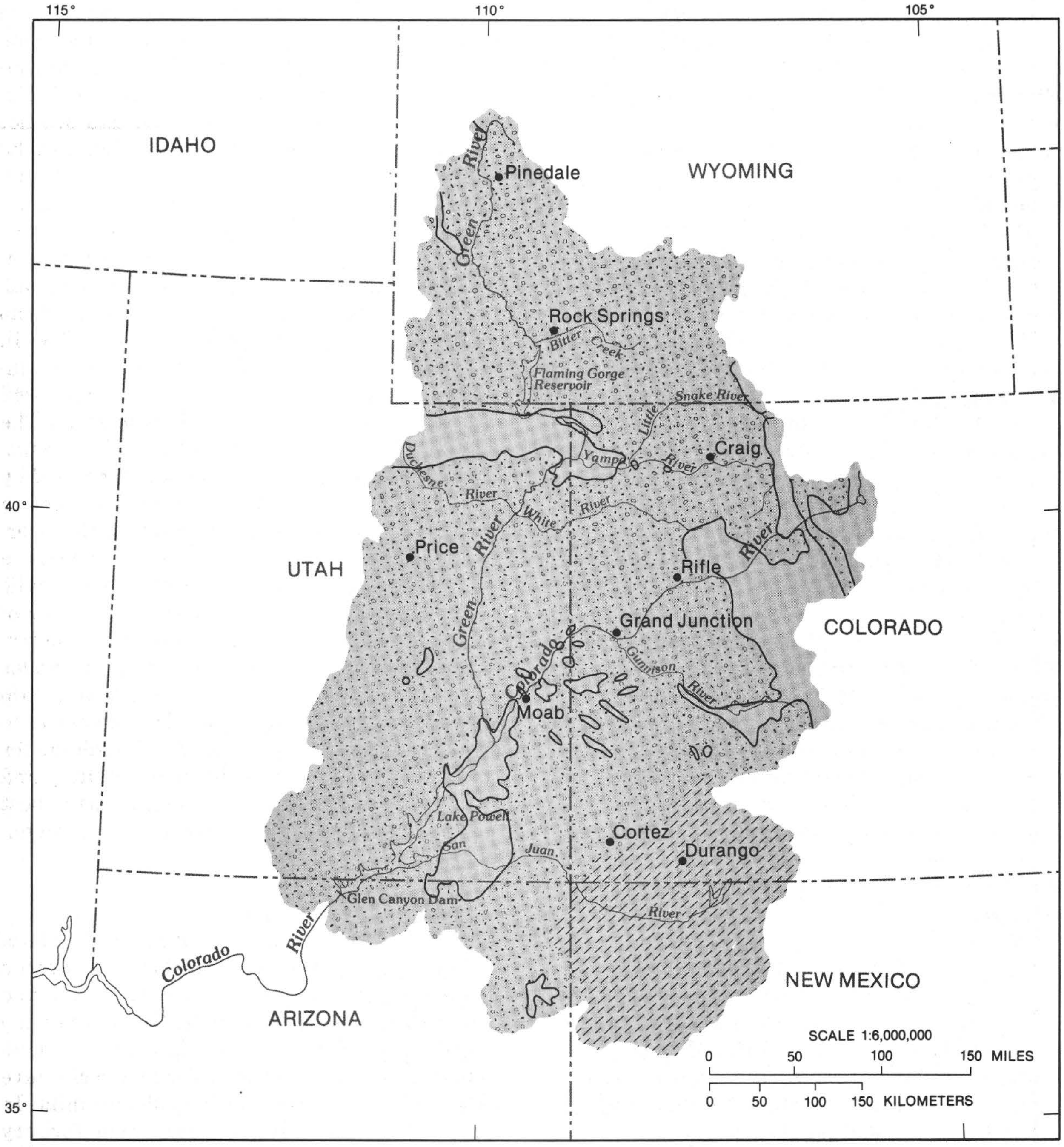
Intrinsic permeability was determined from two Mesozoic rock core samples using solutions that simulated freshwater (dissolved solids of 540 mg/L) and brine (dissolved solids of 70,800 mg/L). Intrinsic permeability values obtained using the brine were 31 and 74 percent greater than those obtained using freshwater. The largest difference in permeability was in the rock core having the greatest quantity of silt- and clay-sized particles, and is thought to be due to flocculation of clay by the saline water with a resulting increase in permeability.

Intrinsic permeability using freshwater was determined from four Mesozoic rock core samples with various pressures on the core to simulate burial depths ranging from 400 to 6,000 feet. A decrease of intrinsic permeability values with increased pressure was evident but small, ranging from 8 to 11 percent. The largest percentage of decrease was in the rock core having the largest mean grain-sized particles.

Two studies have been started to aid quantitative evaluation of the hydrologic characteristics of the Mesozoic aquifers. The first is a comprehensive compilation of hydraulic and lithologic properties of aquifers and confining units. The compilation is essential to the assessment of ground-water recharge and discharge areas and to an understanding of the interformational and intraformational ground-water movement. The second study consists of model simulation. Two models simulating ground-water flow in the Navajo Sandstone have been completed. One model simulates ground-water flow near the San Rafael swell and the other near Lake Powell (fig. 148). Both models are generalized and are not suitable for specific evaluations. Simulation of the Navajo Sandstone near the San Rafael swell supports the concept that the aquifer is primarily recharged on the west side of the swell. Water moves around the swell to the east. Simulation of the Navajo Sandstone near Lake Powell was used to analyze the possible far-reaching effects caused by filling the impounded Lake Powell. This model simulates the local effects of the inundation of the Navajo Sandstone by Lake Powell due to construction of Glen Canyon Dam. The simulation indicates that the hydraulic conductivity of the Navajo Sandstone was increased by fractures. Hydraulic conductivity values used in the model range from 0.25 to 3.5 ft/d. However, fracturing of the Navajo probably occurs near the synclinal axes because in the model the hydraulic conductivity had to be increased by a factor of three to obtain a reasonable match with observed conditions. A third model is being prepared for the region near the Four Corners area (fig. 148) and is a multilayered model. This model is to evaluate various ground-water flow definitions for the Mesozoic aquifers in the modeled area, and to determine which of these definitions is best suited to describe the regional aquifer system.

Cenozoic Rocks

The Cenozoic rocks are sedimentary rocks of Tertiary age. Aquifers of Tertiary age overlie much of the northern half of the study area and are exposed at the surface in the Green River and in the Great Divide-Washakie, Piceance, and Uinta basins (figs. 145 and 149). Tertiary rocks have been divided into four hydrogeologic units. In descending order, they are: (1) an upper Tertiary aquifer which contains sandstone, conglomerate, claystone and tuff of Eocene, Oligocene, Miocene, and Pliocene age; (2) a middle Tertiary aquifer and confining unit which contain claystone, sandstone, shale, limestone, and conglomerate of Eocene and Oligocene age; (3) a lower Tertiary aquifer and confining unit which contain shale and sandstone of Paleocene and Eocene age; and (4) a basal Tertiary aquifer which contains sand-



EXPLANATION

MESOZOIC ROCKS

AREA EXCLUDED FROM STUDY

STUDY AREA

Figure 147.—Areal extent of Mesozoic rocks in the Upper Colorado River Basin (modified from King and Beikman, 1974).

stone, mudstone, shale, and coal of Paleocene and Eocene age.

The hydraulic properties of the Tertiary formations vary markedly from basin to basin. Formations such as the Wasatch of the lower Tertiary aquifer are water yielding in the Green River and Great Divide-Washakie basins, but do not transmit significant quantities of water to wells in the Piceance and Uinta basins. Lake deposits in the Green River Formation of the same hydrogeologic unit form a regional aquifer in the Piceance basin where they are fractured; elsewhere, this hydrogeologic unit is a confining unit. Because of the large variability of the hydraulic properties of the Tertiary aquifers, as well as the relative hydrologic isolation of the major Tertiary basins, the Uinta, Piceance, and Wyoming basins are discussed separately.

In the Uinta basin, the Duchesne River Formation of the middle Tertiary aquifer is most productive; however, it is limited in areal extent to the northern half of the basin. The Uinta Formation of the same hydrogeologic unit is used extensively for water supplies. An inventory of water wells undertaken as part of the study indicates that the Green River and Wasatch Formations are not used extensively for water supply. The North Horn Formation (the basal Tertiary aquifer) is also not used for water supply.

The thickness and depth of burial of the Tertiary aquifer have been mapped from geologic and geophysical logs throughout the Uinta basin as part of the study. The total thickness of Tertiary sediments in the north-central part of the Uinta basin exceeds 15,000 feet. For this reason, few wells have been drilled through the entire Tertiary sequence.

Aquifer characteristics of the Tertiary rocks in the Uinta basin are dependent upon the relative amount of sandstone and degree of fracturing. The Duchesne River and Uinta Formations consist of varying amounts of sand interbedded with shale. However, the most permeable zones are the rocks that were extensively fractured. Estimates of hydraulic conductivity for the Duchesne River Formation range from less than 1 ft/d for unfractured rocks to 500 ft/d for fractured rocks. Work is planned to establish a regional relation between hydraulic conductivity and fracture pattern. The hydraulic conductivity of the formations underlying the Uinta Formation is generally less than 1 ft/d.

The Piceance basin (fig. 145) is the largest area of oil-shale development in the western United States. The important Tertiary aquifer in the Piceance basin is the Uinta Formation and parts of the Parachute Creek Member of the Green River Formation. The depth of burial and thickness of these formations have been mapped in the principal area of oil-shale development, and many hydrologic studies have been completed there. The role of this study in this area is primarily to extend the results of other investigations southward from the area of oil-shale development to the part of the structural basin that lies south of the Colorado River.

In Wyoming, the Fort Union Formation and the Wasatch Formation and their equivalents are the most important hydrogeologic units of the Tertiary aquifer. The Wasatch and Green River Formations interfinger around the periphery of both the Green River and Washakie basins. Sparse well data show that the Wasatch Formation also occurs in the deep parts of these basins, where the formation is completely overlain by confining units of the Green River Formation. The Fort Union and Wasatch Formations, or their equivalents, are found throughout most of the study area in Wyoming. The thickness of formations of the Tertiary aquifer has been mapped utilizing existing data and limited field reconnaissance. However, it has been difficult to identify formation contacts in the subsurface. The similar lithology of the Fort Union and Wasatch Formations has prevented accurate determinations of their contact from geologic and geophysical logs. Accordingly, a generalized map of the combined thickness of the Fort Union and Wasatch Formations has been made. The maximum combined thickness of these units is about 11,000 feet near Pinedale, WY. Areas of thickness greater than 7,000 feet occur near Flaming Gorge Reservoir and in the center of the Great Divide and Washakie basins.

The ability of the Wasatch and Fort Union Formations to transmit and store water is being evaluated. Hydraulic characteristics of the formations were evaluated from sparse aquifer-test, specific-capacity, and drill-stem test data. The most reliable estimates of the aquifer properties have been made in the northern Green River basin. A series of aquifer tests made at different depths indicate that the hydraulic conductivity of the Wasatch Formation, which is about 2 ft/d, is fairly uniform from land surface to a depth of 2,000 feet. However, at 5,000 feet below land surface, the hydraulic conductivity is much lower (7×10^{-3} ft/d) , and below 7,000 feet the hy-

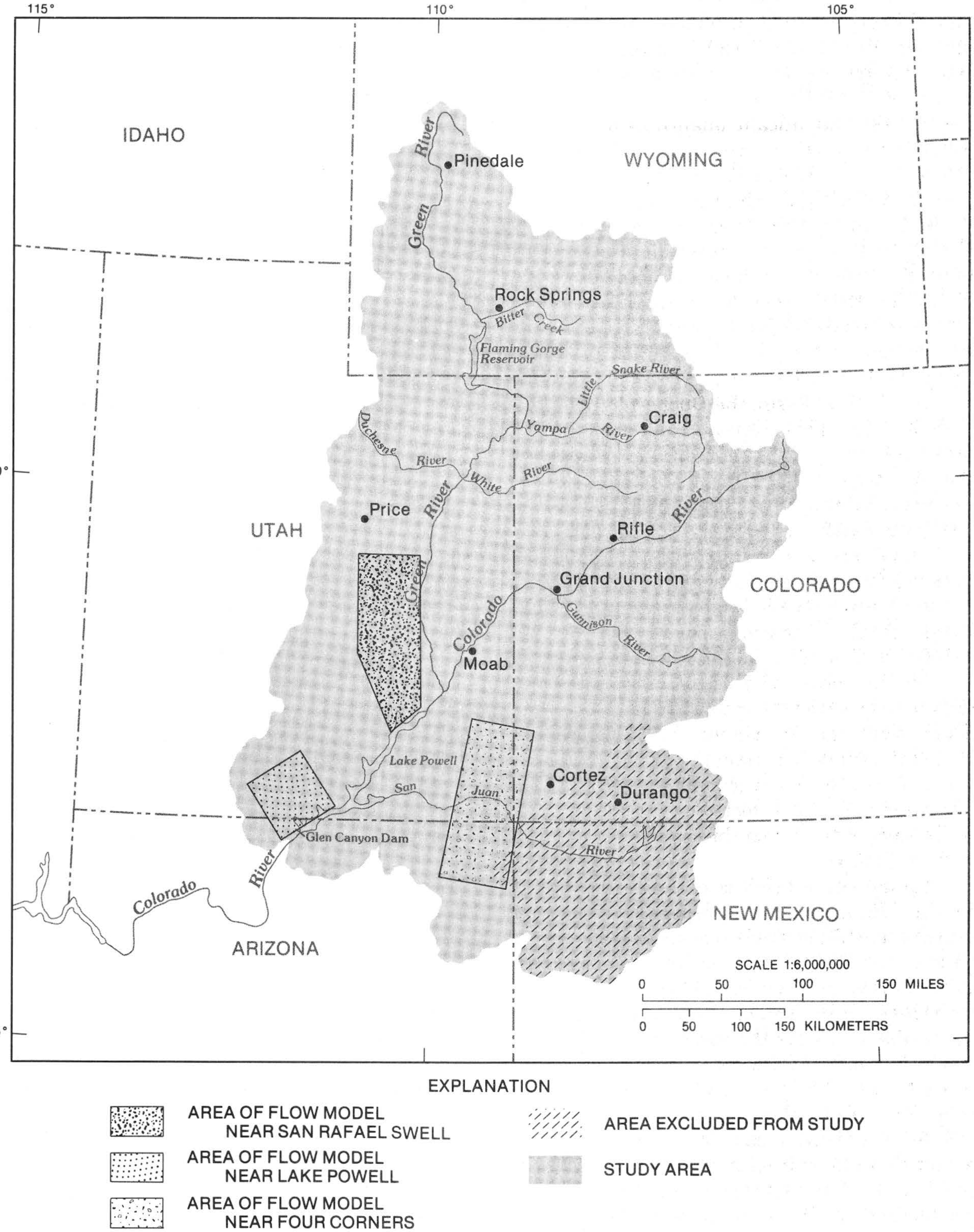


Figure 148.—Location of flow models of the Mesozoic aquifer, Upper Colorado River Basin regional aquifer system.

draulic conductivity of the Wasatch is further reduced (5×10^{-4} ft/d).

Potentiometric surface maps have been prepared in many areas where the Wasatch Formation is exposed at the surface. These maps generally show that recharge occurs on the periphery of the basins and in upland areas, and ground water discharges to streams in the center of the basins. In the Great Divide basin (fig. 145), water in the Wasatch Formation discharges to

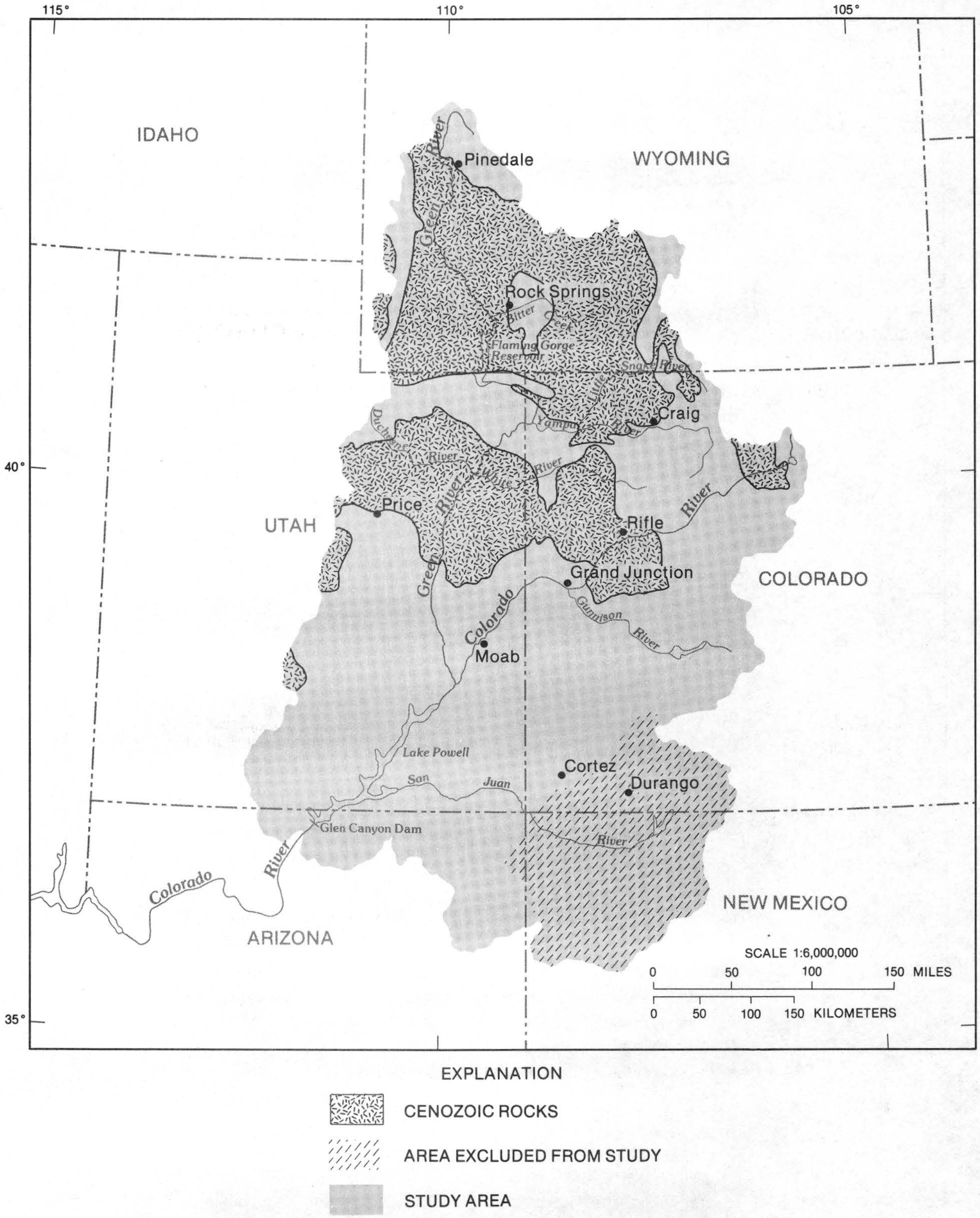


Figure 149.—Areal extent of Cenozoic rocks in the Upper Colorado River Basin (modified from King and Beikman, 1974).

salt flats and phreatophytes. The rate of evapotranspiration is large enough to prevent formation of permanent lakes in the Great Divide basin.

Water-level data are available for the Wasatch Formation in areas overlain by the Green River Formation. Flowing wells are common, and heads are scheduled to be measured. A preliminary examination of available water-level data has shown little variation in heads with depth. The data examined to date primarily represent the Wasatch Formation in outcrop areas. As additional data are collected and interpreted, it is anticipated that head variation with depth will be observed.

Quality of Water

Water-quality investigations in the Upper Colorado River Basin are intended to supplement hydrologic investigations and to serve as an independent check on ground-water flow directions, velocities, and aquifer properties.

A water-quality data base has been made that includes information from the WATSTORE data base of the U.S Geological Survey, from petroleum companies, and from published reports. Maps showing concentrations of dissolved solids have been prepared for the principal hydrogeologic units (Freethey and others, 1984). The water-quality trend provides an independent check of recharge, ground-water movement, and discharge interpreted from potentiometric-surface maps and simulations.

Additional water samples have been collected from the principal hydrogeologic units for analyses of stable isotopes and carbon-14. This isotope-and carbon-14 information will help develop chemical-reaction models. The stable-isotope information provides independent determinations of mixing of water between the hydrogeologic units. This mixing generally occurs between freshwater from recharge areas and brine in the deep hydrogeologic units. Each of these determinations will aid in refining the concepts of ground-water flow in the Upper Colorado River Basin.

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PHASE I REGIONAL AQUIFER-SYSTEM ANALYSIS PROJECTS STARTED IN 1984

CARIBBEAN ISLANDS REGIONAL AQUIFER-SYSTEM STUDY

By Fernando Gómez-Gómez, San Juan, Puerto Rico

INTRODUCTION

The Caribbean Islands regional aquifer system includes Puerto Rico, its offshore islands (Vieques, Culebra, and Mona), and the U.S. Virgin Islands (St. Croix, St. Thomas, and St. John.) However, the regional aquifer system study will investigate only the aquifers underlying the islands of Puerto Rico and St. Croix (fig. 150).

The island of Puerto Rico has an area of about 3,300 mi^2. It consists of a series of east-to-west mountain ranges with a maximum altitude of about 4,400 feet, flanked on the north and south by foothills. Extensive coastal plains as much as 8 miles in width exist along the north and south coasts. Rainfall ranges from 200 inches in the rain forests of the northeast to 35 inches in the lowlands of the southwest. The annual average rainfall is about 75 inches. Streamflow varies seasonally with precipitation and topography (Bogart and others, 1964). The average runoff (25 years of record) is about 23 in./yr (Gómez-Gómez and Heisel, 1980).

HYDROGEOLOGIC FRAMEWORK

The geology of Puerto Rico is complex and varied. The central core of the island consists largely of volcanic and intrusive rocks of late Cretaceous and early Tertiary age. These rocks are interbedded with lava flows and thin beds of limestone, which have been partly recrystallized in many places. The volcanic rocks and interbedded limestones have been complexly faulted, folded, metamorphosed, and intruded by dioritic rocks. Clastic sediments and limestones of Oligocene and Miocene age were deposited to the north and south of the central mountain core. The older clastic sediments are poorly sorted mixtures of gravel, sand, and fine-grained materials. These sediments grade upward into thick beds of relatively pure limestone. The limestones have been subjected to intensive solutional activity at their outcrop areas. A mature karst topography has developed mostly along the north coast (Giusti, 1978).

The principal aquifers in Puerto Rico are as follows:

1. Alluvial deposits of Quaternary age along the coastal areas and in some interior stream valleys.
2. Limestone formations of Middle Tertiary age on the north and south coasts.
3. Volcanic and igneous rocks of Late Cretaceous and Early Tertiary age in the interior.

Six ground-water provinces (McGuinness, 1948) have been defined in Puerto Rico as follows (fig. 151):

(1) North Coast,
(2) South Coast,
(3) East Coast,
(4) West Coast,
(5) Lajas Valley, and
(6) Interior.

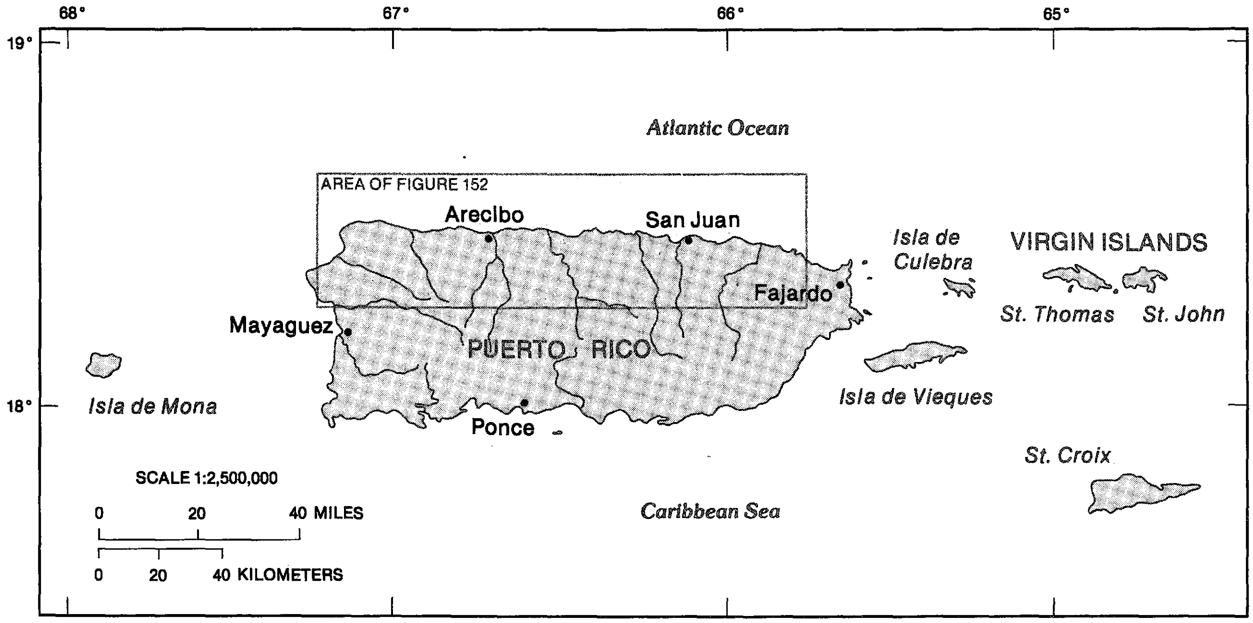


Figure 150.—Study area (shaded) of the Caribbean Islands regional aquifer system.

North Coast Ground-Water Province

The North Coast Ground-Water Province extends for approximately 90 miles in length and averages 9 miles in width. It is comprised primarily of Tertiary limestones which dip north at approximately 5 degrees (fig. 152). The North Coast Ground-water Province has been further subdivided into six areas (fig. 151):

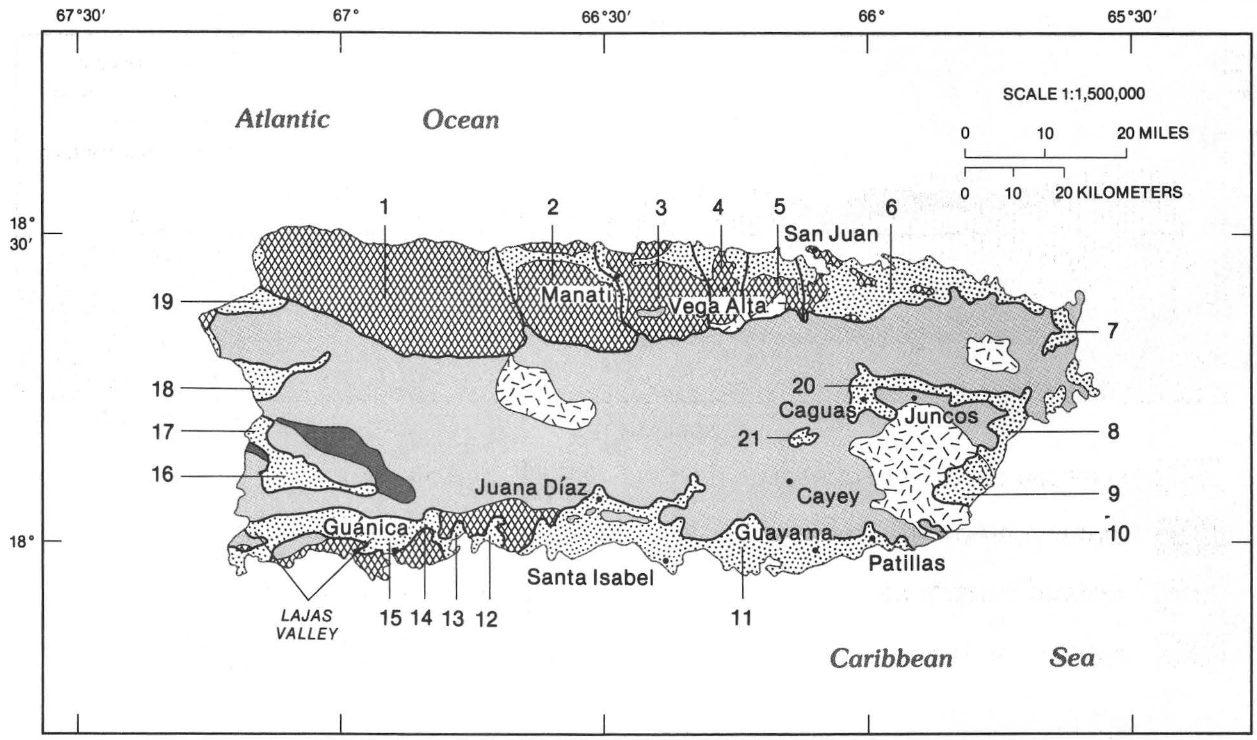
(1) Aguada-Río Grande de Arecibo,
(2) Río Grande de Arecibo-Río Grande de Manatí,
(3) Río Grande de Manatí-Río Cibuco,
(4) Río Cibuco-Río de La Plata,
(5) Río de La Plata-Río de Bayamón, and
(6) Río de Bayamón-Río Grande.

These subdivisions follow general drainage patterns of the principal rivers in the area.

1. Aguada-Río Grande de Arecibo.—This area, from Aguada to Río Grande de Arecibo, is an elevated limestone plateau about 25 miles long and 12 miles in width. It is characterized by rolling hills on the northern half and intense karst development on the southern half. A water-table aquifer occurs within the Aymamón and Aguada Limestones. In the southern part of the outcrop areas, the older rocks of the Cibao Formation and Lares Limestone are a water-table aquifer. Depth to the water table normally exceeds 400 feet below land surface, except in lower areas along the narrow river valleys. Ground-water development has been minimal in this area (less than 10 Mgal/d). Yields to wells range from 50 gal/min in the inland areas to as much as 1,500 gal/min in the vicinity of Río Camuy. There is no information on yields from the older Cibao Formation and Lares Limestone. These formations dip toward the coast to depths of as much as 2,500 feet. Wells within one mile from the coast tapping the water-table aquifer are generally saline.

2. Río Grande de Arecibo-Río Grande de Manatí.—The principal aquifers in this area occur within stream-valley alluvium and limestone rocks. Alluvial deposits of Quaternary age as much as 300 feet thick consist of fine-grained sand, gravel, and clay. Yields to wells range from 50 to 800 gal/min. Water-table and artesian aquifers exist in the limestone units. The water-table aquifer within the Aymamón and Aguada Limestones is the most productive. Yields to wells range from 150 to more than 2,000 gal/min. Within the outcrop areas of the Montebello Limestone Member of the Cibao Formation and Lares Limestone yields of wells are generally poor, seldom exceeding 100 gal/min. However, artesian conditions occur within these formations at depths in excess of 1,500 feet with initial heads of about 400 feet above sea level and well yields in excess of 2,000 gal/min. Ground-water withdrawals from the water-table aquifers are about 20 Mgal/d, and 7 Mgal/d from the artesian system. The quality of the water in the water-table aquifer has been affected by seawater encroachment and chemical spills.

3. Río Grande de Manatí-Río Cibuco.—The principal aquifer in this area occurs in the Aymamón and Aguada Limestones. The artesian system formed by the deeper Montebello Member of the Cibao Formation and Lares Limestone is known to occur toward the western section of the area. The Montebello Limestone, however, pinches out east of the Río Grande de Manatí. There is no information on the possible occurrence of the artesian system east of the town of Manatí. Alluvial deposits are locally important within the Río Grande de Manatí lower valley. In the Río



EXPLANATION

	UNCONSOLIDATED ALLUVIAL AND OLD ALLUVIAL DEPOSITS	QUATERNARY
	LIMESTONE	TERTIARY
	MASSIVE ANDESITIC TUFFS, SHALES, AND STRATIFIED ASH AND TUFF	CRETACEOUS AND TERTIARY
	VOLCANIC AND IGNEOUS ROCKS -- Granitoid intrusives including diorites, quartz diorites, granites, and other holocrystalline types	
	SERPENTINITE	CRETACEOUS?
——	BOUNDARY OF AQUIFER AREAS	

GROUND-WATER PROVINCES

NORTH COAST
1 Aguada to Río Grande de Arecibo
2 Río Grande de Arecibo to Río Grande de Manatí
3 Río Grande de Manatí to Río Cibuco
4 Río Cibuco to Río de La Plata
5 Río de La Plata to Río de Bayamón
6 Río de Bayamón to Río Grande
EAST COAST
7 Fajardo
8 Naguabo - Humacao
9 Yabucoa
10 Maunabo

SOUTH COAST
11 Ponce to Patillas
12 Tallaboa
13 Guayanilla
14 Yauco
15 Guánica
WEST COAST
16 Guanajibo
17 Yaguez
18 Añasco
19 Culebrinas
LAJAS VALLEY
INTERIOR
20 Caguas-Juncos valleys
21 Cayey valley

Figure 151.—General geology of the Island of Puerto Rico and ground-water provinces.

Cibuco Valley sediments are mostly composed of silty-clay and are not productive.

Yields to wells in the water-table aquifer range from 200 to more than 2,000 gal/min. In the artesian aquifer initial heads of 500 feet above sea level and well yields of 1,000 gal/min were reported. Production from the alluvial deposits are usually less than 800 gal/min. Pumpage exceeds 30 Mgal/d. Seawater encroachment occurs in the water-table aquifer near the coast.

4. Río Cibuco-Río de La Plata.—The principal aquifer in the area is also within the

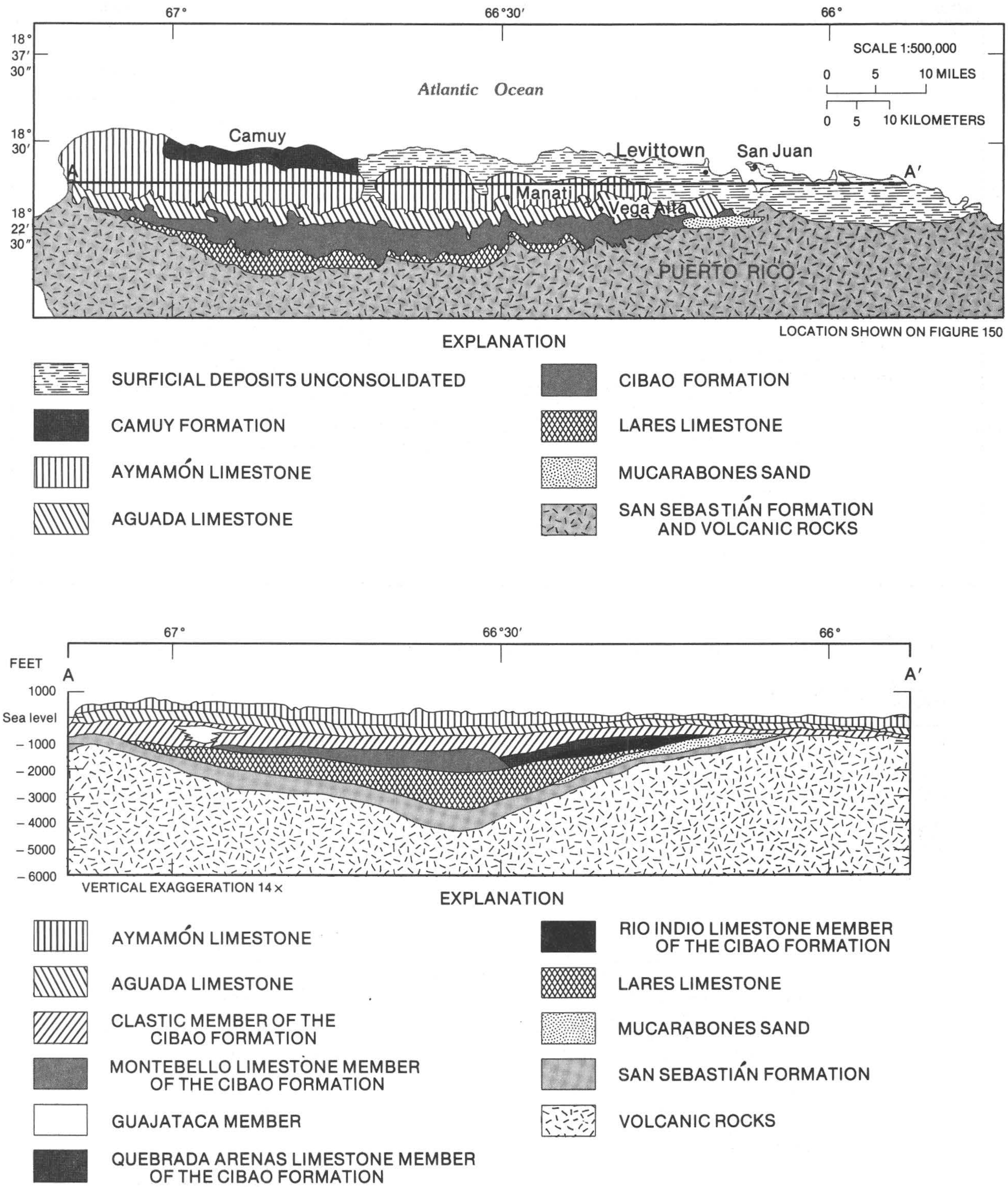


Figure 152.—Generalized subsurface geology of the North Coast Ground-Water Province of Puerto Rico.

Aymamon and Aguada Limestones. Alluvial deposits occur in the upper reaches of the Río Cibuco. The occurrence of an artesian system within the Cibao Formation and Lares Limestone is uncertain. Yields to wells in the water-table aquifer range from 200 to 2,000 gal/min. In the alluvial aquifers, well yields seldom exceed 150 gal/min. Withdrawals in the area exceed 20 Mgal/d. Seawater encroachment occurs in the coastal areas. A major chemical spill has affected a large segment of the aquifer near the town of Vega Alta.

5. Río de La Plata-Río de Bayamón.—The Aguada Limestone forms the principal freshwater aquifer in the area. Most of the overlying Aymamón Limestone contains saline water because of its proximity to the coast. The Cibao Formation, which underlies the Aguada Limestone, is composed mostly of clay and marl and yields little water to wells. The Mucarabones Sand, a local aquifer, underlies the Cibao Formation. Alluvial deposits as much as 200 feet thick occur along the entire section.

Yields to wells from the Aguada Limestone range from 200 to 600 gal/min. Wells tapping the Mucarabones Sand near metropolitan San Juan produce as much as 100 gal/min. The alluvial deposits contain freshwater as a thin lens (about 100 feet thick) floating above a saline water body. Total withdrawals from the area are about 10 Mgal/d. Seawater encroachment occurs in the Campanillas and Levittown areas.

6. Río de Bayamón-Río Grande.—The most productive aquifer in the metropolitan San Juan area is formed by clastic sediments and beach deposits of Tertiary and Quaternary age. The Mucarabones Sand underlies the Aymamón and Aguada Limestones on the western part of the section. Most of the Aymamón and Aguada Limestones contain saline water throughout the area.

Yields to wells from the unconsolidated deposits range from 30 to 150 gal/min. Wells tapping the Aguada Limestone in the eastern part can yield as much as 800 gal/min. Withdrawals are less than 10 Mgal/d. Urban development has reduced recharge to the ground-water system. Seawater encroachment occurs throughout the area.

Recharge to the aquifers along the north coast is mostly from precipitation. Discharge occurs mostly to streams in the coastal areas or to the seabed. Ground-water contamination with toxic organic compounds has been detected in water samples taken from 25 wells. Seawater encroachment as a result of overpumping is becoming widespread near the north coast.

South Coast Ground-Water Province

The South Coast Ground-Water Province consists of an alluvial plain and four alluvial valleys cut into Tertiary Limestones. Volcanic rocks underlie most of the eastern third of the south coast. The alluvial plain extends about 40 miles from Ponce to Patillas, averaging 3 miles in width from the coast to the foothills. The alluviated valleys from west to east include the areas of (1) Guánica, (2) Yauco, (3) Guayanilla, and (4) Tallaboa (fig. 151).

The alluvial plain consists of a series of coalescing fans formed by fast flowing intermittent streams. Sediments of Holocene to Pleistocene age are mostly coarse-grained sand and gravel in the apex of the major fans, becoming finer-grained toward the coast and interfluvial areas. They range in thickness from 100 feet at the shoreline toward the east to as much as 3,000 feet near Santa Isabel.

Yields to wells along the alluvial plain range from 150 gal/min in the interfluvial areas to as much as 3,000 gal/min near Guánica, averaging 500 gal/min. Wells penetrating the Ponce Limestone below alluvium can yield as much as 900 gal/min. Ground-water withdrawals from the alluvial plain are about 120 Mgal/d. Although no significant seawater encroachment occurs, persisting cones of depression in water levels may induce seawater contamination. Organic industrial chemicals have been detected in wells near Patillas, Guayama, and Juana Díaz.

The four alluvial valleys are very similar in composition and hydrogeologic characteristics. The alluvium is as much as 200 feet thick, mostly composed of sand and gravel. Streams flow perennialy in each valley. The Ponce Limestone and Juana Díaz Formation of Tertiary age underly the four valleys. Total thickness of the Tertiary units may be as much as 4,000 feet thick near the coast, thinning to 150 feet near Guánica.

Yields to wells tapping the alluvium range from 150 to 3,000 gal/min. Several wells also tapping the Ponce Limestone yield as much as 2,000 gal/min. However, water from the limestone is more mineralized than from the alluvium, probably due to the existence of connate seawater in the limestone. The Ponce Limestone and Juana Díaz Formation are not considered an aquifer in the outcrop areas. Total ground-water withdrawals from the four alluvial valleys are about 20 Mgal/d. The aquifers were overpumped through 1980, but presently are underutilized due to availability of surface water and reduction in irrigated acreage.

East Coast Ground-Water Province

The East Coast Ground-Water Province includes the alluvial valleys in Fajardo, Naguabo-Humacao, Yabucoa, and Maunabo (fig. 151). The alluvium in most of the areas consists of clay and silt, except in the Yabucoa and Maunabo Valleys, where coarse-grained sand predominates. The deposits range from 100 feet thick near Fajardo to 400 feet near Yabucoa.

Yields to wells range from 100 to 2,000 gal/min. The most productive wells occur in the Yabucoa Valley. Water withdrawals from the area are about 10 Mgal/d. High concentrations of manganese and iron occur in wells at Humacao, Yabucoa, and Maunabo.

West Coast Ground-Water Province

The West Coast Ground-Water Province includes the alluvial valleys of (1) Río Guanajibo, (2) Río Yaguez, (3) Río Grande de Añasco, and (4) Río Culebrinas (fig. 151). The alluvium is mostly detrital clay, sand, silt, and gravel of Quaternary age ranging in thickness from 60 feet to as much as 450 feet. Limestone deposits of Tertiary or Upper Cretaceous age and unknown thickness occur beneath the alluvial deposits in the western Río Guanajibo Valley. Yields to wells tapping the alluvium range from 50 gal/min to as much as 400 gal/min in the Río Guanajibo Valley. Wells tapping the limestone in the Río Guanajibo Valley can produce as much as 1,500 gal/min. Withdrawals are about 8 Mgal/d, most of it in the Río Guanajibo Valley.

Lajas Valley

The Lajas Valley (fig. 151) includes about 40 mi² of volcanic rocks of Cretaceous age in southeastern Puerto Rico. Limestone rocks of Cretaceous age (?) extend throughout the area overlain by unconsolidated deposits of Quaternary age. The thickness of the limestones is unknown, while the clastic deposits are as much as 200 feet thick. Most of the unconsolidated deposits are fine grained and yield little water to wells. Well yields from the limestone can be as much as 2,000 gal/min. However, most of the water in the valley is saline because of low recharge and of connate salty water. Actual withdrawals are less than 1 Mgal/d mostly from wells tapping alluvial fans along the edge of the valley.

Interior Province

The Interior Province (fig. 151) includes the central volcanic core of Puerto Rico, extending from west to east over an area of about 2,000 mi². The principal aquifers in the area occur in sedimentary, volcanic, and intrusive rocks of Late Cretaceous or Early Tertiary age. The most important aquifers are in the alluvial valleys of Caguas-Juncos and Cayey areas. The alluvium consists mostly of clay, sand, and gravel at Caguas-Juncos and clay with rock fragments at Cayey. Thickness of the unconsolidated deposits range from as much as 120 feet near Caguas to less than 35 feet in Cayey. Yields to wells average 200 gal/min in the Caguas-Juncos area and 150 gal/min in the Cayey area. Wells that tap fractured systems in the volcanic rocks can yield from 15 gal/min to as much as 450 gal/min. Withdrawals from the Interior Province are about 15 Mgal/d.

Kingshill Aquifer in St. Croix

The Island of St. Croix, with an area of 82 mi², is underlain by volcanic and sedimentary rocks of Late Cretaceous and Early Tertiary age. These rocks have been intruded by gabbro and diorite of Tertiary age. The east and west parts of the island are separated by a graben filled with volcanic ash overlain by marl and limestone known as the Kingshill Marl (fig. 153).

The Kingshill aquifer in St. Croix is the most important ground-water system in the U.S. Virgin Islands. It covers about 30 mi² in central St. Croix. Locally, it varies from marl to sandstone of Miocene to Oligocene age lying unconformably on bedrock. Alluvial deposits blanket the area to a thickness less than 20 feet except in stream courses where they are 200 feet thick.

Ground water in the Kingshill aquifer occurs under water table conditions in the alluvial deposits and Kingshill Marl. Yields to wells may be as much as 100 gal/min, but are usually about 30 gal/min. Withdrawals are about 800,000 gal/d. Saline water has affected numerous wells. Its source is either saltwater encroachment from the coast or upward coning of underlying saline water.

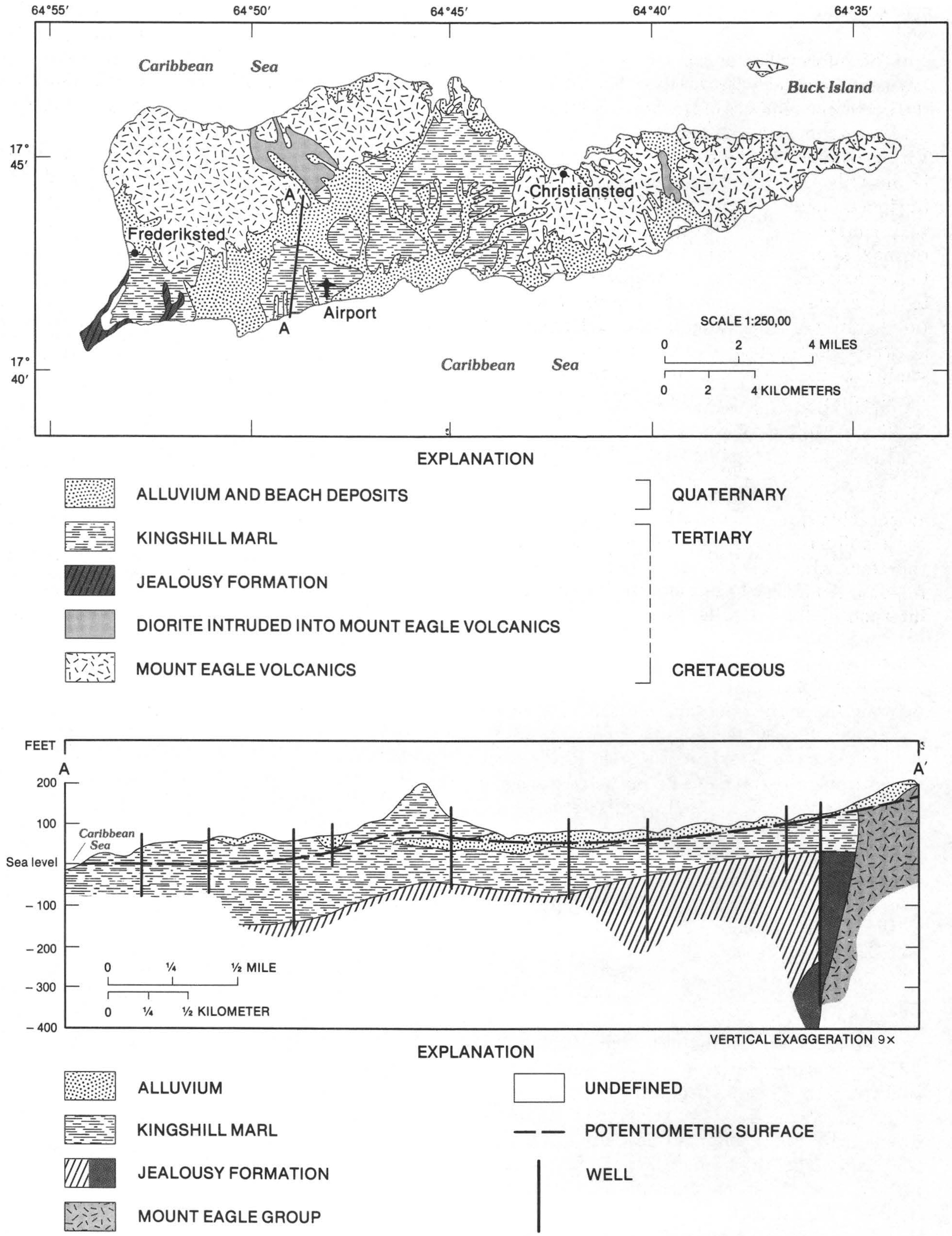


Figure 153.—Generalized subsurface geology of the Island of St. Croix, U.S. Virgin Islands, and a typical cross section of the Kingshill aquifer.

SUMMARY

The information above shows that the most important ground-water areas in Puerto Rico are the limestone aquifers of the north coast and the alluvial aquifers of the south coast. In the U.S. Virgin Islands, the Kingshill aquifer of St. Croix is the only significant ground-water area. The Caribbean Islands regional aquifer system study will concentrate in these three areas. The regional aquifer system study will entail compilation of existing information and development of computer-based flow models to understand the flow systems, to evaluate the potential for seawater encroachment near the coast, and to study the effects of change in irrigation patterns. An ongoing cooperative investigation with the Department of Natural Resources of the Commonwealth of Puerto Rico includes studying the occurrence and movement of ground water in the North Coast Ground-Water Province. This cooperative project involves developing flow models (Torres-Gonzalez and Wolansky, 1984); therefore, the effort of the Caribbean Islands regional aquifer system study on the north coast interacts closely with the cooperative study and will be concentrated on geochemistry as well as the interconnection between streams and the shallow water-table aquifers. The regional aquifer system study also will entail some exploratory drilling to fill the information gap in the data base.

In St. Croix, a ground-water flow model of the Kingshill aquifer will be developed and calibrated. Saltwater encroachment as a result of ground-water development also will be investigated.

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MICHIGAN BASIN REGIONAL AQUIFER-SYSTEM STUDY

By Lindsay A. Swain, Reston, Virginia

Michigan lies in a great structural bedrock depression—the Michigan Basin. The youngest bedrocks are at the center of the basin, and the oldest bedrocks crop out along the basin circumference. The important aquifers in the Michigan Basin are glacial aquifers and sandstone aquifers of the Marshall and Saginaw Formations of Paleozoic age. The Michigan Basin regional aquifer system study is designed to investigate the glacial aquifers and the underlying sandstone aquifers which cover about two-thirds of the lower peninsula of the State of Michigan (fig. 154). The study is scheduled for completion within 4 to 5 years.

Approximately 43 percent of Michigan's population, and nearly 100 percent of the State's domestic water supplies, are dependent on ground water (Bedell, 1982). In 1980, about 220 Mgal/d of ground water were pumped for water supplies (Solley and others, 1983). Of this, about 60 percent of the water was pumped from the glacial aquifers, 25 percent from the Saginaw and Marshall Formations, and the remaining 15 percent from other hydrogeologic units, such as the unnamed redbeds and Grand River Formation of Pennsylvanian age.

Nearly all of the study area is overlain by glacial deposits which range from a few inches to about 1,000 feet. Well yields in the glacial deposits are highly variable, depending on the thickness and characteristics of the deposits, whether the deposits are till, lacustrine sand and clay, or outwash sand and gravel. The distribution and characteristics of the glacial deposits are not well delineated nor is there a good understanding of the interconnection between the glacial deposits and underlying bedrock aquifers.

The sandstone of the Marshall and Saginaw Formations is the major aquifer to supply freshwater to the communities in the basin. In the Lansing-East Lansing area, where the Saginaw Formation is the principal source of water supply, a cone of depression has extended over 100 mi^2. Water levels near the pumping center are as much as 160 feet below the predevelopment level. In the Flint area, where both the Saginaw and Marshall Formations are used for public water supply, water having high chloride concentrations has migrated to the pumping centers; some of the well fields eventually were abandoned because of this migration. Areal and vertical distributions of saline water beneath the freshwater are not known; neither is the origin of the saline water. In places, the saline water may be as shallow as 100 feet below land surface; in other areas, the saline water may be more than 1,000 feet in depth. Where the Marshall Formation is confined, formations both above and below contain evaporite deposits. Water leaking through the confining units may dissolve the evaporites resulting in the accumulation of saline water in the Marshall Formation. Saline water in the Saginaw Formation seems different from that in the Marshall Formation and appears unrelated to the confining units. The predominant area where the Saginaw Formation contains saline water is in the Saginaw Bay area.

To ensure that sufficient ground water can be developed in the basin, it is critical to know the relation between development of ground

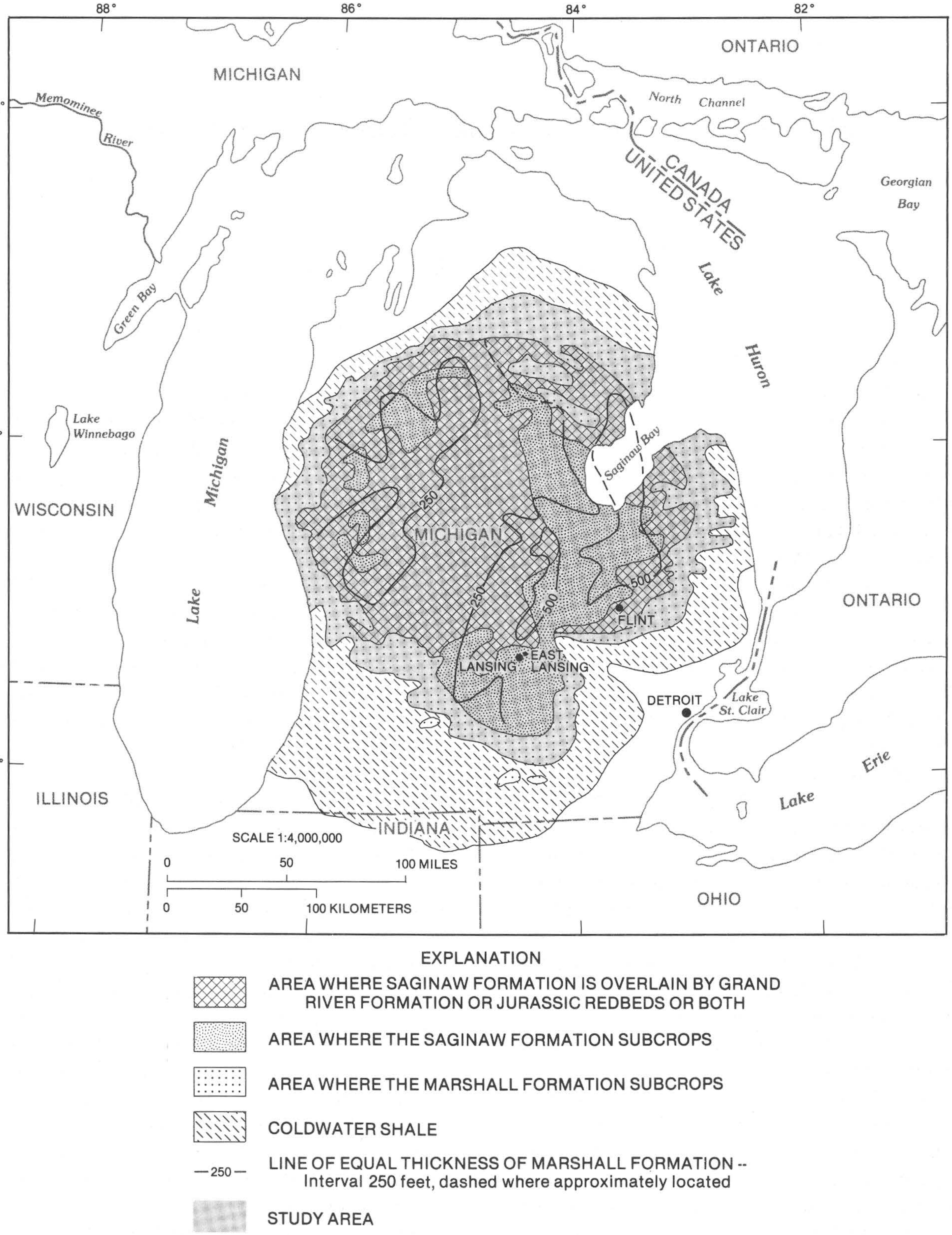

Figure 154.—Study area of the Michigan Basin regional aquifer system.

water in the Michigan Basin and the movement of the saline water. The origin of the saline water is also important as it may be critical to understand the movement of the saline water to the freshwater aquifer system due to development.

The study will use variable density flow models to evaluate all hypotheses and to understand the flow system, from land surface down to a major confining unit of the Coldwater Shale of early Mississippian age, before development and after development. The Coldwater Shale was chosen as the lower boundary of the flow system due to its low permeability. Therefore, the study area is bound by the contact between the Coldwater Shale and the Marshall Formation, the lowermost formation of the studied aquifer system (fig. 154). The established flow models will be used to evaluate the impact of future development.

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SOUTHERN CALIFORNIA ALLUVIAL BASINS REGIONAL AQUIFER-SYSTEM STUDY

By **Peter Martin, San Diego, California**

The southern California alluvial basins regional aquifer system study is scheduled for completion within 4 to 5 years. The study area is shown in figure 155. The study will be conducted in two parts: (1) the first part will produce a comprehensive bibliography and a report that will characterize regional ground-water conditions and identify the major ground-water problems and issues; and (2) the second part of the study will describe and categorize the regional geohydrology of the alluvial basins and analyze the major problems and issues that affect the utilization of ground water.

The geohydrology of the alluvial basins will be described using extensive data files and published reports. In basins where there is a scarcity of data, field investigations may be necessary to collect additional data. An attempt will be made to use geostatistics to extrapolate data from basins with extensive data to basins with limited data of the same category.

The study area includes 88 identified alluvial basins which will be grouped according to common characteristics and relationships. Elements that will be used to categorize the basins include: (1) depositional history of the basin fill, (2) ground-water flow characteristics, (3) recharge and discharge characteristics, and (4) water quality.

Three major water problems or issues have been selected for detailed investigations. They are: (1) saltwater intrusion in coastal basins, (2) flow between aquifer layers, and (3) the quantity and distributuion of recharge in coastal and desert basins.The study plans for the different investigations are discussed below.

Saltwater Intrusion In Coastal Basins—Nearly all of the alluvial coastal basin aquifers in southern California are affected by saltwater intrusion. Data on hydraulic characteristics and water quality in coastal basins will be analyzed to determine the natural geohydrologic and human activities that may affect the rate of saltwater intrusion. Field investigations may be necessary where there are little or no data. Geohydrologic and human activities that will be studied include: the presence of offshore faults and their effect on ground-water movement, the presence of permeable buried channel deposits in otherwise impermeable clay layers, tidal movement of saltwater into stream channels and subsequent vertical movement into underlying aquifers, the presence of multiple aquifer layers and the movement of saltwater between pumped and unpumped aquifers, the distribution and quantity of pumpage, and artificial recharge and other management alternatives used to control saltwater intrusion.The relative importance of the various controlling factors on movement of saltwater will be evaluated using mathematical ground-water flow models. Solute-transport models may be used to simulate saltwater movement due to pumping.

Flow Between Aquifer Layers—In general, there are at least three distinct aquifers in the coastal basins: (1) a shallow perched aquifer, that is virtually unconfined, in the upper part of alluvial deposits of Holocene age; (2) a principal freshwater aquifer, that is generally confined, comprising the lower part of the Holocene

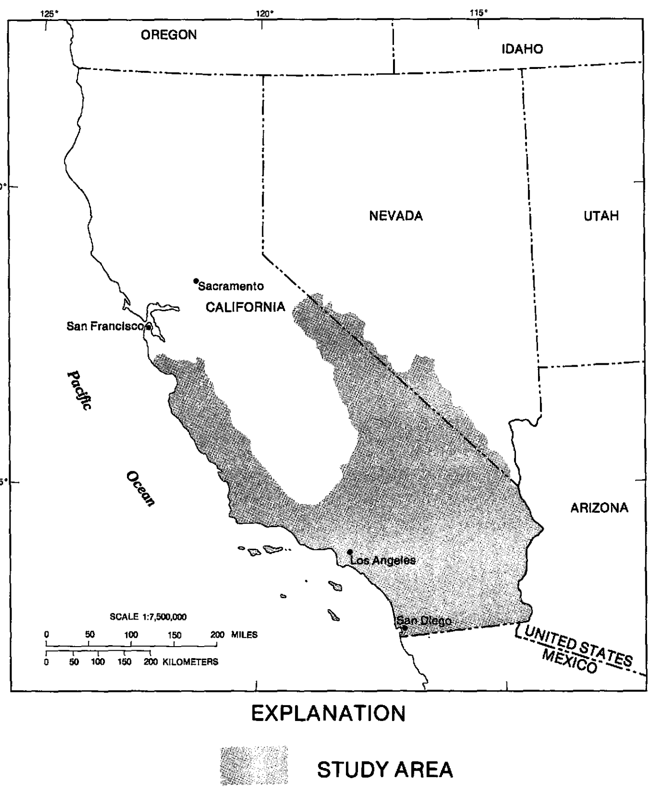


Figure 155.—Study area (shaded) of the Southern California Alluvial Basins regional aquifer system.

alluvium and nearly all sediments of Pleistocene to mid-Pliocene age; and (3) a deep saline-water aquifer that occurs in the consolidated rocks of Tertiary age and that underlies the principal freshwater aquifer. The desert basins also contain at least three distinct aquifers: (1) unconsolidated alluvial deposits of Pliocene to Holocene age that comprise the principal aquifer, (2) loosely consolidated sedimentry deposits of Tertiary age, and (3) consolidated rocks of pre-Tertiary age.

The mathematical models will be used to help gain an understanding of the ground-water flow between aquifer layers in coastal and desert basins. The aquifer system will be divided into several layers to simulate water movement among different water-bearing zones in coastal basins and among the interbedded alluvial and lacustrine deposits in desert basins. The models will be used to identify the parameters that have the greatest control over flow between aquifer layers and to determine the acceptable range in values of these parameters. The models will also be used to help evaluate the potential for water-quality degradation during different levels of stresses in the principal aquifer. Data required for the models will be compiled from existing reports and data files. If necessary, new data will be collected.

Geochemical techniques will also be used to help determine the amount of ground-water flow between aquifer layers. Concentrations of common dissolved ions, trace elements, and isotopes will be used to quantify the flow contributed to the sample wells from each aquifer layer. In some of the basins, this will require sample collection, especially for isotope analyses.

Quantity And Distribution Of Recharge In Coastal Or Desert Basins—In southern California, aquifer recharge is derived from: (1) infiltration of streamflow, (2) infiltration of precipitation, and (3) irrigation return flow. Recharge from streams is usually estimated from the decrease in streamflow between two gaging stations; however, in the arid and semiarid basins of southern California, this method is impractical due to the large number of small streams and the poorly integrated drainage network. Several methods that have been used to estimate infiltration of streamflow include: (1) rainfall runoff relationship, (2) basin characteristics, (3) channel geometry, and (4) model simulation. The differences between methods and the effectiveness of the methods will be evaluated with field observations.

Infiltration of precipitation in coastal basins is usually estimated from rainfall-infiltration functions developed for a particular basin. The transferability of the rainfall-infiltration functions, from basin to basin within each category of the basins, will be evaluated in depth. Direct infiltration of precipitation in desert basins is usually considered negligible. To determine qualitatively if recharge from precipitation is occuring in the desert, soil core samples will be analyzed to determine tritium concentrations with depth.

The quantity and quality of irrigation return flow are major problems affecting the utilization of ground water in southern California. The use of water for irrigation typically adds dissolved constituents in ground water. Consequently, as agricultural activities increase in a basin, the quality of the ground water is degraded. Problems common to agricultural areas include high concentrations of nitrate, dissolved solids, and pesticides. This study will delineate areas where agricultural activities have significantly degraded the ground-water quality. An attempt will be made to relate the degree of contamination in the principal aquifer to the thickness of the unsaturated zone and the history of irrigation in selected basins. In basins where there is a large unsaturated zone and the principal aquifer has not been contaminated, the perched water table will be sampled to determine the vertical migration of irrigation return flow.

Note from the editor—*The activity of the Southern California Alluvial Basin Regional Aquifer-System Study was temporarily suspended in 1985. All resources related to this study were transferred to Central Valley, California, to intensify a phase II (followup) study.*

PHASE II
REGIONAL AQUIFER-SYSTEM ANALYSIS PROJECTS
(followup studies)

CENTRAL VALLEY REGIONAL AQUIFER SYSTEM, CALIFORNIA, PHASE II STUDY

By Robert J. Gilliom, Sacramento, California

A regional study of the Central Valley aquifer system conducted during 1978-82 identified several water-resource issues that are important to water management in the area but could not be adequately addressed within the regional scope of the study. One of the key water issues for the Central Valley is the effect of extensive irrigation on the water quality and flow system of the aquifers. This issue has recently come to the forefront of public, political, and scientific attention, with the discovery that subsurface agricultural drainage water from certain parts of the western San Joaquin Valley contains undesirably high concentrations of some trace elements, particularly selenium.

The U.S. Geological Survey, in cooperation with the U.S. Bureau of Reclamation, is conducting a comprehensive regional study of the sources, occurrence, and movement of selenium and other trace elements in the hydrologic system of the western San Joaquin Valley. Phase II of the Central Valley regional aquifer system study is a complementary part of the overall study plan for the San Joaquin Valley which specifically focuses on (1) detailed evaluation of the hydrogeology of the aquifer system at a shallow depth (less than 200 feet) that may be affected by poor quality of irrigation return flow, and (2) a regional study of the ground-water quality of the entire San Joaquin Valley—both eastern and western parts.

The hydrogeologic evaluation of the aquifer system at a shallow depth in the western San Joaquin Valley is the main focus of Phase II. The study area with the poorest quality of irrigation return flow will consist of a ground-water flow model that relies on the extensive data collections and geochemical process studies that are separately funded as parts of the cooperative study with the U.S. Bureau of Reclamation. Simulation will be used to help characterize the shallow ground-water flow system, to provide quantitative estimates of rates and directions of flow including irrigation return flow, and to help understand subsurface movement of selenium and other chemical constituents. Preliminary ground-water flow models will be constructed at two scales: (1) agricultural field, and (2) alluvial fan (the Panoche fan) which covers about 200-300 mi^2. Each model will provide insight into (1) interaction of the shallow water-bearing zone and the regional aquifer system, and (2) processes responsible for subsurface movement of chemical constituents.

The regional study of ground-water quality in the San Joaquin Valley is the main focus of the Phase II study. The study will be developed by using recently collected data for the western part of the Valley and by extending the same type of data-collection effort for the eastern part of the Valley. The study will focus on major ions, and trace elements, with selective analyses for pesticides. Analyses of data for these sampling programs will involve statistical analyses of the areal distribution of chemical constituents in relation to hydrogeologic and environmental factors potentially affecting the ground-water quality, and application of geochemical models to assess the measured chemistry in relation to estimated equilibrium concentrations.

FLORIDAN REGIONAL AQUIFER SYSTEM, PHASE II STUDY

By Peter W. Bush, James A. Miller, Atlanta, Georgia, and Morris L. Maslia, Doraville, Georgia

INTRODUCTION

The Floridan regional aquifer-system study investigated and described the flow system from a regional and subregional perspective. During the course of that study, local aspects of the system that merited continued or more detailed work were noted but were not dealt with in order to fulfill the broader objectives of the initial study.

The purpose of the Floridan regional aquifer-system phase II study is to investigate some of these local aspects. The phase II study was started in 1983 and is scheduled for completion in 1986. Four investigations are part of the phase II study. The locations of these investigations are shown in figure 156. A brief description of each activity follows.

PETROGRAPHIC STUDY, CENTRAL FLORIDA

A 2,000-foot test well was continuously cored in central Florida during the initial study. As part of the phase II studies, petrography of the cores was studied in detail to determine the effect that original carbonate depositional environment and postdepositional mineralogic change (diagenesis) have had on the porosity and permeability of the Floridan aquifer system. Thin sections of the core were examined with a petrographic microscope. X-ray analysis and scanning electron microscopy were used on selected core samples, and laboratory values of porosity and permeability were obtained from selected core intervals.

Results of this petrographic study show that the primary control on carbonate porosity is the original depositional environment of the rock. Six general depositional environments were identified; only one high-energy, relatively coarse-grained limestone (foraminiferal grainstone) showed appreciable primary porosity and permeability. Dolomite in the core occurs both as a fine-grained primary deposit and as a fine to coarse-crystalline replacement (diagenetic) material. The replacement dolomites show considerable porosity and permeability, indicating that diagenesis is a secondary, but very important, factor in determining carbonate permeability. Anhydrite and gypsum that formed in a tidal-flat environment fill practically all the pore space in the carbonate rocks in certain horizons. Replacement dolomites appear to have formed in a mixed meteoric-seawater zone, during several episodes of dolomitization.

Results of this petrographic study have been summarized in the article "Petrology of Lower and Middle Eocene Carbonate Rocks, Floridan aquifer, central Florida" by P. A. Thayer and J. A. Miller, published in volume 34 of Transactions of the Gulf Coast Association of Geological Societies (1984).

SALTWATER MOVEMENT STUDY, HILTON HEAD ISLAND, SOUTH CAROLINA

The potential exists for saltwater contamination of coastal freshwater supplies in areas where the Floridan aquifer system is unconfined or loosely confined. One such area is Hilton Head Island, a rapidly growing resort area along the southern coast of South Carolina. The freshwater supplies of the island are obtained from wells tapping the Upper Floridan aquifer. In 1984, water quality in the upper part of the Upper Floridan on the island was good. However, concentrations of chloride in water in the lower part of the Upper Floridan are greater than 250 milligrams per liter (mg/L), and water throughout the Upper

Floridan 3 or 4 miles northeast of the island ranges from slightly saline to very saline with depth.

The Upper Floridan aquifer in the vicinity of Hilton Head Island is loosely confined by clastic materials, primarily sand and clay. The clastic materials thin a few miles northeast of Hilton Head Island, and limestone of the Upper Floridan may crop out in the sea. Confinement by clay increases southeast and southwestward from the island.

Before ground-water development, the configuration of the Upper Floridan potentiometric surface suggests that the direction of natural flow

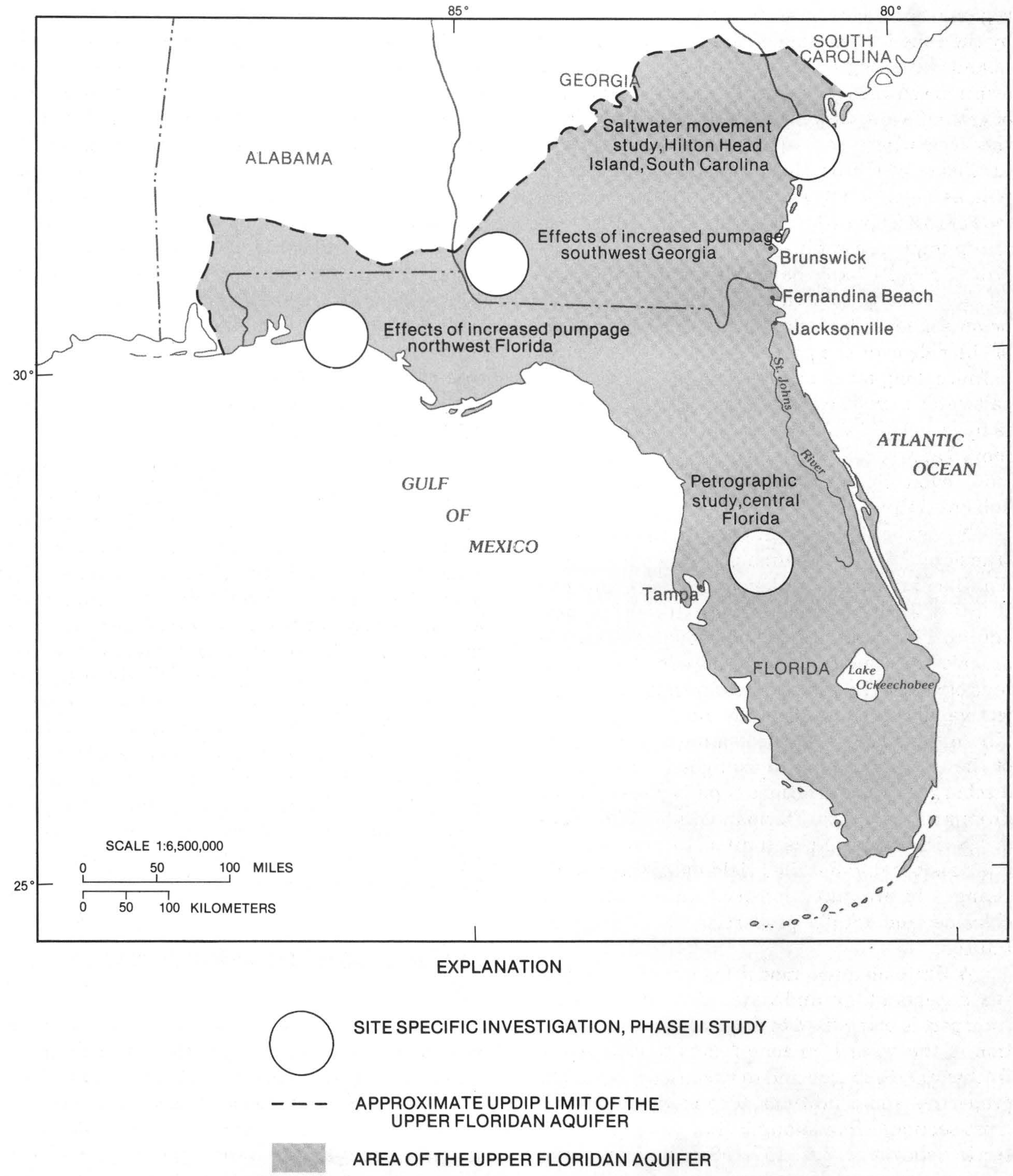


Figure 156.—Site-specific investigations of the Floridan aquifer system, phase II study.

in the aquifer in the vicinity of Hilton Head Island was east-northeastward, from the Savannah area across Hilton Head Island to a natural discharge area coincident with the loosely confined part of the Upper Floridan northeast of the island (fig. 157).

Large withdrawals from the Upper Floridan centered in Savannah have created a deep regional cone of depression. The area influenced by the cone includes the vicinity of Hilton Head Island. Pumping at Savannah has reversed the original seaward gradient. The present direction of ground-water flow is west-southwestward from the area where the aquifer is loosely confined northeast of Hilton Head Island toward Savannah, as suggested by the 1980 potentiometric surface of the Upper Floridan (fig. 158). The Savannah pumping, combined with local pumping on Hilton Head Island, have reduced heads on the island about 10 to 30 feet. By 1980, heads were below sea level, but monitoring of concentrations of chloride over the past 25 years has not shown definite long-term changes in the freshwater-saltwater transition zone beneath the island. It is hypothesized that the transition zone has not moved appreciably from its predevelopment position, even though heads have been lowered substantially.

Nevertheless, the transition zone is in a transient state, presumably seeking a new equilibrium position. A key question is, how fast is the transition zone moving toward a new equilibrium position? Is the movement on the order of geologic time, or hundreds of years, or tens of years? To answer these questions, the objective of the phase II study in the vicinity of Hilton Head Island is to determine the sensitivity of the transition zone to changes in freshwater discharge and to variations of properties and conditions in the Upper Floridan aquifer. The scope of the investigation is limited to a sensitivity analysis because detailed field data (specifically changes in pressure, density, concentrations of chloride, and aquifer properties with depth) are limited.

A finite-element model for simulating density-dependent ground-water flow with solute transport is being used to determine how the position of the transition zone reacts to changes in freshwater discharge and to variations in aquifer properties and conditions. The model simulates cross-sectional flow along a flow path (proposed section shown in figs. 157-158). The first series of simulations are of predevelopment conditions. These simulations are used to test which aquifer properties and boundary conditions control the flow system, and test whether variation of these factors within reasonable limits can produce a simulated flow system compatible with the predevelopment flow system simulated during the initial study of the aquifer system.

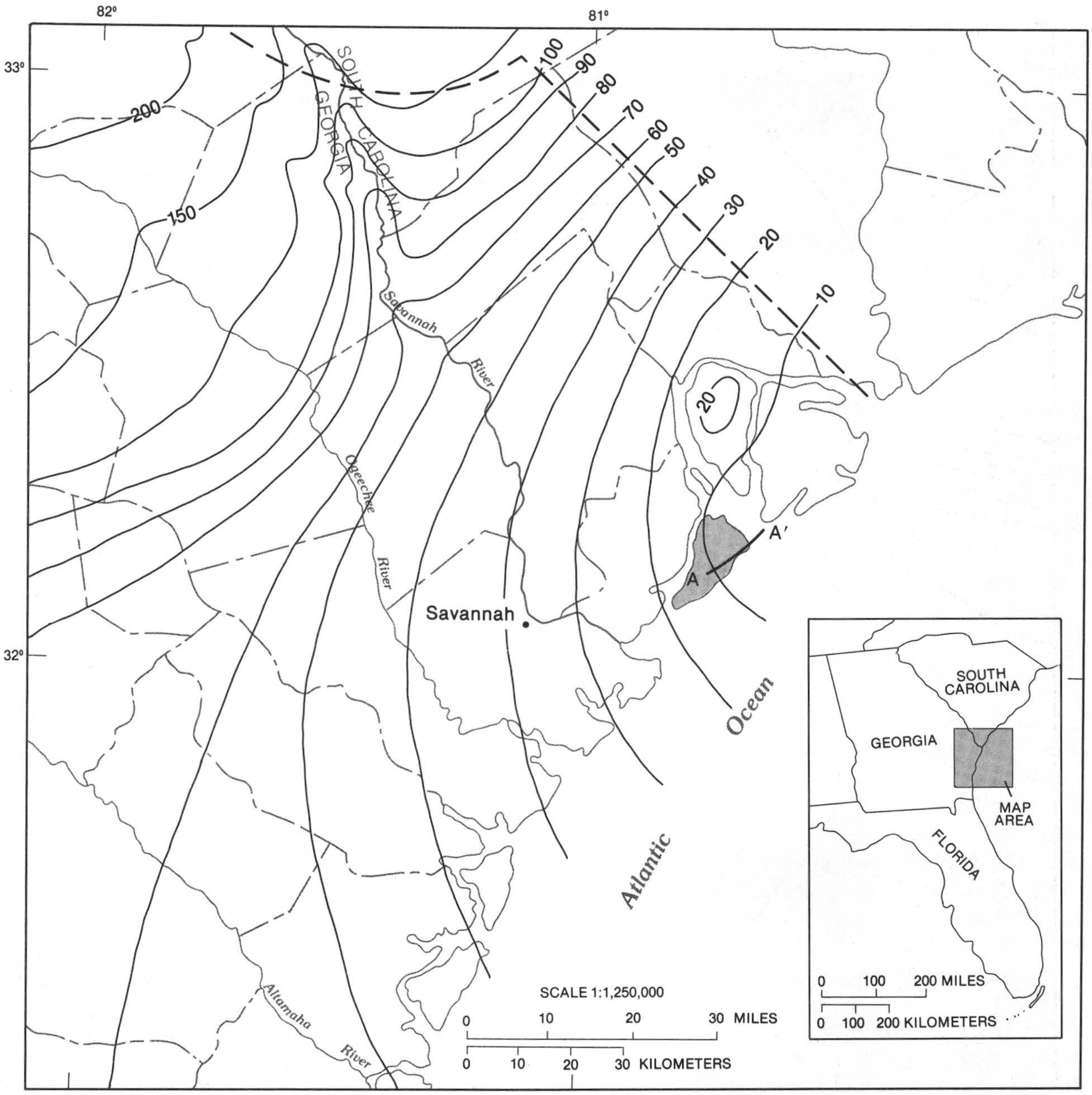
If the series of predevelopment simulations are successful, the results will be used as initial conditions for transient simulations that include solute transport (chlorides) over a timespan equivalent to that of ground-water development, or about 80 to 90 years. The 1984 boundary conditions will be used in a series of pumping periods over the simulated timespan. These simulations will test the hypothesis that it is possible for the transition zone to have remained virtually in its predevelopment position even as flow conditions changed during the 80 to 90 years of development in the area.

EFFECTS OF INCREASED PUMPAGE, SOUTHWEST GEORGIA AND NORTHWEST FLORIDA

The Upper Floridan aquifer in southwest Georgia is relatively thin (generally less than 200 feet), has varying transmissivity, and is loosely confined. In contrast, the Upper Floridan in northwest Florida is relatively thick (generally 400 to 500 feet), has much lower transmissivity, and is tightly confined. Ground-water development in both of these areas is increasing rapidly. Based on the initial study of the Floridan aquifer system in these two areas, a paper will be written to compare each area's response to the 1980 and projected future ground-water development. Emphasis will be placed on discussion of the different flow regimens as well as the methods of analysis necessitated by the very different hydrologic conditions that occur in southwest Georgia and northwest Florida.

REGIONAL POTENTIOMETRIC-SURFACE MAP, 1985

As part of the initial study, an aquifer-wide potentiometric-surface map of the Upper Floridan was constructed (Johnston and others, 1981). The map was based on water-level and pressure-head measurements made simultaneously in May 1980 in more than 2,700 wells tapping the Upper Floridan. It is planned that a similar



EXPLANATION

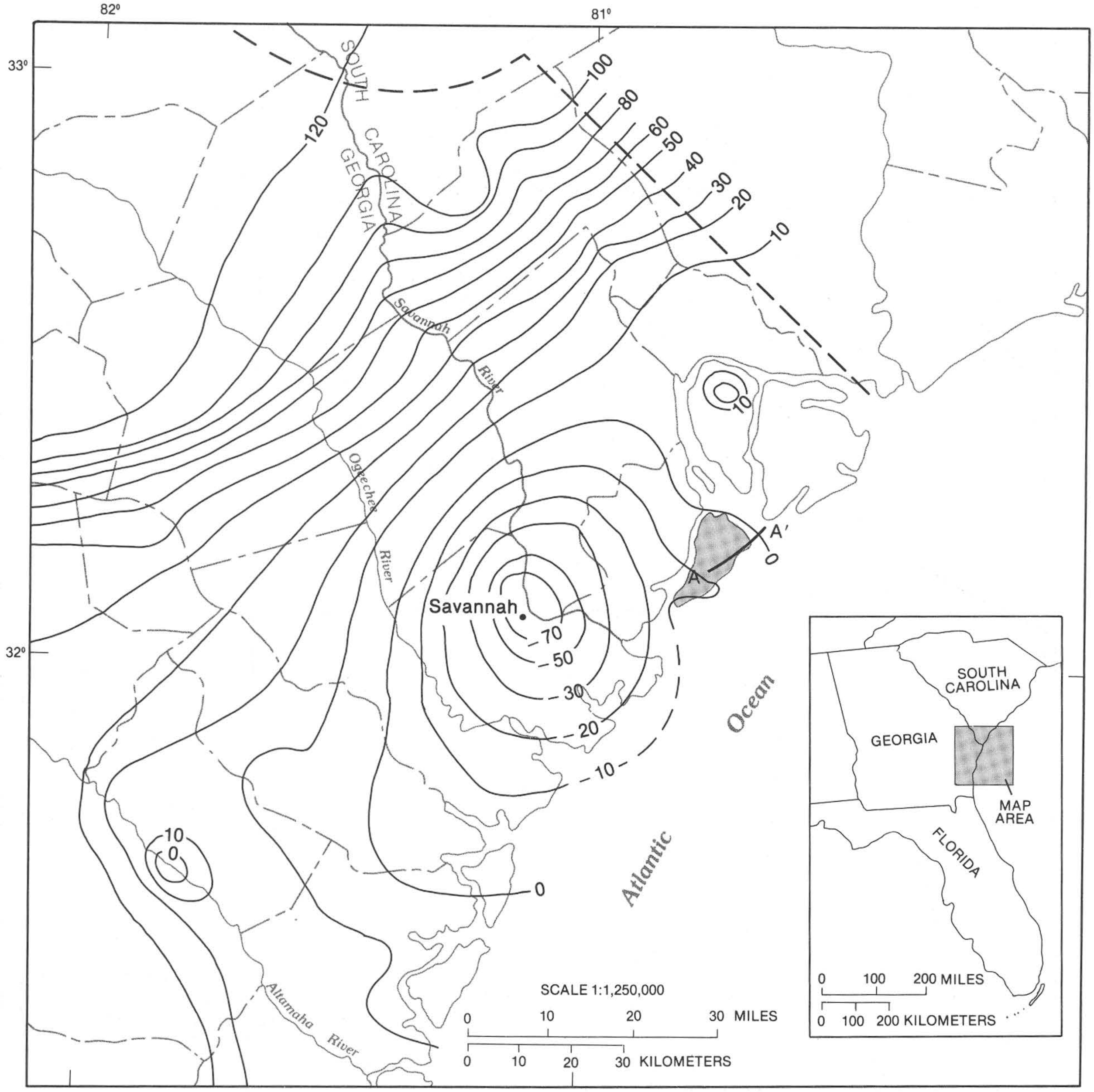
AREA OF HILTON HEAD ISLAND

—20— ESTIMATED PREDEVELOPMENT POTENTIOMETRIC CONTOUR -- Shows altitude of potentiometric surface of the Upper Floridan aquifer. Contour interval, in feet, is variable. Datum is sea level

— — APPROXIMATE UPDIP LIMIT OF THE UPPER FLORIDAN AQUIFER

A——A′ PROPOSED LINE OF SECTION FOR SIMULATION

Figure 157.—Estimated predevelopment potentiometric surface of the Upper Floridan aquifer in the area of Hilton Head Island and vicinity; A-A′ is the proposed line of section for simulation.



EXPLANATION

AREA OF HILTON HEAD ISLAND

—20— MAY 1980 POTENTIOMETRIC CONTOUR -- Shows altitude of potentiometric surface of the Upper Floridan aquifer. Contour interval, in feet, is variable. Datum is sea level

— — APPROXIMATE UPDIP LIMIT OF THE UPPER FLORIDAN AQUIFER

A——A′ PROPOSED LINE OF SECTION FOR SIMULATION

Figure 158.—May 1980 potentiometric surface of the Upper Floridan aquifer in the area of Hilton Head Island; A-A′ is the proposed line of section for simulation.

potentiometric-surface map will be constructed for May 1985 based on a similar mass measurement of water levels in Alabama, Florida, Georgia, and South Carolina. The May 1985 potentiometric-surface map will be evaluated and compared with the May 1980 map.

PUBLISHED REPORT

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REFERENCE CITED

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HIGH PLAINS REGIONAL AQUIFER SYSTEM, PHASE II STUDY

By John B. Weeks, Denver Colorado

INTRODUCTION

The initial High Plains regional aquifer study provided a regional description of the aquifer system and calibrated regional ground-water flow models. The models were calibrated on the basis of water-level changes from predevelopment to 1980. Pumpage and irrigation return flow are two poorly known factors, however, they are critical for simulation. An indirect method for estimating pumpage was developed during the initial study; irrigation return flow was adjusted during model calibrations. The accuracy of pumpage estimates and an independent estimate of irrigation return flow are essential to developing more accurately calibrated flow models.

The ground-water flow models developed during the initial study are capable of projecting future water levels in the aquifer resulting from the strategies proposed by a study of the Economic Development Administration (EDA) of the U.S. Department of Commerce. However, the accuracy of the water- level projections cannot be evaluated unless the accuracy of information on pumpage and irrigation return flow are evaluated, which was not pursued during the initial study. For this reason, a phase II study was started in 1982 and is scheduled for completion in 1986.

EVALUATION OF PUMPAGE AND IRRIGATION RETURN FLOW

During the initial phase of the High Plains regional aquifer-system study, pumpage was estimated on the basis of reported crop acreage, calculated irrigation demand, and reported irrigation efficiency. Pumpage in 1980 was estimated using irrigated acreage determined from Landsat imagery associated with a random sample of measured pumpage and irrigated acreage (Heimes and Luckey, 1983). The procedure was found to be inexpensive for estimating pumpage for large areas; but, the accuracy of the procedure has not been determined.

The estimates of historical pumpage were used in development of the ground-water flow models during the initial study. During model development, part of the simulated historical pumpage was allowed to return to the aquifer as irrigation return flow. The percentage of pumpage that became irrigation return flow varied with time and location, but return flow ranged from 30 to 55 percent of the gross pumpage. Although the percentage of return flow is consistent with reported estimates, a method to estimate the return flow independently is needed to verify the results obtained from model analyses.

Based on existing hydrologic data, two areas were selected to evaluate pumpage and irrigation return flow (fig. 159). One area includes Castro and Parmer Counties in the southern High Plains of Texas. These two counties were selected because the Texas Department of Water Resources and the Underground Water Conservation District No. 1 have long-term annual water-level records. Digital Landsat data are also available for the area since 1974. The second area is in the Upper Republican Natural Resources District (NRD) which includes Chase, Perkins, and Dundy Counties in Nebraska. These three counties were selected because the NRD has installed flowmeters on all irrigation wells. Meter installation began in 1978 and was completed in 1980. Digital Landsat data are available since 1974. The period of analysis for each area will be for about 10 years, 1974-84, depending on the availability of data.

Pumpage Estimates

Pumpage data were collected during the 1983 and 1984 irrigation seasons from about 50 irrigation systems in each of the two areas using portable flowmeters and time-of-operation sensors. In Nebraska, pumpage from nearly all wells was measured by the NRD using inline volumetric flowmeters. The NRD data will be used to measure the accuracy of the pumpage data collected using portable flowmeters. In Texas, no metering program exists to measure pumpage. Therefore, 12 volumetric inline flowmeters were installed as calibration sites for the portable flowmeters.

Pumpage and irrigated acreage data, collected during 1983 and 1984, are being used to estimate total pumpage in each area for the entire period, 1974-84. The estimated total pumpage will then be compared to the measured pumpage in Nebraska to evaluate the accuracy of the estimates. In Nebraska, the total 1983 pumpage measured by portable flowmeters and time-of-operation sensors at 52 sites was within 3 percent of the total pumpage measured by inline flowmeters. This indicates that reasonably accurate measurements can be made with portable flowmeters and time-of-operation sensors.

Irrigated acreage is being mapped in both areas for selected years using Landsat imagery. Irrigated acreage obtained from Landsat is being compiled by the National Mapping Division (NMD) of the U.S. Geological Survey for test areas in Nebraska and Texas for 1974, 1976, 1978, 1980, 1982, and 1983. The acreage data are being digitized and stored in a computerized data base.

Estimates of total pumpage for selected years during 1974-83 are being made for both test areas by estimating the irrigation application rate and multiplying that number by the irrigated acreage determined from Landsat imagery. Two approaches to estimate application rate for the years in which measured data are not available are being evaluated: (1) development of a time series using measured application rates, and (2) development of relations between measured application rates and the calculated consumptive-use.

Irrigation Return-Flow Estimates

Irrigation return flow is being estimated from the difference between pumpage and change in aquifer storage. Change in aquifer storage is being determined from maps of water-level change and specific yield. Maps of water-level change are being prepared for the period 1975 through 1983 for both test areas. Drillers' logs are being used to estimate and prepare maps of specific yield. On a volumetric basis, the difference between total pumpage and change in aquifer storage is assumed to be the irrigation return flow plus changes in recharge and discharge due to natural stresses, if any. The estimates of pumpage and irrigation return flow will be compared to values resulting from the calibrated ground-water flow models.

Maps of specific yield are being prepared for both areas. For the area in Nebraska, lithologic logs were compiled from 532 wells which represent about 18 percent of the estimated 3,000 wells in the area. For the area in Texas, lithologic logs were compiled for Castro, Parmer, and adjoining counties. For Castro and Parmer Counties, 996 logs were processed which represent about 12 percent of the estimated 8,000 wells in the area. For bordering areas in Texas, about 300 additional logs were processed. The lithologic logs were processed by a computer program and averaged specific yield was determined for each log. The average specific yield for each county was estimated to be:

Chase County, NE	12.3 percent
Dundy County, NE	14.7 percent
Perkins County, NE	12.2 percent
Castro County, TX	18.4 percent
Parmer County, TX	19.5 percent

A map of water-level change for the period 1978-82 for the test area in Nebraska was prepared using data from 384 wells. Maps of water-level change for the period 1975-84 for the area in Texas were prepared using data from 125 wells. Maps of water-level change were digitized and stored in a data base. Change in aquifer storage is being computed from that data base.

EVALUATION OF EFFECTS OF FUTURE PUMPING

The results of the EDA study of the High Plains have been published in a report by High Plains Associates (1982); recommendations to the U.S. Congress also have been made by the High Plains Study Council (1982). Additionally, more detailed reports have been published by investigators in each of the six individual States. These reports will be used to estimate future pumpage for the 40-year period from 1980 to 2020.

Ground-water flow models of the High Plains aquifer are being used to evaluate the effects of pumpage on the aquifer from 1980 to 2020. Simulated results will be presented in maps of water-level and saturated-thickness changes.

Finite-difference models of the northern, central, and southern High Plains aquifers have been developed and calibrated during the initial study. During the phase II study, these calibrated models are being used to project future water

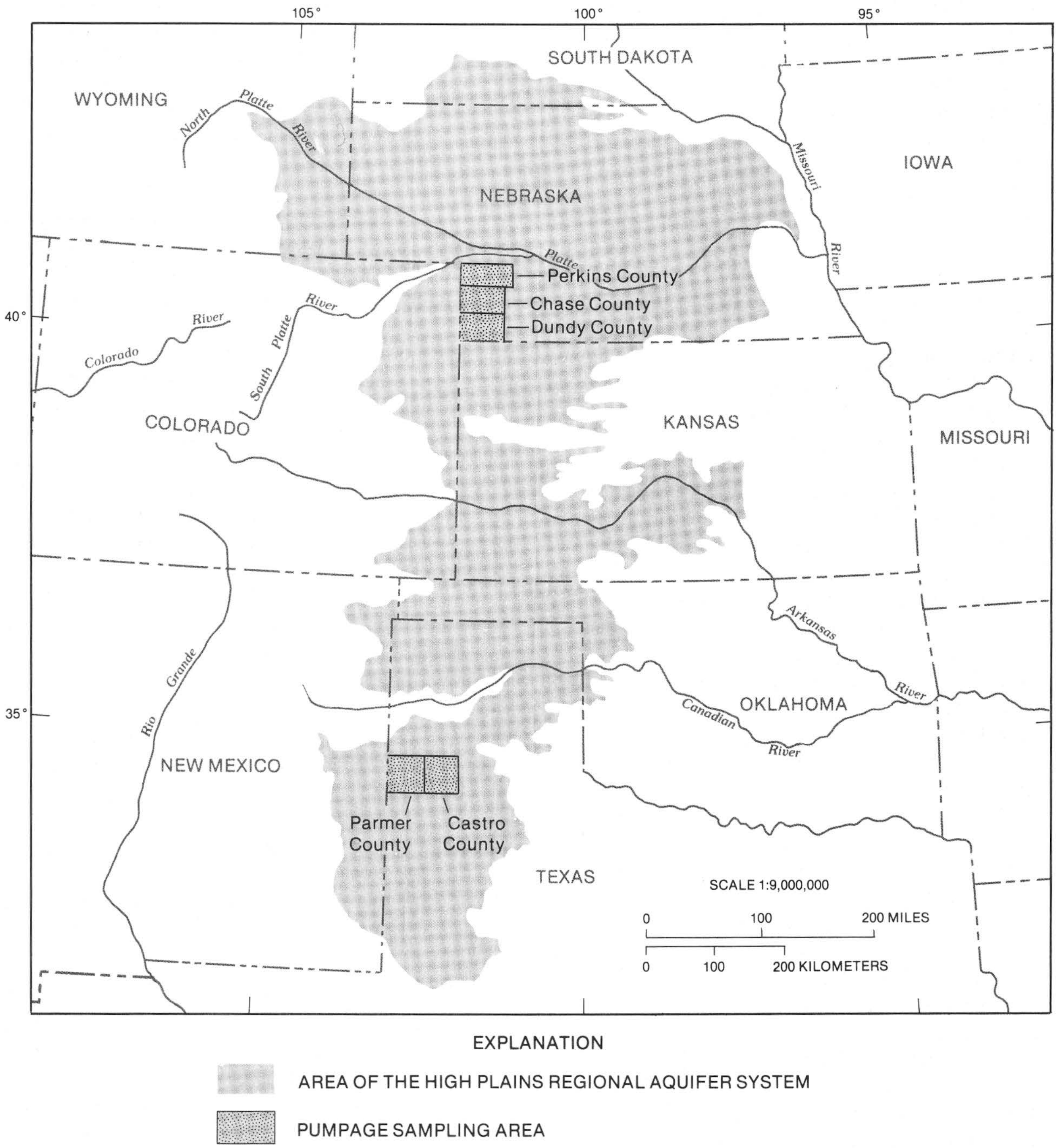

Figure 159.—Study area of the High Plains regional aquifer system, phase II study.

levels under three different development strategies. The strategies are: (1) a baseline condition where future water use is based on the projected need, (2) a voluntary-reduction strategy where water use is slightly reduced from the projected need, and (3) a mandatory-reduction strategy where water use is greatly reduced from the projected need.

Projections of water-level and saturated-thickness changes have been completed for the three strategies in the southern High Plains. For the baseline condition, the total projected pumpage is 113 million acre-ft between 1980 and 2020. The projected pumpage for the same period for the voluntary-reduction strategy is 109 million acre-ft and for the mandatory-reduction strategy is 88 million acre-ft. Under the baseline condition, much of the southern High Plains would experience severe water-level declines; by the year 2020, much of the area in the southern High Plains would have less than 25 feet of saturated thickness remaining. Similar conditions would occur under the voluntary-reduction strategy because the pumpage is only slightly less. However, under the mandatory-reduction strategy, water-level declines would be less severe and approximately 124 million acre-ft of drainable water would remain in aquifer storage by the year 2020. This remaining water in aquifer storage is about 40 percent of the total volume pumped from the aquifer from predevelopment to the year 2020 under the baseline condition.

PUBLISHED REPORT

Stephens, D. M., Heimes, F. J., and Luckey, R. R., 1984, Irrigation data from Chase, Dundy, and Perkins Counties, southwestern Nebraska, 1983: U.S. Geological Survey Open-File Report 84-471, 31 p.

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High Plains Associates, 1982, Six-state High Plains Ogallala aquifer regional resources study, a report to the U.S. Department of Commerce and the High Plains Study Council: High Plains Associates, Austin, TX, 438 p.

High Plains Study Council, 1982, A summary of results of the Ogallala aquifer regional study, with recommendations to the Secretary of Commerce and Congress: High Plains Study Council, 61 p.

SNAKE RIVER PLAIN REGIONAL AQUIFER SYSTEM, PHASE II STUDY

By Gerald F. Lindholm, Boise, Idaho

INTRODUCTION

During phase I of the Snake River Plain regional aquifer-system study, several areas were identified for more detailed study. Long-term regional hydrologic changes were successfully simulated using quasi three-dimensional ground-water flow models. However, in key local areas the desired degree of understanding was not satisfactorily achieved with the large-scale regional flow models. Data need to be collected and incorporated into smaller scale local flow models that will be developed during phase II studies. A stream-aquifer model of the eastern Snake River Plain is also scheduled to be developed. The following are brief descriptions of these key studies.

BIG LOST RIVER BASIN

The Big Lost River basin contributes major quantities of water to the eastern part of the Snake River Plain (fig. 160). Estimates of water yield from most tributary basins have been made during the phase I study. However, all estimates are long-term annual averages that do not account for short-term variations in precipitation and irrigation. Since the 1950's, concern has increased about declines in ground-water levels and decreases in ground-water discharge from the eastern Snake River Plain. Consequently, there is concern about the amounts of water available to the Plain from tributary drainage basins. This phase II study is to address those concerns. A ground-water flow model will be used to test different hypotheses concerning water yield to the Snake River Plain from the Big Lost River basin.

AMERICAN FALLS RESERVOIR AREA

About 1.8 million acre-ft of ground water are discharged annually to the Snake River and American Falls Reservoir between Blackfoot and Neeley (figs. 49 and 160). In the 10-mile reach from Blackfoot to the high-water line of the reservoir, the Snake and Portneuf Rivers receive large quantities of ground water, largely from springs. Discharge data suggest that gains have remained relatively stable since 1912, although irrigation in and upstream from that area has increased significantly. The American Falls area was included in the regional flow model developed during the phase I study. Although a reasonable transient simulation was achieved, the level of detail was inadequate to answer specific questions about the effect of irrigation on ground-water and surface-water relations. A detailed quasi three-dimensional model to simulate ground-water movement in the American Falls Reservoir area will be developed and calibrated. Data will be collected in 1985 on irrigation diversions, return flows, water levels, river gains, pumpage, and water chemistry. Test holes will be drilled and piezometers installed to help define the flow system locally.

GROUND-WATER FLOW MODEL COMBINING THE EASTERN AND WESTERN PARTS OF THE SNAKE RIVER PLAIN

During the phase I study, separate ground-water flow models were developed for the eastern and western parts of the Snake River Plain. The dividing line includes about 40 miles of the Snake River where numerous springs discharge several thousand cubic feet per second from the eastern part of the Snake River Plain (figs. 49 and 160). Water budget analyses indicate that virtually all water in both the eastern and western parts of the Plain can be accounted for without underflow along the dividing line. The Snake River in this area is a regional sink and gains water from both the east and west sides of the river. The hypothesis that there is no underflow along the dividing line needs to be tested. A ground-water flow model that combines parts of the eastern and western Plain would be developed and used for the test. Initial parameter values, including boundary flux, would be obtained from models developed during the phase I study. A fine mesh and several layers would be needed to increase model sensitivity. During the phase I study, springs were simulated either as constant flux or leaky boundary with constant head representing the spring levels. In the phase II study, springs would be simulated as a seepage face. After model calibration, sensitivity tests would be run to determine the validity of assumptions made during the phase I study.

GEOCHEMICAL INVESTIGATIONS OF SOURCES OF WATER TO THE THOUSAND SPRINGS

Sources of water discharging as springs in the thousand Springs area are inadequately defined. During the phase I study, maps and sections were

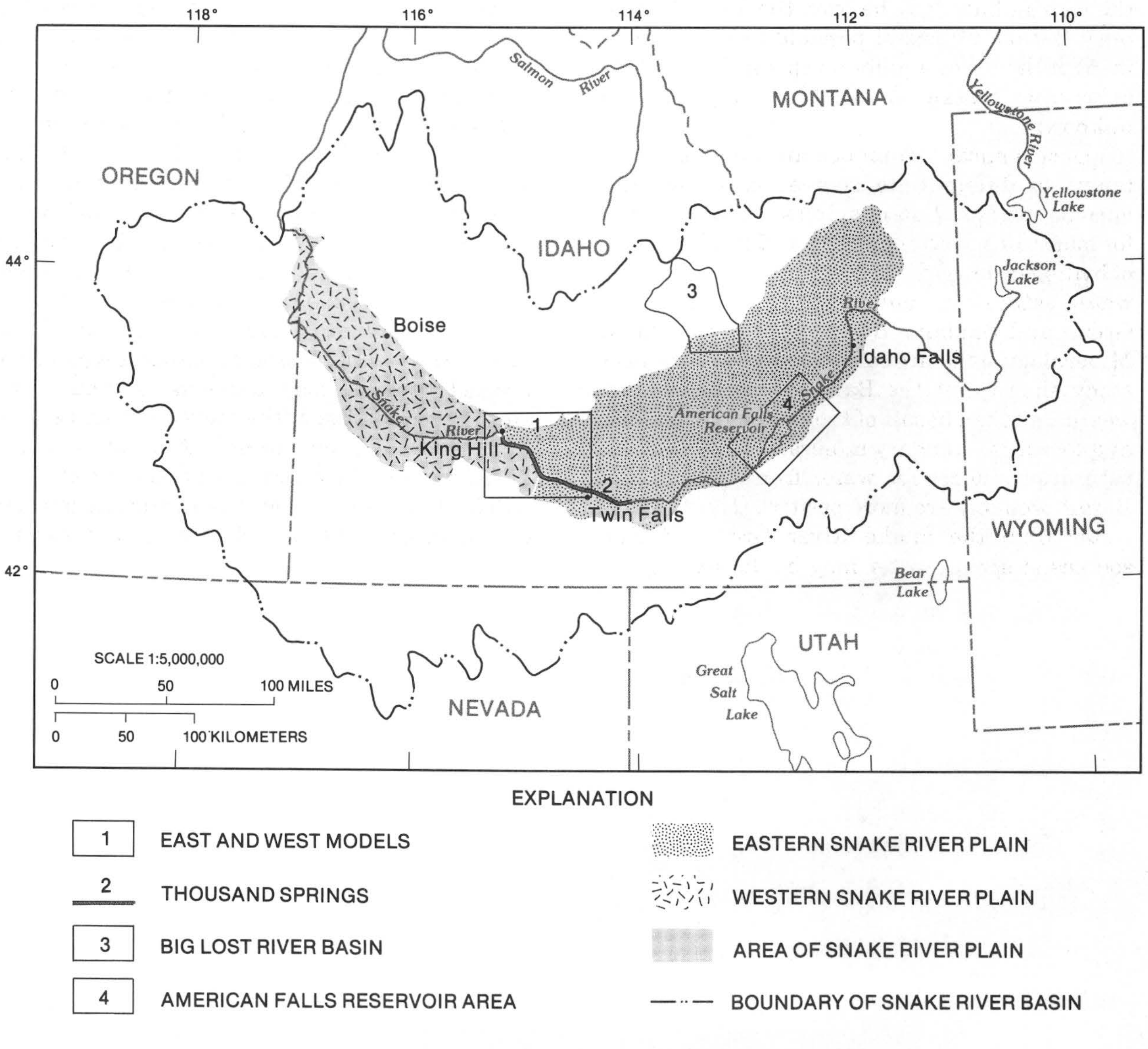

Figure 160.—Study area of the Snake River Plain regional aquifer system, phase II study.

prepared to describe the geology of the north Snake River canyon wall between Milner and King Hill (fig. 49). Geologic controls on spring locations were determined and quantity of discharge estimated. The actual points of emergence of some springs are not always apparent. In places, spring vents are masked by talus deposits; elsewhere, springs issue from the canyon floor as much as 200 feet below other springs. A test hole about 12 miles upgradient from the springs penetrated a basalt aquifer (Tertiary Banbury Basalt), which is separated from the overlying Snake River Plain aquifer by a 200-foot thick zone of sediments. The altitude of the top of the lower aquifer is about the same as the floor of the Snake River canyon. As such, it seems probable that some springs issuing from the canyon floor may be from the Tertiary Banbury Basalt. Effects of possible future development of the lower aquifer on spring flow and the overlying Snake River Plain aquifer are unknown.

Geochemical techniques are proposed to attempt to differentiate sources of water from various springs. Water samples will be analyzed for major ions, trace elements, and stable isotopes of hydrogen, oxygen, carbon, and sulfur. Mineral-water saturation indices for the Snake River Group and Banbury Basalt will be calculated. Mineralogic evidence collected during the initial study shows that the Banbury Basalt is more weathered than basalt of the Snake River Group and contains secondary minerals. Mineral-water saturation indices for water from the Banbury Basalt probably are more positive than those for water from the Snake River Group. If true, sources of spring water may be differentiated.

STREAM-AQUIFER MODEL FOR THE EASTERN PART OF THE SNAKE RIVER PLAIN

Ground- and surface-water systems in the Snake River Plain are closely interrelated. During the phase I study, historical changes in ground water and surface water resulting from 100 years of irrigation were defined. Percolation of large quantities of excess surface water diverted for irrigation raised water levels several tens of feet in much of the Snake River Plain and as much as 200 feet locally. Ground-water discharge subsequently increased, largely as spring flow; flow in the Snake River changed accordingly. The increase in river discharge was appropriated for hydropower generation, irrigation, and other uses. When surface water was fully appropriated, ground water became a feasible alternative source. The result was a reversal of the original trends; ground-water levels began to decline and spring flow decreased. Increased competition for water made holders of junior water rights concerned about the reliability of their supply and a conflict of interest developed between proponents of hydropower generation and irrigation. The relation between ground and surface water is now generally recognized, and concern has developed that increased use of ground water will adversely impact availability of surface water. Many considerations are necessary to properly manage conjunctive use of ground and surface water. One of the tools that can be used is a stream-aquifer model. A stream-aquifer model of the eastern part of the Snake River Plain will be developed to study the hydrologic impacts of the conjunctive use of ground and surface water.

APPENDIX: RASA PROJECT CHIEFS AND THEIR ADDRESSES

Regional Aquifer-System Analysis Program

Sun, Ren Jen, Program Coordinator
Office of Ground Water, Water Resources Division
U.S. Geological Survey, National Center, Mail Stop 411
12201 Sunrise Valley Drive
Reston, Virginia 22092

1. Completed Phase I Regional Aquifer-System Analysis Projects.

Central Valley regional aquifer-system study, California

Bertoldi, Gilbert L., Project Chief (Resigned)
Page, Ronald W., Hydrologist
U.S. Geological Survey
Federal Building, Room W-2235
2800 Cottage Way
Sacramento, California 95825

Floridan regional aquifer-system study

Johnston, Richard H., Project Chief (retired)
Bush, Peter W., Hydrologist
U.S. Geological Survey
649 Federal Bldg.,
300 East 8th Street
Austin, Texas 78701

High Plains regional aquifer-system study

Weeks, John B. Project Chief
U.S. Geological Survey
Bldg. 53, Denver Federal Center
Mail Stop 412
Lakewood, Colorado 80225

Northern Great Plains regional aquifer-system study

Dinwiddie, George A., Project Chief
U.S. Geological Survey, National Center
Mail Stop 410
12201 Sunrise Valley Drive
Reston, Virginia 22092

Northern Midwest regional aquifer-system study

Young, Harley L., Project Chief
U.S. Geological Survey
1815 University Avenue, 2nd Floor
Madison, Wisconsin 53705-4042

Snake River Plain regional aquifer system study

Lindholm, Gerald F., Project Chief
U.S. Geological Survey
230 Collins Road
Boise, Idaho 83702

Southwest alluvial basins regional aquifer-system study

Study in parts of Colorado, New Mexico, and Texas

Wilkins, David W., Project Chief
U.S. Geological Survey
Western Bank Building, Room 720
505 Marquette, N.W.
Albuquerque, New Mexico 87102

Study in southern and central Arizona and adjacent states

Anderson, Thomas W., Project Chief
U.S. Geological Survey
Federal Bldg., FB-44
301 W. Congress Road
Tucson, Arizona 85701

2. Active Phase I Regional Aquifer-System Analysis Projects

Central Midwest regional aquifer-system study

Jorgensen, Donald G., Project Chief
U.S. Geological Survey
1950 Constant Avenue, Campus West
University of Kansas
Lawrence, Kansas 66046

Columbia Plateau basalt regional aquifer-system study

Vaccaro, John J., Project Chief
U.S. Geological Survey
1201 Pacific Avenue, Suite 600
Tacoma, Washington 98402

Great Basin regional aquifer-system study

Harrill, James R., Project Chief
U.S. Geological Survey
Federal Bldg., Room 224
705 North Plaza Street
Carson City, Nevada 89701

Gulf Coastal Plain regional aquifer-system study

Grubb, Hayes F., Project Chief
U.S. Geological Survey
211 East 7th Street, 3rd Floor
Austin, Texas 78701

Northeast glacial regional aquifer-system study

Lyford, Forest P., Project Chief
U.S. Geological Survey
353 U.S. Post Office & Courthouse
Albany, New York 12201

Northern Atlantic Coastal Plain regional aquifer-system study

Meisler, Harold, Project Chief
U.S. Geological Survey
810 Bear Tavern Road
Suite 206
West Trenton, New Jersey 08628

Oahu Island regional aquifer-system study, Hawaii

Ewart, Charles J., Project Chief
U.S. Geological Survey
300 Ala Moana Blvd., Room 6110
Honolulu, Hawaii 96850

Southeastern Coastal Plain regional aquifer-system study

Miller, James A., Project Chief
U.S. Geological Survey
Richard B. Russell Federal Bldg., Suite 772
75 Spring Street, S.W.
Atlanta, Georgia 30303

Upper Colorado River Basin regional aquifer-system study

Cenozoic aquifer system

Glover, Kent C., Project Chief
U.S. Geological Survey
2120 Captial Avenue, Room 4004
Cheyenne, Wyoming 82003

Mesozoic aquifer system

Freethey, Geoffrey, Project Chief
U.S. Geological Survey
Administration Bldg., Room 1016
1745 West 1700 South
Salt Lake City, Utah 84104

Paleozoic aquifer system

Taylor, O. James, Project Chief
U.S. Geological Survey
Bldg. 53, Denver Federal Center
Mail Stop 415, Box 25046
Lakewood, Colorado 80225

Phase I Regional Aquifer-System Analysis Projects started in 1984.

Caribbean Islands regional aquifer-system study

Gómez-Gómez, Fernando, Project Chief
U.S. Geological Survey
GSA Center, Bldg. 652
Highway 28, Pueblo Viejo
San Juan, Puerto Rico 00936

Michigan Basin regional aquifer-system study

Mandle, Richard J., Project Chief
U.S. Geological Survey
6520 Mercantile Way, Suite 5
Lansing, Michigan 48910

Southern California alluvial basins regional aquifer-system study

Martin, Peter, Project Chief
U.S. Geological Survey
5201 Ruffin Road, Suite F. COC Annex
San Diego, California 92123

Phase II Regional Aquifer-System Analysis Projects (followup studies)

Central Valley regional aquifer system, California, phase II study

Gilliom, Robert J., Project Chief
U.S. Geological Survey
Federal Building, Room W-2235
2800 Cottage Way
Sacramento, California 95825

Floridan regional aquifer system, phase II study

Bush, Peter W., Project Chief
U.S. Geological Survey
649 Federal Bldg.
300 East 8th Street
Austin, Texas 78701

High Plains regional aquifer system, phase II study

Weeks, John B., Project Chief
U.S. Geological Survey
Bldg. 53 Denver Federal Center
Mail Stop 412
Lakewood, Colorado 80225

Snake River Plain regional aquifer system, phase II study

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