GEOHYDROLOGY AND SIMULATION OF STEADY-STATE FLOW CONDITIONS IN REGIONAL AQUIFER SYSTEMS IN CRETACEOUS AND OLDER ROCKS UNDERLYING KANSAS, NEBRASKA, AND PARTS OF ARKANSAS, COLORADO, MISSOURI, NEW MEXICO, OKLAHOMA, SOUTH DAKOTA, TEXAS, AND WYOMING

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Geohydrology and Simulation of Steady-State Flow Conditions in Regional Aquifer Systems in Cretaceous and Older Rocks Underlying Kansas, Nebraska, and Parts of Arkansas, Colorado, Missouri, New Mexico, Oklahoma, South Dakota, Texas, and Wyoming

By DONALD C. SIGNOR, JOHN O. HELGESEN, DONALD G. JORGENSEN, *and* ROBERT B. LEONARD

REGIONAL AQUIFER-SYSTEM ANALYSIS—
CENTRAL MIDWEST

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1414—C

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FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

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

The final interpretive results of the RASA Program are presented in a series of U.S. Geological Survey Professional Papers that describe the geology, hydrology, and geochemistry of each regional aquifer system. Each study within the RASA Program is assigned a single Professional Paper number beginning with Professional Paper 1400.

Gordon P. Eaton

Gordon P. Eaton
Director

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ILLUSTRATIONS

[Plates are in pocket]

TABLES

CONVERSION FACTORS

Multiply	By	To obtain
inch	25.40	millimeter
foot (ft)	0.3048	meter
square foot (ft^2)	0.09290	meter squared
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer
millidarcy (mD)	9.87×10^{-16}	square meter
foot per second (ft/s)	0.3048	meter per second
foot per mile (ft/mi)	0.1894	meter per kilometer
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
cubic foot per day (ft^3/d)	0.02832	cubic meter per day
cubic foot per year (ft^3/yr)	0.02832	cubic meter per year
cubic foot per square mile per year [$(ft^3/mi^2)/yr$]	0.0109	cubic meter per square kilometer per year
gallon per day per square foot [$(gal/d)/ft^2$]	40.7427	liter per day per square meter
inch per year (in/yr)	25.40	millimeter per year
acre-foot (acre-ft)	1,233	cubic meter

Chemical concentrations and water temperatures are given in metric units. Chemical concentration is given in milligrams per liter (mg/L). Water temperature is given in degrees Celsius (°C), which can be converted to degrees Fahrenheit (°F) by the following equation:

$$°F = 1.8\ (°C) + 32.$$

SEA LEVEL

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

REGIONAL AQUIFER-SYSTEM ANALYSIS—CENTRAL MIDWEST

GEOHYDROLOGY AND SIMULATION OF STEADY-STATE FLOW CONDITIONS IN REGIONAL AQUIFER SYSTEMS IN CRETACEOUS AND OLDER ROCKS UNDERLYING KANSAS, NEBRASKA, AND PARTS OF ARKANSAS, COLORADO, MISSOURI, NEW MEXICO, OKLAHOMA, SOUTH DAKOTA, TEXAS, AND WYOMING

By DONALD C. SIGNOR, JOHN O. HELGESEN, DONALD G. JORGENSEN, *AND* ROBERT B. LEONARD

ABSTRACT

Regional aquifer systems of Cambrian through Lower Cretaceous sedimentary rocks were the focus of an investigation termed the Central Midwest regional aquifer-system analysis. The study area consists of about 370,000 square miles and extends from the foothills of the Rocky Mountains in Colorado to the Mississippi River in eastern Missouri and from South Dakota to the Ouachita, Arbuckle, and Wichita Mountains of Arkansas and Oklahoma.

Geologically, the study area lies within the stable interior of the North American continent and is dominated structurally by broad basins and arches. Substantial crustal deformation along the Ouachita, Arbuckle, and Wichita Mountains in the south and the Rocky Mountains in the west was the major factor in defining the southern and western boundaries of the study area.

There are two subregions in the study area, the Plains subregion and the Ozark subregion, that have three regional flow systems. Two regional flow systems separated by a thick confining system are present in the Plains subregion; one regional flow system exists in the Ozark subregion.

Within the Plains subregion, one flow system is in Cambrian through Mississippian rocks; the second flow system is in Cretaceous sandstone. The regional geohydrologic units in the Plains subregion are the basement confining unit (Precambrian rocks), the Western Interior Plains aquifer system (Cambrian and Mississippian rocks), the Western Interior Plains confining system (Upper Mississippian and Jurassic rocks), the Great Plains aquifer system (Lower Cretaceous rocks), the Great Plains confining system (Upper Cretaceous and younger rocks), and the High Plains aquifer (Tertiary and Quaternary rocks). The Great Plains confining system and the High Plains aquifer were studied only as they relate hydrologically to the Great Plains aquifer system.

In the Ozark subregion, a freshwater flow system in Lower Paleozoic rocks is laterally adjacent to the saline-water flow system of the Western Interior Plains aquifer system. The regional geohydrologic units in the Ozark subregion are the basement confining unit, the Ozark Plateaus aquifer system, and the Western Interior Plains confining system. The Ozark Plateaus aquifer system consists of the St. Francois aquifer, the St. Francois confining unit, the Ozark aquifer, the Ozark confining unit, and the Springfield Plateau aquifer.

The distribution of hydraulic head in the Western Interior Plains aquifer system indicates flow generally west to east and southeast. Hydraulic heads in the Ozark Plateaus aquifer system indicate nearly radial outflow toward the Plains subregion and subregion boundaries of the Missouri River, Mississippi River, and crustal deformation areas toward the south. Upward leakage occurs near the saltwater-freshwater transition zone where the aquifers in the Plains and Ozark subregions are laterally adjacent.

A numerical ground-water flow model was used to test the conceptualization of flow in the regional aquifers and aquifer systems in the study area. The numerical model used is a finite-difference model termed "The U.S. Geological Survey Modular Ground-Water Flow Model." Five principal geohydrologic units in the regional study area were represented in the model by five model layers. The Western Interior Plains aquifer system and the Ozark Plateaus aquifer system were represented by two model layers; the Western Interior Plains confining system and the Great Plains aquifer system by one model layer each; and a combined Great Plains confining system and High Plains aquifer by one model layer. Calibration of the model was based on closeness of fit of computed hydraulic head to field hydraulic head, analyses of regional outflow and inflow at the boundaries, and comparison to subregional-model hydraulic head and flow.

Comparisons of model-computed hydraulic head to field hydraulic head were made for model layers 2, 4, and 5. Model layer 2 simulated the Great Plains aquifer system. Model layer 4 simulated the upper unit of the Western Interior Plains aquifer system in the Plains subregion and the Springfield Plateau aquifer in the Ozark subregion. Model layer 5 simulated the lower units of the Western Interior Plains aquifer system in the Plains subregion and the St. Francois confining unit and St. Francois aquifer in the Ozark subregion. The average deviation of computed hydraulic head from field hydraulic head was less than 1 foot for the three layers. Effects of variable density due to brines increased average deviation of computed hydraulic head from field hydraulic head to 6 feet lower

for model layer 2, 30 feet higher for layer 4, and 41 feet higher for layer 5.

A sensitivity analysis to evaluate the effects of changes of hydraulic conductivity, leakance between layers, and recharge on computed hydraulic head showed that computed hydraulic head was very sensitive to changes in hydraulic conductivity, slightly sensitive to leakance changes between model layers 4 and 5, but more sensitive to leakance changes between layers 1 and 2, and only moderately sensitive to recharge for the overall model. Changes in recharge data affected computed hydraulic-head values in the Ozark subregion to a large degree. Simultaneously varying recharge and permeability values in the model showed that more than one solution of the model could be computed that would meet the criterion of computed hydraulic head matching calibration hydraulic head. This result was also true of simultaneously varying recharge and leakance between layers. Thus, the model solution was not unique.

Because of the large model-node spacing of 28 miles for the regional-model cells, an error analysis comparing the differences between computed heads and field heads for two node spacings was made. The area used for comparing hydraulic-head data was the Great Plains aquifer system. The Great Plains aquifer system was studied using a subregional model with a uniform node spacing that was one-half of the node spacing and the same orientation as the regional model. The hydraulic characteristics of the regional and subregional models were discretized from the same maps and data bases. The same computer model was used in both regional and subregional applications. From a statistical analysis, the model-node spacing relationship between regional and subregional models did not appear to make a significant difference in computed hydraulic head. The analysis also showed that there was no significant difference between the field hydraulic-head values of the two models. In conclusion, the modeling analysis indicated that:

1. The steady-state flow simulation of the study area reasonably represented the regional flow system.
2. An analysis using only freshwater hydraulic head where variable-density brines are present may lead to erroneous conclusions.

The potential for ground-water development in the study area can be summarized as follows:

1. Water in the Great Plains aquifer system is of questionable or unknown quality, and very little ground-water development has occurred to date. However, the quantity of water in storage is very large, and the aquifer system may be a supplemental source of water to augment the declining supply in the overlying High Plains aquifer.
2. Very little of the Western Interior Plains aquifer system contains water with less than 5,000 milligrams per liter dissolved solids, and therefore, the potential for ground-water development is slight. However, very poor water quality and very sluggish flow indicate that the aquifer system contains potential sites for waste storage or disposal.
3. Large quantities of water can be obtained from the Ozark Plateaus aquifer system. It contains freshwater, and there is a very large amount of water in storage. Simulated recharge to the aquifer system is about 7,000 cubic feet per second, or about 13 percent of the total recharge reaching the water table (55,000 cubic feet per second). The rejected recharge is discharged locally to streams. The large quantity of rejected recharge represents a potential ground-water resource; however, withdrawals of ground water will affect the base flow of streams draining the Ozark subregion.

INTRODUCTION

The study area of the Central Midwest regional aquifer-system analysis (CMRASA) includes about 370,000 mi^2 (fig. 1). It extends from the foothills of the Rocky Mountains in Colorado and Wyoming to the Mississippi River in eastern Missouri, and from South Dakota to the Ouachita, Arbuckle, and Wichita Mountains of Arkansas and Oklahoma.

Three important regional aquifer systems containing both freshwater and saline water were studied. In much of the area, little was known previously about the regional flow and hydrochemistry within the aquifer systems or about hydrologic relations among them.

Within the central United States, four other regional aquifer-system analyses (RASA) overlie or share geographic or hydrologic boundaries with the CMRASA study. These RASA studies, conducted either prior to or concurrent with the CMRASA, are the High Plains (Weeks, 1978), the Northern Great Plains (Dinwiddie, 1979), the Northern Midwest (Steinhilber and Young, 1979), and the Gulf Coast (Grubb, 1984) (fig. 2).

The background, scope, objectives, and approach of the CMRASA are described in the project-planning report by Jorgensen and Signor (1981). The findings of the CMRASA are reported in five chapters: chapter A is the summary chapter, which will collate the important findings reported in other chapters; chapter B describes the geohydrologic framework; chapter C (this report) describes the geohydrology and modeling analysis of the regional aquifer systems; chapter D describes the geohydrologic and model analyses of the Ozark Plateaus aquifer system; and chapter E describes the geohydrologic and model analyses of the Great Plains aquifer system.

PURPOSE AND SCOPE

The purpose of the work discussed in this report is to describe the aquifers, aquifer systems, confining units, and confining systems, and to test a conceptualization of regional flow within the study area by a computer simulation. The scope of the study is limited to rocks ranging in age from Cambrian through Late Cretaceous. The study does not include the flow regime in the Tertiary and Quaternary rocks constituting the High Plains aquifer; however, the computer flow model simulates the effects of Late Cretaceous and younger rock as a boundary condition.

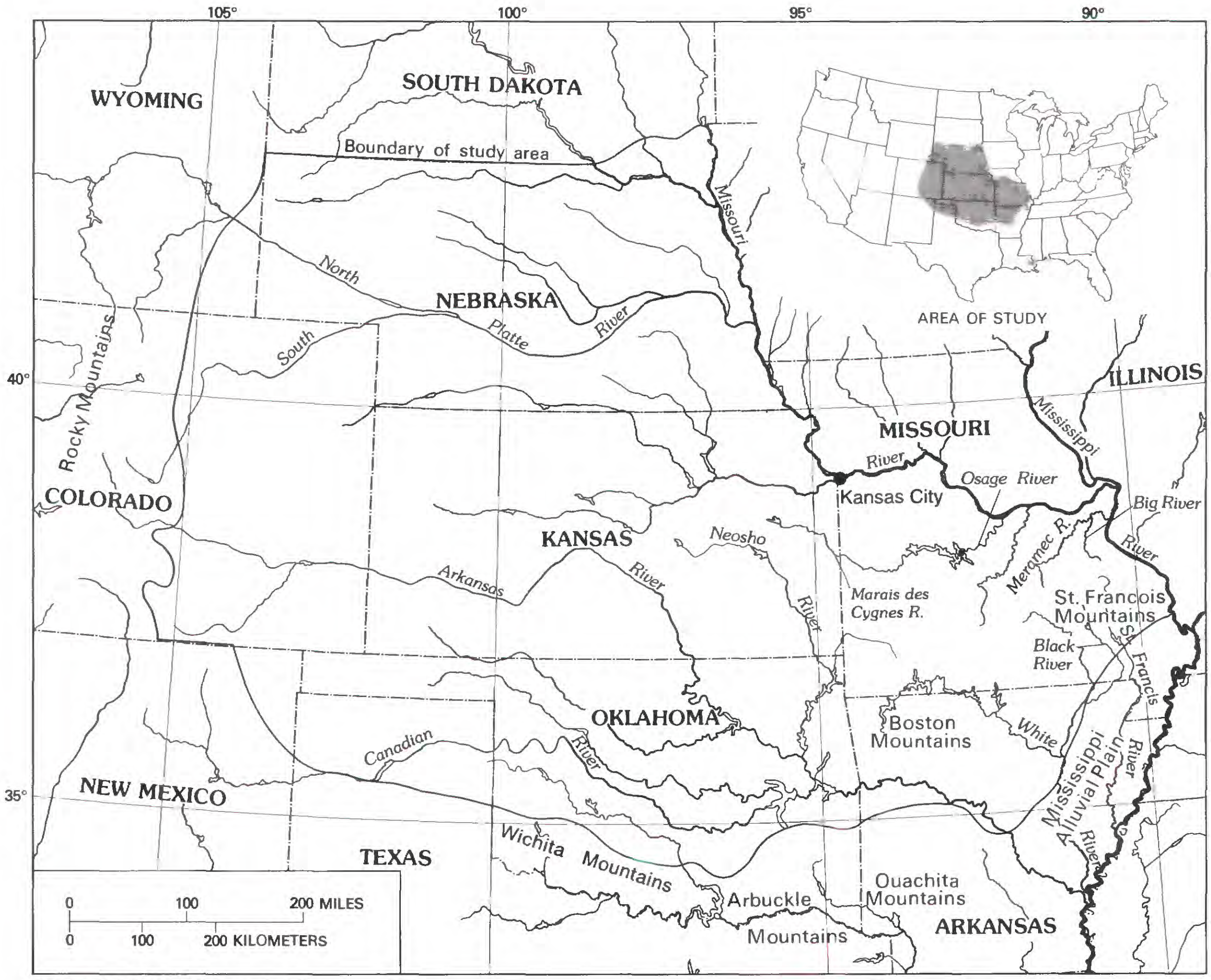

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FIGURE 1.—Location and extent of study area.

PHYSICAL SETTING

The study area includes parts of two major physiographic divisions, the Interior Plains and the Interior Highlands (fig. 3). The Interior Plains includes the Great Plains and the Central Lowlands, and the Interior Highlands includes the Ozark Plateaus and the Ouachita Province. The Ouachita Province includes the Arkansas Valley, which is between the Boston Mountains of the Ozark Plateaus and the Ouachita Mountains south of the study area. The study area is bounded laterally by either major geologic features or by major rivers. Geologic features include the Rocky Mountain Uplifts on the west, the Siouxana Arch on the northeast, a series of uplifts on the south, and the Fall Line (the boundary with the Mississippi Alluvial Plain) on the southeast. Major rivers are the Mississippi and the Missouri Rivers.

Land-surface altitude ranges from less than 500 ft on the Mississippi Alluvial Plain and along the Arkansas Valley to a maximum of about 10,000 ft in the extreme western part of the area in the foothills of the Rocky Mountains (fig. 4). Most of the Interior Plains is relatively flat, whereas the topography of the Ozark Plateaus is hilly and rugged. The Interior Plains area is mainly a mixture of grassland and row crops; much of the Ozark Plateaus is covered by hardwood forest.

Climatic characteristics, including runoff, precipitation, and potential evapotranspiration, are shown in figure 5. Mean annual precipitation for 1931–60, which included two major droughts (1933–37 and 1952–57), ranged from 12 in. in eastern Colorado to more than 48 in. in the Ozark Plateaus (Eagleman, 1976, fig. 3). Mean annual potential

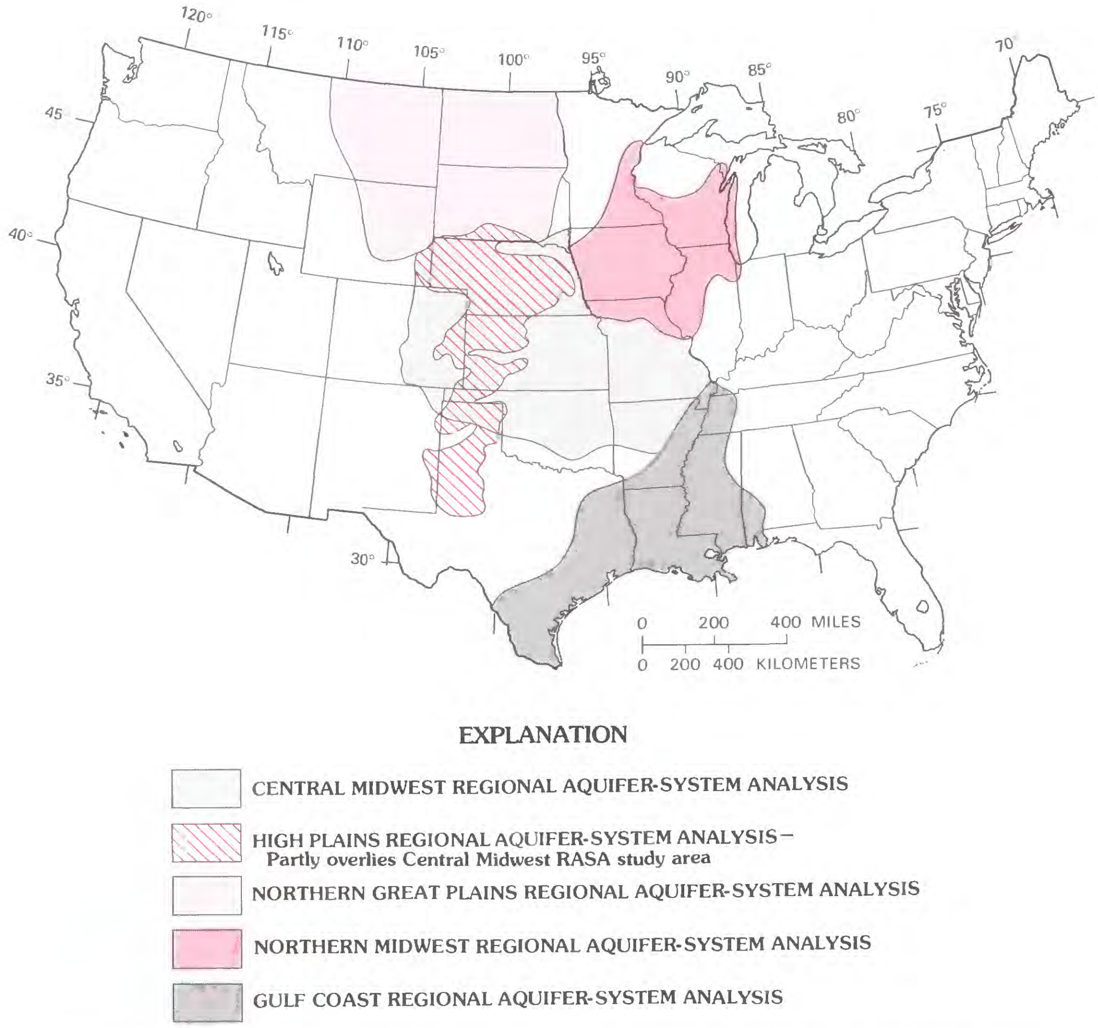


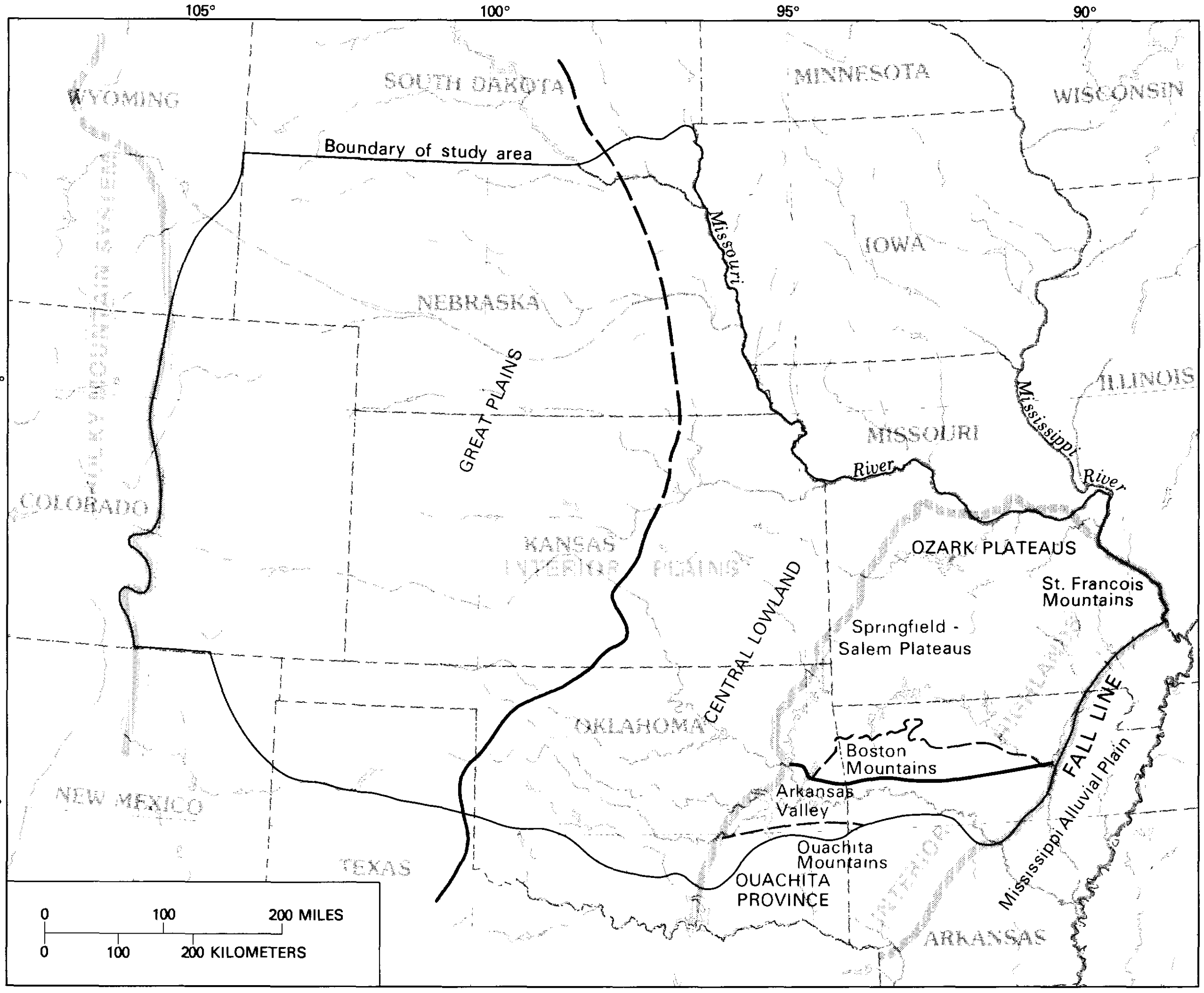
FIGURE 2.—Regional aquifer-system studies contiguous to the Central Midwest regional aquifer-system analysis.

evapotranspiration generally exceeds precipitation (fig. 5). Another CMRASA study (Dugan and Peckenpaugh, 1985) to determine the amount of water passing the root zone and reaching the water table indicates that less than 1 in. is available in Colorado, whereas more than 15 in. of water are available annually for recharge in the Ozark Plateaus (fig. 6). The eastern part of the study area has abundant water resources, especially surface water. Mean annual surface-water runoff during 1931–60 ranged from 1 in. or less in western Kansas and eastern Colorado to more than 15 in. in the Ozark Plateaus (fig. 5). Mean annual runoff exceeded 27 in. during 1951–80 in the Ozark Plateaus (Hedman and others, 1987). Most of the western part of the study area is deficient in surface water but has substantial ground-water resources, mainly in the High Plains aquifer, which overlies some of the aquifers described in this report (fig. 2).

ACKNOWLEDGMENTS

The plan of this investigation was developed with the assistance of personnel from the Arkansas Geological Commission, the Colorado Geological Survey, the Kansas Geological Survey, the Missouri Division of Geology and Land Survey, the Nebraska Conservation and Survey Division, the Oklahoma Geological Survey, and the South Dakota Geological Survey.

The study was aided greatly by a Liaison Committee made up of representatives from State agencies and the U.S. Geological Survey. The Liaison Committee not only disseminated information concerning



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EXPLANATION

BOUNDARY OF PHYSIOGRAPHIC DIVISON—Dashed where approximate

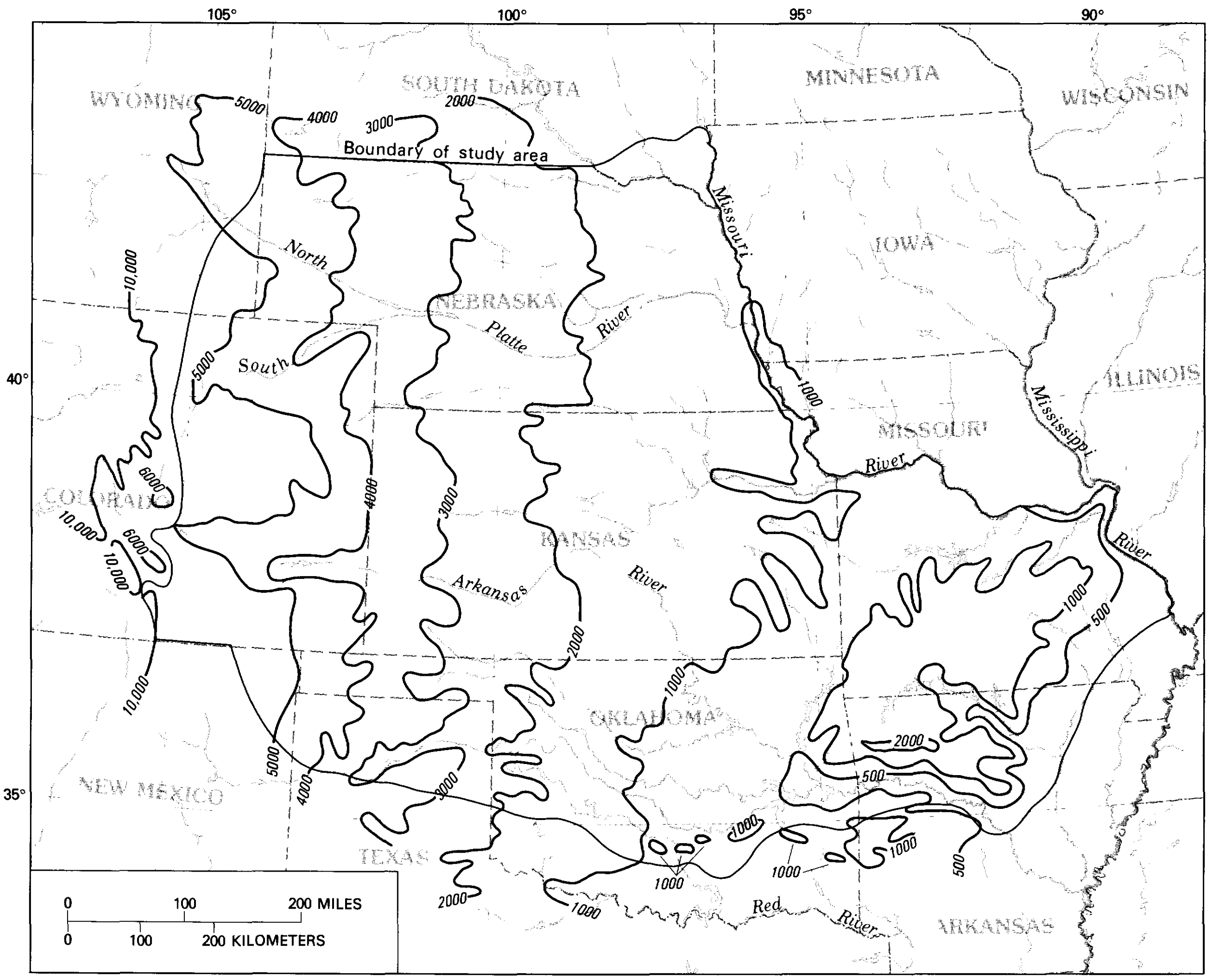
BOUNDARY OF PHYSIOGRAPHIC PROVINCE—Dashed where approximate

BOUNDARY OF PHYSIOGRAPHIC SECTION

FIGURE 3.—Physiographic subdivisions within CMRASA study area (modified from Fenneman, 1946).

the study to State agencies but also helped identify sources of data, defined areas of concern, and made valuable suggestions that aided the study. Committee members included: Orville Wise, Arkansas Geological Commission; Robert Longenbaugh, Office of Colorado State Engineer; William Hambleton, Kansas Geological Survey; Donald Miller, Missouri Division of Geology and Land Survey; Verlon Vrana, Nebraska Natural Resources Commission; and Charles Mankin represented by Robert Arndt, Oklahoma Geological Survey.

The investigation also was served by a stratigraphy advisory group that included: Orville Wise, Arkansas Geological Commission; Richard Pearl, Colorado Geological Survey; P. Allen Macfarlane, Kansas Geological Survey; Thomas Thompson, Missouri Division of Geology and Land Survey; Marvin Carlson, Nebraska Conservation and Survey Division; and Charles Mankin, Oklahoma Geological Survey. This group, along with Claire Davidson of the Geologic Names Committee of the U.S. Geological Survey, contributed significantly to the resolution of problems related to stratigraphic correlation and nomenclature.



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EXPLANATION

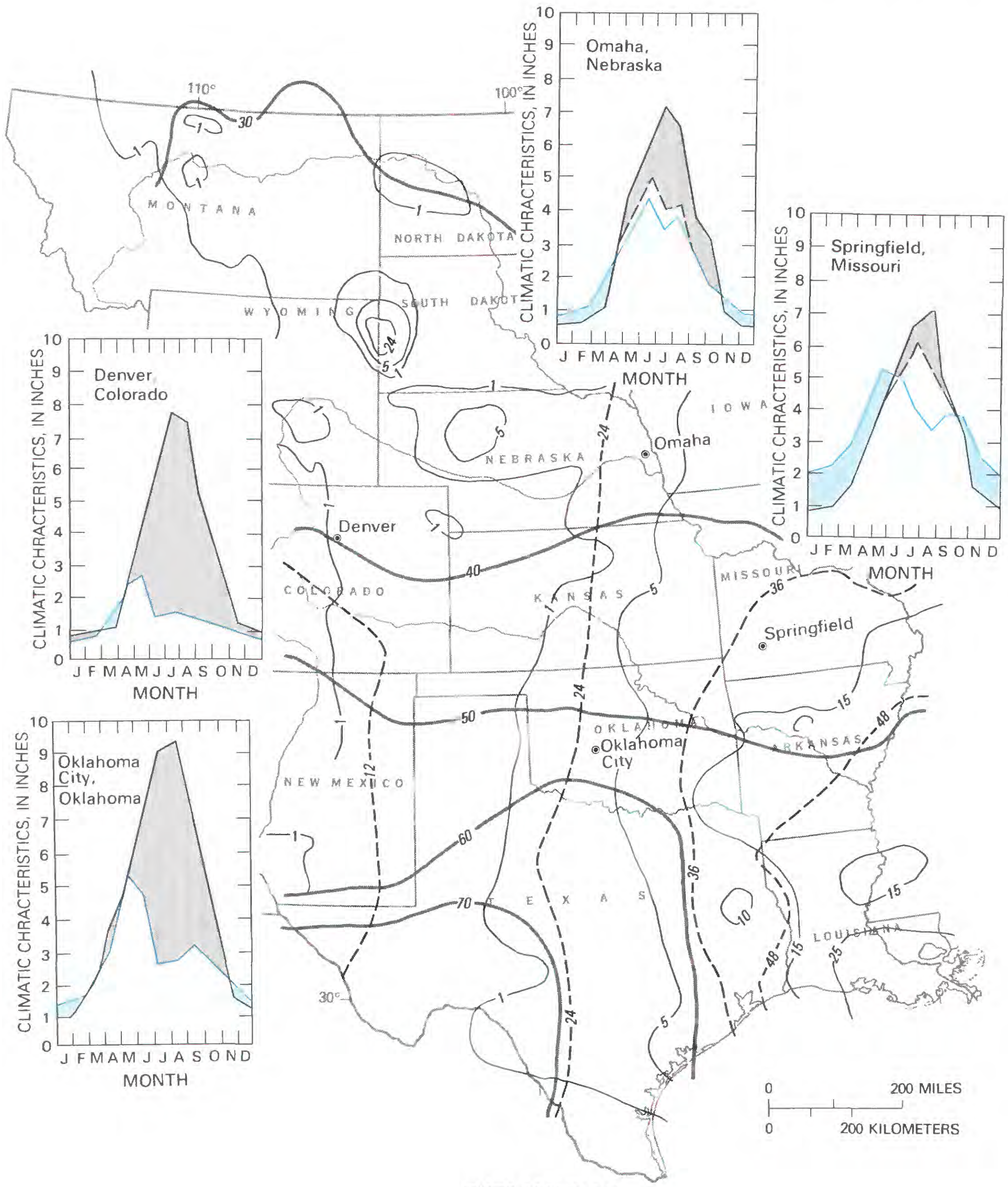
—— 4000 —— TOPOGRAPHIC CONTOUR—Shows altitude of land surface. Contour interval, in feet, is variable. Datum is sea level

FIGURE 4.—Generalized land-surface altitude.

GEOLOGIC SETTING

Gently dipping sedimentary rocks that form broad uplifts and basins occur over most of the study area. However, stratigraphic nomenclature differs markedly among States and structural areas (fig. 7). As noted by Jorgensen and others (1993) hundreds of names have been assigned to rock units within the thick stratigraphic interval. The principal stratigraphic units composing the geohydrologic units discussed in this report are listed in tables 1 and 2.

The study area lies within the stable interior of the North American continent. From Late Proterozoic to present, or since Cambrian time to present, most of the study area has undergone relatively gentle deformation, which involves upwarp and downwarp of the Earth's crust over large areas. Structurally, the study area has been dominated by broad basins and uplifts. Accordingly, most folding of sedimentary rocks has been gentle, and few major fault zones of regional importance occur. However, along the southern and western margins of the study area, substantial crustal deformation of mountain-building force resulted in intense folding and faulting. The lateral change from simple to complex geologic structure was the major factor in defining the southern and western boundaries of the study area. At the boundaries, the change in structure is relatively abrupt in most locations but transitional in others.



EXPLANATION

CLIMATIC CHARACTERISTICS GRAPHS

Moisture deficit

Moisture surplus

Potential evapotranspiration

Precipitation

Actual evapotranspiration

10 LINE OF EQUAL ANNUAL RUNOFF—Interval, in inches, varies

24 LINE OF EQUAL ANNUAL PRECIPITATION—Interval 12 inches

40 LINE OF EQUAL ANNUAL POTENTIAL EVAPOTRANSPIRATION—Interval 10 inches

FIGURE 5.—Climatic characteristics of central United States, 1931–60 (modified from Eagleman, 1976; Farnsworth and Peck, 1982, map 3; Moody and others, 1986; Gerbert and others, 1987).

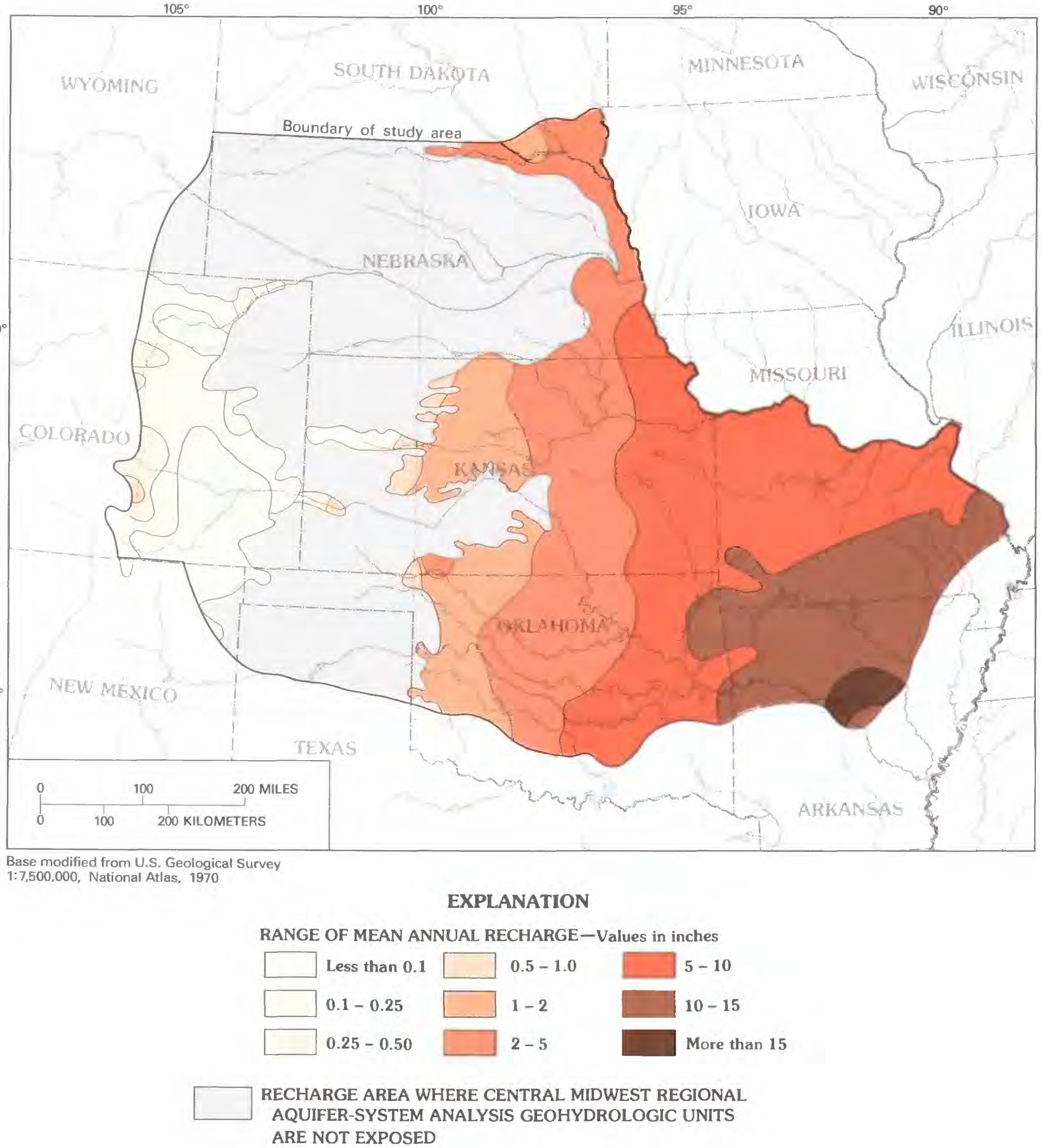
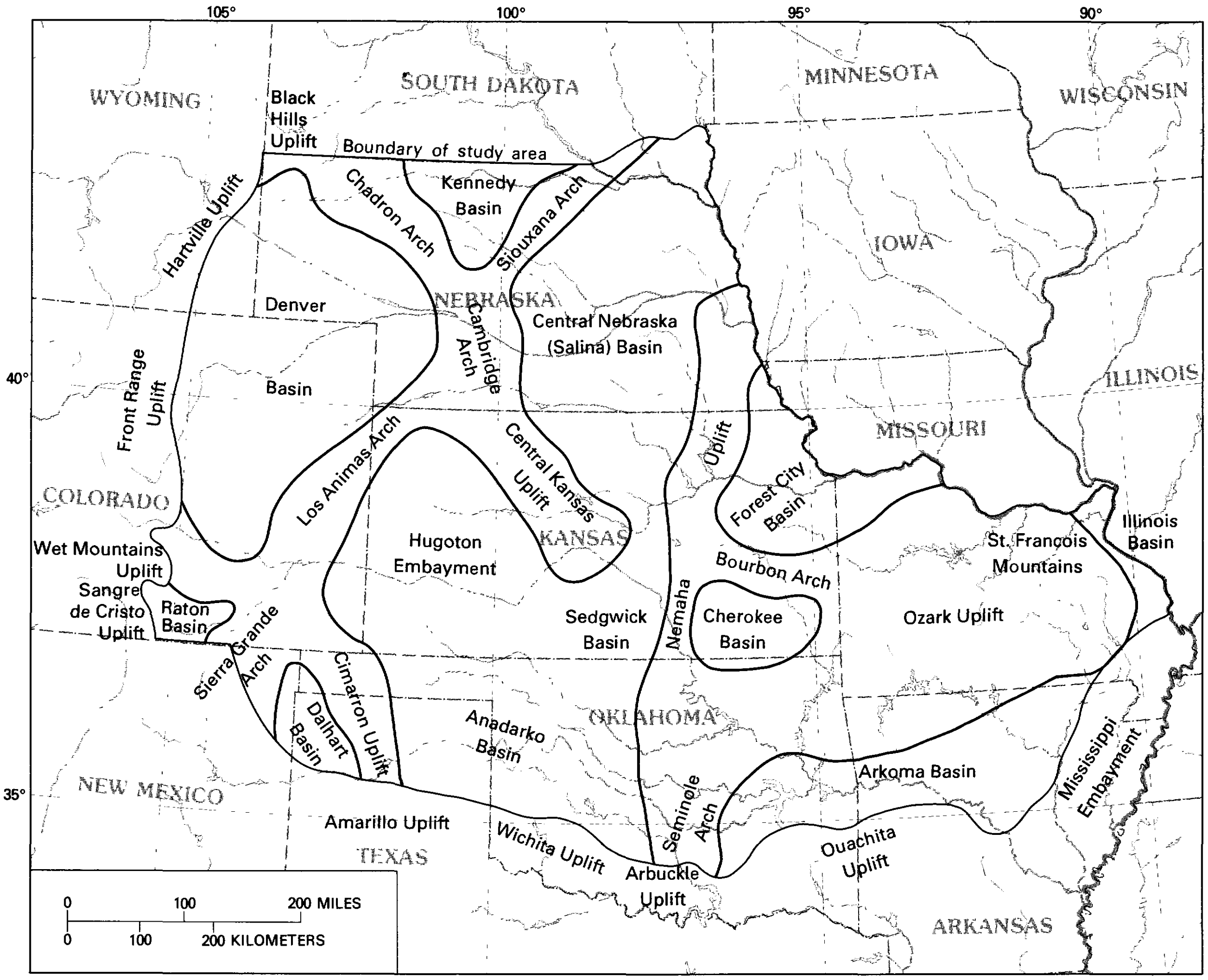


FIGURE 6.—Estimated mean annual recharge to aquifer-outcrop areas in CMRASA geohydrologic units, 1951–80 (from Dugan and Peckenpaugh, 1985).

The present areal distribution of major time-stratigraphic units that crop out in the Central Midwest study area is shown in figure 8. Cross sections (fig. 9) illustrate the general regional continuity of the units; the greatest structural deformation is along the southern and western margins of the study area. Faulting (fig. 10) was substantial in these areas and caused significant offset of rock units. Most faults in the interior of the study area have much less offset than the faults along the southern and western margins.

Precambrian rocks, consisting mainly of igneous and metamorphic rocks, form a basement complex that underlies the Paleozoic and Mesozoic



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EXPLANATION

——— EXTENT OF STRUCTURAL FEATURE

FIGURE 7.—Major structural features of central United States (modified from Condra and Reed, 1959; Carlson, 1963; Merriam, 1963; Oetking and others, 1966, 1967; Renfro, 1973).

sedimentary rocks of interest in this study (tables 1, 2). Deep burial of these rocks at most locations has precluded detailed knowledge of their nature. Merriam (1963) describes Precambrian sedimentary rocks that are only slightly metamorphosed in northeastern Kansas. Seismic reflection studies (Brown and others, 1983) suggest layering and complex structures in the same area. Along the southern boundary in Oklahoma, the basement includes extrusive rhyolite rocks or intrusive granite of Early to Middle Cambrian age (Ham and others, 1964).

Major Precambrian faults in the study area are oriented to the present northeast (fig. 10). The alignment nearly parallels a fault belt called the Colorado lineament (not shown in figs. 7 or 10) described by Warner (1980, fig. 3) and the Sierra Grande, Las Animas, and Siouxana Arches (fig. 7). Other major faults in the area trend northwest in the same general area as the Central Kansas Uplift and the Cambridge and Chadron Arches (figs. 7, 10). Information concerning faults and fractures in Precambrian rocks is of importance because the faults mark weak zones that were sometimes reactivated during later geologic time and the fractures are major paths of ground-water flow in well-indurated rocks.

TABLE 1.—*Generalized correlation of geohydrologic units to stratigraphic units in most of the Plains subregion*
[From Jorgensen and others, 1993]

Geohydrologic unit		Principal rock-stratigraphic unit(s)	Time-stratigraphic unit
High Plains aquifer		Ogallala Formation and unconsolidated deposits	Quaternary and Tertiary
Great Plains confining system		Pierre Shale, Niobrara Formation, Carlile Shale, Greenhorn Limestone, Graneros Shale (includes Lower Cretaceous)	Upper Cretaceous
Great Plains aquifer system	Maha aquifer	Dakota Sandstone, "D" sandstone, "J" sandstone, and equivalent of Newcastle Sandstone	Lower Cretaceous
	Apishapa confining unit	Kiowa Shale and equivalent of Skull Creek Shale	
	Apishapa aquifer	Cheyenne Sandstone and equivalent of Fall River Sandstone and Lakota Sandstone	
Western Interior Plains confining system		Morrison Formation, Sundance Formation, Entrada Sandstone, Dockum Formation, Elk City Sandstone, Doxey Shale, Big Basin Sandstone, Cloud Chief Formation, Day Creek Dolomite, Whitehorse Sandstone, Nippewalla Group, Sumner Group, Chase Group, Council Grove Group, Admire Group, Wabaunsee Group, Shawnee Group, Douglas Group, Lansing Group, Kansas City Group, Pleasanton Group, Marmaton Group, Cherokee Group, Atokan rocks, Morrowan rocks, and Springer Group	Jurassic through Upper Mississippian (Chesterian)
Western Interior Plains aquifer system	Upper aquifer unit	Meramecian, Osagean, and Kinderhookian rocks	Upper Mississippian through Upper Cambrian
	Confining unit	Chattanooga Shale and Woodford Shale	
	Lower aquifer units	Hunton Group, Sylvan Shale, equivalent of Galena Dolomite, Viola Limestone, Simpson Group, Arbuckle Group, and Reagan Group	
Basement confining unit		Mostly igneous and metamorphic rocks	Cambrian and Precambrian

Dake (1930, p. 194) stated that in the Ozark subregion, rhyolite lava and ash, along with granite, granite porphyry, and basic dikes, formed a large land mass during the Precambrian. He further stated that the mass, of nearly 2,000-ft elevation, was deeply eroded and faulted by Cambrian time. Precambrian faulting in the Ozark subregion is reported also by Bridge (1930, p. 136). The core of the St. Francois Mountains is an epizonal granite batholith that was emplaced 1.5 billion years ago and was concomitant with the lowering and raising of sea level.

DEFINITION OF GEOHYDROLOGIC CHARACTERISTICS—METHODOLOGY

A description of the geohydrologic framework includes the extent, thickness, and altitude of the geohydrologic units. It is useful to define the geohydrologic units within the framework of lithostratigraphic units because the rock units are, in general, well established and well known, especially as compared to geohydrologic units. Definition of geohydrologic units (aquifers, aquifer systems, confining units, and confining systems) within the study area is based largely on hydrologic relations and hydraulic properties of the rocks. Aquifers and aquifer systems that contain relatively freshwater have a characteristic potentiometric surface that is continuous and relatively smooth. Therefore, the potentiometric surface of water-bearing units is useful in defining a geohydrologic unit. In addition, maps showing hydrochemical characteristics are useful in delineating a geohydrologic unit because many aquifers have a characteristic water chemistry.

Interpretive maps prepared for this study are based on many sources, and density of data varies considerably depending on the geographic area and

TABLE 2.—*Generalized correlation of geohydrologic units to stratigraphic units in most of the Ozark subregion*
[From Jorgensen and others, 1993]

Geohydrologic unit		Principal rock-stratigraphic unit(s)	Time-stratigraphic unit
Western Interior Plains confining system		Marmaton Group, Cherokee Group, Atokan rocks, Bloyd Shale, Hale Formation, Morrowan rocks, Pitkin Limestone, Fayetteville Shale, and Batesville Sandstone	Middle Pennsylvanian through Upper Mississippian (Chesterian)
Ozark Plateaus aquifer system	Springfield Plateau aquifer	Moorefield Formation, St. Louis Limestone, Salem Limestone, Warsaw Limestone, Boone Formation, St. Joe Limestone Member of Boone Formation, Keokuk Limestone, Burlington Limestone, and Fern Glen Limestone	Mississippian
	Ozark confining unit	Chouteau Group[1] and Chattanooga Shale, Northview Shale, and Hannibal Shale	Lower Mississippian and Upper Devonian
	Ozark aquifer	Clifty Limestone, Penters Chert, Lafferty Limestone, St. Clair Limestone, Brassfield Limestone, Cason Shale, Fernvale Limestone, Kimmswick Limestone, Plattin Limestone, Joachim Dolomite, St. Peter Sandstone, Everton Formation, Smithville Formation, Powell Dolomite, Cotter Dolomite, Jefferson City Dolomite, Roubidoux Formation, Gasconade Dolomite, Gunter Sandstone Member of Gasconade Dolomite, Eminence Dolomite, and Potosi Formation	Middle Devonian through uppermost Cambrian
	St. Francois confining unit	Elvins Group, Derby and Doe Run Dolomites, Davis Formation	Upper Cambrian
	St. Francois aquifer	Bonneterre Dolomite, Lamotte Sandstone, and Reagan Sandstone	
Basement confining unit		Mostly igneous and metamorphic rocks	Precambrian

[1]Designated Chouteau Limestone by U.S. Geological Survey.

the type of information being mapped. Many maps in this report show few or no data points in some areas where previously prepared interpretations were relied upon.

GEOPHYSICAL LOGS

Compilation of maps defining the framework of geohydrologic units was based largely on electric and lithologic logs from test holes for petroleum. However, in areas where the rocks contain freshwater, lithologic and geophysical logs from wells or test holes drilled for water were used. Logs that were representative of an area were selected to be included in the CMRASA data base (Helgesen and Hansen, 1989). Ideally, the log record included rocks of Cretaceous through Precambrian age. That is, logs were selected to be representative of both the areal and vertical scope of the study. In areas for which numerous logs are available, such as near oil and gas reservoirs, only a few of the numerous logs available were selected. In other areas, where data are scarce, most available logs were selected.

DISSOLVED SOLIDS

Dissolved-solids concentration is a measure of the water chemistry of a geohydrologic unit. Many aquifers have a characteristic hydrochemistry that is the result of many relations, including the chemistry

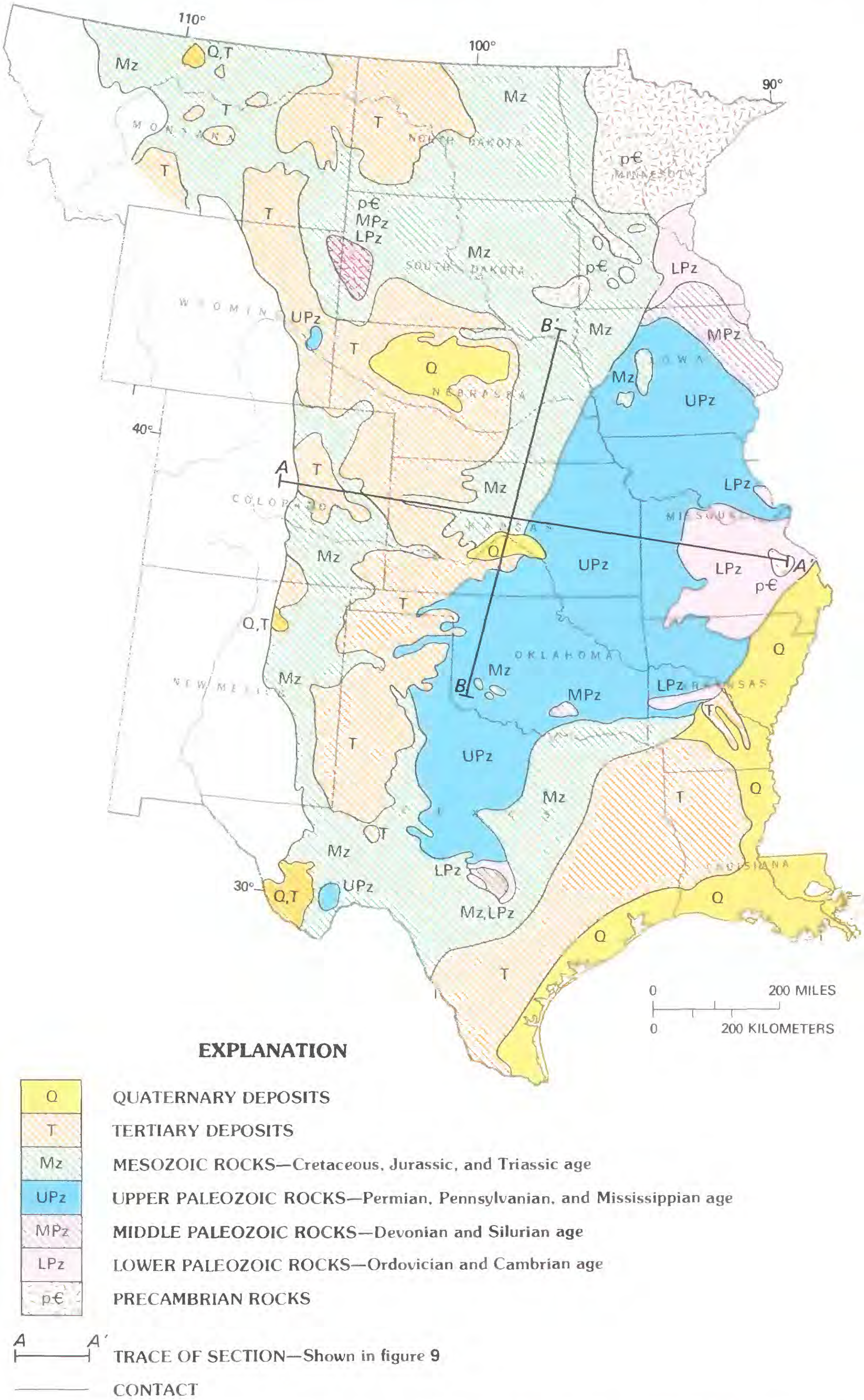


FIGURE 8.—Regional geology of central United States (modified from Kinney, 1966).

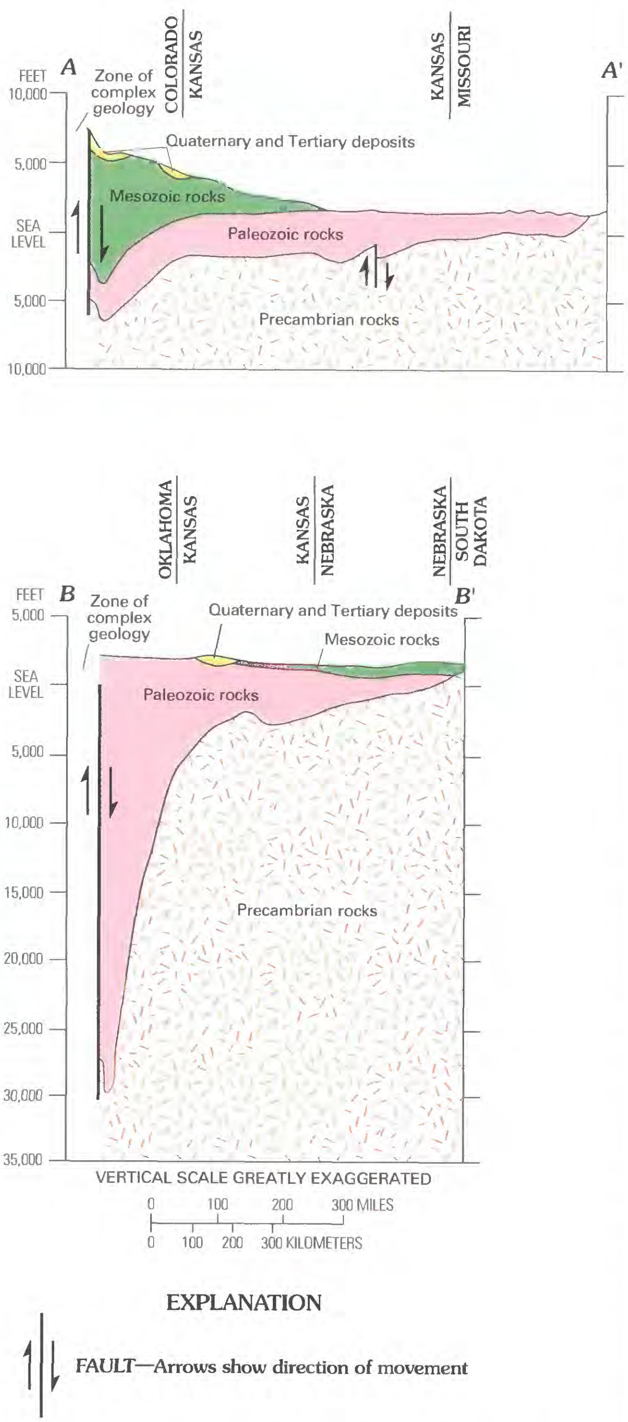


FIGURE 9.—Generalized geologic sections across study area. Traces of sections shown in figure 8.

of the water recharging the aquifers. Maps showing preliminary interpretations of the dissolved-solids concentrations were prepared from data collated into files for the CMRASA (Helgesen and Hansen, 1989).

Most of the chemical analyses of saline water were obtained from oil- and gas-well tests. Large differences between the reported concentrations of dissolved solids typically characterized samples from adjacent sites. The cause of these differences is difficult to determine. It is not known whether the differences are related to areal, vertical, and temporal variability, or to errors resulting from the sampling procedure or methods of chemical analysis.

The exact source of many of the water samples is unknown. Some samples were obtained from drill-stem tests; others are "production" water. Some may be mixtures of formation water with drilling fluids, or with water from zones above or below the zone of interest. Chemical analyses of freshwater from adjacent wells usually did not show large variability.

Hydrochemical data from many sources were entered in the CMRASA "Hydrochemical File." Sources of data for the computer file included the U.S. Geological Survey's National Water Data Storage and Retrieval System (WATSTORE) (Baker and Foulk, 1975), National Uranium Resource Evaluation (NURE) project files (Arendt and others, 1979, p. 11), Petroleum Data System (PDS) brine file maintained at the University of Oklahoma, as well as information from existing publications and other sources. Selected data from this large file were placed in a "Hydrochemical Data Base."

Selection of analyses for inclusion in the data base generally was governed by ionic balance, supporting data, and implied continuity with adjacent analyses. In the absence of analytical data throughout an extensive area, the concentration of dissolved solids was estimated from wireline-geophysical logs by measurement of spontaneous potential or by cross plotting resistivity and porosity data. The procedure for estimating water resistivity and dissolved-solids concentration is described by Jorgensen (1988).

HYDRAULIC HEAD

The configuration of the potentiometric surface is useful in evaluating the continuity of an aquifer and to show areas of recharge and discharge. The potentiometric surface is defined by the altitude of the water level in tightly cased wells. In rocks containing variable-density fluids in which density is not a function of pressure, no potential field can be defined (Jorgensen and others, 1982). However, a surface defined by equivalent freshwater head is useful in evaluating the lateral hydraulic continuity of the water-yielding rocks.

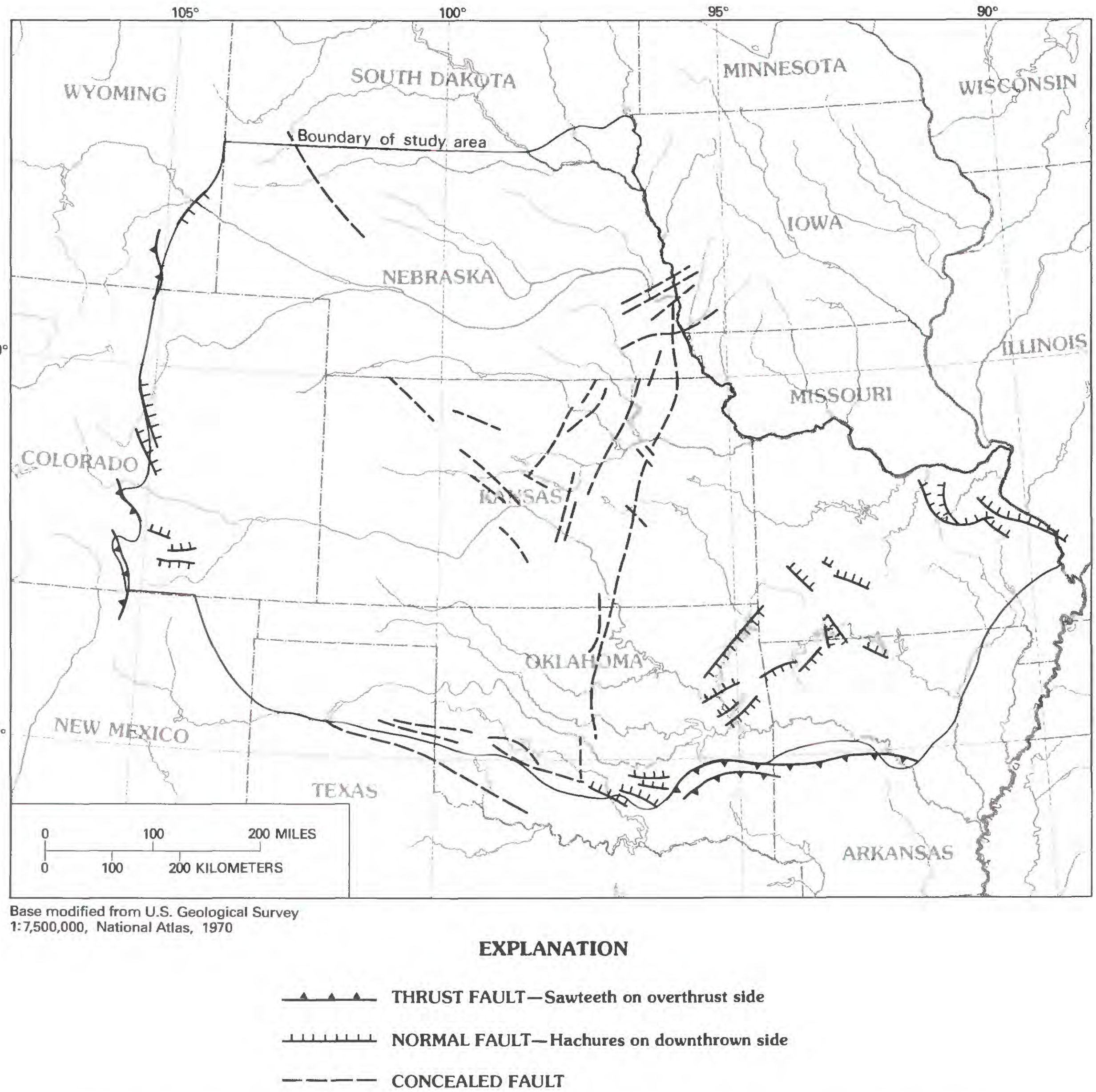


FIGURE 10.—Major faults in study area (modified from King, 1968; Sims, 1985; Krueger and Keller, 1986).

Altitudes of equivalent freshwater head were determined from shut-in pressures of drill-stem tests or by measurements of the altitude of the water level in wells containing freshwater. The following equation was used to calculate equivalent freshwater head (h_e):

$$h_e = z + \frac{p}{dg}, \tag{1}$$

where

- z is the altitude at the centerline of the rock interval tested (L);
- p is the pressure at the centerline of the rock section (FL^{-2});
- d is density of freshwater (ML^{-3}); and
- g is the acceleration due to gravity (LT^{-2}).

Pressure values from the drill-stem tests were adjusted to the centerline of the rock section. The adjustment to pressure (p_a) was calculated by

$$p_a = d_f g\,(z_g - z), \tag{2}$$

where

- d_f is the density of the formation water; and

z_g is the altitude of the pressure gage used for the drill-stem test.

Density of water is a function of the concentration of dissolved solids and temperature. In shallow wells containing freshwater, the correction for density is trivial, and the equivalent freshwater head is considered equal to the altitude of the water level within the well.

Accuracy of the calculated freshwater-head values using results from drill-stem tests is difficult to evaluate because it is dependent on the accuracy of the recording gage and the conditions of the test, including the hydraulic character of the rock. Comparison of data from adjoining drill-stem tests indicates considerable differences. The differences may be due either to unknown inaccuracies as reported in the drill-stem test results, uncertainties in fluid density and temperature, or to areal, vertical, or temporal variations, such as the effects of petroleum production in nearby wells.

Potentiometric surfaces mapped for this study are estimated predevelopment hydraulic-head distributions. In relatively shallow, freshwater parts of the aquifers, considerable water-level data were available, and development generally has not caused large regional water-level declines. In deeper, saline parts of the aquifers, data are scarce in some areas and reflect the complexities of petroleum development, as discussed above, in other areas. Thus, varying degrees of interpretation were required to estimate the predevelopment potentiometric surfaces.

Data of selected water-level measurements in wells were stored in the Ground-Water Site Inventory (GWSI) data base of the U.S. Geological Survey. Results of drill-stem tests were stored in a CMRASA file (Helgesen and Hansen, 1989), "Reservoir Parameter File."

POROSITY

Porosity of saturated rock is the ratio of the volume of interstices in the rock to its total volume (Lohman, 1972, p. 3). Porosity is an important element in both storage and transmissive properties of a geohydrologic unit.

Porosity values were available throughout most of the study area from analyses of the results of drill-stem tests, data from laboratory analyses of test-hole cores, and analyses of production data from oil and gas reservoirs. However, these data were not considered to be definitive of regional values. Cores are very small samples, and for fractured rock, they may not be in their original configuration and they rarely if ever indicate the size of major openings along fractures. Erroneous values may be obtained, especially in regard to repacked samples of unconsolidated material. Reported porosity values for many oil and gas reservoirs are of unknown accuracy and determined by unknown methods. Formation porosity can be determined from borehole wireline-geophysical measurements, such as density (gamma-gamma), neutron, and sonic logs (MacCary, 1978, p. 9).

For the CMRASA, porosity values were determined from borehole-compensated dual-porosity logs (density and neutron) where available. Those data were used to prepare preliminary maps of regional porosity. Porosity values determined from logs were considered to better represent geohydrologic units in which they were made because, generally, the values include the effect of the thickness of the entire unit. The porosity values determined from geophysical logs were for entire rock sections with similar lithologic and hydraulic characteristics and usually with several hundred feet or more of thickness.

A comparison between reported porosities for oil and gas reservoirs and log-derived values on the preliminary regional porosity maps was made for different lithologies and different stratigraphic units. Mapped values correlated well with porosity values of reservoir-rock sections that consisted of Jurassic and Lower Cretaceous sandstone (fig. 11). The good correlation likely results from the dominance of primary

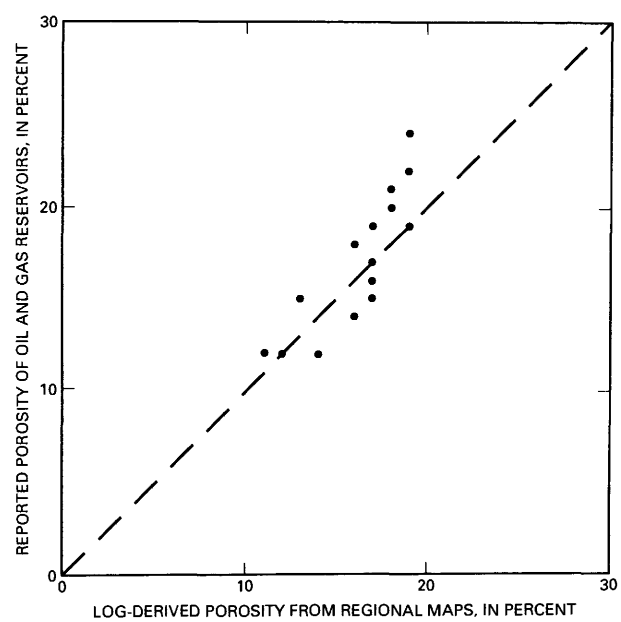


FIGURE 11.—Relation of regional (log-derived) to reported oil- and gas-reservoir porosity of Jurassic and Lower Cretaceous sandstone.

porosity that is more homogeneous in nature than fracture or secondary porosity.

Reservoir porosity values reported in the literature for Cambrian and Devonian rocks, Mississippian limestone, and Pennsylvanian and Permian rocks did not correlate well with regionalized (log-derived) porosity values (figs. 12–14). The poor correlations are attributed partly to the effects of heterogeneous secondary porosity that resulted principally from fracturing and dissolution of rock material. Oil and gas reservoirs in a heterogeneous rock section typically have larger porosity values than average values for the rock section (Ray and others, 1985). In general, oil and gas reservoirs have larger porosity than the regional rock unit within which they occur. This is especially true in fractured rocks. The larger porosity in oil and gas reservoirs generally occurs in thin vertical sections that are usually of limited lateral extent. Therefore, reported porosity values for oil and gas reservoirs are generally site specific and may not be representative of the entire section of the rock unit, either vertically or areally.

Porosity data indicate a general trend of decreasing porosity with depth. This is consistent with Davis' (1969, p. 59) statement and supporting data that permeability and porosity of dense rocks decrease with depth. The relation is well defined by sandstone in the study area, which generally has dominant primary porosity; however, the trend is less defined in carbonate rocks, which are dominated by secondary porosity.

For the deepest rocks in the study area, porosity data of any type or source were not available. Estimates of porosity in these rocks were made based on porosity data of similar rocks at lesser depths and the relation of decreasing porosity with depth. These porosity estimates were used to supplement the geophysical-log porosity values on the preliminary porosity maps of the geohydrologic units. From the combined data, final porosity maps were prepared.

HYDRAULIC CONDUCTIVITY AND INTRINSIC PERMEABILITY

Hydraulic conductivity is the primary criterion for distinguishing between aquifers, which convey water, and confining units, which restrict or confine water movement. Hydraulic-conductivity data for aquifers containing freshwater are largely from aquifer tests using one or more wells that completely penetrate a single aquifer and from estimates based on the specific capacity of the pumping well. Specific capacity is the rate of well discharge per foot of drawdown and is a function of transmissivity and well efficiency. Available specific-capacity data were from wells that were usually open to one or more aquifers but that generally did not penetrate the complete thickness of the

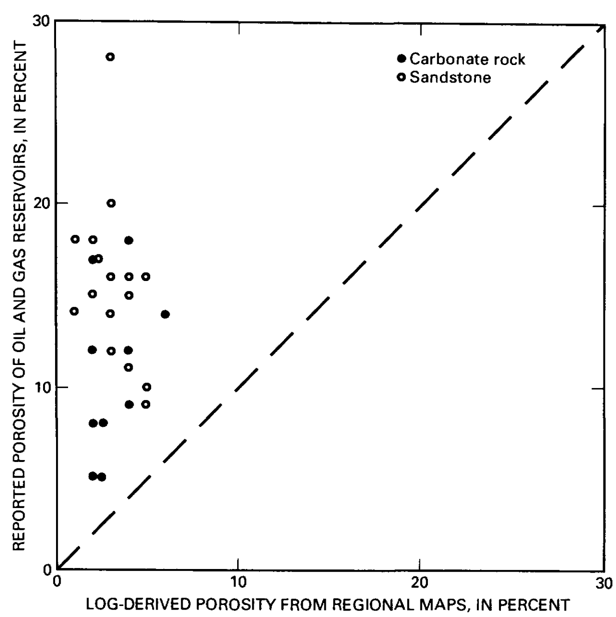


FIGURE 12.—Relation of regional (log-derived) to reported oil- and gas-reservoir porosity of Cambrian and Devonian rocks.

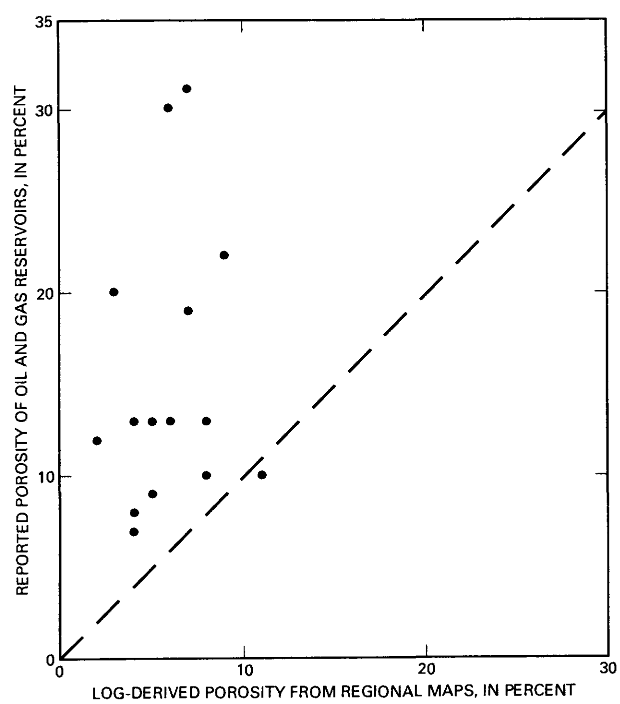


FIGURE 13.—Relation of regional (log-derived) to reported oil- and gas-reservoir porosity of Mississippian limestone.

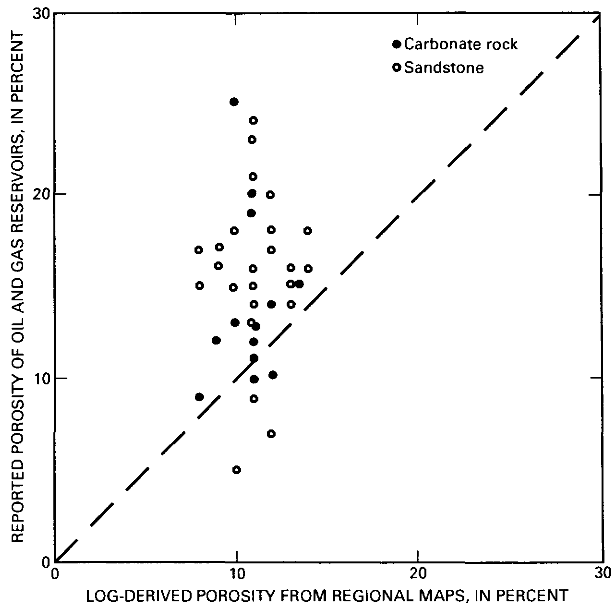


FIGURE 14.—Relation of regional (log-derived) to reported oil- and gas-reservoir porosity of Pennsylvanian and Permian rocks.

lowermost aquifer. Thus, not all specific-capacity data provided accurate estimates for an individual aquifer. Hydraulic-conductivity values determined from aquifer tests or specific-capacity tests are stored in the U.S. Geological Survey's Ground-Water Site Inventory (GWSI) data base.

In rocks containing saline water, virtually no aquifer-test or specific-capacity data were available. In these areas, intrinsic permeability data were obtained from analyses of results from drill-stem tests, from laboratory analyses of rock cores, or from oil-production tests. Commercial quantities of oil are found in reservoirs or traps of relatively permeable rock material. Because nearly all drill-stem tests are conducted in these traps or reservoirs of permeable material, permeability values determined from drill-stem tests may not be representative of the effective regional permeability that is needed for evaluating regional aquifer systems.

Permeability values from cores of fractured rock are difficult to evaluate as to their representativeness of effective regional permeability. For example, permeability values from a 20-ft core of fractured and vuggy dolostone within the study area ranged from less than 1.1×10^{-15} ft^2 to more than 3.2×10^{-11} ft^2 (Jorgensen and others, 1993). The permeability of the fractures in the rock core was not determined because it was not possible to arrange the fractured pieces in the laboratory within the same spacing or in-situ orientation. Both laboratory tests and drill-stem tests generally are conducted on relatively thin rock sections of 20 ft or less. In general, laboratory tests of the thin intervals tested are not representative of the thick regional geohydrologic units. Therefore, a method was developed to estimate the permeability of geohydrologic units using data from boreholes with a specific suite of wireline geophysical logs.

The permeability relation used was:

$$k = \frac{\varepsilon}{S_s^2} \cdot \frac{n^{m+2}}{(1-n)^2}\ , \tag{3}$$

where e is a rock constant;
n is porosity;
m is the cementation factor; and
S_s is the specific surface area (Jorgensen, 1988).

The second set of terms to the right of the equal sign of equation 3 is termed the porosity factor $(\bar{p})$ from which an empirical equation for intrinsic permeability was developed. The equation is:

$$k = (1.828 \times 10^5)\,\bar{p}^{\ 1.10} \tag{4}$$

Intrinsic-permeability values from aquifer and specific-capacity tests and calculated estimates of intrinsic permeability using geophysical-log data and equation 4 were plotted on maps, and lines of equal value were drawn to show regionalized intrinsic permeability. These values were the initial estimates of intrinsic permeability for a digital finite-difference flow model of the geohydrologic units. The modeling procedure itself can be used to refine estimates of rock characteristics, such as permeability, if adequate data are available. The modeling procedure, which will be discussed in detail later in this report, therefore was used to further refine initial estimates.

Regionalized permeability values were compared to site-specific permeability values reported for oil and gas reservoirs, as shown in figures 15–18. The figures show the poor correlation between reported reservoir permeability and regionalized permeability and that differences of two orders of magnitude are common. Permeability values for reservoirs are site specific and were measured in relatively thin sections of rock, such as 20 ft or less, whereas the regionalized permeability values calculated from geophysical-log data were for thick sections, such as 100 to 1,000.ft. Similar to the comparison of regional porosity to reservoir porosity, it is concluded that permeability values determined from maps of estimated permeability based on geophysical-log data more accurately represent regional permeability than reported values of oil- and gas-reservoir permeability.

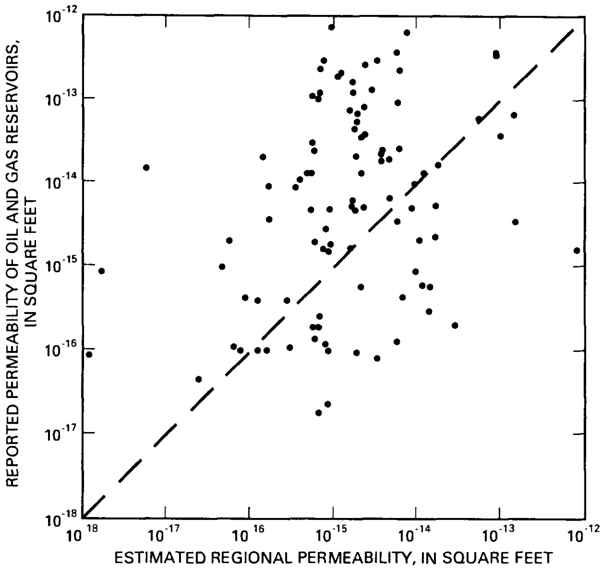


FIGURE 15.—Relation of estimated regional (log-derived) to reported oil- and gas-reservoir intrinsic permeability of Cambrian and Ordovician dolostone.

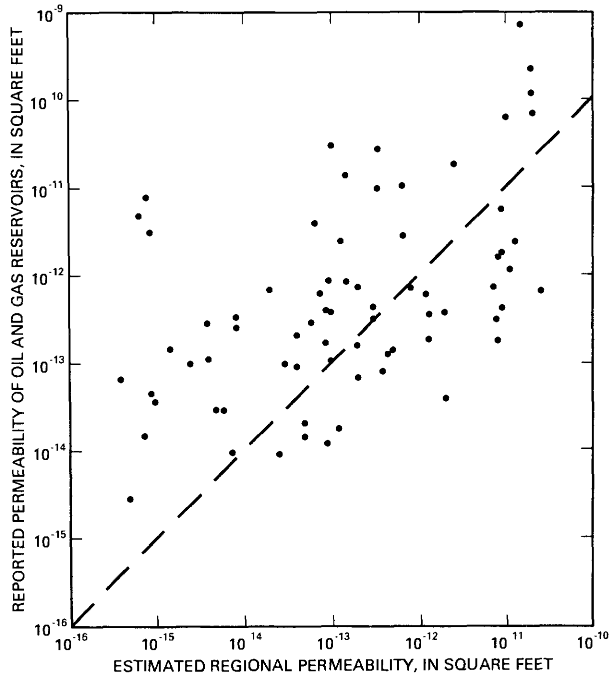


FIGURE 16.—Relation of estimated regional (log-derived) to reported oil- and gas-reservoir intrinsic permeability of Mississippian limestone.

GEOHYDROLOGY

Geohydrology of the study area is described for two subdivisions, which for clarity and convenience are referred to as the Plains subregion and the Ozark subregion. Jorgensen and others (1993) present a geologic description of the major geohydrologic units in the two subregions and discuss the geologic and hydrologic history of the regions. Jorgensen and others (1993) define six regional geohydrologic units in the Plains subregion and three regional geohydrologic units in the Ozark subregion. The correlation of these regional geohydrologic units to rock- and time-stratigraphic units is summarized in tables 1 and 2; subsurface relations and areal extent of the geohydrologic units are shown in figures 19 and 20. Geohydrologic units that were studied in detail are the Great Plains aquifer system, the Western Interior Plains confining system, which is common to both the Plains and Ozark subregions, the Western Interior Plains aquifer system, and the Ozark Plateaus aquifer system. Maps showing the thickness of and altitude of the tops of the principal aquifers and confining units composing these four regional systems are presented by Jorgensen and others (1993). The remaining geohydrologic units were not studied in detail for the CMRASA; however, the High Plains aquifer was studied as the subject of a separate regional aquifer-system analysis (fig. 2) (Weeks, 1978).

This section briefly describes the geology of the principal aquifers and confining units making up the four regional systems and presents information on their hydraulic properties, hydraulic heads, and dissolved-solids concentrations. The basement confining unit of Precambrian rocks is discussed first, and then the two subregions are discussed separately beginning with the geohydrologic units containing the oldest rocks.

The lowermost confining unit in both the Plains and Ozark subregions is the basement confining unit, composed mostly of crystalline rocks (tables 1, 2). The rocks are fractured and yield small quantities of water to wells at many locations, such as in the mountains west of Denver, Colorado, in southeastern South Dakota, and locally where the rocks crop out in the St. Francois Mountains. However, on a regional basis, they are assumed to form the base of ground-water flow in the Plains and Ozark subregions. The top of the basement confining unit ranges from less than 34,000 ft below sea level in southwestern Oklahoma to more than 1,000 ft above sea level in the St. Francois Mountains in eastern Missouri. It is about 500 ft above sea level in southeastern South Dakota and more than 7,000 ft below sea level along the Rocky Mountains in central Colorado (Jorgensen and others, 1993, pl. 3).

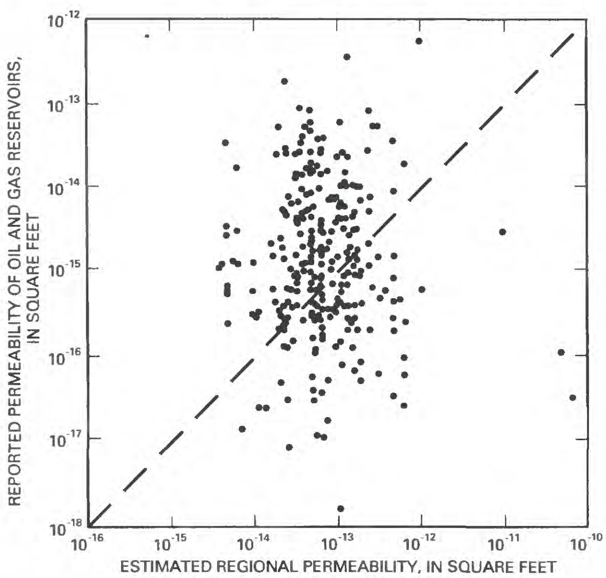


FIGURE 17.—Relation of estimated regional (log-derived) to reported oil- and gas-reservoir intrinsic permeability of Pennsylvanian and Permian sandstone and limestone.

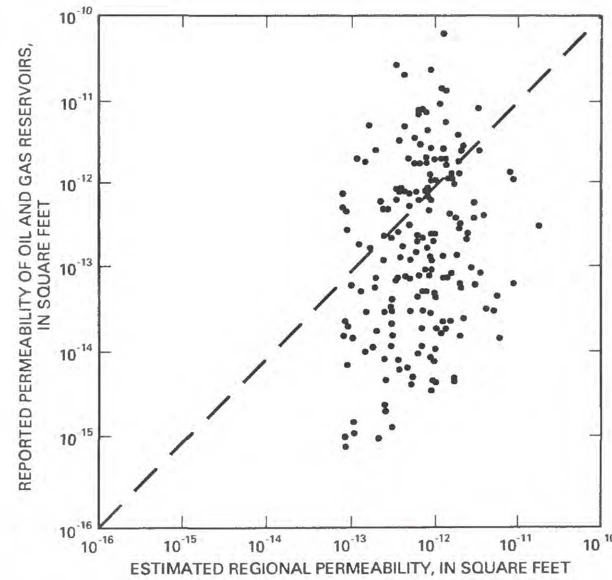


FIGURE 18.—Relation of estimated regional (log-derived) to reported oil- and gas-reservoir intrinsic permeability of Jurassic and Lower Cretaceous sandstone.

PLAINS SUBREGION

The principal geohydrologic units (table 1) in the Plains subregion are the basement confining unit, the Western Interior Plains aquifer system, the Western Interior Plains confining system, the Great Plains aquifer system, the Great Plains confining system, and the High Plains aquifer. These geohydrologic units, with the exception of the previously described basement confining unit, are discussed in the following paragraphs.

WESTERN INTERIOR PLAINS AQUIFER SYSTEM

The Western Interior Plains aquifer system, at most locations, consists of aquifers of permeable limestone, dolostone, and sandstone of Late Cambrian through Late Mississippian age that are separated by slightly permeable shale or dolostone all of which overlay the basement confining unit. The stratigraphic units included in the aquifer system are listed in table 1. Collectively, the aquifer system contains permeable and slightly permeable beds that function regionally as a hydraulic unit.

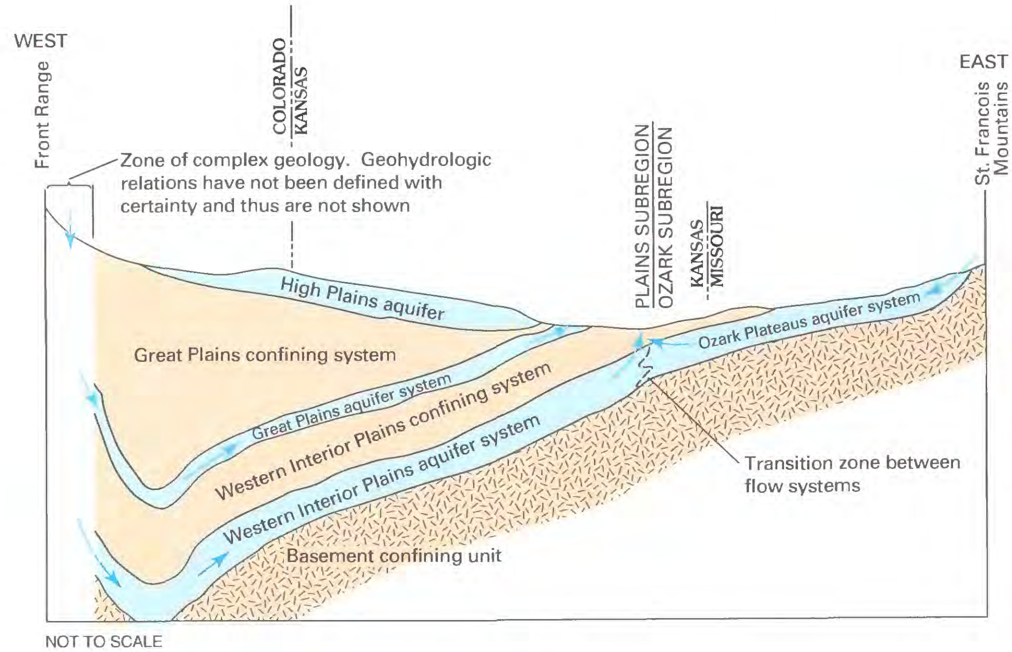
The Western Interior Plains aquifer system extends from the Rocky Mountains eastward to a transitional boundary with the Ozark aquifer system (fig. 20). The aquifer system is not present over large areas of western Nebraska and northern Colorado. The southern boundary of the system along the faulted and folded area in southern Oklahoma is considered to be mostly a no-flow boundary. The regional study boundary in southeast Oklahoma and northwest Arkansas also approximates a no-flow boundary.

The Western Interior Plains aquifer system consists of an upper unit, a confining unit, and lower units. The three units are reasonably distinct at most locations but are within the same hydrologic system. This conceptualization of upper and lower geohydrologic units is implied from equivalent freshwater heads and water chemistry.

Estimated lateral intrinsic permeability of rocks in the Western Interior Plains aquifer system is very small in the western part of the system (figs. 21 and 22) but increases toward the east by about six to seven orders of magnitude.

The top of the lower units in the Western Interior Plains aquifer system generally dips away from the Ozark Uplift and Missouri River toward the Anadarko, Arkoma, and Denver Basins (fig. 7). The lowest altitude is in the Anadarko Basin at more than 26,000 ft below sea level. Near the Arkoma Basin, the altitude of the top is more than 10,000 ft below sea level and near Denver, Colorado, more than 7,000 ft below sea level.

In northwestern Nebraska, northeastern Colorado, and scattered areas of central Kansas, the lower units are absent on the Cambridge Arch, the Chadron Arch, and the Central Kansas Uplift. The lower units are separated laterally from similar carbonate rocks in South Dakota and Wyoming by the Chadron Arch and the Siouxana Arch. The thickest section of the lower units exceeds 10,000 ft in southwestern



EXPLANATION

DIRECTION OF GROUND-WATER FLOW

FIGURE 19.—Cross section showing geohydrologic units from central Colorado to eastern Missouri.

Oklahoma and generally exceeds 1,000 ft in Oklahoma, southern Kansas, and eastern Nebraska.

Estimated lateral intrinsic permeability ranges from $1x10^{-11}$ ft^2 in southeastern Kansas to less than $1x10^{-17}$ ft^2 in the western Anadarko Basin in Oklahoma and less than $1x10^{-18}$ ft^2 in the central Denver Basin in Colorado (fig. 21). A lateral intrinsic permeability of $1x10^{-12}$ ft^2 for rocks making up the aquifer on the Central Kansas Uplift was determined by a ground-water flow model analysis by Carr and others (1986). Figure 21 shows the estimated mean lateral intrinsic-permeability values for that area of the Central Kansas Uplift to be about $1x10^{-12}$ to $1x10^{-15}$ ft^2, a difference of three orders of magnitude. The lateral intrinsic permeability values shown in figure 21 were used as initial estimates for model simulation of the lower units in the Western Interior Plains aquifer system, and the range of variability is discussed in a later section. The very small permeability values are due to the effect of depth on permeability. In general, values of lateral intrinsic permeability less than $1x10^{-15}$ ft^2 are considered typical of confining material. However, the lower units are considered to be regionally distinct as an aquifer in the Western Interior Plains aquifer system. Regionalized porosity generally ranges from 1 to 5 percent for the lower units (fig. 23).

Dissolved-solids concentrations in water from the lower units range from 280,000 mg/L in central Oklahoma to 10,000 mg/L at the western limit and less than 1,000 mg/L in the transition zone to the freshwater flow system of the Ozark subregion (fig. 24). The most distinctive characteristic of the water chemistry in the lower units in the Western Interior Plains aquifer system is the extensive area in which the water contains extremely large concentrations of dissolved solids.

Lines of equal predevelopment equivalent freshwater head for the lower units (plate 1) show a large hydraulic gradient east and northeastward from the Rocky Mountains toward the Central Kansas Uplift; a southeastward gradient between the Central Kansas Uplift and the Missouri River in eastern Nebraska; generally eastward gradients transverse to the Nemaha Uplift in eastern Kansas and east-central Oklahoma; and an area of relatively low equivalent freshwater head in southeastern Kansas and northeastern Oklahoma. Two additional distinct features are apparent from the lines of equivalent freshwater head (plate 1). The first is an area of high equivalent freshwater head in the vicinity of the Anadarko Basin in western Oklahoma and the Oklahoma and Texas Panhandles, which likely is related to an overpressure zone in the sandstone of the Western Interior Plains

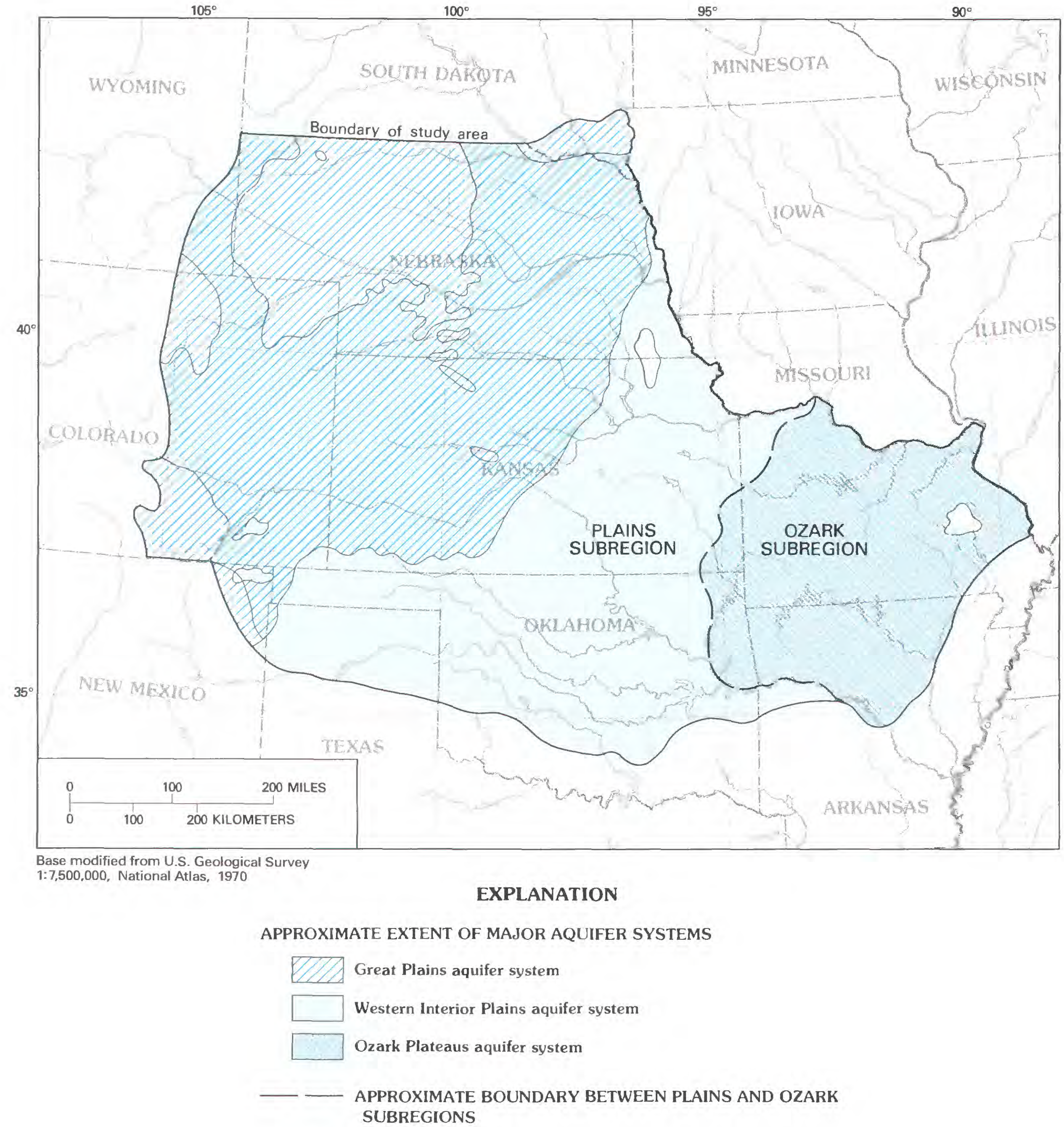


FIGURE 20.—Extent of major aquifer systems in study area.

confining system. The second is the area of contorted contours of equivalent freshwater head in eastern Missouri and adjacent Arkansas, which results from natural recharge to the near-surface rocks and local discharge to streams in this area.

The intermediate confining unit that separates the lower units from the upper unit in the Western Interior Plains aquifer system limits regional flow between the two units. The confining unit is present over most of the study area in Oklahoma, except in the panhandle; in the eastern one-half of Kansas, except over the Nemaha Uplift; and in the extreme southeastern corner of Nebraska. The confining unit is thin relative to other regional units in the study area. The thickness varies generally from about 300 ft thick in northeastern Kansas and southeastern Nebraska to zero where absent.

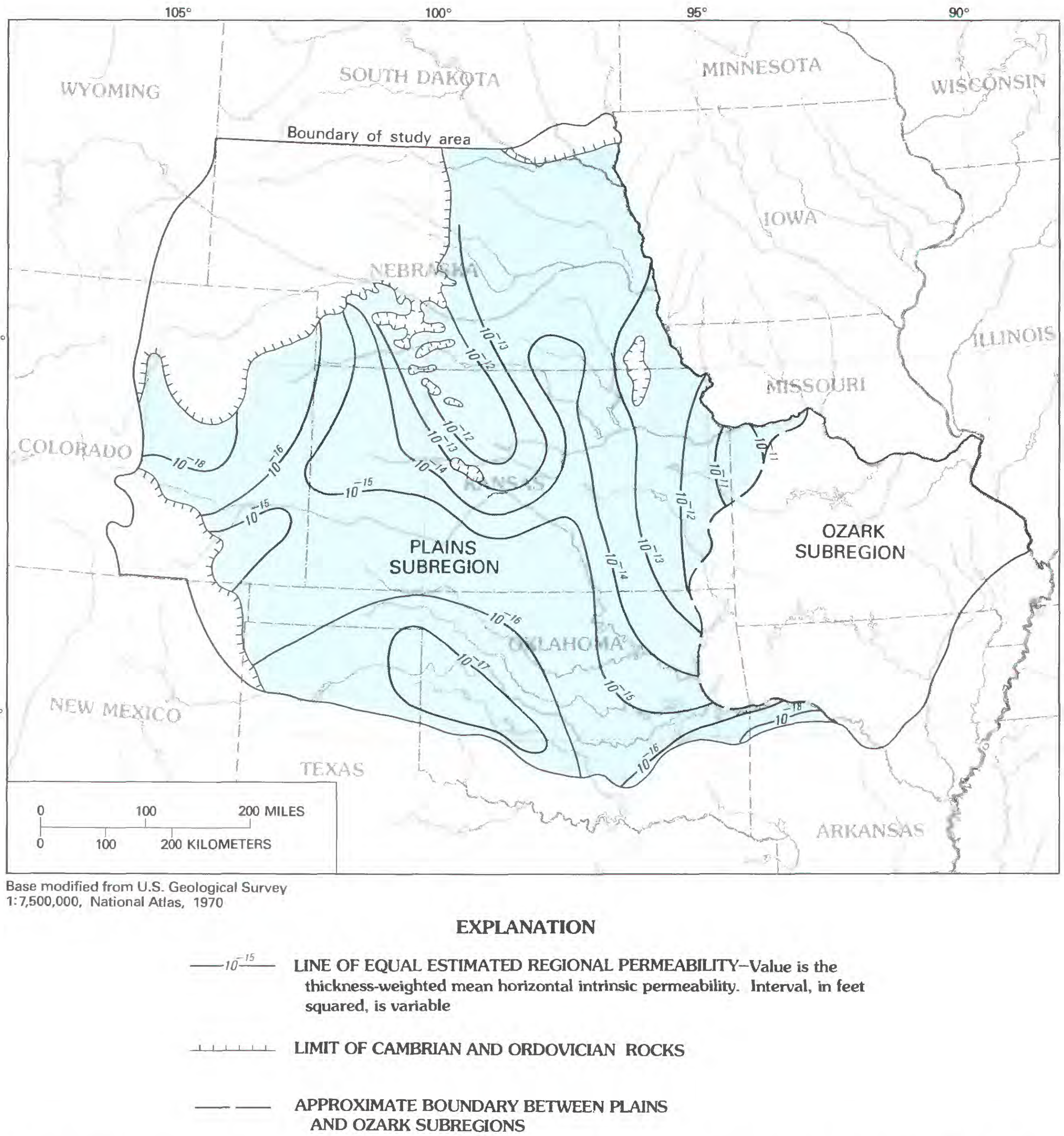


FIGURE 21.—Estimated lateral intrinsic permeability of lower units in the Western Interior Plains aquifer system and the Ozark and St. Francois aquifers in the Ozark Plateaus aquifer system.

The top of the upper unit in the Western Interior Plains aquifer system generally slopes away from the Missouri River and the Ozark subregion toward the Denver, Arkoma, and Anadarko Basins; the lowest altitude is less than -22,000 ft in the Anadarko Basin of southwestern Oklahoma. The upper unit in the study area is absent from areas in southeastern Nebraska and northeastern Kansas over the Nemaha Uplift, in the extreme southeastern corner of South Dakota, and from about central Nebraska west almost to the Wyoming State line. It is also absent from northeastern Colorado to central Kansas where the upper unit is eroded from the Cambridge Arch and Central Kansas Uplift.

The thickest section of the upper unit exceeds 2,500 ft in western Oklahoma where maximum deposition has occurred, but generally the unit has a thickness of several hundreds of feet elsewhere.

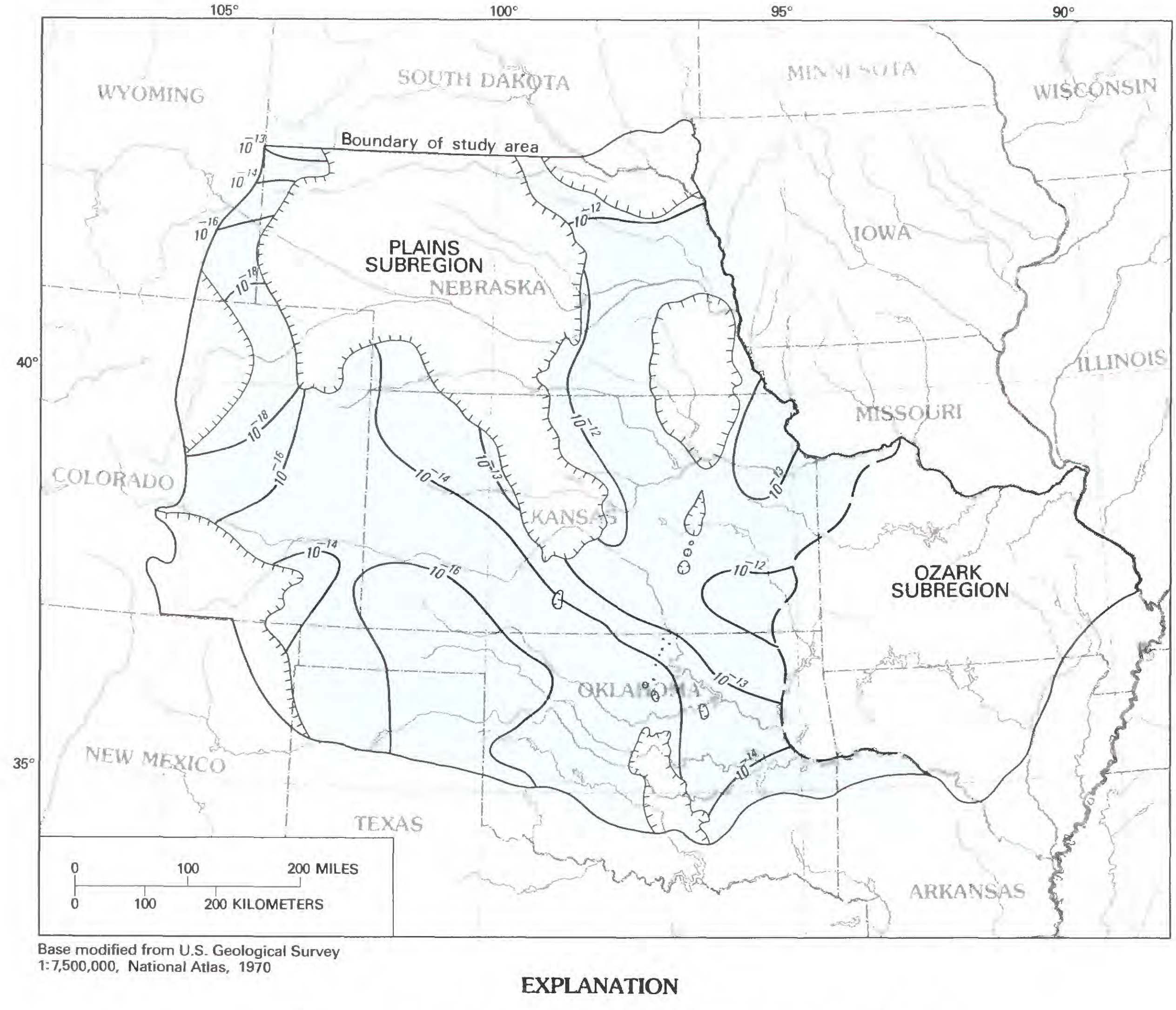


FIGURE 22.—Estimated lateral intrinsic permeability of upper unit in the Western Interior Plains aquifer system and the Springfield Plateau aquifer in the Ozark Plateaus aquifer system.

Additionally, the rocks in the deep basins have not been exposed to erosion and diagenesis associated with uplift. Accordingly, the rock sections in the Anadarko and Denver Basins are less permeable than rocks in the same stratigraphic units in other areas (fig. 22). In contrast, the dolostone and limestone in and adjacent to the Ozark Uplift, the Cambridge Arch-Central Kansas Uplift, and the Nemaha Uplift in southeastern Nebraska and northeastern Kansas are fractured, have solution openings, and are very permeable because they have been exposed many times to weathering for extensive periods.

Estimated values of lateral intrinsic permeability for the upper unit range from $1x10^{-12}$ ft^2 in eastern Kansas to less than $1x10^{-16}$ ft^2 in the Anadarko Basin in Oklahoma and $1x10^{-18}$ ft^2 in the Denver Basin in Colorado (fig. 22). These values were used to estimate hydraulic-conductivity values for model simulation of the upper unit, and their range of variability is discussed in later sections. As with the lower units in the

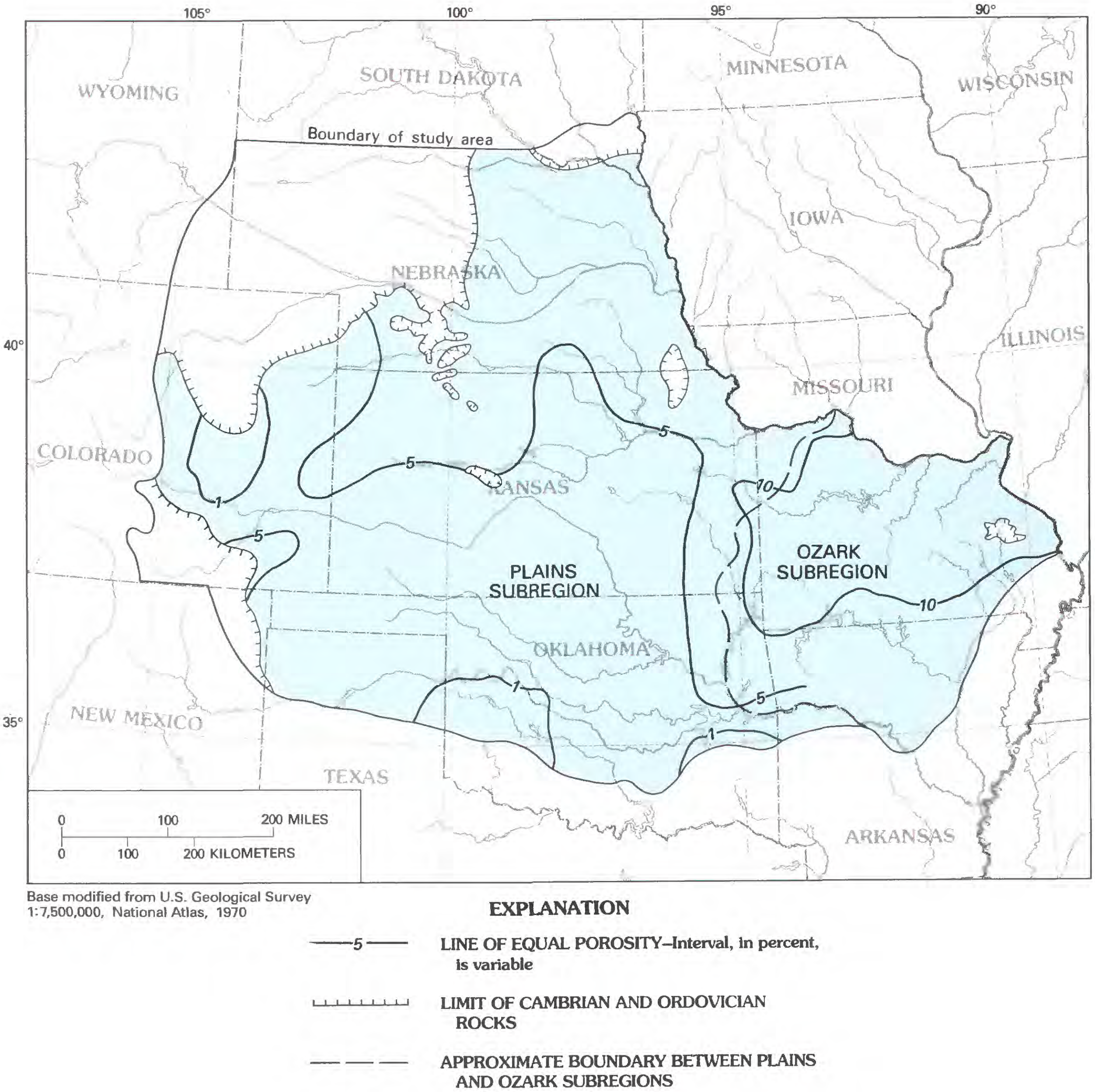


FIGURE 23.—Regional porosity of lower units in the Western Interior Plains aquifer system and the Ozark and St. Francois aquifers in the Ozark Plateaus aquifer system.

Western Interior Plains aquifer system, the very small mean lateral intrinsic-permeability values occur in the deeply buried parts of the unit. Comparison of the intrinsic permeability and porosity of the lower units (figs. 21, 23) to the intrinsic permeability and porosity of the upper unit (figs. 22, 25) indicates that the upper unit is much more permeable and porous than the lower units over much of the study area, although the values are nearly the same where both units are deeply buried.

Regional porosity of the upper unit (fig. 25) shows the same effect of depth of burial as described for permeability. The porosity is small (about 1 percent) in and near major basins not subjected to uplift and erosion and increases to 10–15 percent near the Central Kansas, Nemaha, and Ozark Uplifts.

Dissolved-solids concentrations of water in the upper unit range from more than 250,000 mg/L in small areas of north-central Oklahoma and south-central Kansas to less than 10,000 mg/L in the

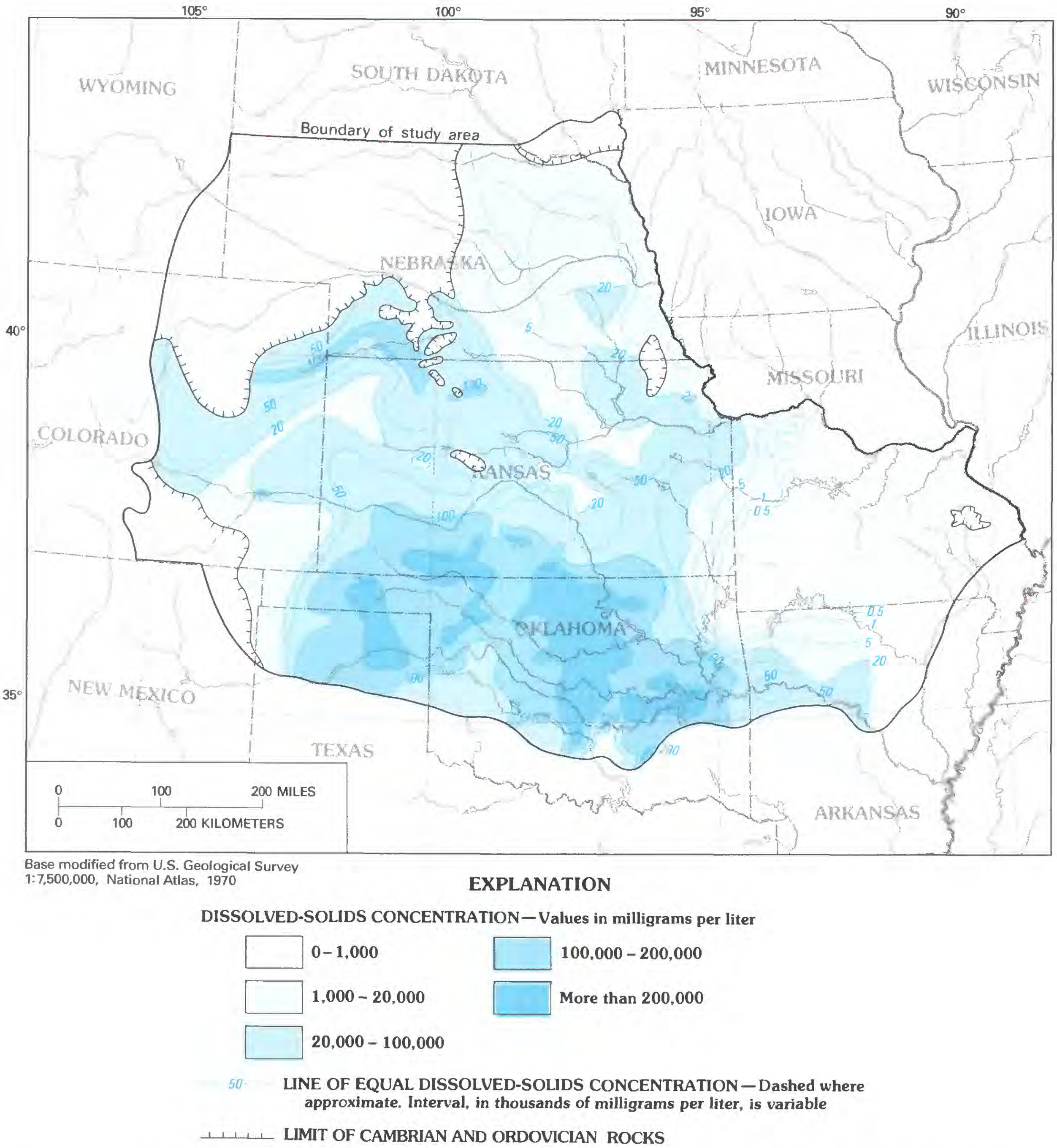


FIGURE 24.—Dissolved-solids concentrations in water from lower units in the Western Interior Plains aquifer system and the Ozark aquifer, the St. Francois confining unit, and the St. Francois aquifer in the Ozark Plateaus aquifer system.

transition zone adjacent to the freshwater flow system in the Ozark subregion and in east-central Nebraska (fig. 26). These large concentrations were expected because the water may represent a paleomarine evaporative brine introduced in the geologic past, a hypothesized, slow flow rate and long contact time, and minimal recharge through the overlying confining system at most locations. The velocity of flow is very slow because in much of the area both intrinsic permeability and hydraulic gradient are small.

Gradients in the upper unit shown on the map of equivalent freshwater head (pl. 2) are similar to those described for the lower units. Gradients are generally east-northeast in the western part of the unit and

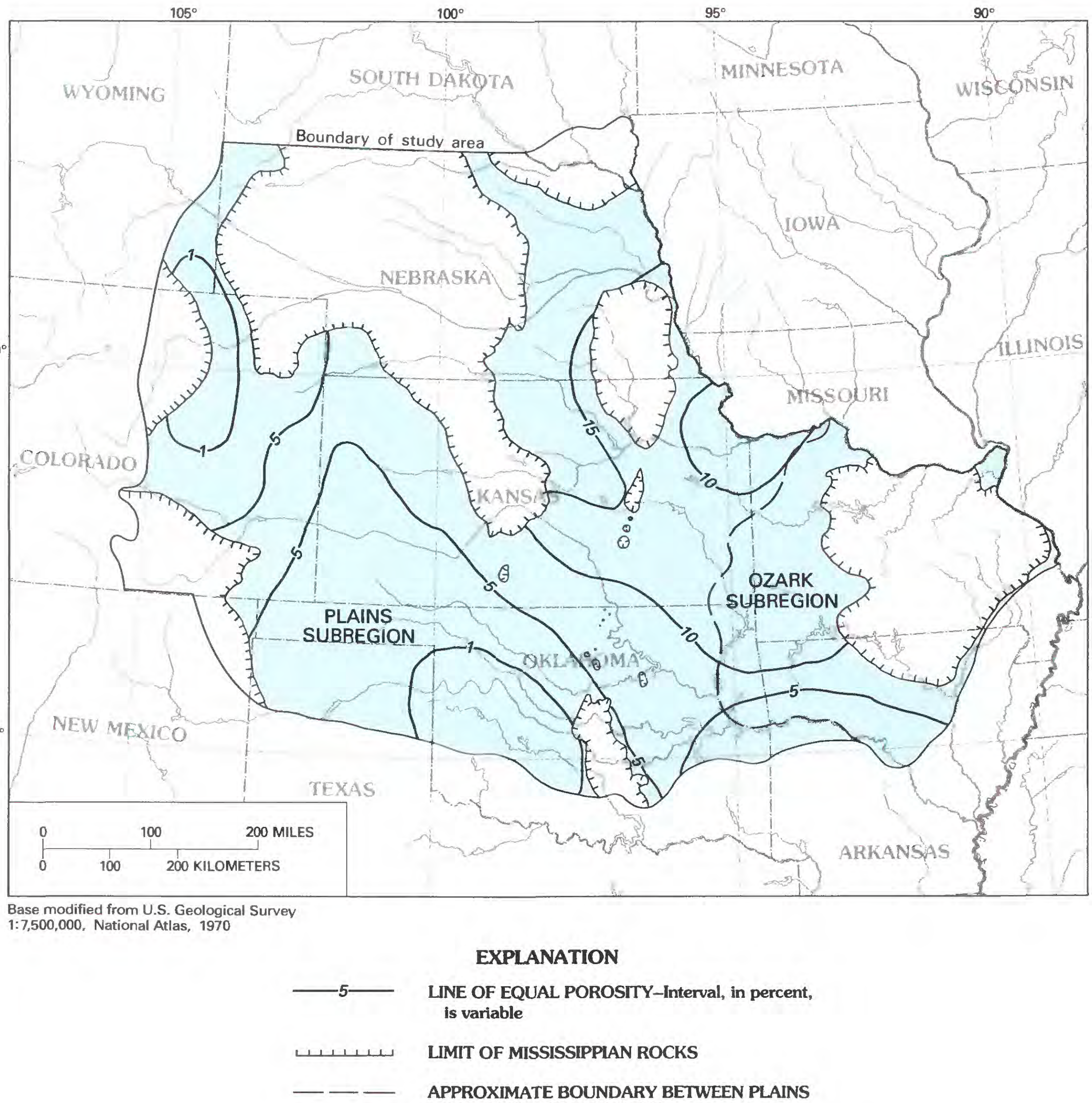
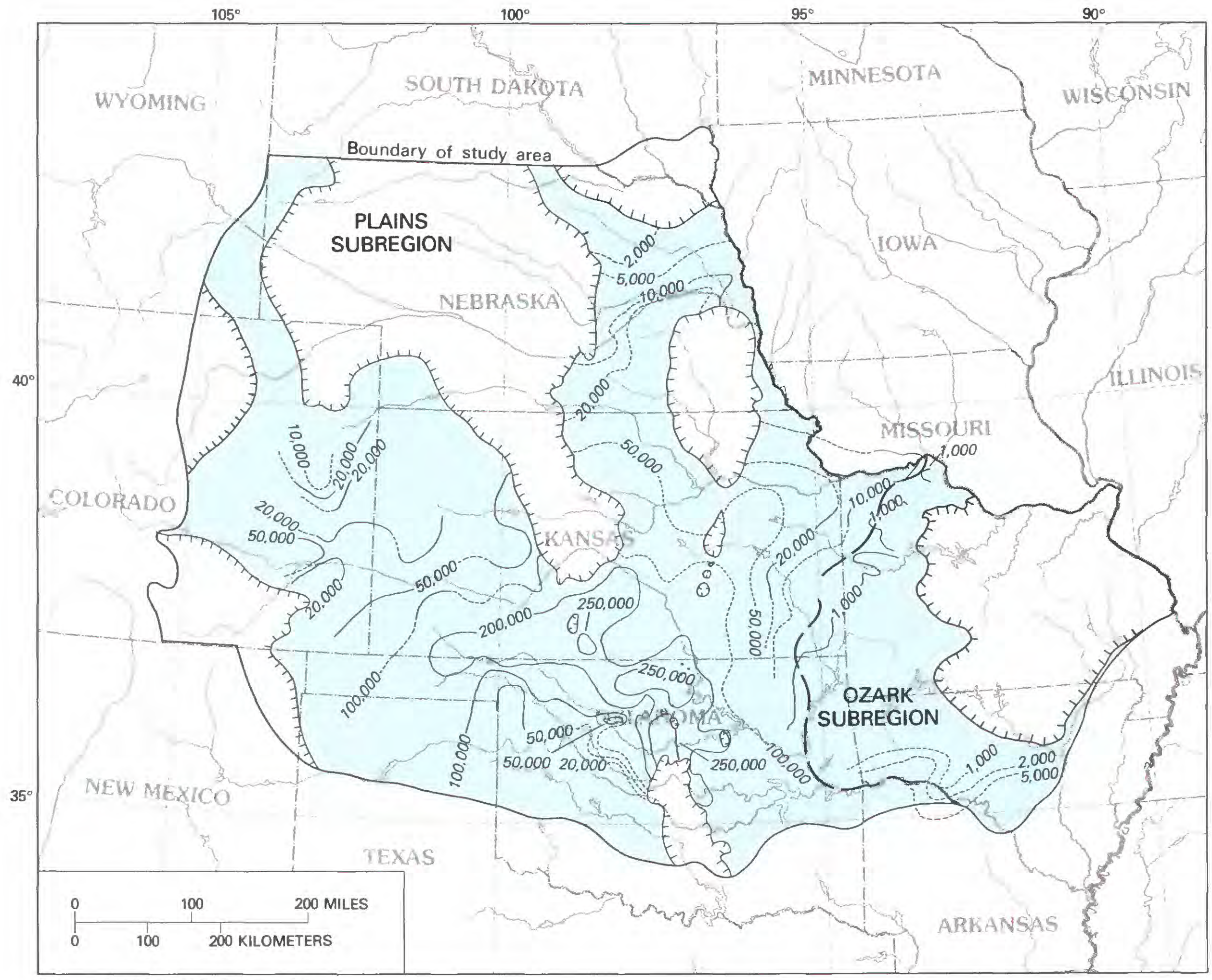


FIGURE 25.—Regional porosity of upper unit in the Western Interior Plains aquifer system and the Springfield Plateau aquifer in the Ozark Plateaus aquifer system.

south-southeast in the northern and eastern parts. Moderately small equivalent freshwater-head values occur in southeastern Kansas and eastern Oklahoma, and areas of large hydraulic head associated with the Anadarko Basin in western Oklahoma and the eastern part of the Texas Panhandle are coincident with an overlying geopressure zone in Morrowan rocks (table 1).

WESTERN INTERIOR PLAINS CONFINING SYSTEM

The Western Interior Plains confining system separates the Great Plains aquifer system and the Western Interior Plains aquifer system. The system restricts vertical flow to and from the Western Interior Plains aquifer system. It is the most extensive unit in the CMRASA study area. The confining system is



Base modified from U.S. Geological Survey 1:7,500,000, National Atlas, 1970

EXPLANATION

—5,000--- LINE OF EQUAL DISSOLVED-SOLIDS CONCENTRATION IN WATER FROM UPPER UNIT IN THE WESTERN INTERIOR PLAINS AQUIFER SYSTEM OR SPRINGFIELD PLATEAU AQUIFER IN THE OZARK PLATEAUS AQUIFER SYSTEM—Dashed where approximate. Interval, in milligrams per liter, is variable

LIMIT OF MISSISSIPPIAN ROCKS

APPROXIMATE BOUNDARY BETWEEN PLAINS AND OZARK SUBREGIONS

FIGURE 26.—Dissolved-solids concentrations in water from upper unit in the Western Interior Plains aquifer system and the Springfield Plateau aquifer in the Ozark Plateaus aquifer system.

bounded on the west by the Rocky Mountains and extends eastward to the Ozark subregion. The confining system also extends north and northeast beyond the study area and south through the faulted and folded zone in southern Oklahoma. At the southeastern limit of the study area, the confining system dips steeply into the Arkoma Basin. The altitude of the top ranges from more than 6,000 ft in northeastern New Mexico to less than -3,000 ft in the Denver Basin. In general, in western Missouri, eastern Kansas, and eastern Oklahoma, there is a west-to-east slope toward a broad physiographic low near the outcrop area adjacent to the western perimeter of the Ozark subregion. The thickness of the confining

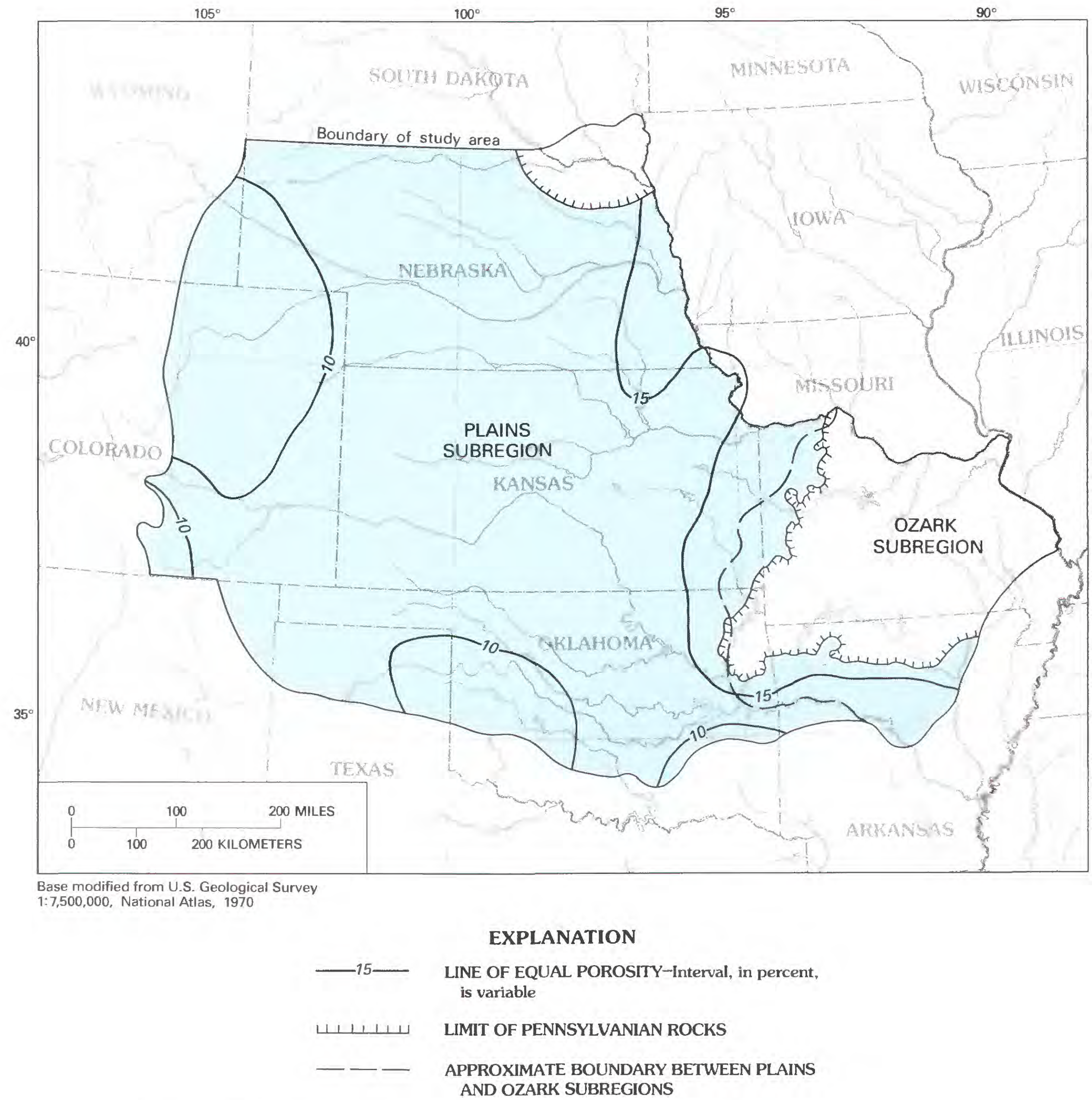


FIGURE 27.—Estimated porosity of sandstone in the Western Interior Plains confining system.

system ranges from zero in northeastern Nebraska and near the Ozark subregion to more than 20,000 ft in the Anadarko Basin of southwestern Oklahoma.

The rocks composing the confining system are of Late Mississippian through Jurassic age but are mostly of Pennsylvanian and Permian age. The Pennsylvanian rocks consist mostly of shale separated by limestone and sandstone layers. In the deeper part of the Anadarko and Arkoma Basins, the ratio of sandstone to limestone increases. In the shallower areas, the ratio of limestone to sandstone exceeds 1. Adjacent to the Amarillo Uplift and Sierra Grande Arch, permeable arkosic gravel is present and is termed "granite wash" because it is believed to have been eroded from a granite uplift.

The Permian rocks of the confining system consist mostly of shale with evaporite deposits, sandstone, and limestone. Adjacent to uplifts, such as the Front Range of the Rocky Mountains, the Sierre Grande, and the Amarillo, permeable arkosic gravel is present. Triassic and Jurassic rocks are mostly shale, sandstone, and limestone.

In general, except in the deeper part of the Arkoma and Anadarko Basins, the sandstone and limestone deposits are aquifer units. The estimated porosity of sandstone, porosity of carbonate rocks, and estimated thickness-weighted mean lateral intrinsic permeability of the confining system are shown in figures 27–29. The aquifer units in the Western Plains confining system are less deeply buried and more permeable than aquifer units in the underlying Western Interior Plains aquifer system; however, leakage through the confining system at most locations is very small because the system includes thick layers of shale and extensive layers of nearly impermeable evaporite deposits (mainly halite and gypsum).

Halite and gypsum deposits are present in the western one-half of Kansas, parts of northwestern Oklahoma, the Oklahoma and Texas Panhandles, eastern Colorado, western Nebraska, and eastern Wyoming (fig. 30). Although not continuous, there appears to be sufficient thickness of overlapping units that the evaporite deposits severely restrict vertical leakage through the confining system; hence, leakage

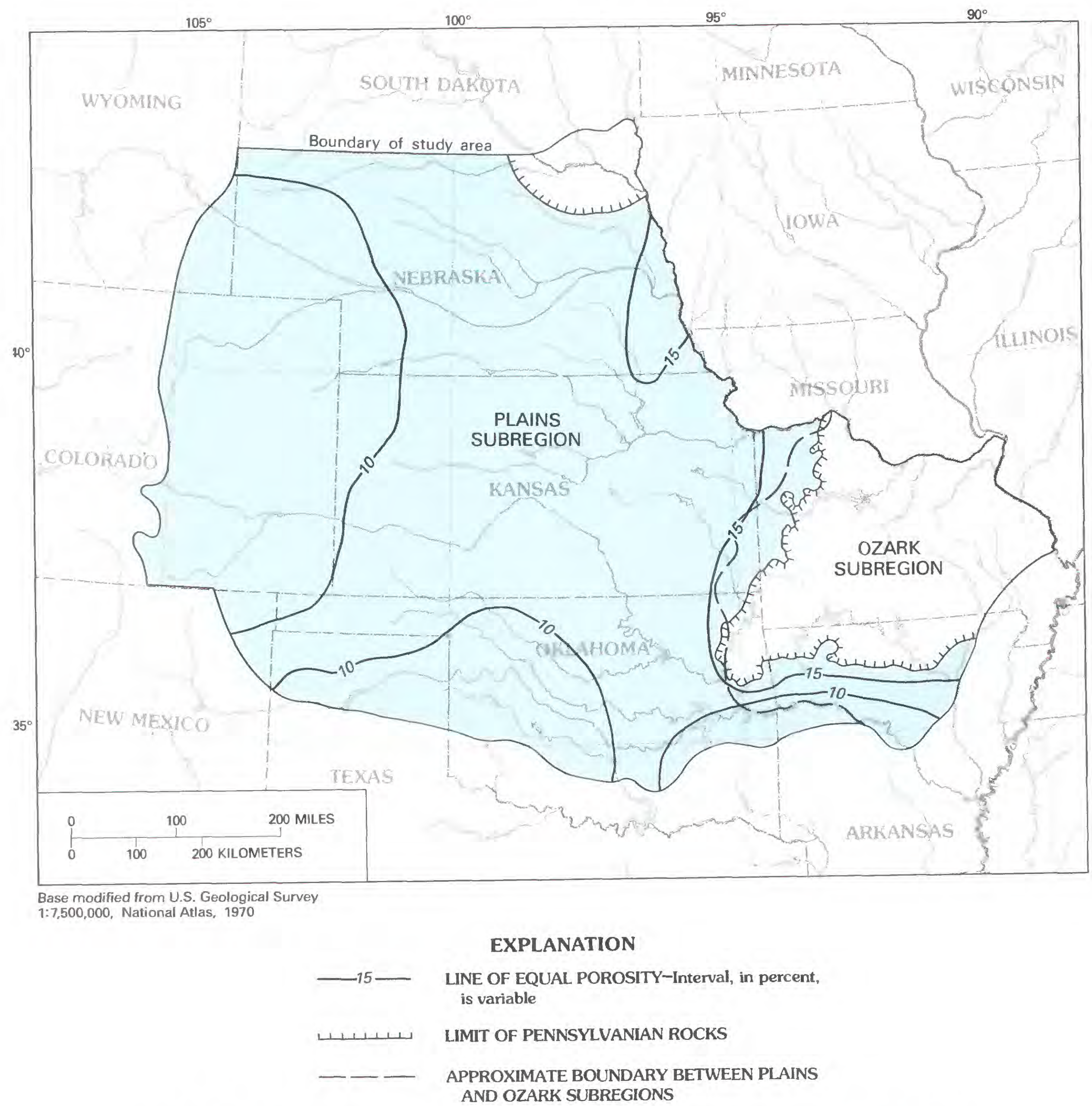
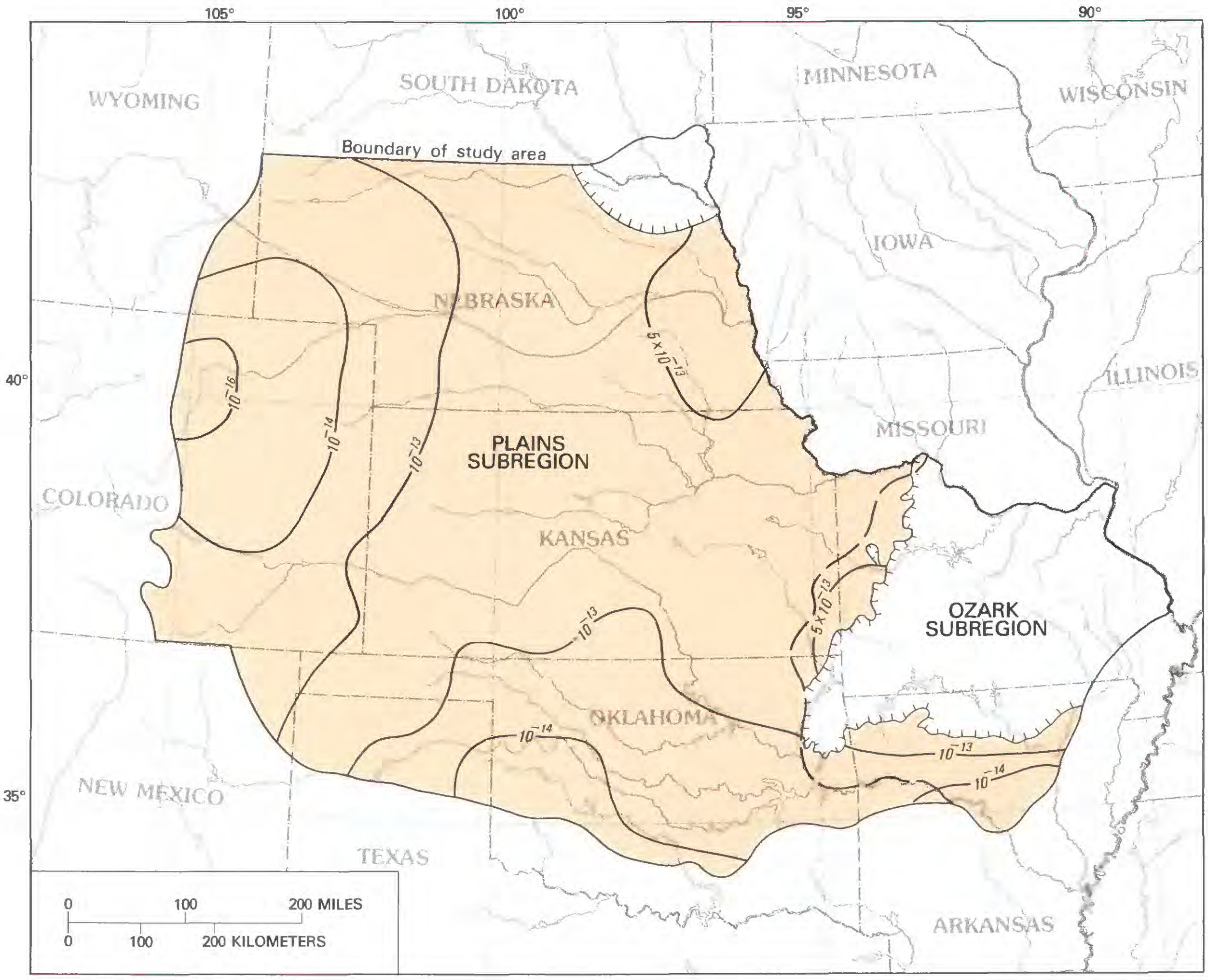


FIGURE 28.—Estimated porosity of carbonate rocks in the Western Interior Plains confining system.



Base modified from U.S. Geological Survey 1:7,500,000, National Atlas, 1970

EXPLANATION

10^{-16} **LINE OF EQUAL ESTIMATED PERMEABILITY—Value is the thickness-weighted mean horizontal intrinsic permeability. Interval, in feet squared, is variable**

EXTENT OF WESTERN INTERIOR PLAINS CONFINING SYSTEM

APPROXIMATE BOUNDARY BETWEEN PLAINS AND OZARK SUBREGIONS

FIGURE 29.—Estimated lateral intrinsic permeability of the Western Interior Plains confining system.

may approach zero. Because of the confining system's large thickness and extent, it not only functions as a confining system but also may transmit small quantities of lateral flow where there is slight intrinsic permeability (fig. 29). However, such lateral flow cannot be defined or quantified in the confining system.

The pressure of pore water in the permeable units of the confining system varies both vertically and horizontally because these units are not in a reasonably distinct hydraulic system (Jorgensen and others, 1993). Very high pressures exist in a geopressure zone near the base of the confining system in the Anadarko Basin in Oklahoma. At this location, equivalent freshwater head exceeds 6,000 ft in Morrowan sand and shale. The head in the underlying upper unit in the Western Interior Plains aquifer system near the Anadarko Basin also is about 6,000 ft because of the overpressure. Pressure also is transferred and attenuated into the underlying lower units in the Western Interior Plains aquifer system where equivalent freshwater head is more than 3,000 ft in the slightly permeable rocks.

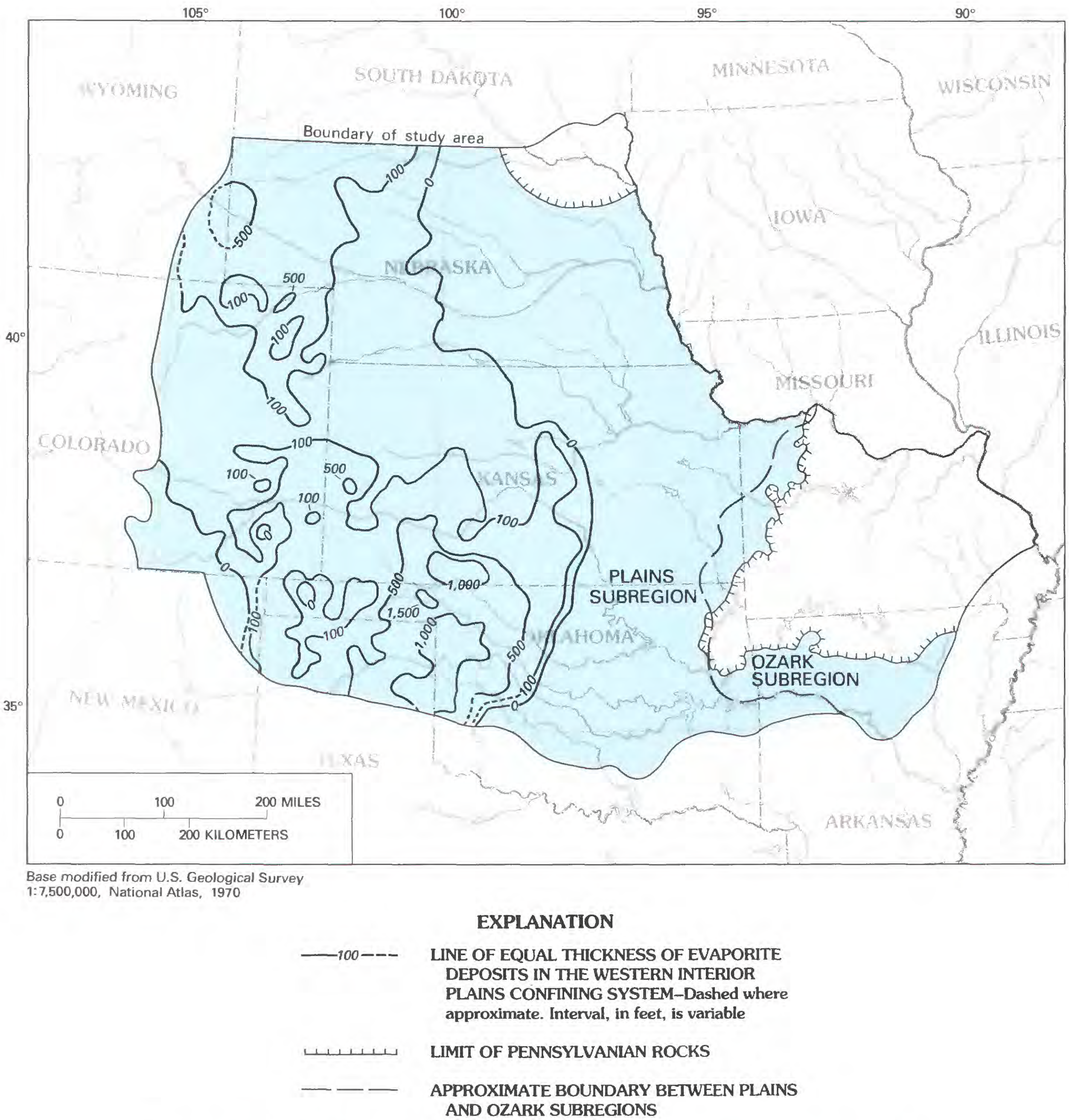
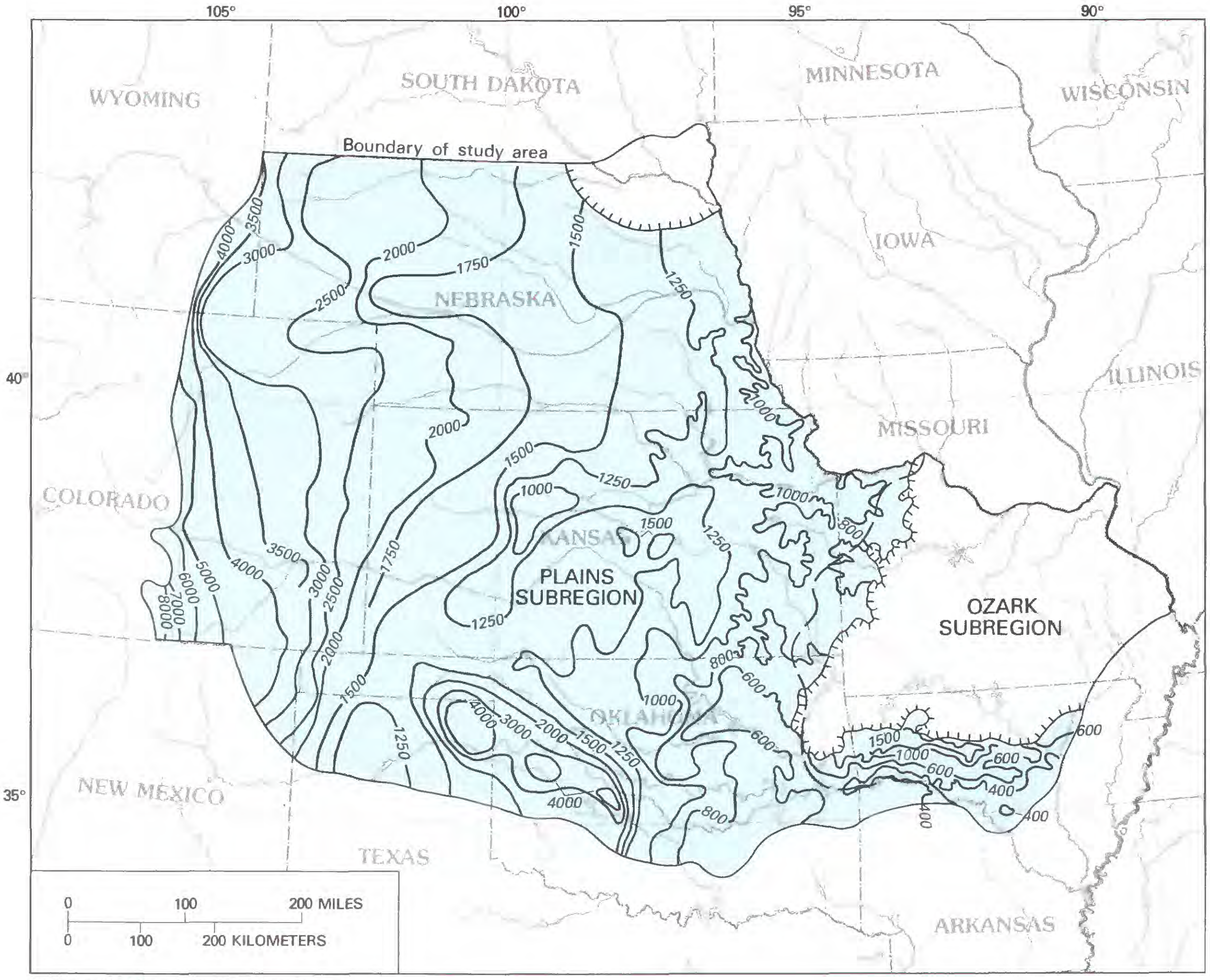


FIGURE 30.—Thickness of evaporite deposits in the Western Interior Plains confining system.

The geohydrologic characteristics of a confining system are such that, where a vertical hydraulic gradient exists between overlying and underlying units, a single hydraulic-head value in the confining unit cannot represent the entire vertical section nor can a laterally continuous surface of equivalent freshwater head be representative of the confining system. Therefore, even though pressure data from drill-stem tests are available for computing an equivalent freshwater head in parts of the confining system, a valid map of the regional equivalent freshwater head for the entire thickness of the confining system cannot be made.

Although the Western Interior Plains confining system is conceived as a confining system, very generalized equivalent-freshwater-head data referenced to the centerline of the confining system were used to map an approximation of freshwater head for the system (fig. 31). Lines of equal equivalent freshwater



Base modified from U.S. Geological Survey 1:7,500,000, National Atlas, 1970

EXPLANATION

—1000— **LINE OF EQUAL APPROXIMATE EQUIVALENT FRESHWATER HEAD OR NEAR-SURFACE FRESHWATER HEAD**—Shows altitude at which water would have stood in a tightly cased well penetrating rocks of the Western Interior Plains confining system. Dashed where approximately located. Interval, in feet, is variable. Datum is sea level

LIMIT OF WESTERN INTERIOR PLAINS CONFINING SYSTEM

— — — **APPROXIMATE BOUNDARY BETWEEN PLAINS AND OZARK SUBREGIONS**

FIGURE 31.—Approximate equivalent freshwater head in buried parts and near-surface freshwater head in outcrop area of the Western Interior Plains confining system.

head in figure 31 are interpretations and serve two purposes, which are to:

1. Provide the altitude of a constant-head boundary in the outcrop area for computer simulation; and
2. Indicate that freshwater-head values in vertically adjacent layers are consistent; that is, the hydraulic head in buried parts of the confining system should be intermediate between hydraulic head in the overlying and underlying units.

The data in the buried part of the confining system indicate consistency. However, in the outcrop area overlying the Anadarko Basin in western Oklahoma, the overpressure in the lower part of the confining system (Morrowan sandstone of earliest Pennsylvanian

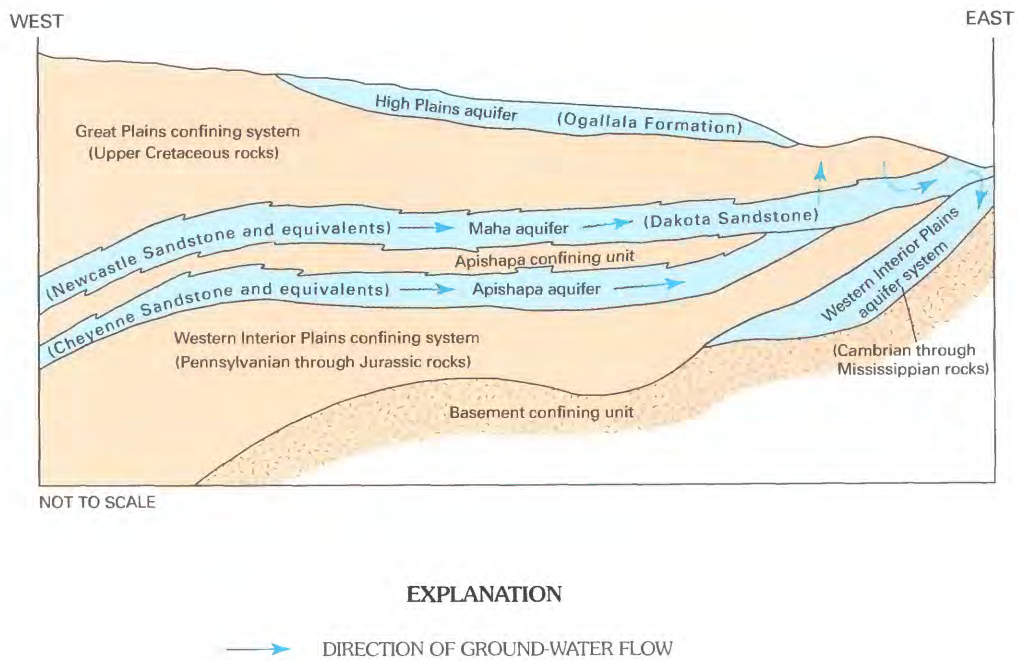


FIGURE 32.—Geohydrologic section showing the Maha and Apishapa aquifers and intervening Apishapa confining unit in the Great Plains aquifer system.

age) cannot be related to hydraulic head in the outcrop or subjacent layers.

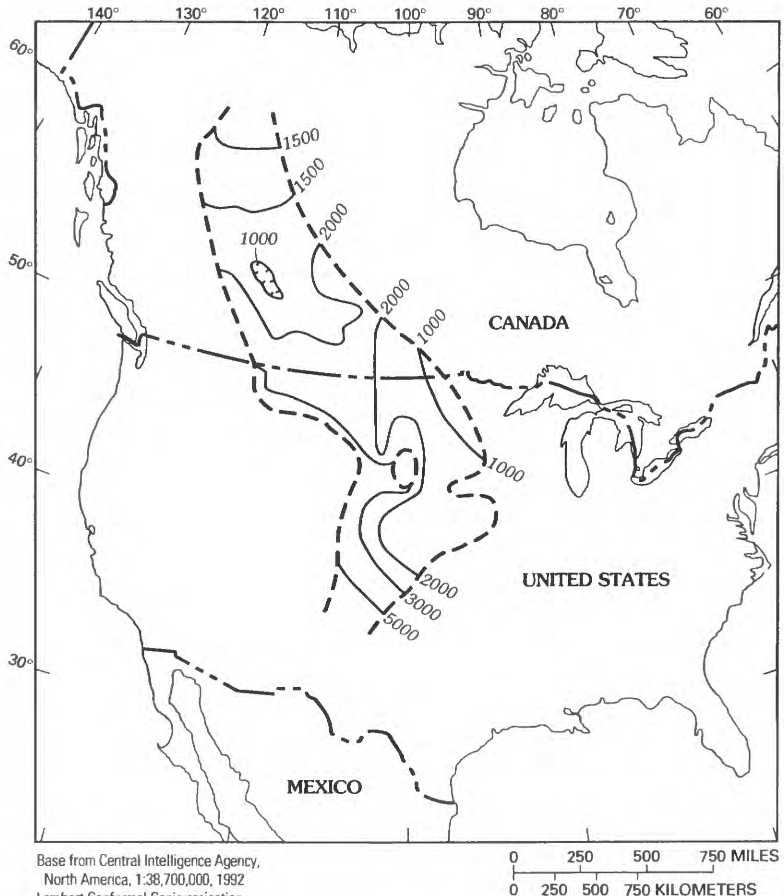
At locations where the confining system crops out, data are presented for the near-surface freshwater head (fig. 31). In general, these heads represent or are affected substantially by the near-surface water table. The altitude of the water table varies a few feet per year, but for the scale of this study can be assumed to be at a constant altitude (fig. 31). The outcrop area of the confining system is extensive and includes both recharge and discharge areas to the underlying aquifers, which depend primarily on land-surface topography and hydraulic-head relations.

GREAT PLAINS AQUIFER SYSTEM

The Great Plains aquifer system consists of two aquifers separated at most locations by a confining unit. These are the Maha and Apishapa aquifers and the intervening Apishapa confining unit (table 1, fig. 32). The aquifer system, which consists mostly of water-bearing Lower Cretaceous sandstone, is one of the most extensive in North America (Helgesen and others, 1982) and extends from near the Arctic Circle in Canada to New Mexico (fig. 33). In the United States, water in the Maha and Apishapa aquifers generally flows from west to east. This was recognized by Darton (1905); however, at that time available data were inadequate for clearly differentiating between the two aquifers at most locations. These two aquifers have been termed the "Dakota aquifer" or "Dakota aquifer system" by various investigators. Others have restricted the term "Dakota aquifer" to either the upper (Maha) aquifer or the lower (Apishapa) aquifer. In this report, the aquifers that consist mostly of Lower Cretaceous sandstone and that are part of the regional flow system are termed the Great Plains aquifer system as defined by Helgesen and others (1993).

The altitude of the top of the aquifer system ranges from more than 6,000 ft in northeastern New Mexico, a recharge area, to 1,000 ft in southeastern South Dakota, a discharge area, to less than -4,000 ft in the Denver Basin.

The Great Plains aquifer system has some large outcrop areas in southeastern Colorado and east-central Kansas. Smaller outcrop areas occur along the Front Range in the Rocky Mountains in Colorado and near the Black Hills Uplift in southwestern South Dakota. In terms of regional flow, the aquifer system receives recharge through the western outcrops. There are indications that vertical leakage through the overlying Upper Cretaceous shale provides some recharge (Neuzil, 1980). Outflow occurs in the eastern outcrops and where the Missouri River incises the unit in southeastern South Dakota, eastern Nebraska, and western Iowa. Also, there is evidence of inflow to and outflow from the overlying material near peripheral areas of the eastern extent of the unit, such as in eastern Nebraska and central Kansas. At those locations, the unit pinches out and is in direct contact with permeable material of overlying units.



Base from Central Intelligence Agency, North America, 1:38,700,000, 1992
Lambert Conformal Conic projection
Standard parallels 37° and 65°

Contours modified from Hitchon (1969), Lobmeyer (1980), and others

EXPLANATION

—3000— **POTENTIOMETRIC CONTOUR–Shows altitude of potentiometric surface of the Great Plains aquifer system. Hachures indicate depression. Contour interval is 500, 1000, and 2000 feet. Datum is sea level**

- - - - - **LIMIT OF GREAT PLAINS AQUIFER SYSTEM**

FIGURE 33.—Great Plains aquifer system in North America (modified from Helgesen and others, 1982).

The Great Plains aquifer system occurs beyond the boundary of the CMRASA study area along the northern State line of Nebraska. The State line is a common border with the Northern Great Plains RASA. The potentiometric surface of the Maha aquifer in the Great Plains aquifer system suggests that flow lines are approximately parallel to the State line, indicating little flow across that boundary, a condition that is consistent with the results of the Northern Great Plains RASA study (Downey, 1986, figs. 40, 45, pl. 3).

The Apishapa aquifer is the lower of the two major aquifers in the Great Plains aquifer system and is composed of permeable, loosely cemented, medium- to very fine-grained sandstone of the Cheyenne Sandstone and equivalents, such as the Fall River and Lakota Sandstone (fig. 32, table 1). The name Apishapa refers to the water-yielding sandstone layers that occur in the vicinity of the Apishapa River and other adjacent areas in southeastern Colorado, northeastern New Mexico, and southwestern Kansas. The maximum thickness of the Apishapa aquifer exceeds 400 ft; however, the typical thickness is between 100 and 200 ft.

The Apishapa confining unit is composed of very slightly permeable shale layers, which at most locations consist of the Kiowa Shale and the equivalent Skull Creek Shale. The confining unit, which separates the Apishapa aquifer from the overlying Maha aquifer, is not as extensive as the Maha aquifer. The Apishapa confining unit is thin, with thicknesses ranging from 10 to 270 ft. The altitude of the top is, in general, about 200 ft less than the altitude of the top of the Great Plains aquifer system.

The Maha aquifer is the upper aquifer unit in the Great Plains aquifer system and is more extensive than the underlying Apishapa aquifer (fig. 32). The aquifer at most locations consists of permeable, loosely cemented, medium- to fine-grained sandstone of the Dakota Sandstone, "D" and "J" sandstones of informal usage, and equivalent Newcastle Sandstone. The altitude of the top of the Maha aquifer and the altitude of the top of the Great Plains aquifer system are the same except where the Maha is absent and the Apishapa aquifer is present in a few locations in southeastern Colorado and northeastern New Mexico. The aquifer thickens from less than 100 ft in eastern Colorado and along its eastern limit to about 650 ft in northeastern Nebraska. Maximum thicknesses exceed 900 ft in small areas of north-central and central Nebraska.

Nearly all of the available hydrologic information relating to permeability, water chemistry, and potentiometric head of the Great Plains aquifer system are for the Maha aquifer. The limited information available for the Apishapa aquifer indicates that potentiometric head does not differ greatly between the two aquifers. The regional predevelopment potentiometric surface for the Maha aquifer is shown in figure 34. The head in the Apishapa aquifer is believed to be similar and closely related. However, in some areas there may be a considerable difference; for example, in South Dakota just north of the study area, a substantial potentiometric-head difference occurs between the aquifers. Also in the recharge areas in southeastern Colorado, the Apishapa aquifer has a lower head than the Maha.

Estimated porosity and lateral intrinsic permeability for both aquifers (figs. 35, 36) are considered to be primary because the aquifer material is mostly a loosely cemented sandstone. Both porosity and permeability decrease with depth.

A generalized distribution of dissolved-solids concentrations in water from the Maha aquifer (Helgesen and Leonard, 1989) is shown in figure 37. Limited water-quality information within the study area indicates that water from the Apishapa in some locations may have a larger dissolved-solids concentrations than water from the Maha aquifer.

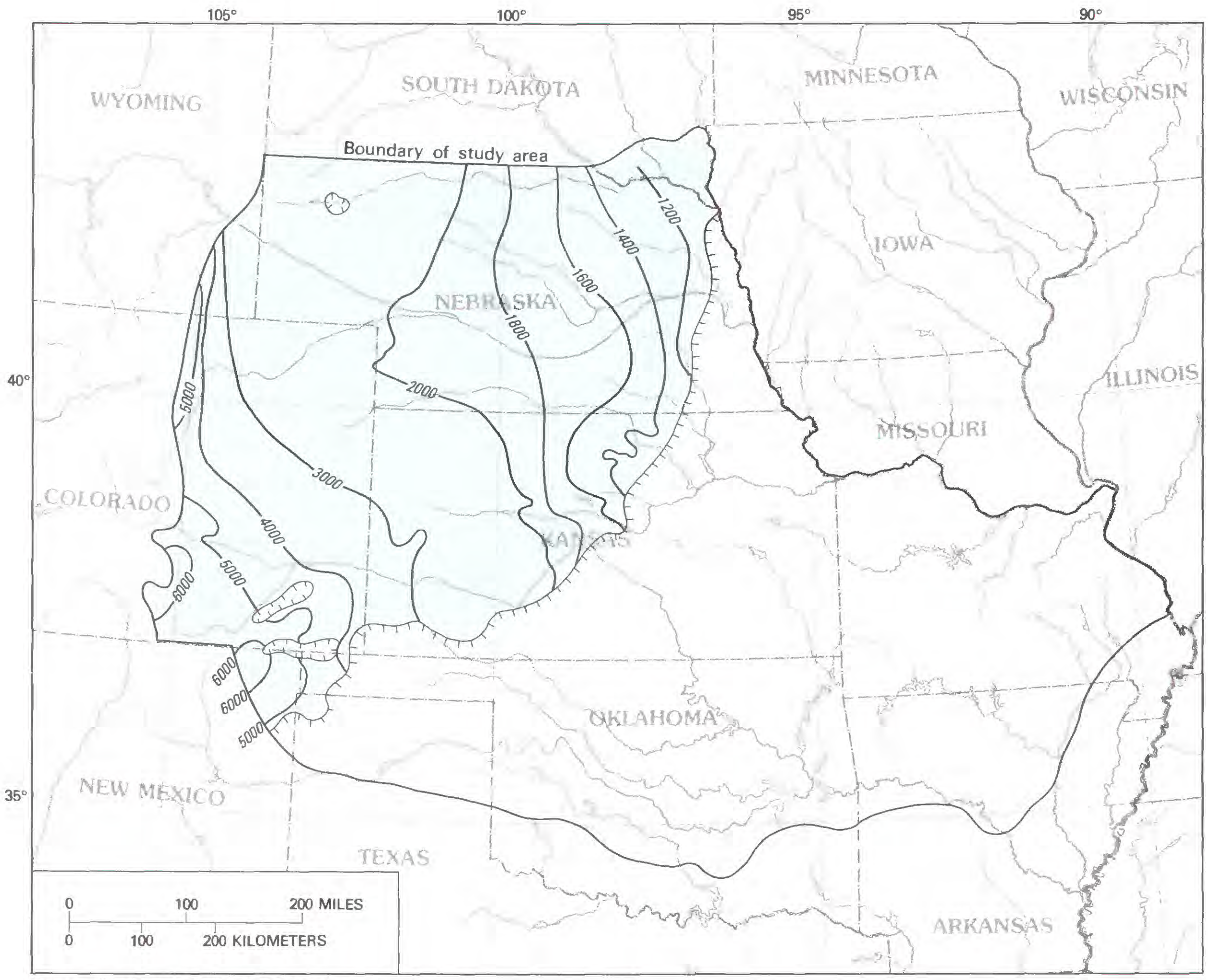
OVERLYING UNITS

The uppermost geohydrologic units for the Plains subregion of the CMRASA consist principally of the Great Plains confining system and the High Plains aquifer, with smaller areas of glacial drift and loess deposits (fig. 38). The geohydrology and digital simulation of the High Plains aquifer are treated in detail by Gutentag and others (1984) and Luckey and others (1984).

Below the High Plains aquifer, the Great Plains confining system, which has a combined thickness of as much as 8,000 ft (fig. 39), is composed of shale including the Upper Cretaceous Graneros, Pierre, and Carlile Shales, as well as very slightly to slightly permeable Tertiary clay and silt. The confining system also includes two extensive but minor aquifers in the Greenhorn Limestone and Niobrara Formation. The confining system at most locations effectively restricts flow between the Great Plains aquifer system and the unconfined High Plains aquifer or other unconfined units. Vertical movement of water from the High Plains aquifer through the Upper Cretaceous shale is dependent on the hydraulic head in the High Plains aquifer, the vertical permeability and thickness of the shale, and the hydraulic head in the Great Plains aquifer system. Leonard and others (1983, fig. 8) show that in the coincident areas of the High Plains aquifer and the Great Plains aquifer system, hydraulic head in the High Plains aquifer is generally higher than in the Great Plains aquifer system (fig. 40). Peripheral to the Upper Cretaceous shale in the Great Plains confining system are areas where permeable materials of the High Plains aquifer are in direct contact with permeable materials of the Great Plains aquifer system. Glacial drift and loess deposits in eastern Kansas and Nebraska in part lie directly on rocks older than Early Cretaceous. The glacial drift is largely heterogeneous material of variable hydraulic character. Loess deposits also are present as a surficial unit and consist of uniform silt-sized permeable material. The glacial drift and loess deposits probably act as leaky confining materials relative to the underlying units.

OZARK SUBREGION

The Ozark subregion generally corresponds to the Ozark Plateaus physiographic province (fig. 41). Ground-water flow, in general, is outward from a topographic high along an axis from the St. Francois Mountains in southeastern Missouri to northwestern Arkansas. The regional flow system is in Paleozoic sedimentary rocks, which are underlain by



Base modified from U.S. Geological Survey 1:7,500,000, National Atlas, 1970

EXPLANATION

—*3000*— **POTENTIOMETRIC CONTOUR—Shows altitude at which water would have stood in tightly cased wells. Contour interval, in feet, is variable. Datum is sea level**

LIMIT OF GREAT PLAINS AQUIFER SYSTEM

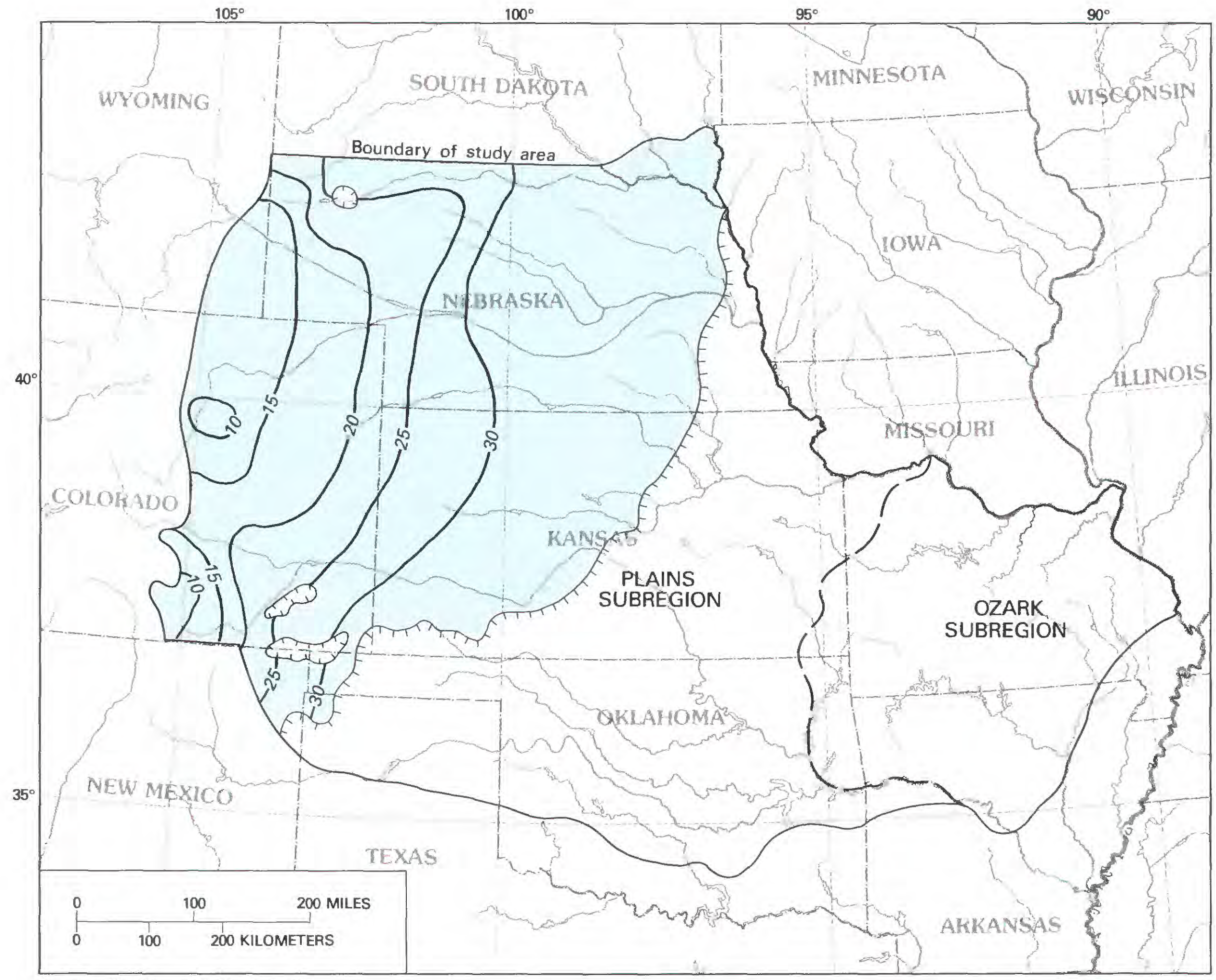
FIGURE 34.—Predevelopment potentiometric surface of the Maha aquifer in the Great Plains aquifer system.

the basement confining unit that is composed mostly of igneous and metamorphic rocks.

The geohydrologic units and model analysis of ground-water flow in the Ozark subregion are described by Imes and Emmett (1994). The following description of the geohydrologic framework of the Ozark Plateaus aquifer system in the Ozark subregion is based on the aquifer and confining-unit designations of that report.

OZARK PLATEAUS AQUIFER SYSTEM

Most ground water in the Ozark subregion is in the Ozark Plateaus aquifer system. The aquifer system contains freshwater and is laterally adjacent to the Western Interior Plains aquifer system of the Plains subregion. The geologic units of the Ozark Plateaus aquifer system are approximately equivalent to those of the Western Interior Plains aquifer system



Base modified from U.S. Geological Survey 1:7,500,000, National Atlas, 1970

EXPLANATION

—20— LINE OF EQUAL POROSITY–Interval, in percent, is variable

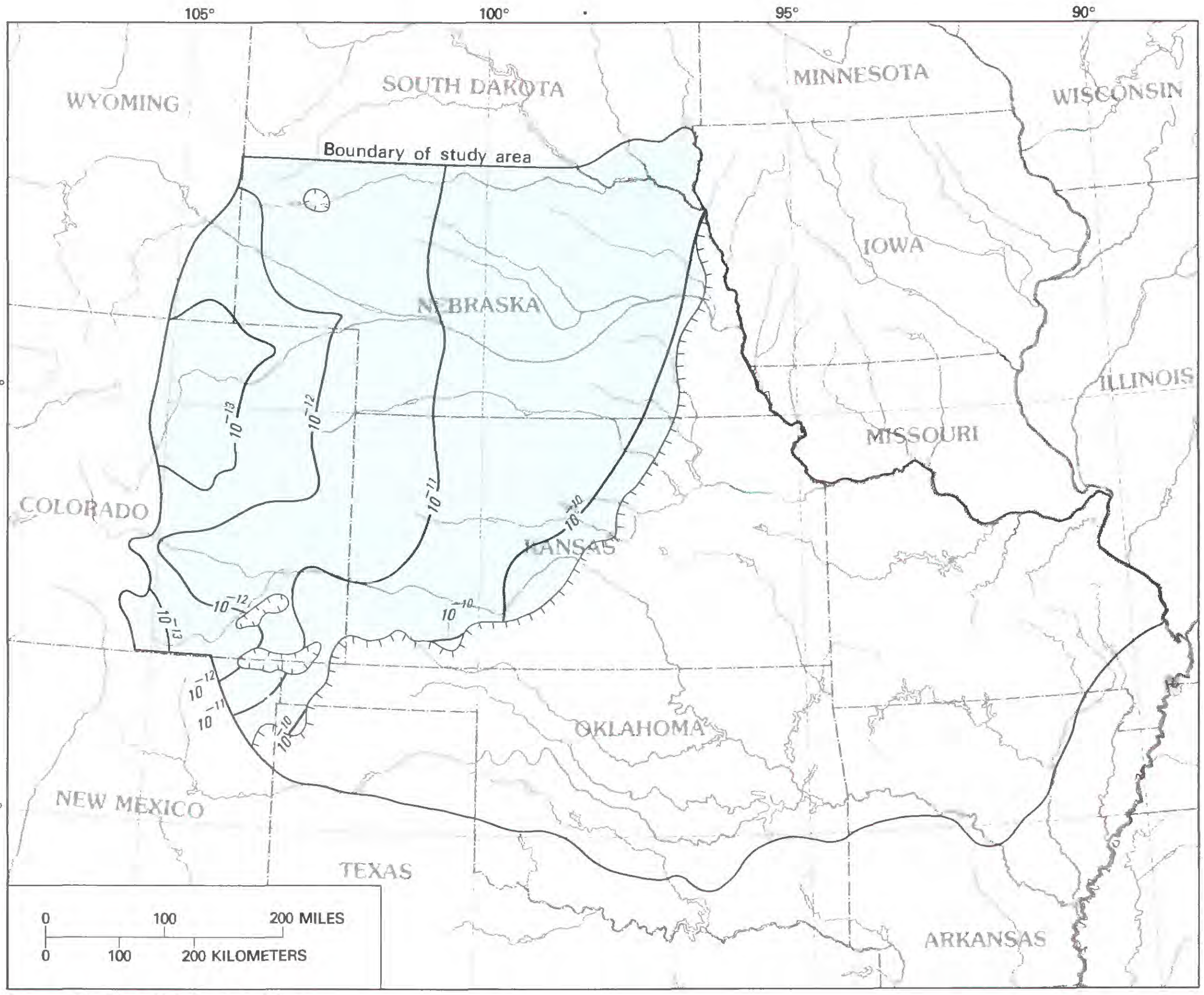
LIMIT OF GREAT PLAINS AQUIFER SYSTEM

— — — APPROXIMATE BOUNDARY BETWEEN PLAINS AND OZARK SUBREGIONS

FIGURE 35.—Estimated porosity of the Apishapa and Maha aquifers.

(tables 1, 2; fig. 19). The two aquifer systems are separated by a no-flow hydrologic boundary, not rock boundaries. In general, the rocks of the St. Francois aquifer, the St. Francois confining unit, and the Ozark aquifer are age equivalents to the Paleozoic rocks of the lower units in the Western Interior Plains aquifer system. The rocks of the Springfield Plateau aquifer are equivalent to the Mississippian rocks of the upper unit in the Western Interior Plains aquifer system. The intervening Ozark confining unit is principally the Chattanooga Shale.

The Ozark Plateaus aquifer system is overlain along the western and southern borders by the Western Interior Plains confining system. The Ozark Plateaus aquifer system consists of rocks that range in age from Late Cambrian through Late Mississippian (table 2). Tops and bottoms of the geohydrologic units do not always conform to geologic-time divisions or boundaries of rock-stratigraphic units but are chosen to delineate groups of rocks having similar hydrologic properties. The geohydrologic units that comprise the Ozark Plateaus aquifer system crop out



Base modified from U.S. Geological Survey 1:7,500,000, National Atlas, 1970

EXPLANATION

10^{-13} **LINE OF EQUAL ESTIMATED PERMEABILITY**—Value is the thickness-weighted mean horizontal intrinsic permeability. Interval, in feet squared, is variable

LIMIT OF GREAT PLAINS AQUIFER SYSTEM

FIGURE 36.—Estimated lateral intrinsic permeability of the Apishapa and Maha aquifers.

in an approximately circular pattern around the St. Francois Mountains.

ST. FRANCOIS AQUIFER

The St. Francois aquifer, the lowermost aquifer in the Ozark Plateaus aquifer system, is composed of Upper Cambrian rocks, primarily sandstone in the western one-third of the Ozark subregion and sandstone overlain by coarse-grained dolostone in the eastern two-thirds of the subregion.

The St. Francois aquifer extends to the boundaries of the Ozark Plateaus aquifer system (fig. 20). Beyond the relatively small outcrop area near the St. Francois Mountains (fig. 41), the aquifer dips into the subsurface and is buried beneath the other units of the Ozark Plateaus aquifer system. The aquifer dips most steeply to the east toward the Illinois Basin (fig. 7) and south toward the Mississippi Alluvial Plain (fig. 41) where it is buried to a depth of more than 5,000 ft. The aquifer occurs in most of the Ozark subregion, although it rarely is penetrated by wells in the southeast part of the area (Arkansas). The St. Francois aquifer is thickest (greater than 1,000 ft) in eastern Missouri beneath the Mississippi Alluvial Plain

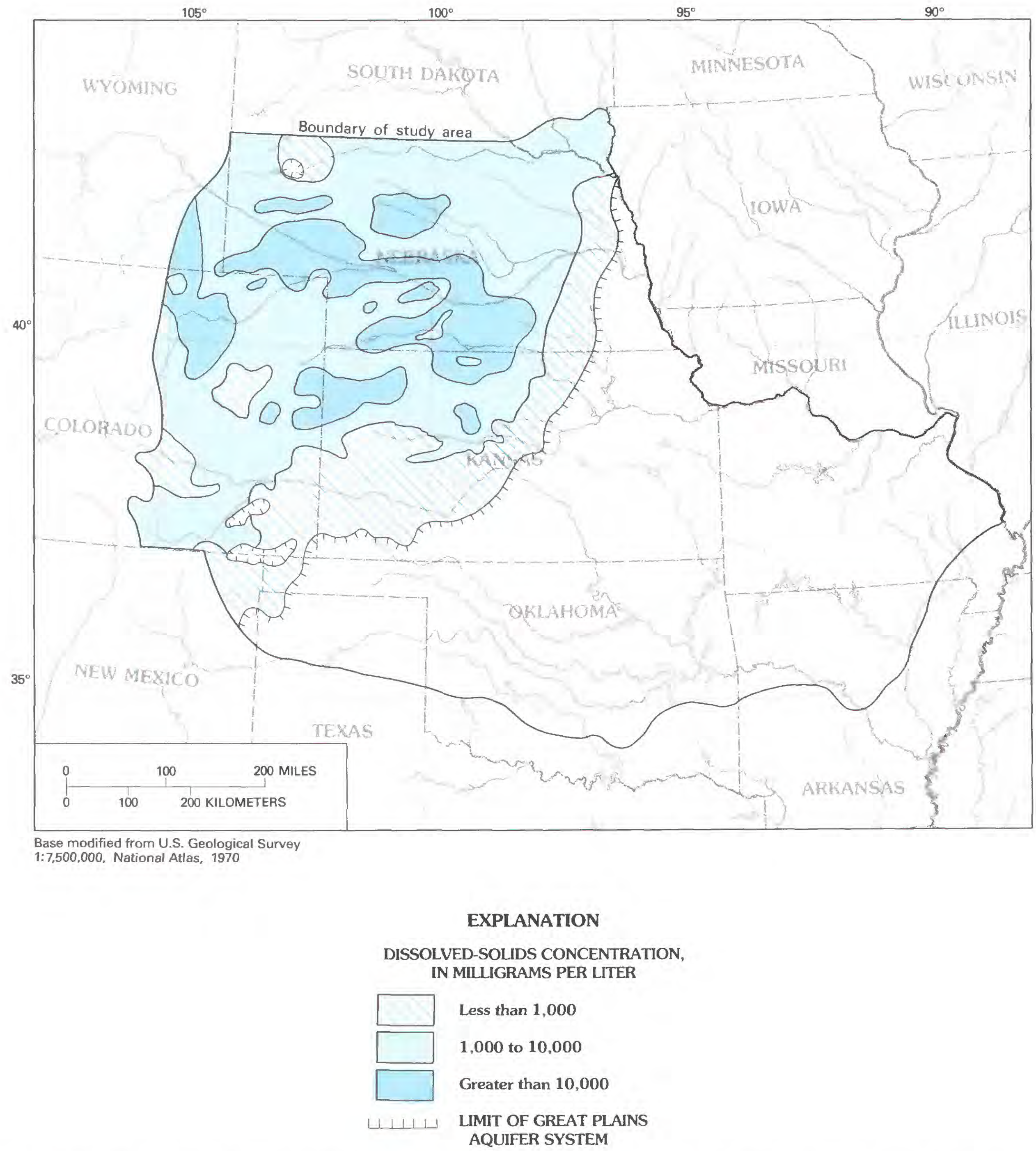


FIGURE 37.—Dissolved-solids concentrations in water from the Maha aquifer in the Great Plains aquifer system (modified from Helgesesn and Leonard, 1980).

but generally is between 100 and 300 ft thick in northern Arkansas.

Estimates of hydraulic conductivity based on specific-capacity data in and near the outcrop area near the St. Francois Mountains range from $1x10^{-4}$ ft/s to less than $1x10^{-6}$ ft/s and tend to decrease away from the outcrop.

Few data are available on water chemistry of the St. Francois aquifer except in the outcrop area. Chemical analyses of water samples collected from the

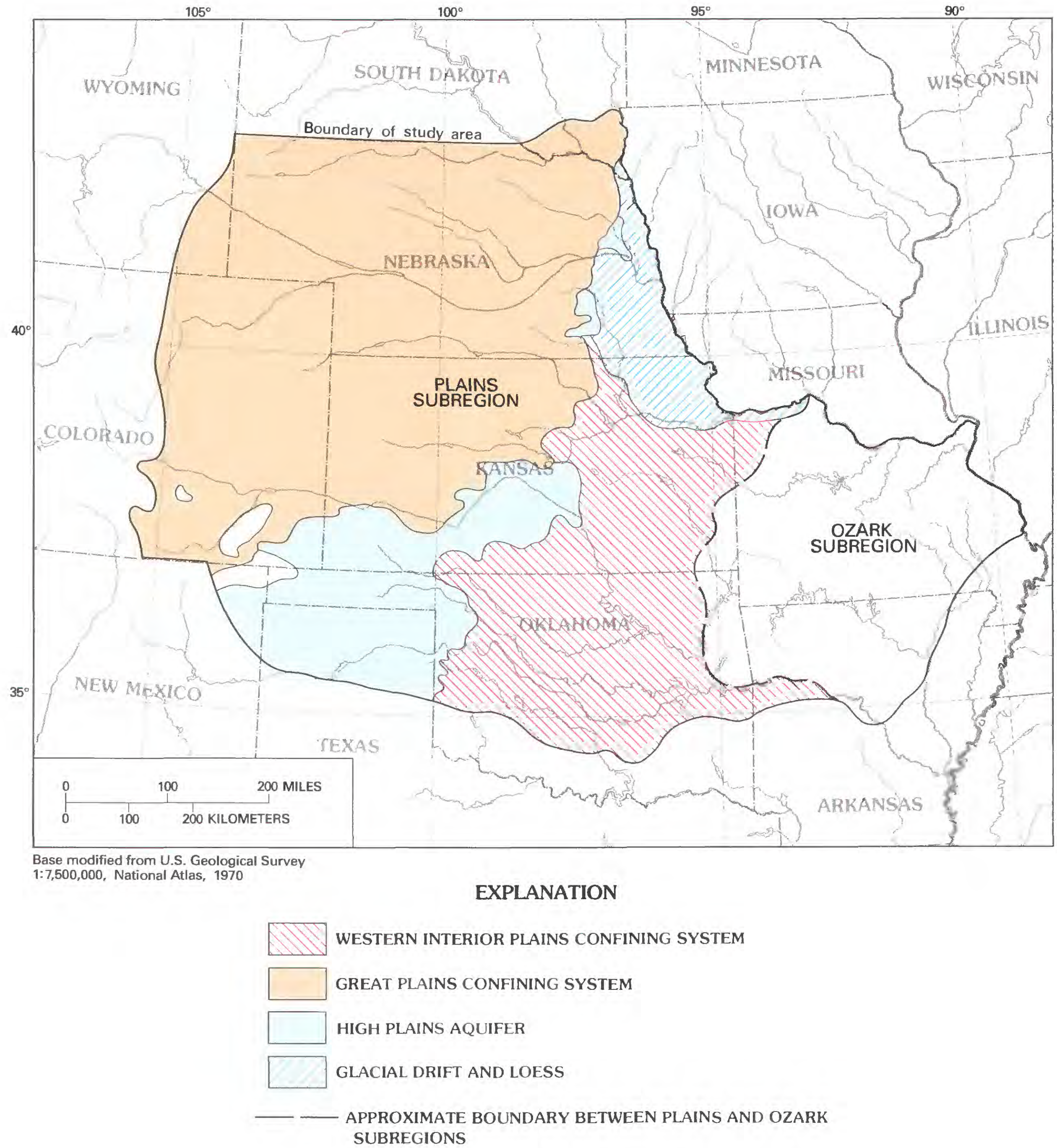
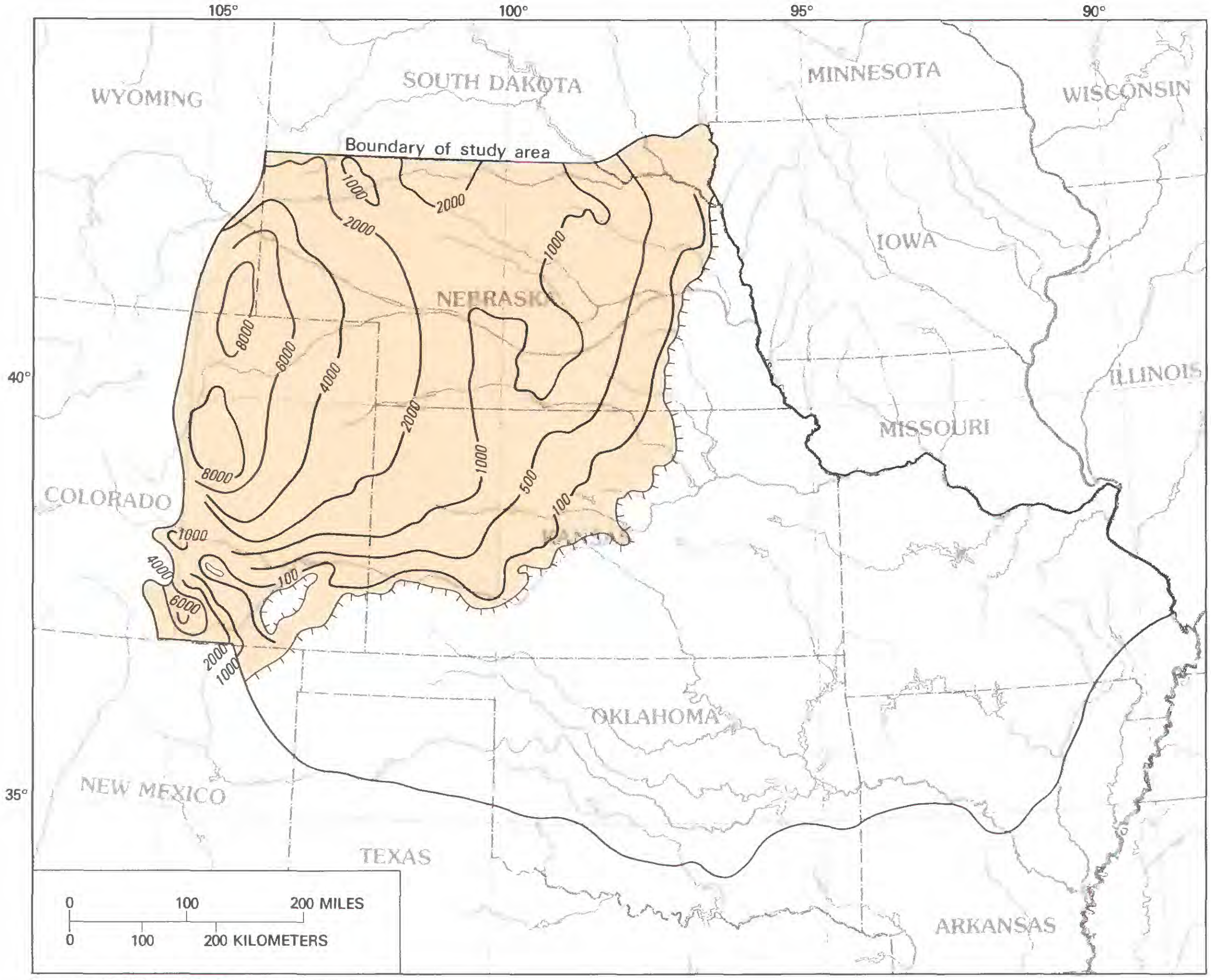


FIGURE 38.—Geohydrologic units overlying regional aquifer systems in Plains subregion.

aquifer in the outcrop area indicate dissolved-solids concentrations from less than 100 to more than 400 mg/L. Dissolved-solids concentrations in water samples from wells that penetrate the deeply buried St. Francois aquifer are not large, but actual concentrations in water from the St. Francois aquifer are not known because the wells are open to overlying geohydrologic units.

Where the St. Francois aquifer crops out, ground-water levels are affected primarily by topography. The aquifer is recharged from infiltration of precipitation through the soil zone. The aquifer is recharged also by flow from the fractured Precambrian rocks near the outcrop area. Hydraulic-head data in the immediate vicinity of the St. Francois Mountains indicate that water from the St. Francois



EXPLANATION

—2000— **LINE OF EQUAL THICKNESS OF GREAT PLAINS CONFINING SYSTEM– Interval, in feet, is variable**

LIMIT OF GREAT PLAINS CONFINING SYSTEM

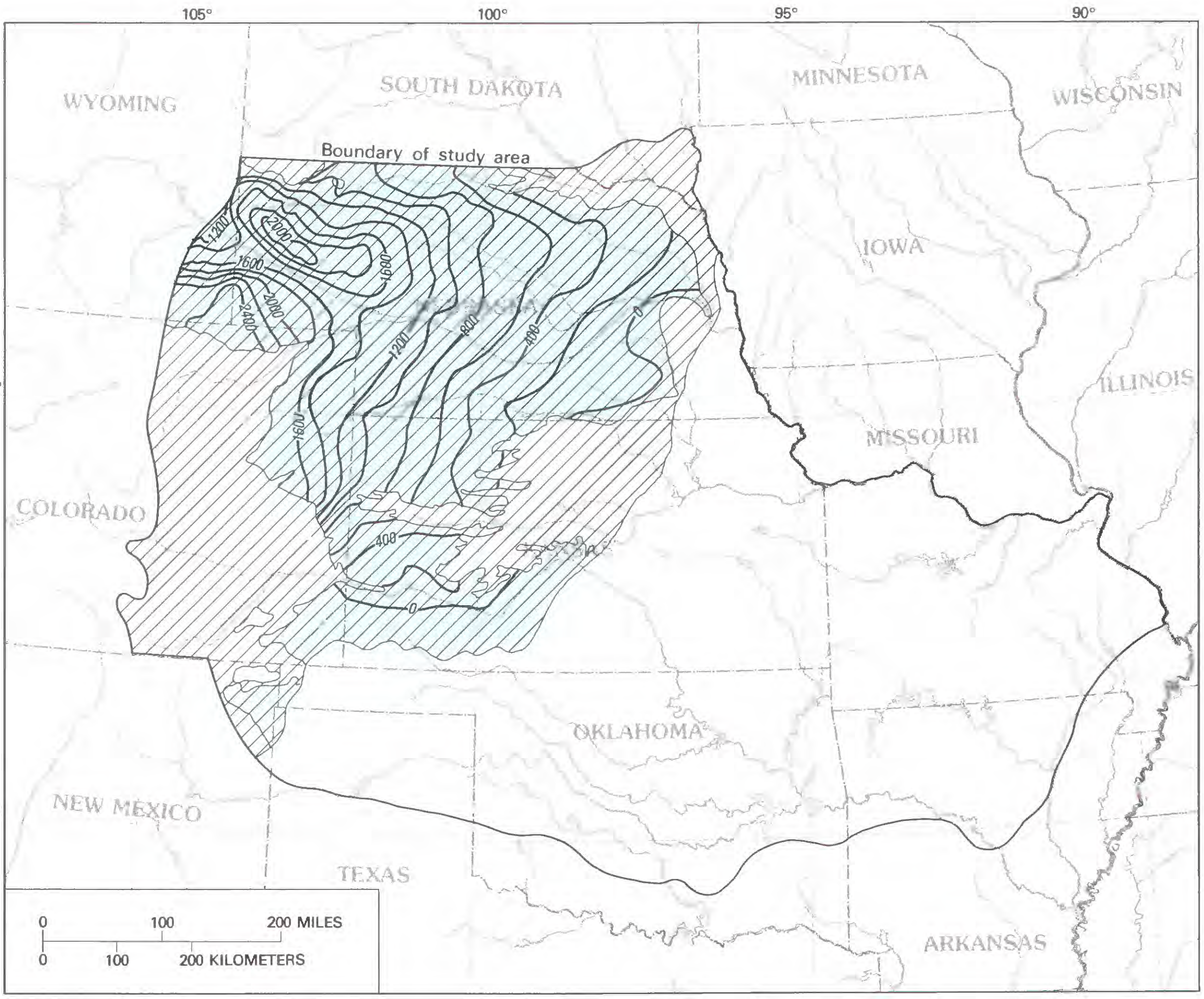
FIGURE 39.—Thickness of the Great Plains confining system.

aquifer discharges upward to overlying aquifers through confining units downdip from the outcrop area and discharges to the St. Francis and Black Rivers to the south and the Big River to the north (fig. 1). In that area, leakance of the confining unit is not sufficiently small to hydrologically isolate the St. Francois aquifer from topographic effects.

ST. FRANCOIS CONFINING UNIT

The St. Francois confining unit restricts flow between the St. Francois aquifer and the overlying Ozark aquifer. The confining unit is composed of shale, dolostone, and limestone. No quantitative measurements of the hydraulic conductivity of the St. Francois confining unit are known (Imes and Emmett, 1994). Imes and Emmett (1994) cite evidence that the hydraulic conductivity near fracture zones in the unit is too large to hydrologically isolate the underlying St. Francois aquifer in those localities. However, in areas where the rocks are not fractured, the unit forms an effective confining unit.

The total outcrop area of the confining unit is about 400 mi^2. The confining unit dips steeply and radially from the St. Francois Mountains (as much as 150 ft/mi), except to the west where the unit dips



Base modified from U.S. Geological Survey
1:7,500,000, National Atlas, 1970

EXPLANATION

GREAT PLAINS AQUIFER SYSTEM

HIGH PLAINS AQUIFER

—1200— LINE OF EQUAL DIFFERENCE IN HYDRAULIC HEAD–
Shows altitude of the potentiometric surface of High Plains aquifer minus (-) the altitude of the potentiometric surface of the Great Plains aquifer system. Interval 200 feet

FIGURE 40.—Differences in hydraulic head between the High Plains aquifer and the Great Plains aquifer system (modified from Leonard and others, 1983).

more gently into the subsurface. Smaller isolated areas within the Ozark subregion where the unit does not occur in the subsurface usually are at or near areas where the St. Francois aquifer is also absent.

The maximum thickness of the St. Francois confining unit, as determined from well logs, is about 900 ft in southeastern Missouri beneath the northwestern edge of the Mississippi Alluvial Plain. The confining unit generally is thicker in the eastern and northeastern parts of the Ozark subregion. The confining unit is thinner in western Missouri, ranging from about 100 to 300 ft, and pinches out a few miles west of the Ozark subregion in eastern Kansas and northeastern Oklahoma.

Shale in the St. Francois confining unit, which in general is nearly impermeable, may occur as distinct layers but usually is distributed throughout the limestone and dolostone beds. The percentage-of-shale content of the St. Francois confining unit ranges from 0 to 60 percent and generally is less than 30 percent

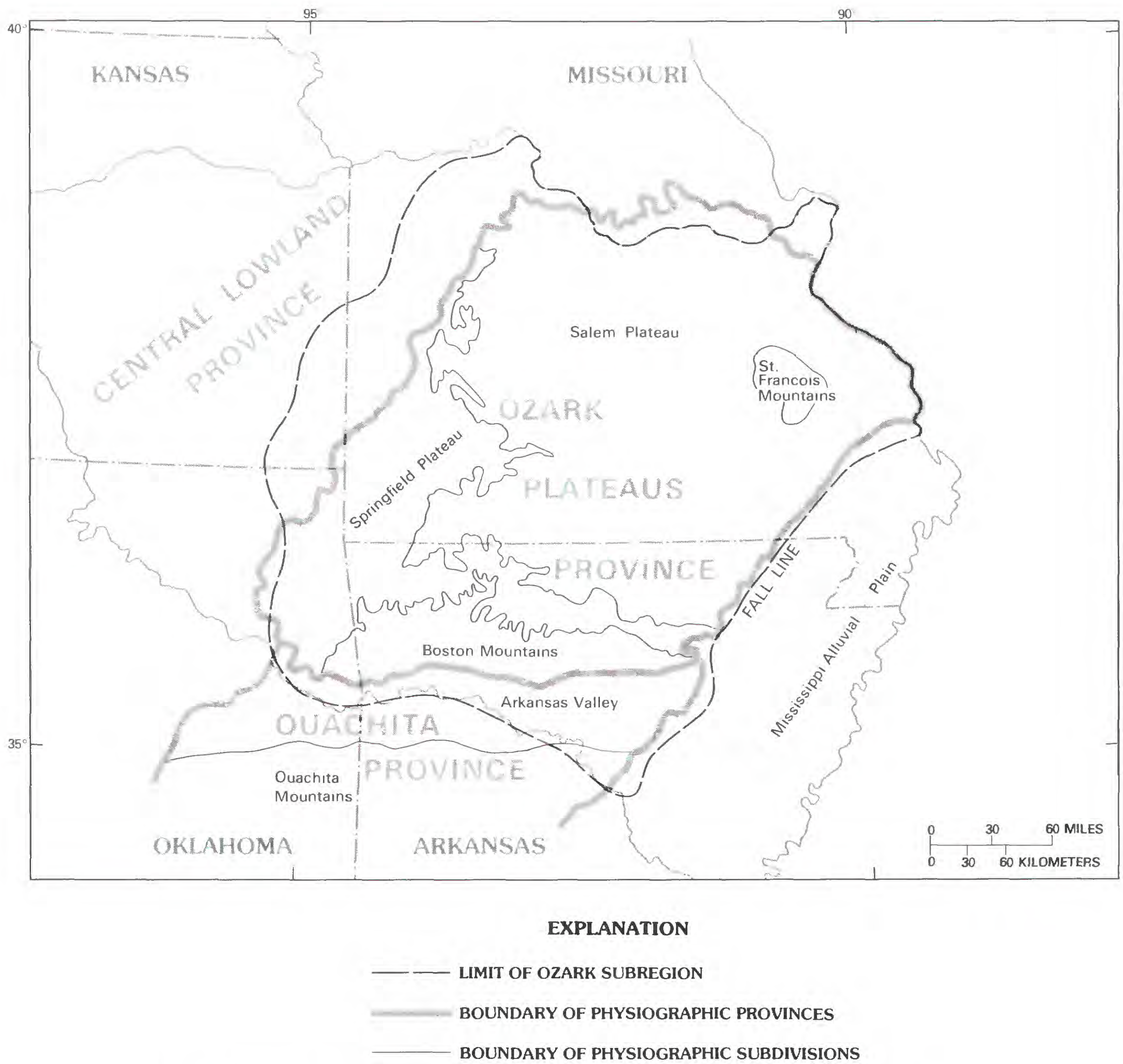


FIGURE 41.—Physiographic features of Ozark subregion (modified from Fenneman, 1946).

(fig. 42). The confining ability of the unit is dependent not only on the shale content but also on the degree of cementation of the carbonate rock and the abundance of fractures and solution openings.

The relatively fine-grained nature of the dolostone and the lack of significant secondary porosity are the probable reasons for the small permeability of the confining unit even in regions devoid of shale. However, the confining unit is more likely to be leaky in the dolomitic areas than in areas with a large shale content. There is evidence from well cores that the upper part of the confining unit is more permeable along a northwest-trending reef zone passing through Wright and Douglas Counties, Missouri. It is possible that the St. Francois and Ozark aquifers are more hydraulically connected along the reef zone. Fracture zones are evident and likely result in increased leakage locally, especially in the vicinity of faults.

OZARK AQUIFER

The Ozark aquifer, the most areally extensive and most intensively pumped aquifer in the Ozark subregion, is composed of a sequence of formations ranging in age from Late Cambrian to Middle Devonian (table 2) that vary considerably in water-yielding capability but collectively act as a regional aquifer. Where buried west and south of the extensive outcrop area, the base of the Chattanooga Shale

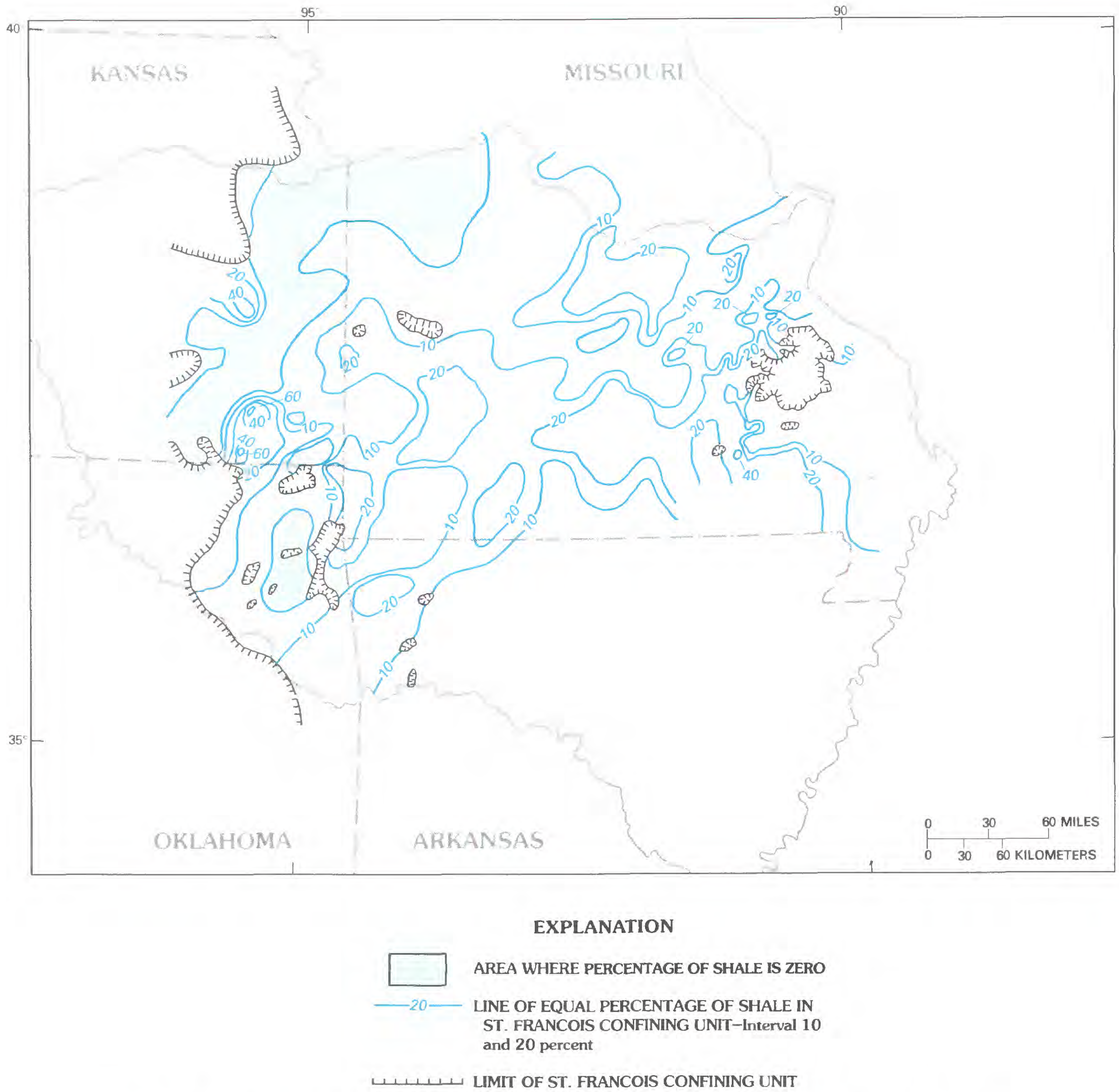


FIGURE 42.—Shale content of the St. Francois confining unit (modified from Jorgensen and others, 1993).

stratigraphically represents the upper boundary of the Ozark aquifer. The aquifer is exposed over a large part of the Ozark subregion and is dissected by streams where exposed. The rocks are highly fractured and moderately faulted.

Where the aquifer dips into the subsurface, it generally is overlain by Upper Devonian rocks; however, in the northern part of the Ozark subregion (Gasconade County, Missouri, and vicinity), the aquifer is in direct contact with as much as 200 ft of Pennsylvanian-age shale. To the southeast in the Mississippi Alluvial Plain, the aquifer is overlain by thick deposits of Cretaceous, Tertiary, and Quaternary sediment.

The Ozark aquifer is the thickest geohydrologic unit in the Ozark subregion. In southeastern Kansas, southwestern Missouri, and northeastern Oklahoma, aquifer thicknesses generally range from about 800 to 1,500 ft. Thicknesses are less uniform over short distances in Oklahoma where Precambrian topographic relief affected the thickness of rocks that typically are part of the Ozark aquifer.

The hydraulic conductivity of the Ozark aquifer has been estimated from specific-capacity data and inferred from density of fractures and development of solution openings. In general, the more the rocks have been uplifted, the more they have been fractured, the more dissolution has occurred along fractures, and correspondingly, the greater the hydraulic conductivity. Areas of greatest hydraulic conductivity are concentrated along an east-west line passing through the St. Francois Mountains where the hydraulic conductivity may be as much as $1x10^{-3}$ ft/s. In areas where dissolution has occurred and springs are numerous, the hydraulic conductivity is extremely large and locally exceeds $1x10^{-1}$ ft/s (Jorgensen and others, 1993). The hydraulic conductivity decreases to the south to as small as $1x10^{-8}$ ft/s near the southern boundary of the study area.

The Ozark aquifer is a complex sequence of formations (table 2) of different lithologies and a wide range of porosity and permeability values. The lithologies include dolostone, limestone, sandstone, chert, and shale. Dolostone is the predominant rock type. Dissolution of carbonate rock along fractures and bedding planes, including karst development, has enhanced the permeability of the aquifer. Sandstone is present as massive bodies in some formations. The sandstone is usually clean, well sorted, and permeable; however in Arkansas, thick clean sandstone is well cemented and, where unfractured, is much less permeable.

The limestone and chert of Silurian and Devonian age are the uppermost rocks in the Ozark aquifer in the north-central and southern part of the Ozark subregion. The rocks are less permeable than the older rocks in the aquifer and generally are not a source of water supply. The smaller permeability reflects a lack of secondary porosity.

Water in most of the Ozark aquifer is unconfined, and its flow is controlled by the altitude of the water table, which generally is related closely to the land-surface altitude. Throughout the outcrop area, ground water moves toward the major rivers and their tributaries (fig. 43). Hundreds of springs have developed. Flow to most of the springs originates in the upland intervalley areas. Water moves through solution channels along fractures downward to the water table near the valleys and discharges along streams. The Mississippi River Valley is a major discharge area for the aquifer. The Missouri River Valley along the northern edge and east of Saline County, Missouri, is also a major discharge area. To the west of Saline County, the aquifer is confined, and the Missouri River at that location is not a major discharge area. In west-central Missouri, there is evidence that water from the Ozark aquifer discharges into the Marais des Cygnes and Osage Rivers (fig. 43) (Imes and Emmett, 1994). In eastern Kansas and southward to northeastern Oklahoma, flow in the Ozark aquifer is outward to the boundary of the Ozark subregion. Westerly flowing ground water discharges through overlying deposits into the Neosho River Valley (fig. 43). The western boundary of the aquifer is also approximately coincident with the eastward extent of Pennsylvanian shale; this shale affects, in part, the vertical leakage upward out of the Ozark Plateaus aquifer system. South of the Boston Mountains, there is limited flow southward toward the Arkansas River Valley (fig. 43). The quantity of flow southward is limited because of the slight permeability of the rocks in that area. The dissolved-solids concentrations (fig. 43) indicate the extent of freshwater around parts of the periphery of the Ozark aquifer. Concentrations increase toward the west and are approximately coincident with the eastern extent of the Pennsylvanian shale and the approximate boundary between the Plains and Ozark subregions (fig. 24).

OZARK CONFINING UNIT

The Ozark confining unit is the uppermost confining unit of the Ozark Plateaus aquifer system and consists mostly of Upper Devonian Chattanooga Shale and Lower Mississippian Chouteau Limestone. The semipermeable rocks of the Ozark confining unit impede the movement of water between the Ozark aquifer and the overlying Springfield Plateau aquifer. However, local lithologic and structural variations can increase the leakance to the degree that confinement is minimal and does not restrict flow between the Ozark and the Springfield Plateau aquifers to any significant degree. The leakance is the result of the large permeability of the rocks or thinning of the confining unit. The confining unit is not present everywhere in the Ozark subregion; it is missing from the Salem Plateau, an area that includes approximately the eastern two-thirds of the Ozark subregion (fig. 41). In extreme eastern Missouri, formations of Ordovician, Silurian, and Devonian age are included as part of the confining unit. It is probable that some of these formations in eastern Missouri, several of which are thick limestone formations, constitute the western edge of another geohydrologic unit in another flow system outside of the Ozark subregion.

The confining unit does not occur in the subsurface in small areas of southeastern Kansas, southwestern Missouri, northeastern Oklahoma, and northern Arkansas, probably due to nondeposition. The confining unit does not occur in the subsurface over a large area in the northern part of the area (Gasconade County, Missouri, and vicinity) and along the Mississippi Alluvial Plain where the rocks that typically form the unit have been removed by erosion.

Thickness of the Ozark confining unit ranges from 0 to more than 1,600 ft in extreme eastern Missouri.

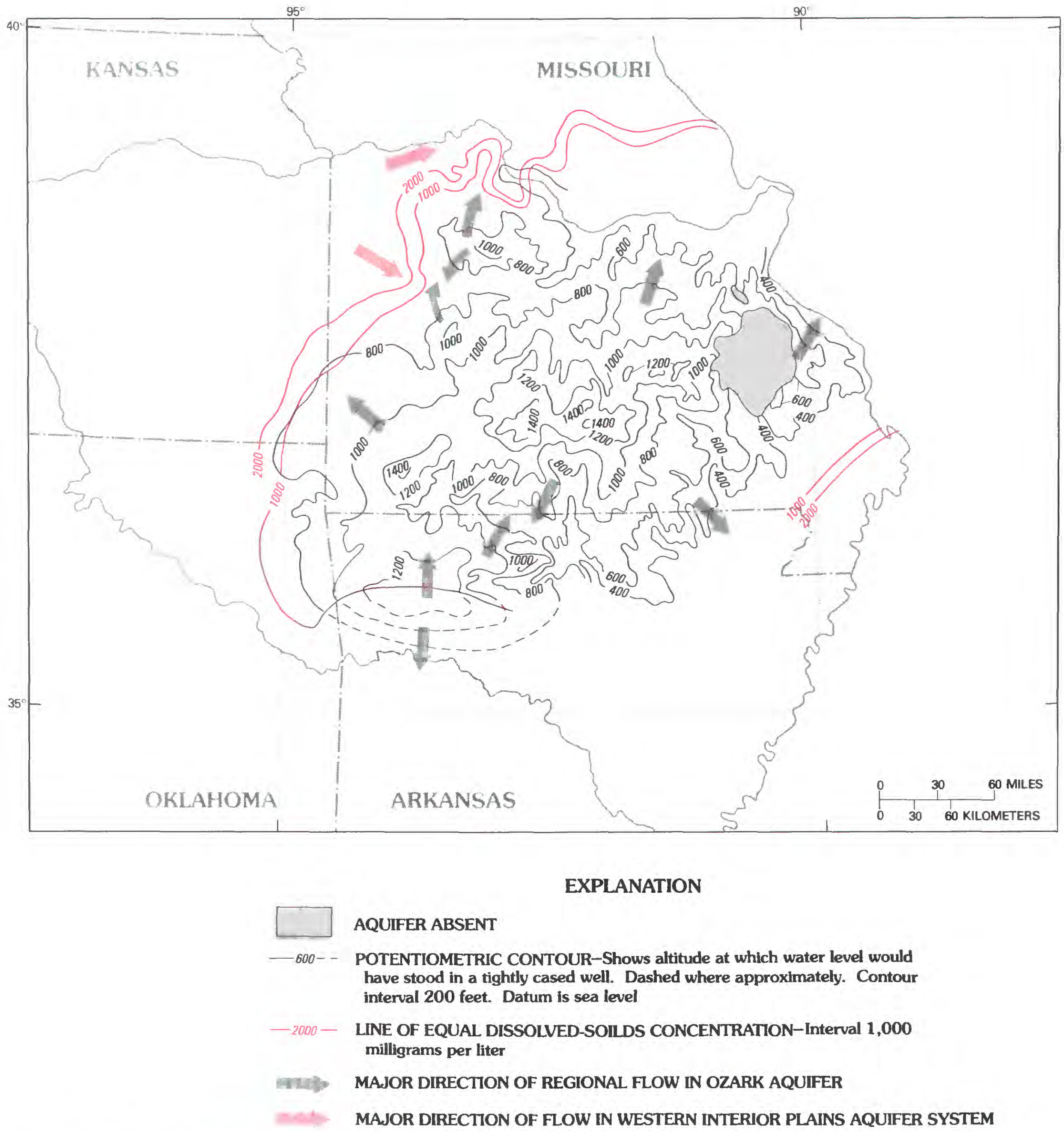


FIGURE 43.—Predevelopment potentiometric surface and dissolved-solids concentrations in water from the Ozark aquifer.

However, at most locations, the unit is less than 100 ft thick and is relatively thin in comparison to other geohydrologic units in the Ozark subregion.

Along the northwestern boundary of the Ozark subregion, the confining unit is lithologically more complex than elsewhere in the subregion. Limestone predominates in this sequence, but shale also is present in large quantities. Sandstone is present in small quantities. The thickness of the major shale formations, the Maquoketa Shale, Orchard Creek Shale, Chattanooga Shale, and the Hannibal Shale, may total as much as 100 ft; additionally, many of the limestone formations also contain shale. In the northern and northwestern parts of the Ozark subregion, the lithology varies from mostly shale to mostly limestone. Southward, the confining unit is represented solely by the Chattanooga Shale, which is predominantly shale with small amounts of sandstone.

The shale content of the Ozark confining unit, which is an indicator of the confining unit's effectiveness, ranges from near 0 to 100 percent. In southwestern Missouri and southeastern Kansas, the percentage of shale in the confining unit changes abruptly within short distances and ranges from less than 20 to 100 percent. In much of southwestern and southern parts of the Ozark subregion, shale constitutes 100 percent of the confining unit.

SPRINGFIELD PLATEAU AQUIFER

The Springfield Plateau aquifer is composed of a sequence of permeable Mississippian limestone. The aquifer is the uppermost unit of the Ozark Plateau aquifer system and at most locations overlies the Ozark confining unit.

The outcrop of the aquifer corresponds approximately with the Springfield Plateau (fig. 41) from which it derives its name. The aquifer is used extensively as a source of stock and domestic water in the outcrop area, usually in combination with the underlying Ozark aquifer. The Springfield Plateau aquifer is present also in the subsurface south of the Boston Mountains. In that area, the aquifer contains water with more than 1,000 mg/L dissolved solids probably because near-surface permeability has not developed and because slow-moving water dissolves minerals as it passes through the slightly permeable rocks. The aquifer is unused in that area because of both the small yield to wells and large dissolved-solids concentrations.

The aquifer crops out around the western and southern perimeter of the Salem Plateau and along the southwestern extension of the Ozark Uplift axis. The aquifer is not present in the area adjacent to the Mississippi Alluvial Plain and in the subsurface in small areas near the western edge of the Springfield Plateau and the southern and southwestern edge of the Ozark subregion in Arkansas and Oklahoma. The Springfield Plateau aquifer has a relatively uniform thickness of 100 to 400 ft throughout most of the subregion but is much thicker to the northeast in St. Louis County, Missouri, and in a narrow area along the Mississippi River in Perry County where a thickness of 1,500 ft is shown.

Hydraulic-conductivity values from aquifer tests or specific-capacity tests of the Springfield Plateau aquifer are not available (Jorgensen and others, 1993).

The Springfield Plateau aquifer is recharged in its outcrop areas. In general, water moves locally from high intervalley recharge areas to low discharge areas along the streams that dissect the area. Regionally, lateral flow in the Springfield Plateau aquifer is believed to be similar to that in the underlying Ozark aquifer. In the western part of the Ozark subregion, the aquifer receives upward discharge from the Ozark aquifer. In the eastern part of the subregion, hydraulic gradients are reversed, and water is discharged downward to the underlying Ozark aquifer.

The Springfield Plateau aquifer is overlain along the western and southern boundaries by confining material, which is mostly Pennsylvanian shale. This overlying confining material is the eastern edge of the Western Interior Plains confining system. This confining system is composed of layers of very slightly permeable shale and layers of moderately permeable limestone, sandstone, and coal. The limestone layers have some fractures and slight permeability except near the land surface where dissolution has increased the permeability.

SIMULATION OF GROUND-WATER FLOW

A numerical ground-water flow model was used to test the conceptualization of flow in the regional aquifers and aquifer systems in the study area. Numerical flow models simulate ground-water flow and response of a ground-water system to stress by using information on aquifer properties, boundary conditions, and human-induced development to solve mathematical equations that quantify directions and rates of ground-water flow, water-level changes, stream-aquifer interactions, and effects of wells (Bachmat and others, 1980, p. 19).

The numerical model selected for this study is a finite-difference model termed, "The U.S. Geological Survey Modular Ground-Water Flow Model" (McDonald and Harbaugh, 1988). It is designed to simulate three-dimensional movement of ground water of constant density through porous material. Hereafter in this report, the numerical model will be referred to as the model.

The aquifer or aquifer systems simulated may be composed of heterogeneous and anisotropic materials, may have steady-state or transient flow regimes, and may have irregular boundaries. The model has the capability of using either of two solution procedures [strongly implicit (SIP) or slice-successive overrelaxation (SSOR)] to approximate the solution of the following equation:

$$\frac{\partial}{\partial x}\left(K_{xx}\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_{yy}\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_{zz}\frac{\partial h}{\partial z}\right)-W = S_s\frac{\partial h}{\partial t}, \qquad (5)$$

where

x, y, and z are Cartesian coordinates aligned along the major axes of hydraulic conductivity K_{xx}, K_{yy}, K_{zz} (LT^{-1});

h is the hydraulic head (L);

W is volumetric flux per unit volume and represents sources or sinks of water (T^{-1});

S_s is the specific storage of the porous media (L^{-1}); and

t is time (T).

The term "modular" refers to the computer-program structure of the model, which consists of a main program and a series of independent subroutines called modules. The modules are grouped into packages that deal with specific features of the hydrologic system to be simulated. Packages included in the model are those for simulating block-centered flow, rivers, natural recharge, wells, drains and springs, evapotranspiration, and head-dependent flux boundaries (termed general-head boundaries in the model documentation).

MODEL FRAMEWORK

The solution of equation 5 requires that the Cartesian coordinates of the model grid be aligned with the principal directions of the hydraulic-conductivity tensor. Ward (1968, p. 21) indicates a major structural lineament orientation in south-central Kansas that strikes N. 50°–70°E., with a mean strike of N. 60°E. A second set of lineaments oriented at an average value of 93° from the first set of lineaments strikes northwest between extremes of N.24°–56°W., with a mean and mode strike of N. 35°W. Ward (1968) also points out that the joint system studied appears to extend from southern Oklahoma to northeastern Kansas and western Missouri. Ward (1968) concluded that the joints may have formed as a result of northwest, horizontal, compressive forces. The lineaments are, in general, fractures. Because hydraulic conductivity tends to be greatest along the directions of greatest fracturing, the model grid was oriented to match the two principal directions of the fractures.

The model grid was oriented to coincide with the principal directions of lineaments, N. 35°W., from a point in east-central Kansas at the juncture of latitude 39°N. and longitude 96°W. As a consequence, the orthogonal column orientation from that juncture is N. 55°E. It is believed that the grid orientation for the model reasonably corresponds to the principal hydraulic-conductivity tensor of the aquifer system to be simulated. The model grid has 28 rows, 33 columns, and a uniform node spacing of 28 mi (fig. 44).

RELATION OF MODEL LAYERS AND GEOHYDROLOGIC UNITS

Five principal geohydrologic units in the regional study area are represented in the model as layers (fig. 45). From top to bottom:

Layer 1.—Surficial units, which consist of the High Plains aquifer (Gutentag and others, 1984), glacial drift, and the Great Plains confining system.

Layer 2.—The Great Plains aquifer system.

Layer 3.—The Western Interior Plains confining system.

Layer 4.—The upper unit in the Western Interior Plains aquifer system in the Plains subregion and the Springfield Plateau aquifer in the Ozark subregion.

Layer 5.—The lower units in the Western Interior Plains aquifer system in the Plains subregion and a combined unit in the Ozark subregion consisting of the Ozark aquifer, the St. Francois confining unit, and the St. Francois aquifer.

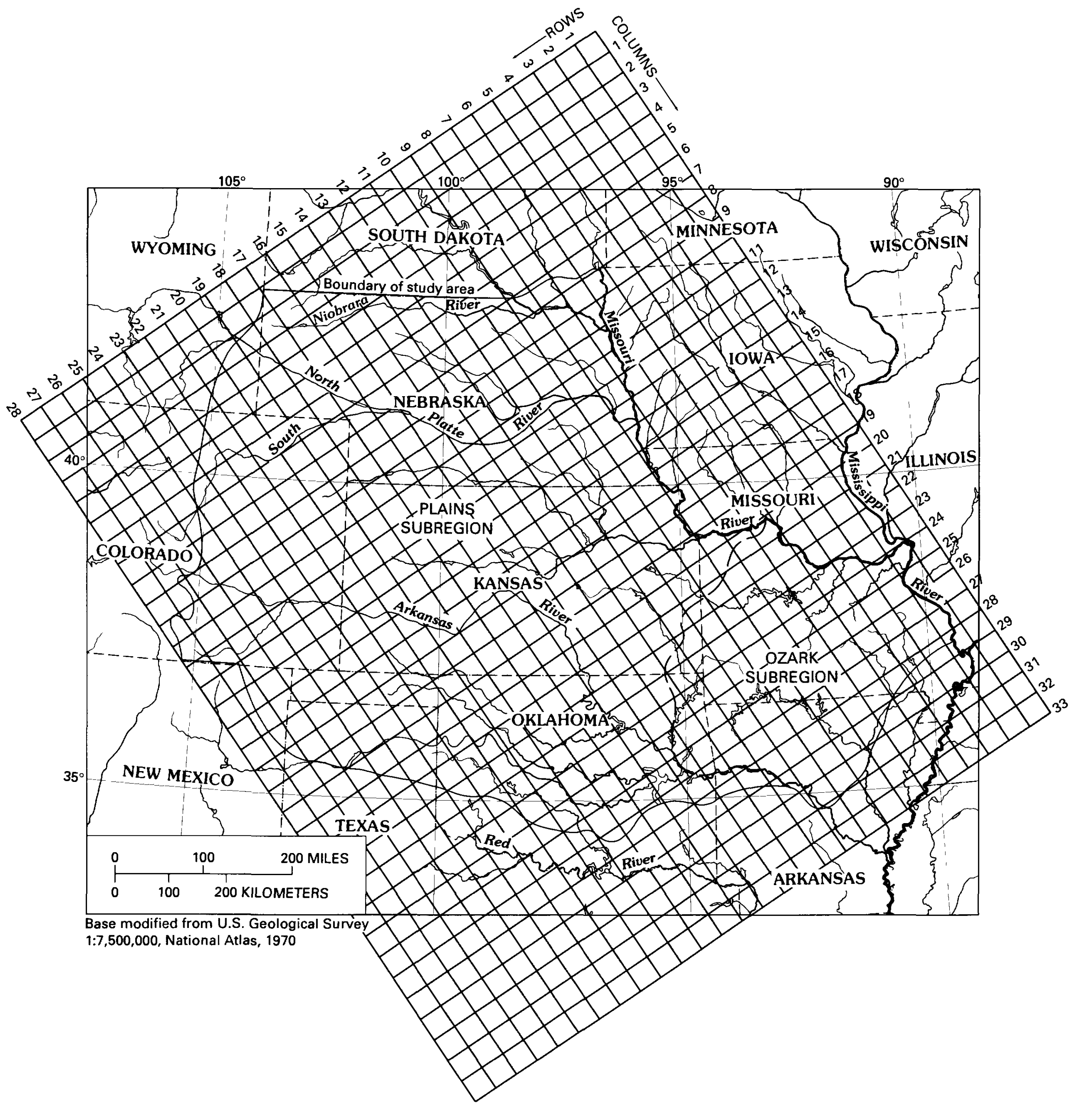
The regional model layer 4, the upper unit in the Western Interior Plains aquifer system, continues into the Ozark subregion as an equivalent layer simulating the Springfield Plateau aquifer. The rocks that are equivalent to the lower units in the Western Interior Plains aquifer system are the Ozark aquifer, the St. Francois aquifer, and the intervening St. Francois confining unit. In some areas there is a predominantly shale confining unit that impedes water flow between layers 4 and 5. This unit includes the Chattanooga Shale that occurs between the upper and lower units of the Western Interior Plains aquifer system and the Ozark confining unit that occurs between the Springfield Plateau aquifer and the Ozark aquifer.

BOUNDARY CONDITIONS OF MODEL LAYERS

Four types of boundary conditions affecting flow were used in the regional model. They were: (1) no flow, (2) specified head, (3) general-head boundary (head-dependent flux), and (4) specified flow. A description of actual geohydrologic boundary conditions and the assignment of corresponding lateral and vertical boundary conditions for each layer in the model are described in the following sections.

LAYER 1

The top layer of the flow model (fig. 46) simulates a composite hydrologic unit consisting of small areas of alluvium and glacial drift, the High Plains aquifer, plus shale of the underlying Great Plains confining system. Layer 1 is treated as a water source or sink and a confining upper boundary to the Great Plains aquifer system. Therefore, a specified-head boundary condition was used for all cells in layer 1. The data



EXPLANATION

— — APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 44.—Location of model grid and study area.

for the specified hydraulic heads were obtained by discretizing a predevelopment potentiometric-surface map of the geohydrologic unit overlying the regional aquifer systems. These data are represented by a machine plot of the potentiometric surface of model layer 1 (fig. 47).

The leakance coefficient between layer 1 and layer 2 was computed at all locations where the Great Plains confining system is present using confining-system thickness and the value for vertical hydraulic conductivity. The thickness of the confining system was computed as the difference between the altitude of the base of the High Plains aquifer (Gutentag and others,

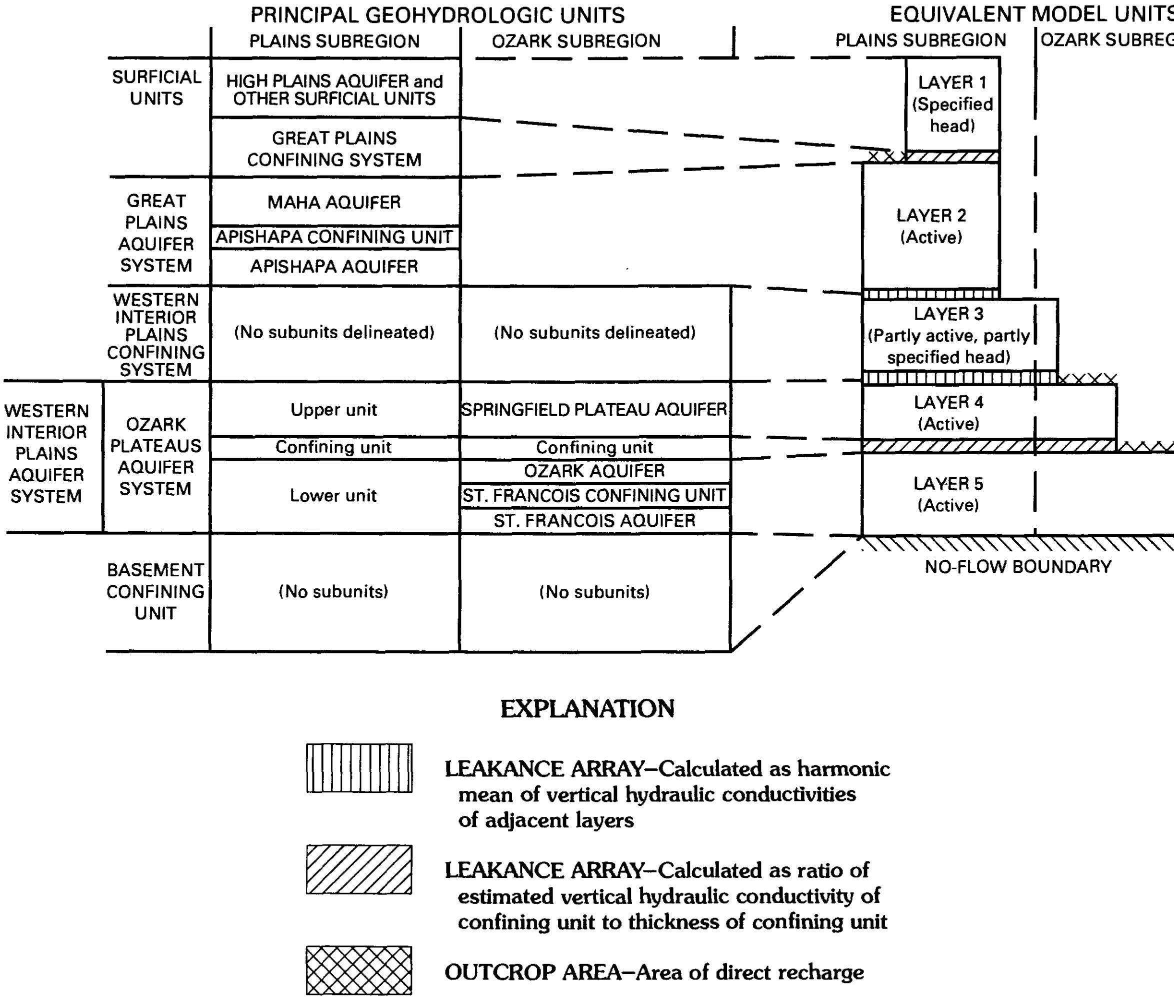
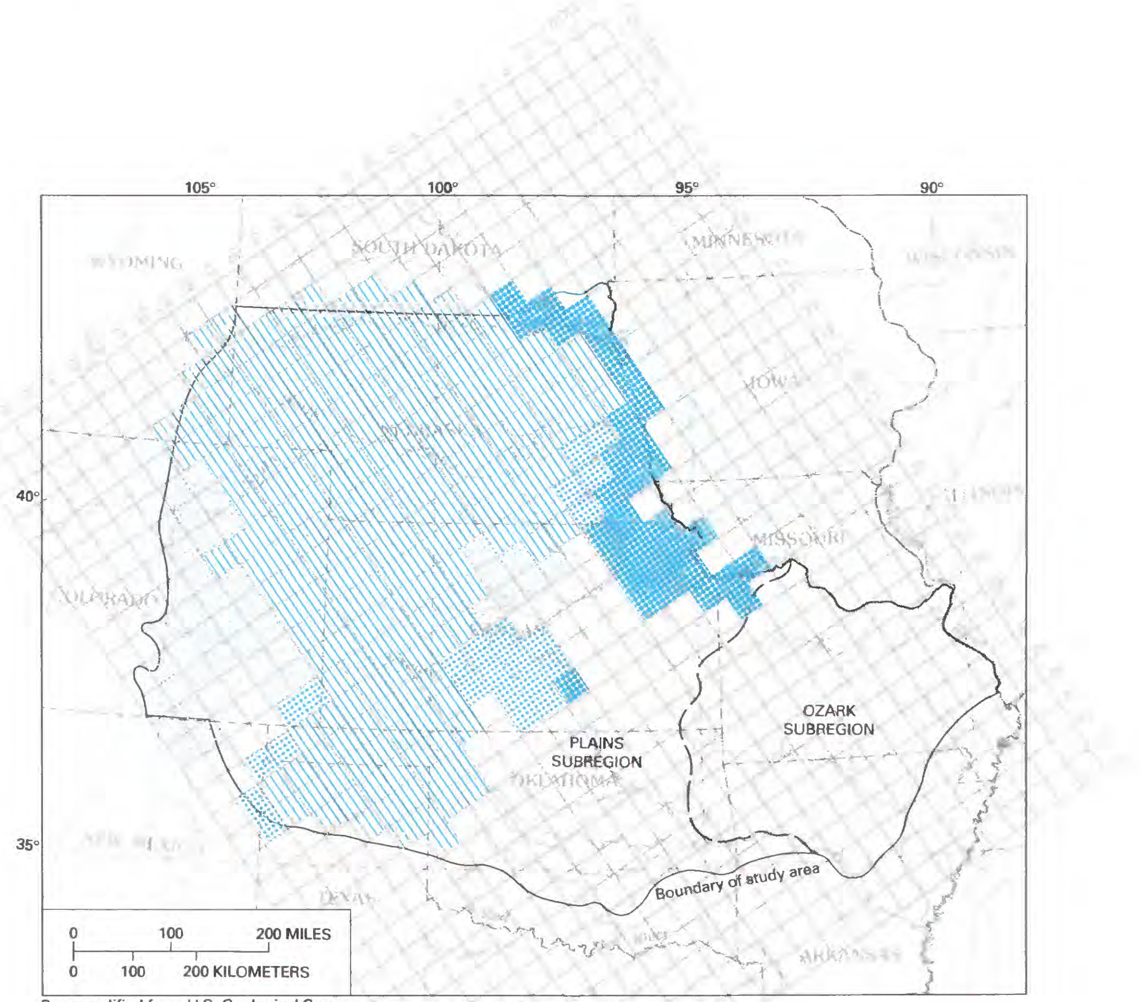


FIGURE 45.—Generalized model layers and relation to principal geohydrologic units in Plains and Ozark subregions (boundary conditions of model layers shown in figures 46 and 48–51).

1984, fig. 6, p. 18) and the top of the Great Plains aquifer system (Helgesen and others, 1993, pl. 4). The glacial drift is predominantly slightly permeable clay; therefore, it also is considered a confining unit. Where glacial drift is present and where the Upper Cretaceous shale crops out, the confining-system thickness is calculated as the difference between the potentiometric head in layer 1 (fig. 47) and the altitude of the top of one of the three underlying geohydrologic units—the Great Plains aquifer system, the Western Interior Plains confining system, or the Western Interior Plains aquifer system.

An estimate of the vertical hydraulic conductivity of the Great Plains confining system needed for computation of leakance factors was obtained by multiplying the lateral hydraulic conductivity of the Great Plains aquifer system (fig. 36) by $1x10^{-6}$. This constant multiplier is suggested by consideration of the 4-to-6 order-of-magnitude difference between hydraulic conductivity of shale and cemented sandstone (table 3). Davis (1969, p. 70, table 4) gives a value of $1.1x10^{-10}$ ft/s for the hydraulic conductivity of Cretaceous shale (lateral or vertical not specified). Bredehoeft and others (1983, p. 20, table 3) give a modeled best-fit value for vertical hydraulic conductivity of Cretaceous shale in South Dakota as $5x10^{-11}$ ft/s. The hydraulic conductivity of the Great Plains aquifer system generally would require a multiplier of about $1x10^{-6}$ to approximate the published estimates of hydraulic conductivity of shale, which is consistent with the information in table 3.

The intrinsic permeability of the Great Plains aquifer system tends to decrease with depth; that is, the thicker the overburden, the smaller the intrinsic permeability (figs. 36, 39). A similar permeability-depth relation also exists for Upper Cretaceous shale



Base modified from U.S. Geological Survey 1:7,500,000, National Atlas, 1970

EXPLANATION

ALLUVIUM AND GLACIAL DRIFT–High Plains aquifer absent

Extent of High Plains aquifer:

- HIGH PLAINS AQUIFER–Great Plains confining system absent
- HIGH PLAINS AQUIFER AND GREAT PLAINS CONFINING SYSTEM

Extent of Great Plains confining system:

- HIGH PLAINS AQUIFER AND GREAT PLAINS CONFINING SYSTEM
- GREAT PLAINS CONFINING SYSTEM–High Plains aquifer absent

SURFICIAL UNITS (Layer 1) ABSENT

— — APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 46.—Model layer-1 grid and areal extent of alluvium and glacial drift, the High Plains aquifer, and the Great Plains confining system.

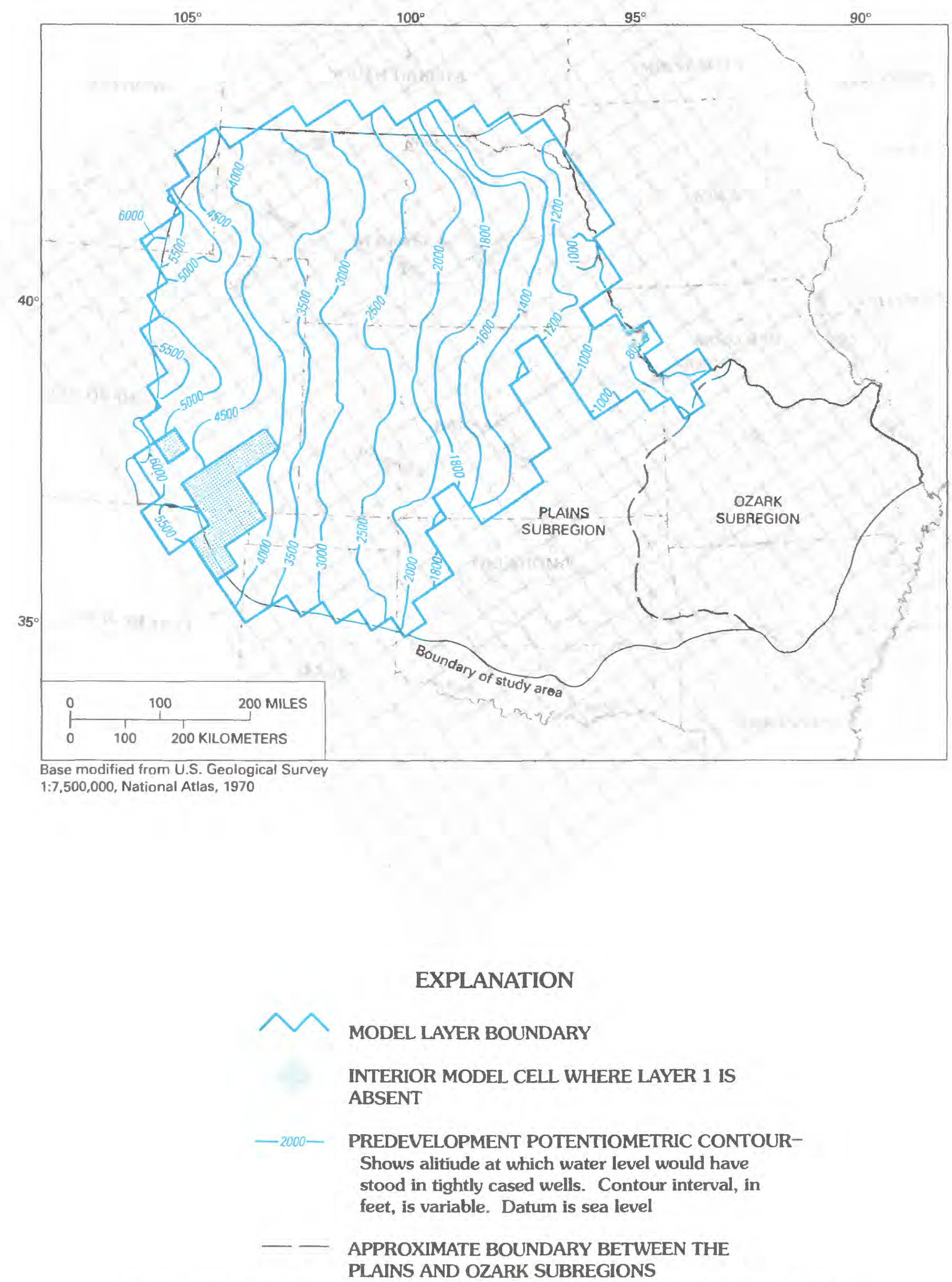


FIGURE 47.—Potentiometric surface of model layer 1 (machine plot of digitized data).

Table 3. Commonly observed ranges of permeability in several rock types (from Davis, 1969)

[Hydraulic conductivity (K) at 1.005 centipoise]

Square meter	10^{-19}	10^{-16}	10^{-12}	10^{-9}	10^{-7}
Square foot	10^{-18}	10^{-15}	10^{-11}	10^{-8}	10^{-6}

Millidarcy	10^{-4}	10^{-3}	10^{-2}	10^{-1}	1	10^{1}	10^{2}	10^{3}	10^{4}	10^{5}	10^{6}	10^{7}	10^{8}

Unconsolidated rocks	Unweathered clay and unweathered glacial drift.	Very fine sand; silt; mixtures of sand, silt, and clay; unstratified clay.	"Clean" sand; mixtures of "clean" sand and gravel.	"Clean" well-sorted gravel or cobble.
Consolidated rocks	Shale at depths greater than 250 feet; evaporite deposits; limestone at depths greater than 15,000 feet.	Sandstone; dolostone; limestone; fractured rocks; shale at depths less than 250 feet.	Extensively fractured rock; weathered limestone or dolostone.	Rocks with large openings.
Permeability characteristics	Very slightly permeable.	Slightly permeable.	Permeable	Very permeable.

Gallons per day per square foot	10^{-5}	10^{-4}	10^{-3}	10^{-2}	10^{-1}	1	10^{1}	10^{2}	10^{3}	10^{4}	10^{5}	10^{6}
Foot per day	10^{-6}	10^{-5}	10^{-4}	10^{-3}	10^{-2}	10^{-1}	1	10^{1}	10^{2}	10^{3}	10^{4}	10^{5}
Foot per second	10^{-11}	10^{-10}	10^{-9}	10^{-8}	10^{-7}	10^{-6}	10^{-5}	10^{-4}	10^{-3}	10^{-2}	10^{-1}	1

(Neuzil and others, 1984, p. 116, fig. 5). Assuming that the relation of permeability to depth for both shale and sandstone of the Great Plains aquifer system generally is valid, multiplying lateral hydraulic conductivity of the Great Plains aquifer system by a constant to determine vertical hydraulic conductivity of the Great Plains confining system also relates vertical hydraulic conductivity of the confining system to depth of burial.

Where the Great Plains confining system and glacial drift are not present, and permeable units such as alluvium and the High Plains aquifer are in direct contact with permeable materials of the Great Plains aquifer system and other sediments in layer 2, the computed leakance coefficient is the harmonic mean of vertically adjacent layers for the uniformly spaced model grid. In those locations, the thickness of layer 1 is computed also as the difference between the potentiometric head in layer 1 and the altitude of the top of the immediately underlying unit.

The vertical hydraulic conductivity used in computing leakance coefficients where the Great Plains confining system (Upper Cretaceous shale) and glacial drift are not present (and, therefore, permeable materials of layers 1 and 2 are in direct contact) was based on assumed ratios of vertical-to-horizontal permeability. Generally, vertical permeability is less than horizontal permeability in unfractured sedimentary rocks.

Gutentag and others (1984, p. 23) conclude that, on a regional scale, the sediments comprising the High Plains aquifer are distributed randomly in the vertical section. Cronin (1964, p. 33) presents observations that the individual beds or lenses of silt, sand, gravel, and clay of the Ogallala Formation (the major geologic unit of the High Plains aquifer) in the southern High Plains of Texas and New Mexico are not continuous over wide areas. This is generally true of the High Plains aquifer. Because there are clay lenses that apparently are not areally extensive, a vertical-to-lateral ratio of hydraulic conductivity in model layer 1 was specified as 0.1. This hydraulic-conductivity ratio was applied to areas in layer 1 where permeable materials directly overlie the Great Plains aquifer system or other underlying aquifers.

LAYER 2

The second layer of the flow model (fig. 48) simulates flow in the Great Plains aquifer system, which includes the Maha and Apishapa aquifers and the Apishapa confining unit. General-head, constant-flow (recharge), and no-flow boundaries are used in layer 2 for simulation of the Great Plains aquifer system.

Along the eastern extent of the Great Plains aquifer system in northeastern Nebraska, central and western Kansas, the Oklahoma Panhandle, and the Texas Panhandle, the aquifer system pinch-out is well defined. No-flow boundaries in model layer 2 simulate that condition. For each cell that simulates outcrop areas of the Great Plains aquifer system, general-head and constant-flow (recharge) boundaries are used to account for both stream controls and direct recharge from precipitation. The aquifer system crops out in a small area at its eastern extent and in a much larger area in southeastern Colorado and northeastern New Mexico (fig. 48).

Model cells representing areas along the Front Range of the Rocky Mountains in New Mexico, northern Colorado, and southeastern Wyoming are no-flow boundaries (fig. 48) because those cells represent the extent or lateral termination of the Great Plains aquifer system. Those cells representing outcrop locations in those areas are general-head and recharge boundaries. Outcrop locations in northern Colorado and southeastern Wyoming are in narrow bands where the rocks of the aquifer system were uplifted and exposed. There are no data available to indicate the amount of inflow that occurs, but because of the small amount of surface exposure, inflow or recharge is likely to be small.

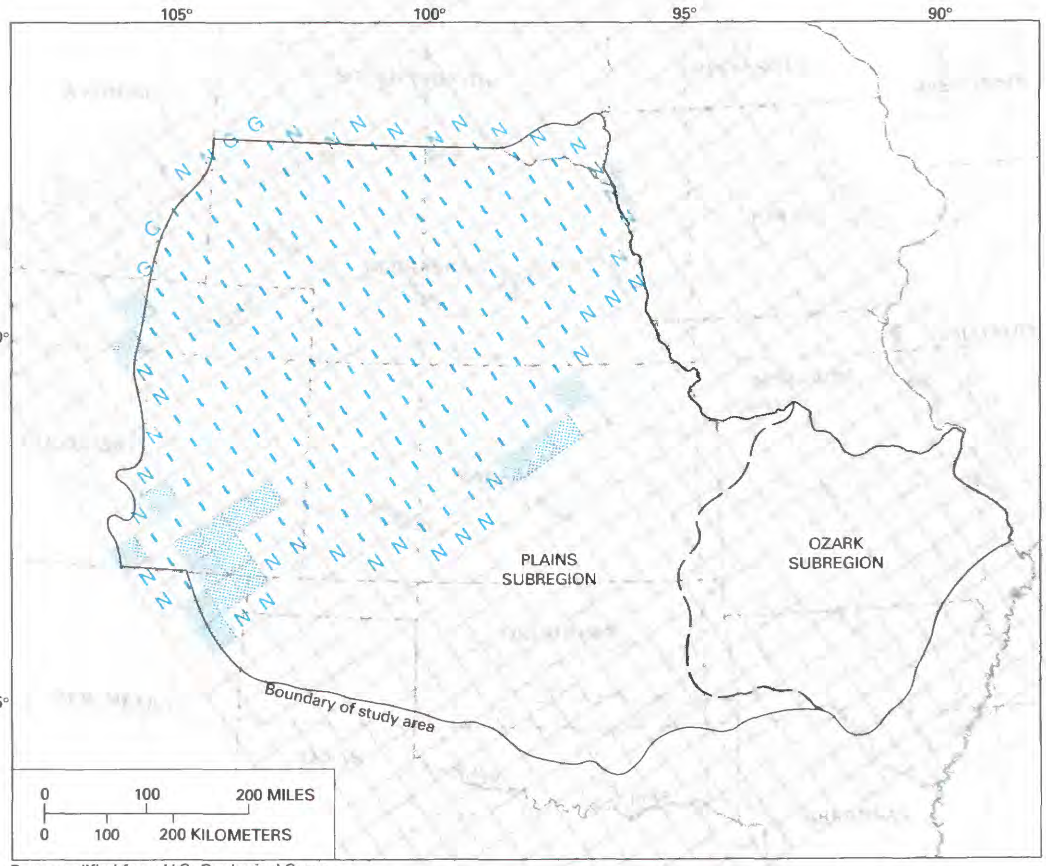
At the northwest corner of Nebraska and southwest corner of South Dakota, and in the Black Hills Uplift just north of the study area where the Great Plains aquifer system crops out, the potentiometric-surface map (fig. 34) indicates the potential for recharge. Miller and Rahn (1974) estimate about 5 ft^3/s of recharge for the entire outcrop area of the Dakota Sandstone (the major component of the Maha aquifer) on the east side of the Black Hills; however, most of that recharge remains in South Dakota and does not enter the study area. Case (1984, p. 158) estimated ground-water movement from Wyoming into South Dakota in the aquifer south of the Black Hills as 0.7 ft^3/s. Cells representing that area are specified as general-head boundaries; however, recharge is expected to be small.

The Great Plains aquifer system extends beyond the limit of the regional study area along the State line between Nebraska and South Dakota and at the northeast corner of Nebraska. In general, the potentiometric-surface map (fig. 34) implies flow parallel to or nearly parallel to the South Dakota-Nebraska State line, indicating little or no flow across the Nebraska State line. Therefore, a no-flow boundary condition is specified for cells in layer 2 along the South Dakota-Nebraska State line.

The ratio of vertical-to-lateral hydraulic conductivity for layer 2 was assumed to be 1:20. The basis for this assumption is the stratification of the Great Plains aquifer system. Helgesen and others (1982) indicate that the Great Plains aquifer system has substantial lithologic heterogeneity consisting of interbedded sandstone, siltstone, and mudstone, which commonly exhibit local facies changes associated with terrestrial and nearshore marine environments of deposition. Because of these conditions, vertical hydraulic conductivity would be relatively small as compared to lateral hydraulic conductivity. The ratio of vertical-to-lateral hydraulic conductivity is required to compute the vertical hydraulic conductivity and thus the leakance coefficient between layer 2 and overlying units where permeable materials are in direct hydraulic contact. However, for most of the area of the Great Plains aquifer system, vertical leakage is controlled by the Great Plains confining system (fig. 32) represented in model layer 1 and evaporite deposits included in the Western Interior Plains confining system represented in layer 3 (fig. 30).

LAYER 3

The third layer of the numerical model represents the Western Interior Plains confining system, a thick section of mostly Pennsylvanian and Permian rocks (fig. 49). The rocks are interbedded shale, limestone, sand, and evaporite deposits. The predominant hydraulic characteristic of the confining system is very small vertical permeability owing to rocks such as extensive halite and gypsum deposits. Because of the generally very small vertical permeability, the Western Interior Plains confining system is simulated by model layer 3 as a vertical-flow restriction in model cells representing areas where the system is buried. For cells representing outcrop areas, the model layer is an upper confining boundary and specified head. Zero leakance between layer 2 and layer 3 is used in the model to simulate areas where evaporite deposits are present (fig. 30). A non-zero leakance coefficient



Base modified from U.S. Geological Survey 1:7,500,000, National Atlas, 1970

EXPLANATION

OUTCROP AREAS

General-head and recharge boundaries

BURIED AREAS–Any cell that has a common face with an inactive cell and has no other imposed boundary condition is termed a "no-flow" boundary

G General-head boundary

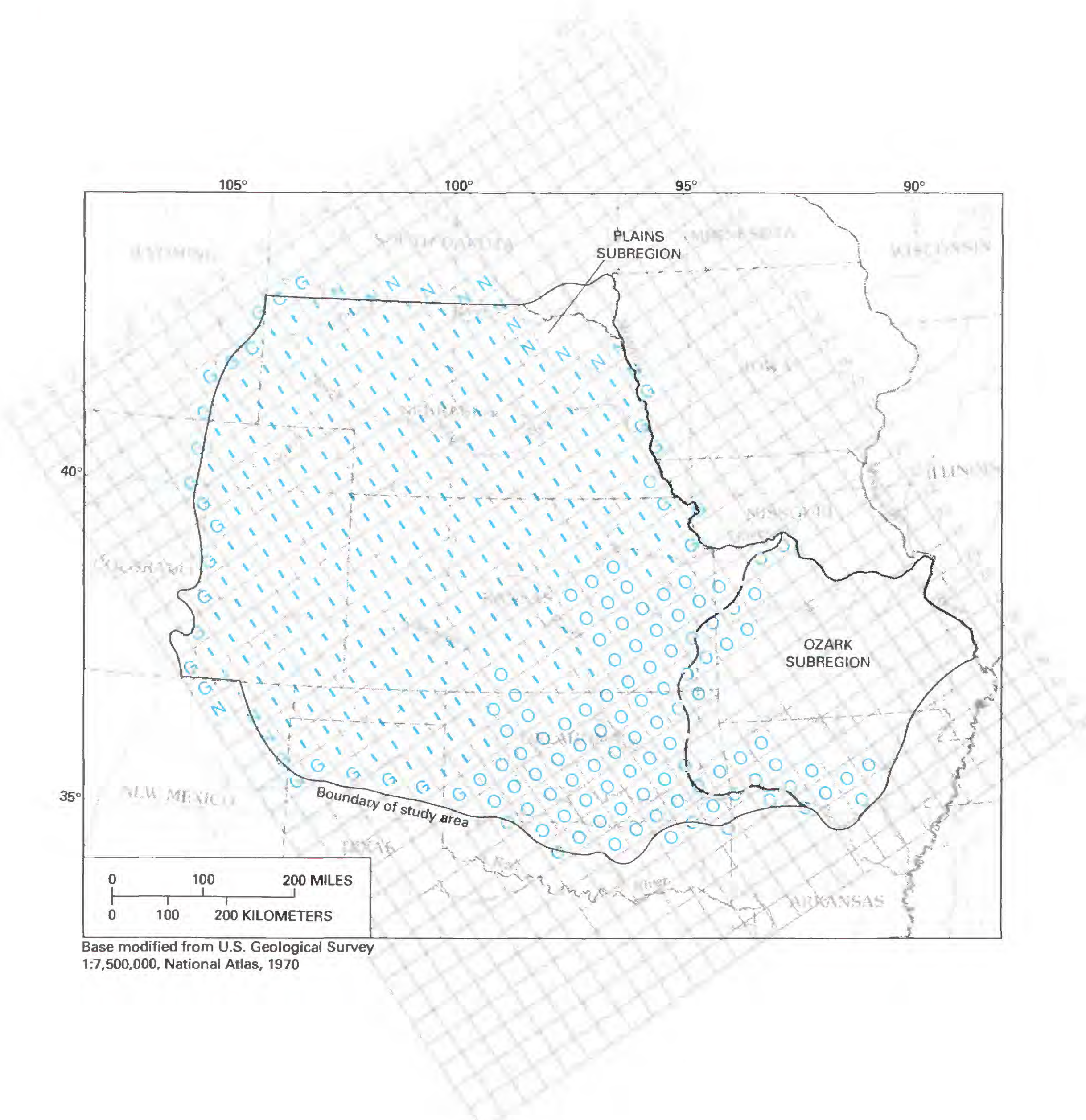
N No-flow boundary

ACTIVE CELL WITHOUT A BOUNDARY CONDITION

INACTIVE CELL

APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 48.—Model layer-2 grid representing areal extent and boundary conditions of the Great Plains aquifer system.



EXPLANATION

- OUTCROP AREA–Specified-head boundary
- GENERAL-HEAD BOUNDARY
- NO-FLOW BOUNDARY–Any cell that has a common face with an inactive cell and has no other imposed boundary condition is termed a "no-flow" boundary
- ACTIVE CELL WITHOUT A BOUNDARY CONDITION
- INACTIVE CELL
- APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 49.—Model layer-3 representing areal extent, outcrop area, and boundary conditions of the Western Interior Plains confining system.

between layers 2 and 3 is computed on the basis of mapped intrinsic permeability of the Western Interior Plains confining system (fig. 29) to represent an area where evaporite deposits are not present, assuming a vertical-to-lateral hydraulic-conductivity ratio of 1:1.67x10^6. However, leakance coefficients between layer 3 and underlying layer 4 were computed only on the basis of the Western Interior Plains confining system vertical-to-lateral hydraulic-conductivity ratio of 1:4.17x10^7 and the thickness of layer 3.

Layer 3 of the model simulates flow in an outcrop area second in size only to layer 1. The specified head in the cells that represent the outcrop area are the shallow ground-water levels (approximate water table) (fig. 31). An extremely small leakance coefficient is used between layers 3 and 4. Thus, the specified heads of layer 3 can simulate a source or sink of water to or from the near-surface water table or to or from underlying aquifers represented by layer 4 or 5. Because the initial hydraulic head for cells simulating the outcrop area is specified, that area of the model "intercepts" all other inputs or outputs; thus no flows related to springs, drains, streams, recharge, or withdrawals affect the simulation in these cells.

The computed hydraulic head represents the entire thickness of the confining system at its centerline. The hydraulic head in a confining unit actually varies in the vertical, and a single value cannot represent the hydraulic head throughout the entire thickness. Therefore, the model-computed hydraulic head cannot be correlated directly to measured water levels unless the measured data were acquired at the centerline of the confining unit. Simulation of steady-state conditions should result in computed hydraulic-head values that are midway between the values in the overlying and underlying layers where a vertical connection is allowed. However, where no vertical connection with layer 2 above the confining unit is allowed, computed hydraulic head is affected slightly by the calculated leakage to or from underlying layer 4.

General-head boundaries are specified at the lateral extent of layer 3 from the northwest corner of Nebraska southward to northern New Mexico. Also, cells representing the extent of the confining system across the Texas Panhandle and 30 mi into Oklahoma are general-head boundaries. In eastern Nebraska and northeastern Kansas, general-head boundaries were used for cells near the Missouri River where water levels are nearly constant and hydraulic continuity exists. Along the State line between Nebraska and South Dakota, the study boundary was simulated as no flow.

The rationale for these boundary conditions was based on preliminary data from drill-stem tests that show hydraulic gradients from west to east generally paralleling the western extent of the confining system in the Panhandle of Texas. The hydraulic-head contours are normal to the Nebraska-South Dakota State line, indicating flow approximately parallel to the State line. There is no information on inflows or outflows at that location in the Western Interior Plains confining system, and specification of general-head boundaries allowed the model to compute flow rates that were consistent with the available hydraulic-head data.

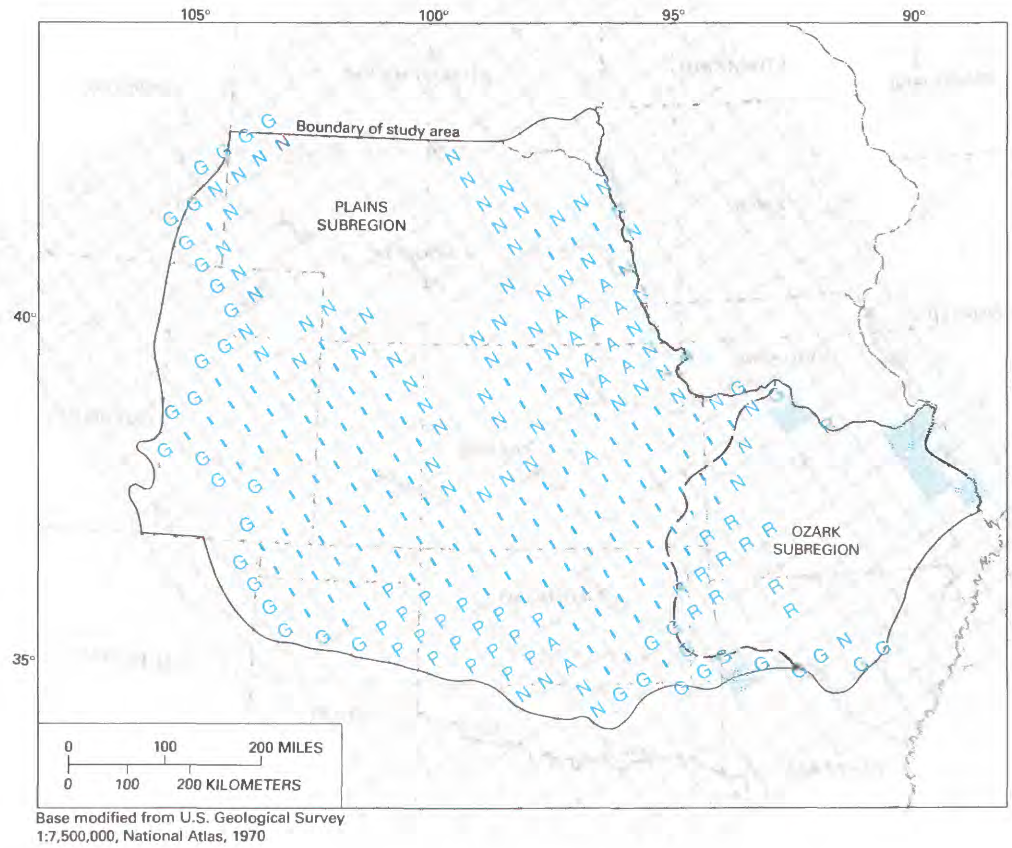
Data collected for the Western Interior Plains confining system show a geopressure area in the Anadarko Basin of west-central Oklahoma. Specifically, the geopressure exists in the lowest rocks in a very thick section of the Western Interior Plains confining system. The source of the high pressure is not simulated in model layer 3, but in layer 4 as explained below.

LAYER 4

The upper unit in the Western Interior Plains aquifer system and the Springfield Plateau aquifer in Mississippian-age rocks are simulated in the model by layer 4 (fig. 50). The Mississippian-age rocks are predominantly limestone with some sandstone and shale layers. The rocks have slight permeability where deeply buried and are permeable in outcrop areas in the Ozark subregion.

Cells that simulate outcrop areas in layer 4 are subjected to a constant-flow boundary condition by the recharge option of the model. No-flow boundaries are specified for the cells at the limit of layer 4 (fig. 50), principally where the rocks of the upper unit in the Western Interior Plains aquifer system pinch out. General-head boundaries (fig. 50) were used along the southern and western edges where the upper unit generally extends to the study-area limits.

In the Anadarko Basin of west-central Oklahoma, there is a geopressure zone (pl. 2) that results from overpressuring in the overlying layer 3. The overpressuring cannot be readily explained. Because the pressure exists immediately above the upper unit in the Western Interior Plains aquifer system and vertically much closer to the centerline of the rocks simulated in layer 4 than to the centerline of the rocks simulated by layer 3, the pressure source is more effectively simulated in layer 4. Simulation is by a general-head boundary using the hydraulic-head values digitized from map data (pl. 2) as the controlling hydraulic head and an arbitrary large head-dependent flux coefficient. Using the general-head option of the model allows head-dependent lateral flow in layer 4 and vertical flow to layers 3 and 5 due to large hydraulic-head



EXPLANATION

OUTCROP AREAS

General-head and recharge boundaries

R Recharge boundary

BURIED AREAS–Any cell that has a common face with an inactive cell and has no other imposed boundary condition is termed a "no-flow" boundary

G General-head boundary

P Overpressure, general-head boundary

A Geohydrologic unit absent

N No-flow boundary

ACTIVE CELL WITHOUT A BOUNDARY CONDITION

INACTIVE CELL

— — APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 50.—Model layer-4 grid representing areal extent and boundary conditions of the upper unit in the Western Interior Plains aquifer system and the Springfield Plateau aquifer in the Ozark Plateaus aquifer system.

values from an overpressure. The imposed conditions produce the effect of overpressure on the simulated flow in layer 4.

LAYER 5

The lower units in the Western Interior Plains aquifer system are simulated by model layer 5 (fig. 51). The rocks simulated are primarily thick sections of dolostone, limestone, and sandstone of Devonian, Silurian, Ordovician, and Cambrian age. Permeability is very small in deep sections in the western part of the study area and in the Anadarko Basin, increasing by about three orders of magnitude across Kansas and Nebraska from south to north to slightly permeable (fig. 20). The rocks are more permeable in the Ozark Uplift area, particularly where the Ozark and St. Francois aquifers crop out and have been subjected to weathering.

No-flow boundaries are specified at the northwestern extent of layer 5 in northeastern Colorado and central and northeastern Nebraska. Toward the northeastern part of the study area, the lower units in the Western Interior Plains aquifer system continue beyond the study area under the Missouri River trench into western Iowa and central Missouri. Where the Missouri River is incised directly into these lower units, hydraulic connection exists between the river and these units. The exchange of water between the river and the lower units is simulated using a general-head boundary. General-head boundaries also are used to simulate the continuation of the lower units generally to the southern and western extents of the study area. Recharge (constant-flow) cells, ranging from -393 to 544 ft^3/s, simulate the outcrop area, the rates of which are based on an analysis presented by Jorgensen and others (1989a, b).

SIMULATION OF VARIABLE-DENSITY FLOW

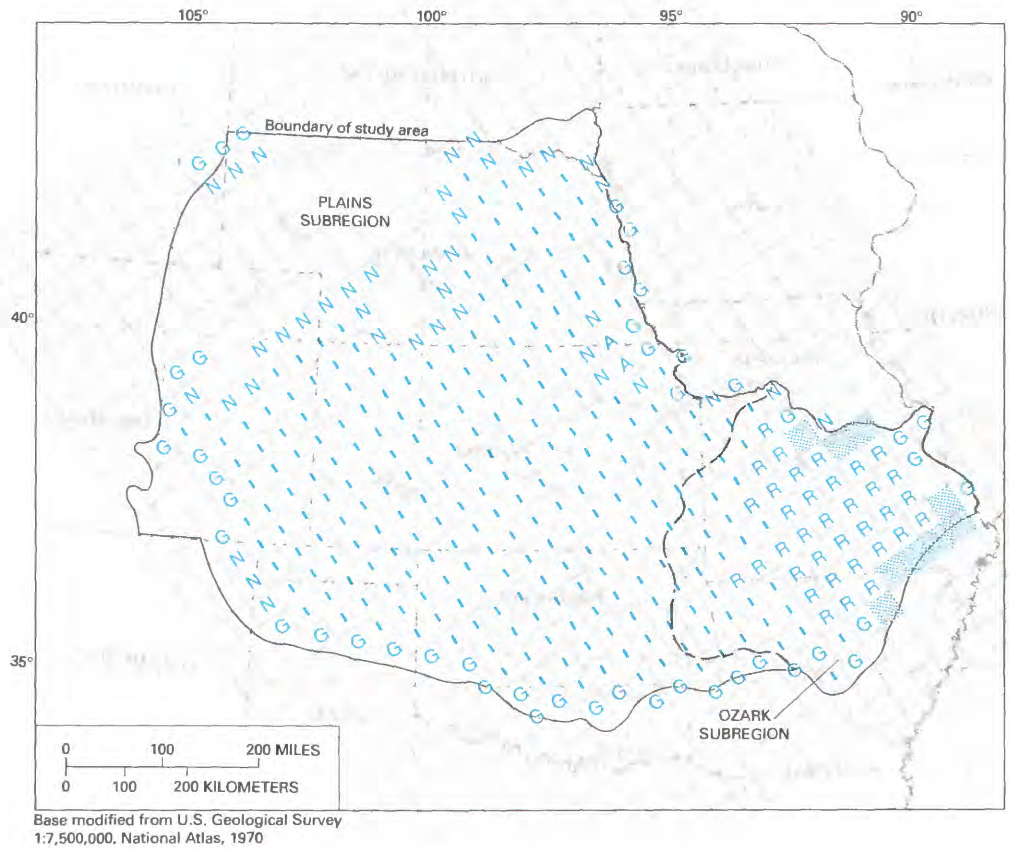
The modeled region includes geohydrologic units that contain water with large concentrations of dissolved solids. Water at temperatures exceeding mean annual temperature also is present, especially at great depths. Large concentrations of dissolved solids in water cause it to have greater density than freshwater. Because temperatures at great depths generally exceed the mean annual temperature, water density is reduced. These two conditions may have offsetting effects on deeply buried aquifers containing water having large concentrations of dissolved solids; however, it is more likely that these conditions result in a variable fluid density that is different from uniform-temperature freshwater conditions. The flow equation (eq. 5) is based on a mathematical development valid only for a constant, uniform-density fluid. Weiss (1982) developed a computer program that generates source and sink input to ground-water models to enable them to simulate the effect of variable-density ground-water flow. For application to the CMRASA digital model, Weiss' code was modified to handle a multilayer system; however, the computational technique was unchanged.

MODEL CALIBRATION

Calibration of the model herein refers to the process in which the model-input information needed for simulation is modified (or calibrated) such that the model is able to simulate measured or independently calculated data of the geohydrologic system being modeled. Neither is it meant that the model simulations are unique nor that the model will accurately simulate conditions other than those specified for the calibration. Calibration was accomplished by matching model-computed hydraulic-head values to either measured hydraulic head or, in lieu of measured head, independently estimated hydraulic head. Thus, the measured or estimated head, collectively referred to herein as "field hydraulic head," is that which is compared to the model-computed hydraulic head to evaluate a simulation. The field hydraulic-head distributions used for calibration are assumed to represent steady-state (predevelopment) conditions. As discussed in the section entitled "Hydraulic Head," considerable water-level data are available in the shallow parts of aquifers. However, head data are scarce in deeper, saline parts of the aquifers, and the predevelopment potentiometric surfaces are, in part, estimated.

CALIBRATION CRITERIA

Calibration criteria of the model were based on closeness of fit of model-computed hydraulic head to field hydraulic head. Regional outflow and inflow at the boundaries and subregional-model hydraulic-head values and flows also were considered. A criteria of average difference from the field hydraulic head of ± 1 ft and a standard deviation of 300 ft was specified for calibrating hydraulic head in layers 2, 4, and 5. Layer 1 is a specified-head source-sink layer (fig. 47); therefore, computed hydraulic-head values for that layer were the field hydraulic-head values. In layer 3, the outcrop area was a specified-head boundary (fig. 49); accordingly, the field hydraulic-head values were the computed hydraulic-head values for those particular nodes. Hydraulic-head values in layer 3



EXPLANATION

OUTCROP AREAS

General-head and recharge boundaries

R Recharge boundary

BURIED AREAS–Any cell that has a common face with an inactive cell and has no other imposed boundary condition is termed a "no-flow" boundary

G General-head boundary

A Geohydrologic unit absent

N No-flow boundary

ACTIVE CELL WITHOUT A BOUNDARY CONDITION

INACTIVE CELL

— — APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 51.—Model layer-5 grid representing areal extent and boundary conditions of the lower units in the Western Interior Plains aquifer system and the Ozark and St. Francois aquifers in the Ozark Plateaus aquifer system.

where it is a subsurface unit were computed by the model. However, adjustments to the leakance arrays controlling leakage to and from layer 3 were made with the objective of obtaining computed hydraulic-head values in layers 2, 4, and 5 within the calibration criteria, and consideration of the computed hydraulic-head values in layer 3 was incidental to that objective.

The largest estimated regional water-budget components were for surface recharge. Estimated maximum values of recharge are indicative of a water-resource potential, but it is unlikely that amount of recharge would occur on a sustained basis. Lateral flow at aquifer boundaries simulated by general-head boundaries was fairly large at the boundary near the Mississippi Alluvial Plain and the Missouri and Mississippi Rivers. Vertical leakage, although generally small for individual cells, was significant for entire layers, particularly in aquifers in the Ozark subregion where a vigorous, freshwater flow system predominates.

Calibration criteria for water-budget components were not defined because of the extent of the regional study area and the lack of data. However, if the calibration criteria for matching simulated hydraulic heads to field hydraulic heads can be met using aquifer-characteristic values within the prototype-to-model range of variability, the model-computed components of the budget, such as outflow to rivers and extent boundaries, may be assumed to reasonably simulate the water budget. From the recharge analysis in the Ozark subregion, quantities of recharge and discharge were established on a cell-by-cell basis.

Geohydrologic units represented by layers 4 and 5 crop out in the Ozark subregion, and the outcrop areas are recharged directly from precipitation. Most of the net flow to the aquifers in the Ozark subregion is subsequently discharged to major rivers, such as the Missouri, Mississippi, and Arkansas, and to vertical leakage in the western part of the Ozark subregion and across the boundary between the Ozark subregion and the Mississippi Alluvial Plain.

In recharge areas of the Great Plains aquifer system represented by the regional model layer 2, net recharge values were uniformly reduced by the same factor as for the Ozark subregion. The stream-aquifer interaction values, however, were unknown. Some outcrop-area recharge rates, such as those occurring in hogback areas along the Rocky Mountain Front Range, were analyzed by Helgesen and others (1993), and those values were used.

HYDRAULIC HEAD

Comparison of computed hydraulic heads from the model to the field hydraulic heads for layers 2, 4, and 5, was made numerically (table 4) and graphically (figs. 52–57). The effects of variable density will be discussed separately.

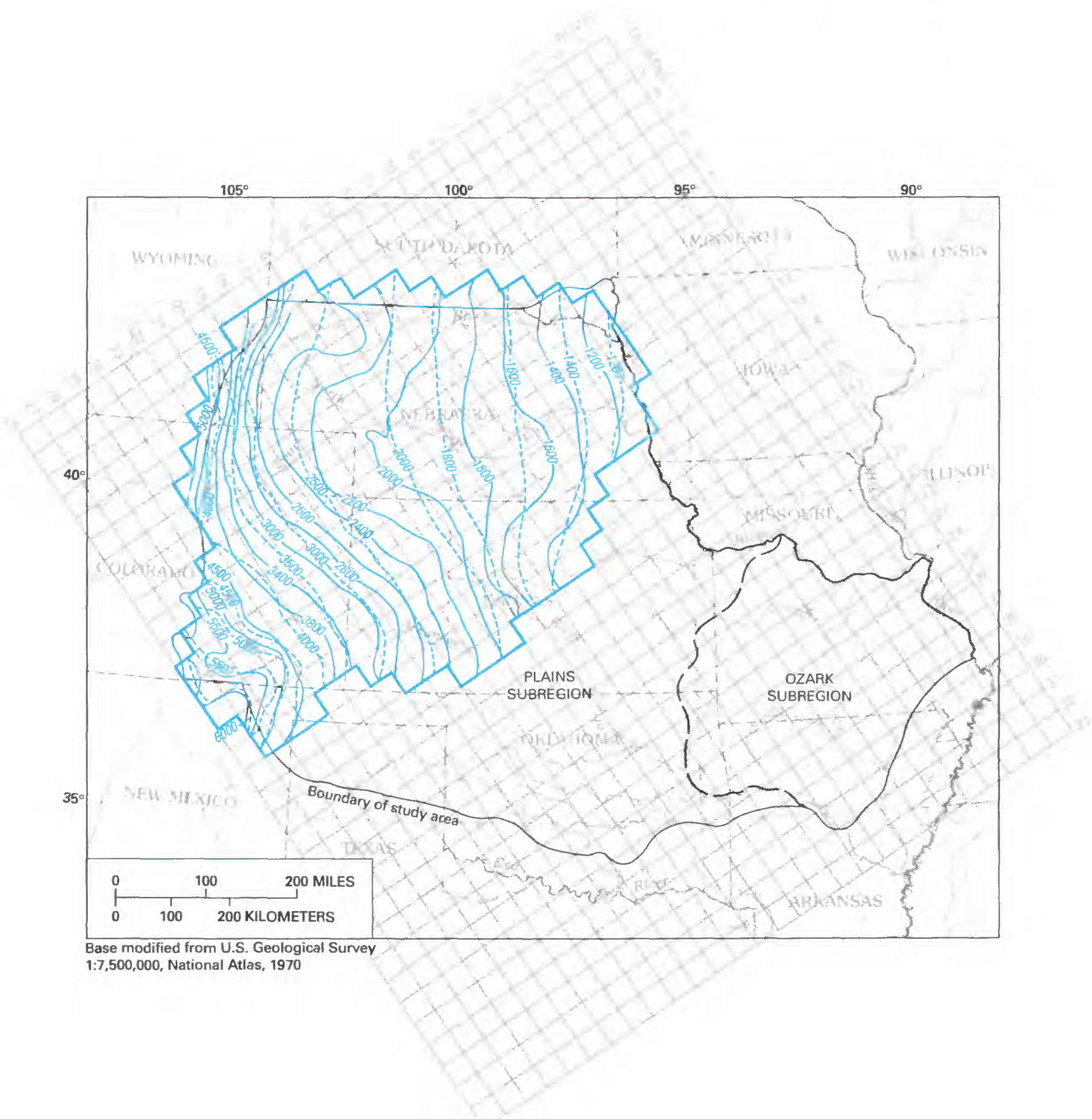
For layer 2, the average deviation from field hydraulic-head values (table 4) is less than 0.2 ft. The standard deviation of about 160 ft indicates, considering the average deviation from field values (0.2 ft) is in essence zero, that 95 percent of all deviations from the field hydraulic head are less than 320 ft. Another way to evaluate the variability is that 95 percent of all values of computed hydraulic head lie in the interval of the average plus or minus two standard deviations (Snedecor, 1957, p. 36).

There are four areas of major deviation from field hydraulic heads for layer 2, as shown in figures 52 and 53. In two areas, model-computed head values are 300 to 400 ft higher than the field values. These areas are located in east-central and northern Colorado where the Great Plains aquifer system is deeply buried. Two areas have computed head values significantly smaller than the field head values. One area is a strip in central Colorado along the Rocky Mountains where most of the lateral boundaries for layer 2 are no-flow boundaries. Along the mountains, the Great Plains aquifer system dips toward the east, and the field hydraulic-head values may not truly define the hydraulic-head distribution in the layer. Also, the McDonald and Harbaugh (1984) model assumes

TABLE 4.—*Statistical comparison of model-computed hydraulic head to field hydraulic head*

		Statistics for differences between computed and field hydraulic head					
		Average deviation [1] (feet)		Mean square deviation (square feet)		Standard deviation (feet)	
Model layer	Number of values	Constant density	Variable density [2]	Constant density	Variable density	Constant density	Variable density
2	248	0.17	-5.87	24,956	24,331	157.97	155.98
4	301	-.82	30.22	54,785	48,720	234.06	220.73
5	382	-.13	40.82	90,561	80,276	300.93	283.33

[1]Negative average indicates that the computed hydraulic-head average is less than the field hydraulic head.
[2]Model computation includes "pseudosources" (Weiss, 1982).



EXPLANATION

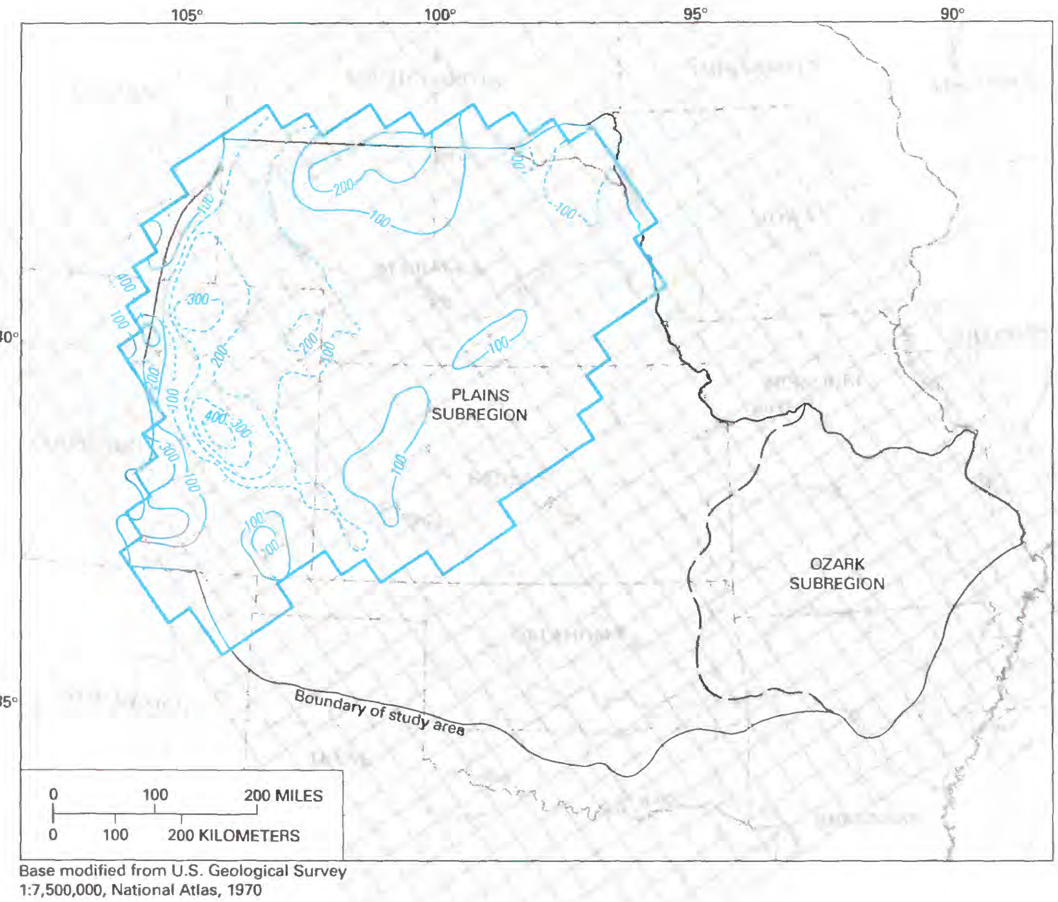
MODEL-LAYER BOUNDARY

2500 MODEL-COMPUTED HYDRAULIC-HEAD CONTOUR–Interval, in feet, is variable. Datum is sea level

2500 FIELD HYDRAULIC-HEAD CONTOUR–Interval, in feet, is variable. Datum is sea level

APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 52.—Model-computed and field hydraulic head for model layer 2 representing the Great Plains aquifer system (predevelopment steady-state conditions).



EXPLANATION

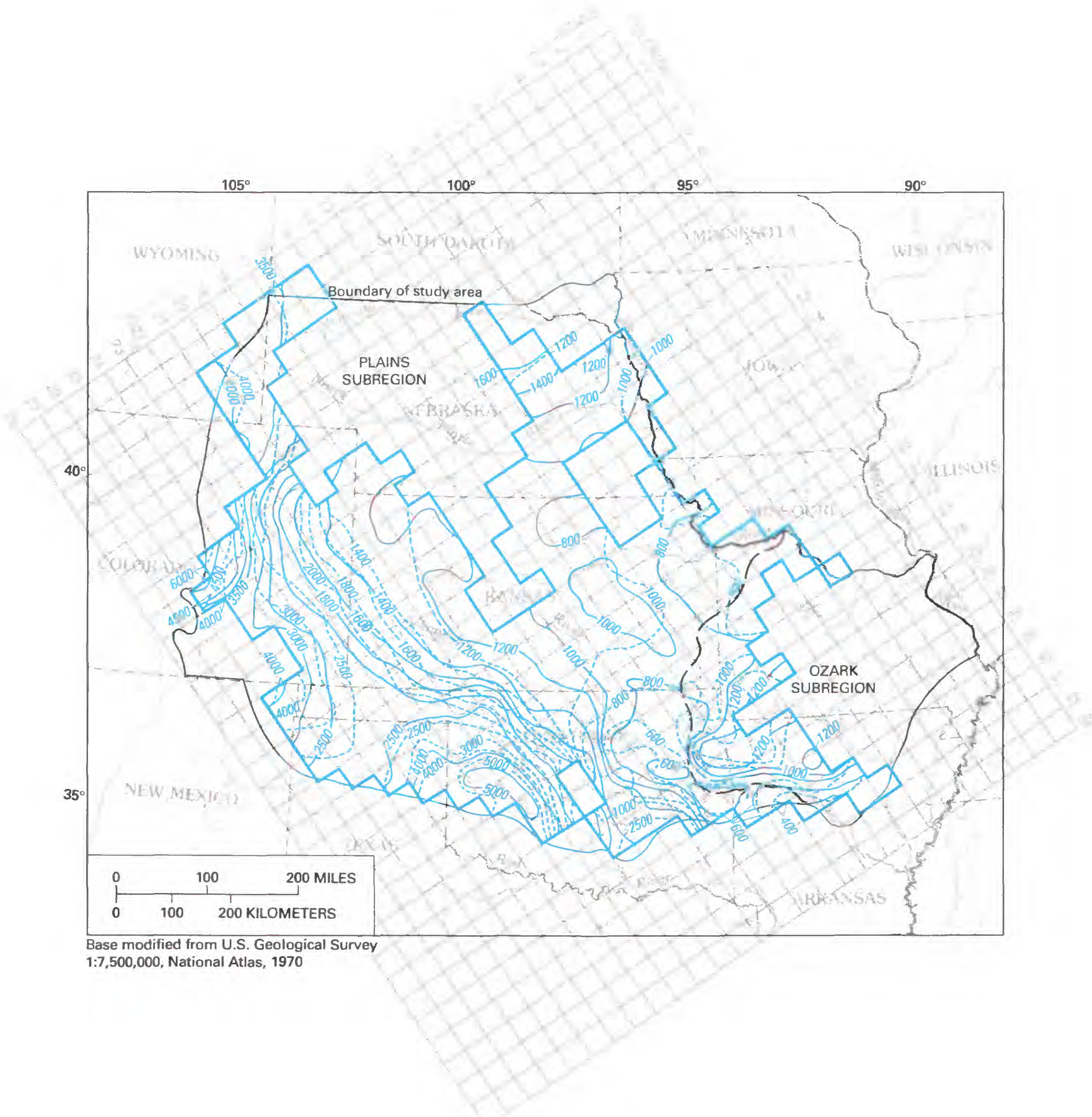
MODEL-LAYER BOUNDARY

—100— LINE OF EQUAL DIFFERENCE BETWEEN MODEL-COMPUTED AND FIELD HYDRAULIC HEAD (MODEL-COMPUTED HEAD IS BELOW FIELD HEAD)—Interval, in feet, is variable

---100--- LINE OF EQUAL DIFFERENCE BETWEEN MODEL-COMPUTED AND FIELD HYDRAULIC HEAD (MODEL-COMPUTED HEAD IS ABOVE FIELD HEAD)—Interval 100 feet

— — APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 53.—Difference between model-computed and field hydraulic head for model layer 2 representing the Great Plains aquifer system.



EXPLANATION

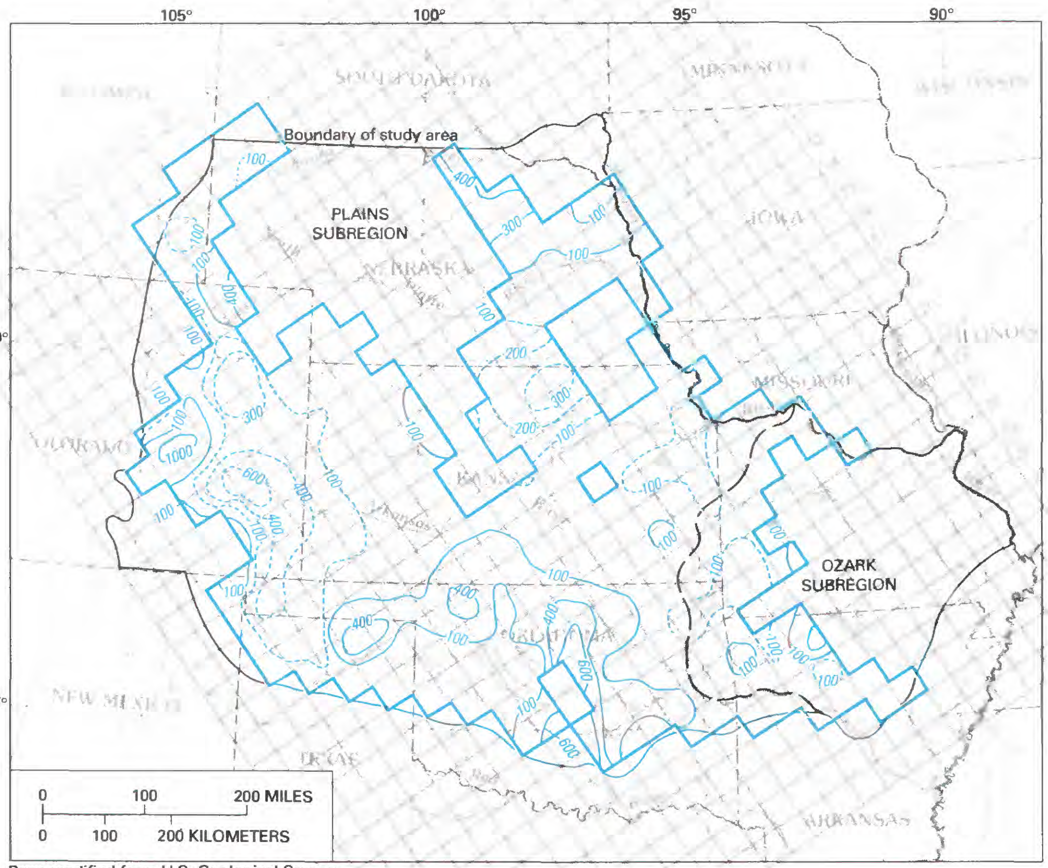
MODEL-LAYER BOUNDARY

---1000--- MODEL-COMPUTED HYDRAULIC-HEAD CONTOUR–Interval, in feet, is variable. Datum is sea level

—1000— FIELD HYDRAULIC-HEAD CONTOUR–Interval, in feet, is variable. Datum is sea level

— — APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 54.—Model-computed and field hydraulic head for model layer 4 representing the upper unit in the Western Interior Plains aquifer system and the Springfield Plateau aquifer in the Ozark Plateaus aquifer system (predevelopment steady-state conditions).



Base modified from U.S. Geological Survey
1:7,500,000, National Atlas, 1970

EXPLANATION

MODEL-LAYER BOUNDARY

LINE OF EQUAL DIFFERENCE BETWEEN MODEL-COMPUTED AND FIELD HYDRAULIC HEAD (MODEL-COMPUTED HEAD IS BELOW FIELD HEAD)—Interval, in feet, is variable

LINE OF EQUAL DIFFERENCE BETWEEN MODEL-COMPUTED AND FIELD HYDRAULIC HEAD (MODEL-COMPUTED HEAD IS ABOVE FIELD HEAD)—Interval, in feet, is variable

APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 55.—Difference between model-computed and field hydraulic head for model layer 4 representing the upper unit in the Western Interior Plains aquifer system and the Springfield Plateau aquifer in the Ozark Plateaus aquifer system.

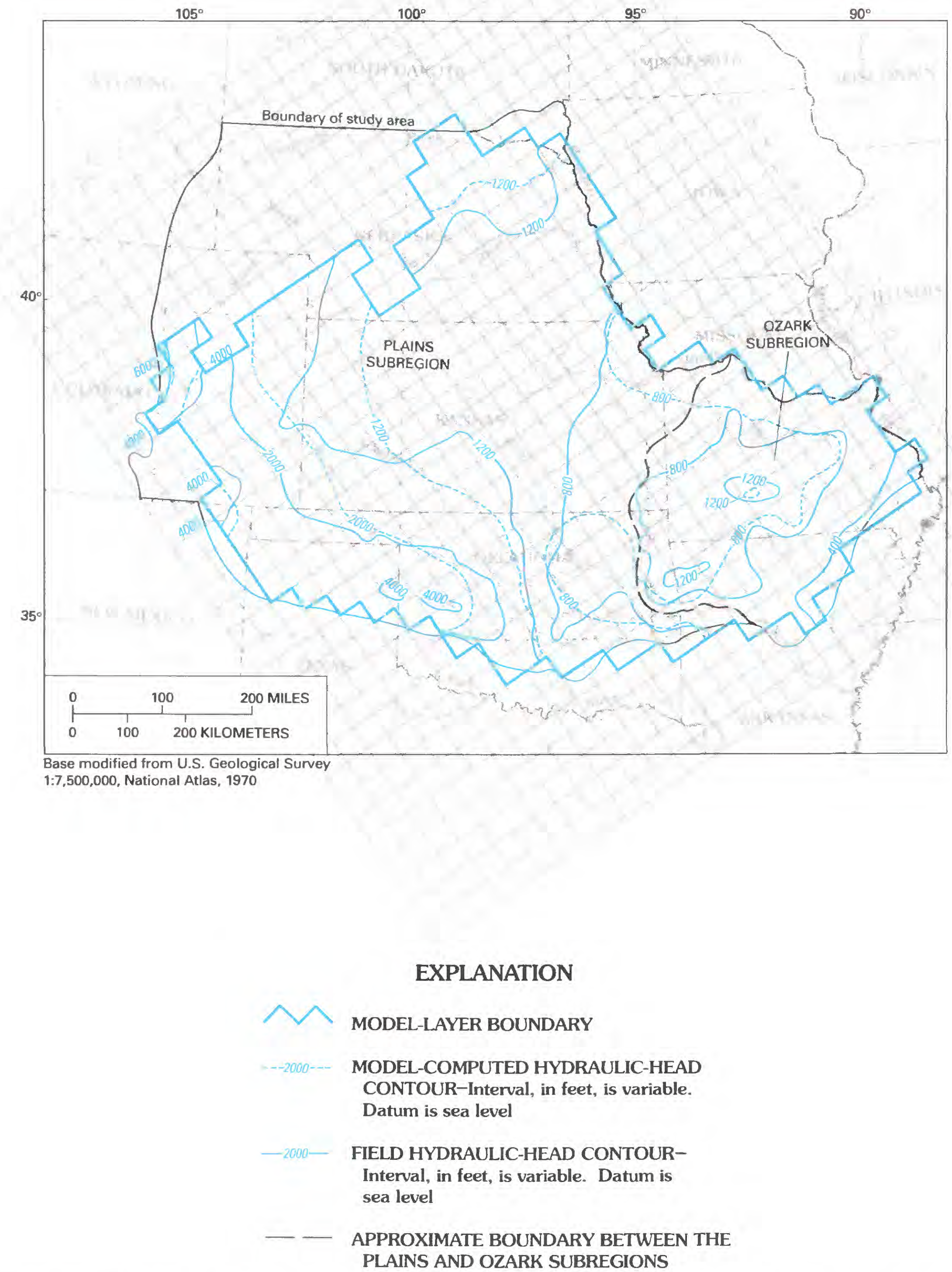
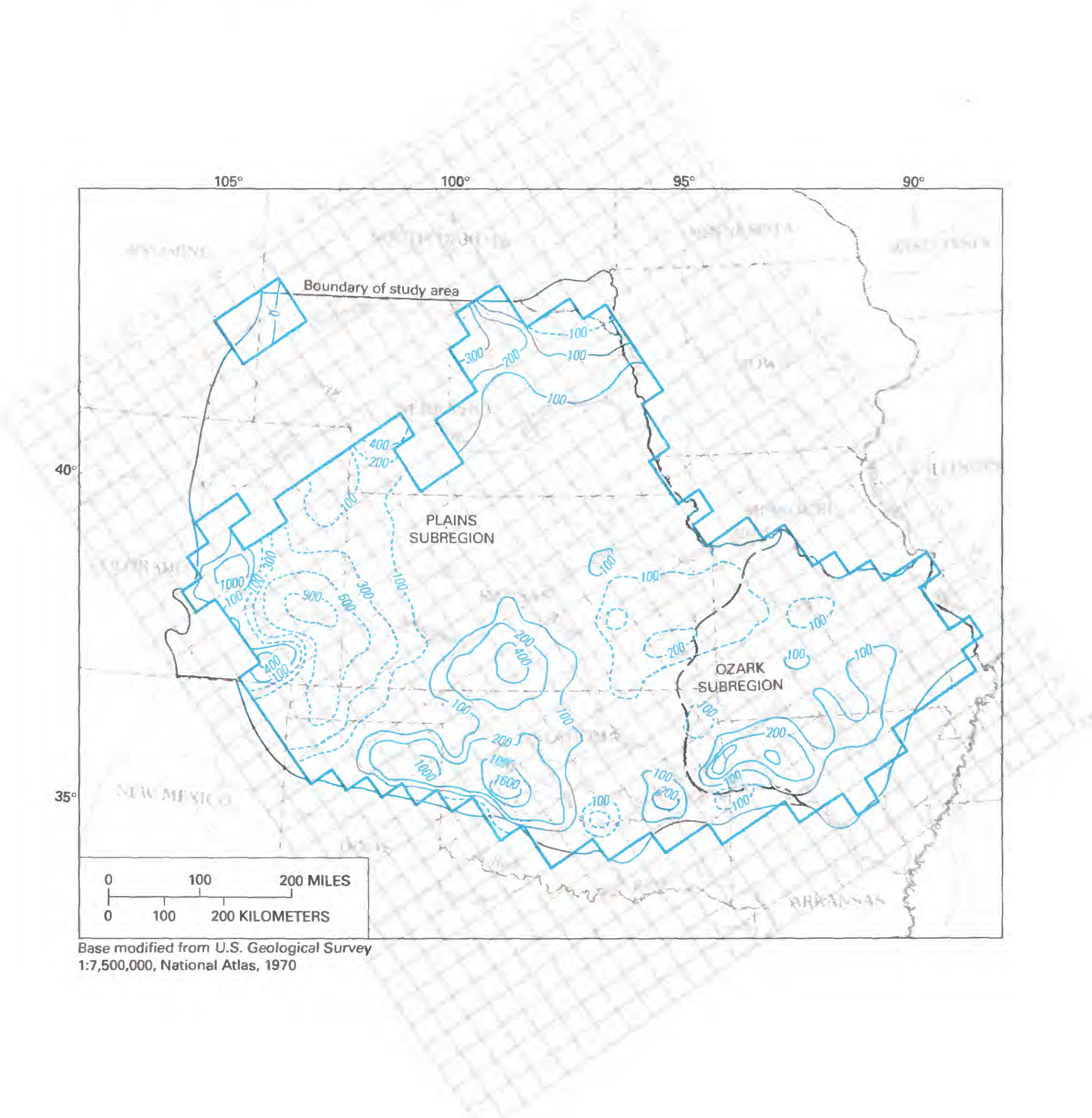


FIGURE 56.—Model computed and field hydraulic head for model layer 5 representing the lower units in the Western Interior Plains aquifer system and the Ozark and St. Francois in the Ozark Plateaus aquifer system (predevelopment steady-state conditions).



EXPLANATION

MODEL-LAYER BOUNDARY

—100— LINE OF EQUAL DIFFERENCE BETWEEN MODEL-COMPUTED AND FIELD HYDRAULIC HEAD (MODEL-COMPUTED HEAD IS BELOW FIELD HEAD)–Interval, in feet, is variable

---100--- LINE OF EQUAL DIFFERENCE BETWEEN MODEL-COMPUTED AND FIELD HYDRAULIC HEAD (MODEL-COMPUTED HEAD IS ABOVE FIELD HEAD)–Interval, in feet, is variable

— — APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 57.—Difference between model-computed and field hydraulic head for model layer 5 representing the lower units in the Western Interior Plains aquifer system and the Ozark and St. Francois aquifers in the Ozark Plateaus aquifer system.

lateral flow, which is not true for steeply dipping strata. Therefore, the poor match of model-computed hydraulic head to field hydraulic head along the Rocky Mountain Front Range is expected. The second area of smaller computed heads is in north-central Nebraska along the Nebraska-South Dakota State line. The model cells along the State line are specified also as no-flow boundaries because it was assumed that flow in the aquifer system is parallel or nearly parallel to the State line. The flow direction or path was inferred from the predevelopment potentiometric surface (fig. 34). Field hydraulic-head values indicate inflow from the outcrop area near the Black Hills Uplift at the extreme northwest corner of Nebraska, and the boundary conditions of the model specified that condition. However, the model-computed hydraulic-head values did not match the field head values. It is concluded that, along the western one-half of the Nebraska-South Dakota State line, inflow across the line may be adequate to maintain higher heads in layer 2 than those calculated by the model. Over the remaining area of layer 2, differences between field and model-computed head values are less than 100 ft. In general, it is concluded that the flow system was simulated accurately in regard to matching field hydraulic heads.

The average deviation of model-computed head values from field heads for layer 4 is about -0.8 ft (table 4). The standard deviation is slightly more than 234 ft, indicating that 95 percent of all model-computed head values are within 468 ft of the field head values. Field head values are matched well over most of Missouri, Kansas, and parts of Oklahoma and Nebraska, but model-computed values in large areas in Oklahoma, Colorado, and Nebraska differ from field values by two standard deviations or more (figs. 54, 55). In north- central Oklahoma, western Oklahoma, the Texas Panhandle, Colorado, and southwestern Nebraska, the computed-head values poorly matched the field values. However, the configuration of the two head distributions are similar (fig. 54). In central Oklahoma, a major discrepancy between computed and field head values occurs on the eastern flank of the Anadarko Basin, extending northward from an area where the aquifer material represented by layer 4 is not present. Model-computed head is less than field head by at least 600 ft in this vicinity. North of the Anadarko Basin south into the Palo Duro Basin of the Texas Panhandle and in northeastern Colorado and southwestern Nebraska, there are areas where the model-computed head is 400 ft less than the field head.

In the areas of west-central Oklahoma and the eastern Texas Panhandle peripheral to the Anadarko Basin, the field hydraulic-head values are based on data for an area in which geopressure exists; thus, the computed hydraulic-head values do not match for that reason. In the Anadarko Basin itself (west-central Oklahoma and the eastern Texas Panhandle), the model-computed heads match the field heads almost exactly because of the general-head boundary condition imposed on the model nodes in that locality. A match of head values using the general-head boundary condition indicates that a very small inflow or outflow will maintain the model-computed head at the field head. The small permeability (figs. 21, 29) of the geohydrologic units making up layer 4 in the Anadarko Basin causes very small flows to have a large effect on the hydraulic head because the units have only a slight capacity for transmitting water. Of the 16 cells in the Anadarko Basin area with a general-head boundary condition, 4 cells have flows exceeding $1x10^{-2}$ ft^3/s, with a maximum single node flow of $2.6x10^{-2}$ ft^3/s. The net flow of about $2x10^{-2}$ ft^3/s into the 16-cell area (about 12,600 mi^2) of the Anadarko Basin is negligible for that large an area; that small a flow rate is beyond the limits of accuracy for the model input.

Where the upper unit in the Western Interior Plains aquifer system, represented by model layer 4, pinches out because of the Cambridge Arch, Central Kansas Uplift, and Nemaha Uplift (figs. 7, 50), no-flow boundaries are specified in the model. In northeast Nebraska north and west of the Nemaha Uplift, the computed head values are as much as 400 ft less than the field head values (fig. 55). In the model area representing the Central Kansas Uplift, the model-computed head is as much as 300 ft higher than the field head. Field head values in northeastern Nebraska and north-central Kansas indicate fairly uniform hydraulic head with a head change from west to east of only about 100 ft for that area (fig. 54). Freeze and Witherspoon (1967, p. 634) conclude that buried stratigraphic pinch-outs can cause recharge or discharge areas where they would not be anticipated on the basis of water-table configuration. However, specifying general-head boundary conditions at the extent of layer 4 to represent the area surrounding the Nemaha Uplift would fix or control the head in almost 70 percent (31 of 46) of the cells in that area. Computed heads at nodes could readily be forced to fit the field heads to provide model-computed inflow or outflow data for the pinch-out boundaries. However, there are no measured flow data for comparison with model-computed flow rates.

Computed heads for layer 5 have an average deviation of 0.13 ft less than the field heads (table 4). The standard deviation is about 301 ft, the largest standard deviation of the three calibrated layers. The match of model-computed heads to field heads of the lower units in the Western Interior Plains aquifer system is fairly good over the regional study area, with the exception of the Anadarko Basin in west-central Oklahoma and the area in the eastern Panhandle of Texas and east-central Colorado. In those two locations, localized extremes of deviation occur (figs. 56, 57). For example, in the Anadarko Basin, computed hydraulic heads in two cells are more than 1,600 ft less than the field heads. However, at that location in the Anadarko Basin, a significant thickness of Devonian or Mississippian shale (Chattanooga Shale) is present. Because the simulated vertical leakance of the shale unit is related by a constant to the hydraulic conductivity of the underlying layer, the possibility arises that the hydraulic-conductivity values used in layer 5 for the cells representing the Anadarko Basin are too small. Similarly, the constant relating lateral hydraulic conductivity in layer 5 to the vertical hydraulic conductivity of the shale may be too small.

The effect of the Chattanooga Shale may be illustrated by another example. Where the Chattanooga Shale is present, the difference in computed hydraulic head between layers 4 and 5 in vertically adjacent model cells representing the aquifer in the Anadarko Basin area is almost 7,200 ft. The extreme field head difference at that location between layers 4 and 5 is 6,200 ft. A model-computed head difference of 7,200 ft is maintained primarily by the small vertical permeability of the intervening confining unit. However, in vertically adjacent cells of layers 4 and 5 in east- central Colorado, the computed heads are the same because of the absence of the Chattanooga Shale.

The model-computed head values (fig. 57) are smaller than the field head values in cells representing layer 5 in northwest Arkansas near the eastern State line of Oklahoma. This is an area where the Ozark confining unit is present between the Ozark and Springfield Plateau aquifers, and the Western Interior Plains confining system (layer 3) is also present. The area is a transition area where layer 4 crops out, and in laterally adjacent cells, layer 3 overlies layer 4. However, for one cell, layers 3, 4, and 5 are all present, but layer 5 crops out in a laterally adjacent cell. The model-computed and measured head values are well matched in layer 4 (fig. 54) but not matched for layer 5 (fig. 56). This mismatch may indicate that the leakance specified for the Ozark confining unit for that area is too small. That the leakance for the confining unit is too small is substantiated because the field head values in layer 4 are larger than those in layer 5 and more leakage between layers 4 and 5 would increase the hydraulic head in layer 5. Because the model- computed heads in outcrop areas of layer 5 are fairly well matched to field heads, the transmissivity and recharge values are considered sufficiently accurate, which further supports a conclusion that the Ozark confining unit may be more leaky than simulated.

Effects of Variable Density

Model computations show little difference between hydraulic head computed with and without specification of variable density for layer 2. The average deviation of computed head values is almost -6 ft (table 4). The standard deviation for the variable-density simulation is almost 2 ft less than the constant-density computation. The differences for layer 2 in the two sets of calculated hydraulic-head values are only slightly discernible on a plot of hydraulic head at the scales shown in figure 52. The differences in the two sets of head values is illustrated in figure 58, which shows that the effects of variable density occur only in northeastern Colorado, extreme southwestern Nebraska, and southeastern Wyoming. The model-computed heads for that area are generally smaller than the field hydraulic head, with a maximum difference of about 130 ft at a cell representing an area on the Front Range of the Rocky Mountains in Colorado. The area of layer 2 affected by variable density represents a deeply buried, thick section of the Great Plains aquifer system containing water with large concentrations of dissolved solids. Also, the permeability of the aquifers represented is slight (fig. 36), which amplifies the effect of even small values of variable density. For the remainder of layer 2, there are several reasons that variable density has little effect on the computation of hydraulic head. These are: (1) the lateral density differences represented by the data are small (fig. 37); (2) the density differences between the freshwater of layer 1 and layer 2 are small; (3) the Great Plains confining system above and the Western Interior Plains confining system below isolate a major part of the aquifer material represented by model layer 2 from the effect of vertical density differences; and (4) the specified-head boundary assigned everywhere in model layer 1 produces a general effect.

Model computation of hydraulic head for layer 4 that included the effects of variable density resulted in an average difference between computed hydraulic head and field hydraulic head of about 30 ft (table 4). Principal areas of hydraulic-head differences between values calculated with and without the effects of variable density are located in southeastern Kansas and eastern and central Oklahoma (fig. 59).

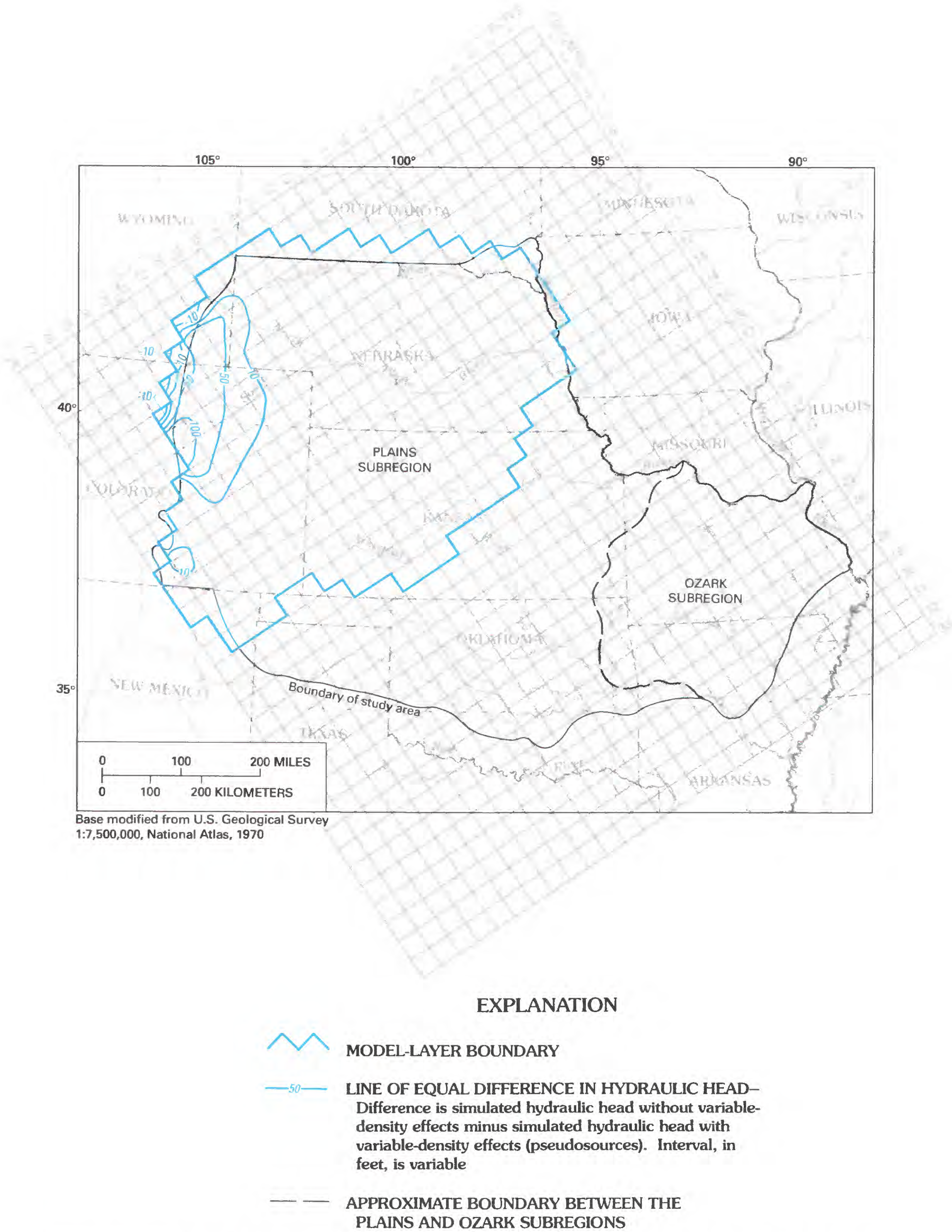


FIGURE 58.—Difference between simulated steady-state hydraulic head with and without effects of variable density (pseudosources) for model layer 2.

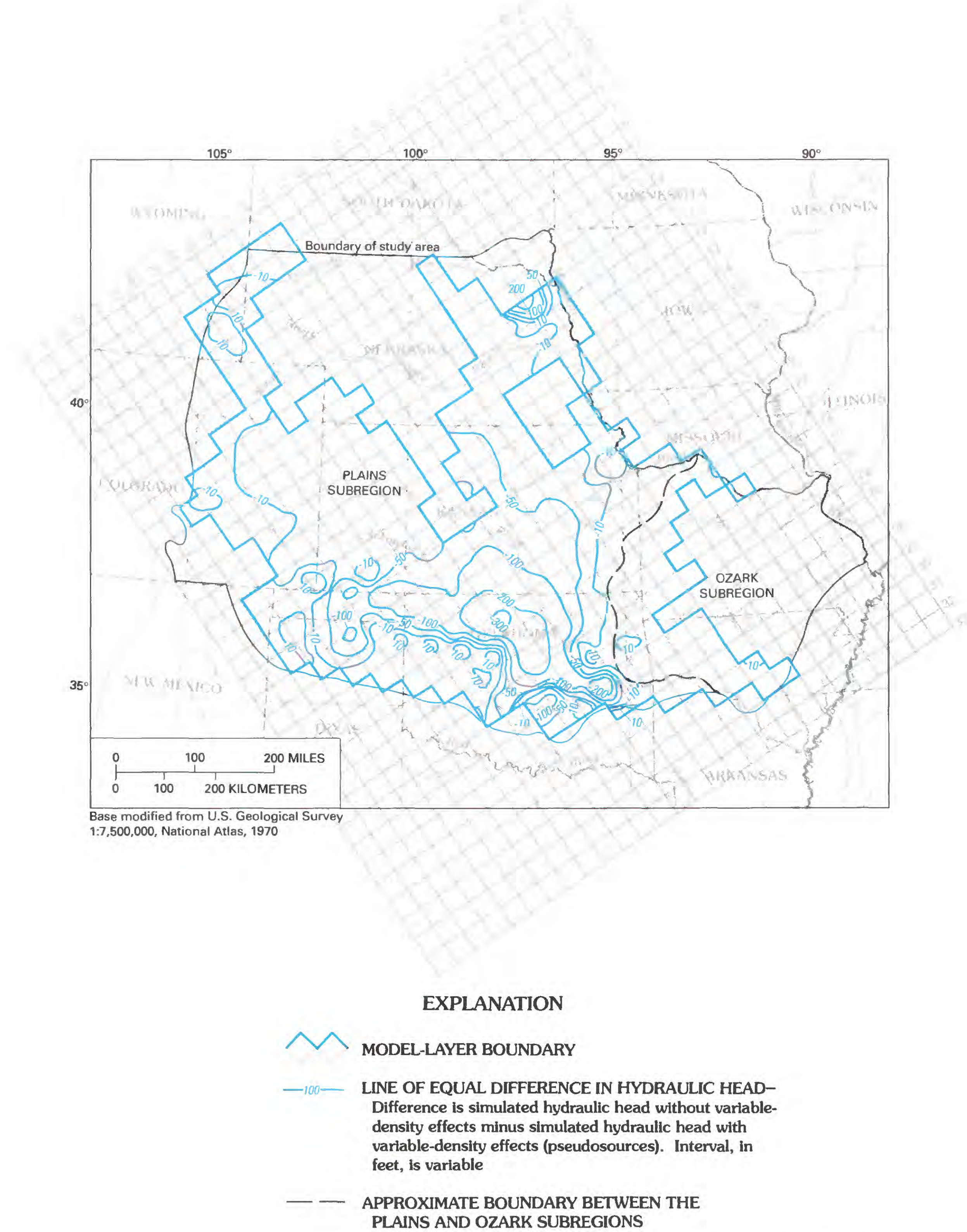


FIGURE 59.—Difference between simulated steady-state hydraulic head with and without effects of variable density (pseudosources) for model layer 4.

In eastern Oklahoma, the maximum head difference between computed and field hydraulic head exceeded 250 ft at a location where the flux due to variable density was not particularly large ($1.67\ ft^3/s$). However, the flux value was derived principally from lateral density differences in the freshwater-saltwater transition zone between the Ozark subregion and the Plains subregion. An example of the effect of slight permeability is in north-central Oklahoma where model layer 4 represents an aquifer in the Anadarko Basin. A calculated hydraulic head for one cell of the model affected by variable density exceeded the hydraulic head calculated without this effect by slightly more than 300 ft. This hydraulic-head difference due to variable density is greater than that in eastern Oklahoma, but it is a hydraulic-head difference that resulted from a much smaller variable-density effect represented by a pseudosource ($2.1x10^{-3}\ ft^3/s$). Although the variable-density effects in the Anadarko Basin area are represented by very small computed flux values, the resulting computed hydraulic-head values for the Anadarko Basin area were a major contribution to the greater-than-average difference between model-computed and field hydraulic head than was the case with computed hydraulic head that did not include the effects of variable density.

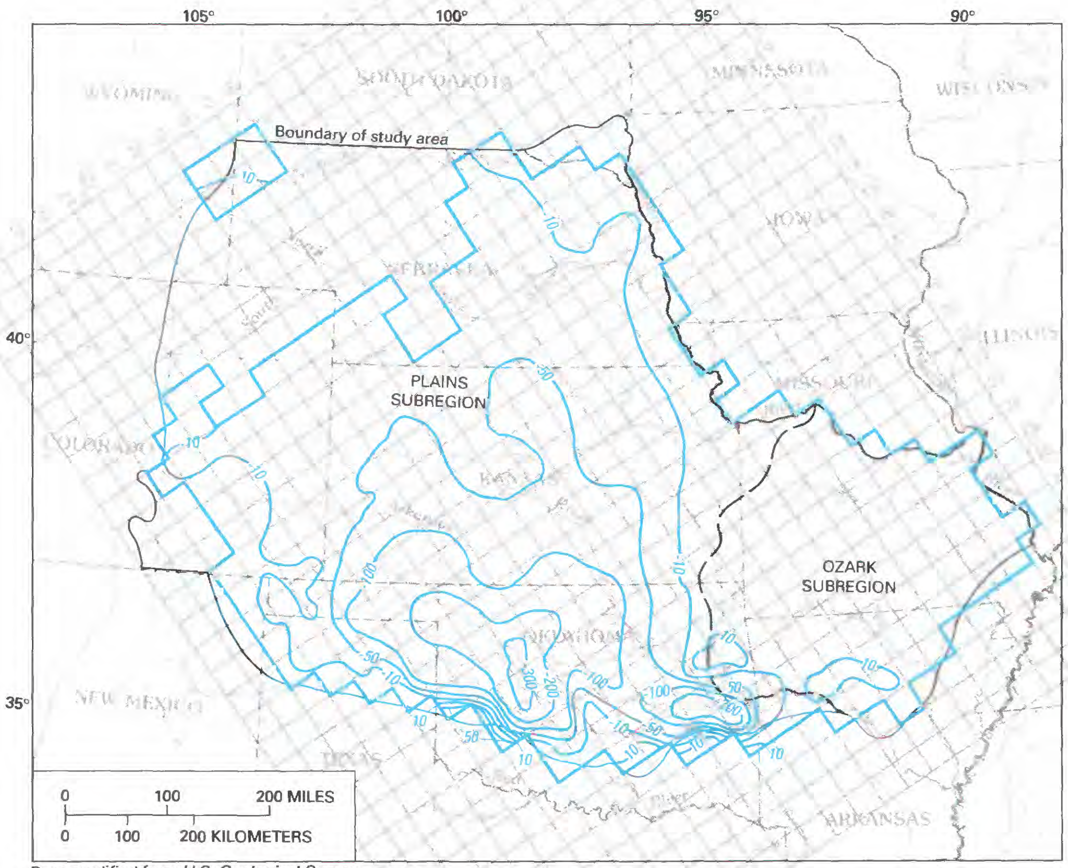
The average difference between the computed and field hydraulic-head values for layer 5 (when the model included the effects of variable density) is about 41 ft (table 4). Hydraulic-head differences between values calculated with and without the effects of variable density are shown in figure 60. Hydraulic-head differences for layer 5 result predominantly from lateral density differences in extreme southeastern Kansas, eastern Oklahoma, and almost all of the study area in northwestern Arkansas. Fluxes of less than $1.0\ ft^3/s$ that result from vertical density differences, although small, cause fairly large differences in calculated hydraulic head. The large hydraulic-head differences between the two calculation techniques not only result from the effect of variable density but also from the effect of extremely small permeability. The calculated hydraulic head also can be smaller than that calculated without consideration of variable density, which was the result in model layer 5 representing an area of an aquifer in the Anadarko Basin in west-central Oklahoma.

The areal configuration of computed hydraulic head for the study with consideration of variable density differs little from the configuration of hydraulic head computed without considering variable density, except in areas of very small hydraulic conductivity. Computation of hydraulic head for the two conditions, however, result in some "local" hydraulic- head differences that would cause different lateral flow directions for the two conditions. Flow vectors show no direction differences for model layer 2, but in layers 4 and 5 there are changes illustrated by figures 61 and 62. For layer 4 (fig. 61), simulation of the area between the Nemaha Uplift and the Central Kansas Uplift (fig. 7) and the area southeast of the Central Kansas Uplift toward eastern Oklahoma exhibits flow-direction changes. For layer 5 (fig. 62), flow-direction changes are limited to cells representing the area in the Central Kansas Uplift and extreme eastern Oklahoma at the freshwater-saltwater transition zone. The only flow-direction changes that would indicate a significant change in the overall flow regime are those in layer 5 over the Central Kansas Uplift. Flow-direction reversals imply a compensating vertical-flow change. For the flow simulation without consideration of variable density, the majority of vertical-flow vectors in western Kansas and eastern Colorado west of the Central Kansas Uplift indicated flow from layer 4 to layer 5. However, considering variable density by introducing pseudosources, the majority of vertical-flow vectors indicated flow from model layer 5 to layer 4 in that same area.

Evaluating the effect of variable density by the introduction of pseudosources (Weiss, 1982) in the regional model clearly indicated that not including variable density and only using freshwater head in computation may lead to conclusions different from those when variable density is considered regarding flow and possibly erroneous conclusions as postulated by Jorgensen and others (1982). Even with the large cell size of the regional model, both lateral and vertical flow-direction changes were indicated in some areas, which supports the necessity of considering variable density.

Recharge

A recharge model for the regional study area was developed by Dugan and Peckenpaugh (1985). Inputs to the recharge model were data on soil permeability, available water capacity of the soil, topography, consumptive water requirements of various vegetation types, monthly precipitation, and computed monthly potential evapotranspiration (PET). The PET values were obtained using the Jensen-Haise method (Jensen and Haise, 1963). Recharge was computed using type vegetation environments for the study area during a 30-year period, 1951–80. The first environment was a representation of vegetation conditions during the period based on 1978 vegetation statistics. The second vegetation environment for the period was termed



Base modified from U.S. Geological Survey 1:7,500,000, National Atlas, 1970

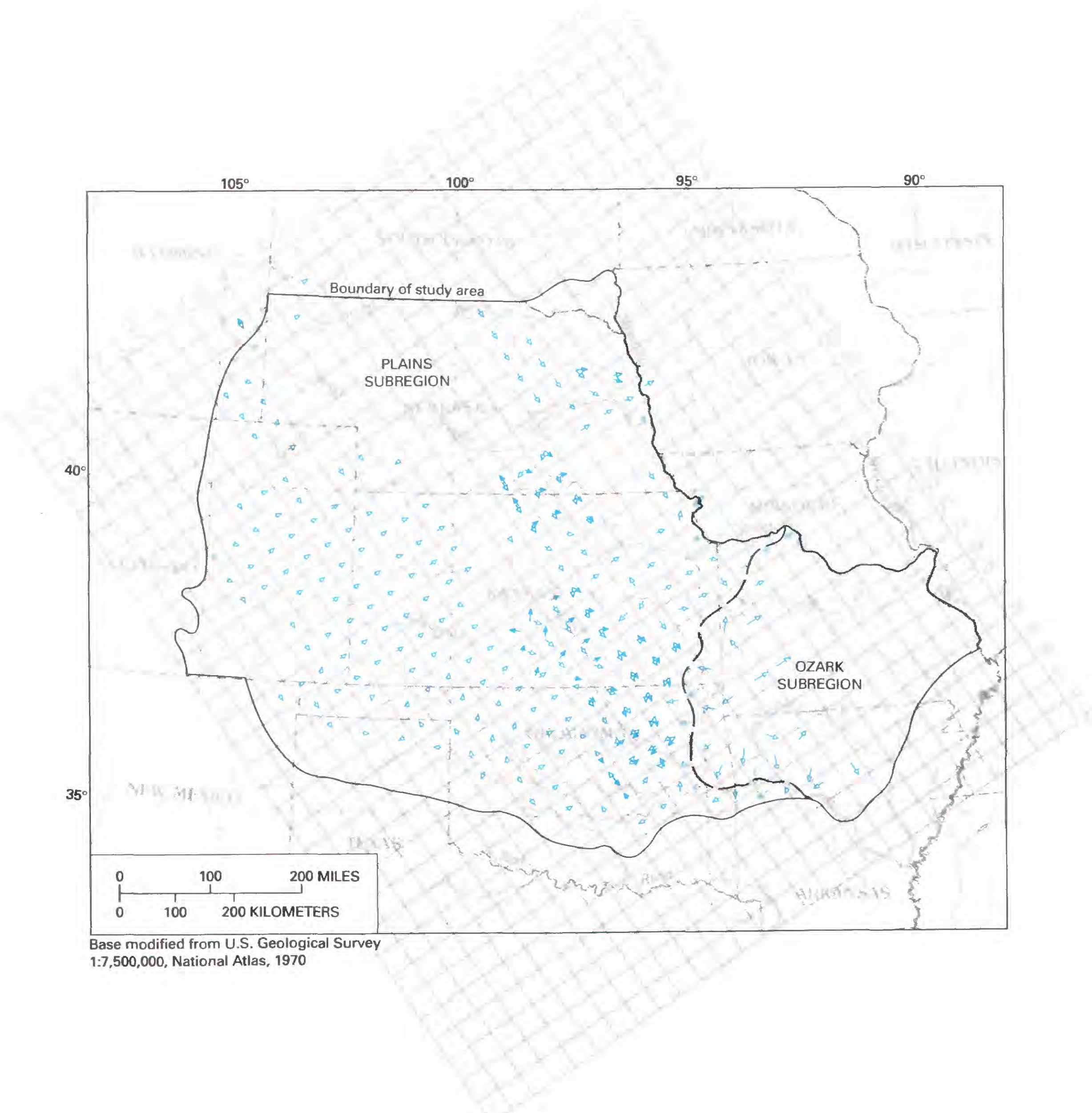
EXPLANATION

MODEL-LAYER BOUNDARY

—10— LINE OF EQUAL DIFFERENCE IN HYDRAULIC HEAD– Difference is simulated hydraulic head without variable-density effects minus simulated hydraulic head with variable-density effects (pseudosources). Interval, in feet, is variable

— — APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 60.—Difference between simulated steady-state hydraulic head with and without effects of variable density (pseudosources) for model layer 5.



EXPLANATION

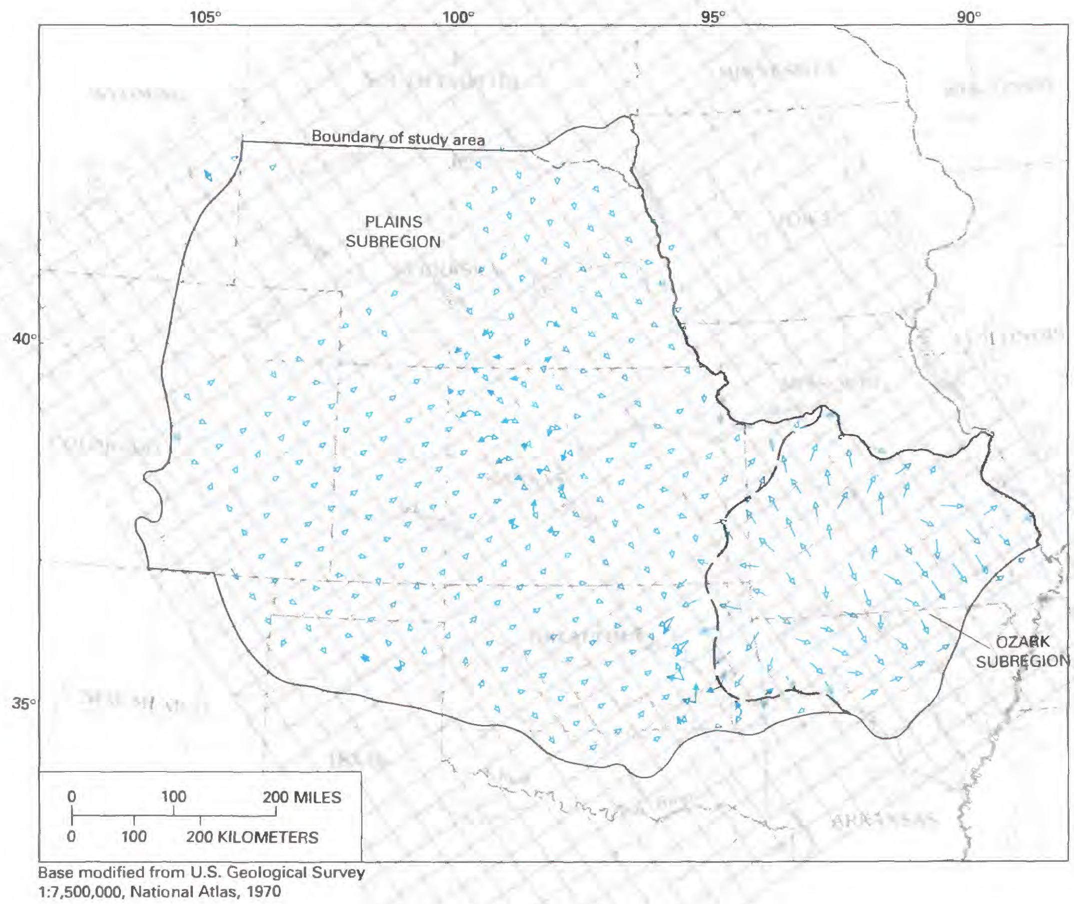
LATERAL FLOW VECTORS–Length of shaft indicates relative magnitude

Variable density

Constant density

APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 61.—Effects of variable density on simulated flow vectors for model layer 4.



EXPLANATION

LATERAL FLOW VECTORS–Length of shaft indicates relative magnitude

Variable density

Constant density

APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 62.—Effects of variable density on simulated flow vectors for model layer 5.

"pre-agricultural" (Dugan and Peckenpaugh, 1985, p. 68), with woodland the same as in 1978 but with cultivated land or cropland treated as grassland. The results show that recharge differences between the two vegetation environments are not significant except in localized areas.

Recharge patterns within the region indicate that climate, particularly precipitation, is a controlling factor. Data show that the proportion of precipitation that potentially may be recharged decreases as precipitation decreases. A scatter diagram of computed mean annual recharge versus precipitation for 356 regional model-grid cells shows a close relation between precipitation and recharge (fig. 63). The relation becomes approximately linear where mean annual precipitation exceeds 35 in. and recharge exceeds 5 in/yr. The spatial relation of recharge to rainfall is illustrated by figures 5 and 6. These figures confirm the similar patterns for rainfall and recharge, and figure 6 shows a maximum of about 15 in. of annual recharge in central Arkansas where mean annual precipitation is about 48 in. (fig. 5). In Colorado and New Mexico, the recharge model predicted extremely small recharge. Dugan and Peckenpaugh (1985, p. 72) conclude that the small recharge is closely related to large PET, solar radiation, percentage of possible sunshine, and small relative humidity.

Outcrops in the Ozark subregion are the principal recharge areas simulated in the regional flow model. In modeling extensive aquifers, it has been observed that the effective amount of flow to the water table required for a successful simulation of observed water levels may be significantly less than that calculated from independent data sources. For example, Jorgensen and others (1989b, p. 683) use Dugan and Peckenpaugh's (1985) data to estimate that for the Ozark subregion, the annual precipitation is approximately 190,000 ft^3/s, and from that, potential recharge to the water table in the aquifer system of the Ozark subregion is about 45,000 ft^3/s.

Ground-water flow discharging from aquifers to 10 major streams in southern Missouri, which likely is only part of the discharge in the Ozark subregion, was reported as 4,169 ft^3/s (Harvey, 1980, p. 50, table 4) (that does not include discharge to the Missouri or Mississippi Rivers). Therefore, one can conclude that recharge probably exceeds 10 percent and may be as much as 20 percent of the potential recharge. Thus, recharge was expected to exceed 4,500 ft^3/s and be 9,000 ft^3/s or more for the Ozark subregion in this study.

Using the technique developed by Jorgensen and others (1989a) for determining the net amount of recharge within a model cell that leaves the cell as

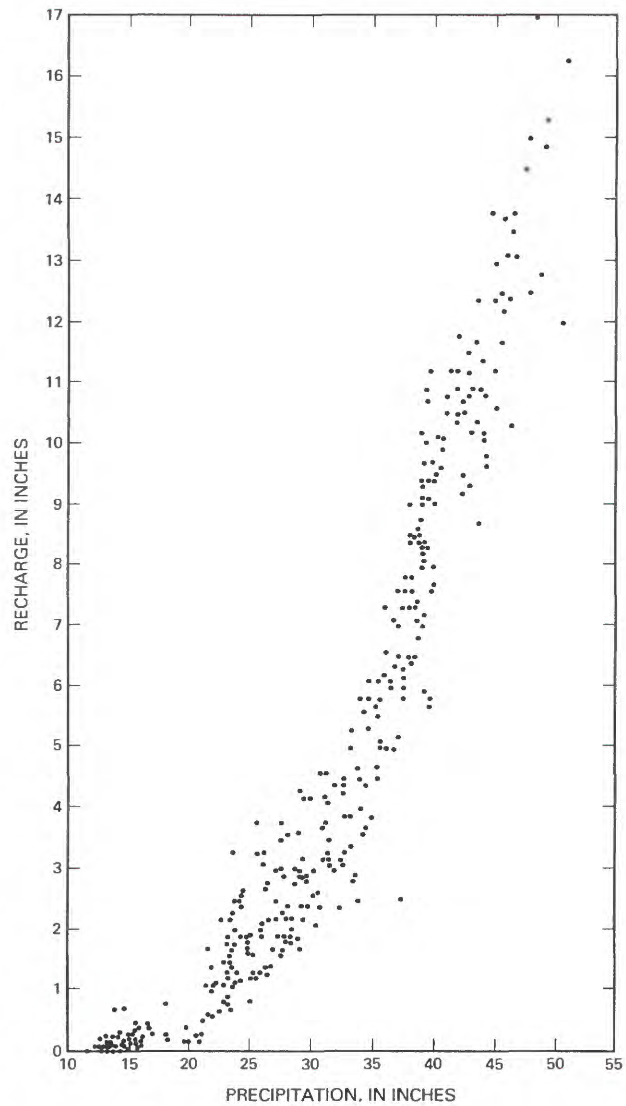


FIGURE 63.—Computed mean annual recharge versus mean annual precipitation, by model-grid element (from Dugan and Peckenpaugh, 1985, fig. 35, p. 73).

regional flow, the minimum net recharge to regional aquifer units in the Ozark subregion was determined to be only about 2,100 ft^3/s. Thus, the net recharge was computed to be about 5 percent of the potential recharge for a three-layer model of the Ozark Plateaus aquifer system. The model-grid spacing was 14 by 14 mi. An advantage of the technique is that the recharge is distributed in the model cells on a rational basis of topographic effects on runoff and stream-aquifer interaction. The 2,100 ft^3/s (the minimum) represents recharge that cannot (and thus, has no potential to) be intercepted by sinks within the model cells but would move to the aquifer as part of net recharge. The remaining recharge, in general, is intercepted and carried away in streamflow. However, there is part of recharge that does have the potential to be intercepted, yet it still may become a part of the net recharge and enter the regional aquifer flow system. In the recharge analysis by Jorgensen and others (1989a, b), the part of actual recharge that becomes net recharge instead of being intercepted by streams was varied by applying what is termed herein an "interception factor." The interception factor defines that part of total water-table recharge within a model-cell area that exits the cell and thus is not a part of the simulated aquifer flow. The factor is unknown and, initially, can only be estimated. Its maximum value would be 1.

Values for recharge to the water table in outcrop areas modeled in the region are listed in table 5. The values were derived from recharge modeling by Dugan and Peckenpaugh (1985) for the whole region and the analyses of Jorgensen and others (1989a,b) applied to the Ozark subregion. The net flow shown for the outcrop areas modeled is water-table recharge minus local aquifer discharge to streams. The data imply that although the recharge through the outcrops of the Ozark and St. Francois aquifers (represented by model layer 5) are always more than twice the recharge for the Springfield Plateau aquifer (represented by model layer 4), the net flow amounts are similar when the interception factor is less than 1.

The analysis gives a recharge budget (table 5) for the regional study outcrop areas for field purposes ranging between 7,399 and 10,135 ft^3/s. For the Ozark subregion, independently determined aquifer-stream interaction outflow for the Springfield Plateau, Ozark, and St. Francois aquifers (sum of the outflows from layer 4 and 5 only) is 3,761 ft^3/s, giving a net flow range for calibration purposes of 3,536 to 6,223 ft^3/s. Note that an interception factor of 1 giving a recharge rate of 1,836 ft^3/s was not believed to provide a field recharge value but does provide a minimum possible value. Net flow to the water table through outcrops for aquifers simulated by layer 5 is negative for an interception factor of 1 (table 5). For that case, local aquifer discharge to streams exceeds recharge to the water table. This condition could exist only if ground-water discharge would be supplied from a remote source, which is not the hydrologic case for the system.

The recharge budget that gave the best simulation based on matching computed hydraulic head to field hydraulic head is given also in table 5. The total recharge for the region as a model input is 6,947 ft^3/s. The upper bracket of the recharge calibration values,

TABLE 5.—*Simulated recharge to outcrops of the regional model area represented by model layers 2, 4, and 5 for several interception factors*

Model layer	Interception factor[1]	Recharge (cubic feet per second)	Aquifer-stream interaction outflow[2] (cubic feet per second)	Net flow (cubic feet per second)
2	1.00	0	80	-80
4		499	446	43
5		1,337	3,315	-1,978
Total		**1,836**	**3,841**	**-2,005**
2	0.85	102	80	22
4		2,227	446	1,781
5		5,070	3,315	1,755
Total		**7,399**	**3,841**	**3,558**
2	0.78	151	80	71
4		3,077	446	2,631
5		6,907	3,315	3,592
Total		**10,135**	**3,841**	**6,294**
		Final Simulation Values		
2	0.86	93	80	13
4		2,087	446	1,641
5		4,767	3,315	1,452
Total		**6,947**	**3,841**	**3,106**

[1]The interception factor defines that part of total recharge that exits a cell and does not become recharge that is simulated in the model.

[2]Aquifer-stream interaction data for layer 2 from Helgesen and others (1993). For layers 4 and 5, data are from Hedman and others (1987) and the analysis of Jorgensen and others (1989a, b).

10,135 ft^3/s, obtained by using an interception factor of 0.78, exceeded the best model simulation value by 46 percent. Use of the upper-bracket recharge value gave such extremely large model-computed hydraulic heads that unrealistic transmissivity adjustments would have been necessary to simulate field hydraulic heads in the Ozark subregion. The lower bracket of the calibration recharge also gave model-computed heads greater than the field hydraulic heads, but only minor adjustments of recharge values were necessary to obtain a reasonable match between computed and field head values in the Ozark subregion.

In the central and western part of the regional study, outcrop areas of the Great Plains aquifer system represented by layer 2 generally are more scattered. They were not subjected to the net-flow-per-cell analysis, but modified recharge data of Dugan and Peckenpaugh (1985) were entered directly into the model. Total recharge to layer 2 of the model (table 5) is about 1 percent of the total for the region or 93 ft^3/s, of which 83 ft^3/s occurred in cells representing areas in north-central Kansas and 10 ft^3/s through the remaining areas in Oklahoma, Colorado, New Mexico, Wyoming, and Nebraska. In the western part of the study area, recharge rates were generally less than 1 ft^3/s per model cell, values consistent with the recharge-precipitation relation shown in figure 63.

SENSITIVITY ANALYSIS

The effect on model results of systematically varying data entered into the model illustrates the relative importance of the data to the calibration criteria of the model. The primary calibration criterion of the regional study was the closeness of fit of model-computed to field hydraulic head. A measure of the closeness of fit is the mean head difference between computed and field head values. Luckey and others (1984, p. 48) point out that the mean hydraulic-head difference (water-level residual) is only one of several possible field criteria. A small mean hydraulic-head difference could be obtained from a model simulation as a result of a balance of errors, but the possibility exists that compensating errors cannot be identified. The mean water-level residual would give no indication of the magnitude of those errors, and even though the field criterion was met, the model may not be calibrated properly. The standard deviation of the differences between the model-computed head and the field head is a measure of hydraulic-head-difference variability that is not affected by compensating errors. Also, the standard deviation of the differences emphasizes the larger differences (Luckey and others, 1984, p. 48).

A sensitivity analysis was performed on the CMRASA regional model to evaluate the changes in the mean and standard deviation of the hydraulic-head differences that would result from general changes in model inputs. Hydraulic conductivity, leakance between layers, and recharge were changed over a range of values after first determining a calibration in which the average difference of model-computed head from field head was less than 1 ft. The input data were changed by using a single-value multiplier; thus, areal distribution of the data was not affected. The sensitivity of model-computed hydraulic-head values to a general change of model inputs is shown by the changes in mean hydraulic-head differences and standard deviations, figures 64, 65, 66.

Model-computed head is quite sensitive to changes in aquifer hydraulic conductivity (fig. 64). Figure 64 shows that computed hydraulic head is much more sensitive to decreases in hydraulic conductivity than to increases in hydraulic conductivity.

The mean difference (between simulated and field head) and standard deviation of the hydraulic-head difference are not minimal at the same hydraulic conductivity for model layer 5. The standard deviation, as a sensitivity indicator, shows that computed hydraulic head becomes very sensitive primarily at hydraulic-conductivity reductions from calibration hydraulic conductivity of 40 percent or more.

Standard deviations for model layers 2 and 4 are near minimum at the field hydraulic conductivity and are fairly insensitive to hydraulic-conductivity increases from that point. However, the standard deviations are sensitive to hydraulic-conductivity reductions of 40 percent or more as in the case of model layer 5.

Computed hydraulic-head values in model layers 4 and 5 are not very sensitive to changes in leakance coefficients (fig. 65). The principal reasons that the variation of leakance coefficient does not greatly affect computed hydraulic head in model layers 4 and 5 are the simulations of evaporite deposits in model layer 3 as a restriction to vertical flow, the large outcrop area simulated by model layer 5, and the relatively small differences in hydraulic head between layers 4 and 5.

Layer 2 is very sensitive to leakage from layer 1 (fig. 65). Increasing the leakance coefficient causes more vertical inflow than can be dissipated laterally without an excessive increase of hydraulic head. This situation occurs partly because model layer 3 represents a zero-leakage layer where evaporites are present in an area coinciding with most of layer 2. Conversely, the computed hydraulic-head values are smaller when

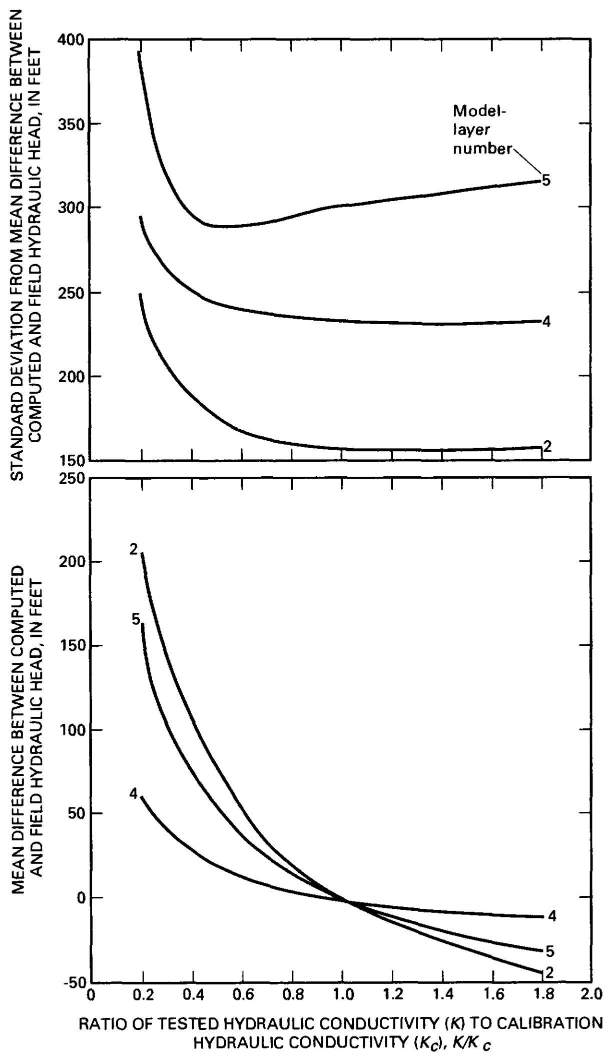


FIGURE 64.—Sensitivity of difference between model-computed and field hydraulic head to change in aquifer hydraulic conductivity in CMRASA geohydrologic units.

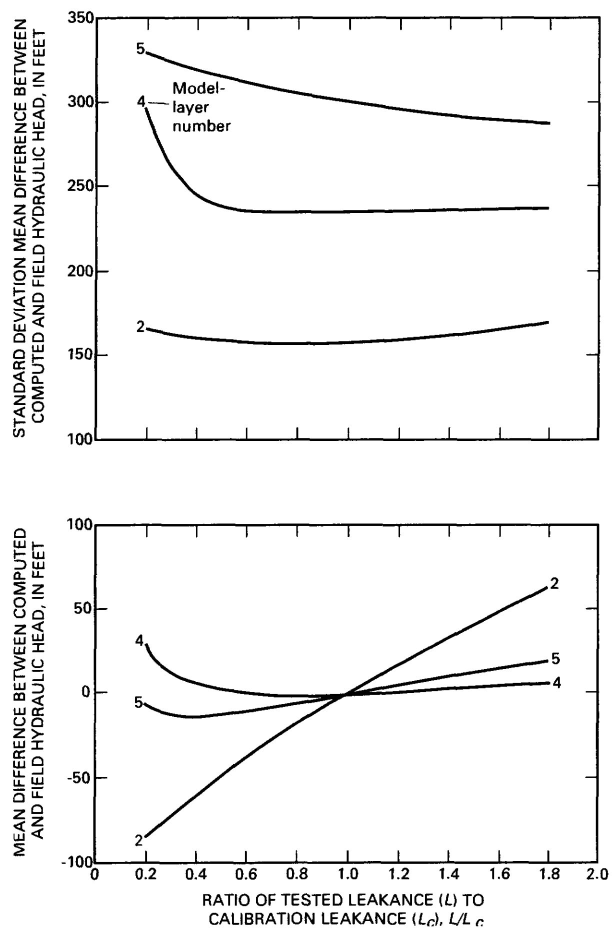


FIGURE 65.—Sensitivity of difference between model-computed and field hydraulic head to change in leakance coefficients between layers in CMRASA geohydrologic units.

the leakance coefficient, and thus inflow, are decreased. The sensitivity of computed hydraulic head to increased or decreased leakage may be termed typical for a confined aquifer.

Model-computed hydraulic-head values are only moderately sensitive to changes in recharge for the regional area (fig. 66). Because constant-flow and general-head boundaries were specified in layer 2 in all recharge areas (fig. 48), varying recharge had no effect on hydraulic-head computation for that layer. There were major changes in computed hydraulic head, however, in the Ozark subregion where recharge was applied to cells in layers 4 and 5 that represent outcrops of the Ozark Plateaus aquifer system. Among cells of layers 4 and 5 representing confined parts of the aquifer system, only those laterally adjacent to cells representing outcrop areas were affected. Therefore, the mean differences between computed and field hydraulic head were small because differences over the whole area were included when computing the mean.

Mean differences in the Ozark subregion alone are much greater (fig. 67). In the Ozark subregion, for values of recharge used in the regional-model field simulation (where the ratio of tested recharge rate to field recharge rate is 1.0), the mean difference between computed and field hydraulic head is almost 23 ft greater than the value for layer 4, representing the whole regional area, and almost 10 ft greater than the value for layer 5 (fig. 67). To obtain the mean difference between computed and field hydraulic head of less than 1 ft in the regional model, the mean differences for the Ozark subregion are cancelled by a negative mean difference for the remainder of the regional area.

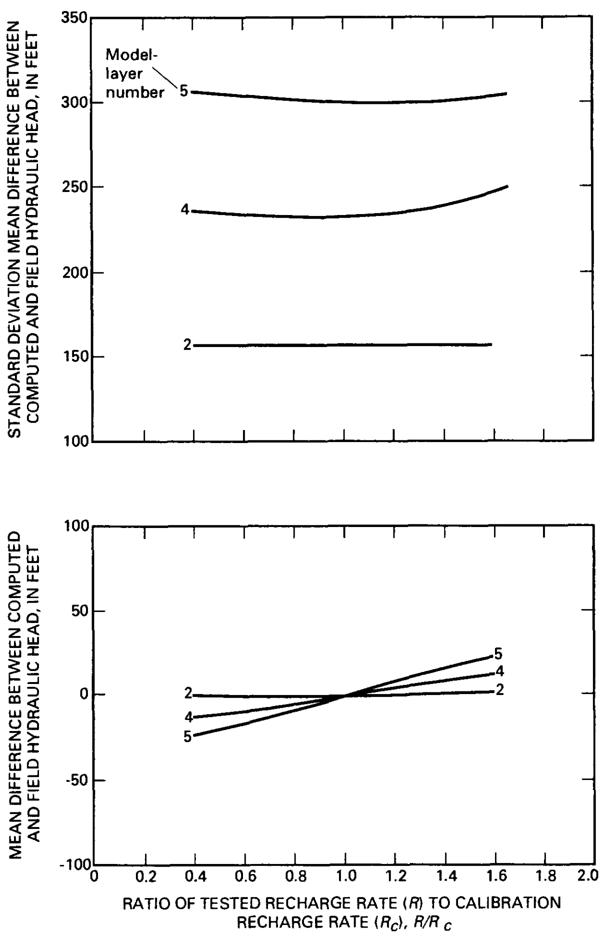


FIGURE 66.—Sensitivity of difference between model-computed and field hydraulic head to change in recharge rate in CMRASA geohydrologic units.

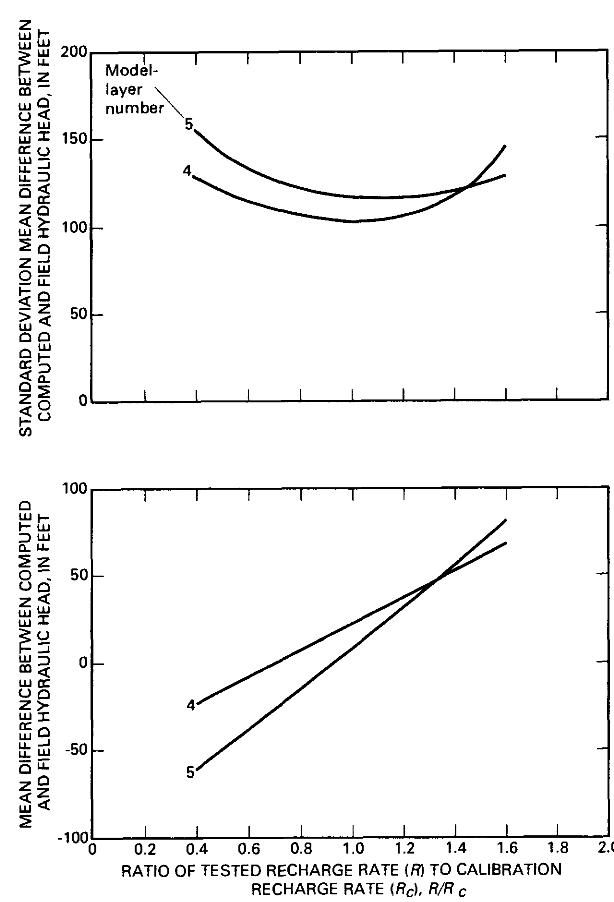


FIGURE 67.—Sensitivity of difference between model-computed and field hydraulic head to change in recharge rate in CMRASA hydrologic units, Ozark subregion.

Varying two model inputs simultaneously and plotting the mean difference between computed and field hydraulic head illustrates the nonunique characteristic of a model calibration. Varying the hydraulic conductivity and recharge simultaneously shows an almost linear effect on the difference between model-computed and field hydraulic head for layers 4 and 5. Mean differences of less than 1 ft result for recharge and hydraulic-conductivity ratios ranging from about 0.4 and 0.6, respectively, to about 1.4 and 1.6, respectively (figs. 68, 69).

Varying recharge and leakance simultaneously showed that two values of leakance would give zero mean differences between computed and field hydraulic head for different values of recharge (figs. 70, 71). For layer 4, the relation is almost symmetrical. A similar relation exists for the two model inputs in layer 5 (fig. 71), but the results are not as symmetrical as for layer 4.

The large outcrop area simulated by cells in layer 5 affects the leakance-recharge relation because leakance would not occur in the cells representing the outcrop area of the Ozark and St. Francois aquifers in the Ozark subregion. Therefore, leakance would have less effect on mean differences between computed and field hydraulic head; however, recharge would have a greater effect.

Summarizing the relative importance of model-input data to field data, the sensitivity analysis curves (figs. 64–66) show that changes in hydraulic-conductivity values from those used for the calibration resulted in the greatest change in model-computed values of hydraulic head. Changes of leakance coefficient resulted in significant changes in the mean difference between computed and field values of hydraulic head for model layer 2 and standard deviation from the mean for model layer 4. However, the leakance coefficient was generally less important than changes in hydraulic conductivity. Differences between computed and field hydraulic head were least

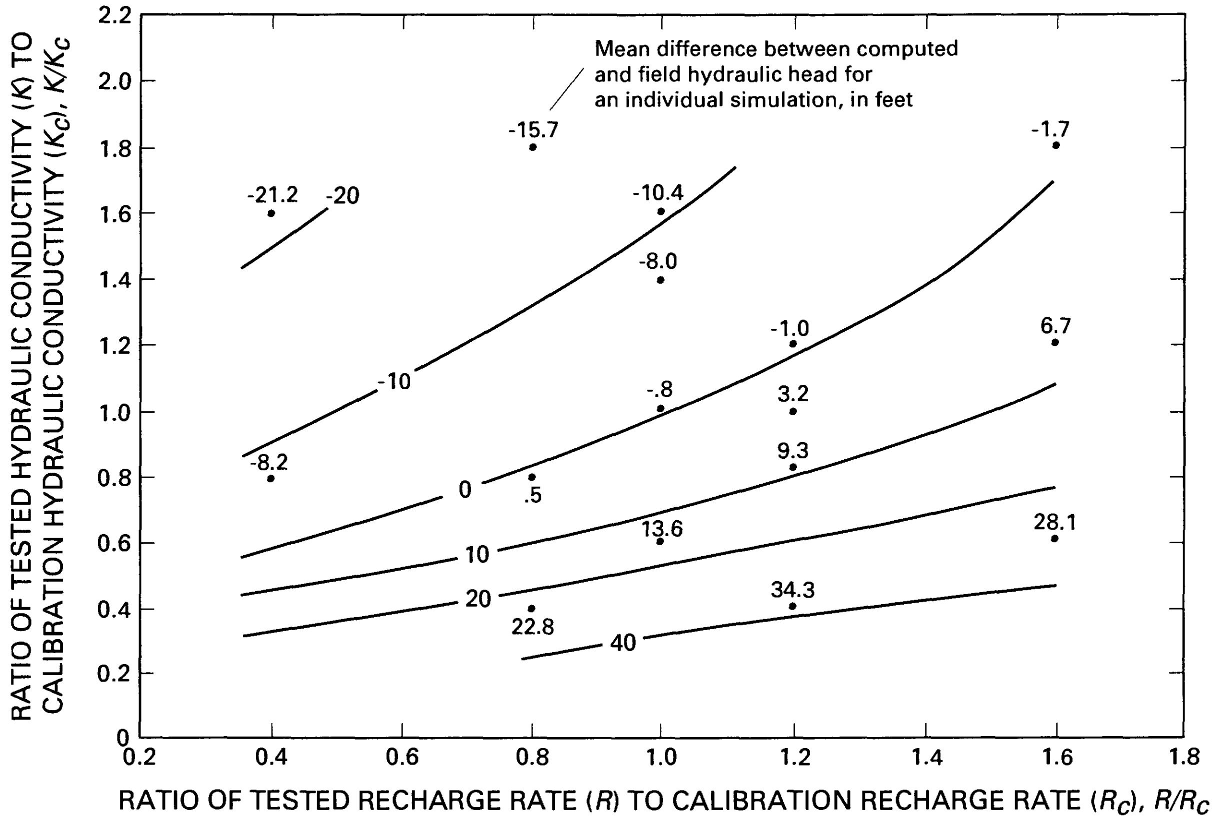


FIGURE 68.—Sensitivity of difference between model-computed and field hydraulic head to simultaneous change in recharge rate and hydraulic conductivity in CMRASA model layer 4.

sensitive to change in recharge relative to calibration values.

ERROR ANALYSIS

McDonald and Harbaugh (1988) point out that any property, such as hydraulic conductivity, of a modeled aquifer associated with a node applies to or is distributed uniformly over the extent of a cell area. This is of concern for large nodes as it could introduce error. The node spacing for the regional study of 28 mi is large because the study area is very large. Additionally, the data available for deep geohydrologic units are sparse. Also affecting selection of a large node spacing was the fact that two subregional studies, a model of the Great Plains aquifer system and a model of the Ozark Plateaus aquifer system, were made at a node spacing of one-half the regional node spacing, allowing study in greater detail. In general, more data are available for those aquifer systems in the subregional areas.

The question of discretization errors resulting from the large node spacing of this study was raised by the investigators during the course of the study. Karplus (1958, p. 104) points out that one would be tempted to assume that the accuracy of any finite-difference solution to a flow equation would be increased by decreasing the node spacing. However, not only the grid spacing but also the nature of the function being analyzed affect the accuracy of the solution. An analysis shows that, if the derivatives of the fourth order or greater of a continuous function are zero (or very small), the second derivative of a function, such as the flow equation (eq. 5) represents an accurate expression regardless of the node spacing (Karplus, 1958, p. 104–105). Estimating the error inherent in a second-order finite-difference approximation of the solution of a differential equation can be accomplished by computing an estimate of the fourth-order derivative (Karplus, 1958, p. 106):

$$\frac{d^4\Phi}{dX^4} \cong \frac{1}{\Delta X^4}\left(\Phi_9 + \Phi_{10} - 4\Phi_1 - 4\Phi_2 + 6\Phi_0 \right), \quad (6)$$

where

d is the differential operator;
Φ is the potential (L);
X is the dimensional direction (L);
ΔX is the node spacing (L); and

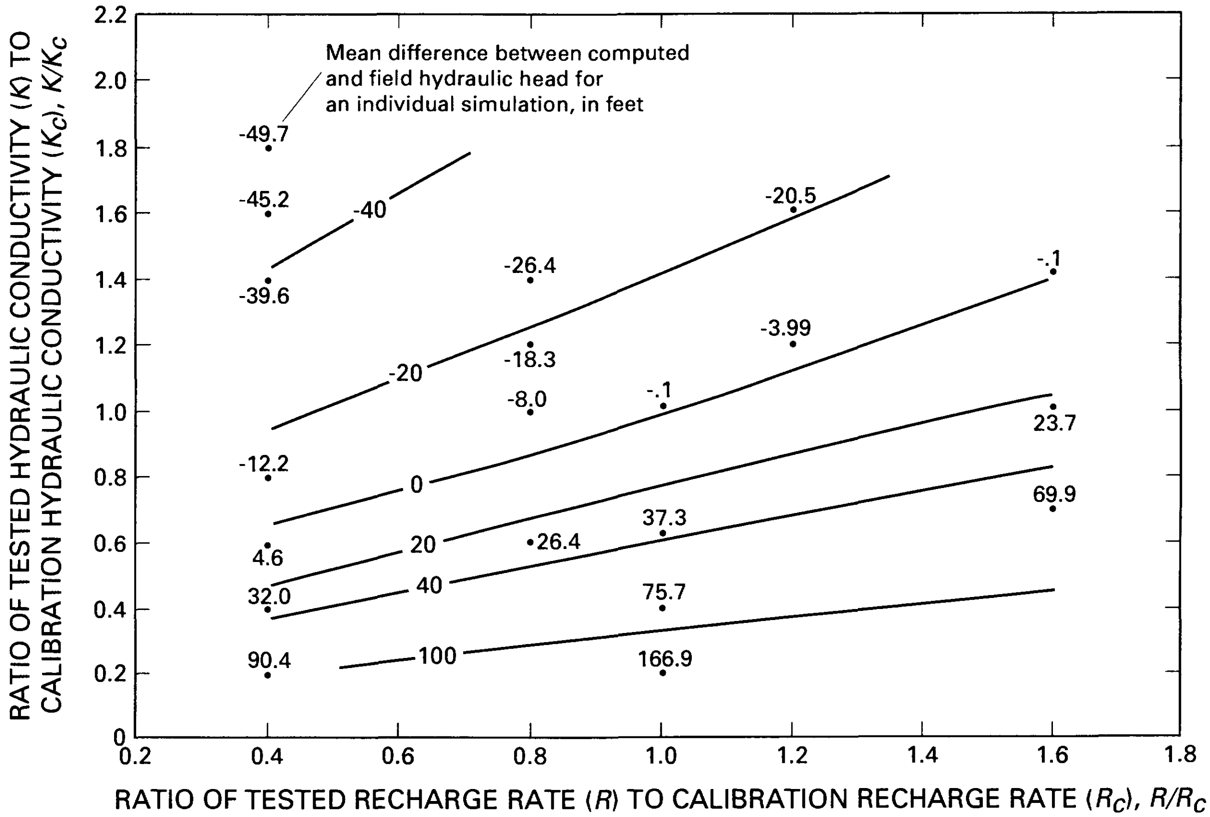


FIGURE 69.—Sensitivity of difference between model-computed and field hydraulic head to simultaneous change in recharge rate and hydraulic conductivity in CMRASA model layer 5.

subscripts "9, 10, 1, 2, and 0" designate five potential values along the x-dimension ΔX distance apart, with Φ_0 between Φ_1 and Φ_2 and Φ_9 and Φ_{10} beyond Φ_1 and Φ_2 (Karplus, 1958, p. 100, fig. 4.20). In general, the hydraulic-head gradients encountered in the regional study area are sufficiently small so that the sum of the potentials in equation 6 are very small in relation to ΔX^4. Therefore, the value of equation 6 is very small. From the preceding premise concerning the fourth-order derivative, the error in solving the flow equation of the model using a large node spacing is sufficiently small that it can be ignored.

However, Weiss (1986, p. 31–35, and p. 39) made an analysis of errors due to node spacing and concluded that the difference in simulated hydraulic head that could be attributed to different node size was surprisingly large considering the small cell size selected. The original model used by Weiss (1986) had a 3-mi cell size. Simulated hydraulic heads using a grid of 1- by 1-mi cells were as much as 90 ft less than simulated hydraulic heads using the larger node spacing in an outcrop area. Recharge 50 percent greater than that of the 3- by 3-mi grid model was required by the 1- by 1-mi grid model to simulate approximately the same hydraulic head near the outcrop.

Ames (1977, p. 24) emphasizes in an error analysis that truncation error is of the finite-difference equation and not of the solution. This point is emphasized because the boundary conditions are essential to the correct solution just as they are when using analytical methods. It is further pointed out that replacement of a continuous function by discretely spaced variables leads to discretization error (Ames, 1977, p. 24). Also, one can readily observe that curved physical boundaries of a study area approximated by a series of square or rectangular cells can be better approximated by a series of small cells than by a few large cells. The errors evoked by large cells at a physical boundary compared to error in a simulation using smaller cells for the same study area are also discretization errors.

An analysis comparing model-computed and field hydraulic-head differences for different node spacings was made for this study. The hydraulic characteristics of the model cells in the regional and subregional numerical models were drawn and discretized from the same maps and data bases. The same computer code was used in all cases for computation.

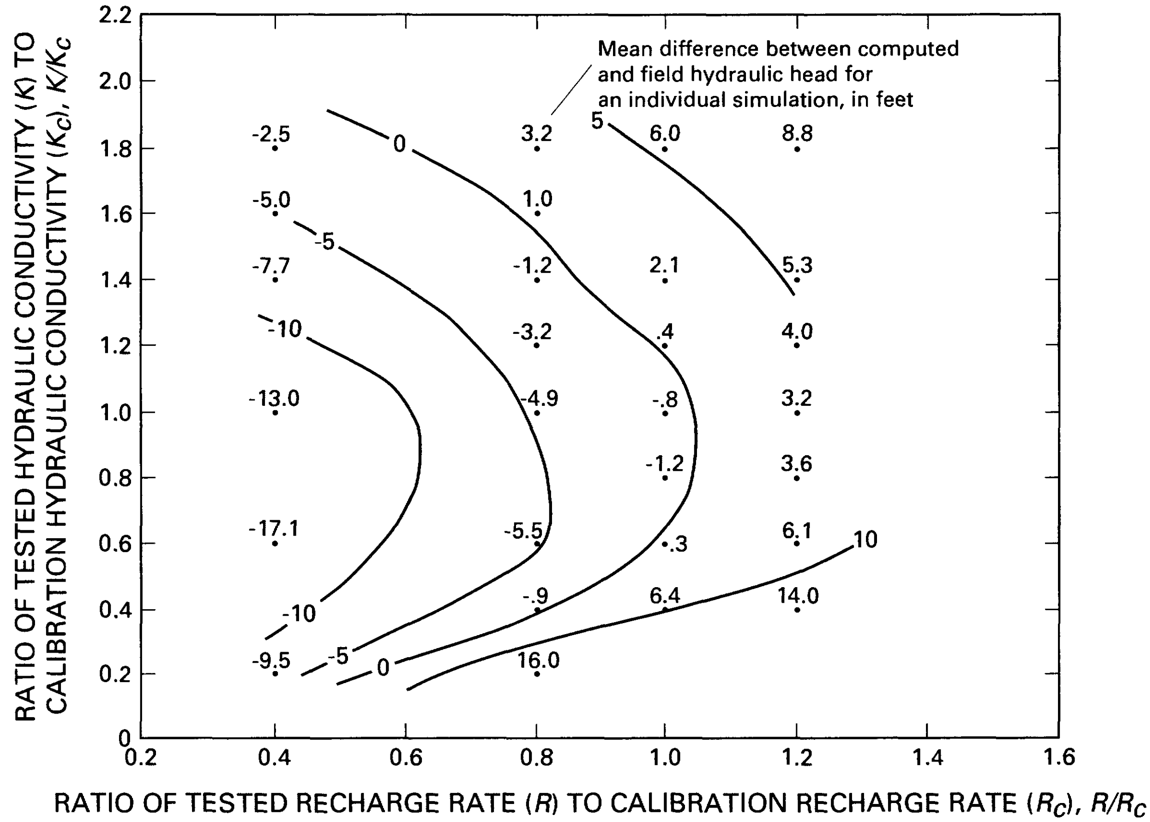


FIGURE 70.—Sensitivity of difference between model-computed and field hydraulic head to simultaneous change in leakance and recharge rate in CMRASA model layer 4.

The area chosen for comparing simulation results that were computed with different node spacing is that encompassing the Great Plains aquifer system. The area in regional model layer 2 in relation to the subregional model grid is shown by figure 72. The Great Plains aquifer system model grid, 46 rows and 35 columns, is included in the area. The data in the Great Plains aquifer system model were discretized to represent model cell areas 14 mi on a side located in quadrants of the regional-model cell areas of 28 mi on a side. Thus, there are no data points in the Great Plains aquifer system model that have exactly the same coordinates of data in the regional model. A data point "located" at a cell center or node in the regional model would be approximated by an average of the four subregional model-cell values within the regional model-cell area.

Comparison of simulated hydraulic-head values computed by the two models was made using a statistical analysis. Field hydraulic-head values for the two systems also were compared using the same analysis. The comparison method used was a test of differences between hydraulic-head values by pairing computed head values and pairing field head values representing the same hydrologic-unit area. An assumption concerning the differences between paired values is that the differences represent a population with a normal distribution of values. The objectives of the analyses were to learn the size of the mean value of each of the two populations of hydraulic-head difference and, particularly, whether the means are different or not different from zero (Snedecor, 1957, p. 49).

The results of three analyses of each of two sets of data are shown in table 6. The largest number of hydraulic-head comparisons in the regional model was termed "all nodes." For every regional-model node where one to four subregional-node values were present, the hydraulic-head value in the regional-model node was compared to the average of the values in the subregional-model nodes. "Full-data nodes" are those where all subregional nodes within a regional-model node have data, and "interior nodes" are comparisons for regional-model nodes that are not located on the unit-extent or study boundaries. The vast majority of interior nodes are nodes that also have data in each of the subregional-model nodes.

The mean difference between field hydraulic-head values for both models on a cell-by-cell basis is small,

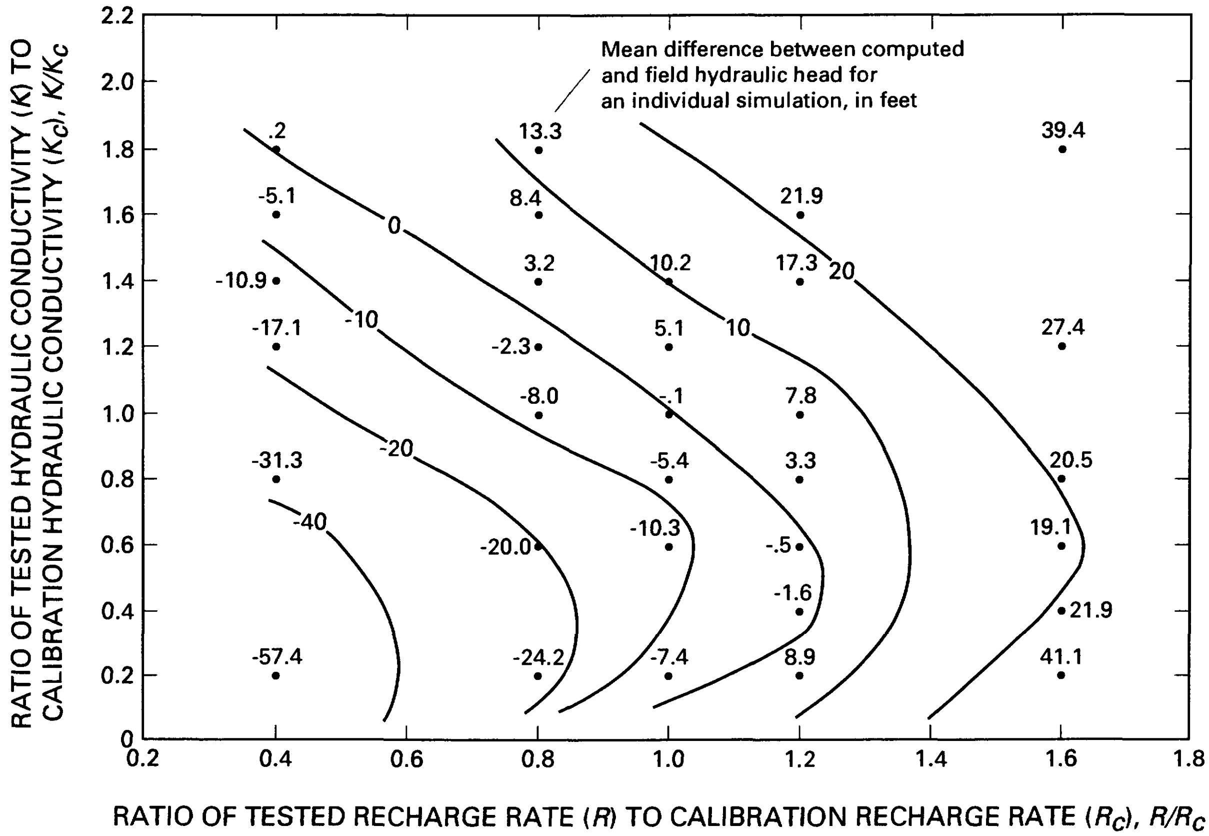
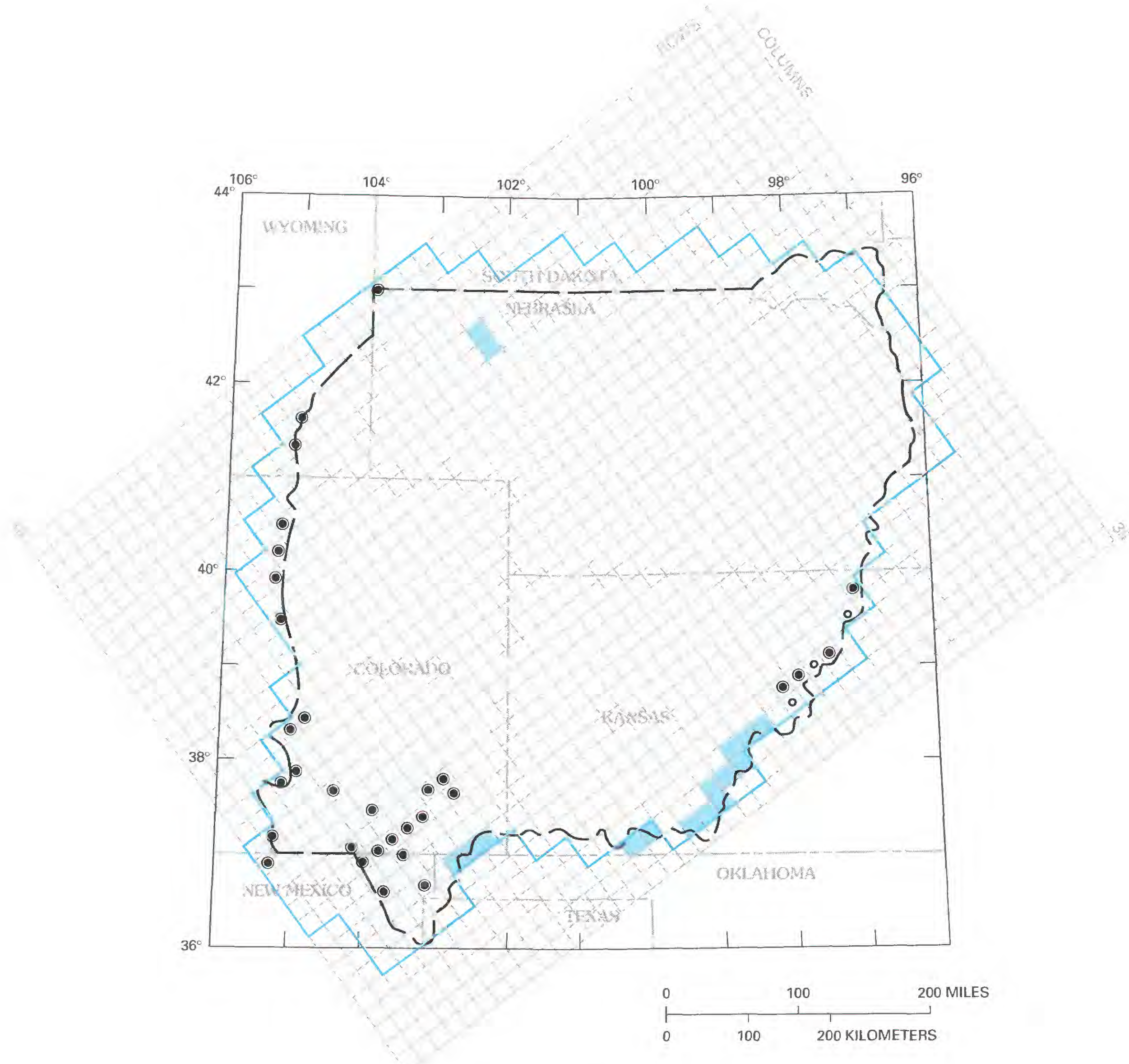


FIGURE 71.—Sensitivity of difference between model-computed and field hydraulic head to simultaneous change in leakance and recharge rate in CMRASA model layer 5.

ranging from 0.57 to 1.42 ft, with standard deviations ranging from 40.28 to 65.88 ft. The field hydraulic-head values are from the same maps and data bases, so the mean differences should approach zero. The standard deviations of the differences represent variations due to judgment and technique of different individuals involved in digitizing and applying the data to represent model-cell areas. The small mean differences show that field values have approximately compensating positive and negative differences. The means of the differences for computed hydraulic head are more variable but range only from 2.91 ft for all nodes to 12.50 ft for full-data nodes. The standard deviations of the differences are similar, ranging from 171.48 to 215.79 ft. The mean difference for "all nodes," which include boundary nodes, is the least because in the model, boundary nodes are constrained by imposed boundary conditions; therefore, they vary less from field values. Conversely, results for the "full-data" and "interior" nodes are more variable because of fewer constraints imposed on those cells in the simulation.

In these cases, the statistical inference from the analyses is that the hypothesis of zero value for the mean of the differences between hydraulic-head values in the models with different-size nodes may be accepted with a significant probability that the hypothesis is correct. The model node-spacing relation between the regional and subregional models does not appear to significantly affect the conclusions that can be made about the computed hydraulic-head values. Thus, the computed hydraulic-head values for the two models do not appear to be significantly different, and it is concluded that the use of a model simulation for the regional area with a smaller node spacing on the order of one-half the spacing used would not have significantly affected the accuracy of the results, a conclusion consistent with the analysis of Karplus (1958, p. 104–105). Luckey and Stephens (1987, p. 31), in a similar analysis comparing 10-mi node spacing (100-mi^2 grids) to 5-mi node spacing (25-mi^2 grids) in a regional study of the High Plains aquifer, determined that "* * * the same general conclusions about the operation of the hydrologic system



EXPLANATION

SUBREGIONAL MODEL

RECHARGE CELL IN OUTCROP AREA

STREAM CELL IN OUTCROP AREA

CELL THAT CONNECTS VERTICALLY ADJACENT LAYERS

REGIONAL MODEL LAYER 2

RECHARGE CELL

MODEL EXTENT

BOUNDARY OF STUDY AREA

FIGURE 72.—Subregional model grid and relation of the Great Plains aquifer system subregional model to regional model layer 2.

TABLE 6.—*Results of statistical analyses comparing hydraulic-head data from the regional and subregional models of the Great Plains aquifer system*

[The statistics are of hydraulic-head differences of paired data. >, greater than]

Hydraulic-head data	Number of data pieces	Mean difference $(\bar{x})$, in feet	Standard deviation (S), in feet	Sample of standard error $(S\bar{x})$	Computed t[1]	Probability of greater mean difference
			All nodes			
Field hydraulic heads	254	1.42	40.28	2.53	0.562	>0.50
Model-computed hydraulic heads	250	2.91	171.48	10.845	.268	>.50
			Full-data nodes			
Field hydraulic heads	216	.57	65.88	4.480	.126	>.50
Model-computed hydraulic heads	216	12.50	193.22	13.15	.951	.33
			Interior nodes			
Field hydraulic heads	217	1.33	56.18	3.81	.346	>.50
Model-computed hydraulic heads	214	11.61	215.79	14.75	.79	.43

[1] $t = \dfrac{\bar{x}}{S_{\bar{x}}}$ (Snedecor, 1957, p. 50).

would have been reached," if the smaller grid spacing would have been used instead of the larger spacing.

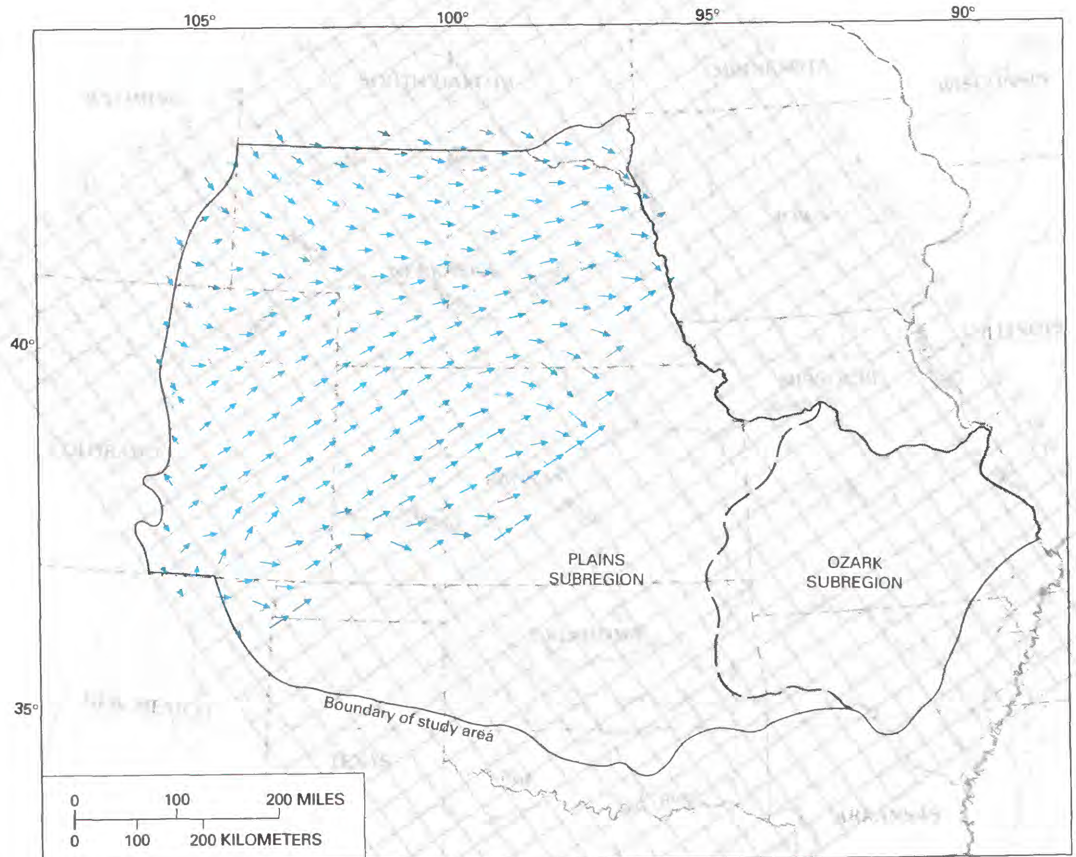
RESULTS OF SIMULATIONS

FLOW

Computed head values have the same general areal distribution as field hydraulic-head values developed from data (figs. 52, 54, 56). Therefore, the simulated regional flow system generally conforms to the initial conception of the flow patterns. A lateral flow-velocity vector and vertical flow-velocity vector were determined at each node for each model cell for all of the units that were simulated. All velocities referred to are Darcy velocities (specific discharges). In the case of lateral flow, flow velocities at model nodes in each direction were derived from simulated volumetric flow rates divided by the harmonic mean of cross-sectional areas at individual vertical cell faces. Then, the two velocity values thus determined for opposing cell faces were averaged. Lateral velocities along the two "horizontal" grid directions then were used to compute a resultant velocity vector. The length of the vector arrows on the flow-direction plots (figs. 73–75) are scaled logarithmically to indicate magnitude. This applies only to the "shaft" of the arrow (the arrowheads or points are all the same size). Therefore, a "shaft" of a flow vector that is twice the length of another vector indicates one order of magnitude (10 times) greater velocity. These data from the model simulation for layers 2, 4, and 5 are shown in figures 73–75. Vertical flow directions are shown in figures 76–79. The relative magnitude of flow also is indicated.

For model layer 2 (fig. 73) representing the Great Plains aquifer system, simulated flow is mostly eastward across Nebraska, with northeasterly components across eastern Colorado and in the western parts of Kansas. Recharge areas are indicated for southeastern Colorado, along the Front Range of the Rocky Mountains, and in the Black Hills Uplift area in southwestern South Dakota. In north-central Kansas, positive recharge was input to model cells representing the outcrop area. However, general-head boundary conditions imposed on the cells allowed flow towards the outcrop cells, and thus a net discharge is indicated by the flow vectors. This is consistent with the model flow budget that indicates an outflow for that area of about 53 ft^3/s.

Vertical flows for layer 2 (figs. 76, 77) indicate upward flow to layer 1 coincident with most of the eastern extent of the Great Plains aquifer system and along the Missouri River in northeastern Nebraska.



Base modified from U.S. Geological Survey
1:7,500,000, National Atlas, 1970

EXPLANATION

LATERAL FLOW VECTOR–Length of shaft indicates relative magnitude

APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS

FIGURE 73.—Lateral flow-velocity vectors for model layer 2 representing the Great Plains aquifer system.

Vertical flows are indicated beyond the extent of the Great Plains aquifer system in northeastern Kansas, southeastern Nebraska, and a short distance in the alluvium along the Missouri River. This is an area of glacial drift and alluvium represented by layer 1, and simulated vertical flows are to or from underlying layers that represent subcrops, such as model layer 3.

Downward flow to layer 2 is greatest near the layer boundaries. The greatest vertical flow is in those locations because the Great Plains confining system is

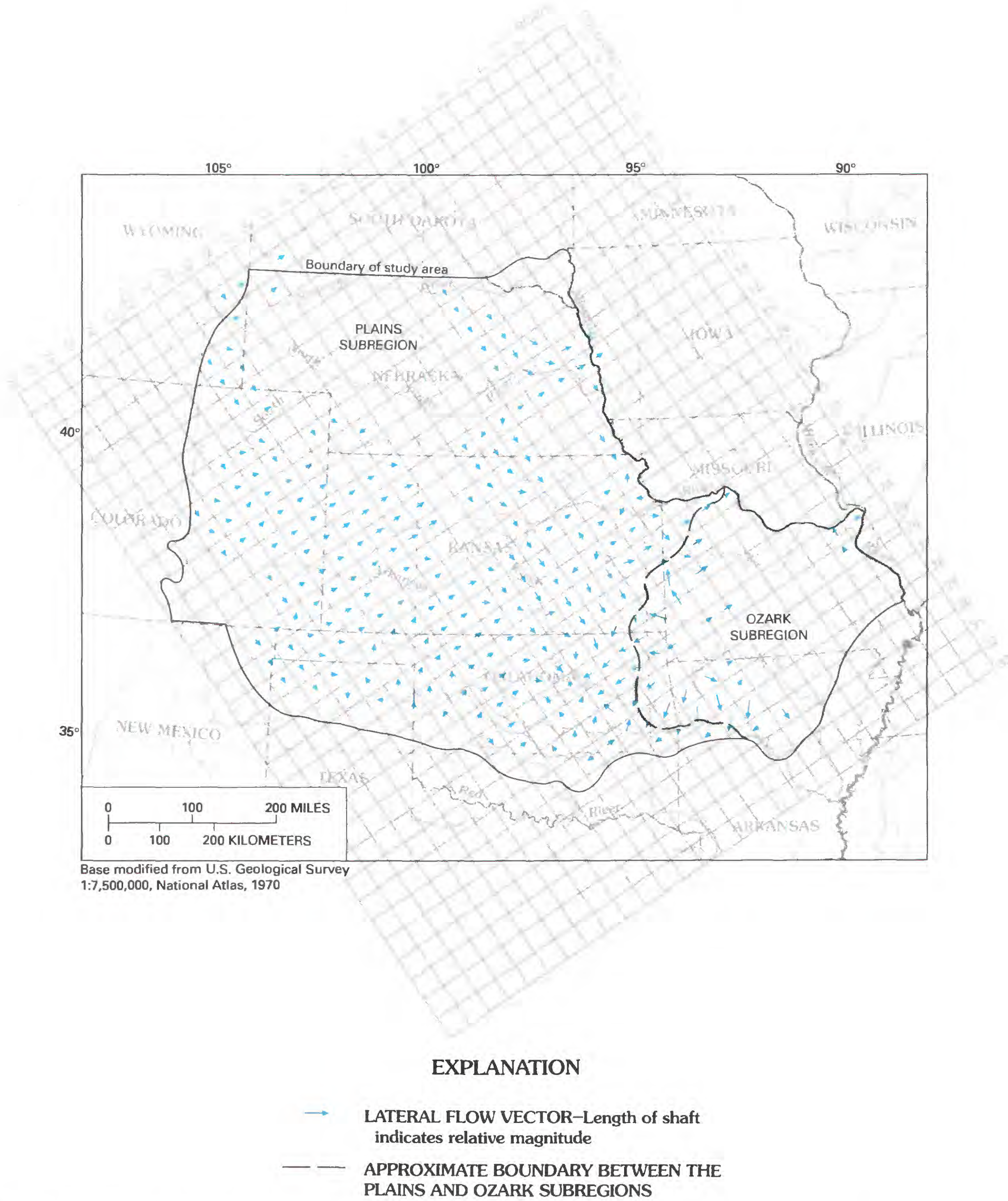


FIGURE 74.—Lateral flow-velocity vectors for model layer 4 representing the upper unit in the Western Interior Plains aquifer system and the Springfield Plateau aquifer in the Ozark Plateaus aquifer system.

thin or not present. Where the permeable materials of the High Plains aquifer and the Great Plains aquifer system are in direct contact, the simulation indicates that the interchange readily occurs.

Vertical leakage of the Great Plains aquifer system, between layer 2 and layer 3, is shown in figure 77. Most of the area has no flow indicated because zero vertical leakance is specified to simulate

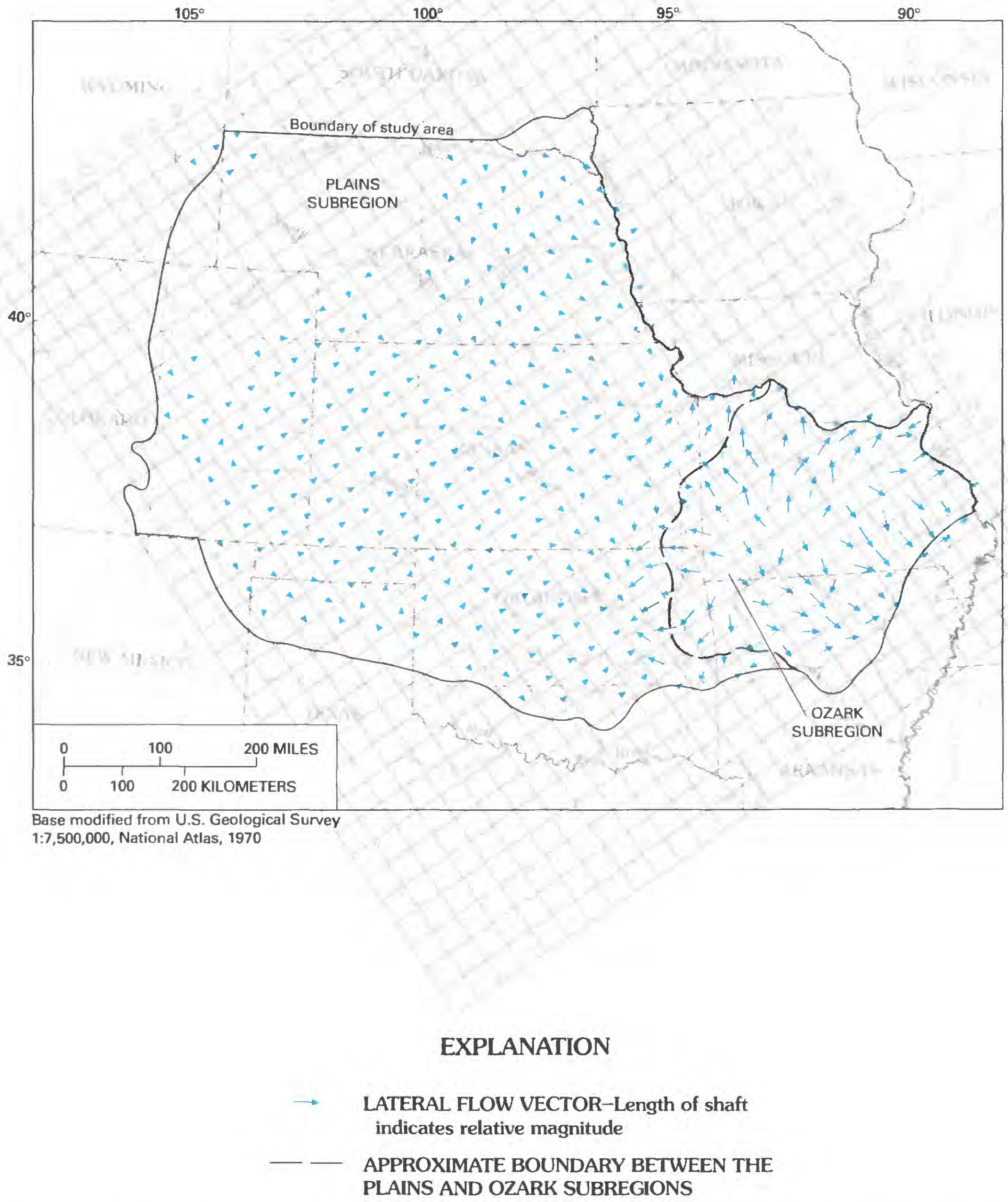


FIGURE 75.—Lateral flow-velocity vectors for model layer 5 representing the lower units in the Western Interior Plains aquifer system and the combined Ozark aquifer, St. Francois confining unit, and St. Francois aquifer in the Ozark Plateaus aquifer system.

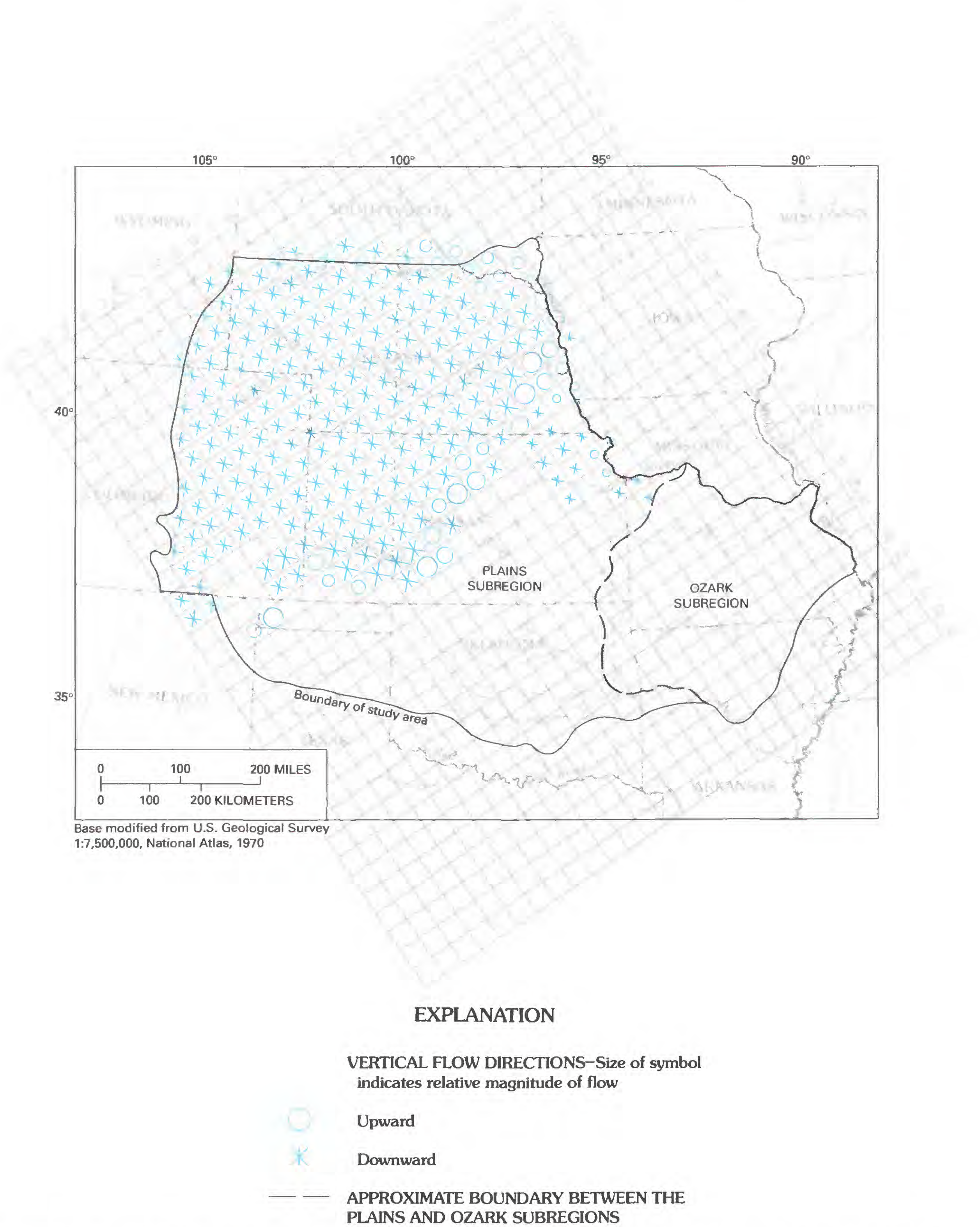
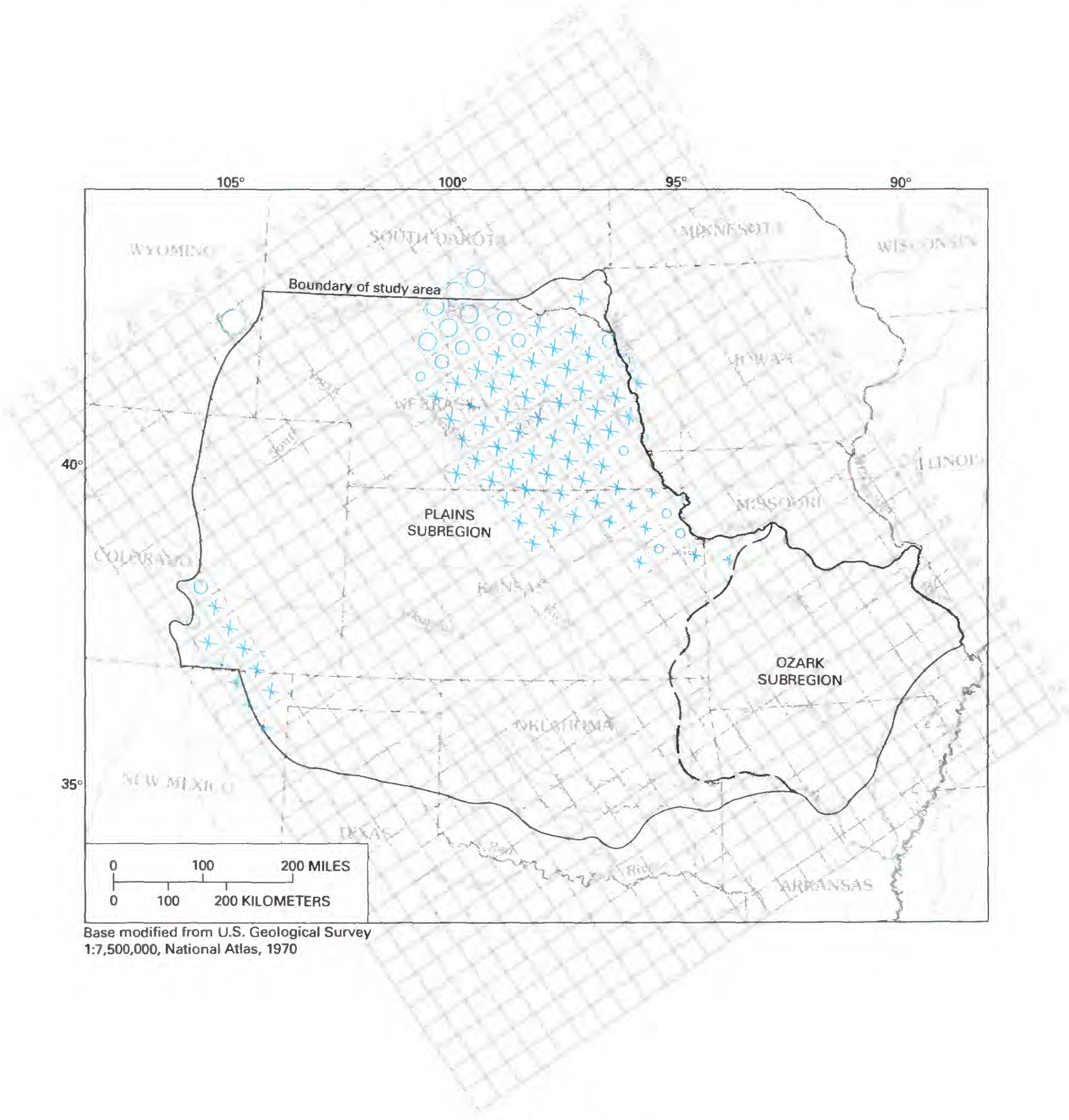


FIGURE 76.—Vertical flow directions for model layer 1 simulating water movement between surficial units and underlying units, principally between the High Plains aquifer and the Great Plains aquifer system represented by model layer 2.



EXPLANATION

VERTICAL FLOW DIRECTIONS—Size of symbol indicates relative magnitude of flow

- ○ **Upward**
- ✳ **Downward**

— — **APPROXIMATE BOUNDARY BETWEEN THE PLAINS AND OZARK SUBREGIONS**

FIGURE 77.—Vertical flow directions between model layers 2 and 3 simulating water movement principally between Great Plains aquifer system and the Western Interior Plains confining system (large no-flow area in model layer 2 results from assigning zero vertical leakance for evaporite deposits in model layer 3).

evaporite deposits in the Western Interior Plains confining system. Where evaporite deposits are not present, vertical flow is indicated, but the quantity is small. An area of significant vertical flow from layer 3 to layer 2 is indicated in north-central Nebraska. As an example to illustrate the small vertical flow, the largest volume of vertical flow in a single model cell between layers 2 and 3 was computed to be about $2x10^{-2}$ ft^3/s. On a daily basis, that amount is about 1,700 ft^3/d or about 631,000 ft^3/yr. Cell size is 28 by 28 mi or 784 mi^2. The leakage then would be about 805 $(ft^3/mi^2)/yr$. A surface-penetration depth for 805 $(ft^3/mi^2)/yr$ is about $3.5x10^{-4}$ in/yr. For some considerations, such as geochemical reactions occurring over geologic time, this quantity could be significant.

Simulated lateral flow in the upper unit in the Western Interior Plains aquifer system (layer 4) shows the diversity of flow resulting from irregular layer-extent boundaries (fig. 74). Flow in eastern Colorado and western Kansas is principally toward the Cambridge Arch and the Central Kansas Uplift. In eastern Nebraska, flow is to the southeast, but flow "splits" around the Nemaha Uplift where the upper unit of the aquifer system does not occur. Along the west and south sides of the Nemaha Uplift, the simulated flow moves southeast and along the southern boundary of the uplift and the model-layer boundary northeastward toward the Missouri River in northeastern Kansas. However, flow also is indicated toward the southeast along the east side of the Central Kansas Uplift into east-central Oklahoma. Flow toward the northeast in western Oklahoma is shown into Kansas and then south back toward eastern Oklahoma. The radial flow of the Ozark Plateaus aquifer system from the Ozark Uplift toward the west and south is illustrated by the vector plot and shows the area in southeastern Kansas where flow directions converge. This results in the freshwater-saltwater transition zone that occurs at the confluence of the Ozark Plateaus aquifer system and the Western Interior Plains aquifer system.

Vertical flows simulated for layer 4 are shown in figures 78 and 79. Because layer 3 is treated as a confining system, flow volumes between layer 3 and 4 (fig. 78) generally are small. The simulation shows upward flow from layer 4 to layer 3 in an extensive area of the western Ozark subregion, throughout most of Oklahoma, all of the Texas Panhandle, and southeastern Colorado. Simulated upward flow over a significant part of that area (western Oklahoma and eastern Texas Panhandle) is due to modeling the overpressure zone in the Anadarko Basin, a zone which also is overlain by evaporite deposits. Upward flow to layer 3 from layer 4 is, however, particularly large at the periphery of the Ozark subregion. This is the area where the Springfield Plateau aquifer, which is represented by model layer 4, discharges water upward. In east-central Colorado, most of Kansas, and the eastern one-half of Nebraska, vertical flow is downward. However, there is a significant area in the central part of Nebraska and the Central Kansas Uplift where the upper unit in the Western Interior Plains aquifer system is not present, and simulated vertical flows shown are between layer 3 and layer 5.

Vertical flow directions simulated between aquifers represented by model layers 4 and 5 are shown in figure 79. There is a clear pattern of large upward flows to layer 4 at the boundary of the Ozark subregion and large downward flows to layer 5 from layer 4 in the interior of the Ozark subregion. Along the Missouri River in eastern Nebraska, a majority of nodes show upward flow to layer 4 indicating discharge to the river. (Recall that flow from layer 3 down to layer 4 predominated along the Missouri River in that locality.) From the extreme southeastern corner of Nebraska along the Missouri River to about 60 mi east of Kansas City (fig. 1), the simulation shows downward flow from layer 4 to layer 5, but thereafter, with only one exception, fairly large upward flows occur from layer 5 to layer 4 along the Missouri and Mississippi Rivers, which simulates discharge to those rivers.

Flow from layer 5 to layer 4 (fig. 79) is indicated in central Oklahoma extending in a narrow band across central Kansas into Nebraska. Location of the band is approximately bracketed between the Central Kansas Uplift and the Nemaha Uplift. Along that band from south-central Nebraska through north-central Kansas, the upper unit in the Western Interior Plains aquifer system is present. Just west of the band of upward flow cells is the Central Kansas Uplift where the rocks represented by layer 4 are not present. Thus, the simulated downward flow is from the Western Interior Plains confining system represented by layer 3 to layer 5 representing the lower units in the Western Interior Plains aquifer system.

Because layer 4 is modeled as overpressured in the Anadarko Basin of west-central Oklahoma, simulated flow is into layer 5. In the Texas Panhandle and a large area in southwestern Kansas, some upward flow is indicated from layer 5. However, the simulated hydraulic-head values for those areas of layer 4 and layer 5 are nearly the same (figs. 54, 56), implying that vertical flow is small. The reason these simulated hydraulic-head relations exist in the panhandle area is that the areal extents of layers 4 and 5 across the Texas Panhandle are coincident and are modeled using

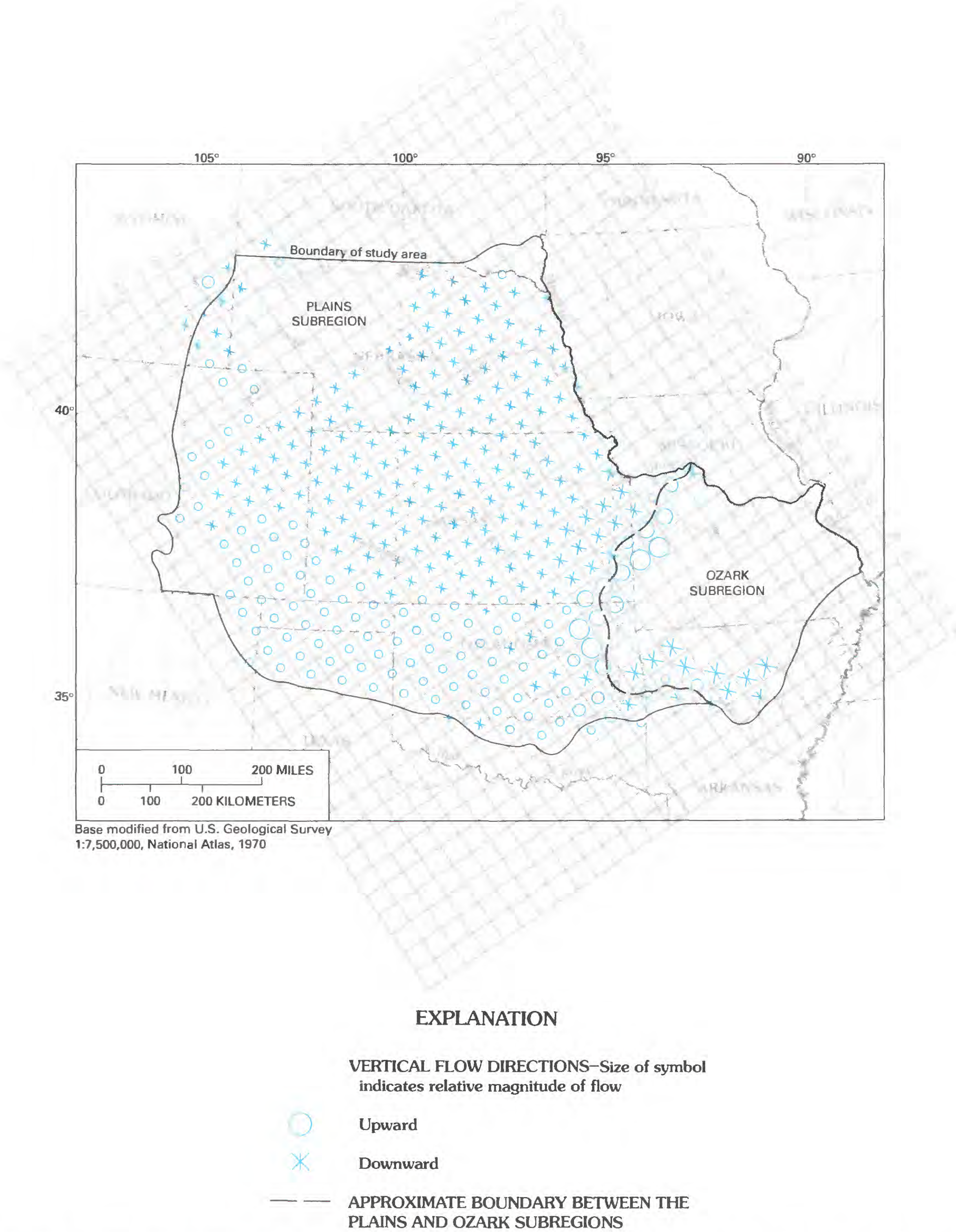


FIGURE 78.—Vertical flow directions between model layers 3 and 4 simulating water movement between Western Interior Plains confining system and the upper unit in the Western Interior Plains aquifer system and between the Western Interior Plains confining system and the Springfield Plateau aquifer in the Ozark Plateaus aquifer system.

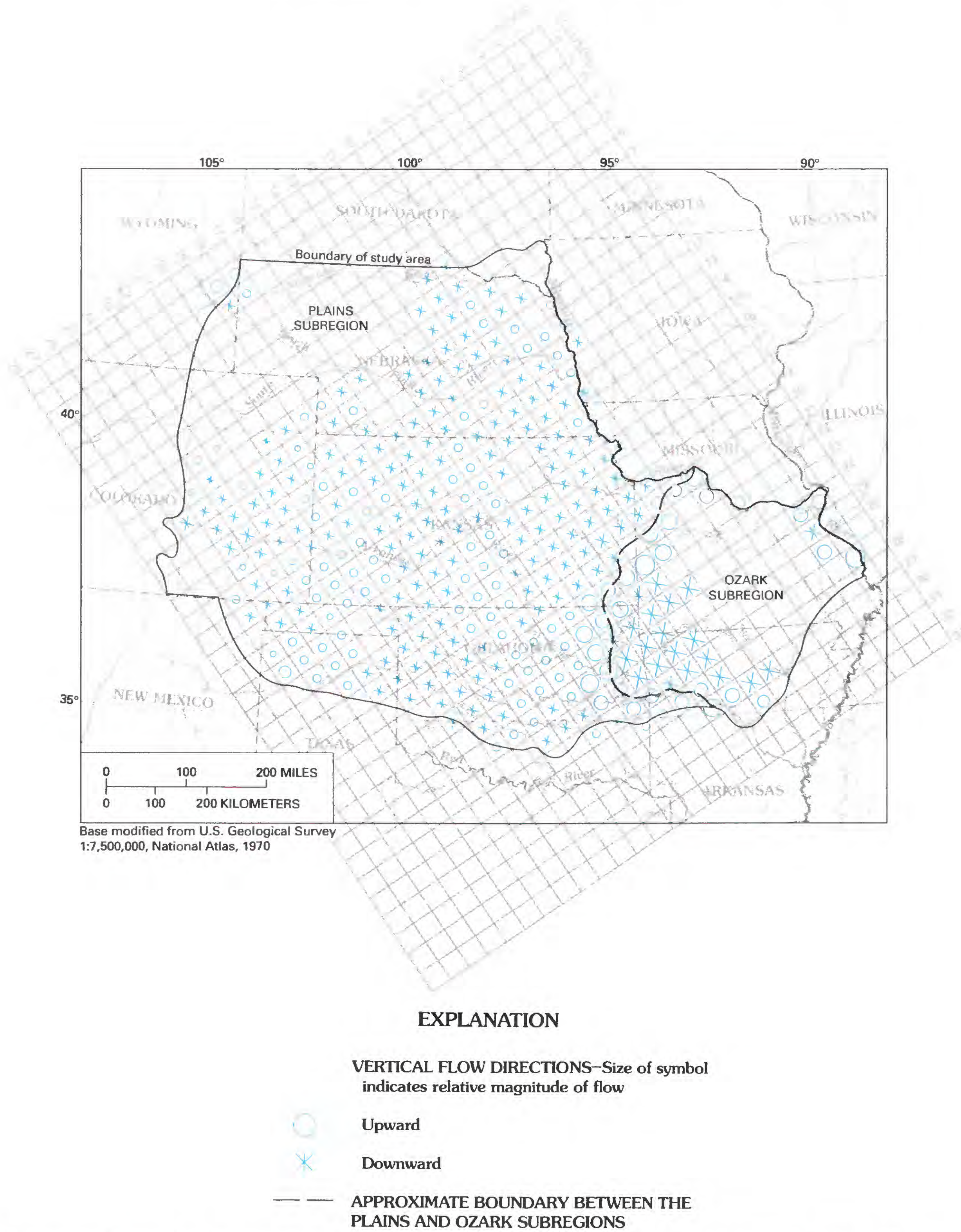


FIGURE 79.—Vertical flow directions between model layers 4 and 5 simulating water movement between the upper and lower units in the Western Interior Plains aquifer system and between the Springfield Plateau aquifer and the combined Ozark and St. Francois aquifers in the Ozark Plateaus aquifer system (no-flow area in the Ozark subregion is where model layer 5 simulates an outcrop area).

general-head boundaries to simulate continuation of the units beyond the extent of the study area (figs. 50, 51). Therefore, the initial condition in which field hydraulic heads in layer 5 slightly exceeded heads in layer 4 persists when the hydraulic heads are computed in the simulation. The extent of model layers and boundary conditions imposed in eastern Colorado and northwestern Kansas results in both upward and downward flow in adjacent cells in that area. Variations of aquifer characteristics, steep hydraulic-head gradients, and the occurrence of mountain and basin conditions affect the simulation accuracy and consistency because of the large cell areas.

Lateral flow-velocity vectors for layer 5 shown in figure 75 have a pattern very similar to those for layer 4 (fig. 74) in the western part of the regional area. This is expected because the computed hydraulic heads for the two layers are similar and because, in locations of the model where the confining unit is absent between the upper and lower units in the Western Interior Plains aquifer system, the layers are modeled with a large leakance coefficient between them. The lateral flow vectors in layer 5 are affected by large permeability and recharge in the outcrop area causing the outflow pattern from the Ozark subregion. The same flow pattern near the Ozark subregion boundary in southeastern Kansas shown for layer 4 also is shown by the flow pattern for layer 5 and results from the simulation of that area as a boundary between separate freshwater and saltwater flow systems.

A comparison can be made between the magnitude of lateral flow vectors in layers 4 and 5 in southwestern Kansas where the units represented by these layers are deeply buried and the magnitude of lateral flow-velocity vectors in the Ozark subregion where these units are closer to the surface. A representation of flow in the deep sections by model layers 4 and 5 indicates nearly stagnant or extremely sluggish flow, whereas flow velocities in the Ozark subregion are many orders of magnitude greater (figs. 74, 75). For example, an average interstitial velocity across a model-cell face for layer 5 located in western Kansas was computed to be 3.2×10^{-10} ft/s compared to an average interstitial velocity of 1.4×10^{-5} ft/s in the Ozark subregion, a difference of about five orders of magnitude.

Lateral flow for layer 5 in Kansas and Nebraska is affected by the Cambridge Arch and Central Kansas Uplift. From southwestern Kansas, flow appears almost uniform toward the northeast, then at the Central Kansas Uplift, flow turns toward the southeast paralleling the uplift (fig. 75). In northeastern Kansas, the flow moves to the northeast toward the Missouri River. In the eastern one-half of Nebraska, flow is indicated to the south just east of the Cambridge Arch moving east and southeast toward the Missouri River, and also is affected by the Nemaha Uplift. Flow in Oklahoma is generally toward the northeast, meeting with the outflow from the Ozark subregion. Upward flow from layer 5 to layer 4 is shown in eastern Oklahoma (fig. 79), primarily as a result of the model simulation of the boundary conditions of the two flow systems.

MODEL BUDGET

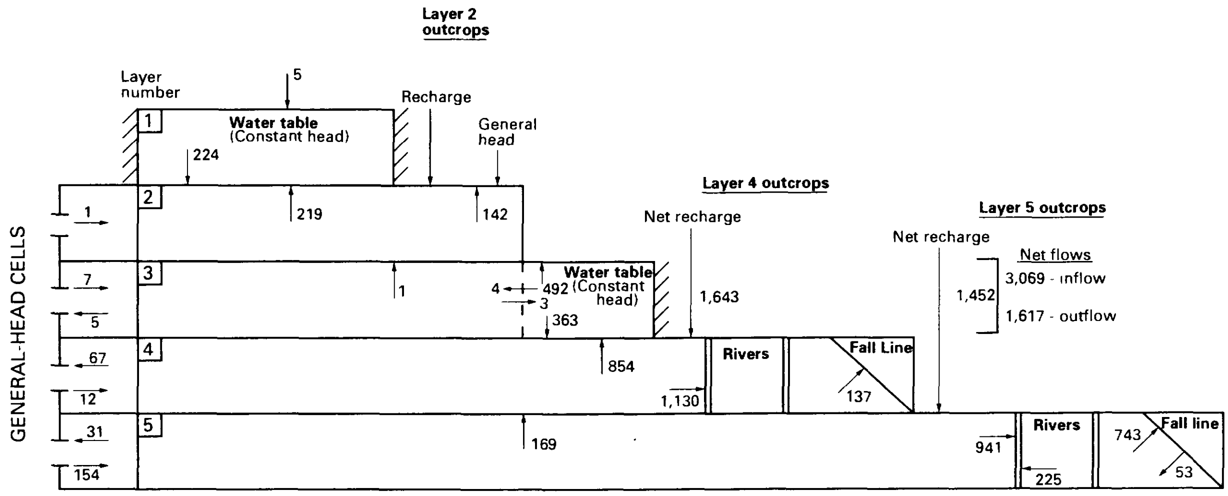
The model-computed budget results from the imposed aquifer properties, geometry, and boundary conditions, coupled with the computed hydraulic head. The budget for all layers is shown in figure 80.

The specified-head condition for the surficial units (layer 1) resulted in a simulated downward flow of 224 ft^3/s from layer 1 to layer 2 and simulated upward flow from layer 2 to layer 1 of 219 ft^3/s.

The overall hydrologic budget for layer 2 (the Great Plains aquifer system) is about 361 ft^3/s in and about the same amount out (fig. 80). The representation of the Great Plains aquifer system (layer 2) has about 93 ft^3/s recharge imposed (table 5), most of which (83 ft^3/s) is in north-central Kansas with general-head boundaries specified to allow the simulation to compute recharge values necessary to balance the flow. The model simulation, from specification of general-head boundaries, computed an additional recharge of 42 ft^3/s, a major part of which (27 ft^3/s) was for the southeastern Colorado, northeastern New Mexico area. In north-central Kansas, discharge at outcrops was computed to be 136 ft^3/s for a net outflow at those model cells of 53 ft^3/s.

The Western Interior Plains confining system (layer 3) has only about 1 ft^3/s interchange with the Great Plains aquifer system represented by layer 2, and there are very small lateral flows. The major interchanges from layer 3 occur in cells that simulate the outcrop area of constant-head boundary conditions near the Ozark subregion. There, vertical leakance allows upward movement of water from the Ozark Plateaus aquifer system and some from the Western Interior Plains aquifer system (fig. 78).

In regard to the constant-head cells (fig. 80), inflow to the system (to layer 4) is 363 ft^3/s, and outflow from the system (to layer 3) is 854 ft^3/s for a net leakage from layer 4 to layer 3 of about 491 ft^3/s. All but about 1 ft^3/s of the flow from layer 4 to layer 3 is within and in proximity to the Ozark subregion.



EXPLANATION

1,130 DIRECTION OF FLOW—Number indicates rate of flow, in cubic feet per second

Note: Fall Line refers to the boundary with the Mississippi Alluvial Plain

FIGURE 80.—Simulated regional hydrologic budget.

Recharge to that part of layer 4 that represents the Springfield Plateau aquifer in the Ozark subregion was input as 1,643 ft^3/s (fig. 80). A major outflow determined from the simulation for the Springfield Plateau aquifer is 1,130 ft^3/s to rivers, 772 ft^3/s to the Missouri and Mississippi Rivers, and about 355 ft^3/s from model cells in the vicinity of the Canadian and Arkansas Rivers in eastern Oklahoma and northwestern Arkansas (fig. 1). An outflow across the Fall Line (toward the Mississippi Alluvial Plain) of 137 ft^3/s is indicated. Vertical-flow vectors (fig. 79) between layer 4 and layer 5 representing flow between the Springfield Plateau aquifer and the Ozark and St. Francois aquifers in the Ozark subregion and within about 40 mi beyond the western Ozark subregion boundary indicate a large vertical exchange. In this area, the model-computed flow between layers 4 and 5 is 1,366 ft^3/s from layer 4 to 5 and 1,527 ft^3/s from layer 5 to layer 4, resulting in net upward flow of 161 ft^3/s. The model budget for the region (fig. 80) indicates a net vertical flow from layer 5 to layer 4 of 169 ft^3/s; therefore, most of that vertical flow for the regional model occurs in the cells representing the Ozark subregion and its vicinity.

Model budget results for layer 5 (fig. 80), the combined Ozark aquifer, St. Francois confining unit, and St. Francois aquifer in the Ozark subregion, indicates a combined aquifer discharge to the Missouri and Mississippi Rivers of 941 ft^3/s and combined inflow from rivers of 225 ft^3/s. Simulated inflow and outflow at the study boundary for layer 5 (the lower units in the Western Interior Plains aquifer system) are 154 and 31 ft^3/s, respectively. These flows are principally along the study-area boundaries in Oklahoma, the Texas Panhandle, and southeastern Colorado (fig. 51).

Flow was generally out of the Ozark subregion; the net simulated flow across the Fall Line (fig. 3) to the Mississippi Alluvial Plain for layers 4 and 5 is 827 ft^3/s. Imes and Emmett (1994) present evidence that most ground water entering the Mississippi Alluvial Plain across the Fall Line discharges upward into alluvial material within a short distance. Permeable rocks of the Ozark aquifer (represented by layer 5 in the Ozark subregion) are in direct contact with permeable alluvium and Cretaceous-age sand in a narrow band approximately 10-mi wide along the Fall Line. Southeast of the band, a nearly impermeable Cretaceous clay separates the permeable Cretaceous sand and underlying Paleozoic rocks from the alluvial material. Some freshwater may flow beneath the clay and eventually leak upward into the alluvium, but it is likely that most ground water moves into the alluvium. Also, the few dissolved-solids measurements indicate freshwater does not travel far beneath the alluvial plain (Imes and Emmett, 1994). Water that crosses the Fall Line from the Ozark Plateaus aquifer system probably discharges into marshes, shallow water tables, drains, and the Black and White Rivers (fig. 1) that are adjacent and practically parallel to the Fall Line for a major part of its length.

WATER SUPPLY AND WATER-STORAGE POTENTIAL

GREAT PLAINS AQUIFER SYSTEM

The Great Plains aquifer system in the regional study area is generally overlain by a thick, slightly permeable confining material and by the very productive High Plains aquifer. Because large quantities of water of usable quality are available from the High Plains aquifer and water in the Great Plains aquifer system is of questionable or unknown quality, very little development for water supply has occurred in relation to the amount of water stored. Also, because the Great Plains aquifer system is deeply buried in northeastern Colorado, northwestern Kansas, and throughout most of Nebraska, it has not been considered prior to this study to have significant potential for development (Helgesen and others, 1982, p. 410). However, a unit saturated thickness exceeding 700 ft in Nebraska indicates a very large quantity of water in storage. Water stored in the Great Plains aquifer system based on thickness and porosity (figs. 34, 35) is estimated to be almost 12 billion acre-ft (table 7) in the regional study area. The amount of water that the aquifer system might yield is estimated to be one-fourth that amount or about 3 billion acre-ft if nonyielding clay layers and specific yield are considered—still a very large resource. (Note: In table 7 the amount listed for South Dakota, 190 million acre-ft, is not representative of the storage in the Great Plains aquifer system for the whole State but represents only the small part of the State included in the CMRASA.) Helgesen and others (1982, p. 410) point out that in the regional study area, the Great Plains aquifer system (historically referred to in the literature as the "Dakota") has been an important source of water for irrigation, domestic, or stock use for many years in parts of Kansas and southeastern Colorado. The droughts of the 1970's increased attention to the Great Plains aquifer system as a source for irrigation water in Kansas, eastern Nebraska, and northwestern Iowa.

Water from the Great Plains aquifer system may be a source to supplement water withdrawn for irrigation from the High Plains aquifer. This conclusion is based on small dissolved-solids concentrations (fig. 37) and simulated vertical flow between the surficial units (layer 1) and the Great Plains aquifer system (layer 2) (fig. 76). Potable water occurs along the eastern and southeastern extent of the Great Plains aquifer system, a condition consistent with the vertical exchange (both upward flow and downward flow) of water between the two layers as indicated by the steady-state model simulation. Along the eastern extent of the Great Plains aquifer system, the natural flow system may be effective in mixing natural recharge, water in the Great Plains aquifer system, and water in the High Plains aquifer. This aspect is particularly important to a location in Kansas where the largest area of interchange is indicated by the model simulation in an area of large irrigation demand and concern for dwindling supplies. Additional study of the effect on the quantity and quality of water flowing into and out of the High Plains aquifer would be of value at locations where it is presently occurring and at locations where interchange might be induced further. Previous studies have dealt with parts of the issue to a large degree (U.S. Geological Survey, 1966; Lobmeyer and Weakly, 1979; Kume and Spinazola, 1982; Spinazola and Dealy, 1983; Kume, 1984; Watts, 1989), but in light of information developed in the present study, further investigations would be of value.

WESTERN INTERIOR PLAINS AQUIFER SYSTEM

Ground water stored in the Western Interior Plains aquifer system, based on saturated thickness and regionalized porosity (figs. 23, 25), amounts to more than 7 billion acre-ft of water (table 7). Because very little of the aquifer system contains water with less than 5,000 mg/L dissolved solids, the water has very limited use, and its potential for development as a water supply is slight.

Results of flow simulation in model layers 4 and 5 representing units in the Western Interior Plains aquifer system, however, indicate extremely slow fluid velocities in the deeply buried units and that the system is nearly isolated. Because of the degree of isolation, very slow-moving water, and poor water quality, the aquifer system at least in some areas has characteristics that indicate potential for waste storage or disposal. Although this study did not consider waste disposal, consideration of the depth, slow groundwater movement, and extremely large dissolved-solids concentrations in the ground water (brines) suggests that waste storage or disposal is feasible. As an example, lateral flow velocities simulated in layer-5 model

TABLE 7.—*Water in storage in geohydrologic units in the regional study area*
[Data in millions of acre-feet; ---, no data. Total ground water in storage. Drainage water would be less if nonyielding clay beds and specific yield of permeable units are taken into consideration]

State	Geohydrologic unit		
	Great Plains aquifer system	Western Interior Plains aquifer system	
		Upper unit	Lower units
	Plains subregion		
Arkansas	---	---	71
Colorado	1,553	188	223
Kansas	2,717	818	1,631
Missouri	---	29	121
Nebraska	7,035	174	766
New Mexico	54	8	9
South Dakota	190	---	8
Oklahoma	9	625	2,099
Texas	---	137	215
Wyoming	244	51	---
Total	11,802	2,030	5,143

State	Geohydrologic unit			
	Springfield Plateau aquifer[1]	Ozark aquifer	St. Francois aquifer	Combined St. Francois-Ozark aquifer[2]
		Ozark subregion[3]		
Arkansas[4]	30	1,775	51	1,825
Arkansas[5]	50	1,113	39	1,152
Kansas	25	127	7	134
Missouri	118	3,329	544	3,873
Oklahoma	45	445	8	454
Total	268	6,789	649	7,438

[1]Regional model layer 4. [2]Regional model layer 5.
[3] Data from Leo F. Emmett, U.S. Geological Survey, written commun., 1987.
[4]Freshwater. [5]Saline water.

cells in southwestern Kansas are extremely slow. The model computed a quantity of flow across a model-cell face in the northeasterly direction to be $9\text{x}10^{-6}$ ft^3/s (layer 5, row 21, column 16; flow from row 21 to 20 along column 16). The simulated face area of that model cell is $1.53\text{x}10^{8}$ ft^2, and porosity is about 5 percent (fig. 23). Therefore, a pore velocity for modeled flow would be about $1.2\text{x}10^{-12}$ ft/s, which is $4\text{x}10^{-5}$ ft/yr or about 40 ft in a million years. Even for three orders of magnitude greater velocity, the travel would be only about 7.6 mi in a million years.

In considering long-term waste storage, other important factors besides velocity of travel should be evaluated. Plans for storage of wastes near areas of prior tectonic activity would need to include the potential for future tectonic activity. There is general tectonic stability in the Western Interior Plains; however, tectonic activity has occurred in the Rocky Mountains and across the southern boundary of the regional study area, such as the Cimarron Uplift, the Amarillo-Wichita-Criner Uplifts, and the Arbuckle Mountains. Also, the Central Kansas Uplift and similar features indicate areas of potential tectonic activity as a result of prior tectonic events. Although the overpressure zone in the Anadarko Basin is in an area of very slight permeability and water movement is extremely slow, the potential for upward movement exists because of overpressure. Waste storage in an overpressured zone probably is not desirable. However, the integrity of the overlying confining material, the very slow water movement, and the thickness of the confining system overlying the Western Interior Plains aquifer system indicate that the aquifer system has potential for long-term waste storage or disposal in some areas.

OZARK PLATEAUS AQUIFER SYSTEM

The results of this study quantify the large potential of the water resources of the Ozark subregion. Freshwater in storage in the Ozark subregion of the CMRASA is estimated to exceed 7.5 billion acre-ft (table 7) (L.F. Emmett, U.S. Geological Survey, written commun., 1987). Also, a total model-recharge input (table 5) of 6,947 ft^3/s is but a small part of the approximately 55,000 ft^3/s of potential recharge from precipitation through the soil zone to the water table for the entire Ozark subregion. Thus, more than 87 percent of the potential recharge is "rejected," that is, intercepted locally by streams as indicated by the regional model analysis. The rejected water is a ground-water resource that could be utilized. Increased pumping of ground water would result in increased recharge to the regional aquifers and decreased discharge to streams in their outcrop areas.

A major point to be considered is that, of the water resources actually available for planning and development in the Ozark subregion, the model evaluates only a relatively small component of the water that reaches the water table and is not discharged locally (not intercepted within model cells). Water budgets from ground-water models have been used as a basis for planning and development; however, serious errors can result if available resources not accounted for in the model are ignored. The Ozark subregion is not unique in that some tens of thousands of cubic feet per second of rainfall and rejected or intercepted recharge could be withdrawn and removed from the ground-water flow system without materially affecting the ground-water flow system. However, such withdrawals would affect the surface-water resource. For example, if the 6,947 ft^3/s of recharge that was simulated by the model was used in the future, this would not mean that all the aquifer contribution to the base flow of stream would be eliminated. Instead, less recharge from infiltration to the water table would be intercepted by streams. Ground-water contribution to streams would be reduced, and the overland part of the streamflow would be affected only slightly. The ground-water resources of the area generally are little used. Harvey (1980, p. 1) concluded that because only a small part of available ground water in the Springfield-Salem Plateaus of southern Missouri and northern Arkansas is used (and large increases in use are not predicted), it is unlikely that declines in ground-water levels of extended duration will occur in the near future.

The regional model computed flow to the Missouri and Mississippi Rivers of 1,488 ft^3/s in the Ozark subregion. Flows were 1,088 ft^3/s to the Missouri River and 400 ft^3/s to the Mississippi River. These computed flows are minor in relation to actual mean annual streamflows, such as 177,800 ft^3/s, the mean annual discharge of the Missouri River at St. Louis for 1951–80 (Hedman and Jorgensen, 1990, table 2). However, results of an analysis of the gain in mean annual flow to the Missouri River from the Missouri River Valley aquifer between Waverly and St. Louis, Missouri, for that 30-year period averaged 1,110 ft^3/s (Hedman and Jorgensen, 1990, table 2), a value comparable to the 1,088 ft^3/s computed by the model for subjacent regional geohydrologic units. Analysis by Hedman and Jorgensen (1990) includes gaged inflow, ungaged inflow, valley recharge, city discharge, city intake, water-surface evaporation, consumptive use, and Missouri River Valley aquifer underflow. Large boundary sources and sinks (such as the Missouri and Mississippi Rivers) often are simulated as unknowns so that the resulting computed flows compensate for other imposed conditions, such as specified-head and head-dependent flux. The implications are that analyses of gains and losses of major streams should be carefully analyzed particularly if the sources or sinks are boundaries between studies. The comparability of model-computed river gain to an independent analysis of river gain (gain of the Missouri River between Waverly and St. Louis, Missouri) is an indication of good correspondence between the model and the independent streamflow study.

SUMMARY

The study area of the Central Midwest regional aquifer-system analysis (CMRASA) extends from the foothills of the Rocky Mountains in Colorado to the Mississippi River in eastern Missouri, and from South Dakota to the Ouachita, Arbuckle, and Wichita Mountains of Arkansas and Oklahoma. The purpose of this report is to describe the aquifers, aquifer systems, confining units, confining systems, and the results of computer simulation of regional flow. The study is limited to rocks ranging from Cambrian through Lower Cretaceous age and, thus, does not include the flow regime in younger rocks constituting the High Plains aquifer and other surficial units. However, effects of Upper Cretaceous and younger units are included in the simulation as an upper boundary.

Structural deformation is not severe over most of the study area, which lies within the stable interior of the North American continent. Geologic structure is dominated by broad basins and arches; however, along the southern and western boundaries, substantial

crustal deformation resulted in intense folding and faulting. These complex structures define the southern and western boundaries of the study area. Generally, time-stratigraphic units are continuous where present without severe structural deformation and offset of rock units except at those boundaries.

Regional estimates of porosity were determined from geophysical logs. A comparison between reported porosity (from analyses of drill-stem tests, data from laboratory analyses of rock cores, and analyses of production data from oil and gas reservoirs) and the regionalized porosity (from geophysical logs) indicated good correlation for loosely cemented fine- to medium-grained sandstone with dominantly primary porosity but poor correlation for rocks such as limestone in which secondary porosity dominates. Reported porosity values generally are not representative of entire stratigraphic sections because testing was done in specific oil and gas reservoirs, which typically have greater than average porosity. Preliminary maps of porosity were made and modified to provide an estimate of porosity for basins based on decreases in porosity with depth.

Few permeability data were available for most of the regional study area. In rocks containing saline water, permeability data were obtained from analyses of drill-stem tests, laboratory analysis of rock cores, and oil-production tests. A method was developed to estimate the permeability of geohydrologic units from geophysical logs that provided estimated values of porosity. Permeability values from aquifer tests and calculated permeability values using data from geophysical logs were plotted and contoured on maps to show regionalized permeability. Correlation between site-specific permeability values from oil-reservoir tests and calculated permeability values using geophysical logs was poor. Differences of two orders of magnitude were not unusual because site-specific data represent small sections of rock, whereas geophysical logs may represent the entire section of the geohydrologic unit.

Geohydrology of the study area is described on the basis of two subdivisions referred to as the Plains subregion and the Ozark subregion. In the Plains subregion, six regional geohydrologic units have been defined, and in the Ozark subregion, three regional geohydrologic units have been defined. Geohydrologic units that were studied in detail were the Great Plains aquifer system, the Western Interior Plains confining system, which is common to both the Plains and Ozark subregions, the Western Interior Plains aquifer system, and the Ozark Plateaus aquifer system.

Hydraulic-head distribution indicates that flow in the Western Interior Plains and Great Plains aquifer systems is generally west to east and southeast in the regional saline-water system. Hydraulic head in the Ozark Plateaus aquifer system indicate nearly radial outflow from the Ozark Uplift toward the Plains subregion, the Missouri River, the Mississippi River, and the southern boundary of the study area. Recharge in the Plains subregion occurs at the western boundary near the Rocky Mountains and Black Hills Uplift and through some outcrop areas of the Great Plains aquifer system in southeastern Colorado and north-central Kansas. Comparatively large quantities of water are recharged in the Ozark subregion through the outcrop areas. Upward movement from both freshwater and saline-water flow systems probably occurs near the saltwater-freshwater transition zone in western Missouri, southeastern Kansas, northeastern Oklahoma, and northwestern Arkansas.

A numerical ground-water flow model was used to test the conceptualization of flow in the regional aquifers and aquifer systems in the study area. The model used was the U.S. Geological Survey Modular Ground-Water Flow Model, which was designed to simulate three-dimensional movement of ground water of constant density through porous media. The model grid for the study area utilized 28 rows and 33 columns uniformly spaced at 28 mi. Rows are oriented coincident with the principal direction of structural lineaments, N. 35°W. from a point in east-central Kansas. The lineaments are, in general, fractures, and it is believed that the grid orientation reasonably corresponds to the principal hydraulic-conductivity tensor of aquifers in the flow system to be simulated.

Five geohydrologic units in the regional study area are represented in the model as layers. They are:

Layer 1.—High Plains aquifer, glacial drift, and the Great Plains confining system;

Layer 2.—Great Plains aquifer system.

Layer 3.—Western Interior Plains confining system.

Layer 4.—The upper unit in the Western Interior Plains aquifer system in the Plains subregion and the Springfield Plateau aquifer in the Ozark Plateaus aquifer system in the Ozark subregion.

Layer 5.—The lower units in the Western Interior Plains aquifer system in the Plains subregion and the combined Ozark aquifer, the St. Francois confining unit, and the St. Francois aquifer in the Ozark Plateaus aquifer system in the Ozark subregion.

The model was based on a mathematical development for which a constant, uniform density was assumed. However, some deep geohydrologic units contain water with large concentrations of dissolved solids. Because this saline water has greater density than freshwater, consideration was given to the effects of variable density on flow. The effect of variable-density water on flow was simulated using a technique that compensates for variable density in the constant-density flow computation.

Calibration of the model was based on comparison of model-computed hydraulic head to measured or estimated hydraulic head (collectively referred to in the report as "field hydraulic head"). The general areal configuration of computed hydraulic-head values showed little difference between simulations that included the effect of variable density and simulations that assumed constant density. However, some "local" hydraulic-head differences caused computed flow-velocity vectors to indicate significant differences in direction. No direction differences were indicated for model layer 2, but in layers 4 and 5 there were changes primarily in areas of large density differences and very small permeability. Indication of flow-direction changes supports the necessity of considering the effects of variable density.

A major component of the regional steady-state hydrologic budget is potential recharge in the Ozark subregion. A recharge model for the regional study area developed areal recharge potential based on hydrologic properties of the soil, vegetation type, monthly precipitation, and computed monthly potential evapotranspiration for a 30-year period (1951–80). Using a technique of distributing recharge potential within a model cell, the minimum recharge in the Ozark subregion was determined to be about 5 percent of the potential recharge; however, stream-discharge data attributed to ground-water flow for only part of the Ozark subregion indicated that recharge probably exceeds 10 percent of potential recharge and may be as much as 20 percent. Recharge potential distributed within a model cell coupled with cell-by-cell estimates of outflow or inflow because of stream-aquifer interactions gave an estimate of net flow per cell that was used as input to the model. The recharge budget that gave the best-fit model simulation for the whole region was a recharge of about 7,000 ft^3/s, of which 93 ft^3/s were into layer 2, about 2,100 ft^3/s were into layer 4, and about 4,800 ft^3/s were into layer 5. Outflow to rivers and streams for layers 4 and 5 totaled about 3,800 ft^3/s; thus, the net recharge to the region was about 3,200 ft^3/s. Only about 1 percent of the net recharge modeled for the region goes into layer 2.

A sensitivity analyses of the model showed that average differences between model-computed and field hydraulic-head values were most sensitive to changes in hydraulic conductivity. Differences were not very sensitive to changes in leakance between layers because computed hydraulic-head differences between layers 4 and 5 were small. Also, for much of the regional study area, evaporite deposits in the Western Interior Plains confining system were simulated as impermeable in layer 3; thus, layers 4 and 5 were isolated from layer 2. The changes in leakance between layers in and near the Ozark subregion resulted in a small effect because the large outcrop area of the Ozark aquifer (layer 5) offered little area for vertical leakage.

Computed hydraulic heads were moderately sensitive to recharge for the overall regional model. Because of the modeling technique for layer 2, hydraulic-head values were insensitive to changes in recharge; any change was compensated for by the general-head boundary condition. The effects of recharge were major for changes in computed hydraulic head in the Ozark subregion where recharge was applied to cells in layers 4 and 5 that represented outcrop areas of the Springfield Plateau aquifer and the Ozark aquifer. The buried parts of those layers were affected only near the outcrop areas. Therefore, the effects of recharge on the average difference between computed and field hydraulic head were small because differences for the whole region were included when computing the average.

Varying two model characteristics simultaneously, such as hydraulic conductivity and recharge, illustrated a nonunique characteristic of the model calibration. For model layers 4 and 5, average differences between computed and field hydraulic head of less than 1 ft were obtained with recharge rates from about 40 percent of the calibration values and with hydraulic conductivity from about 60 percent of the calibration values to about 160 percent for both values. Varying recharge and leakance values simultaneously produced a double-value relation of differences between computed and field hydraulic head. The relation was symmetrical for layer 4 in that for a recharge of 80 percent of the calibration value, a leakance factor of about 40 percent or 160 percent gave an average difference between computed and field hydraulic head of about 1 ft. A similar relation occurred for layer 5; however, it was somewhat less symmetrical.

The question of discretization errors resulting from the large (28 mi) node spacing of the regional model was investigated. An error analysis compared computed and field hydraulic heads of the regional model with 28-mi node spacing to computed and field hydraulic heads of the Great Plains aquifer system subregional model with 14-mi node spacing. There were no significant differences between model results using the two different node spacings.

Model-computed hydraulic heads indicated regional flow systems that generally conformed to conceptual flow patterns. Variations of flow directions

Harvey, E.J., 1980, Ground water in the Springfield-Salem Plateaus of southern Missouri and northern Arkansas: U.S. Geological Survey Water-Resources Investigations Report 80–101, 66 p.

Hedman, E.R., and Jorgensen, D.G., 1990, Surface- and ground-water interaction and hydrologic budget of the Missouri River Valley aquifer between Yankton, South Dakota, and St. Louis, Missouri: U.S. Geological Survey Hydrologic Investigations Atlas HA–721, 1 sheet, scale 1:1,500,000.

Hedman, E.R., Skelton, John, and Freiwald, D.A., 1987, Flow characteristics for selected springs and streams in the Ozark subregion, Arkansas, Kansas, Missouri, and Oklahoma: U.S. Geological Survey Hydrologic Investigations Atlas HA–688, 4 sheets, scale 1:750,000.

Helgesen, J.O., and Hansen, C.V., 1989, Description of data files compiled for the Central Midwest regional aquifer-system analysis: U.S. Geological Survey Open-File Report 89–42, 37 p.

Helgesen, J.O., Jorgensen, D.G., Leonard, R.B., and Signor, D.C., 1982, Regional study of the Dakota aquifer (Darton's Dakota revisited): Ground Water, v. 20, no. 4, p. 410–414.

Helgesen, J.O., and Leonard, R.B., 1989, Geohydrology of the Great Plains aquifer system, central United States, *in* Swain, L.A., and Johnson, A.I., eds., Regional aquifer systems of the United States—Aquifers of the midwestern area: American Water Resources Association Monograph Series 13, p. 179–190.

Helgesen, J.O., Leonard, R.B., and Wolf, R.B., 1993, Hydrology of the Great Plains aquifer system in Nebraska, Colorado, Kansas, and adjacent areas: U.S. Geological Survey Professional Paper 1414–E, 80 p.

Hitchon, Brian, 1969, Fluid flow in western Canada sedimentary basin, Part 2—Effect of geology: Water Resources Research, v. 5, no. 2, p. 460–469.

Imes, J.L., and Emmett, L.F., 1994, Geohydrology of the Ozark Plateaus aquifer system in parts of Missouri, Arkansas, Oklahoma, and Kansas: U.S. Geological Survey Professional Paper 1414–D, 127 p.

Jensen, M.E., and Haise, H.R., 1963, Estimating evapotranspiration from solar radiation: American Society of Civil Engineers, Journal of Irrigation and Drainage Division, v. 89, no. IR 4, p. 15–41.

Jorgensen, D.G., 1988, Using geophysical logs to estimate porosity, water resistivity, and intrinsic permeability: U.S. Geological Survey Water-Supply Paper 2321, 24 p.

Jorgensen, D.G., Gogel, Tony, and Signor, D.C., 1982, Determination of flow in aquifers containing variable density water: Ground Water Monitoring Review, v. 2, no. 2, p. 40–45.

Jorgensen, D.G., Helgesen, J.O., and Imes, J.L., 1993, Regional aquifers in Kansas, Nebraska, and parts of Arkansas, Colorado, Missouri, New Mexico, Oklahoma, South Dakota, Texas, and Wyoming—Geohydrologic framework: U.S. Geological Survey Professional Paper 1414–B, 72 p.

Jorgensen, D.G., and Signor, D.C., 1981, Plan of study for the Central Midwest regional aquifer-system analysis in parts of Arkansas, Colorado, Kansas, Missouri, Nebraska, New Mexico, Oklahoma, South Dakota, and Texas: U.S. Geological Survey Water-Resources Investigations Report 81–206, 28 p.

Jorgensen, D.G., Signor, D.C., and Imes, J.L., 1989a, Accounting for intracell flow in models with emphasis on water-table recharge and stream-aquifer interaction—I, Problems and concepts: Water Resources Research, v. 25, no. 4, p. 669–676.

____1989b, Accounting for intracell flow in models with emphasis on water-table recharge and stream-aquifer interaction—II, A procedure: Water Resources Research, v. 25, no. 4, p. 677–684.

Karplus, W.J., 1958, Analog simulation solution of field problems: New York, McGraw-Hill, 434 p.

King, P.B., compiler, 1968, Tectonic map of North America, *in* National Atlas: U.S. Geological Survey, 1 sheet, scale 1:7,500,000.

Kinney, D.M., 1966, Geologic map of the United States: U.S. Geological Survey, 1 sheet, scale 1:7,500,000.

Krueger, J.M., and Keller, G.R., 1986, Interpretation of crustal structure from regional gravity anomalies, Ouachita Mountains area and adjacent Gulf Coastal Plain: American Association of Petroleum Geologists Bulletin, v. 70, no. 6, p. 667–689.

Kume, Jack, 1984, Geohydrology and chemical quality of water in Middle and Upper Jurassic and Lower Cretaceous rocks, western Kansas: U.S. Geological Survey Water-Resources Investigations Report 84–4045, 54 p.

Kume, Jack, and Spinazola, J.M., 1982, Geohydrologic data from sandstone aquifers in southwestern Kansas: U.S. Geological Survey Open-File Report 82–868, 112 p.

Leonard, R.B., Signor, D.C., Jorgensen, D.G., and Helgesen, J.O., 1983, Geohydrology and hydrochemistry of the Dakota aquifer, central United States: Water Resources Bulletin, v. 19, no. 6, p. 903–911.

Lobmeyer, D.H., 1980, Preliminary potentiometric-surface map showing freshwater heads for the Lower Cretaceous rocks in the Northern Great Plains of Montana, North Dakota, South Dakota, and Wyoming: U.S. Geological Survey Open-File Report 80–757, 1 sheet.

Lobmeyer, D.H., and Weakly, E.C., 1979, Water in the Dakota Formation, Hodgeman and northern Ford Counties, southwestern Kansas: Kansas Geological Survey Irrigation Series 5, 41 p.

Lohman, S.W., 1972, Ground-water hydraulics: U.S. Geological Survey Professional Paper 708, 70 p.

Luckey, R.R., Gutentag, E.D., Heimes, F.J., and Weeks, J.B., 1984, Digital simulation of ground-water flow in the High Plains aquifer in parts of Colorado, Kansas, New Mexico, Oklahoma, South Dakota, Texas, and Wyoming: U.S. Geological Survey Professional Paper 1400–D, 57 p.

Luckey, R.R., and Stephens, D.M., 1987, Effects of grid size on digital simulation of ground-water flow in the Southern High Plains of Texas and New Mexico: U.S. Geological Survey Water-Resources Investigations Report 87–4085, 32 p.

MacCary, L.M., 1978, Interpretation of well logs in a carbonate aquifer: U.S. Geological Survey Water-Resources Investigations Report 78–88, 30 p.

McDonald, M.G., and Harbaugh, A.W., 1988, A modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Techniques of Water-Resources Investigations, book 6, chap. A1, 548 p.

Merriam, D.F., 1963, Geologic history of Kansas: Kansas Geological Survey Bulletin 162, 317 p.

Miller, R.H., and Rahn, P.H., 1974, Recharge to the Dakota Sandstone from outcrops in the Black Hills, South Dakota: Bulletin of the Association of Engineering Geologists, v. 11, no. 3, p. 221–234.

Moody, D.W., Chase, E.B., and Aronson, D.A., 1986, National water summary 1985, Hydrologic events and surface-water resources: U.S. Geological Survey Water-Supply Paper 2300, 506 p.

Neuzil, C.E., 1980, Fracture leakage in the Cretaceous Pierre Shale and its significance for underground waste disposal: Baltimore, Maryland, Johns Hopkins University, Ph.D. dissertation, 150 p.

Neuzil, C.E., Bredehoeft, J.D., and Wolff, R.G., 1984, Leakage and fracture permeability in the Cretaceous shales confining the Dakota aquifer in South Dakota, *in* Jorgensen, D.G., and Signor, D.C., eds., Geohydrology of the Dakota aquifer—Proceedings of the First C.V. Theis Conference on Geohydrology, Lincoln, Nebraska, October 5–6, 1982: Worthington, Ohio, National Water Well Association, p. 113–120.

Oetking, Philip, Feray, D.E., and Renfro, H.B., 1966, Geological highway map of the Mid-continent region, Kansas, Missouri, Oklahoma, and Arkansas: American Association of Petroleum Geologists, Map 1, 1 sheet, scale 1:1,875,000.

____1967, Geological highway map of the southern Rocky Mountain region, Utah, Colorado, Arizona, and New Mexico: American Association of Petroleum Geologists, Map 2, 1 sheet, scale 1:1,875,000.

Ray, M.B., Brown, Alton, and Schmoker, J.W., 1985, Discussion of empirical relation between carbonate porosity and thermal maturity—An approach to regional porosity prediction: American Association of Petroleum Geologists Bulletin, v. 69, no. 11, p. 2021–2030.

Renfro, H.B., 1973, Geological highway map of Texas: American Association of Petroleum Geologists, Map 7, 1 sheet, scale 1:1,875,000.

Sims, P.K., 1985, Precambrian basement map of the northern Mid-continent, U.S.A.: U.S. Geological Survey Open-File Report 85–604, scale 1:1,000,000, 16 p.

Snedecor, G.W., 1957, Statistical methods (5th ed.): Ames, Iowa, The Iowa State College Press, 534 p.

Spinazola, J.M., and Dealy, M.T., 1983, Hydrology of the Ogallala aquifer in Ford County, southwestern Kansas: U.S. Geological Survey Water-Resources Investigations Report 83–4226, 58 p.

Steinhilber, W.L., and Young, H.L., 1979, Plan of study for the Northern Midwest regional aquifer-system analysis: U.S. Geological Survey Water-Resources Investigations Report 79–44, 20 p.

U.S. Geological Survey, 1966, Ground water in the Cimarron River basin, New Mexico, Colorado, Kansas, and Oklahoma: Lawrence, Kansas, U.S. Geological Survey Open-File Report, 51 p.

Ward, J.R., 1968, A study of the joint patterns in gently dipping sedimentary rocks of south-central Kansas: Kansas Geological Survey Bulletin 191, pt. 2, 23 p.

Warner, L.Q., 1980, Colorado lineament, *in* Kent, H.C., and Porter, K.W., eds., Colorado geology: Denver, Colorado, Rocky Mountain Association of Geologists, Symposium on Colorado, Guidebook, p. 11–21.

Watts, K.R., 1989, Potential hydrologic effects of ground- water withdrawals from the Dakota aquifer, southwestern Kansas: U.S. Geological Survey Water-Supply Paper 2304, 47 p.

Weeks, J.B., 1978, Plan of study for the High Plains regional aquifer-system analysis in parts of Kansas, Nebraska, New Mexico, Oklahoma, South Dakota, Texas, and Wyoming: U.S. Geological Survey Water-Resources Investigations Report 78–70, 28 p.

Weiss, Emanuel, 1982, A model for the simulation of flow of variable-density ground water in three dimensions under steady-state conditions: U.S. Geological Survey Open-File Report 82–352, 59 p.

____1986, Ground-water flow in the Navajo Sandstone in parts of Emery, Grand, Carbon, Wayne, Garfield, and Kane Counties in southeast Utah: U.S. Geological Survey Water-Resources Investigations Report 86–4012, 77 p.