GEOHYDROLOGY OF THE SOUTHEASTERN COASTAL PLAIN AQUIFER SYSTEM IN ALABAMA

REGIONAL AQUIFER-SYSTEM ANALYSIS

U. S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1410-H

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By Michael Planert, John S. Williams, *and* Sydney S. DeJarnette

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FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

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

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

Robert M. Hirsch

Robert M. Hirsch
Acting Director

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[Plates are in pocket]

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CONVERSION FACTORS AND VERTICAL DATUM

Multiply	*By*	*To obtain*
inch (in)	25.4	millimeter
foot (ft)	0.3048	meter
foot per mile (ft/mi)	0.1900	meter per kilometer
foot squared per day (ft^2/d)	0.09290	meter squared per day
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer
gallon per minute (gal/min)	0.0630	liter per second
million gallons per day (Mgal/d)	0.04381	cubic meter per second

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

GEOHYDROLOGY OF THE SOUTHEASTERN COASTAL PLAIN AQUIFER SYSTEM IN ALABAMA

By MICHAEL PLANERT, JOHN S. WILLIAMS, and SYDNEY S. DEJARNETTE

ABSTRACT

The Coastal Plain sediments in southeastern Alabama have been divided into five aquifers: Lisbon, Nanafalia-Clayton, Providence-Ripley, Eutaw, and Tuscaloosa. These aquifers may contain two or more permeable zones that locally could be considered individual aquifers; the zones are sufficiently connected hydraulically (based on water levels in wells and data from geophysical logs), however, that they are considered to function as a single aquifer on a regional scale.

The study area encompasses about 46,500 square miles; its limits were determined by physical and hydrological boundaries of the aquifer system. A finite-difference model was constructed to simulate ground-water flow under both steady-state (predevelopment) and transient (pumping) conditions. Because the model was designed to simulate only regional flow, only major rivers were simulated; smaller streams that are considered part of local- and intermediate-flow systems were not simulated.

The variable aquifer properties used in model calibration were transmissivity, leakance, recharge, and riverbed conductance. The model was calibrated to potentiometric heads derived from historical water-level measurements made prior to initiation of substantial ground-water withdrawal in an area, and to streamflow data.

The predevelopment ground-water flow patterns in updip areas indicated by the simulated potentiometric surfaces agreed well with conceptual flow models developed prior to the study. However, the simulation indicated a different pattern for downdip, confined areas. Specifically, a large inflow of ground water appears to enter the study area from both its eastern and western boundaries.

The approach to the transient simulation was to duplicate the response to pumping from 1965 through 1981. This interval was divided into three pumping periods—1965 through 1970, 1971 through 1975, and 1976 through 1981. The model used four sources of water for simulation of pumping—capture of ground-water flow to rivers, induced flow from outside the model boundaries (head-dependent flux boundaries), capture of water from the uppermost aquifer, and water derived from storage within the aquifer. For the final 2 years of the transient simulation (the duration of the last time step in the model simulation, 1980–81), and with pumpage simulated as 175 million gallons per day, 52 percent was derived from reduced ground-water flow to rivers, 19 percent from outside the model boundaries, 10 percent from the uppermost aquifer, and 19 percent from storage. The relatively large contribution from storage indicates that the aquifer system had not reattained equilibrium conditions as of the end of 1981.

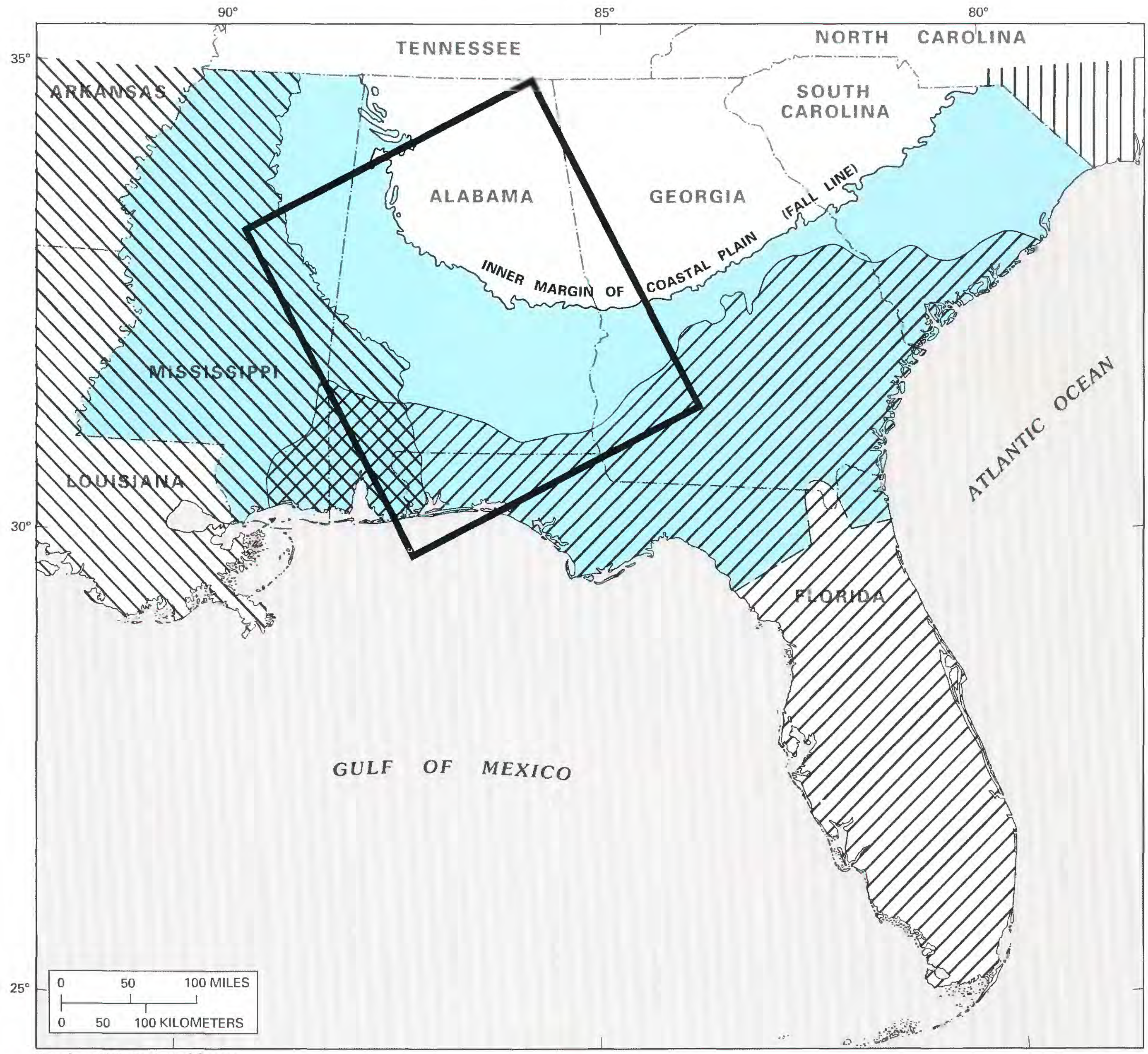
Two additional simulations were made to determine the length of time needed to approach equilibrium. Given the same pumping scheme, an additional pumping period of 50 years was simulated. The rate of water derived from storage in the final time step was 8 percent. Additional drawdowns that may be expected for pumping centers in Alabama after 50 years, with no increase in withdrawal rates, are 3 feet at Dothan and Montgomery and in the Fort Rucker-Enterprise area, and 2 feet at Selma and Demopolis.

Simulated withdrawal rates of 1981 were increased by 37.5 percent at Montgomery, Dothan, and Selma and 10 percent at all other Alabama pumping centers, and the model was again run for an additional 50 years beyond 1981. Under this scheme, the rate of water derived from storage during the final time step was 9 percent of the total. Estimated additional nodal drawdowns at Dothan, Montgomery, Fort Rucker-Enterprise area, Selma, and Demopolis were 37, 33, 17, 12, and 4 feet, respectively.

INTRODUCTION

In 1980, the U.S. Geological Survey began a study of the Southeastern Coastal Plain aquifer system as part of its Regional Aquifer-System Analysis (RASA) program. The Southeastern Coastal Plain aquifer system is composed of the sand aquifers in rocks of Tertiary and Cretaceous age in Alabama, Georgia, Mississippi, and South Carolina. Three other RASA studies bound the Southeastern Coastal Plain study (Sun, 1986)—the Floridan aquifer system study to the south, which addressed carbonate aquifers in rocks of mostly Tertiary age in Alabama, Florida, Georgia, and South Carolina; the Northern Atlantic Coastal Plain RASA study to the north of South Carolina, which addressed the unconsolidated sediments of Tertiary and Cretaceous age from North Carolina to New York; and the Gulf Coast RASA study to the west, which addressed the unconsolidated sediments of Quaternary to Tertiary age in Alabama, Arkansas, Illinois, Kentucky, Louisiana, Mississippi, Missouri, Tennessee, and Texas (fig. 1).

The Southeastern Coastal Plain study area was divided into four subregions, which allowed evaluation of the aquifer system within each State. The subregional



Base from U.S. Geological Survey
National Atlas, 1970

EXPLANATION

Aquifer systems—Where aquifer systems overlap, two or more patterns are shown

Southeastern Coastal Plain

Floridan

Mississippi embayment and coastal lowlands

Northern Atlantic Coastal Plain

Boundary of Alabama subregional study area

FIGURE 1.—Relations among adjacent regional aquifer systems, and location and extent of the Alabama subregional study area. (Modified from Barker, 1986.)

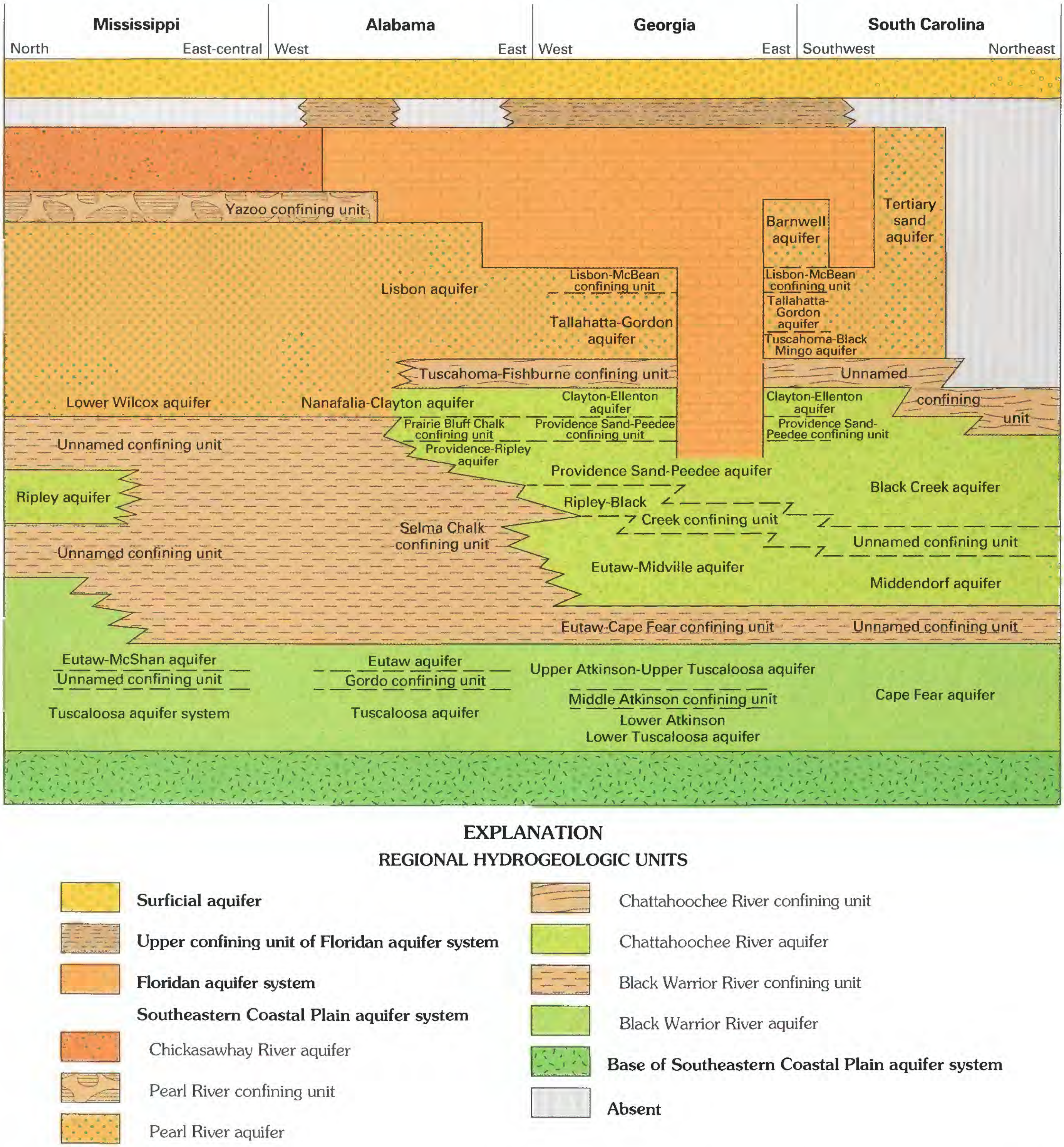


FIGURE 2.—Relations among regional and subregional hydrogeologic units in the Southeastern Coastal Plain aquifer system (from Miller and Renken, 1988).

studies address local subdivisions of the aquifer system in greater detail than the regional study (fig. 2). Data collected and compiled in these subregional studies were then integrated to define the regional framework of geology and ground-water flow for the four-State study area. The findings of the Southeastern Coastal Plain RASA study are described in U.S. Geological Survey Professional Paper 1410 (A-H), which consists of eight chapters. Chapter H (this report) describes the geohydrology of the Alabama subregion, whose extent is shown in figure 1.

The individual chapters of Professional Paper 1410 are as follows:

Chapter	*Subject*	*Authors*
A	Summary	J.A. Miller
B	Geohydrologic Framework	R.A. Renken
C	Regional modeling and hydrology	R.A. Barker and Maribeth Pernik
D	Geochemistry	R.W. Lee
E	South Carolina subregion	W.R. Aucott
F	Georgia subregion	R.E. Faye and G.C. Mayer
G	Mississippi subregion	M.J. Mallory
H	Alabama subregion	Michael Planert, J.S. Williams, and S.S. DeJarnette

PURPOSE AND SCOPE

This report describes the geohydrology of the Southeastern Coastal Plain aquifer system in Alabama. The objectives of the study were to (1) delineate and describe the geohydrologic framework of the aquifer system in Alabama, (2) describe the flow system prior to development, (3) analyze the historical effects of ground-water development, and (4) estimate the potential effects of future ground-water development. The hydrologic analysis required extension of the study area beyond the boundaries of Alabama; most of the area investigated lies within the State, however, and most discussions in this report are concerned with conditions in Alabama.

The study relied, for the most part, on existing data to define the geohydrologic framework. As part of the study, however, two wells were drilled in Alabama where data were lacking. A well inventory and a mass water-level measurement were made to obtain information needed to define regional aquifers and to prepare potentiometric surface maps. Historical data on ground-water pumping were obtained from files of the U.S. Geological Survey and the Geological Survey of Alabama.

GENERAL DESCRIPTION OF THE STUDY AREA

Alabama can be divided into two geologic provinces —the Appalachian province and the Coastal Plain province (Adams and others, 1926). The irregular boundary between these two provinces, a continuation of the "Fall Line" of the Atlantic States, marks the inner margin of the Coastal Plain sediments (fig. 1). The study area for the Alabama subregion encompasses about 46,500 mi^2; its limits were determined by physical and hydrological boundaries of the aquifer system. North of the inner margin of the sediments and in northwestern Alabama, the aquifer system is absent as a result of erosion and removal of the sediments. Beyond the boundaries in Georgia, southern Alabama, and Florida, the rate of ground-water flow in the aquifers is small and may be assumed to be negligible, either at the downdip limit of the freshwater flow system or at ground-water divides between major rivers.

The geologic history of the Southeastern Coastal Plain of the United States has been dominated by repeated transgressions and regressions of the sea. In the study area, the resultant distribution of rock types reflects erosional and depositional events. Sediments studied and discussed in this report range in geologic age from the Cretaceous through the Tertiary Periods. In the study area, Upper Cretaceous rocks were deposited in fluvial, deltaic, or shallow-marine environments, and accordingly range in character from gravel and coarse sand, through completely interbedded sand and clay, to thick sequences of massive chalk. Tertiary rocks generally were laid down in shallow-marine water and are mostly sand, clay, and marl, with local deposits of limestone. The relative resistance of each of these formations to erosion controls the shape of the land surface in the study area, because the geology, physiography, and topography are intimately related.

PHYSIOGRAPHY AND TOPOGRAPHY

The following discussion is taken from Lineback (1973), Moore (1977), and Sapp and Emplaincourt (1975). The study area lies within the East Gulf Coastal Plain section of the Coastal Plain physiographic province. The East Gulf Coastal Plain is underlain by unconsolidated to semiconsolidated sediments that dip gently toward the Gulf of Mexico. Although the entire land surface can be considered a young to mature area of undulating low relief, several resistant formations have formed low ridges or lines of hills called cuestas. Cuestas are formed when gently dipping formations are eroded and a characteristic shape results—a steep face or scarp on one side of the ridge, and a gentle slope that follows the dip of the formation on the other side. These cuestas are separated by lowland areas. The alternating ridges and lowlands are called a belted plain. Altitudes are variable, ranging from sea level at the coast to approximately 800 ft at the inner margin of the Coastal Plain.

Prominent physiographic districts within the Coastal Plain in Alabama are, from north to south, the Fall Line Hills, Black Prairie Belt, Chunnenuggee Hills, Southern Red Hills, Lime Hills, Dougherty Plain, Southern Pine

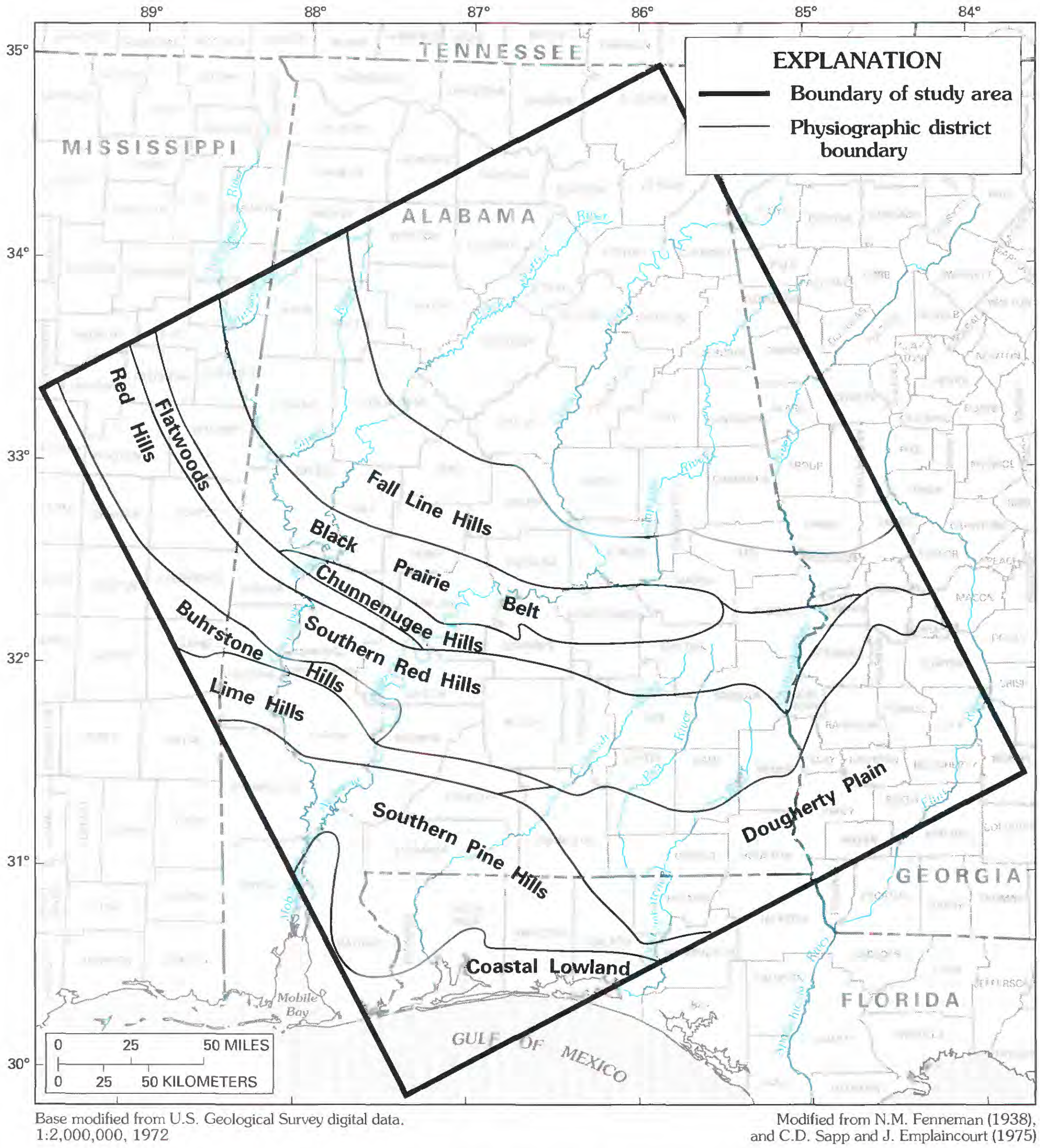
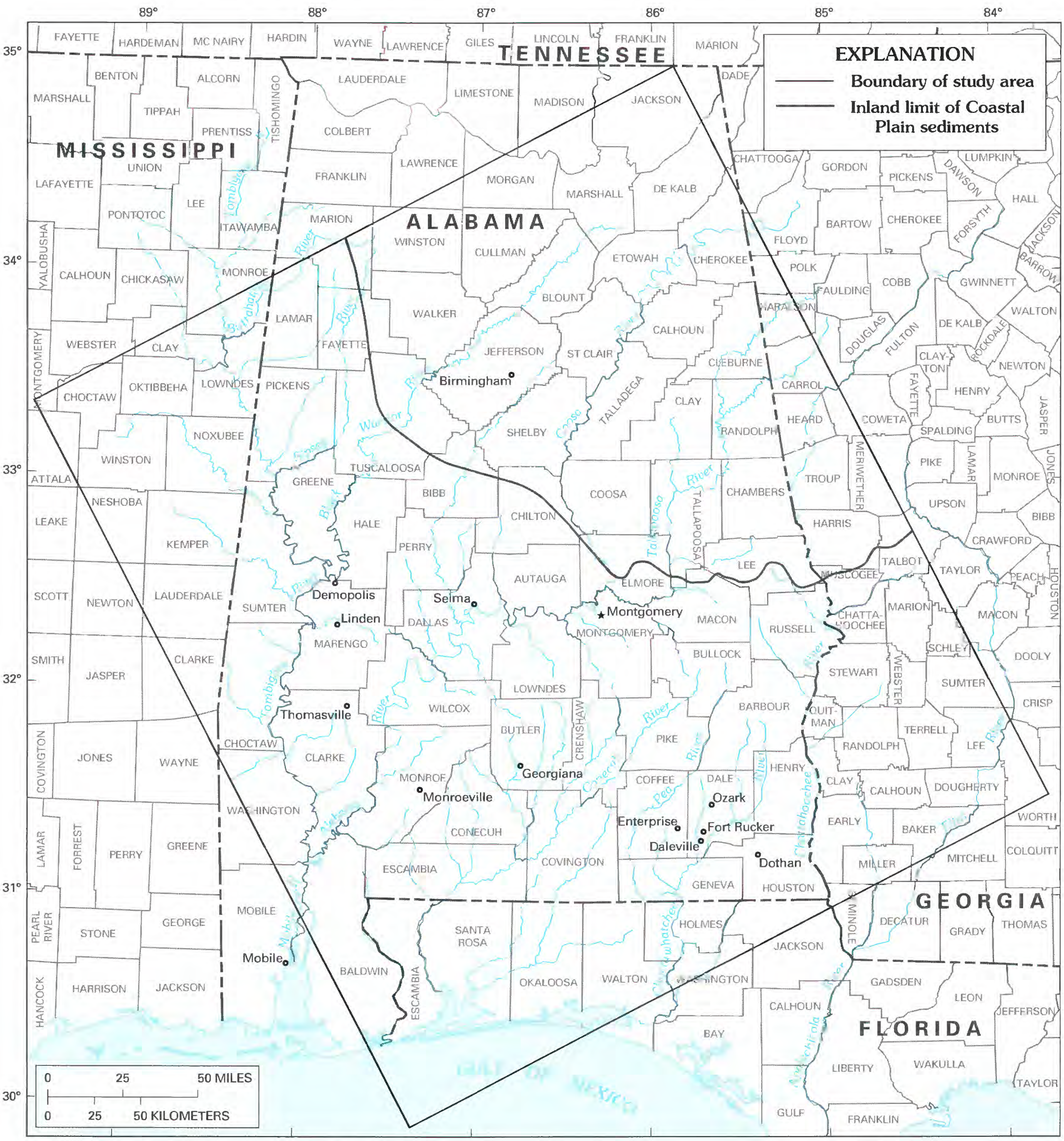


FIGURE 3.—Physiographic districts and features of the study area.

Hills, and Coastal Lowlands (fig. 3). Although no major rivers originate in the Alabama portion of the Coastal Plain, many large rivers, including the Tombigbee, Black Warrior, Alabama, and Chattahoochee, cross the region (fig. 4).

Fall Line Hills District.—The Fall Line Hills district is underlain by sand and clay. The surface is a dissected upland, with a few broad, flat ridges separated by valleys ranging from 100 to 200 ft deep. The district occupies a zone where streams descend from resistant sedimentary and crystalline rocks to the less resistant Coastal Plain deposits. In western Alabama, the maximum width of the district is about 50 mi (fig. 3). Altitudes range from more than 700 ft in northwestern Alabama to about 250 ft along the northern border of the Black Prairie Belt.

Black Prairie Belt District.—The Black Prairie Belt district occupies a 20- to 25-mi-wide, crescent-shaped area encompassing approximately 8,000 mi^2 which extends from near the Georgia border westward through Alabama and then northward through Mississippi (fig. 3). The district, developed mainly on chalk and marl, is characterized by an undulating plain of low relief. In eastern Alabama altitudes are generally more than 250 ft, and in western and central Alabama altitudes are



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 4.—Major rivers draining the Coastal Plain within the study area.

about 200 ft. The Black Prairie Belt is not present in extreme eastern Alabama because the chalk on which it is developed is replaced by clastic sedimentary rocks.

Chunnenuggee Hills District.—The Chunnenuggee Hills district is a series of sand hills and cuestas that in western Alabama is represented by a 5-mi-wide band (fig. 3) called the Ripley Cuesta. In eastern Alabama where the chalk of the Black Prairie Belt intertongues with more resistant clay, siltstone, and sandstone, the Chunnenuggee Hills district widens to about 25 mi. Most of the hills and cuestas of the Chunnenuggee Hills rise 100 to 200 ft above the level of the Black Prairie Belt to the north. The hills extend eastward from near the Mississippi-Alabama border across most of the State.

The Flatwoods subdistrict, a lowland generally about 5 to 8 mi wide, extends from northern Mississippi to just east of the Alabama River. The flat-lying, relatively smooth surface of the Flatwoods has an altitude of about 200 ft. Dark clay of the Porters Creek Formation underlies the subdistrict.

Southern Red Hills District.—The Southern Red Hills district is about 20 mi wide in western Alabama and increases to 50 mi wide in eastern Alabama (fig. 3). One of the unique features of the Southern Red Hills is its nearly flat surface, which slopes gently to the south and west. The altitude at the inner edge of the district is about 600 ft throughout its length. Streams cutting through the plain produce local relief of several hundred feet. Large areas of undissected uplands remain near the southern edge of the belt.

The Buhrstone Hills subdistrict is a cuesta which extends from the Pearl River in Mississippi to about the middle of Alabama. The 10-mi-wide, hilly belt rises 300 to 400 ft above the nearby streams and is considered the most rugged topographic region in the Alabama Coastal Plain. Resistant, siliceous claystone and sandstone underlie the subdistrict.

Lime Hills District.—The Lime Hills district extends from near the Alabama-Mississippi border in a belt 5 to 30 mi wide from southwestern Choctaw County into western Covington County (fig. 3). In some places, the topography of the district approaches that of the Buhrstone Hills. The rugged topography is the result of resistant limestone which underlies the district. In the western part of the Lime Hills, the relief is 200 to 250 ft from valley floors to ridge crests. The eastern part of the belt in Monroe and Conecuh Counties is less rugged and has a relief of 100 to 150 ft.

Dougherty Plain District.—The Dougherty Plain district in southeast Alabama (fig. 3) is a westward continuation of a limestone upland in Georgia. The district is underlain by undifferentiated limestone residuum, bedded sand and clay, and surficial terrace material. Active dissolution of the limestone has resulted in subsurface capture of the smaller streams, especially in extreme southeast Alabama. The topography is that of a low cuesta, more dissected in south-central Alabama than in the southeast. Altitudes range from 100 ft to more than 300 ft, and local relief is generally less than 80 ft. The district is the eastward equivalent of the Lime Hills district.

Southern Pine Hills District.—The Southern Pine Hills district is a cuestalike, elevated, southward-sloping plain (fig. 3). The plain is developed on estuarine deposits to the north and on sand and gravel to the south. Altitudes range from 400 ft in the north to about 100 ft a few miles inland from the Gulf of Mexico. The relief is as much as 250 ft in the northern part, where streams draining eastward to the Tombigbee River and westward to the Alabama River drop to base level over relatively short distances. The southern part of the district has smaller relief and is characterized by rounded hills.

Coastal Lowland District.—The Coastal Lowland district is a flat to gently undulating, locally swampy plain underlain by terrigenous deposits (fig. 3). It includes a mainland plain indented by many tidal streams and fringed by tidal marsh and offshore barrier islands. The landward edge of the district is defined by the base of the Pamlico scarp at an altitude of 25 to 30 ft. The scarp marks the elevation of an ancient shoreline at a higher stand of the sea.

DRAINAGE

The major rivers draining the Coastal Plain in Alabama are the Tombigbee, the Black Warrior, the Alabama, and the Chattahoochee (fig. 4). The Tombigbee River flows within the Black Prairie Belt as it enters Alabama until it is joined by the Black Warrior River at Demopolis, where its course changes to the southwest and it crosses other physiographic districts. The Alabama River is formed by the confluence of the Coosa and Tallapoosa Rivers, and it flows primarily westward within the Black Prairie Belt from Montgomery to Selma. At Selma, it turns and flows southwestward until it joins with the Tombigbee River to form the Mobile River. The Chattahoochee River lies on the border between Alabama and Georgia and flows due south through the Coastal Plain.

Three major rivers start in the Coastal Plain and drain water primarily stored in the Tertiary deposits—the Choctawhatchee, the Pea, and the Conecuh (fig. 4). All three rivers rise in the Chunnenuggee Hills district and flow in a southwesterly direction through the study area.

CLIMATE

Alabama is classified as subtropical humid in climate, with hot, humid summers and relatively mild winters (Trewartha, 1968). Severe cold weather is rare in central and southern Alabama, and freezing temperatures usually do not continue for more than 48 consecutive hours.

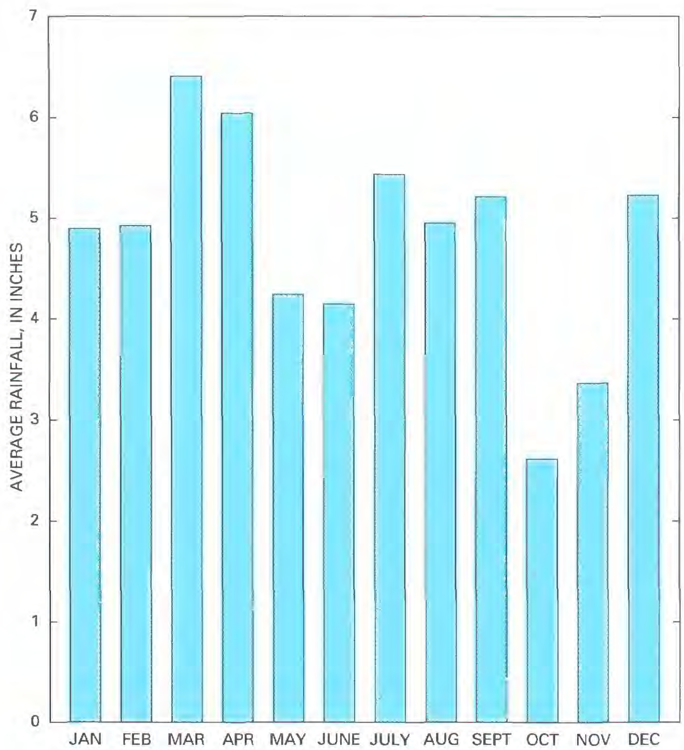


FIGURE 5.—Monthly mean rainfall in the study area, 1951–80. (Data from National Oceanographic and Atmospheric Administration, 1985.)

January is the coldest month, and July is the warmest. Precipitation occurs almost entirely as rain. Summer rainfall is controlled by local thunderstorms moving inland from the Gulf of Mexico, and winter precipitation is controlled primarily by continental air masses moving south from the midwestern part of the continent. Average annual rainfall for stations in the Coastal Plain in Alabama is about 55 in/yr. The range in monthly mean rainfall (fig. 5) for 1951–80 was from 2.6 inches in October to 6.4 inches in March (National Oceanographic and Atmospheric Administration, 1985).

PREVIOUS INVESTIGATIONS

Studies of regional geology of the Coastal Plain of Alabama have been published by many investigators; however, only Barksdale and others (1976) addressed the geohydrology of the entire Alabama Coastal Plain. Most counties in the Coastal Plain have reports that address ground-water availability locally. Reports that discuss the geology of the entire sequence of rocks in the Coastal Plain of Alabama are as follows: Bollman (1968) discussed Cretaceous and Tertiary exposures in west-central Alabama. Copeland and others (1976) discussed Cretaceous and Tertiary faults in southwestern Alabama. Cushing and others (1964) discussed the general geology of the Mississippi embayment. Jones (1967) compiled a guidebook of the geology of the Coastal Plain of Alabama. Monroe (1956b) discussed reverse faulting in the Coastal Plain of Alabama.

The following reports describe the geology of the Cretaceous rocks of Alabama: Applin and Applin (1947) correlated Upper Cretaceous subsurface stratigraphy and structure in Alabama, Georgia, and north Florida. Eargle (1946, 1948a) correlated the pre-Selma Upper Cretaceous in Mississippi and Alabama. He also discussed the Cretaceous of east-central Alabama (1948b) and produced a map of the Selma Group in east Alabama (1950). McGlamery (1944, 1955) described the Upper Cretaceous formations of west-central Alabama. McGlamery and Hastings (1960) described selected outcrops in the Selma Group near Montgomery, Ala. Monroe (1941, 1956a) discussed the Selma Group in Alabama and the pre-Selma Cretaceous (1955, 1964); he also (1946) correlated outcropping Upper Cretaceous formations in Alabama and Texas. Scott (1968) described facies changes in the Selma Group in central and east Alabama. Stephenson (1928) described structural features in the Atlantic and Gulf Coastal Plains. Stephenson and Monroe (1938) reported on the stratigraphy of the Upper Cretaceous Series in Mississippi and Alabama. Copeland (1972) described a field trip of the Upper Cretaceous in central Alabama.

The following reports focus on the geology of the Tertiary sediments: Copeland (1966) reported on facies changes in the Tertiary of Alabama. Hosman and others (1968) discussed Tertiary aquifers in the Mississippi embayment. Moore and Joiner (1969) reported on the subsurface geology of southeast Alabama. Moore (1971) discussed the subsurface geology of southwest Alabama. Various authors have produced geologic maps of individual counties in Alabama. These maps have been published by the Geological Survey of Alabama and are listed in the references at the end of this report.

The following are reports that discuss ground-water hydrology or quality in the Coastal Plain: Avrett (1968) compiled ground-water-quality data for Alabama. Callahan (1964) discussed the yield of aquifers in the river basins of the Southeastern Coastal Plain. Cushing and others (1970) discussed the availability of water in the Mississippi embayment. Cushing (1966) mapped the altitude of the base of freshwater in the Coastal Plain aquifers of the Mississippi embayment. Ellard (1977) mapped fresh and slightly saline ground water in the Coastal Plain of Alabama. O'Rear (1964) discussed ground-water levels in Alabama for 1959 and 1960. Powell and others (1964) discussed water problems associated with oil production in Alabama.

Reports that focus on the geohydrology and ground-water quality of the Tertiary sediments include the following: LaMoreaux (1949) discussed ground water in southeast Alabama. Ivey (1957) discussed the geology and ground water of the Monroeville, Ala., area. Various authors have also produced reports published by the Geological Survey of Alabama on the geology and ground-water resources of individual counties based on data collected in the late 1950's through the 1960's, and these are listed in the references at the end of this report.

The following reports have been produced by this study in Alabama: Davis and others (1983) described a test well in the Nanafalia aquifer in Choctaw County, Ala. Planert and Sparkes (1985) estimated the hydraulic conductivity of the confining bed between the Eutaw and Gordo aquifers in Marengo County, Ala. Williams, Planert, and DeJarnette (1986a, b, c) and Williams, DeJarnette, and Planert (1986a, b) presented maps of the potentiometric surface and ground-water withdrawals for the Eutaw, Lisbon, Providence-Ripley, Tuscaloosa, and Nanafalia-Clayton aquifers, respectively, in Alabama. Davis (1988) defined the geohydrologic framework for the Coastal Plain rocks within the Alabama subregional study area.

ACKNOWLEDGMENTS

The authors thank the landowners, municipalities, and industries in the study area for access to their wells and to data about their wells. The authors are also grateful for the cooperation received from the Geological Survey of Alabama and the Alabama Department of Environmental Management for access to their data.

METHOD OF STUDY

The hydrogeologic framework of the aquifer system was determined primarily from geophysical logs and descriptions of well cuttings and cores on file with the U.S. Geological Survey and the Geological Survey of Alabama. The geophysical logs were also used to calculate the depth to water that contains concentrations of 10,000 mg/L (milligrams per liter) of dissolved solids, which was considered the limit of fresh ground-water flow. Two wells were drilled in areas for which data were lacking. A well inventory and mass water-level measurements were made to obtain data needed to delineate aquifers and to determine the configuration of the potentiometric surfaces of water in the aquifers. Three aquifer tests were conducted in the study area. The values of aquifer properties determined by the tests were used in conjunction with estimates of transmissivity made from specific-capacity tests as initial or rough estimates of the aquifer properties in Alabama. After the data were assembled, a digital model of the aquifer system was constructed to evaluate the initial estimates of the aquifer properties and to refine them through iterative simulations. The model analysis was used to determine the regional ground-water flow patterns for the Coastal Plain sediments in Alabama and to evaluate the effects of pumping on the ground-water flow system.

GEOLOGIC FRAMEWORK

GEOLOGIC SETTING

Coastal Plain sedimentary rocks are present in southern and western Alabama. These sedimentary rocks are underlain by metamorphic, igneous, and sedimentary rocks of Paleozoic and early Mesozoic age. The Coastal Plain sedimentary rocks are the product of cyclic invasion and retreat of ancient seas, and were deposited from Jurassic through Holocene time under marine, marginal marine, and nonmarine conditions. Deeply buried sedimentary rocks of Jurassic age are not considered part of the regional aquifer system being studied. They were excluded because the study focused on the fresh ground-water flow system and the Jurassic rocks are known to contain water having a dissolved-solids concentration of greater than 10,000 mg/L (Renken, 1984).

STRATIGRAPHY

The section of rocks studied in this aquifer evaluation ranges in age from Early Cretaceous through Tertiary (pl. 1). The rocks have a combined thickness of approximately 3,500 ft. Most of the following discussion is taken from Adams and others (1926), Copeland (1966, 1968), Copeland and others (1976), Jones (1967), MacNeil (1946), and Scott (1968).

CRETACEOUS ROCKS

Rocks of Early Cretaceous Age

Rocks of Early Cretaceous age do not crop out in Alabama. At their updip extent in the subsurface, these sediments overlie sedimentary, metamorphic, and igneous rocks of Paleozoic age. Downdip, they overlie sedimentary rocks of Triassic and Jurassic age. Pink nodular limestone fragments and red and green shale, sometimes found near the top of the Early Cretaceous sediments, distinguish them from the massive sand of the overlying Coker Formation of Late Cretaceous age. In places, the upper part of the Early Cretaceous sediments contains massive beds of coarse to very coarse sand and fine gravel (Davis, 1988).

Rocks of Late Cretaceous Age

The Late Cretaceous formations of Alabama crop out in the northern part of the Coastal Plain province in a belt 50 to 75 mi wide and about 275 mi long which trends westward in the eastern part of the State and northwestward in the western part of the State (fig. 6).

The formations are composed of deposits of sand, gravel, clay, and chalk; most were deposited in relatively shallow marine waters, but some were deposited by streams on low plains that bordered the coast. A considerable part of the basal formation of the series was deposited by streams. The chalk was formed as a calcareous muddy ooze on the bottom of a relatively clear sea of moderate depth. These beds dip south toward the Gulf of Mexico or south and west toward the Mississippi embayment at low angles, about 20 to 50 ft/mi. The deposits can be divided into 10 formations, which are, in ascending order, the Coker and Gordo Formations of the Tuscaloosa Group, the McShan Formation, the Eutaw Formation, the Mooreville Chalk, the Blufftown Formation, the Demopolis Chalk, the Ripley Formation, the Prairie Bluff Chalk, and the Providence Sand of the Selma Group.

The Tuscaloosa Group is composed of the Coker and Gordo Formations in western Alabama and is designated the Tuscaloosa Group undifferentiated in eastern Alabama, where the two formations cannot be separated. It crops out along the Fall Line in a band 10 to 30 mi wide (fig. 6). The group consists of irregularly bedded sand, clay, and gravel that generally attain a thickness of about 1,000 ft. The gravel occurs chiefly in the basal beds of the formations near their contact with the underlying rocks (Adams and others, 1926). In the eastern part of the State, the group is more regularly laminated.

The Coker Formation of the Tuscaloosa Group consists of varicolored, unconsolidated beds of clay, sand, and gravel that range from 25 to 100 ft thick and were deposited in a fluvial to deltaic environment. The coarser sand beds and the gravel beds are near the bottom of the formation. The gravel contains large amounts of quartz pebbles derived from conglomerate of the underlying Pottsville Formation of Paleozoic age. The upper part of the Coker consists chiefly of clay and sandy clay containing lenses of micaceous and glauconitic sand. Massive clay beds, as thick as 80 ft, are present in the Coker in Fayette County.

The lower half of the Gordo Formation of the Tuscaloosa Group is predominantly a gravelly sand and is the main gravel-bearing part of the Tuscaloosa Group. It also contains some mottled clay. The upper part of the Gordo consists essentially of lenticular beds of mottled clay, carbonaceous clay, and crossbedded sand that locally contains gravel. The great amount of gravel in the lower part of the Gordo indicates an uplift of the source area in northern Alabama and Tennessee. The gravel and coarse-grained sand probably were deposited on a broad, low-lying alluvial plain. After the initial influx of gravel, sea level rose intermittently, and marine to marginal marine clay accumulated over extensive areas. Occasionally, the clay beds were buried by blankets of sand. Downdip, in south-central and southwestern Alabama, the entire Gordo Formation is marine.

The distinction between the McShan and Eutaw Formations is based on the character of the glauconite found in outcrops, with the glauconite in the Eutaw Formation being coarser and of a darker green color. The formations are difficult to distinguish in the subsurface. Because of this difficulty, the McShan has been mapped as part of the Eutaw Formation in Alabama (Charles Copeland, Geological Survey of Alabama, oral commun., 1986); therefore, in this report the Eutaw Formation includes sediments of the McShan Formation. The Eutaw Formation crops out in a belt parallel to, and immediately south of, the Tuscaloosa Group (fig. 6). The Eutaw is composed predominantly of glauconitic, fine- to medium-grained, micaceous sand, which is commonly crossbedded and was deposited in shallow marine waters. The maximum thickness of the Eutaw is about 400 ft. Throughout the lower half of the formation, the sands are interstratified with subordinate thin laminae and massive layers of clay. The upper 100 ft of the Eutaw Formation in central and western Alabama, which consists chiefly of massive glauconitic sand, with indurated calcareous beds in the uppermost part, has been named the Tombigbee Sand Member. In eastern Alabama, all but the basal part of the Eutaw consists of calcareous carbonaceous clay and sandy limestone beds. The basal part of the unit consists of glauconitic sand and reworked sand, clay, and gravel from the underlying Tuscaloosa Group undifferentiated.

The Selma Group consists largely of calcareous strata between the top of the Eutaw Formation and the base of the Tertiary System. The Selma Group crops out south of, and parallel to, the Eutaw Formation in a belt ranging in width from 25 to 35 mi. The different formations making up the Selma Group are mostly clay and chalk in western Alabama, but they grade eastward by facies change into sand and clay.

The Mooreville Chalk, the basal unit of the Selma Group in central and western Alabama, overlies the Eutaw Formation in unconformable to gradational contact and consists of an unnamed lower member and an upper Arcola Limestone Member. Calcareous clay, marl, and clayey chalk of the unnamed lower member thicken from about 260 ft in western Alabama to about 600 ft in central Alabama. To the east, the member thins and merges or grades into clastic rocks that are part of the Blufftown Formation. The Blufftown, which is exposed along the Chattahoochee River in Russell and Barbour Counties, unconformably overlies the Eutaw Formation and consists of about 500 ft of sand and sandy clay that is partly calcareous and indurated in the upper half of the unit. The Arcola Limestone Member of the Mooreville is about 10 ft thick and consists of thin beds of impure, dense, fossiliferous limestone separated by beds of calcareous clay. The Arcola, along with the lowermost part of the unconformably overlying Demopolis Chalk, also grades eastward into the Blufftown Formation.

The Demopolis Chalk thins from about 500 ft in Sumter County to about 420 ft in Montgomery County, where it becomes silty and micaceous in part and merges with the Cusseta Sand Member of the Ripley Formation and the upper part of the underlying Blufftown Formation. The Demopolis is mostly pure, massively to thinly bedded, fossiliferous chalk, except for the lowermost part of the formation, which consists of thin beds of marly chalk.

The Ripley Formation in Alabama consists of the lower Cusseta Sand Member (pl. 1), restricted to eastern Alabama and ranging up to 200 ft thick, and an upper, unnamed member that extends across the entire State and ranges in thickness from 35 to 225 ft. Where the Cusseta crops out in Russell and Barbour Counties, it consists of medium to coarse sand that locally contains clay pebbles at the base but fines upward and becomes micaceous. Westward, in Montgomery County, the Cusseta Sand Member becomes calcareous and contains interbedded sandy chalk as the horizon changes into the Demopolis Chalk. The upper member of the Ripley consists of sand and calcareous clay, with local, thin beds of fossiliferous sandstone. In western Alabama, the member is finer grained and highly clayey in its lower part, and locally is eroded to about 35 ft in Sumter County; to the east, it is generally coarser grained, crossbedded, and highly fossiliferous in places, and is about 180 to 250 ft thick.

The Prairie Bluff Chalk overlies the Ripley Formation in western and central Alabama and grades eastward into the Providence Sand (pl. 1). The Prairie Bluff is thickest (up to about 125 ft thick) in Lowndes County in central Alabama, and thins to about 10 ft in Marengo County. The Prairie Bluff consists of pure, massive chalk in its lower part, but it grades upward into fine sand and locally is micaceous near the top of the formation.

The Providence Sand consists of the lower Perote Member, which is chiefly thin-bedded silty clay and fine-grained, micaceous, carbonaceous sand, and an upper, unnamed member that is composed of crossbedded fine- to coarse-grained sand and mottled clay. The Providence interfingers with the Prairie Bluff Chalk in Lowndes County and thickens eastward to as much as 300 ft. The Providence is mostly deltaic in outcrop areas but becomes marine downdip.

TERTIARY ROCKS

The Tertiary formations of Alabama consist predominantly of marine clastic rocks that are transitional between equivalent deltaic rocks in Mississippi and carbonate strata of the Florida peninsula. Accordingly, many of the Tertiary units become increasingly calcareous to the east and southeast, and several grade completely into limestone. Except for the uppermost Tertiary formations, which are fluvial in part, Tertiary strata in Alabama were deposited in marine to marginal marine environments.

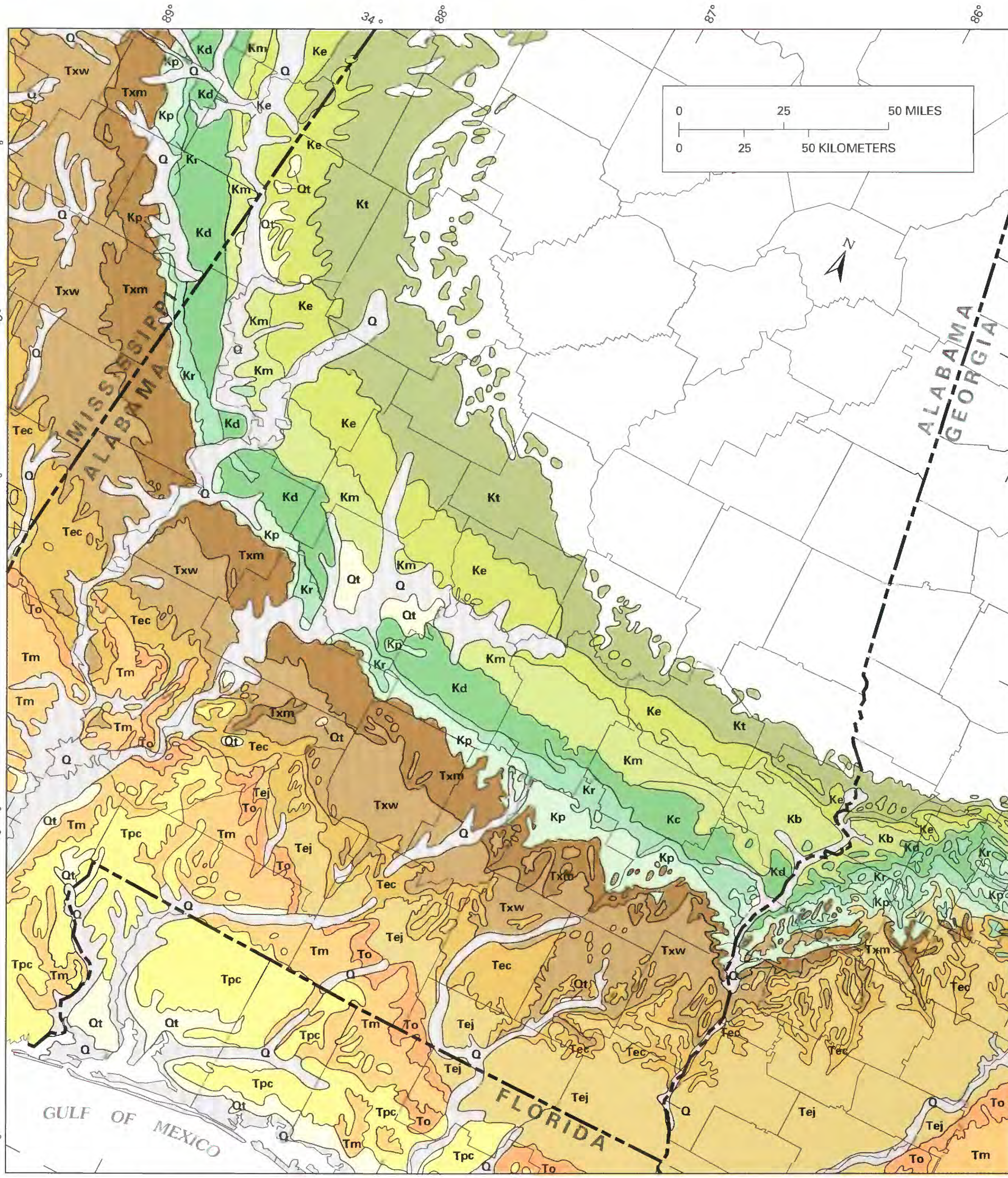
Rocks of Paleocene Age

Paleocene rocks in Alabama have been divided into six formations (pl. 1). From oldest to youngest, they are the Clayton, Porters Creek, Naheola, Nanafalia, Baker Hill, and Tuscahoma Formations. The Clayton Formation unconformably overlies rocks of Cretaceous age, and a basal sand or conglomerate commonly marks the Tertiary-Cretaceous contact. In western Alabama, the Clayton is very thin (5–20 ft thick) and consists of chalky marl and limestone. In Wilcox, Butler, and Crenshaw Counties in central Alabama, the Clayton attains a thickness of 150 ft and is divided into the lower Pine Barren Member, consisting of laminated silt and sand, and the upper McBryde Limestone Member, composed of massive impure chalk and granular limestone that become clayey and sandy in Butler County. The Clayton in eastern Alabama consists of a lower sandy zone, a middle zone of hard, fossiliferous limestone, and an upper zone of calcareous clay or argillaceous limestone.

The Porters Creek Formation consists mostly of massive, nonmarine clay that reaches a thickness of about 450 ft in western Alabama and thins to 150 ft or less in the central part of the State. A bed of glauconitic shell marl ranging in thickness from 5 to 15 ft marks the top of the formation and is called the Matthews Landing Marl Member. East of Crenshaw County, beds that correlate with the Porters Creek, if present, are included in the Clayton Formation because of similar lithologies.

The Naheola Formation is divided into the lower Oak Hill Member, consisting of laminated sandy silt, silty clay, and fine-grained sand, with lignite locally prominent in the uppermost part, and an upper Coal Bluff Marl Member of variable thickness that consists of glauconitic sand and sandy marl. The Naheola is thickest in western Alabama and thins rapidly east of Butler County in the central part of the State where, if present, it is included in the Porters Creek Formation.

Three members make up the Nanafalia Formation in Alabama: the basal Gravel Creek Sand Member, a coarse-grained sand with minor gravel and clay pebbles; an unnamed middle member, which consists of glauco-



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

Modified from Bennison (1975)

FIGURE 6.—Geologic units cropping out in the Coastal Plain of Alabama, Mississippi, and Georgia within the modeled area. (Reprinted from American Association of Petroleum Geologists and published with permission.)

Group	Geologic unit: West	Geologic unit	Geologic unit: East	Map symbol
		Undifferentiated stream and beach deposits		Q
		Undifferentiated alluvium and terrace deposits		Qt
		Citronelle Formation		Tpc
		Catahoula Sandstone		Tm
		Undifferentiated Oligocene Formations		To
	Yazoo Formation	Moodys Branch Formation	Ocala Limestone	Tej
	Gosport Sand	Lisbon Formation Tallahatta Formation		Tec
Wilcox Group		Hatchetigbee and Bashi Formations Tuscahoma Formation Nanafalia and Baker Hill Formations		Txw
	Naheola Formation Porters Creek Formation	Clayton Formation		Txm
Selma Group	Prairie Bluff Chalk		Providence Sand	Kp
Selma Group		Ripley Formation		Kr
Selma Group	Demopolis Chalk		Cusseta Sand	Kd / Kc
Selma Group	Mooreville Chalk		Blufftown Formation	Km / Kb
	McShan Formation	Eutaw Formation		Ke
Tuscaloosa Group		Gordo Formation Coker Formation		Kt

FIGURE 6.—Continued.

nitic sand and sandy marl; and the upper Grampian Hills Member, which is chiefly clay containing minor glauconitic sand. The middle member is locally referred to as the "Ostrea thirsae beds" because of the local abundance of this pelecypod in the member. The Nanafalia Formation crops out in a band across the entire width of Alabama and consists of two prominent fining-upward sequences in the eastern part of the State. In southwestern Alabama, the formation is about 250 ft thick.

The updip equivalent of the Nanafalia Formation in eastern Alabama is called the Baker Hill Formation (Gibson, 1982a). It consists of massively bedded kaolinitic and bauxitic clay and crossbedded micaceous sand with minor amounts of carbonaceous clay.

The Tuscahoma Formation consists chiefly of laminated fine-grained sand and clayey silt, but it also contains fossiliferous glauconitic sand and marl units that are recognizable over long distances. One such unit in the lower part of the formation is the Greggs Landing Marl Member, which is traceable from Choctaw County eastward to Georgia. A second marl unit, the Bells Landing Marl Member, is present near the top of the Tuscahoma, but it is not as extensive as the Greggs Landing Marl Member. The basal beds of the Tuscahoma vary greatly in character, from gravelly glauconitic sand to massive clayey sand to a thick sequence of unconsolidated white sand.

Rocks of Eocene Age

Eocene rocks in Alabama have been divided into eight formations (pl. 1). From oldest to youngest, they are the Bashi, Hatchetigbee, Tallahatta, and Lisbon Formations, the Gosport Sand, the Moodys Branch Formation, the Yazoo Formation, and the Ocala Limestone. The Bashi Formation is a thin marl that formerly was considered a member of the Hatchetigbee Formation but recently has been raised to formation rank (Gibson, 1982b). The Hatchetigbee Formation consists of laminated fine-grained sand and clay that are similar to the beds of the Tuscahoma Formation.

The Tallahatta Formation disconformably overlies the Hatchetigbee and consists of siliceous claystone, called "buhrstone" in old reports, with some beds of glauconitic sand and sandstone. Locally, in Choctaw County, the Meridian Sand Member is recognizable at the base of the Tallahatta. In central and eastern Alabama, the Tallahatta consists largely of sand and sandy clay; downdip, it becomes calcareous. The Lisbon Formation consists chiefly of calcareous, glauconitic sand, marl, and sandy clay, with a few nonmarine sand beds in Choctaw County. Minor carbonaceous clay and crossbedded sand are present in the middle part of the Lisbon. The Lisbon thins downdip and to the east. The Gosport Sand consists of fine- to coarse-grained, glauconitic, fossiliferous sand with interfingering wedges of carbonaceous clay. The Gosport disconformably overlies the Lisbon Formation and is 30 ft thick or less in outcrop. The Gosport Sand is not recognizable east of Monroe County.

The Moodys Branch Formation consists of fossiliferous, calcareous, glauconitic sand and sandy marl where it crops out in western Alabama. In central Alabama, the Moodys Branch consists of sand and marl beds, with glauconitic sandy limestone prominent. The Yazoo Formation lies conformably on the Moodys Branch and is divided into four members in western Alabama (pl. 1): (1) calcareous clay of the North Twistwood Creek Clay Member, (2) calcareous sand of the Cocoa Sand Member, (3) limestone and marl of the Pachuta Marl Member, and (4) calcareous clay of the Shubuta Member. The Yazoo Formation grades eastward into the Ocala Limestone (pl. 1), with the Ocala consisting of fossiliferous, locally glauconitic, sandy limestone and calcareous sand. The Ocala thickens from less than 25 ft at the Tombigbee River to more than 140 ft in eastern Alabama.

Rocks of Oligocene Age

Oligocene rocks in Alabama have been divided into ten formations. From oldest to youngest, they are (1) the Red Bluff Formation, which consists of glauconitic limestone overlain by glauconitic, calcareous clay and silty clay with thin beds of sand; (2) the Forest Hill Formation, a westward equivalent of the Red Bluff that consists of sand and lignitic clay; (3) the Bumpnose Formation, an eastward equivalent of the Red Bluff that consists of a greenish-gray, chalky limestone; (4) the Marianna Formation, which consists of porous chalky limestone and includes glauconitic limestone and calcareous sand in the bottom part in western Alabama; (5) the Mint Spring Formation, a partial equivalent of the Marianna, which consists of fossiliferous, glauconitic sand and clayey sand; (6) the Glendon Formation, an irregularly indurated limestone; (7) the Byram Formation, a sandy, glauconitic, fossiliferous marl; (8) the Bucatunna Formation, a sand and carbonaceous clay; (9) the Chickasawhay Limestone, which consists of glauconitic, soft marl and harder beds of limestone; and (10) the Paynes Hammock Formation, which consists of sand and clay with some beds of fossiliferous marl.

Rocks of Miocene and Pliocene Age

Miocene rocks in Alabama include the Catahoula Sandstone, consisting of sand with minor amounts of clay, and undifferentiated overlying sand and clay strata. The Pliocene deposits consist of crossbedded sand with lenses of clay in the Citronelle Formation.

QUATERNARY DEPOSITS

Quaternary deposits consist largely of alluvial and terrace sands.

STRUCTURE

Structural features in the Coastal Plain of Alabama include the Livingston fault zone in Sumter and Marengo Counties and structural features related to the Mississippi Interior Salt Dome Basin in southwestern Alabama (fig. 7).

The Livingston fault zone interrupts the regional dip of the formations that make up the Selma Group in a long narrow belt extending southeastward through parts of Sumter and Marengo Counties (Monroe, 1941; Monroe and Hunt, 1958; Newton and others, 1961). The strata are broken by a series of parallel horsts and grabens that

FIGURE 7.—Structural features in the Coastal Plain of Alabama. (Modified from Moore, 1971.)

strike generally N. 70° W. and are bounded by high-angle reverse faults. In addition to the Selma Group, the faults probably affect the underlying Eutaw and Gordo Formations, based on evidence of movement of ground water (Gardner, 1981).

Southwestern Alabama occupies the easternmost extension of the Mississippi Interior Salt Dome Basin. Some of the geologic structures observable in Lower Cretaceous or younger sediments in this basin are the result of movement of the underlying Louann Salt of Jurassic age (Kidd and Wilson, 1971). Salt at depth acts as a plastic medium and will move into zones of weakness in response to sediment loading. Structures formed as positive features by salt swells or domes and as collapse-type features, such as grabens, where salt was removed, are present in southwestern Alabama. Salt movement associated with these structures was sporadic, with alternating dormant and active periods. Isopach maps of geologic units overlying the salt reflect the periods of salt activity, showing thickening of the units where subsidence resulted from salt removal and thinning of the units where positive or domal movement occurred.

The most prominent structural features within the salt dome basin in southwestern Alabama are the Hatchetigbee anticline, the Jackson fault, the Klepac dome, the Mobile graben, and the Gilbertown, Pollard, and Bethel fault zones (fig. 7). Other important structures include the domes at Citronelle, South Carlton, Chatom, and McIntosh. A more subtle, less defined feature having possible regional significance is the Wiggins uplift. The domes are related to salt movement. The Wiggins uplift and the Mobile graben and other fault zones may have been formed by other tectonic mechanisms (Miller, 1982).

A complex, north-south-oriented fault system known as the Mobile graben extends from Jackson, Ala., south to Mobile Bay (fig. 7). The Jackson fault is the northernmost fault on the east flank of the graben system. Movement along faults within this system may have resulted in the confluence of the Alabama and Tombigbee Rivers and the formation of Mobile Bay.

In Clarke, Choctaw, and Washington Counties, faults can be observed on the surface, and movement of some faults has been recent enough for the grabens to be expressed as topographic lows. Displacement at the surface on these faults generally ranges from less than 100 ft to about 200 ft. Along Gilbertown fault zone, the faults have a maximum surface displacement of about 150 ft, but in the subsurface, displacement of the Eutaw Formation ranges from 350 ft to 900 ft at depths of between 3,500 and 4,000 ft. Deposition contemporaneous with faulting resulted in increased thicknesses of sediments in the downthrown blocks.

GEOHYDROLOGIC FRAMEWORK

The Southeastern Coastal Plain aquifer system within the RASA study area is composed of sediments that can be divided into a series of aquifers and confining units (pl. 1). The sediments are unconsolidated sand, clay, and gravel and semiconsolidated to consolidated limestone and chalk. The beds dip gently toward the Gulf of Mexico from southwestern Georgia to central Alabama, and from there into eastern Mississippi toward the axis of the Mississippi embayment. Each formation is offlapped coastward by the next younger formation, and their eroded edges are exposed in an updip to downdip succession of older to younger arcuate belts. Updip-to-downdip variation in proportions of sand, clay, limestone, or chalk within the formations was caused by the succession of depositional environments from continental to nearshore to marine. Some formations also exhibit significant lateral lithologic variation.

The Coastal Plain sediments within the Alabama RASA subregional study area are divided into five aquifers, named for the principal water-yielding geologic units that compose them. They are, in descending order, the Lisbon, Nanafalia-Clayton, Providence-Ripley, Eutaw, and Tuscaloosa aquifers (figs. 8–12). The aquifers are separated by five confining units: the Yazoo, Tuscahoma, Prairie Bluff, Selma, and Gordo (pl. 1). Each of the aquifers contains one or more permeable zones that may be considered individual aquifers on a local scale, but that are sufficiently connected hydraulically, based on water levels in wells and geophysical logs, to be considered a single aquifer on a larger scale.

YAZOO CONFINING UNIT

The Yazoo confining unit (pl. 1) consists of clay, limestone, and residuum in the Yazoo Formation and the upper part of the Moodys Branch Formation. Confinement by this unit decreases from west to east, and the underlying Lisbon aquifer may be essentially unconfined east of the Choctawhatchee River in Alabama (Walter, 1976).

LISBON AQUIFER

The Lisbon aquifer includes the lower part of the Moodys Branch Formation, the Gosport Sand, the Lisbon, Tallahatta, Hatchetigbee and Bashi Formations (all Eocene) and the upper sands of the Tuscahoma Formation (Paleocene). One or more of these formations may be absent or may not be part of the aquifer at any one geographical location. The Lisbon aquifer is composed mostly of unconsolidated sand and clay beds, but locally it contains claystone or carbonate rocks. The recharge area of the aquifer extends across the study area in a slightly curving band 20 to 30 mi wide (fig. 8). The Lisbon aquifer, although widespread in extent and a reliable source of water, generally yields no more than 200 to 500 gal/min to large wells.

The Lisbon aquifer, as herein defined, includes both the Upper Wilcox and Lisbon aquifers of Walter (1976). He separated those aquifers on the basis of a layer of clay and claystone within the Tallahatta Formation in western Alabama that he considered an effective confining bed. However, because the clay is absent over much of the study area, and because available data are insufficient to confirm whether the layer is an effective, regionally extensive confining unit, this report considers both of Walter's named aquifers as belonging to the same aquifer, while acknowledging the possibility of further subdivision.

TUSCAHOMA CONFINING UNIT

The Tuscahoma confining unit (pl. 1) is formed by clay beds in the middle part of the Tuscahoma Formation. The degree of confinement provided by the clay beds decreases from east to west, probably allowing vertical flow between the Nanafalia-Clayton and Lisbon aquifers in western Alabama.

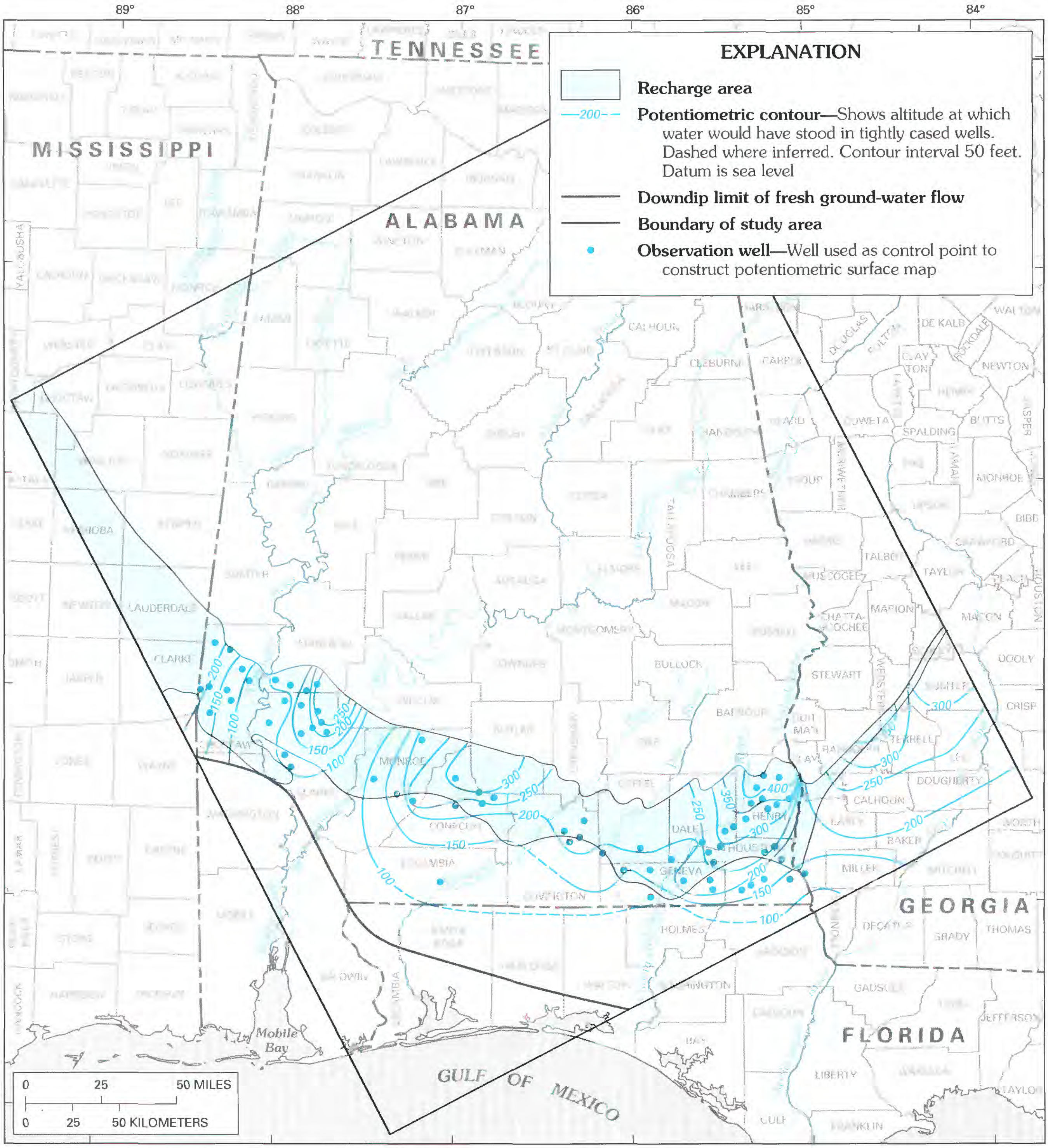
NANAFALIA-CLAYTON AQUIFER

The Nanafalia-Clayton aquifer includes the basal sands of the Tuscahoma Formation, the whole of the Nanafalia, Baker Hill, and Naheola Formations, and the upper part of the Clayton Formation, all of Paleocene age. However, one or more of these formations is absent at any one geographical location. The Nanafalia-Clayton aquifer is composed mostly of unconsolidated sand and clay beds, but it includes carbonate rocks in the Clayton Formation. The recharge area of the aquifer extends across the study area in a slightly curving band that is as wide as 25 mi in eastern and central Alabama, where the aquifer is thickest, and as narrow as 5 mi in eastern Mississippi and western Georgia, where the thickness of water-bearing strata is not as great (fig. 9).

The Nanafalia-Clayton aquifer is one of the more productive aquifers of the Alabama Coastal Plain sediments, with potential well yields of approximately 200 to 2,800 gal/min from large public-supply wells in Alabama (Shamburger, 1976). Highest yields occur in southeastern Alabama and southwestern Georgia.

PRAIRIE BLUFF CONFINING UNIT

The Prairie Bluff confining unit (pl. 1) is formed by the Paleocene Porters Creek and Clayton Formations plus the Cretaceous Prairie Bluff Chalk. The limestone and



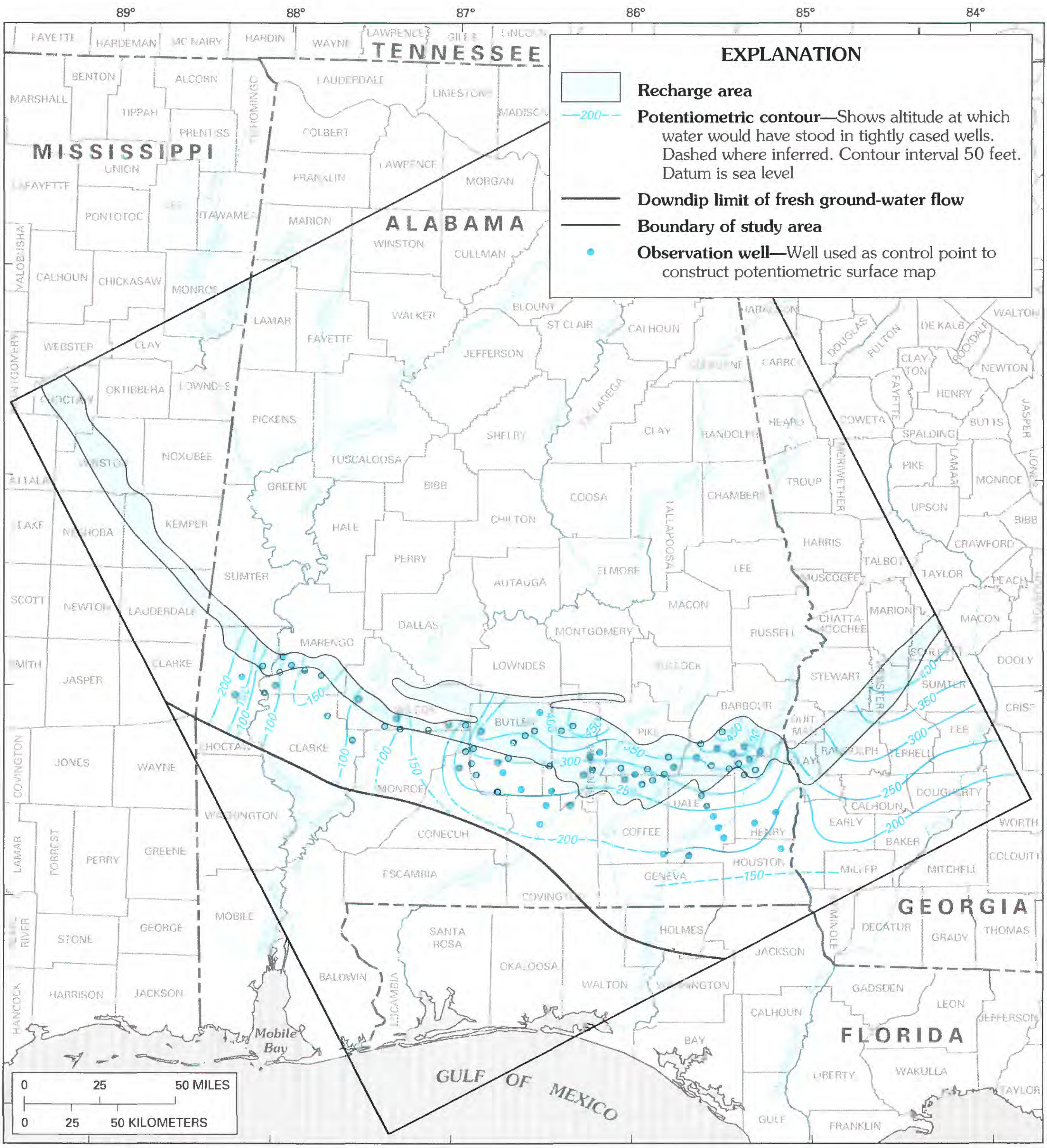
Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 8.—Predevelopment potentiometric surface, area of recharge, and downdip limit of ground-water flow based on dissolved-solids concentrations of 10,000 mg/L in water of the Lisbon aquifer.

clay beds in the confining unit provide varying degrees of confinement, probably allowing leakage between the Providence-Ripley aquifer and the overlying Nanafalia-Clayton aquifer in some areas.

PROVIDENCE-RIPLEY AQUIFER

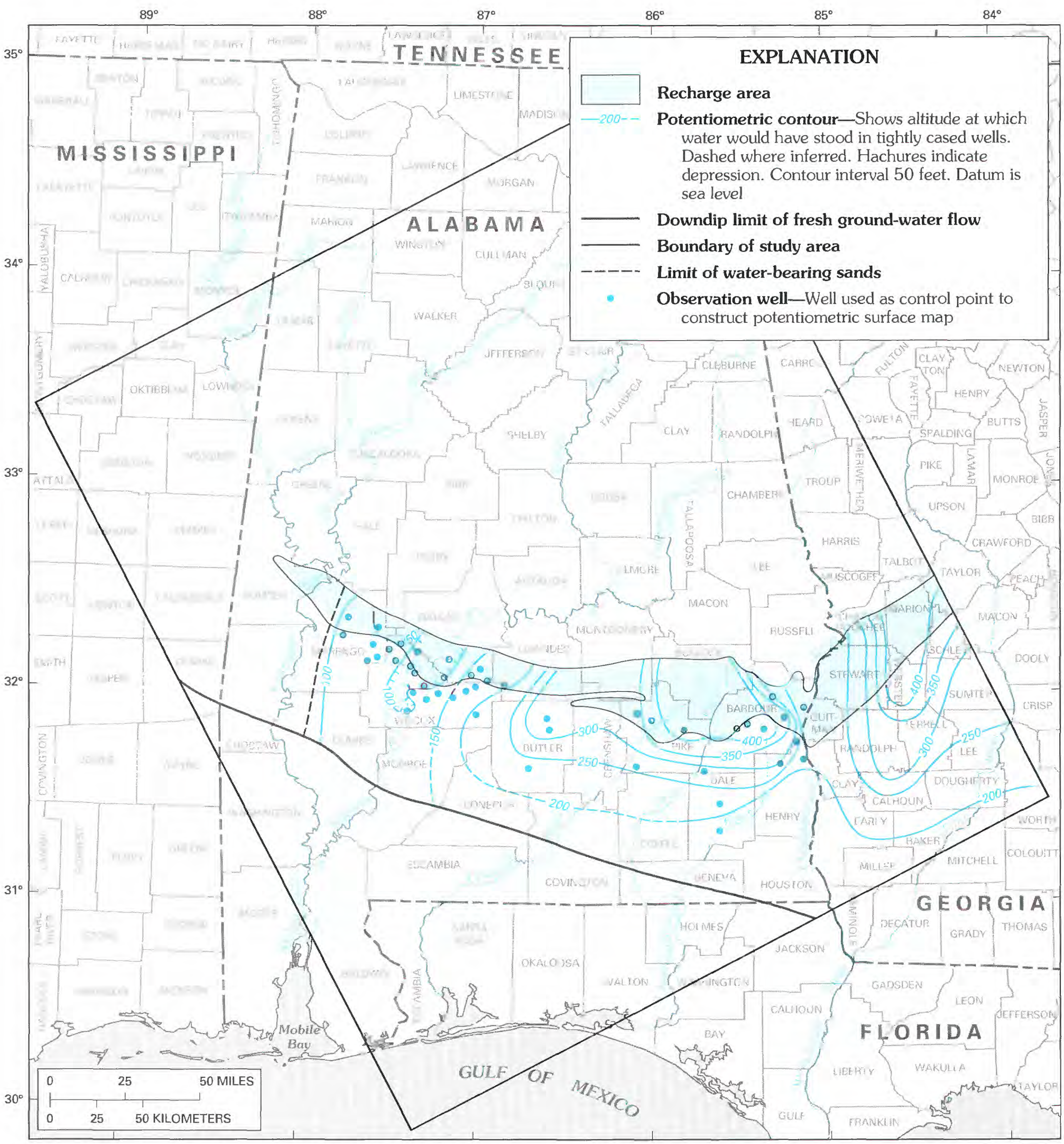
The Providence-Ripley aquifer includes the Providence Sand, the Ripley Formation (including the Cusseta



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 9.—Predevelopment potentiometric surface, area of recharge, and downdip limit of ground-water flow based on dissolved-solids concentrations of 10,000 mg/L in water of the Nanafalia-Clayton aquifer.

Sand Member), and the Blufftown Formation. These formations are composed of consolidated and unconsolidated sand, sandstone, and minor clay beds of Cretaceous age. The recharge area of the Providence-Ripley aquifer extends across the study area in a slightly curving, eastwardly widening band that is as much as 30 mi wide in eastern Alabama (fig. 10). Water-bearing sands in the Providence-Ripley aquifer pinch out in

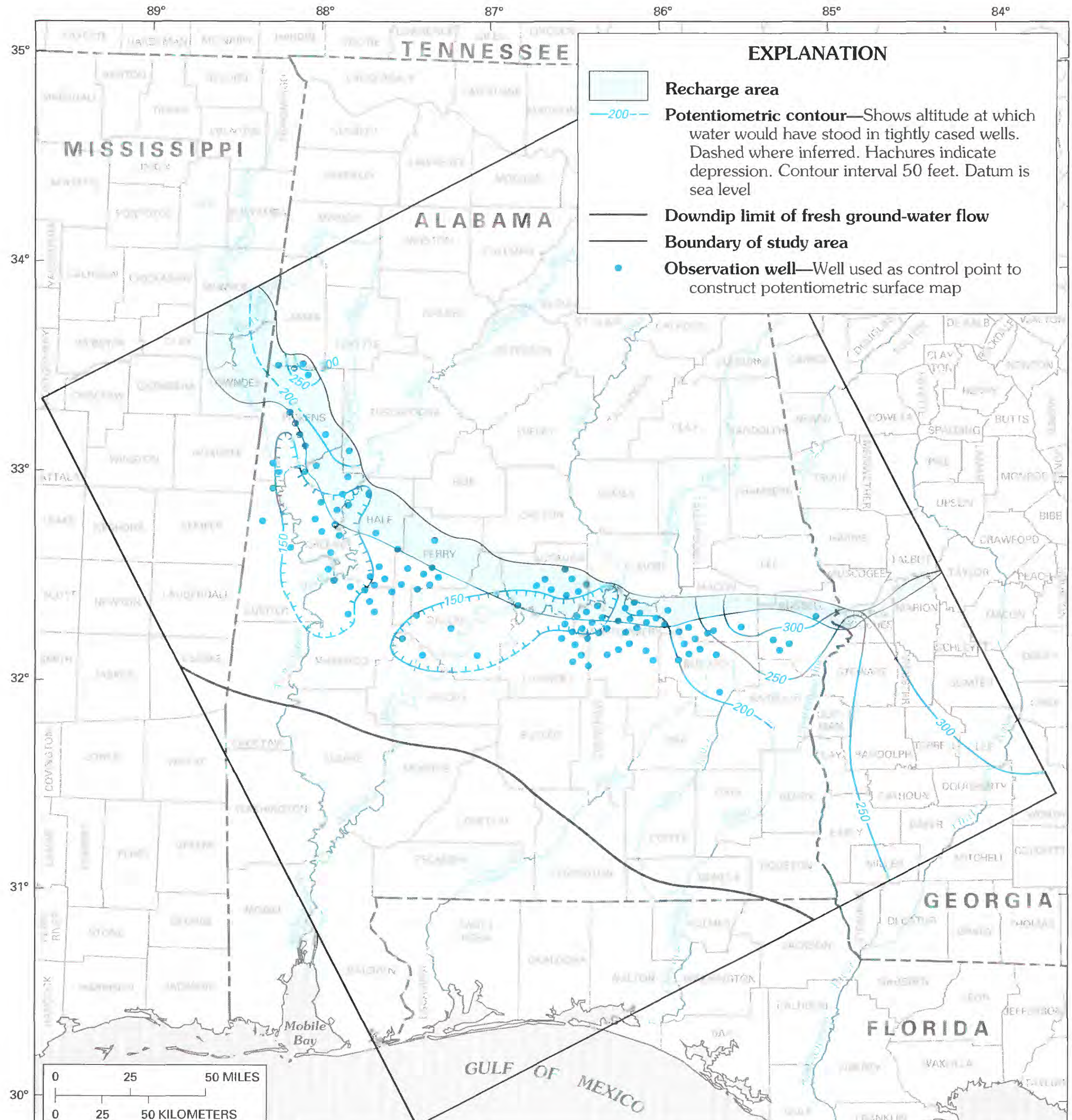


Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 10.—Predevelopment potentiometric surface, area of recharge, and downdip limit of ground-water flow based on dissolved-solids concentrations of 10,000 mg/L in water of the Providence-Ripley aquifer.

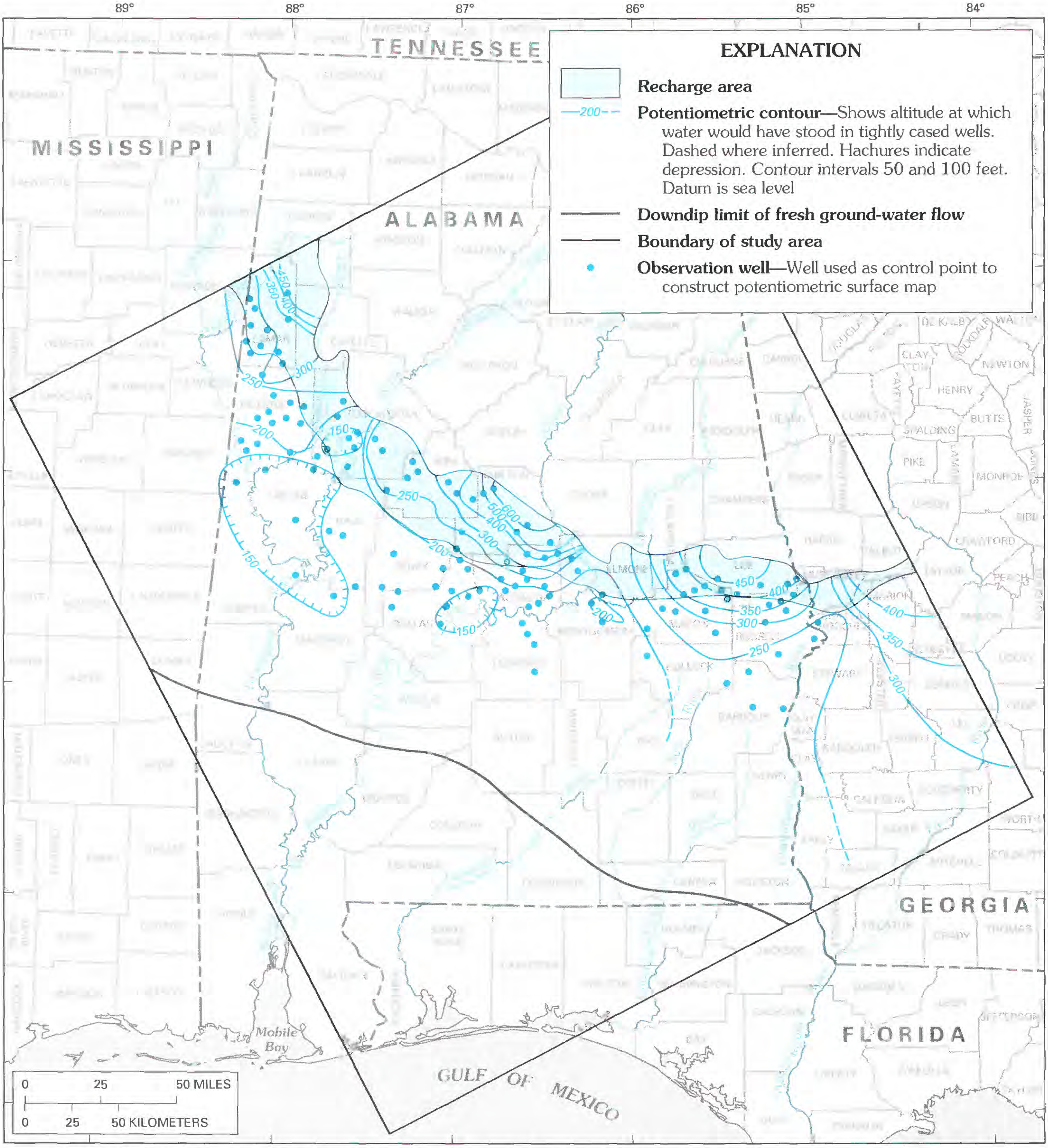
Marengo County, Ala., where the Ripley Formation is composed primarily of clay and the Providence Sand is replaced by the Prairie Bluff Chalk.

The transmissivity of the Providence-Ripley aquifer is low in western Alabama but improves east of Butler County, Ala. This aquifer is very productive in western



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 11.—Predevelopment potentiometric surface, area of recharge, and downdip limit of ground-water flow based on dissolved-solids concentrations of 10,000 mg/L in water of the Eutaw aquifer.



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 12.—Predevelopment potentiometric surface, area of recharge, and downdip limit of ground-water flow based on dissolved-solids concentrations of 10,000 mg/L in water of the Tuscaloosa aquifer.

Georgia. Potential yields for large wells in Alabama range from about 100 to 1,400 gal/min (Lipp, 1976).

SELMA CONFINING UNIT

From western Mississippi to central Alabama, the Cretaceous Demopolis and Mooreville Chalks of the Selma Group make up the Selma confining unit (pl. 1), which overlies the Eutaw aquifer. The chalks range in thickness from about 600 ft to more than 800 ft in the western and central parts of the study area. In eastern Alabama and southwestern Georgia, clays in the upper part of the Eutaw Formation that underlie the Blufftown Formation become the predominant confining bed of the Eutaw aquifer. This confining unit forms, or at least contains, competent confining beds that, in general, hydraulically separate the Eutaw aquifer from the overlying aquifers across most of the study area, except in extreme eastern Alabama and western Georgia, where the confining beds abruptly become thinner.

EUTAW AQUIFER

The Eutaw aquifer includes the Eutaw Formation, which is extensive across the study area, and the McShan Formation in eastern Mississippi and extreme western Alabama. Both are composed of sands, clays, and minor sandy limestones of Cretaceous age. The Eutaw aquifer is composed of a regionally extensive basal sand, and isolated sand beds in its upper part. The recharge area of the aquifer extends across the study area in an arcuate band 2 to 20 mi wide (fig. 11). The updip limit of the Eutaw aquifer is the outcrop of the Gordo Formation and the Tuscaloosa Group. Varying degrees of hydraulic connection exist between the Eutaw aquifer and these underlying formations.

The permeability and thickness of the sands that make up the Eutaw aquifer generally decrease from west to east. From eastern Mississippi to western Alabama, the Eutaw aquifer consists of a lower 300 ft of sand with interbedded clay layers, and an upper 75 to 100 ft of sand interbedded with clay. In eastern Montgomery County, the lower beds thin, and from Macon County eastward, the formation consists of only the Tombigbee Sand Member of the Eutaw Formation. In west-central Alabama, large quantities of water can be obtained from the lower part of the Eutaw aquifer, whereas yields from the Tombigbee Sand Member are relatively low, and the water in many places contains high concentrations of chloride and fluoride. Yields from large wells in the Eutaw aquifer range from about 100 gal/min in Russell County, eastern Alabama, to 1,500 gal/min in Dallas County, west-central Alabama.

GORDO CONFINING UNIT

A marine clay in the upper part of the Gordo Formation separates the Tuscaloosa aquifer from the overlying Eutaw aquifer in the western part of the study area (pl. 1). This confining unit is prominent in eastern Mississippi and western Alabama but is absent in eastern Alabama and western Georgia. However, even though electric logs from eastern Alabama do not show a definite confining unit, the high clay content of the aquifer material in the eastern part of the study area results in some hydrologic separation of the permeable zones.

TUSCALOOSA AQUIFER

The Tuscaloosa aquifer includes the Coker Formation and part of the Gordo Formation from eastern Mississippi to central Alabama and the undifferentiated Tuscaloosa Group in western Alabama (pl. 1). Locally, Lower Cretaceous rocks containing freshwater are also included in this aquifer. All rocks are composed of unconsolidated sediments of Cretaceous age. The recharge area of the Tuscaloosa aquifer extends across the State in a slightly curving band 10 to 25 mi wide having a generally northwest-southeast orientation (fig. 12). The aquifer is underlain by Lower Cretaceous, largely unconsolidated rocks or pre-Cretaceous consolidated rocks that contain water having total dissolved-solids concentrations of greater than 10,000 mg/L. It is overlain by a confining unit that consists of the upper parts of the Gordo Formation and equivalent rocks. Lithologies within the Tuscaloosa aquifer consist of micaceous sands, crossbedded sands, varicolored clays, and gravel beds containing chert and quartz pebbles. The total thickness of water-bearing zones within the aquifer ranges from about 600 to 900 ft in the outcrop area to about 1,100 ft in the subsurface (Kidd, 1976), varying considerably depending on the amount of clay within the section. The clay content within the sand beds increases rapidly east of Montgomery, Ala., with a resulting reduction in the transmissivity of the aquifer. This is evidenced by the lower yields of wells in Macon and Bullock Counties (Scott, 1960b, 1961). Yields of large public-supply wells range from about 500 gal/min in Bullock and Barbour Counties, eastern Alabama, to 1,900 gal/min in Lamar County, western Alabama. A marine clay bed exists over most of the extent of the Coker Formation, and few wells penetrate this clay downdip from the Gordo Formation outcrop, so the water-bearing sands in the lower part of the Coker Formation may not be included in the aquifer.

PREDEVELOPMENT STEADY-STATE FLOW SYSTEM

Recharge to the aquifer system originates as infiltration of precipitation on the outcrops of the various aquifers. Most of this recharge drains to smaller streams, with relatively little residence time in the aquifer, is transpired by vegetation, evaporates, or flows downward through the aquifer and intervening confining units into an underlying aquifer. The balance continues to flow downdip into the confined parts of the aquifer, generally toward lower land-surface elevations. Because ground-water flow is controlled by the forces of gravity, the flow usually moves to the lowest potentiometric elevation in the system. Normally, from the deep, confined parts of the aquifer, ground water discharges to large rivers that act as regional drains or, upon encountering a relatively impermeable downdip boundary, flows vertically into overlying aquifers. However, there are extremely competent confining units within the Southeastern Coastal Plain aquifer system that isolate some of the major rivers from the deep, confined parts of the aquifers. Flow that originated in the recharge areas at the basin divides of these aquifers originally moved downdip; when it encountered the impermeable boundary within the aquifer downdip, the overlying confining unit prevented upward movement and the ground water had to move laterally, approximately along the strike, to a point where the flow could be diverted updip to discharge at a river in the outcrop of the aquifer. This pattern of regional flow is illustrated by the predevelopment potentiometric surfaces of the aquifers.

The predevelopment flow patterns within the five major aquifers in the Alabama Coastal Plain are shown by the potentiometric surface maps in figures 8 through 12. The five aquifers are separable, the lower two from the upper three, on the basis of somewhat different ground-water flow patterns that result from differing hydrogeologic conditions. In all five aquifers, however, the two major factors believed to determine flow patterns are discharge to rivers and a presumed no-flow boundary downdip resulting from a combination of low-permeability rocks and highly mineralized water. For example, the Chattahoochee, Alabama, and Tombigbee Rivers are prominent drains for all the aquifers, and the low gradients and smooth characteristics of the potentiometric surfaces in figures 8 through 12 reflect the damming effect of the decrease in permeability downdip. Steeper gradients would be expected for aquifers having such relatively low transmissivities if appreciable water was being discharged at the downdip limits.

The two lowermost aquifers in the system are the Tuscaloosa and Eutaw aquifers. They are isolated from the overlying aquifers, except updip in the easternmost part of the study area, by the Selma confining unit, which is the most effective confining unit in the study area. Flow patterns within these two aquifers were similar, reflecting a degree of hydraulic interconnection. Within the study area, ground water entered the Tuscaloosa and Eutaw aquifers primarily by direct recharge updip in the outcrop areas, and to a lesser extent by flow from the east and west in the confined zone. Some of the water that entered the system by recharge in the outcrops was immediately drained to nearby streams or was removed by evapotranspiration; the remainder moved downdip into the confined zone. Downdip movement of ground water was eventually impeded by low-permeability rocks and highly mineralized water. Because vertical movement is restricted by the overlying Selma confining unit, the water first flowed laterally along the strike of the aquifer and then flowed updip, discharging to the regional drains in the outcrop area of the Eutaw aquifer.

The regional drains for the Tuscaloosa and Eutaw aquifers are the Tombigbee-Black Warrior and Alabama River systems (which represent the topographically lowermost points for these aquifers in the study area), and to a lesser degree the Chattahoochee River. While the Tombigbee-Black Warrior and Alabama Rivers drain a relatively large part of the study area, the Chatahoochee River receives flow from the Eutaw and Tuscaloosa aquifers only in the recharge areas adjacent to where the river crosses the outcrops of those two aquifers (figs. 11, 12). The potentiometric maps indicate significant depressions in the potentiometric surface around the two major drains, even where the aquifers are confined (figs. 11, 12). These potentiometric lows probably are due to vertical leakage into the Eutaw aquifer from the Tuscaloosa aquifer, and then to the rivers from the Eutaw aquifer through fractures in the Selma confining unit, which underlies the rivers (Gardner, 1981).

The three uppermost aquifers in the study area, the Providence-Ripley, Nanafalia-Clayton, and Lisbon aquifers, likewise have mutually similar flow patterns (figs. 8–10) and are hydraulically connected over much of their mutual extent. The flow patterns within the three aquifers are somewhat different from those in the Eutaw and Tuscaloosa aquifers, for several reasons. First, there is no major overlying confining unit as effective as the Selma confining unit overlying the Eutaw aquifer, so vertical leakage is less restricted. Second, recharge and discharge areas are, in general, relatively larger, and drains exert a more significant immediate influence on ground-water flow. Third, the density of major drains is greater in the upper three aquifers, resulting in generally shorter flow paths.

The sources of water entering the Providence-Ripley, Nanafalia-Clayton, and Lisbon aquifers are essentially the same as those for the Tuscaloosa and Eutaw aquifers; however, no water enters the Providence-Ripley aquifer from the west because the aquifer terminates within the study area. The major drains are the Tombigbee, Alabama, Conecuh, Choctawhatchee, and Chattahoochee River systems. In contrast to the condition in the Eutaw and Tuscaloosa aquifers, the Chattahoochee River is a more significant drain for the upper three aquifers. This is because of (1) the river's relatively lower elevation within the outcrops of those aquifers, (2) the fact that aquifers become more transmissive from west to east in the upper three aquifers (opposite the condition in the lower two aquifers), and (3) the large combined outcrop area in the upper three aquifers. The result is that a significant amount of ground water entering the study area from the east by way of the Providence-Ripley, Nanafalia-Clayton, and Lisbon aquifers discharges to the Chattahoochee River. In contrast, most of the flow entering the Eutaw and Tuscaloosa aquifers from the east in downdip areas flows under the Chattahoochee River and continues northwestward to discharge into the Alabama River (figs. 11, 12).

SIMULATION OF GROUND-WATER MOVEMENT

DESCRIPTION OF THE FLOW MODEL

The model used to simulate the ground-water flow system of the Alabama subregional study area and the entire Southeastern Coastal Plain aquifer system was the U.S. Geological Survey's modular finite-difference ground-water flow model (McDonald and Harbaugh, 1984), which enables a quasi-three-dimensional simulation of flow components within a multilayer aquifer system.

The model simulates the regional flow system by approximating potentiometric surfaces defined by head measurements from deep wells. It also simulates the flow to streams provided by discharge from the deep parts of the aquifers. Simulations were made to approximate both predevelopment conditions and transient conditions since significant pumping began. Aquifer characteristics derived from the initial calibration of the steady-state (predevelopment) model were used as initial conditions for the transient model. The additional stresses provided by simulated pumping in the transient model allowed refinement of the aquifer characteristics. These refined characteristics were then used to recalibrate the steady-state model.

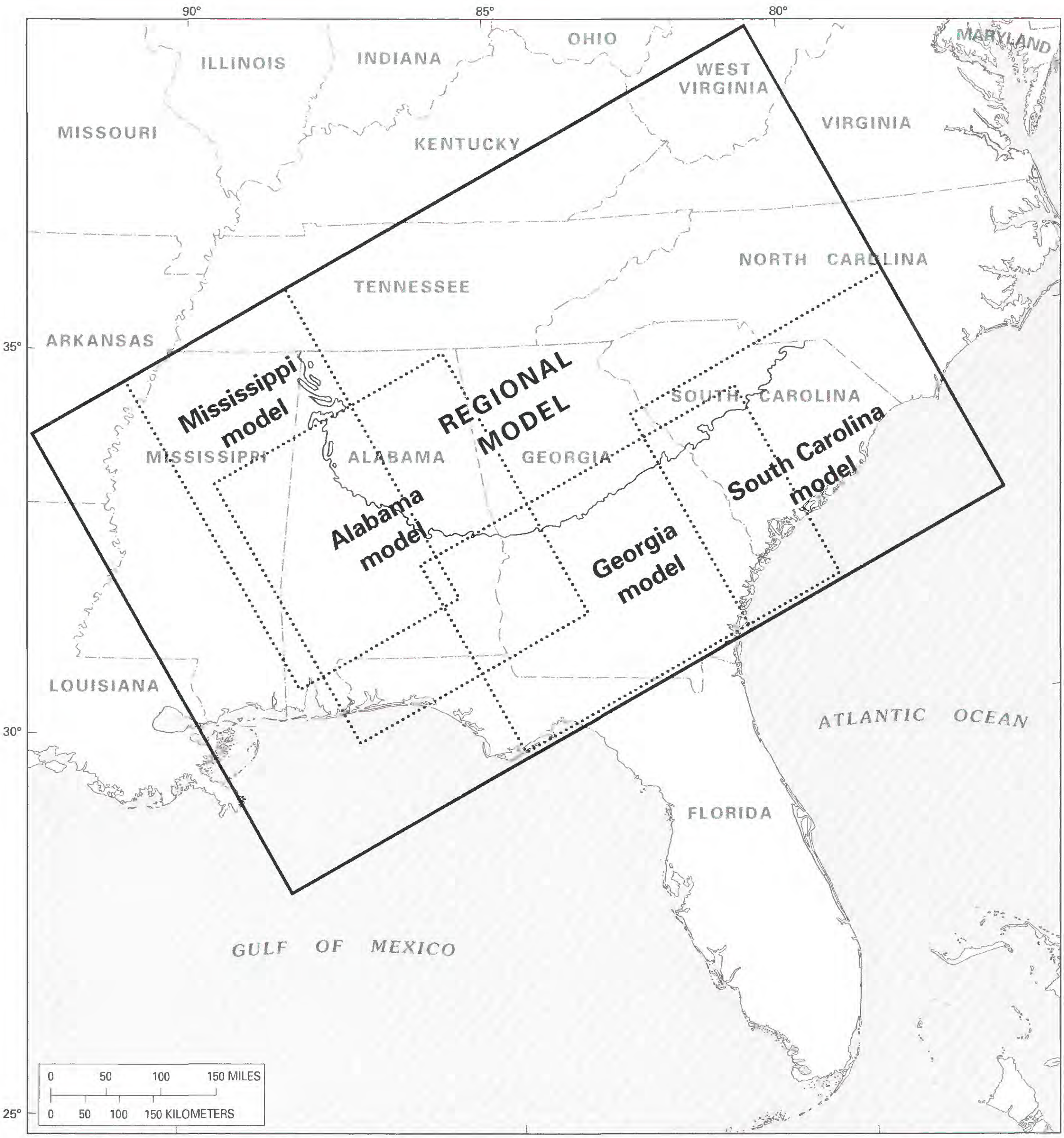
To ensure continuity of the simulated aquifer properties between subregional models, each subregion of the Southeastern Coastal Plain RASA study overlapped the adjacent subregions (fig. 13). Modeling was independent from subregion to subregion, but differences in calibrated values in overlap areas of the various subregional models were resolved by mutual agreement. To provide an overview of the flow system in the Southeastern Coastal Plain, and to ensure the compatibility of the different subregional models, a regional model was constructed using the data from the subregional models (Barker and Pernik, chap. C of this Professional Paper). The regional model had a coarser mesh than the subregional models, each cell of the regional model having four times the cell area of the subregional models.

The horizontal finite-difference grid for the Alabama subregional model consisted of 68 rows and 60 columns, with a node spacing of 4 mi, each cell representing an area of 16 mi^2 (fig. 14). The flow system was modeled in six layers—five active layers (aquifers) and an overlying source-sink layer. Simulated vertical flow between adjacent model layers was controlled by intervening confining units in which vertical leakage could be varied areally. The model layers corresponded to the following named aquifers (A) and confining units (C), which were numbered sequentially from shallowest to deepest, as follows:

Layer number	*Layer name*
A1	Coastal lowlands aquifer system of the Gulf Coast RASA study and the Upper Floridan aquifer (source-sink layer)
C1	Yazoo confining unit
A2	Lisbon aquifer
C2	Tuscahoma confining unit
A3	Nanafalia-Clayton aquifer
C3	Prairie Bluff confining unit
A4	Providence-Ripley aquifer
C4	Selma confining unit
A5	Eutaw aquifer
C5	Gordo confining unit
A6	Tuscaloosa aquifer

Constant-head boundaries, head-dependent flux boundaries, no-flow boundaries, and river nodes were applied for each model layer, as shown in figures 15 through 20. Boundary conditions for the model were chosen to coincide as closely as possible with assumed no-flow boundaries, with ground-water divides, or with the estimated water-table head in the overlying aquifer, represented by the source-sink layer (layer A1) (fig. 15).

The updip limit of each aquifer and the downdip limit of freshwater were modeled as no-flow boundaries, with the downdip limit based on either a delineated line, where water in the aquifer has a dissolved-solids concentration of 10,000 mg/L, or an extreme reduction in permeability (figs. 16–20). The northeast model boundary approximately coincides with the ground-water divide between the Flint and Chattahoochee Rivers. It was designated a constant-head boundary for the steady-state model, and a head-dependent flux boundary for the transient model. The southeast boundary approximately coincides with the downdip limit of the Providence-



Base modified from U.S. Geological Survey National Atlas, 1970

EXPLANATION

——— Boundary of regional model

········· Boundary of subregional model

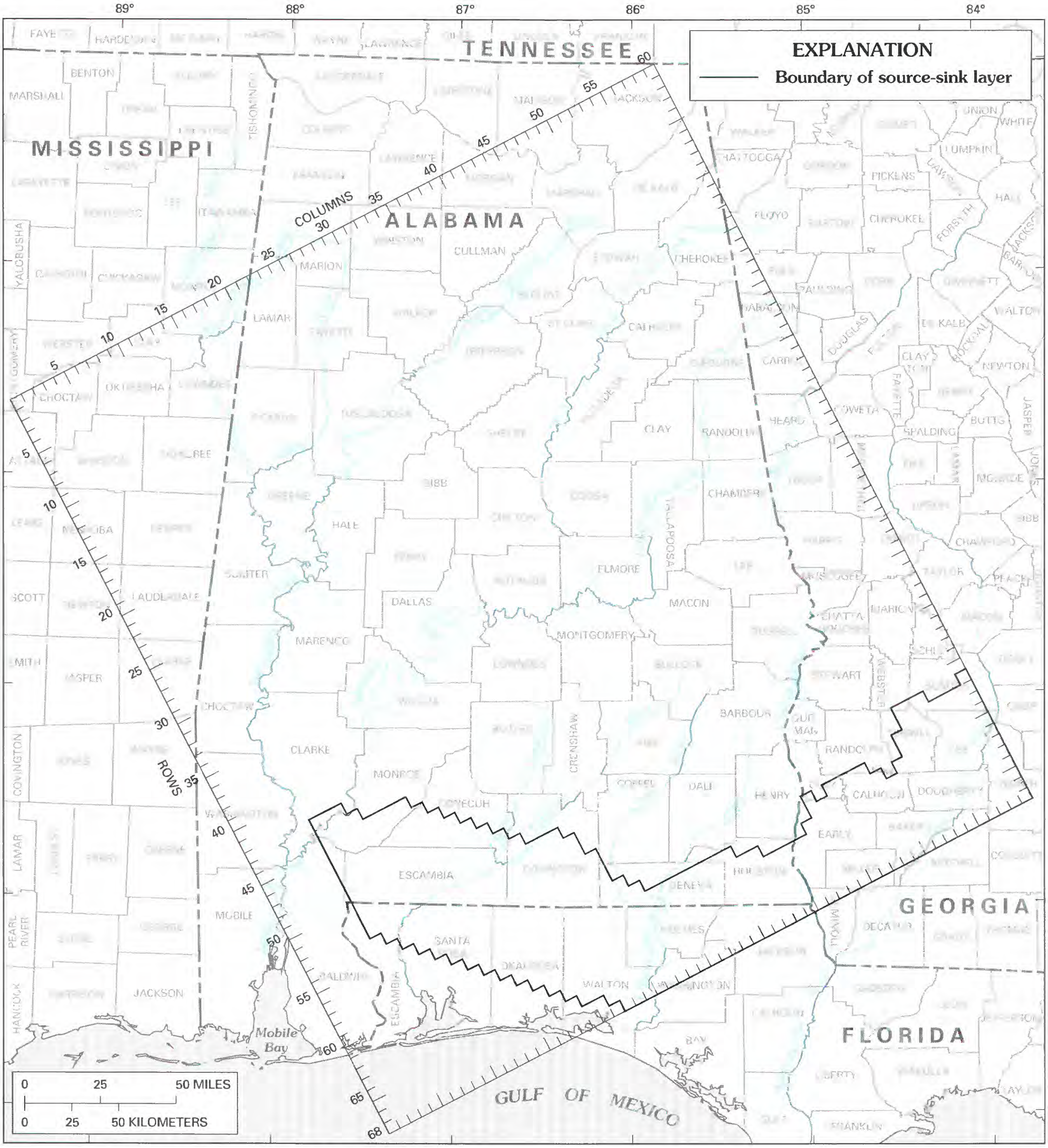
——— Inner margin of Coastal Plain

FIGURE 13.—Relations among regional and subregional model areas covering the Southeastern Coastal Plain aquifer system. (From Miller and others, 1986.)

Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 14.—Model grid, showing boundary of simulated flow region and river nodes.

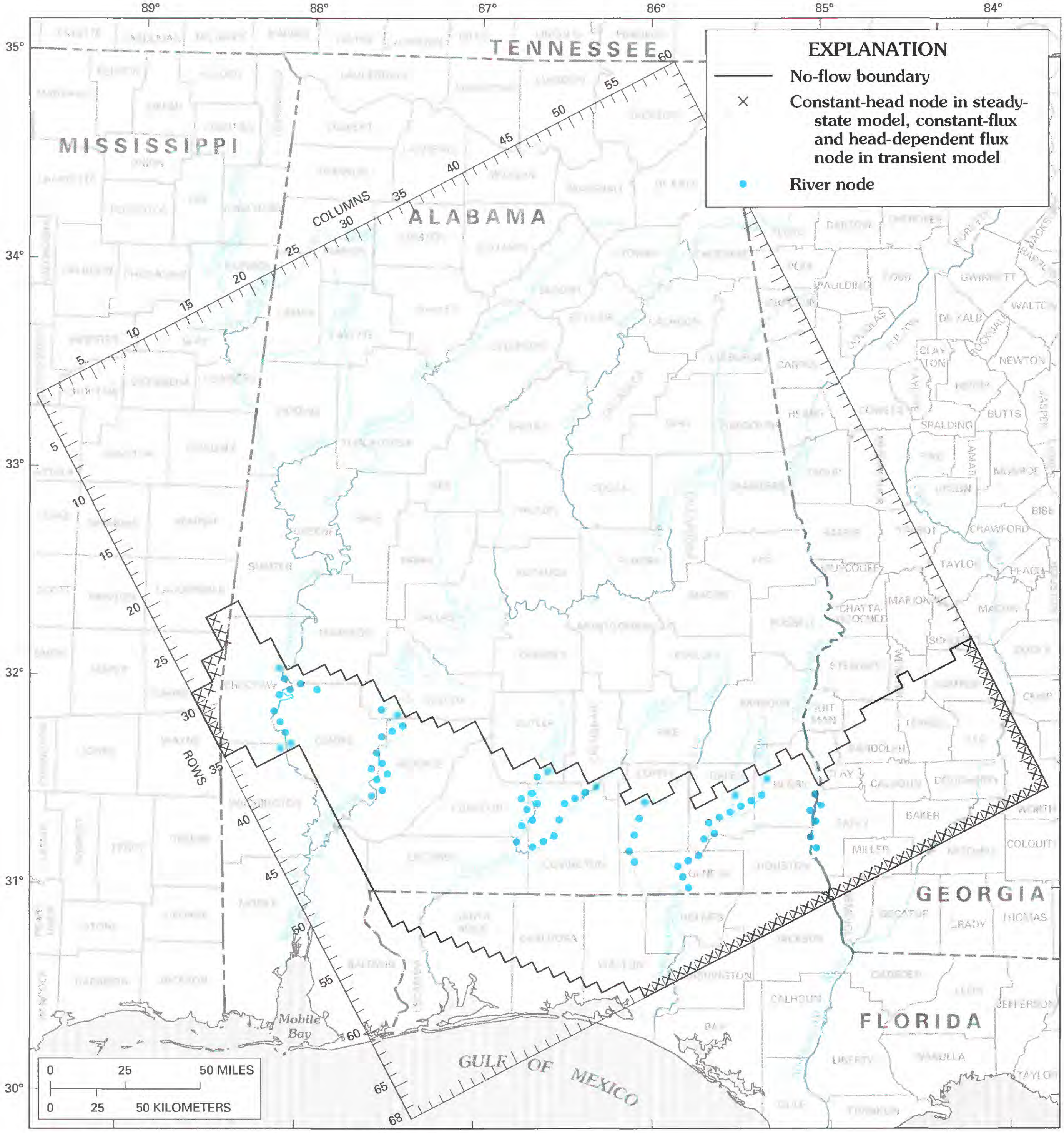
Ripley, Eutaw, and Tuscaloosa aquifers (model layers A4, A5, and A6) and was designated no-flow in those layers. The Lisbon and Nanafalia-Clayton aquifers extend beyond the southeast boundary. Accordingly, this boundary was designated constant head in model layers A2 and A3 for the steady-state model, and a head-dependent flux boundary for the transient model. The northwest model boundary, which approximately



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 15.—Boundary of model layer A1 (Upper Floridan aquifer and coastal lowlands aquifer system). All nodes represent constant head.

coincides with a ground-water divide near the Buttahatchee River, was designated a constant-head boundary for the steady-state model, and a head-dependent flux boundary for the transient model. No natural hydrologic boundaries are present near the southwest limits of the model, so the simulation was extended sufficiently westward to include the area of the aquifer system drained by the main stem of the Tombigbee River. The

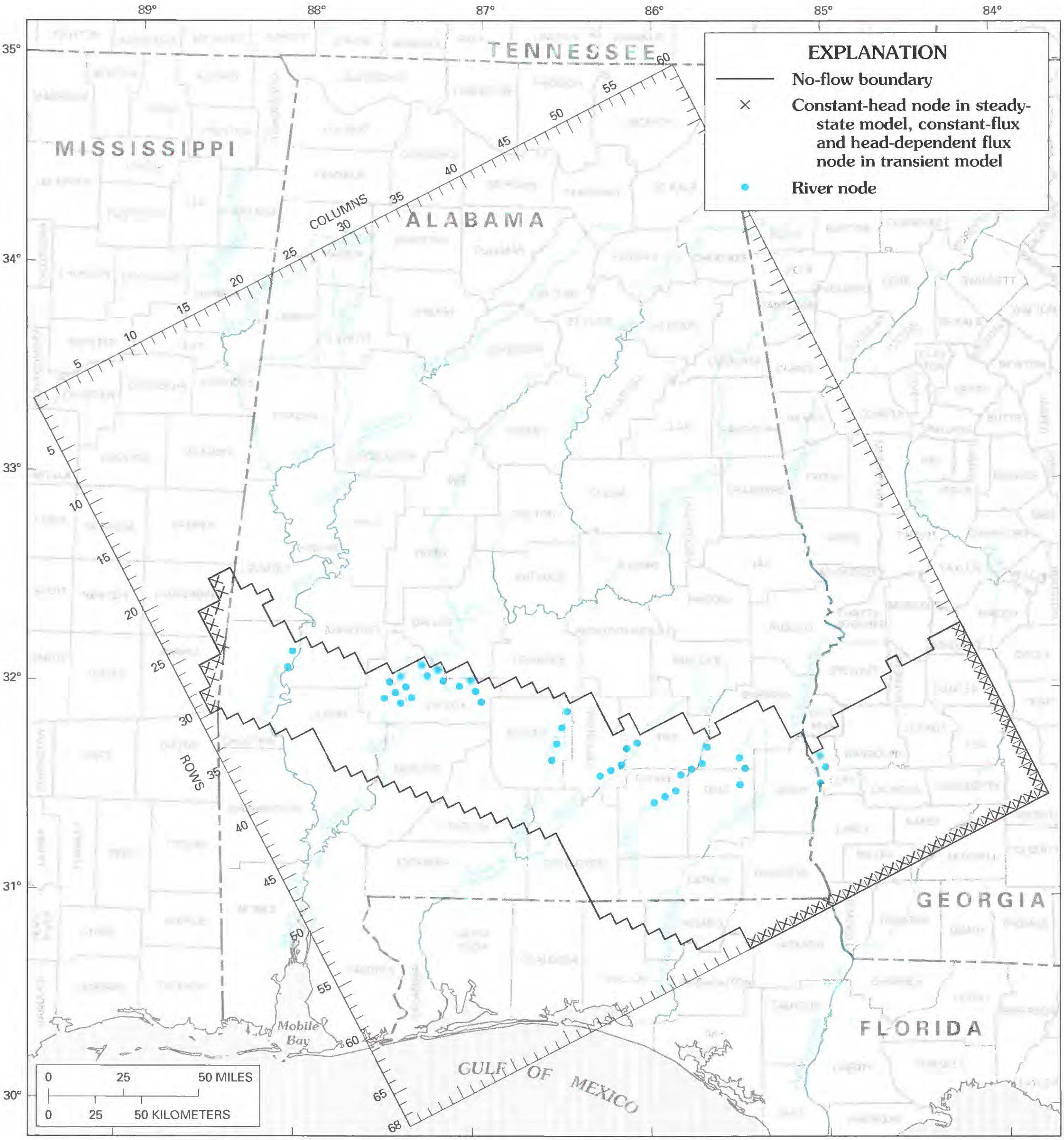


Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 16.—Boundary conditions for model layer A2 (Lisbon aquifer).

resultant boundary was designated constant head in the steady-state model, and head-dependent flux in the transient model.

Because water-table conditions are present in the parts of the coastal lowlands aquifer system of the Gulf Coast RASA study and the Upper Floridan aquifer that lie within the study area, it was assumed that the head in the source-sink layer A1 would remain constant throughout the steady-state and transient simulations. The ratio of the specific yield of the coastal lowlands aquifer



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 17.—Boundary conditions for model layer A3 (Nanafalia-Clayton aquifer).

system and Upper Floridan aquifer to the storage coefficient of the Lisbon aquifer is probably about 1,000 to 1. This means that head declines of at least 1,000 ft in the Lisbon aquifer would be necessary to produce a 1-ft decline in those overlying sediments. Because head declines in most areas of the Alabama Coastal Plain aquifer system are less than 100 ft, the elevation of the water table in the overlying sediments should be only slightly affected by pumping from the underlying aquifers. Therefore, layer A1 was designated a constant-

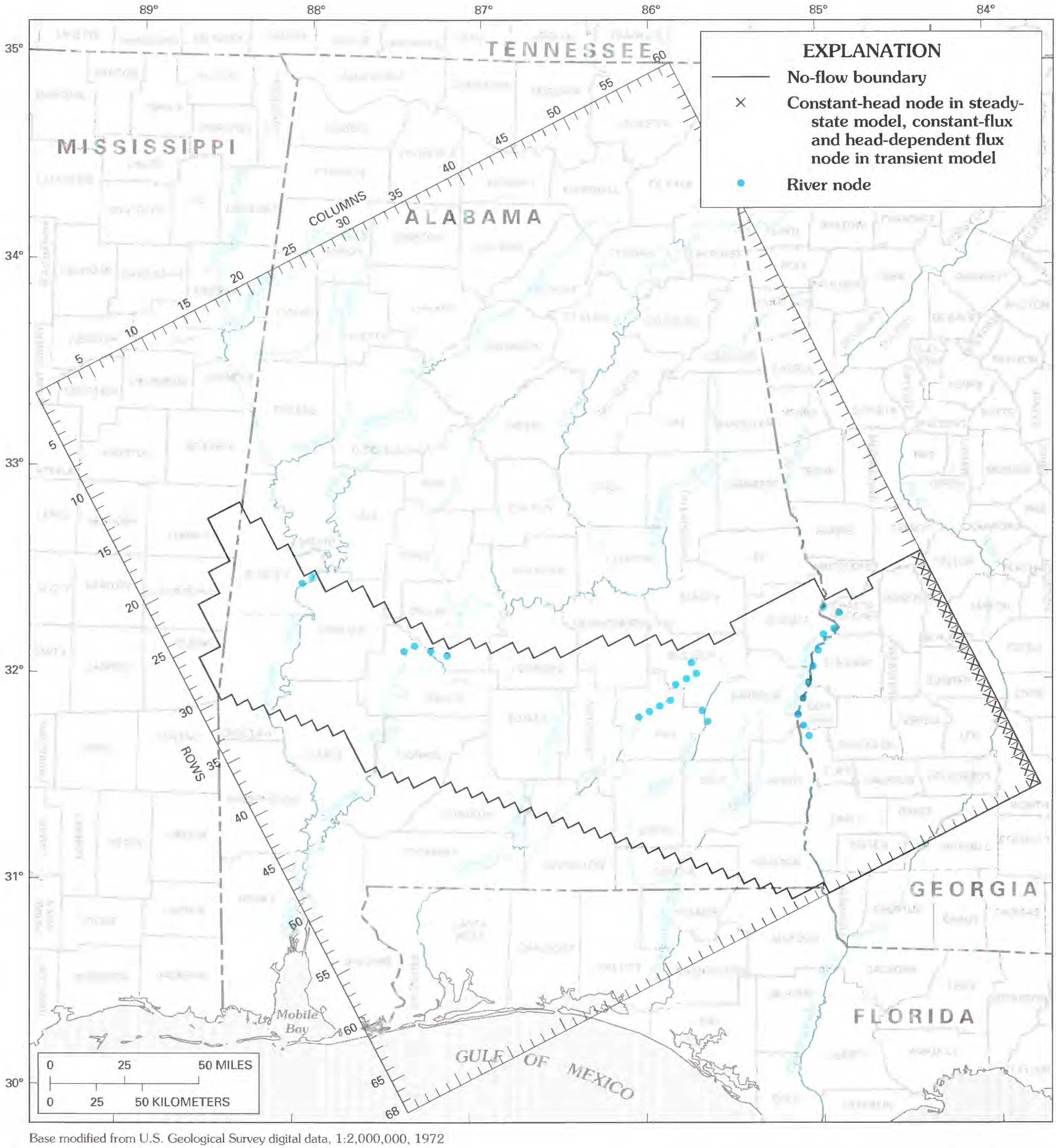
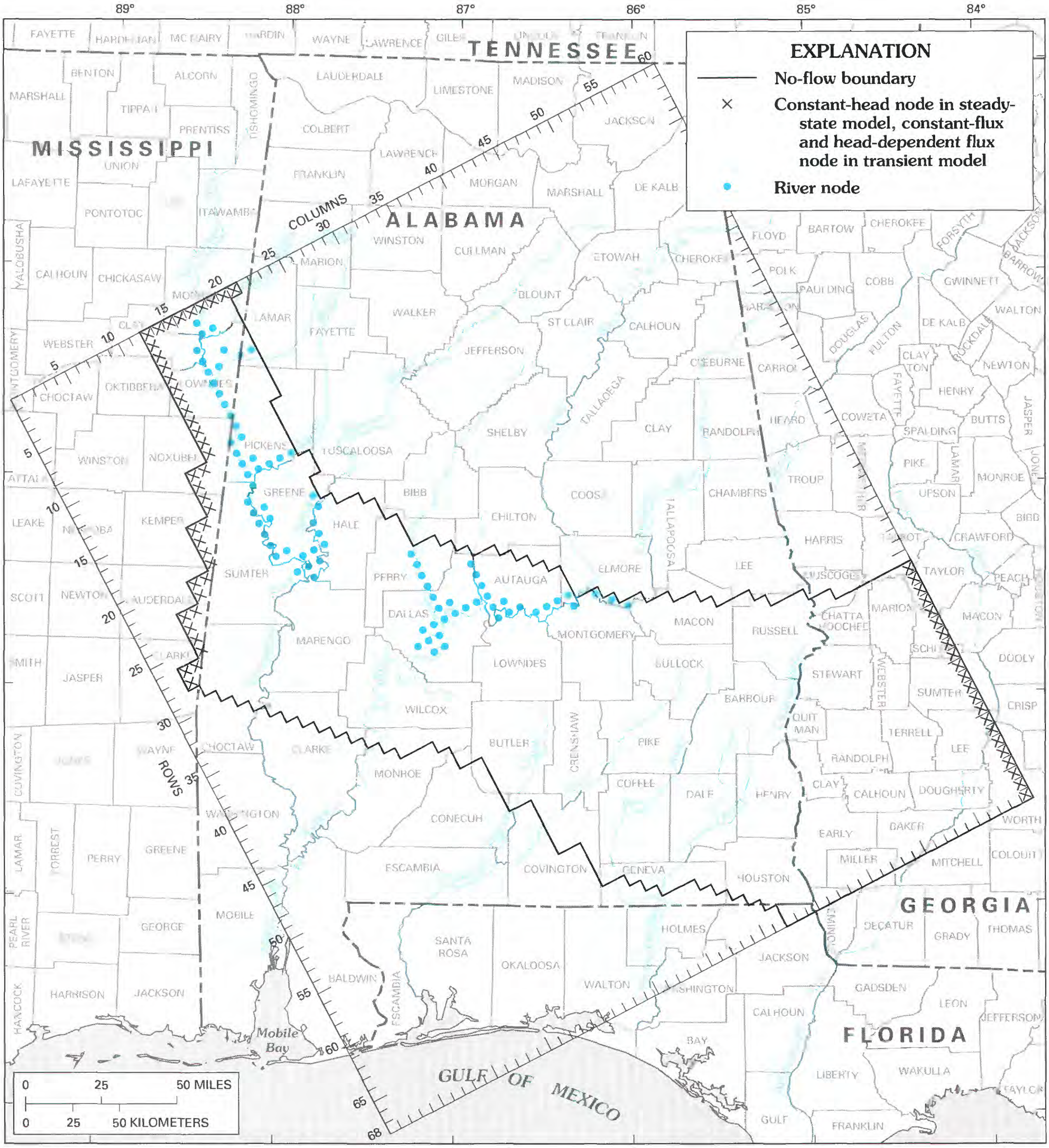


FIGURE 18.—Boundary conditions for model layer A4 (Providence-Ripley aquifer).

head boundary. Potentiometric heads for the Upper Floridan aquifer were taken from a simulation of predevelopment heads in the Floridan (Maslia and Hayes, 1988), but they were modified for central and western Alabama, where actual historical head data differed from heads in that simulation. Heads in the coastal lowlands aquifer system were derived from published data.

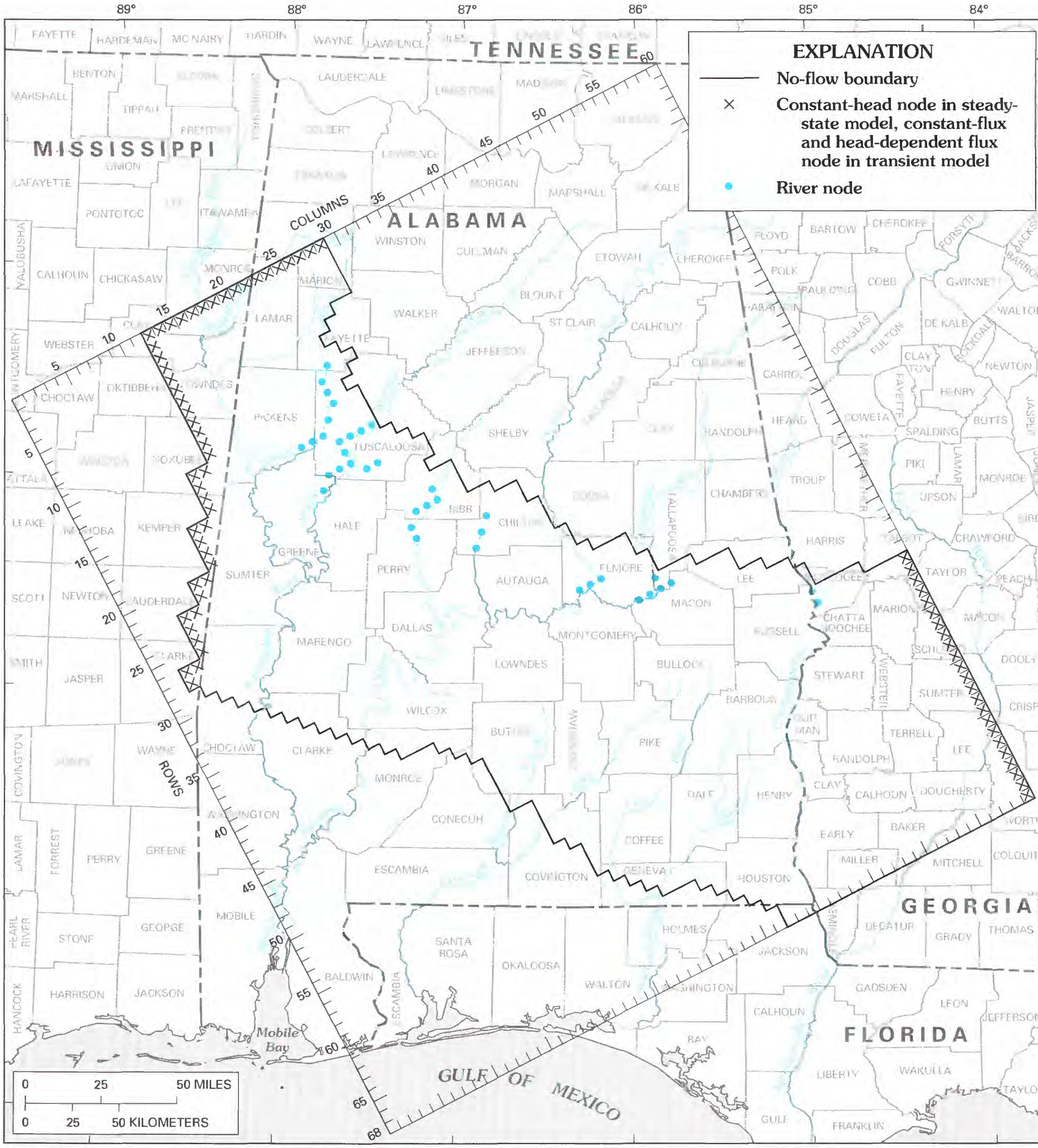
The aquifer properties varied during model calibration were transmissivity, leakance, recharge, and riverbed conductance. Initial estimates of transmissivity were



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 19.—Boundary conditions for model layer A5 (Eutaw aquifer).

derived from aquifer-test data, specific-capacity data, or geophysical logs. Insufficient data prevented reliable areal maps of transmissivity to be produced for the entire study area, so initial approximations were made according to the estimated thickness of aquifer material for each layer. Leakance was estimated using aquifer-test data and geophysical logs, and was initially input as a single, areawide value for each model confining layer (1×10^{-8}/d (day) for model layer C4, the Selma confining layer, and 1×10^{-7}/d for the other confining layers). As



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 20.—Boundary conditions for model layer A6 (Tuscaloosa aquifer).

modeling progressed, leakance was varied areally to improve the simulation within limits permitted by geological evidence. Also, in areas of overlap with the adjacent Mississippi and Georgia subregional models, transmissivity and leakance were adjusted as necessary to compromise between those two models and the Alabama subregional model.

Initial recharge estimates were made using preliminary results from the regional model, while initial riverbed conductances (*Criv*) were approximated by the equation

$$Criv = K'LW/M$$

where

K'=vertical hydraulic conductivity of the riverbed,
L=length of the river reach in a model node,
W=average river width in a model node, and
M=riverbed thickness.

Riverbed thickness was assumed to be 1 ft, while vertical hydraulic conductivity was assumed to be approximately one-tenth of the hydraulic conductivity of the aquifer materials drained by the river. The length and width of river reaches within a grid cell were estimated from U.S. Geological Survey 7.5-minute topographic maps.

The methodology employed in modeling the Southeastern Coastal Plain aquifer system assumed that ground-water flow is separable into three systems: local, intermediate, and regional (Toth, 1963). Further, it was reasoned that, owing to the coarse grid mesh being used in the model, local and intermediate flow systems could not be simulated because their activity is within a cell. Therefore, the model simulated only the regional (deep) flow system. This is the portion of the ground-water flow system recharged by water percolating deep into the ground that does not discharge to small streams in outcrop areas of the aquifer. Flow paths in the deep system tend to be long; under natural conditions, hundreds to thousands of years can elapse before water discharges to wetlands, rivers, or the sea.

Because of the coarseness of the grid (16 mi^2 per cell) used in the subregional model, small, shallow streams in recharge areas that receive water primarily from local and intermediate flow systems were not included in the model. Only the major drains (rivers) were simulated, because they are the ones by which most of the regional flow is discharged from the ground-water system. The choice of streams to include in the model was somewhat arbitrary, but was initially based on whether a stream had a pronounced effect on the potentiometric surface (as defined by water levels in deeper wells) and on whether a stream maintained flow during periods of drought.

The rate of regional discharge to a stream is difficult to estimate because during average climatic conditions the contribution to total streamflow represented by regional discharge is relatively small. Calculations using base-flow-separation techniques may include some portion of flow from the intermediate system, resulting in overestimation of regional discharge. However, ground-water discharge to streams during periods of extreme drought is likely to represent predominantly regional discharge. For that reason, the estimation of regional flow to streams was based on minimum, 1-day streamflow data acquired during the most historically severe drought in Alabama, that of 1954.

Flow from tributary streams not considered regional drains, and therefore not simulated, was subtracted out, and the remainder was considered to represent regional base flow. Minimum flow generally approximated the 100-percent flow-duration statistic, and that statistic was used as an estimate of regional flow for gaged streams for which 1954 data were not available. For some streams, flow-duration data were nonexistent or were based on only a very short period of record. Regional base flow for these streams was determined by estimating the discharge due to regional base flow in adjacent streams on a per-square-mile basis. This value was then multiplied by the drainage area of the stream basin in question to approximate the discharge due to regional flow.

STEADY-STATE SIMULATION

CALIBRATION PROCEDURE

The steady-state simulation was an approximation of predevelopment, long-term average conditions in the Coastal Plain of Alabama and parts of the adjacent States of Mississippi and Georgia. The calibration process involved attempting to match potentiometric heads obtained prior to significant ground-water extractions in the vicinity of wells from which the head data were obtained, and to approximate the discharge from the regional ground-water flow system to the regional drains.

Potentiometric heads representative of the regional flow system generally do not fluctuate significantly under natural conditions, either seasonally or from year to year. Therefore, water-level measurements made during most years in wells that tap the deeper aquifer are not significantly affected by pumping or severe drought and are adequately representative of long-term, average conditions for a model of this scale.

The criterion for a head match was set at 25 ft. That is, if the simulated aquifer head in a particular node was within ±25 ft of an observed head, the simulated head was considered to be within acceptable limits. No absolute criterion was set for matching base flow for given reaches of the regional drains. Rather, stream discharge was used solely as a guide to calibration. This was done because (1) some of the regional drains have been regulated during the entire period of record for streamflow and (2) the method of determining the component of base flow contributed by regional ground-water flow was imprecise, both because selection of regional drains is arbitrary and because the portion of base flow due to regional flow versus the portion due to local or interme-

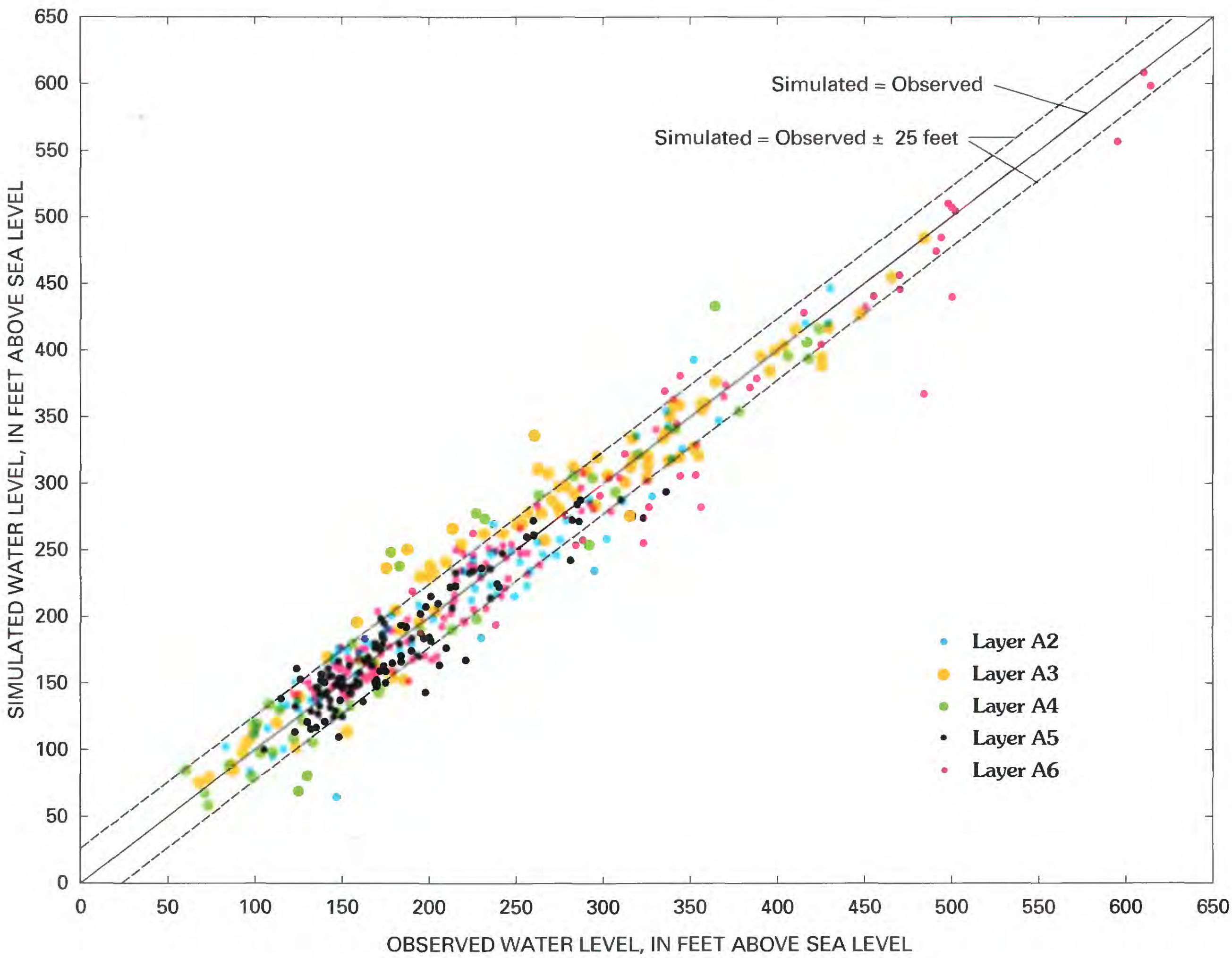


FIGURE 21.—Observed and simulated predevelopment potentiometric heads.

diate flow could not be determined precisely. In the latter case, it was not certain that the minimum base flows recorded during the 1954 drought represented only regional ground-water flow. Therefore, the steady-state model solution was based heavily on the ability to match the predevelopment potentiometric heads. This lack of precise values for ground-water discharge to streams suggests that the control factors of calibration of the steady-state simulation are not as precise as they could be.

No attempt was made to rigorously compare a water budget calculated from field data with the model-simulated budget. This was because (1) neither lateral nor source-sink boundary fluxes were well known, (2) the regional component of ground-water recharge was unknown, and (3) base flow could be only roughly estimated. Lateral boundary fluxes between subregional models were examined by using the regional model, and these fluxes were used as a guide to approximate reasonable flux values at the boundaries. Recharge was varied as necessary (within a maximum rate of 2.8 in/yr (inches per year) per node) to attain acceptable head matches and to approximate calculated streamflows. With base-flow, boundary-flux, and recharge values being uncertain, it was felt that the calculation of a reliable water budget from field data for the Alabama subregion was impractical. (See Wait and others, 1986, for a regional water budget.)

CALIBRATION RESULTS

The overall match between simulated and observed predevelopment heads for all layers was 365 of 451 heads (81 percent) within the 25-ft criterion (fig. 21). In many cases there was no discernable pattern to the heads that could not be matched—they were scattered among other data points where heads were matched (pls. 2–6). The most reasonable explanations for this phenomenon are (1) the observed head data were incorrect, either because of an incorrect land-surface datum measurement or because the water-level measurement was incorrect owing to obstructions in the well, (2) the heads did not represent regional heads (that is, the wells may have been screened in an interval at which heads were representative of the local flow system and not of the regional flow system), or (3) the head measurements were from wells affected by nearby pumping.

In cases in which there was a discernible pattern to the distribution of matched heads, the causes may have been (1) the scale limitations imposed by the coarse grid prevented simulation of hydrologic conditions in the actual system or (2) head measurements were from wells in areas where conditions were not representative of predevelopment conditions.

Scale limitations may have hindered calibration in updip, recharge areas and in the vicinity of stream valleys. In both cases, hydraulic gradients changed at a greater rate than could be adequately simulated, given the 4-mi node spacing. This was particularly apparent in model layer A3 (Nanafalia-Clayton aquifer) (pl. 3) in the Choctawhatchee River valley in southeastern Alabama and in the model layer A6 (Tuscaloosa aquifer) (pl. 6) outcrop area north of the Alabama River in Montgomery, Autauga, and Elmore Counties in central Alabama. When such conditions existed, it was necessary to forego matching a few heads to obtain the greatest overall match.

Historical heads used in model calibration were chosen, when possible, to be representative of predevelopment conditions. However, because predevelopment head data near some pumping centers were unavailable, it was necessary to use water-level measurements that may have been affected by pumping. If water levels near these pumping centers had declined significantly, the simulated potentiometric surface would have been higher than the observed potentiometric surface near these areas. Pumping predates historical water-level measurements in the Nanafalia-Clayton aquifer near Thomasville in Clarke County, in the Providence-Ripley aquifer near Fort Rucker and Ozark in Dale County, Enterprise in Coffee County, and Georgiana in Butler County, and in the Eutaw aquifer near Linden in Marengo County, possibly explaining why simulated heads were higher than observed heads in those localities.

A comparison of estimated and model-simulated base flows in unregulated rivers is given in table 1. Success at matching the streamflows was mixed. As noted earlier, the potential error in estimating regional discharge to rivers may be large, and this may account for disparities between estimated and simulated base flows.

TABLE 1.—*Relation between observed and simulated steady-state base flow in unregulated streams*

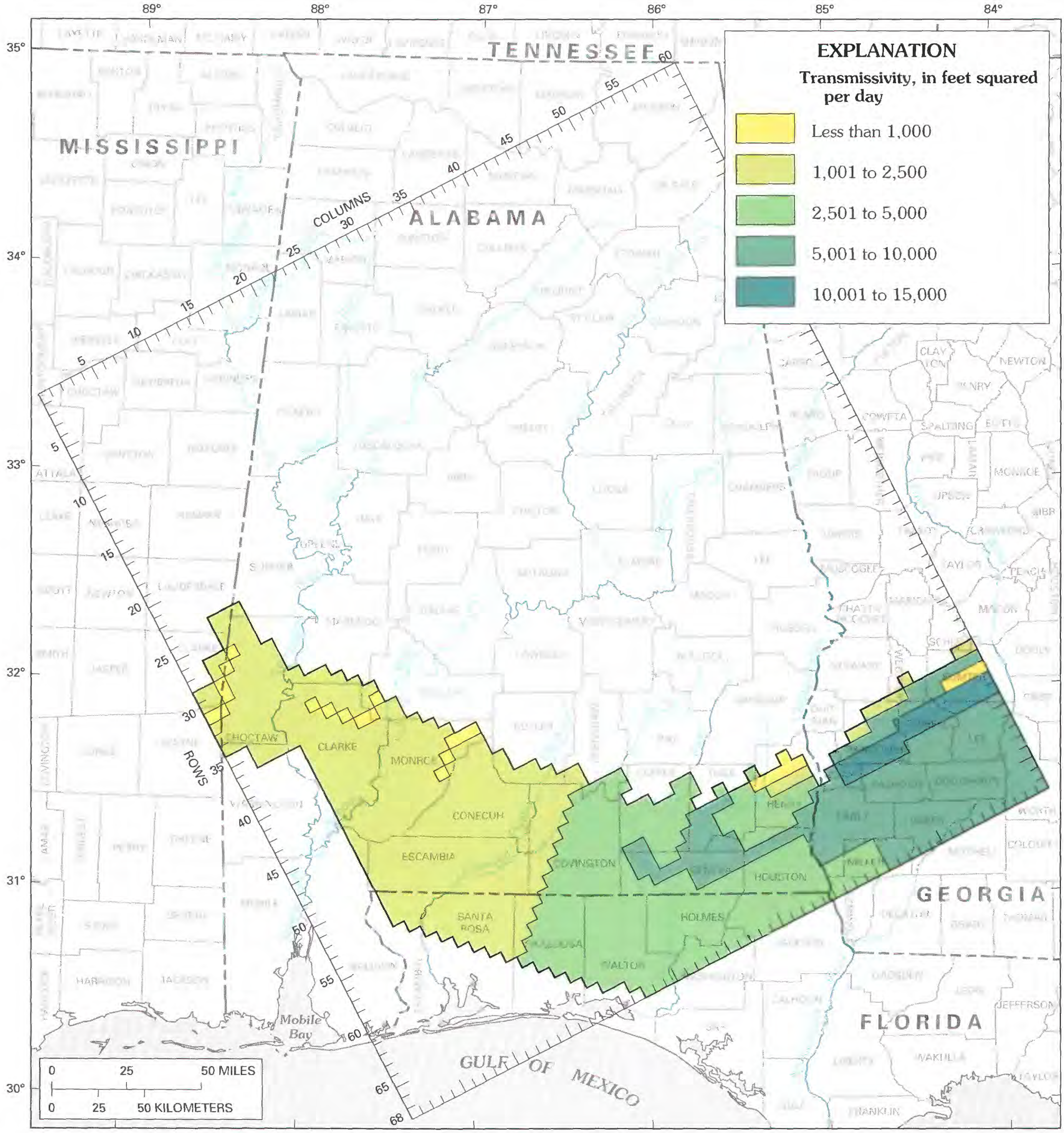
Stream	Observed flow (cubic feet per second)	Simulated flow (cubic feet per second)
Sipsey River	8.5	12
Mulberry Creek	27	11
Cahaba River	84	36
Conecuh River	53	66
Choctawhatchee River	110	102
Pea River	25	52

SIMULATED REGIONAL FLOW REGIME

The model-simulated steady-state potentiometric surfaces are shown on plates 2 through 6. The ground-water flow patterns indicated by the simulated potentiometric surfaces agree well with conceptual flow models of updip areas developed prior to the model analysis. However, the results of simulation suggest a different flow pattern for downdip, confined areas that, in large part, had not been previously considered. Specifically, a large influx of ground water appears to enter the study area from both east and west. Gardner's (1981) model of equivalent layers A5 and A6 (Eutaw and Tuscaloosa aquifers) in western and central Alabama showed similar patterns in those areas, as suggested by this model (pls. 5, 6). However, until the larger scale model constructed for this study was initiated, it had not been recognized that flow perpendicular to the dip of the aquifers occurs in all aquifers and across the entire study area. Locally, flow may be updip, particularly in the valleys of major streams at the point where an aquifer first becomes confined. Note, for example, the closed contours along the Chattahoochee River in all layers (pls. 2–6). These results, of course, assume a correct choice of boundary conditions, the most important of which is the downdip no-flow boundary, whose damming effect forces ground water to discharge to either overlying layers or regional drains. Because the regional confining beds appear to be very effective, little ground water is discharged upward to the overlying aquifers. Therefore, most flow in the downdip areas of these aquifers is horizontal, generally toward major rivers in the unconfined outcrop areas of the aquifer.

SIMULATED TRANSMISSIVITY

The areal distribution of model-derived transmissivity for each model layer is shown in figures 22 through 26. Simulated transmissivities generally did not vary appreciably from initial values; the majority of the changes were refinements to initial values in updip areas and accommodation with values supplied from adjacent models. The values shown reflect changes to the simulated transmissivity made during the transient model calibration, when the addition of pumpage stresses to the model

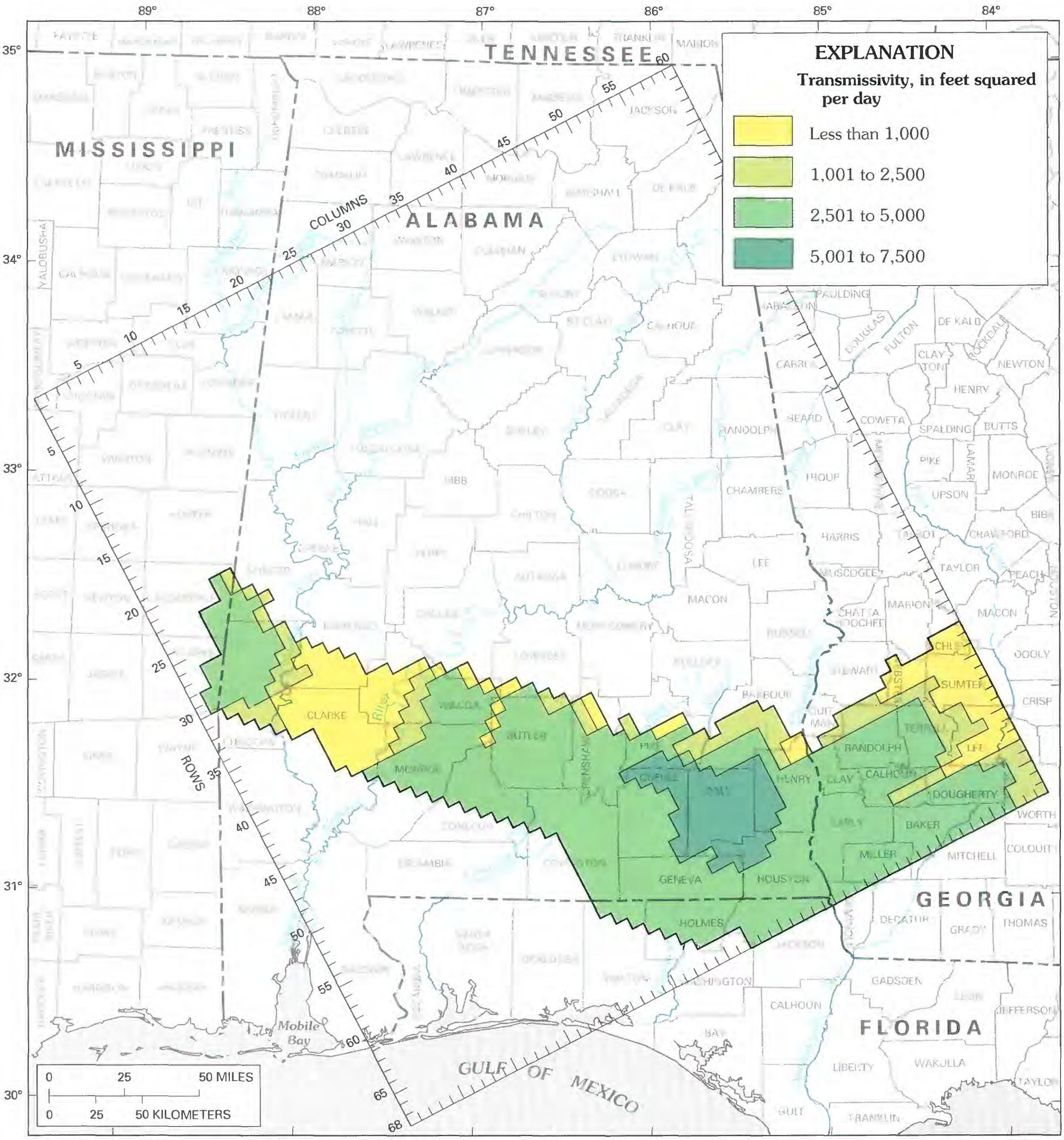


Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 22.—Simulated transmissivity in model layer A2 (Lisbon aquifer).

provided a test of the values derived from the initial steady-state model calibration.

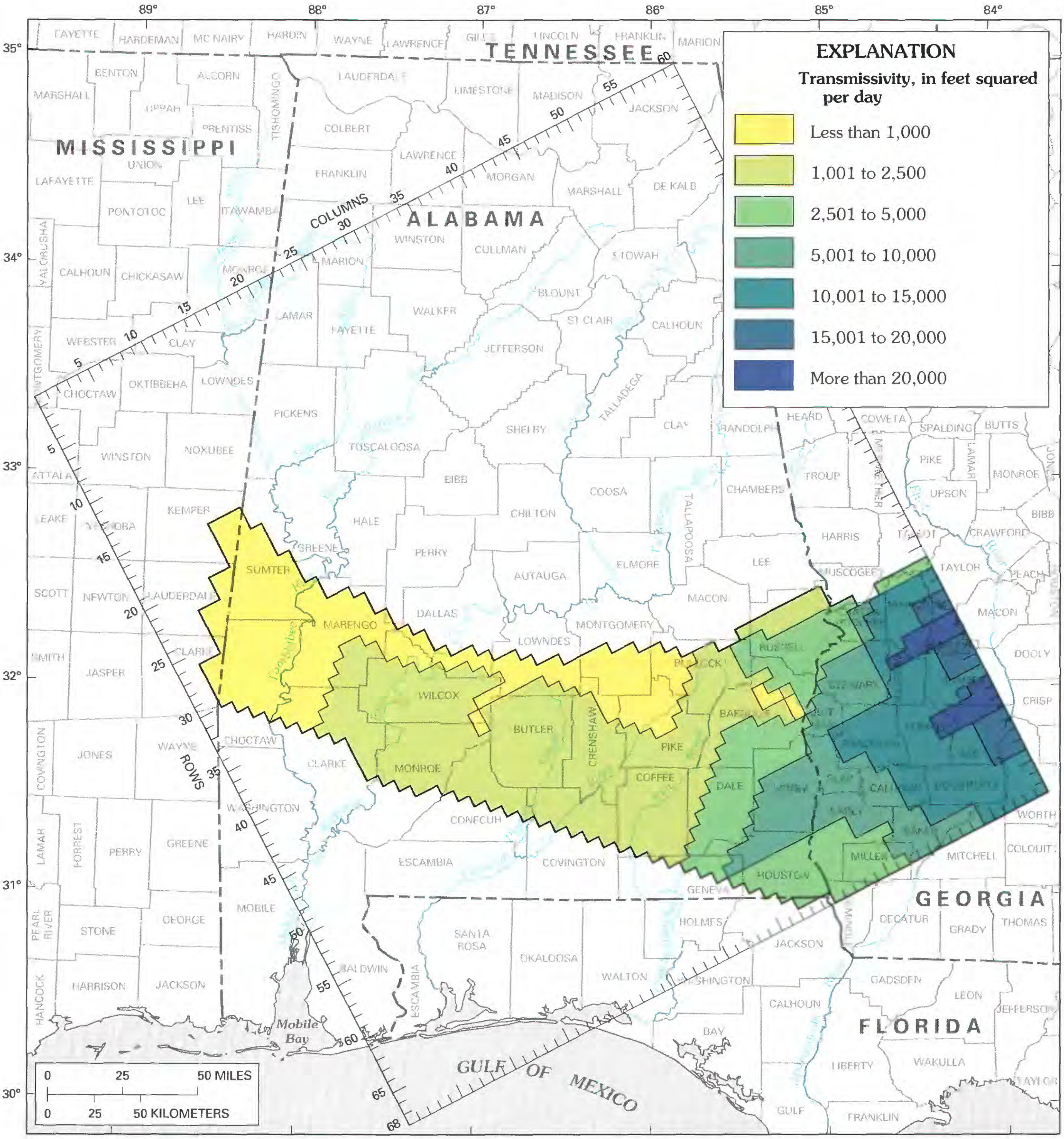
The transmissivity maps show a general trend of west-to-east increase in model layers A2, A3, and A4 (Lisbon, Nanafalia-Clayton, and Providence-Ripley aquifers) and decrease in model layers A5 and A6 (Eutaw and Tuscaloosa aquifers). The transmissivity distributions reflect the marked facies changes that are known to be present. In the case of the Eutaw and Tuscaloosa aquifers, the west-to-east decrease in transmissivity is due to



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 23.—Simulated transmissivity in model layer A3 (Nanafalia-Clayton aquifer).

a decrease in sand thickness along with a decrease in hydraulic conductivity. Conversely, in the Nanafalia-Clayton and Providence-Ripley aquifers, the aquifer materials increase in both thickness and hydraulic conductivity from west to east. The west-to-east increase in transmissivity in the Lisbon aquifer probably is due to a gradual increase in total sand thickness within the aquifer, although the various formations within the aquifer may vary individually in sand thickness from place to place. Simulated transmissivities for the entire Alabama



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 24.—Simulated transmissivity in model layer A4 (Providence-Ripley aquifer).

subregional model ranged from 100 to 24,000 ft^2/d, and within Alabama from 100 ft^2/d in model layer A5 (Eutaw aquifer) to 11,500 ft^2/d in model layer A6 (Tuscaloosa aquifer).

SIMULATED LEAKANCE

Leakance patterns generally conformed to initial concepts of vertical conductance within the aquifer system. However, leakance values occasionally were varied

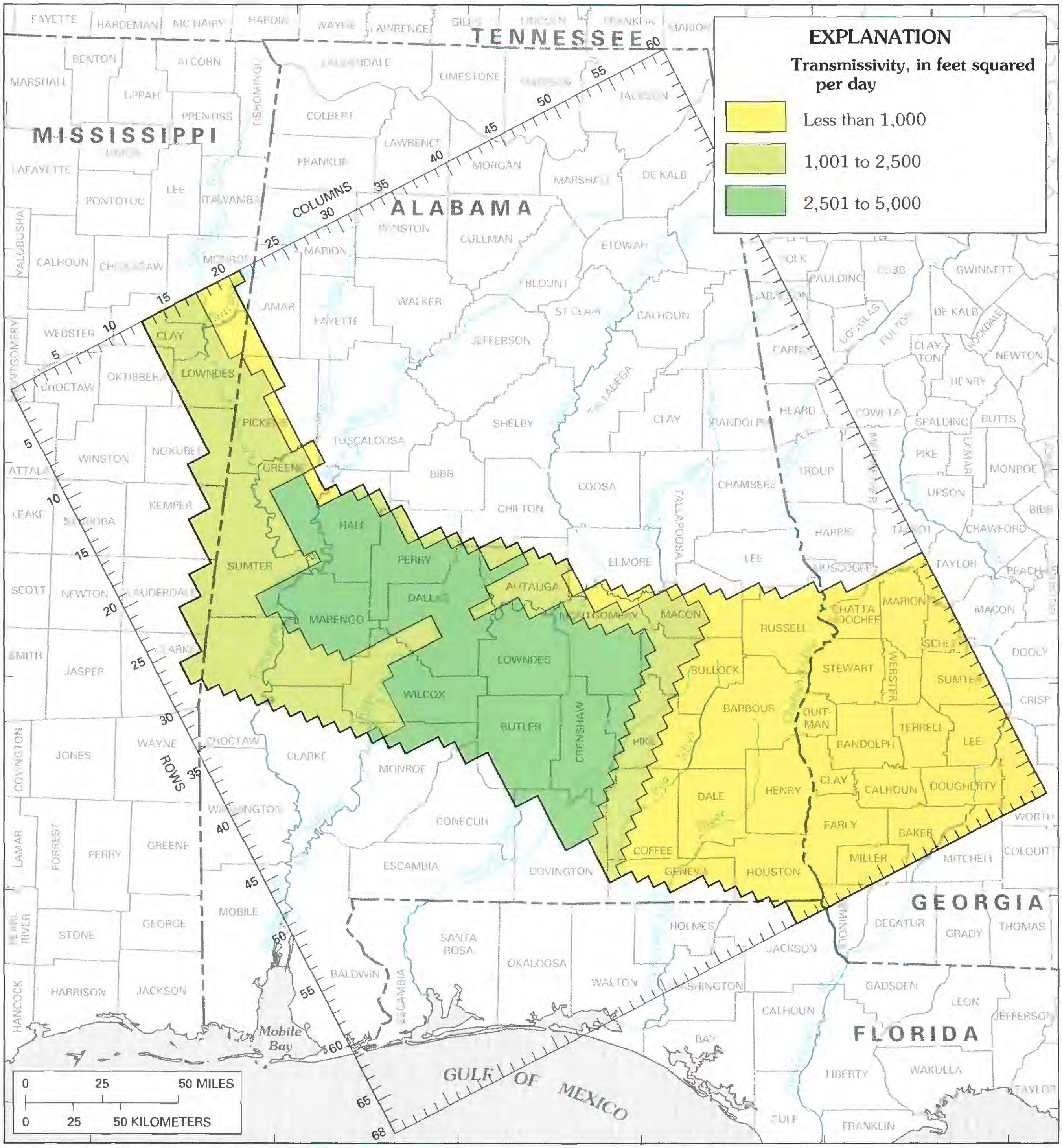


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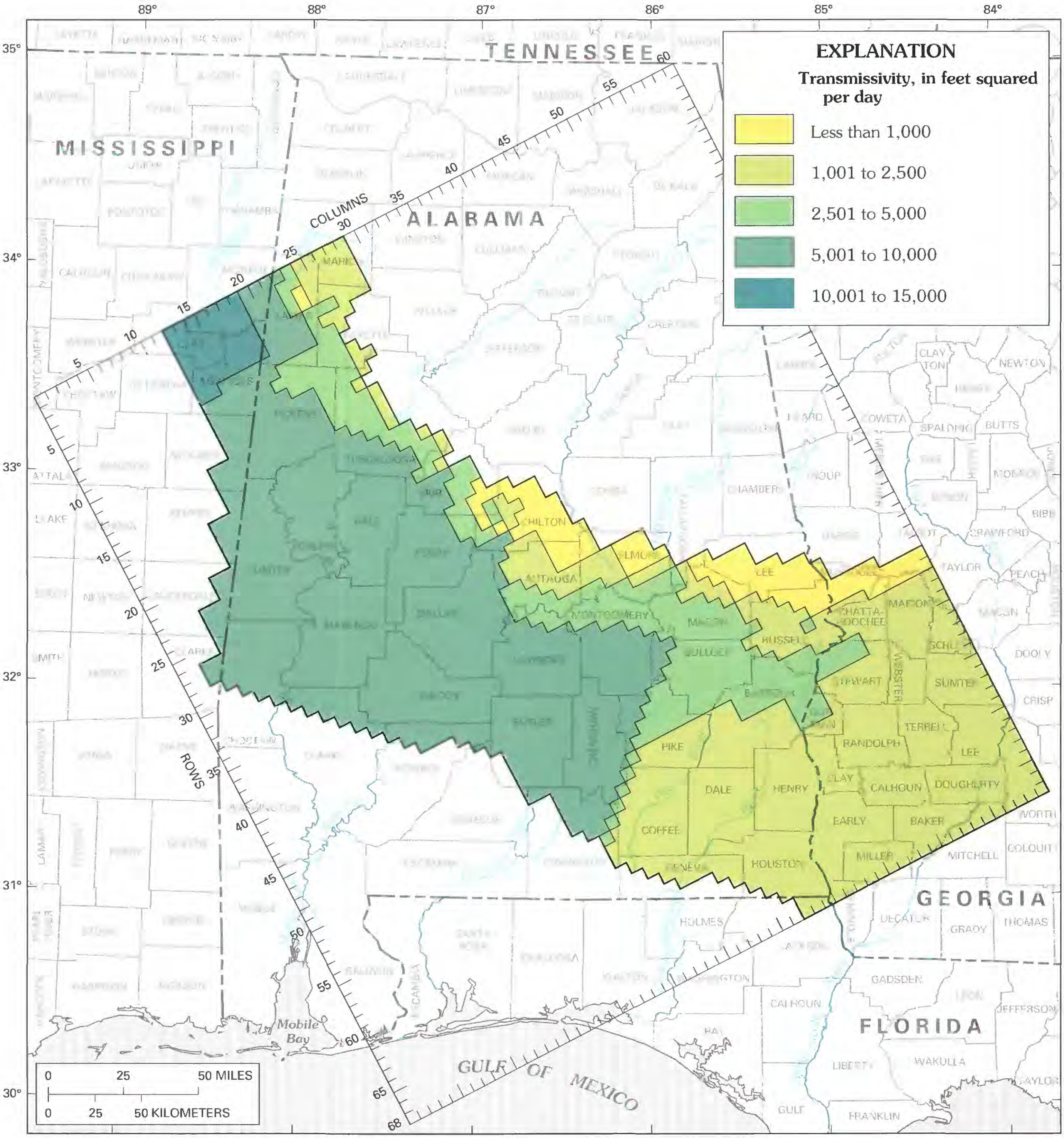
FIGURE 25.—Simulated transmissivity in model layer A5 (Eutaw aquifer).

markedly during model calibration, with values in some areas being changed considerably over relatively short distances. This may indicate geohydrologic conditions in the aquifer system that differ areally. Also, leakance values were increased at updip river nodes to lower head in an underlying, confined aquifer when it was impossible to reproduce the usually steepening potentiometric gradients observed near rivers by other means without adversely affecting overall model calibration. This indicates that confining beds under rivers in updip areas may



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 26.—Simulated transmissivity in model layer A6 (Tuscaloosa aquifer).

allow more vertical leakage than in other areas, possibly because the rivers are entrenched in fault zones, or because the confining beds were eroded during periods when river base levels were lower than at present.

Overall leakance values for nodes not affected by these adjustments for updip river nodes ranged from about 5×10^{-5} to 1×10^{-8}/d for the entire Alabama subregional model. However, in Alabama the range was 7×10^{-5} to

1×10^{-8}/d. The lowest values occurred in model layer C4 (Selma confining unit), which represents in large part the Demopolis and Mooreville Chalks. The highest values occurred in model layer C1 (Yazoo confining unit) in southeastern Alabama and southwestern Georgia. In that area, layer C1 represents the confining unit between the Lisbon aquifer and the overlying Upper Floridan aquifer, which was modeled as a source-sink layer. The relatively high leakance values in model layer C1 suggest that the two aquifers are hydraulically well connected, and that the Lisbon aquifer may be under nearly water-table conditions in much of southeastern Alabama.

Another area of apparently unusually high hydraulic connection is between the Eutaw and Tuscaloosa aquifers along the Alabama and Tombigbee Rivers. This connection is suggested by the potentiometric surfaces in the area, which show a potentiometric low in both aquifers, even where one or both are confined (pls. 5, 6). To simulate the potentiometric low, it was necessary to allow upward leakage from model layer A6 (Tuscaloosa aquifer) to model layer A5 (Eutaw aquifer), and from model layer A5 (Eutaw aquifer) to the rivers. Where the Eutaw aquifer is confined by an extensive outcropping area of the Selma confining unit, river nodes were simulated in contact with the aquifer to provide the leakance necessary to lower the potentiometric head.

SIMULATED RIVERBED CONDUCTANCE

Simulated riverbed conductances ranged from 175 to 75,000 ft^2/d. The lowest values occurred in nodes where the rivers were relatively narrow or where, in model layer A5 (Eutaw aquifer), the rivers run on the chalk of confining unit C4 (Selma confining unit). The highest values occurred where large rivers drain primarily sand aquifers. Conductance values were varied as necessary to control river-aquifer flux and potentiometric head. The calibrated values represent a decrease from initial values: conductance values derived from the equation $Criv=K'LW/M$ generally proved too high when riverbed conductivity and thickness values within an expected range ($K'=1$ to 1×10^{-2} ft/d, $M=1$ ft) were used, allowing excessive flux from the aquifer to the river.

The possible sources of error inherent in the calculation of riverbed conductance are several. Among them are (1) an inappropriate concept of what determines the value of riverbed thickness in the equation, (2) variability of riverbed materials, and thus of vertical hydraulic conductivity, along a reach, (3) error in estimating stream length or width within a node, and (4) error in assigning stream stage in a node.

Simulated values for riverbed conductance were about two orders of magnitude lower than initial estimates. The most likely cause of this discrepancy was that the initial value of streambed thickness used in the equation was too small. Aquifer heads represent the average over the entire aquifer thickness, and therefore the vertical resistance to flow from the midpoint of the aquifer to the streambed must be considered. With aquifer thicknesses generally ranging between 200 and 500 ft, the simulated values of riverbed conductance obtained from model calibration seem more reasonable than the initial estimates. However, given that values are usually varied as necessary to meet streamflow or potentiometric head criteria, riverbed conductance needs to be considered as a model-generated property dependent on the model in which it is used. The simulated values from this model may not be transferable to other models having different initial assumptions and conditions. The range of values reported here should be considered in that light.

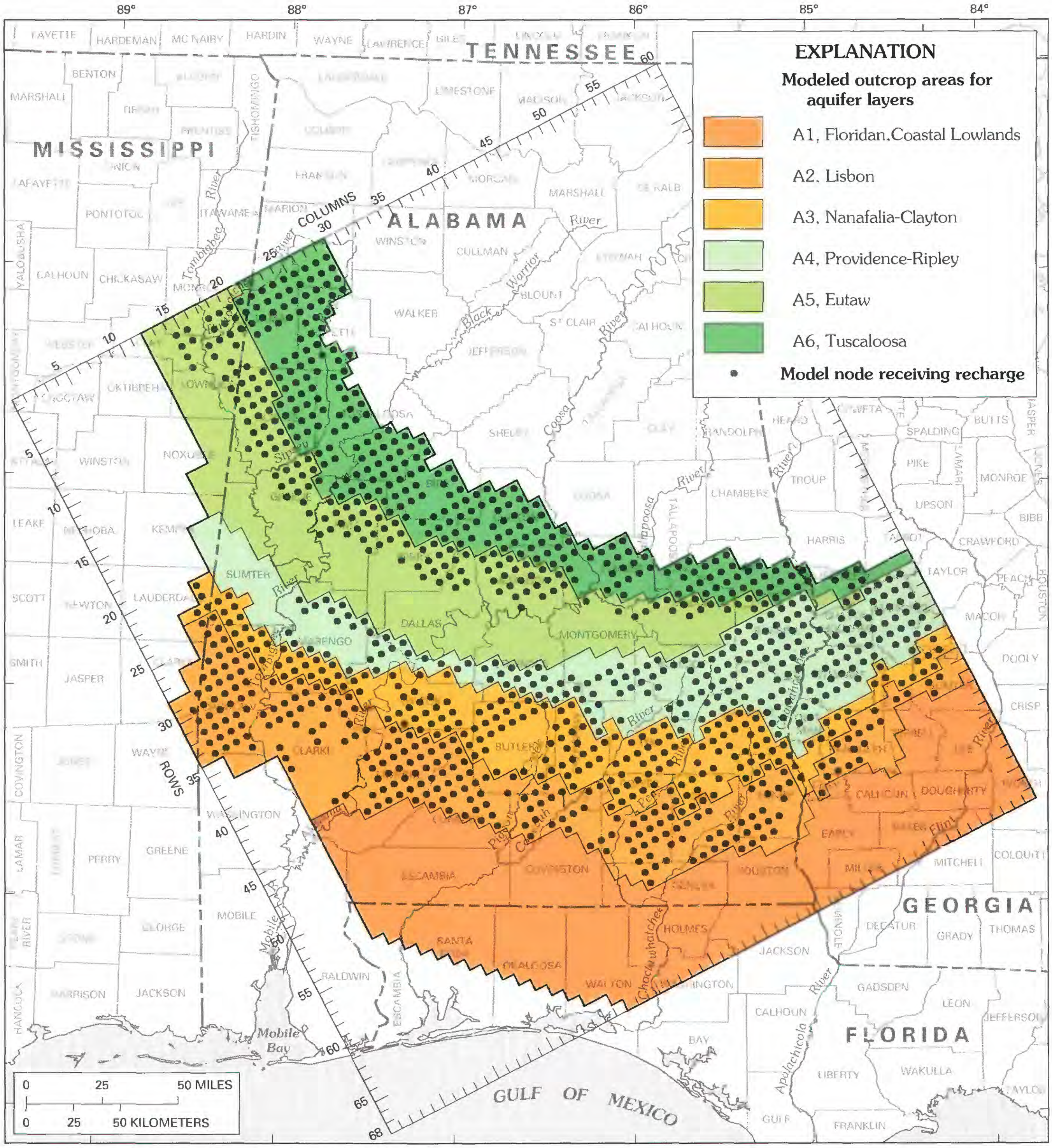
SIMULATED RECHARGE

Recharge was varied during simulation to approximate the observed potentiometric gradients and streamflow in regional drains. The initial recharge pattern of 1 in/yr, applied uniformly over all aquifer outcrops, was adjusted considerably as modeling progressed. The final average recharge to the model was 0.62 in/yr, but average amounts differed among model layers (table 2), from 0.45 in/yr in model layer A4 (Providence-Ripley aquifer) to 0.91 in/yr in model layer A2 (Lisbon aquifer).

The distribution of recharge in the calibrated model is shown in figure 27. Not all the outcrops of the modeled aquifers received recharge; part of the modeled outcrops coincided with confining units and nodes that contained streams that did not receive recharge. In general, more recharge was applied to nodes coinciding with interstream areas than to nodes coinciding with stream valleys and wetlands. This agrees conceptually with what is expected in regional ground-water systems where there are gaining streams. The downward hydraulic gradient is much greater in interstream areas, allowing greater recharge rates (Heath, 1983, p. 20–21). Recharge to lowlands in stream valleys is more likely to be captured by the local flow system and discharged rapidly to

TABLE 2.—*Recharge rates applied during simulation and recharge areas of model layers*

Model layer	Average recharge rate (inches per year)	Maximum recharge rate (inches per year)	Recharge area (square miles)
A2	0.91	2.8	4,368
A3	.67	1.7	3,104
A4	.45	.5	3,568
A5	.47	.8	2,480
A6	.52	1.3	4,224
Entire model	0.62		17,744



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 27.—Outcrop areas of the aquifers as modeled and the distribution of nodes receiving recharge.

streams and rivers than is recharge that enters the aquifer in areas farther removed from such drains. Also, many areas of wet lowlands in the study area contain significant stands of hydrophilic phreatophytes which remove large amounts of water from the aquifer by transpiration, capturing some of the recharge that might otherwise have been incorporated in the deeper ground-water flow system.

SIMULATED PREDEVELOPMENT WATER BUDGET

The model-simulated predevelopment water budget (fig. 28) shows that recharge from precipitation is the largest source of input to the model in all model layers except A5 (Eutaw aquifer). The most significant source of flux from constant heads is the source-sink layer (layer A1, the coastal lowlands aquifer system and Upper Floridan aquifer). Rivers contribute an insignificant amount to the input budget, but they are the primary means by which water leaves the model. Most of the simulated flow out of the modeled study area is to areas assigned constant heads and is upward to the source-sink layer.

Confining layer C4 (Selma confining unit), with a total vertical flux of 18 Mgal/d, retards leakage between aquifer layers to a larger degree than any other confining layer, resulting in two relatively distinct flow systems within the model. The upper system is composed of model layers A2 through A4 (Lisbon, Nanafalia-Clayton, and Providence-Ripley aquifers), while layers A5 and A6 (Eutaw and Tuscaloosa aquifers) make up the lower system. This corroborates evidence offered by the potentiometric surfaces, which indicate different flow patterns for the same two aquifer groupings. Confining layer C2 (Tuscahoma confining unit) provides the next greatest degree of hydraulic separation between aquifers, with a total vertical flux of 24 Mgal/d. These results agree well with the framework for the four-State regional model of the Southeastern Coastal Plain, which combines model layers A3 and A4 (Nanafalia-Clayton and Providence-Ripley aquifers), and model layers A5 and A6 (Eutaw and Tuscaloosa aquifers) of the Alabama subregional model (and comparable layers of the other subregional models), to create, with the addition of model layer A2 (Lisbon aquifer), three active layers. The finer grid of the subregional models enables simulation of smaller scale variation in aquifer characteristics horizontally and vertically, making subdivision of the regional framework practical.

SENSITIVITY ANALYSIS

The model was tested for sensitivity to changes in the various model inputs that were adjusted during calibration. Each model input (transmissivity, leakance, riverbed conductance, and recharge) was varied individually within all model layers by 0.1, 0.5, 2.0, and 10.0 times the calibrated value while all other inputs were held constant. The results of the sensitivity analysis are given in figures 29 through 32, which show the effects of variation of the model inputs in three ways: first, by changes in the mean residuals (the mean of the differences between observed and simulated heads) for each model layer; second, by changes in the mean absolute residuals (the mean of the absolute values of the residuals); and third, by changes in the percentages of calibration heads matched by model layer. The mean of the individual residuals is an estimate of the overall fit of the simulated potentiometric surface to that defined by the heads measured in observation wells. The statistical significance of the mean increases with increasing numbers of observation wells. The mean absolute residual is a measure of the average magnitude of errors at nodes for which there were observed heads.

The graphs of model input change versus change in the measures of sensitivity indicate a greater overall sensitivity to changes in recharge than to other aquifer properties (model inputs) and the least sensitivity to leakance. The graphs also show that model calibration was least affected by large (order of magnitude) changes in leakance compared with other aquifer properties.

Although a sensitivity analysis such as that performed in this study provides a general idea of model sensitivity to changes in model inputs (aquifer properties), it cannot demonstrate the effects of interaction between these properties, nor can it show the relative differences in degree of head change among different areas of the model. For example, a simultaneous increase in recharge and decrease in transmissivity will result in much greater head change than variation in either property taken alone. Likewise, for this model at least, variation in either of these inputs was likely to result in the greatest head changes in areas where the aquifer was simulated as being relatively thin. However, a rigorous analysis that would demonstrate model sensitivity to areal variation of aquifer properties or their interaction would be impractical for a model of this size and complexity.

TRANSIENT SIMULATION

CALIBRATION PROCEDURE

The calibration of the transient simulation was an iterative procedure. It involved adjusting the model inputs that represent aquifer characteristics to approximate the changes in the potentiometric surface of aquifers in the Alabama Coastal Plain that have occurred over time in response to pumping stresses. The solution from the steady-state simulation was used for the initial conditions, with the addition of pumpage and storage properties. The values representing the aquifer characteristics derived from the calibrated transient model were then incorporated in the steady-state model, which was recalibrated using the revised values.

For the transient simulation, the boundary conditions at lateral boundaries that had represented ground-water

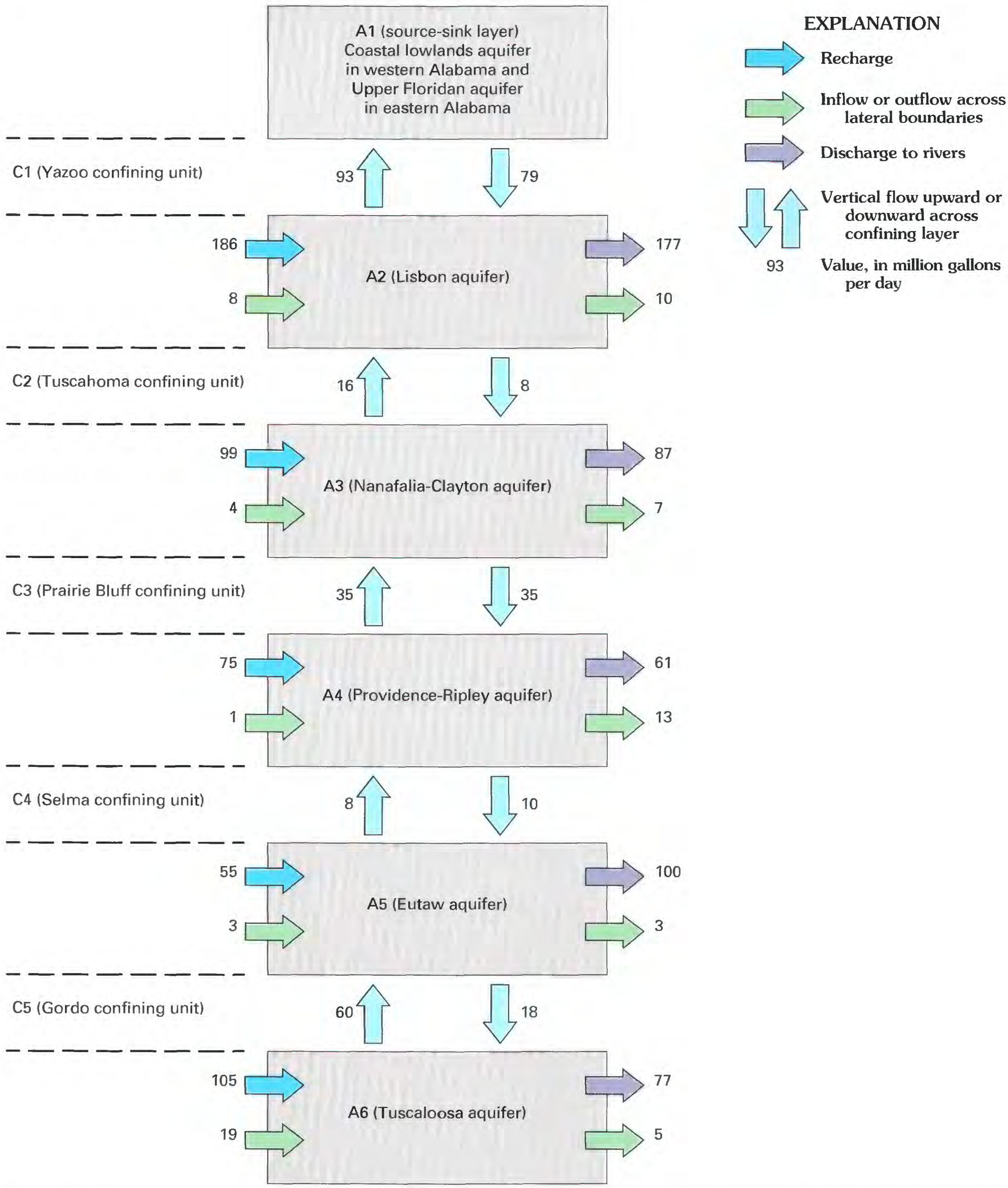


FIGURE 28.—Simulated predevelopment water budget for the Alabama subregion of the Southeastern Coastal Plain aquifer system. Units are rounded to the nearest million gallons per day. (Inflow and outflow may not balance exactly because of error of rounding.)

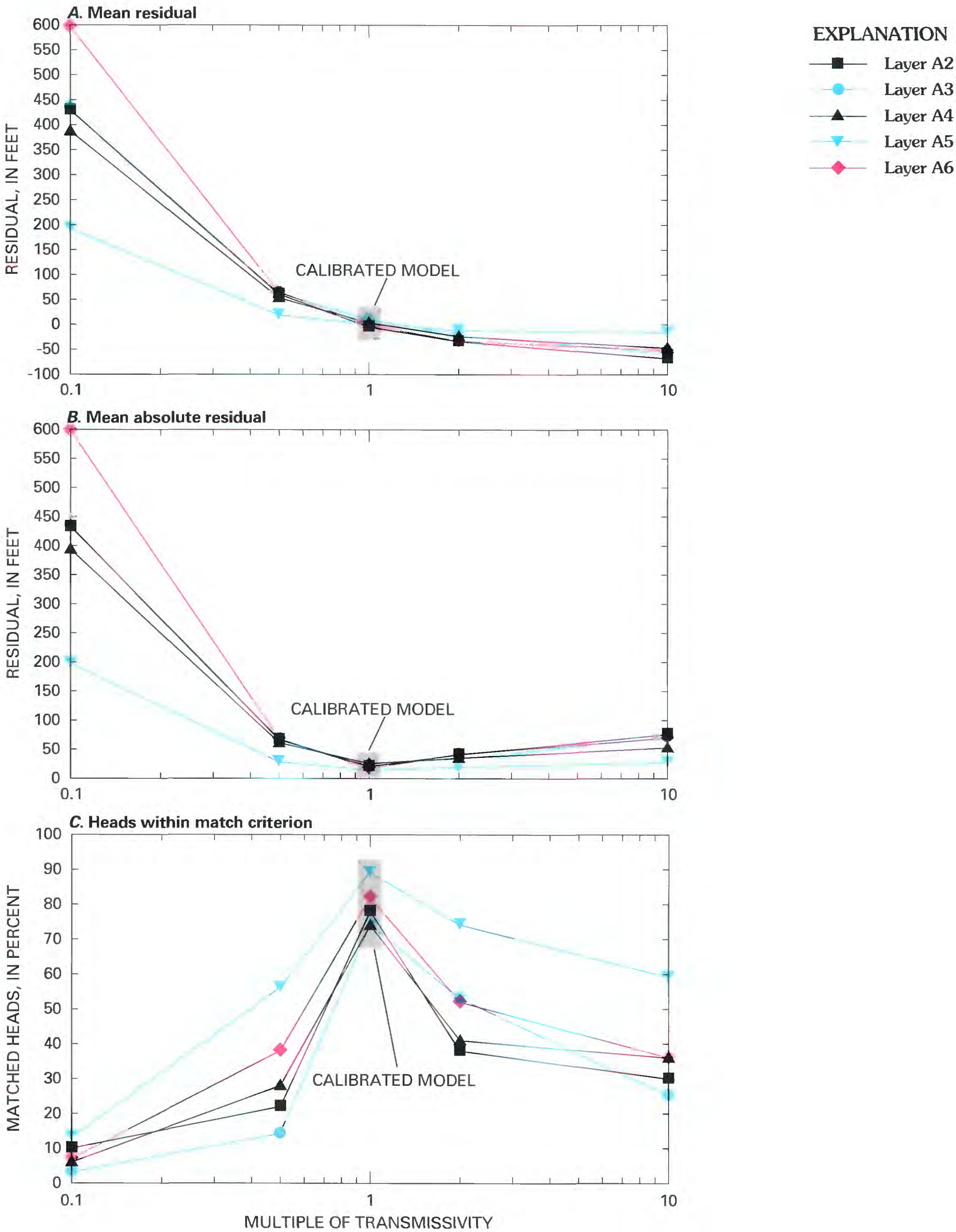


FIGURE 29.—Sensitivity of model layers to changes in transmissivity.

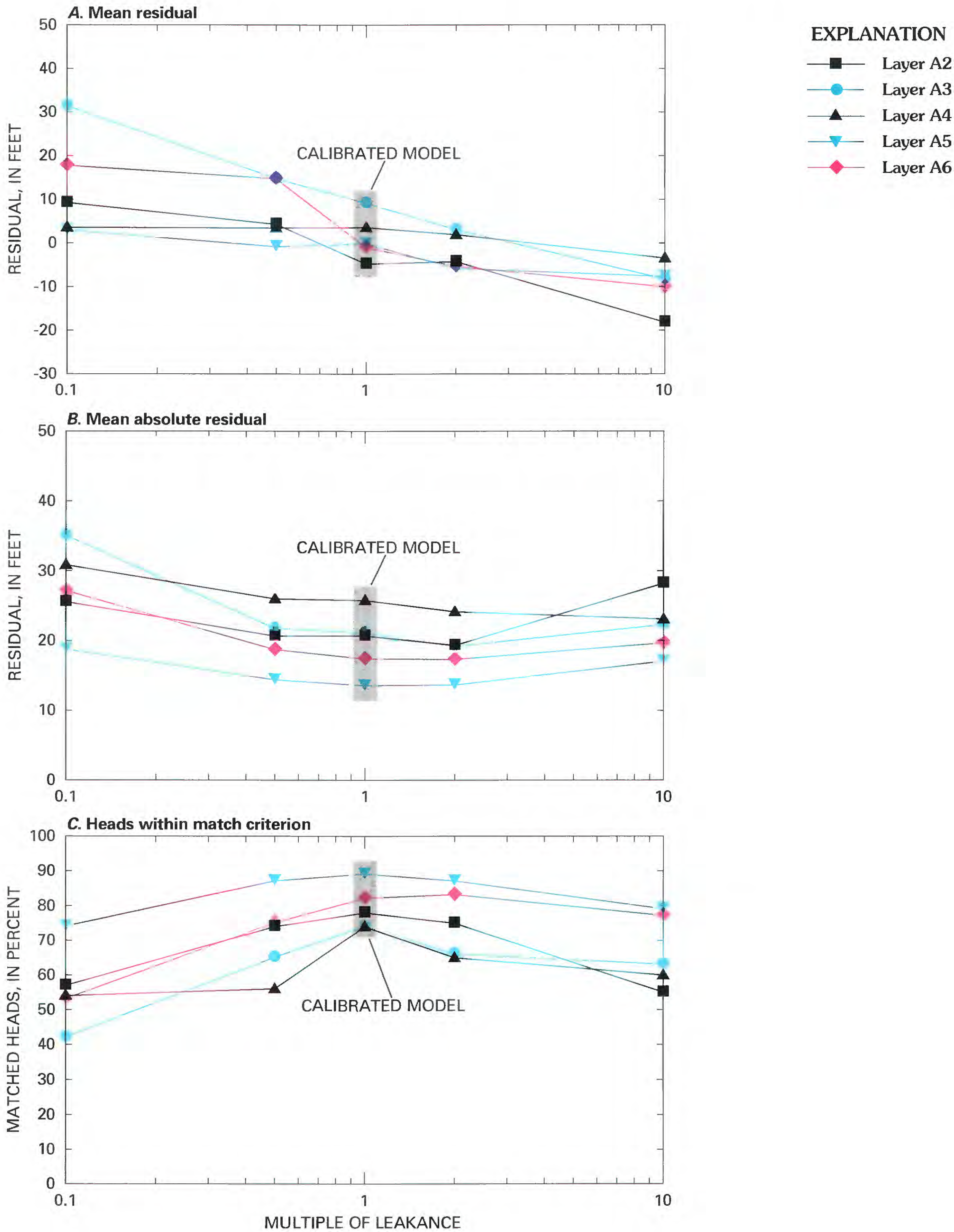


FIGURE 30.—Sensitivity of model layers to changes in leakance.

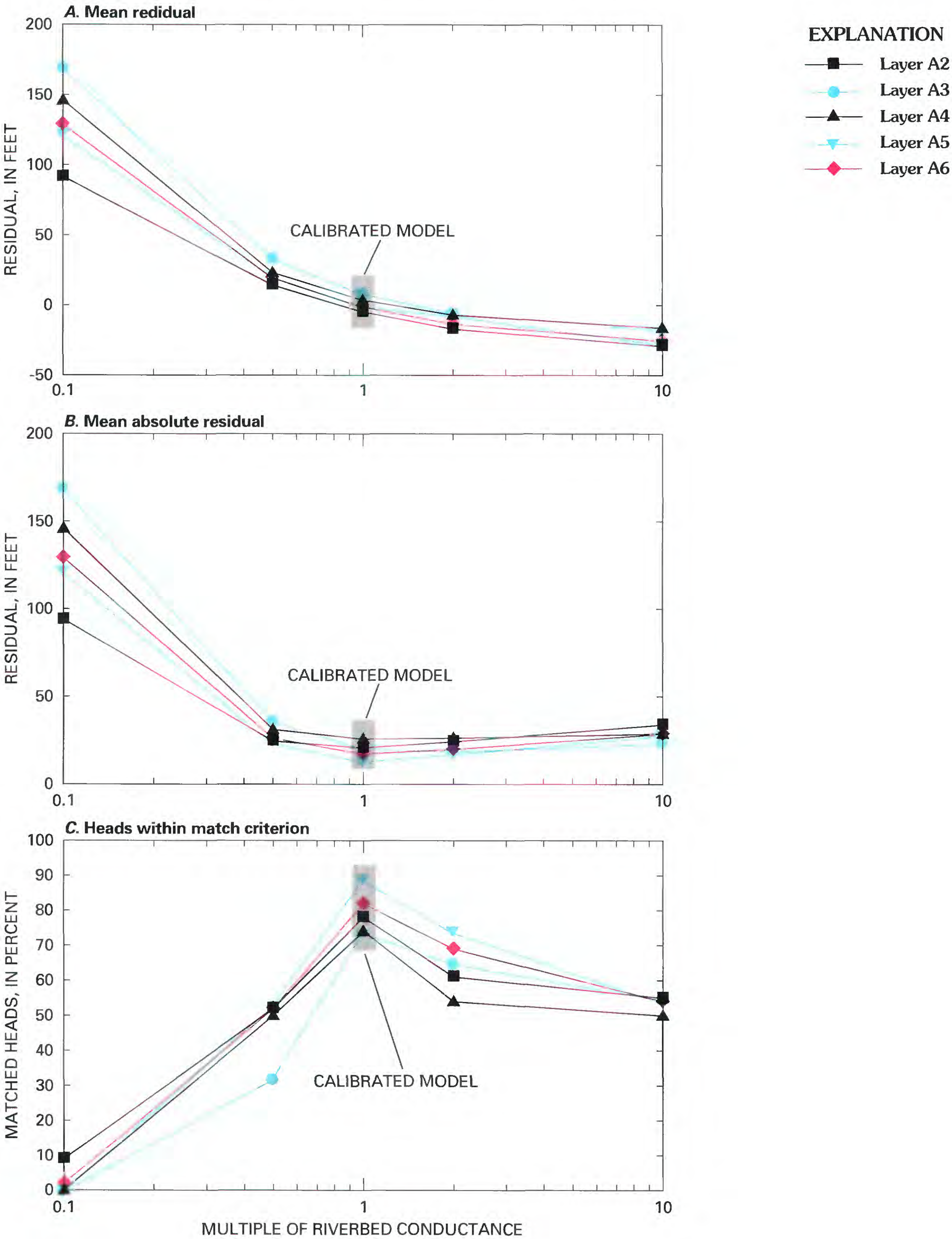


FIGURE 31.—Sensitivity of model layers to changes in riverbed conductance.

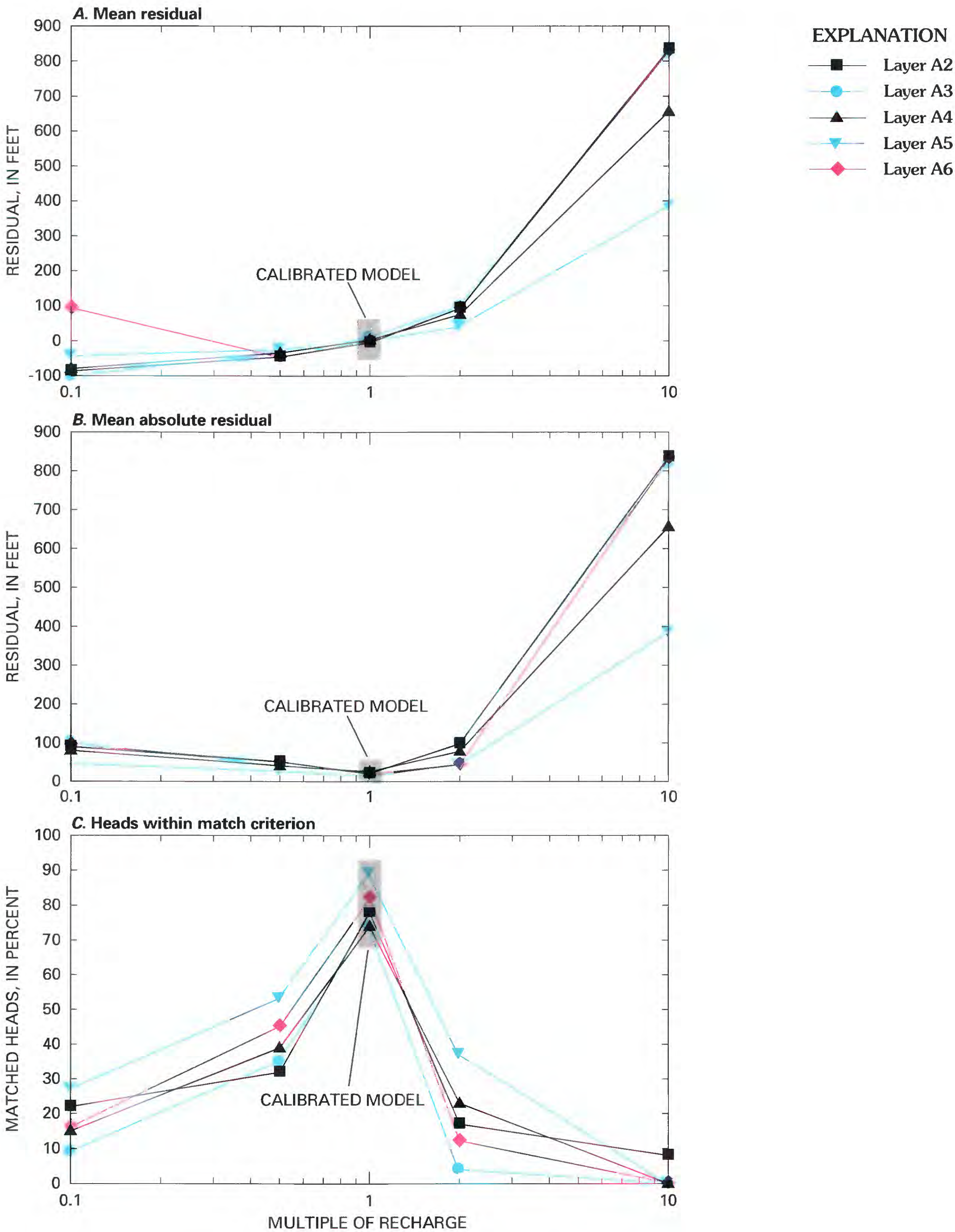


FIGURE 32.—Sensitivity of model layers to changes in recharge.

TABLE 3.—*Simulated pumpage applied during each stress period*

Stress period	Years	Withdrawal rate, by model layer (million gallons per day)					
		Layer 2	Layer 3	Layer 4	Layer 5	Layer 6	Total
1....................	1955–64	7.96	15.78	11.04	35.49	32.14	102.41
2....................	1965–70	9.12	20.17	13.05	33.69	26.73	102.76
3....................	1971–75	10.32	24.10	15.13	35.98	33.10	118.63
4....................	1976–81	17.29	47.40	21.95	35.19	52.87	174.70

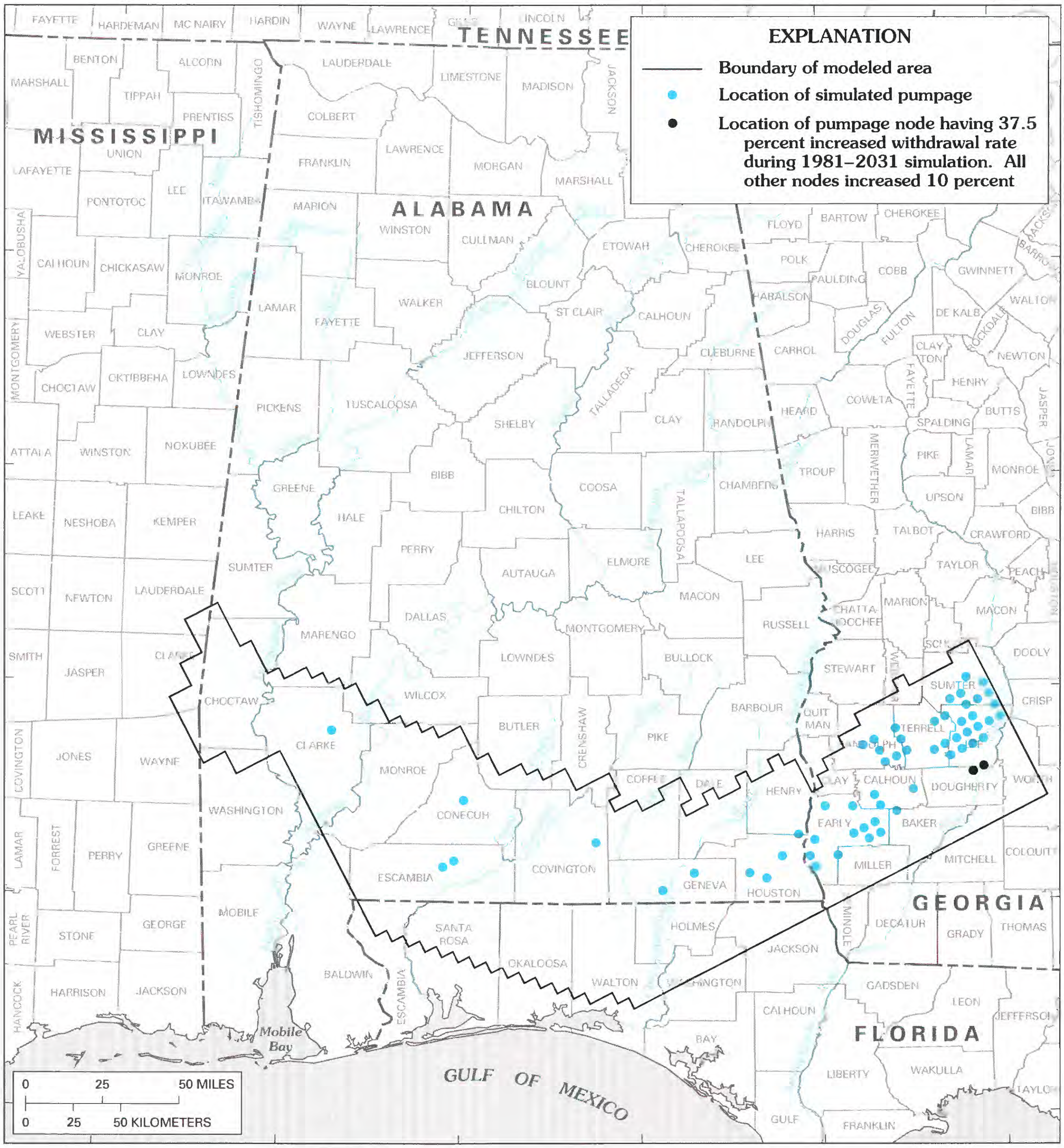
divides in the steady-state model were modified to provide for the fact that pumpage may alter the pattern of ground-water flow, thus changing the positions of ground-water divides. These boundaries were simulated with two conditions simultaneously. The first was constant flux, in which the flux defined by the constant heads during the steady-state simulation was placed along the boundary to simulate the flow present prior to the onset of pumping. The second was head-dependent flux, in which a simulated change in flux across the boundary could occur if pumpage were to lower heads along that boundary. The head-dependent boundary condition requires that a head be chosen that would remain unaffected by pumping at some distance beyond the modeled area. The head chosen for each node was the same as the head in the active node on the boundary. With this condition, no additional flux would occur until pumpage effects reached the boundary. The assigned distance to the dependent head was the distance to the nearest natural discharge boundary, a major river.

Storage values for the transient simulation were initially applied uniformly, with a storage coefficient of 1×10^{-4} for the confined areas of the aquifer and a specific yield of 0.1 for the first row of updip nodes in each aquifer. It was believed that only at the most updip nodes of the aquifers would ground water be under water-table conditions, and only there could a specific yield be justified for storage—wells drilled at a distance of 4 mi downdip (the distance of one grid block) probably would produce water under confined conditions. The values for storage properties were varied during calibration, keeping within limits defined by the few aquifer tests available for the area. The range in values for storage coefficient was 5×10^{-4} to 8×10^{-5}, and the range in values for specific yield was 0.1 to 0.2.

Initial estimates of pumping rates included (1) major municipal and industrial use in all aquifers, (2) discharges from flowing wells in the western parts of the Eutaw and Tuscaloosa aquifers, and (3) discharges from flowing wells in Crenshaw and Pike Counties for the Providence-Ripley aquifer along the Conecuh River valley. Historical pumping for public supply in Alabama was estimated from population data when actual records were not available (Crownover, 1987). After modeling began, it became evident that not all stresses were being simulated. Wells supplying water for catfish farms had been neglected, and the water use for the farms had to be estimated. Estimates of total catfish-farm pumpage came from water-use publications for 1970 (Pierce, 1972) and 1982 (Baker, 1983). Distribution of the catfish-farm withdrawals was determined from topographic maps, with nodal pumping rates estimated by the density of catfish ponds. Figures 33 through 37 show the locations of simulated pumpage, by model layer, and table 3 gives the total pumping rate, for each stress period by layer.

Detailed water-level monitoring in the Coastal Plain aquifers of Alabama did not begin until the compilation of county water-availability reports in the 1950's and 1960's. Therefore, documentation of aquifer response to historical pumping from predevelopment conditions was not available on a statewide basis. The approach to calibration of the transient model was to attempt to approximate the response to stresses from 1965 through 1981, the period during which more extensive water-level and ground-water withdrawal data were collected. A preliminary stress period was simulated, with pumpage modeled at 1965 rates, to adjust the water levels for aquifer responses to pumping prior to 1965. It was not known how long water levels would react to the pumping, but it was assumed that the pumping rate was small enough that a short lead-in period could be used. The initial stress period was varied for lengths of 5, 10, or 15 yr prior to 1965; the best results (matching of drawdown rates in the next stress period) were obtained with a lead-in stress period of 10 yr. Three remaining stress periods were simulated (1965 through 1970, 1971 through 1975, and 1976 through 1981), the length of each period being determined by significant changes in the pumping rates for the larger water users. Withdrawal rates for Mississippi and Georgia were supplied by the staffs of those subregional studies (M.J. Mallory and R.E. Faye, U.S. Geological Survey, written commun., 1986), and changes in rates were adjusted as necessary to conform to the Alabama stress periods.

The initial approach in the calibration of the transient model was to attempt duplication of drawdown rates in observation wells during each stress period. This would have eliminated bias in the drawdown simulated during the first stress period. However, it was soon evident that for certain areas this method would not work. For wells



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 33.—Locations of nodes containing simulated pumpage in model layer A2 (Lisbon aquifer).

having continuous record, the success at duplicating drawdown rates was good. However, for many wells having only intermittent measurements, consistent results were not obtained. Therefore, the approach to calibration was modified to duplicating drawdown rates when possible, but at least approximating total drawdown for areas that had only intermittent measurements.

Early in the calibration process, the transient simulation produced poor results for some areas. One difficulty

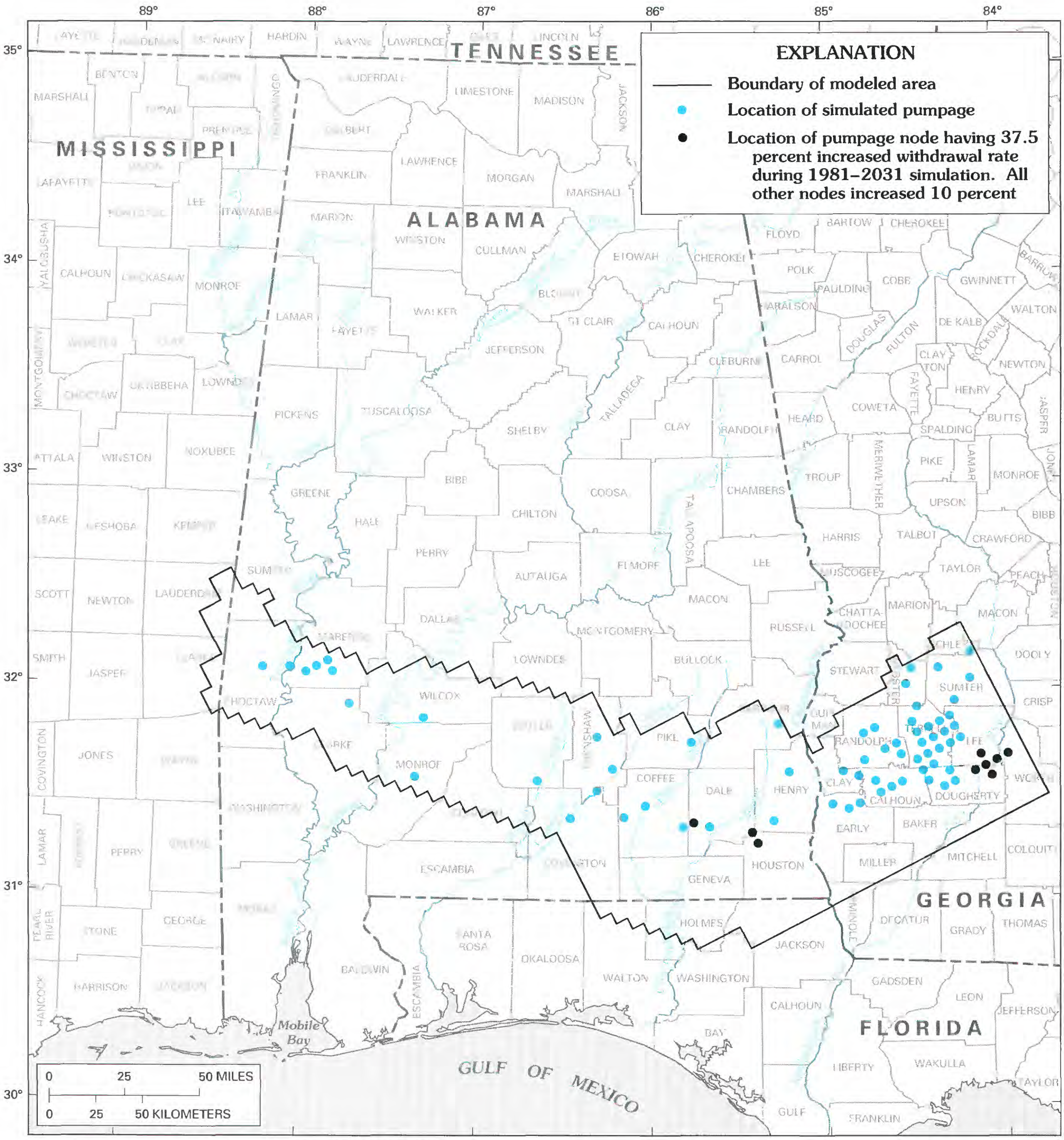


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FIGURE 34.—Locations of nodes containing simulated pumpage in model layer A3 (Nanafalia-Clayton aquifer).

was an inability to approximate water-level changes in several wells in which actual levels fell early in the simulation period, then rose later, despite the fact that pumping rates in these wells did not change. Initially, it was thought that these rises might have resulted from measurement error. However, the phenomenon coincided with the construction of dams on nearby rivers, which undoubtedly would raise the water level in the dammed rivers to which the aquifers discharge. This offers the most plausible explanation for the water-level

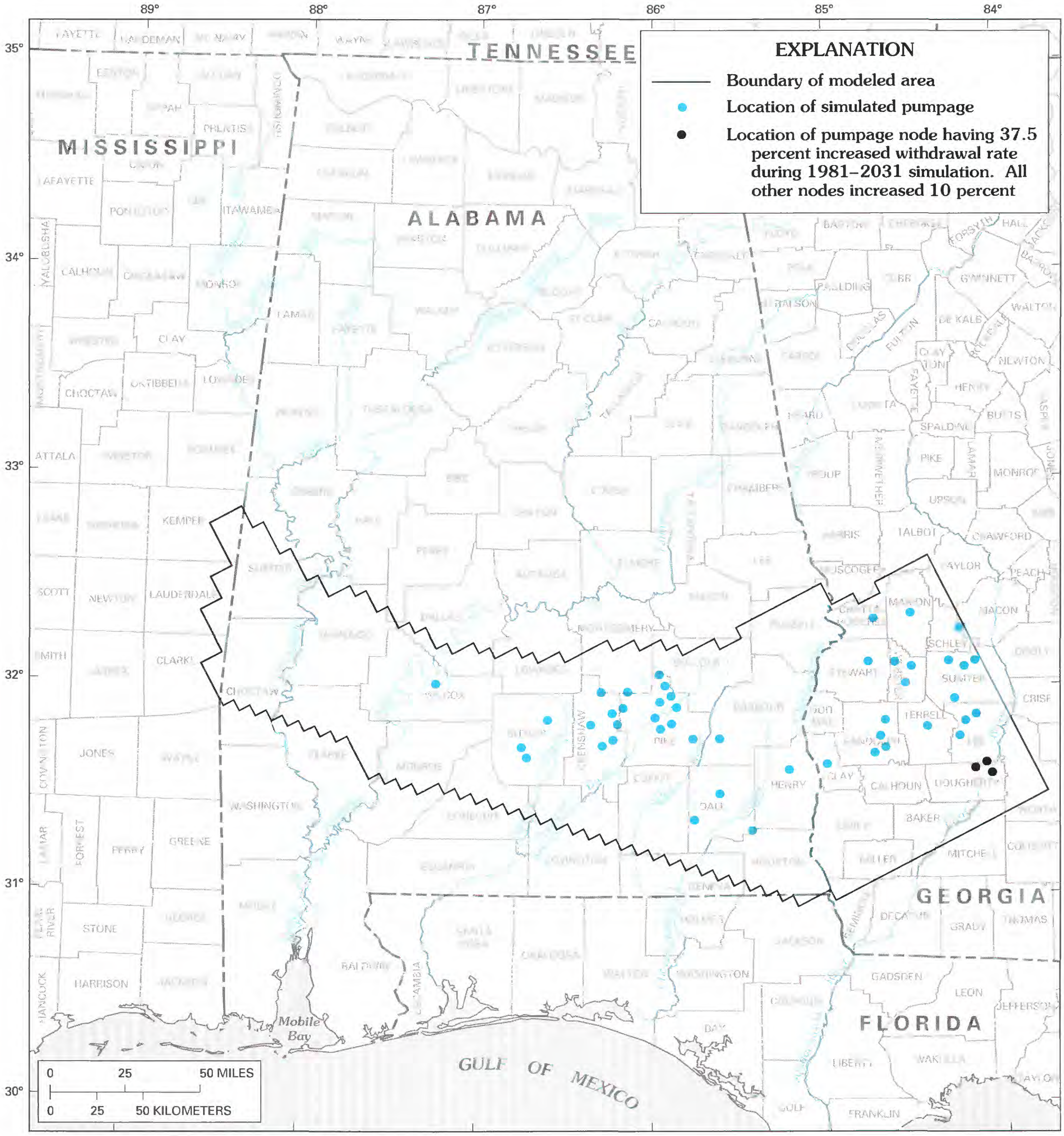

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FIGURE 35.—Locations of nodes containing simulated pumpage in model layer A4 (Providence-Ripley aquifer).

rises. As a result, river stage was adjusted as appropriate to reflect changes in pool elevations on the Tombigbee, Black Warrior, Alabama, and Chattahoochee Rivers during the simulation period.

Another difficulty during the transient calibration was an inability in many areas of model layers A2, A3, and A4 (Lisbon, Nanafalia-Clayton, and Providence-Ripley aquifers) to achieve sufficient simulated drawdowns to

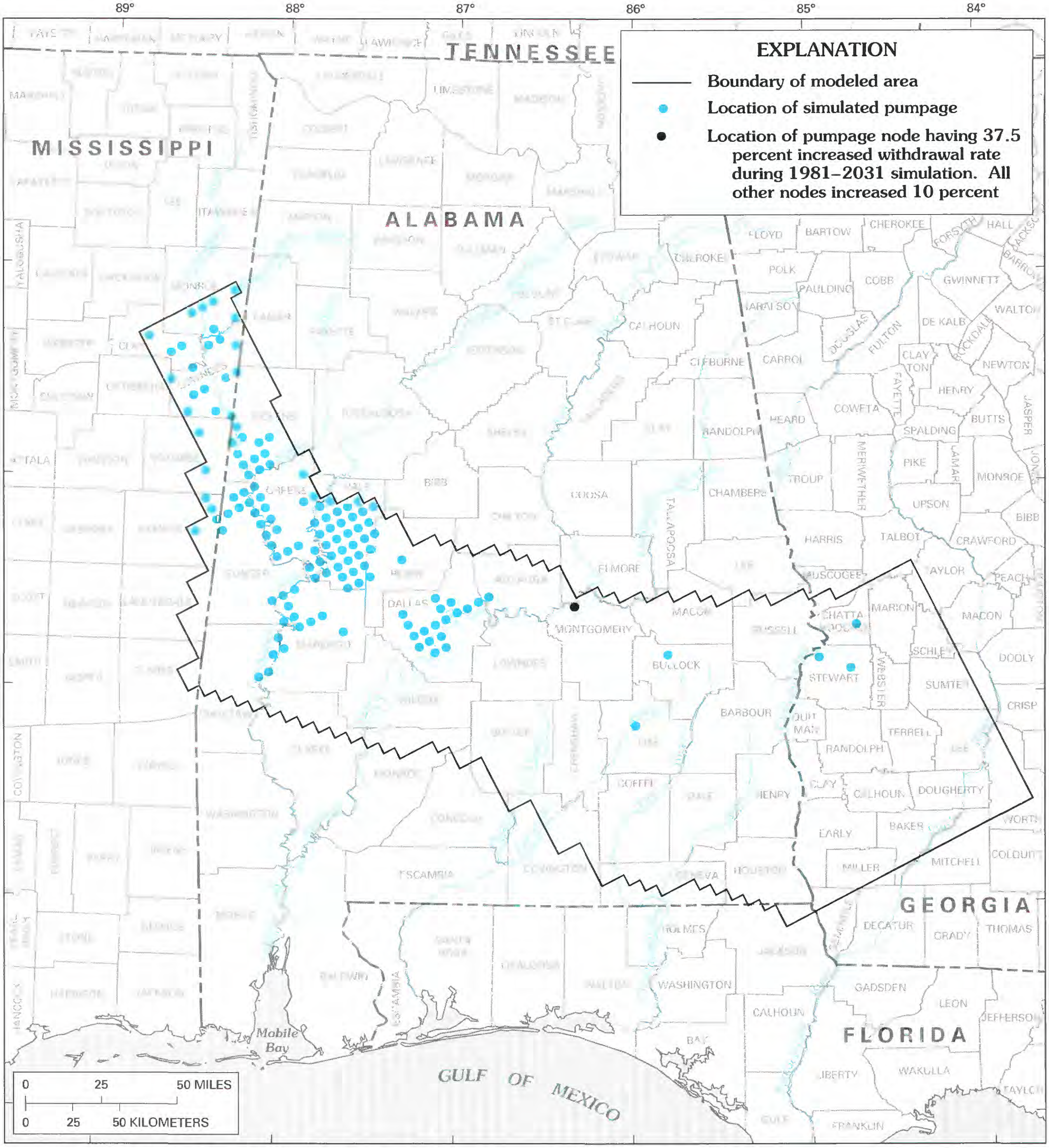

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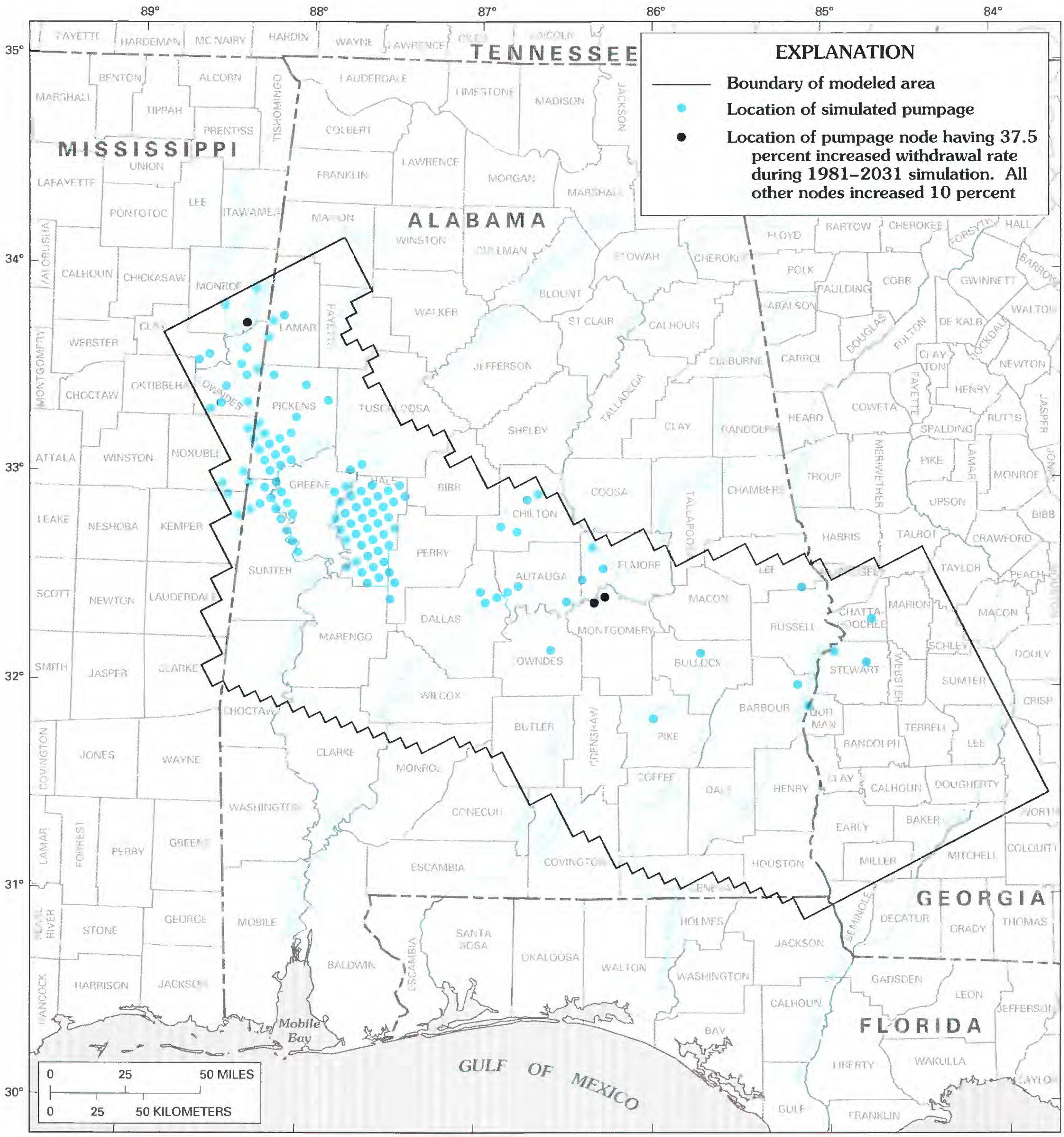
FIGURE 36.—Locations of nodes containing simulated pumpage in model layer A5 (Eutaw aquifer).

approximate those actually occurring in the aquifer system. Simulation under steady-state conditions with pumpage included, so that the model would solve for maximum potential drawdown, showed that sufficient drawdown could not be achieved with the transmissivity values being used. (That is, the transmissivity values



Base modified from U.S. Geological Survey digital data. 1:2,000,000, 1972

FIGURE 37.—Locations of nodes containing simulated pumpage in model layer A6 (Tuscaloosa aquifer).

were too high.) Errors in the initial transmissivity values, which were derived from the predevelopment simulation, in many cases may have resulted from insufficient data on base flow for the regional drains, errors in estimating base flow due to regional discharge, or a paucity of observed head data.

In the three uppermost aquifers, transmissivity values were reduced in the western two-thirds of the model area

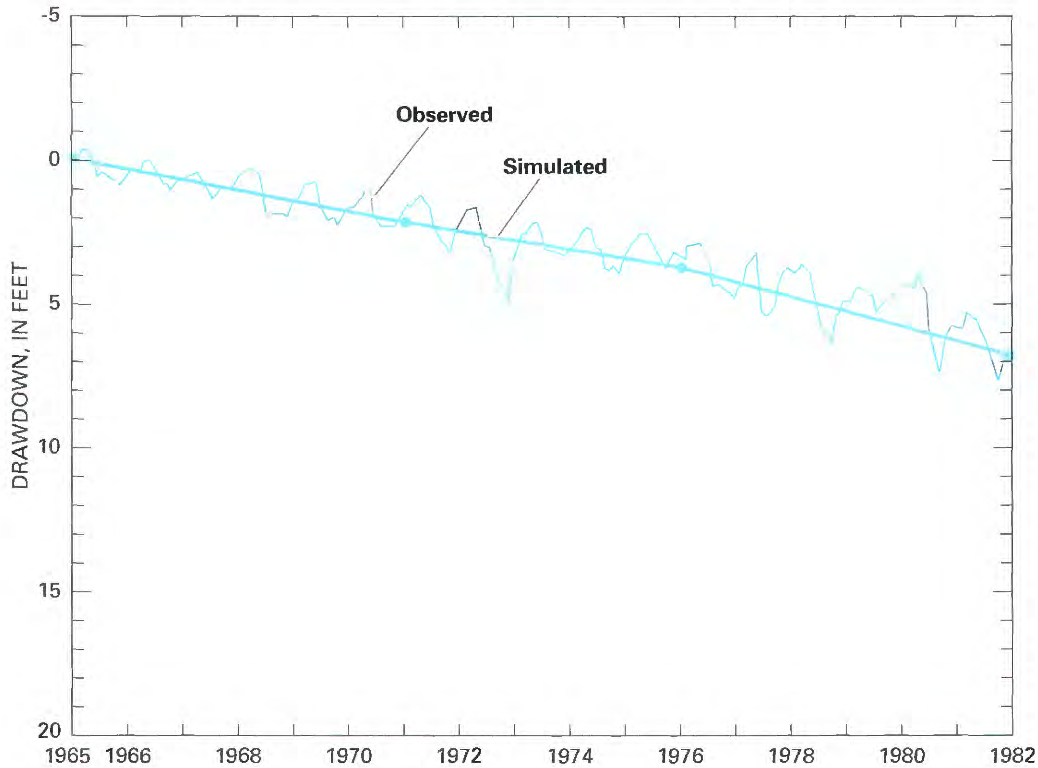


FIGURE 38.—Observed and model-simulated drawdowns in observation well GRE-3, Eutaw aquifer, 1965–81. Well location shown in figure 47.

to achieve sufficient total drawdown. In model layer A2 (Lisbon aquifer), transmissivity was reduced by about 25 percent from the western boundary to the area approximately representing the Pea River. In model layer A3 (Nanafalia-Clayton aquifer), transmissivity was reduced by about 20 percent across Marengo and northern Clarke Counties in Alabama. In model layer A4 (Providence-Ripley aquifer), transmissivity was reduced by about 20 percent from the Tombigbee River through Butler County, Ala., and by as much as 33 percent across Pike and Coffee Counties, Ala.

CALIBRATION RESULTS

Relatively long term water-level records are available for 67 wells in the Coastal Plain of Alabama: 60 with intermittent data and 7 with continuous record. Of the 67 wells, 7 showed no change in water level during the period 1964 through 1981 and 40 showed more than 5 ft of change. The remaining 20 wells showed drawdowns of between 0 and 5 ft. Hydrographs for the period 1965 through 1981 for six continuous-record wells showing more than 5 ft of drawdown, along with the simulated drawdowns for the three stress periods, are presented in figures 38 through 43. Of 34 intermittent-record wells that had more than 5 ft of change, 6 matched the change within 1 ft (with the maximum change being 17 ft) and 3 others matched the change within 2 ft.

Figures 44 through 48 show lines of equal simulated drawdown for each model layer for the period 1965 through 1981, as well as actual total drawdown data obtained from wells for which water-level data near the beginning and end of the period were available. General patterns of drawdown were approximated fairly well, but in a few areas the simulated drawdown exceeded that suggested by observed data. However, the actual potentiometric heads are comparable to simulated heads. This disparity—that the simulated drawdowns apparently were excessive, yet the simulated and measured heads in were comparable—may be due to poor head matches for these areas in the steady-state simulation. That is, the simulated steady-state heads that were used as starting heads for the transient simulation were larger than the measured steady-state heads. Because ground-water development predated water-level measurements in some areas, some of the measured heads used for the steady-state calibration may reflect prior stresses in those areas. Therefore, the simulated steady-state potentiometric surface may have been closer to the true predevelopment potentiometric surface than the surface defined by the observed measurements. Thus, the "excessive" drawdown simulated by the transient model may in fact be the correct amount.

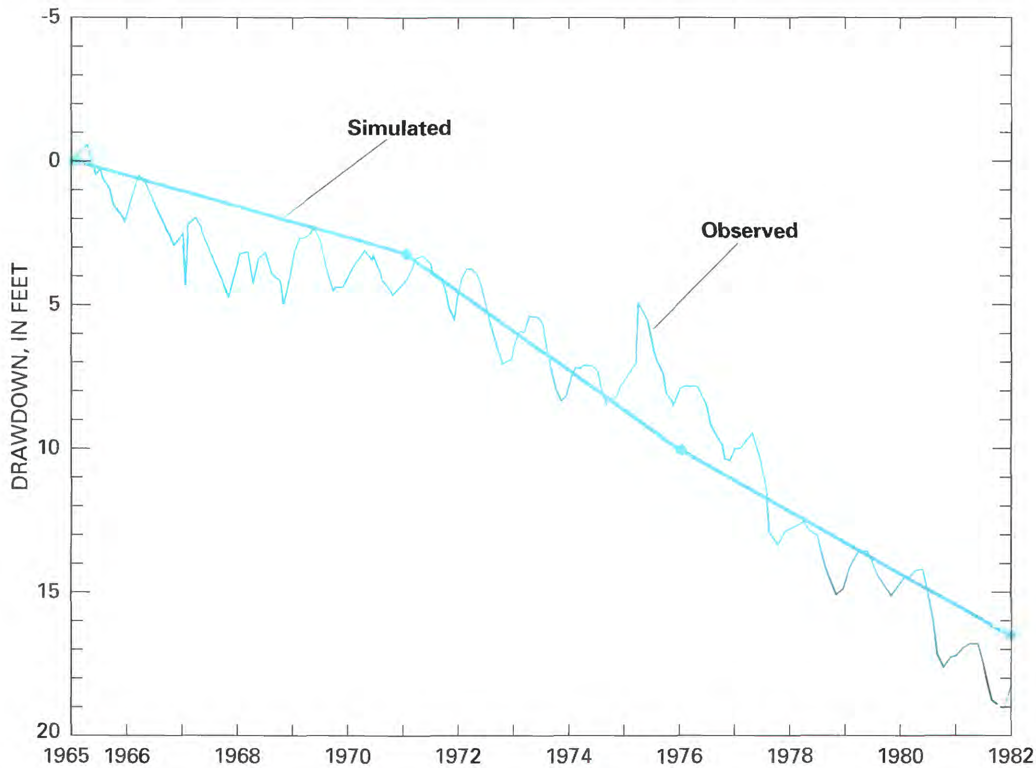


FIGURE 39.—Observed and model-simulated drawdowns in observation well HAL-1, Eutaw aquifer, 1965–81. Well location shown in figure 47.

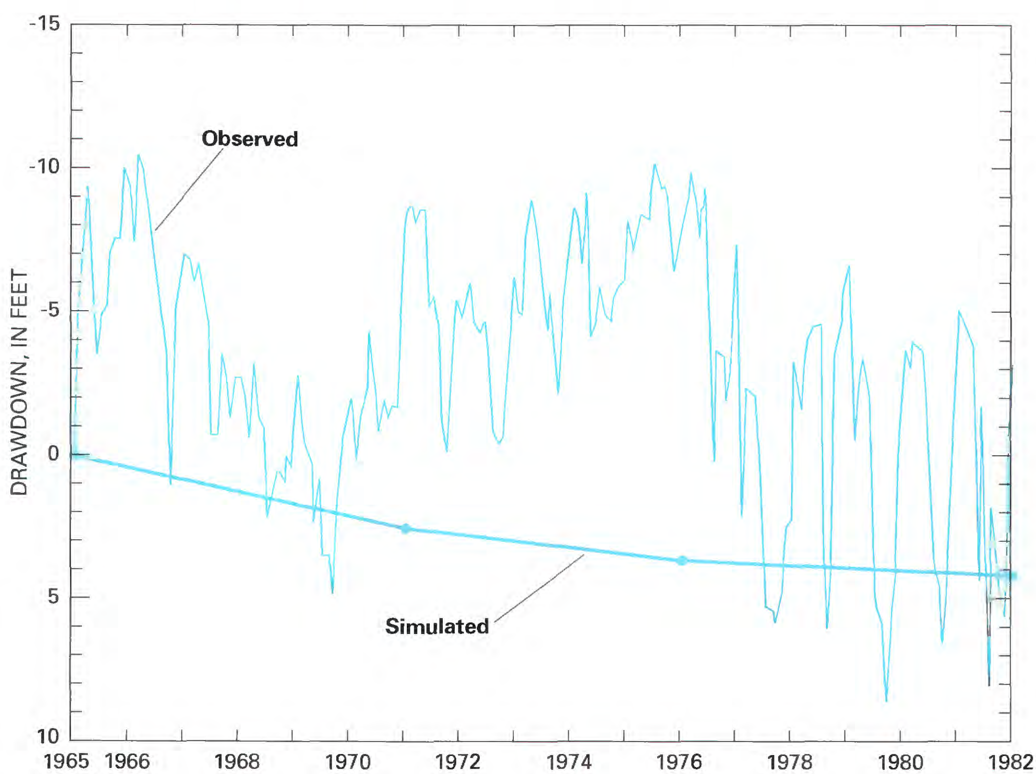


FIGURE 40.—Observed and model-simulated drawdowns in observation well MAG-1, Eutaw aquifer, 1965–81. Well location shown in figure 47.

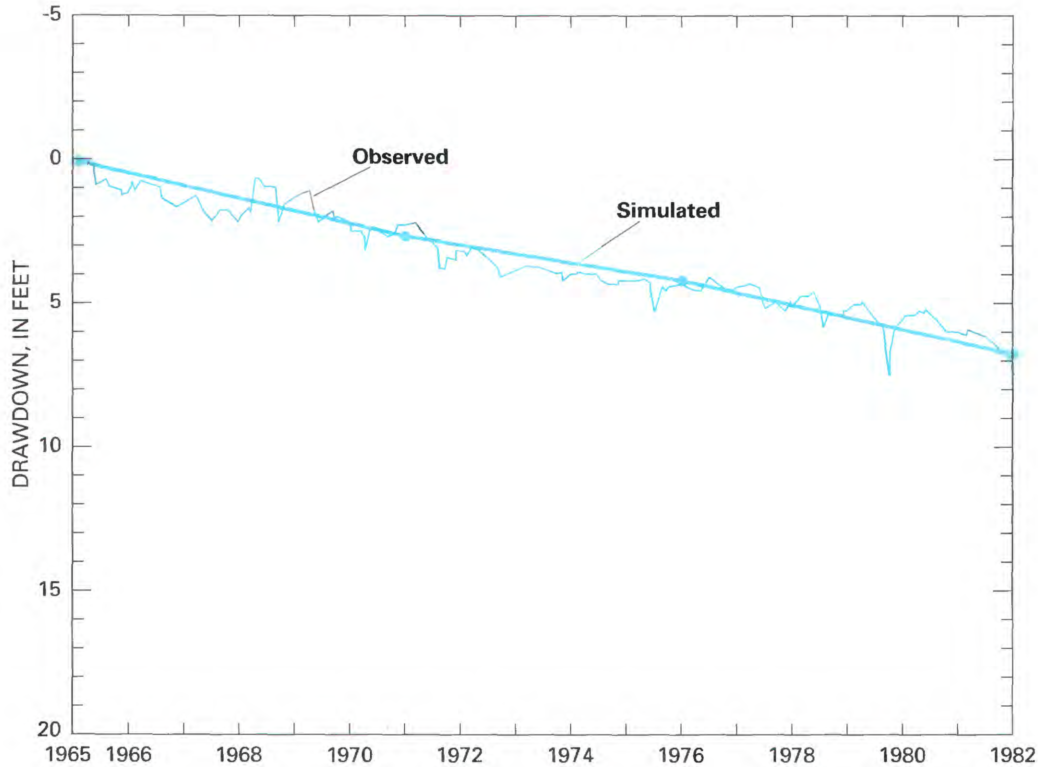


FIGURE 41.—Observed and model-simulated drawdowns in observation well MAG-2, Eutaw aquifer, 1965–81. Well location shown in figure 47.

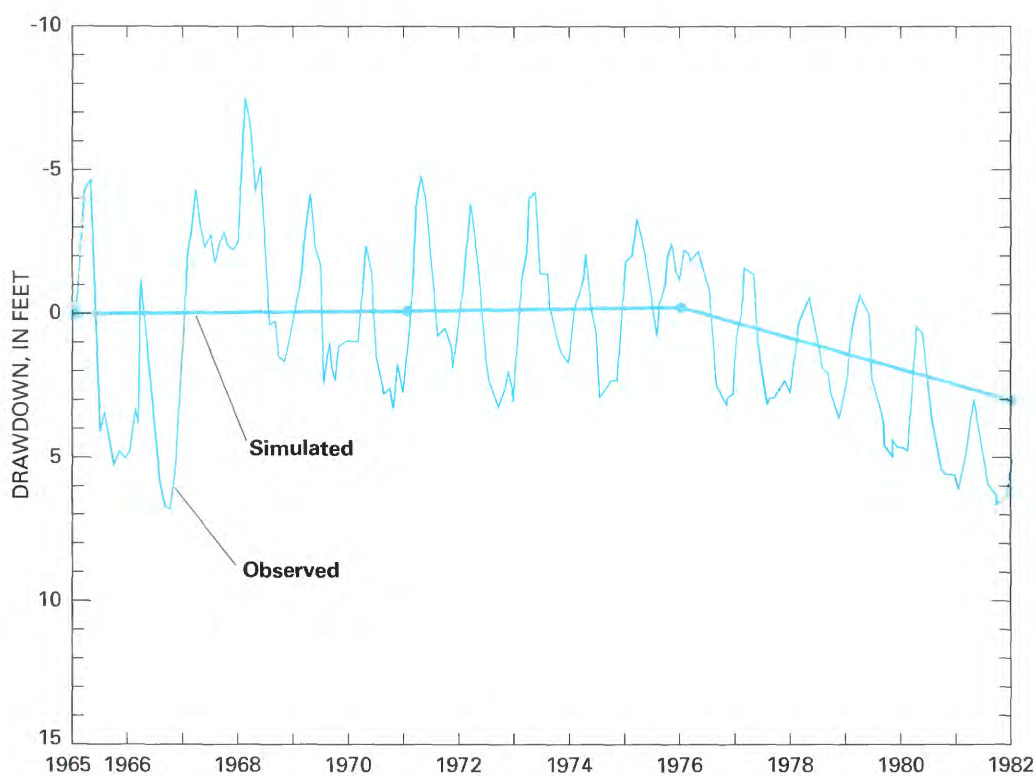


FIGURE 42.—Observed and model-simulated drawdowns in observation well MTG-3, Eutaw aquifer, 1965–81. Well location shown in figure 47.

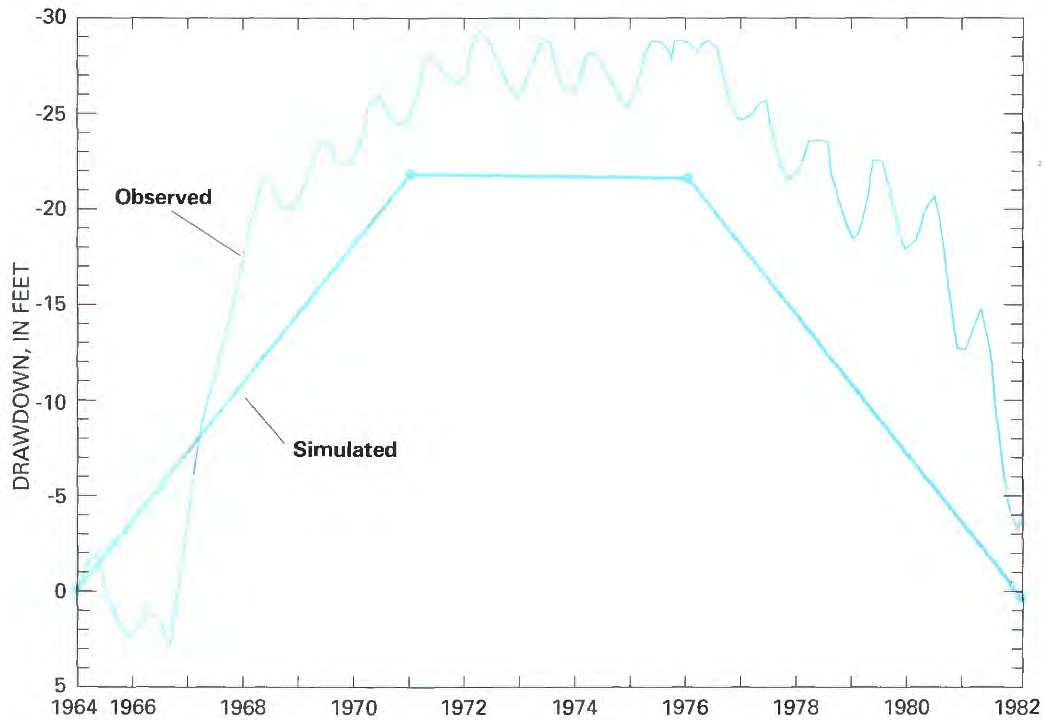


FIGURE 43.—Observed and model-simulated drawdowns in observation well MTG-5, Tuscaloosa aquifer, 1965–81. Well location shown in figure 48.

COMPARISON OF SIMULATED WATER BUDGETS UNDER PREDEVELOPMENT AND 1981 CONDITIONS

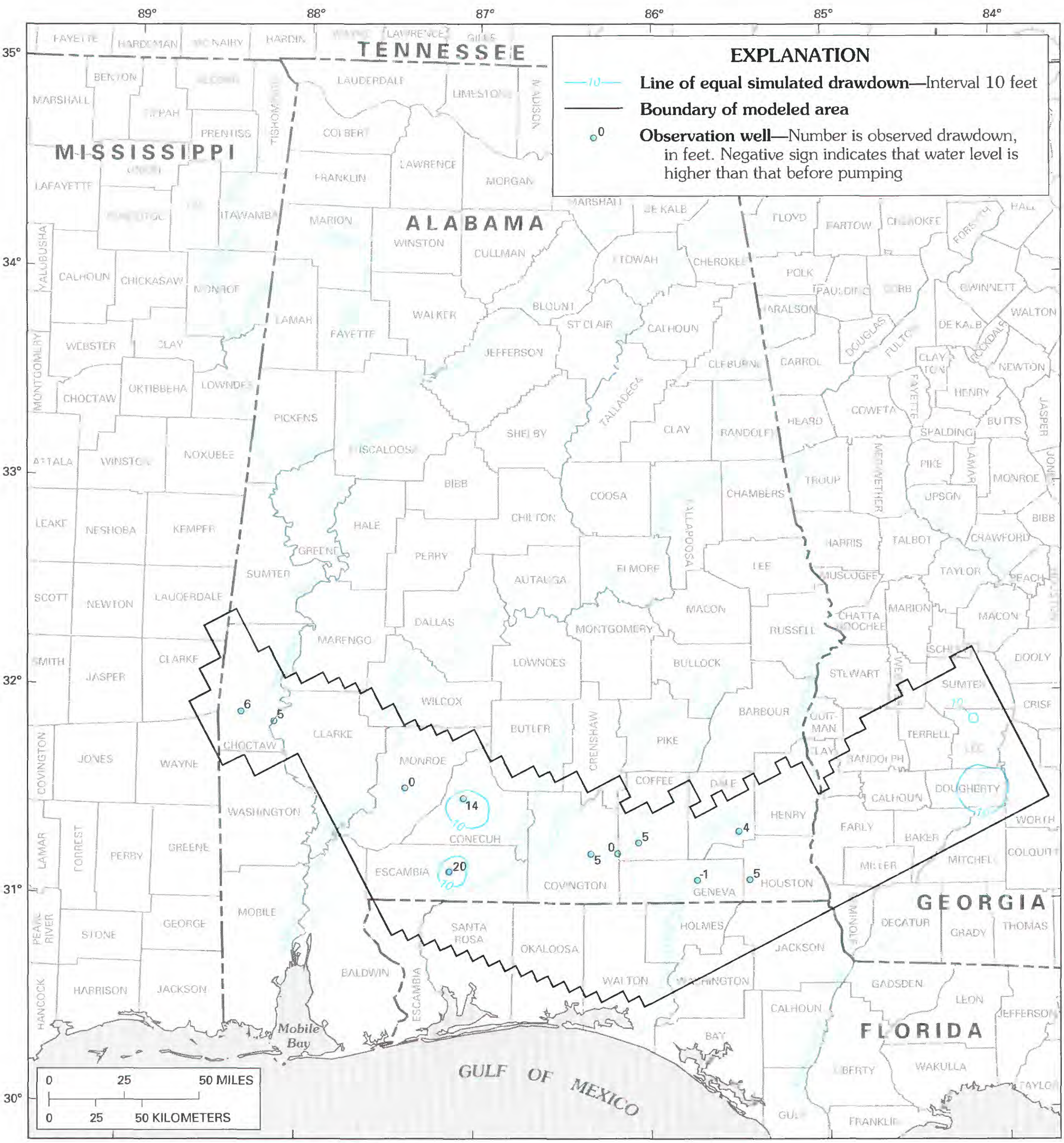
A comparison of the simulated predevelopment water budget for the aquifer system with the simulated water budget under 1981 conditions (fig. 49) indicates that a pumping rate of 175 Mgal/d induced an additional 80 Mgal/d of water into the ground-water flow system. The sources of the pumpage were (1) capture of water formerly lost to layer A1, the source-sink layer (coastal lowlands aquifer system and Upper Floridan aquifers) (net 17 Mgal/d, or 10 percent), (2) inducement of additional water through the lateral model boundaries (net 33 Mgal/d, or 19 percent), (3) reduction in aquifer storage (34 Mgal/d, or 19 percent), and (4) capture of water discharging to rivers (91 Mgal/d, or 52 percent). The large contribution of water to the pumping rate from storage indicates that the simulated aquifer system had not reattained equilibrium conditions as of the end of 1981.

The contribution of capture of water formerly lost to layer A1, the source-sink layer (coastal lowlands aquifer system and Upper Floridan aquifer) is almost completely to the pumpage from model layers A2, A3, and A4 (Lisbon, Nanafalia-Clayton, and Providence-Ripley aquifers), providing about 25 percent (22 of 87 Mgal/d) of the total pumpage for those three model layers. This is because model layers A5 and A6 (Eutaw and Tuscaloosa aquifers) are nearly isolated from the three upper model layers.

PREDICTIVE SIMULATIONS

Two additional simulations were made to determine the length of time needed to approach new equilibrium conditions under the present scheme of pumping. The final stress period was extended 20 yr in one simulation and 50 yr in a second simulation (total simulation times of 47 and 77 yr). The volume of water derived from storage decreased to 12 percent of the pumping rate in the first simulation and 8 percent in the second simulation. Additional drawdowns that may be expected after the additional 50 yr at the current pumping rates at pumping centers in Alabama are 3 ft at Dothan, 3 ft at Montgomery, 3 ft at Fort Rucker-Enterprise, Dale and Coffee Counties, 2 ft at Selma, and 2 ft at Demopolis. These values represent regional drawdowns and do not account for well interference that may affect rates of drawdown on a local scale.

A third simulation was performed in which pumping rates were increased by 37.5 percent of the 1981 pumping rate at the large pumping centers of Montgomery, Selma, and Dothan, and by 10 percent at all other Alabama pumping centers. This resulted in a model-wide increase of 18 percent of the 1981 pumping rate. The simulation was again run an additional 50 yr (a total simulation time of 77 yr). Under this pumping scheme,

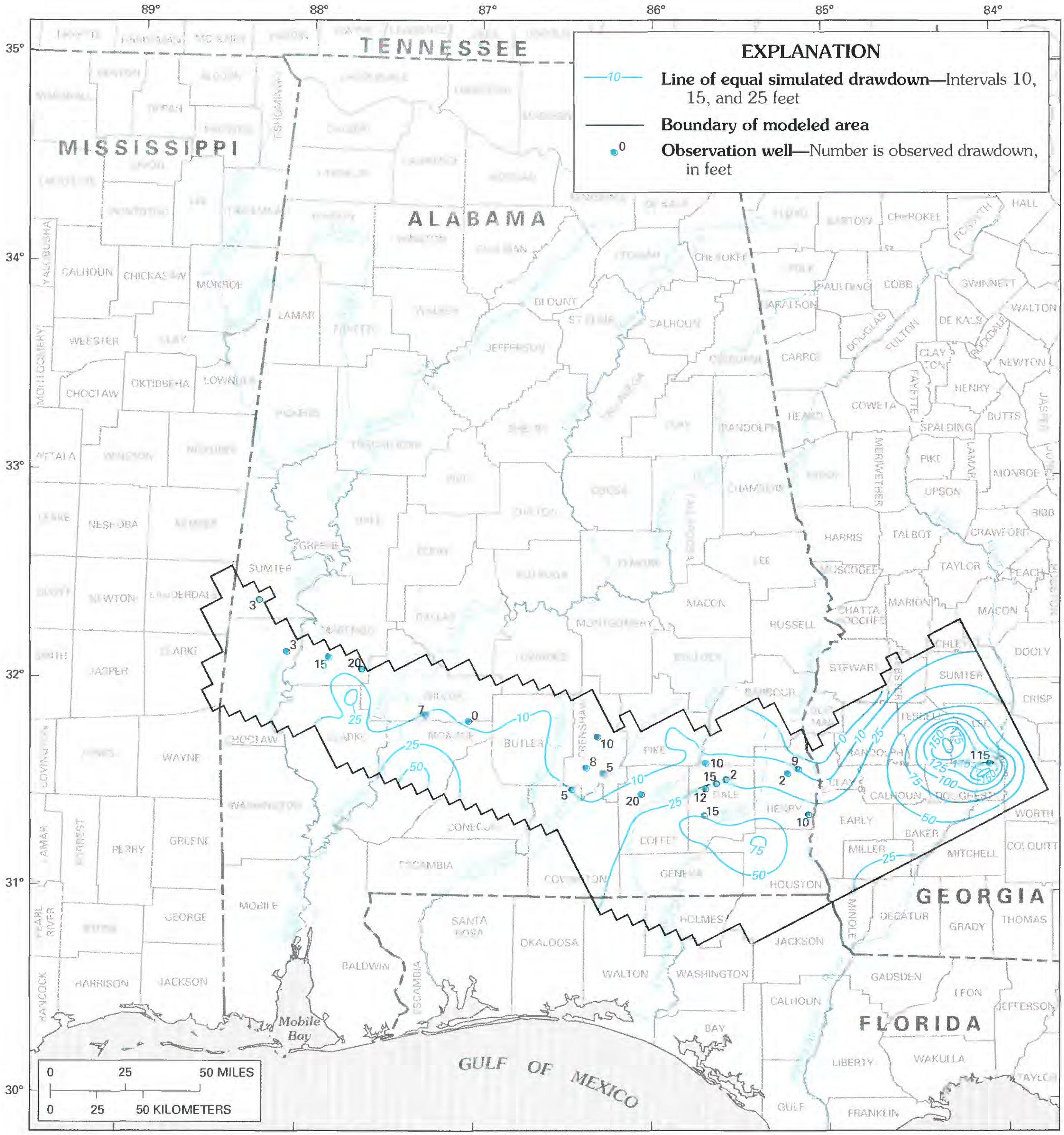


Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 44.—Simulated drawdown in model layer A2 (Lisbon aquifer) and actual drawdowns in selected observation wells, 1965–81.

the volume of water derived from storage during the final time step was 9 percent of the total. Drawdowns at some pumping centers increased markedly under the new pumping scheme (figs. 50–54). The predicted additional nodal drawdowns at Dothan, Montgomery, Fort Rucker-Enterprise, Selma, and Demopolis after 50 yr at the increased withdrawal rates were 37, 33, 17, 12, and 4 ft, respectively.

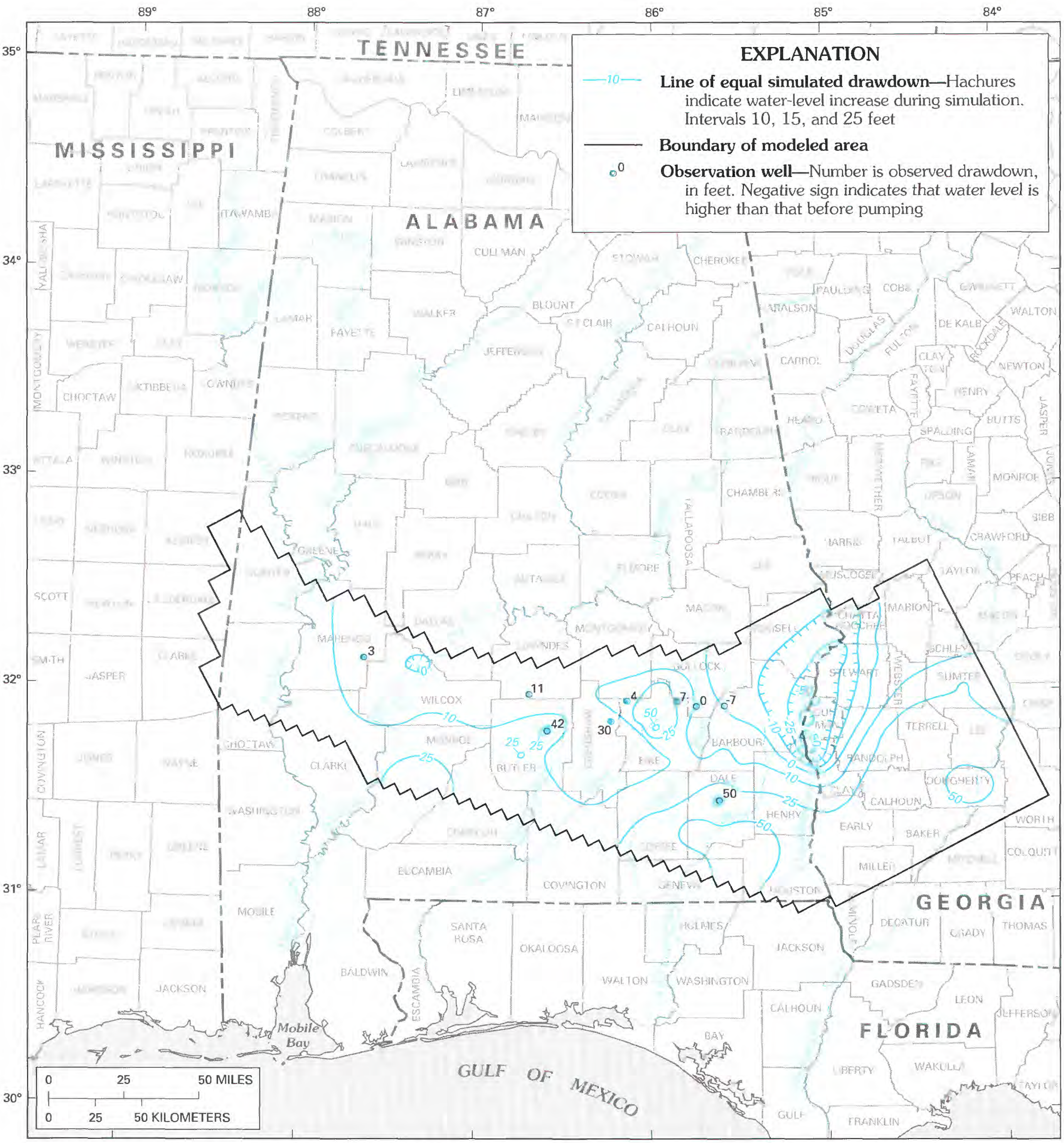
The simulations indicated that the system might approach equilibrium conditions when the volume of



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 45.—Simulated drawdown in model layer A3 (Nanafalia-Clayton aquifer) and actual drawdowns in selected observation wells, 1965–81.

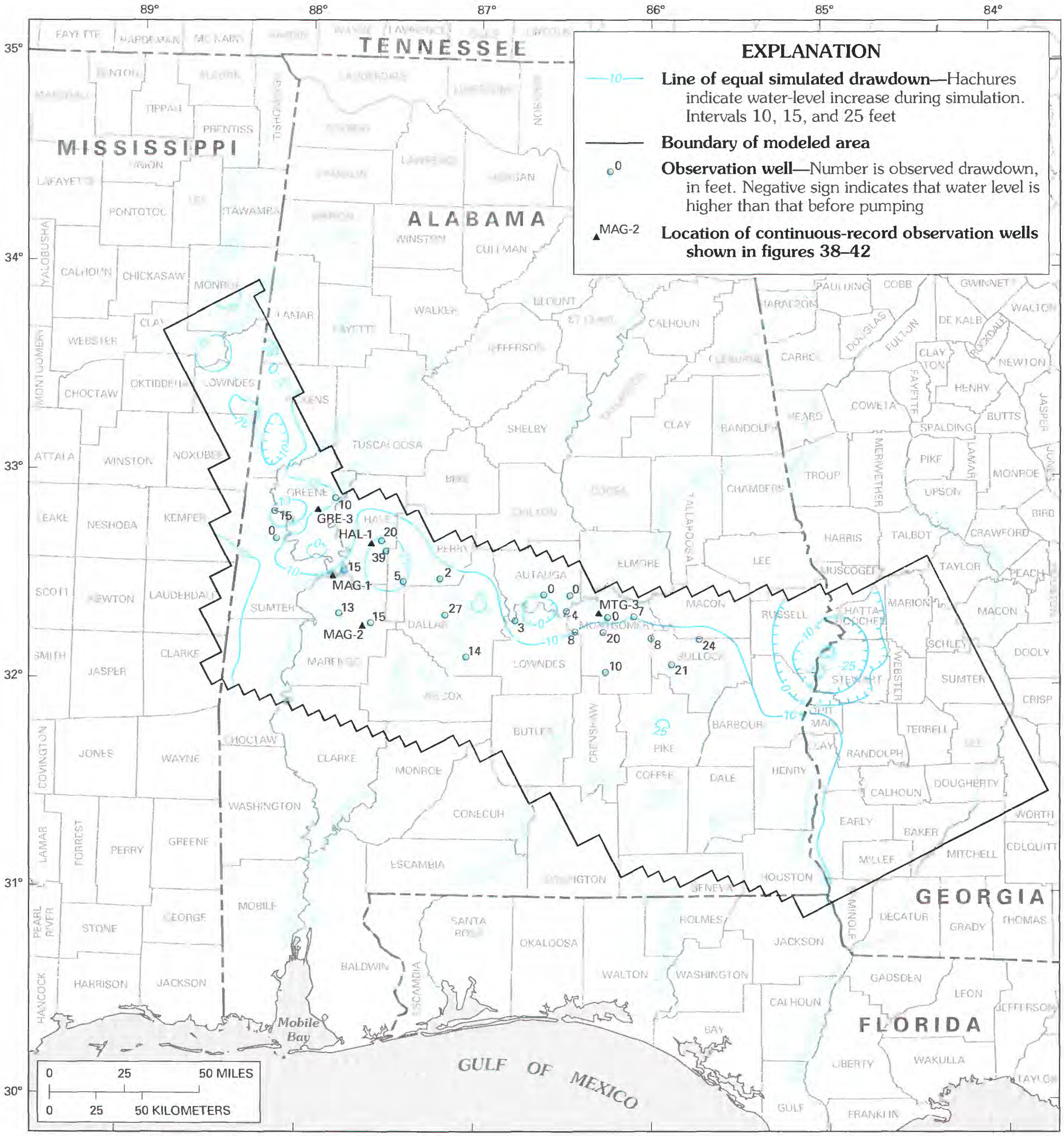
water taken from storage falls below 10 percent of the pumping rate. Without an increase in pumpage, the major pumping centers experienced only 3 ft of additional drawdown in 50 yr while the volume of water supplied from storage declined from 18 to 8 percent. Furthermore, by examining the amounts of decline during the first and final time steps of the 50-yr simulations, it can be inferred that the aquifer system under the increased-pumpage simulation was also approaching equilibrium conditions.



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 46.—Simulated drawdown in model layer A4 (Providence-Ripley aquifer) and actual drawdowns in selected observation wells, 1965–81.

To determine when equilibrium is actually attained and where no pumpage is derived from ground-water storage, the percentage of water derived from storage versus the simulated period of time can be plotted and the resultant curve extrapolated. Figures 55 and 56 show the change in storage for both 50-yr simulations (no increase and an 18-percent overall increase over 1981 pumpage). At the end of the simulations, the curves have essentially the same slope, confirming that the aquifer system is reacting in the same manner for both simula-

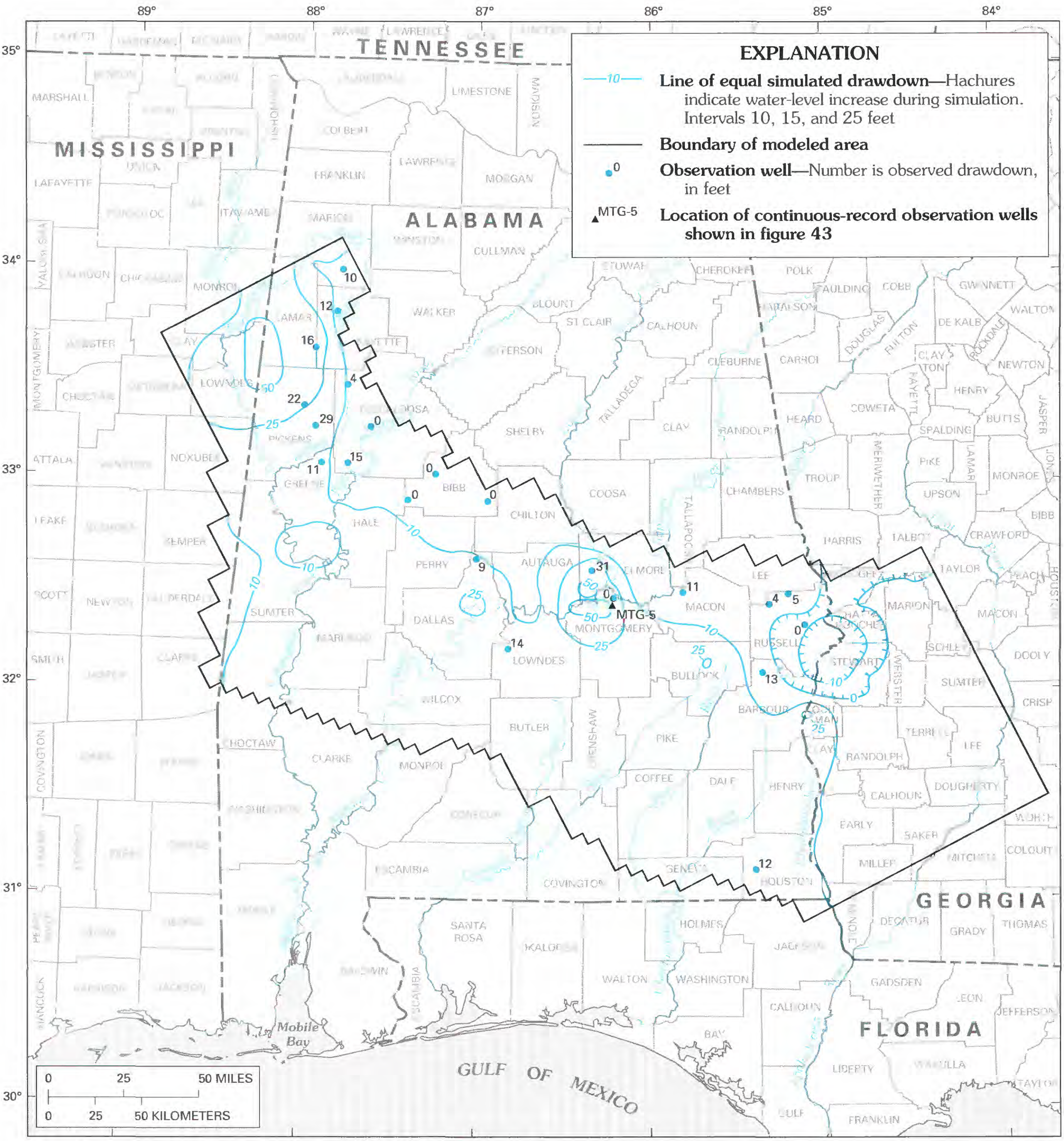


Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 47.—Simulated drawdown in model layer A5 (Eutaw aquifer) and actual drawdowns in selected observation wells, 1965–81.

tions despite the differences in pumpage. Assuming a linear relation between the final two time steps of a 2-percent change in storage for approximately 20 yr, it would require an additional 80 yr beyond the 50-yr stress period to reach true equilibrium, which would put the date at 2111.

Pumping centers in southeastern Alabama (Fort Rucker-Dothan area) can be used to demonstrate that



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 48.—Simulated drawdown in model layer A6 (Tuscaloosa aquifer) and actual drawdowns in selected observation wells, 1965–81.

the aquifer system is approaching a new equilibrium. This area has experienced the greatest water-level declines in Alabama and is likely to continue to be the most heavily pumped area in the near future. Table 4 provides information on the additional drawdowns that may be expected after 50 yr, given the two conditions simulated (no increase in pumping and a 37.5-percent increase over the 1981 pumping rates at major pumping centers). The seven node locations correspond to the pumping centers at Dothan, Fort Rucker, Daleville, and

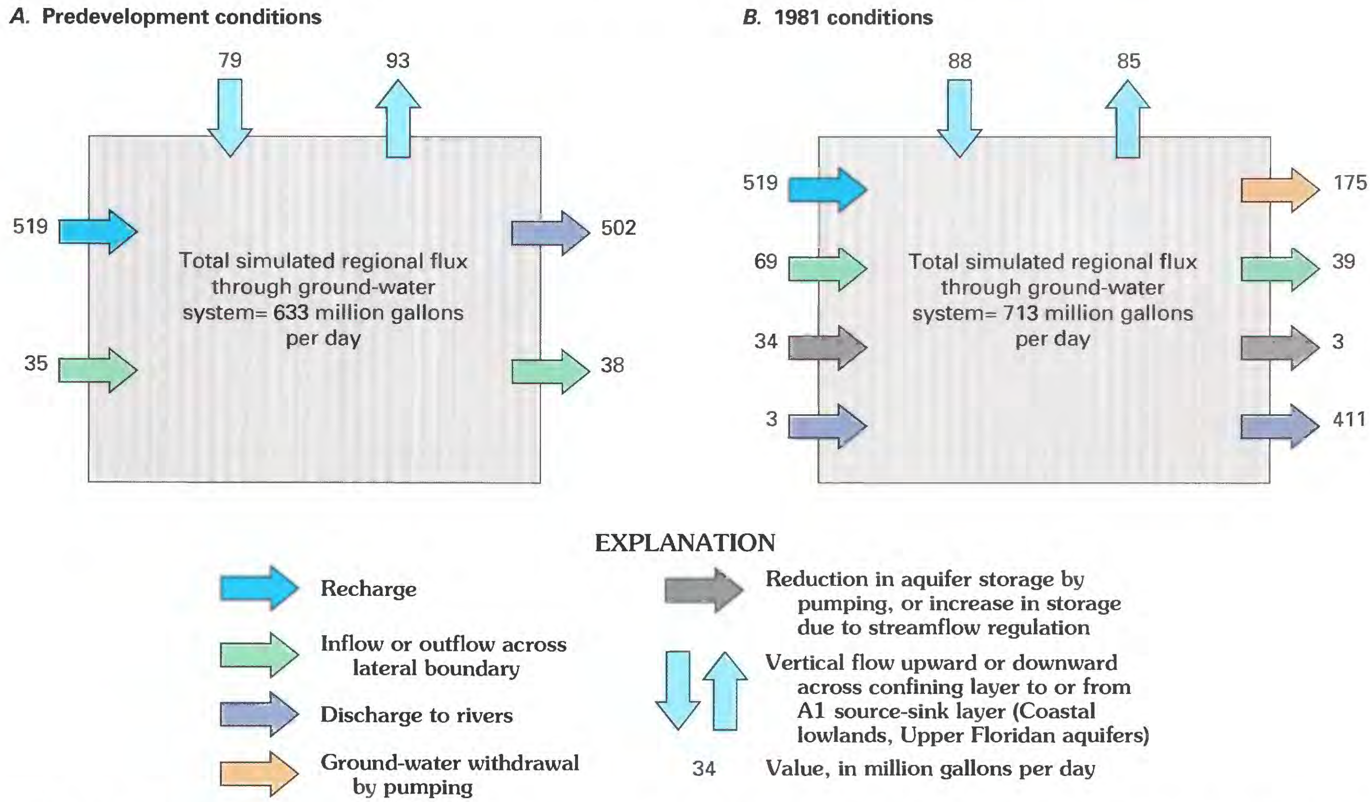


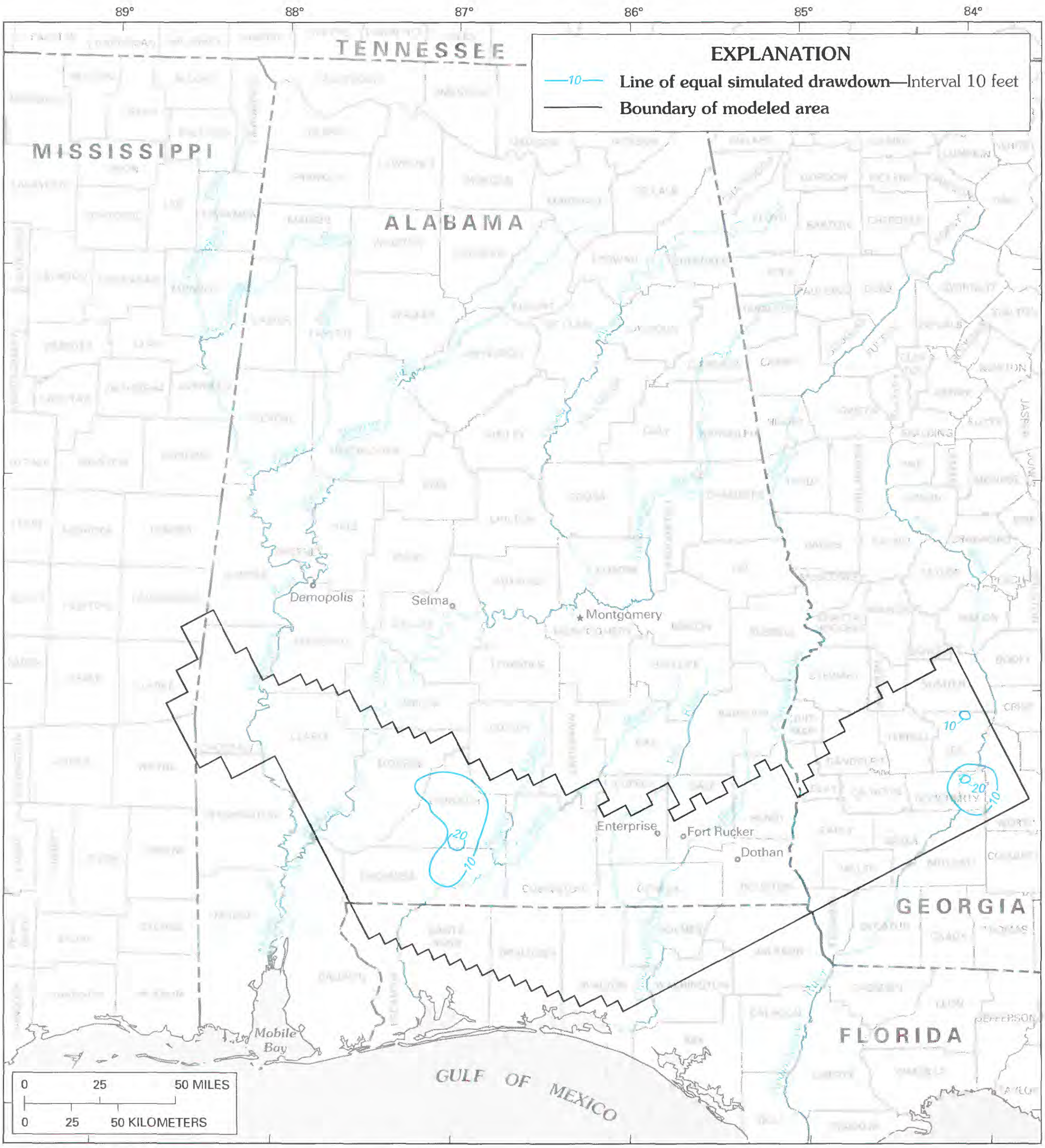
FIGURE 49.—Simulated water budget for the Alabama subregional model, showing ground-water flow-system components and flux rates under predevelopment and 1981 conditions.

Enterprise, Ala. The first time step for this final pumping period in the model was approximately 4 yr (1982–85); the maximum drawdown during these 4 yr was about 1.5 ft at Dothan with no increase in pumping rates, and 34 ft with a 37.5-percent increase over 1981 pumping rates. The final time step simulated approximately 20 yr of pumpage (2012–31); the additional maximum drawdown, again at Dothan, was about 0.3 ft with no increase over 1981 pumping rates, and 0.4 ft with the 37.5-percent increase over 1981 pumping rates. Because the additional drawdowns during the final 20 yr of simulation are so small (a yearly rate of 0.02 ft/yr), the

TABLE 4.—*Simulated drawdown, in feet, in southeastern Alabama under differing pumping schemes*

Node location			Drawdown				
			1982–85		2012–2031		1982–2031
			Pumping rate				
Row	Column	Layer	Same as 1981	Increased rate[1]	Same as 1981	Increased rate[1]	Increased rate[1]
			Dothan				
60	37	A3	1.29	19.04	0.27	0.36	25
61	37	A3	1.44	33.57	.28	.37	37
60	37	A4	1.31	23.85	.29	.37	27
			Fort Rucker				
57	33	A3	.90	9.13	.24	.31	13
57	33	A4	.95	9.40	.25	.32	13
			Daleville				
58	34	A3	1.01	8.78	.25	.32	13
			Enterprise				
57	32	A3	.92	9.09	.22	.29	13

[1]Rate of increase: Dothan, 37.5 percent; Fort Rucker, Daleville, Enterprise, 10 percent.

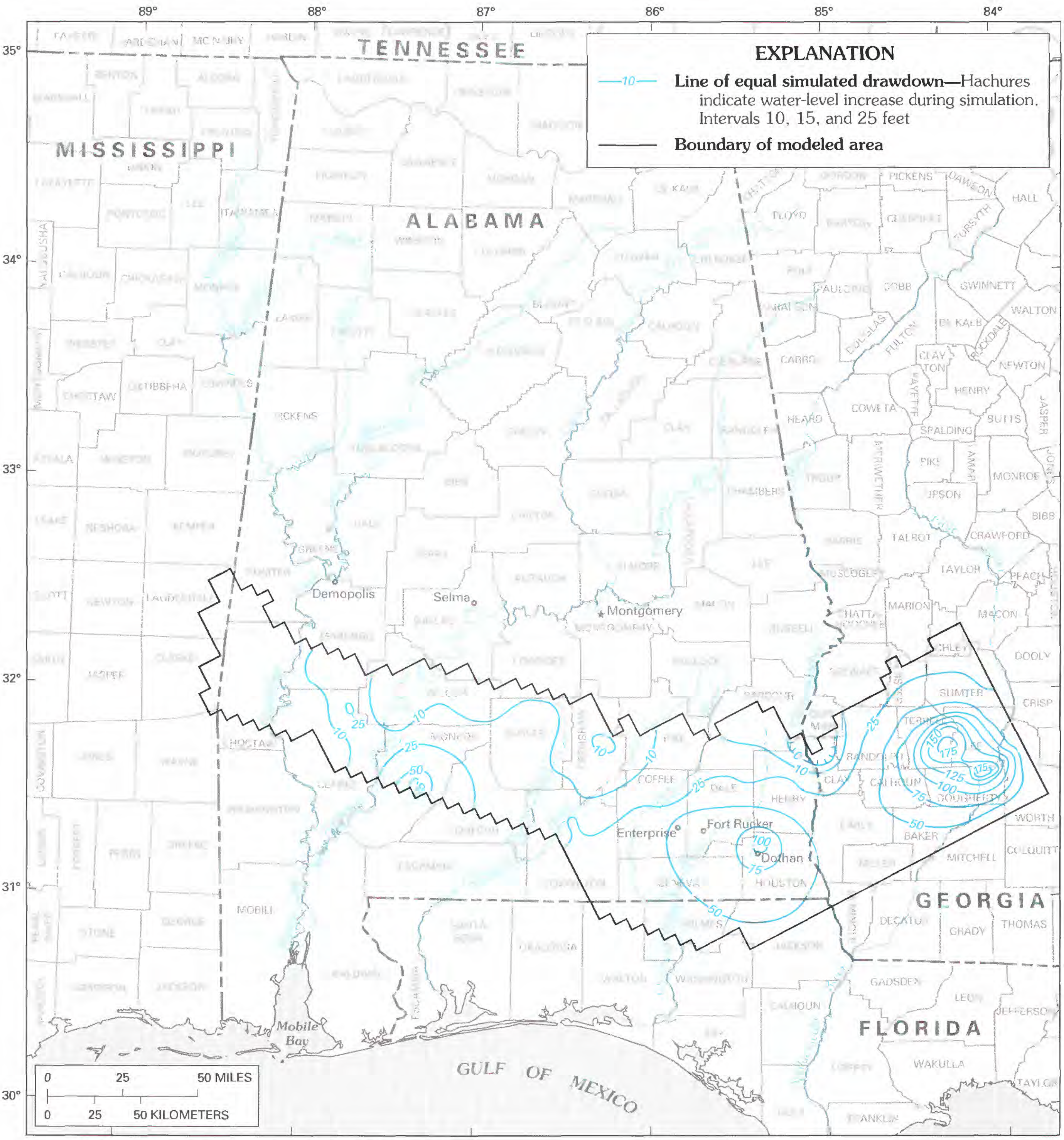


Base modified from U.S. Geological Survey digital data, 1:2,000,000. 1972

FIGURE 50.—Simulated drawdown in model layer A2 (Lisbon aquifer) after 50 yr of projected increased withdrawals.

authors conclude that for practical purposes the aquifer system would attain equilibrium conditions within the simulated 50 yr.

It should also be noted that, in the Dothan-Fort Rucker area, the maximum drawdown (34 ft) due to an increase over 1981 pumping rates at the end of 4 yr is approximately 90 percent of the total drawdown (37 ft) for the entire 50-yr stress period, 1982–2031. This clearly demonstrates that with an increase in pumping rate, water is initially taken from storage near the pumping



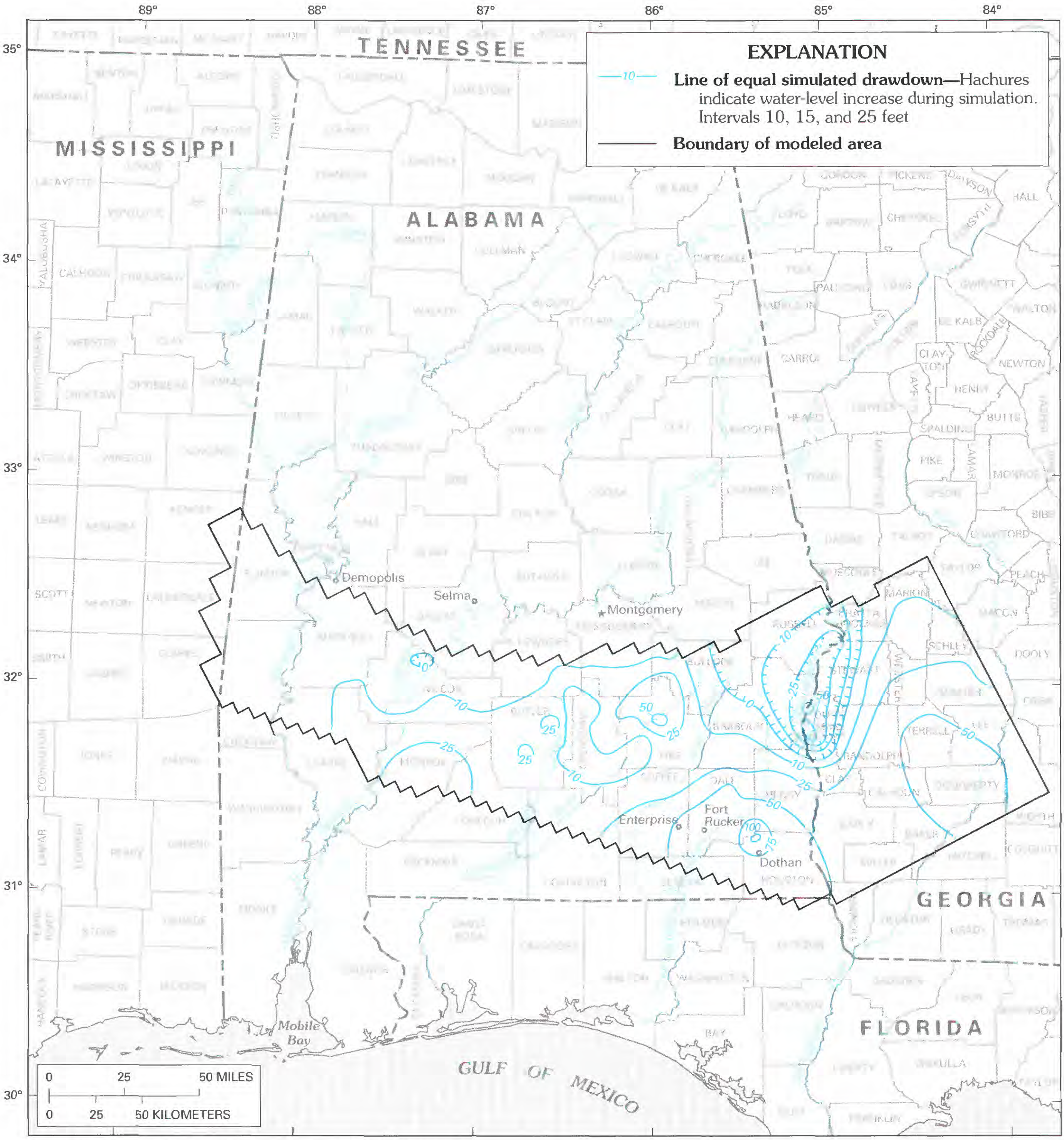
Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 51.—Simulated drawdown in model layer A3 (Nanafalia-Clayton aquifer) after 50 yr of projected increased withdrawals.

centers to create the increased hydraulic gradient in the vicinity of the pumping centers that is needed to supply water to the wells, but that as the area affected by pumping spreads, the rate of water-level change diminishes as the wells draw water from an expanding area of influence.

SUMMARY

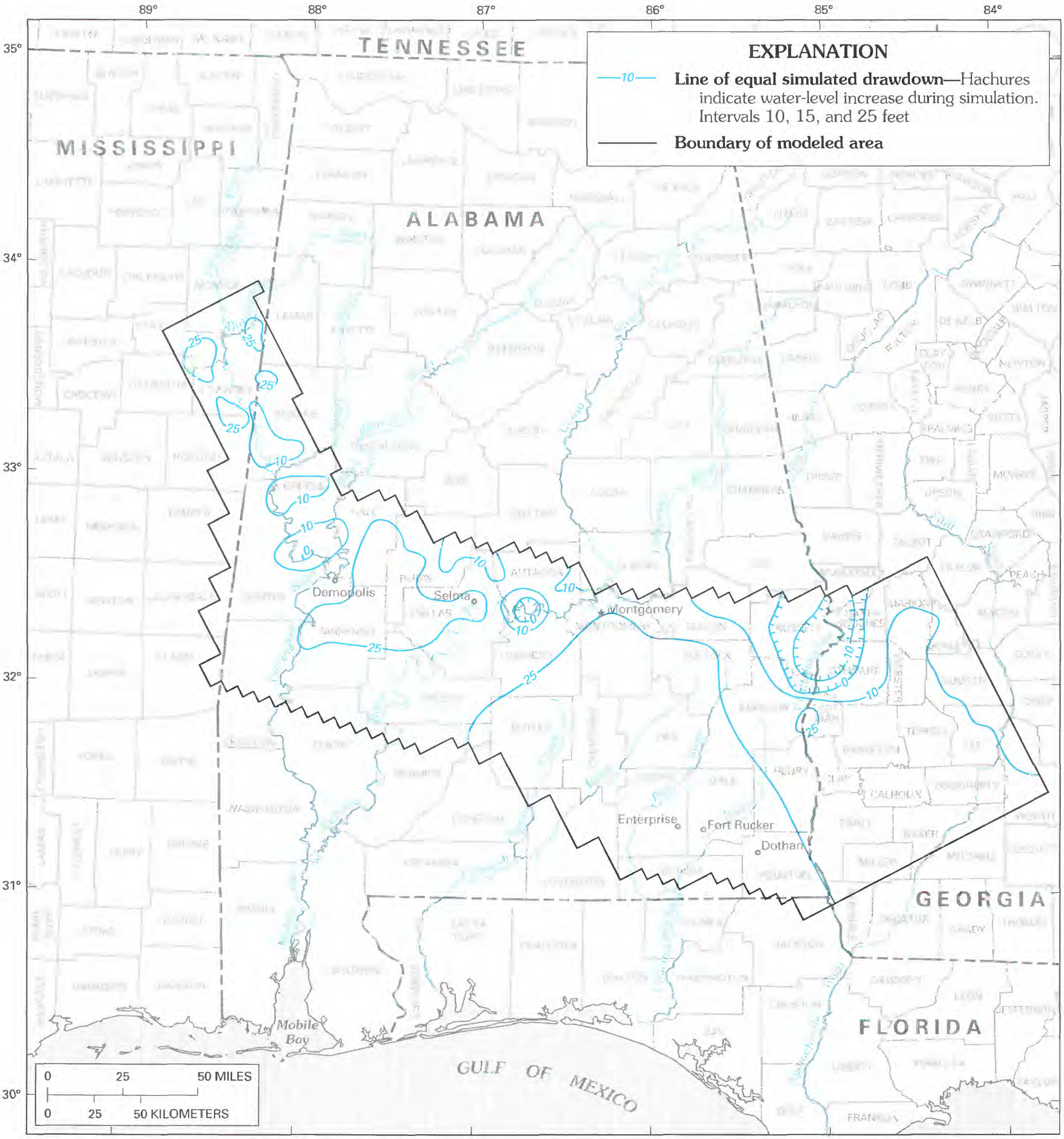
In 1980, the U.S. Geological Survey began a study of the Southeastern Coastal Plain aquifer system as part of its Regional Aquifer-System Analysis (RASA) program. The Southeastern Coastal Plain aquifer system is com-



Base modified from U.S. Geological Survey digital data. 1:2,000,000, 1972

FIGURE 52.—Simulated drawdown in model layer A4 (Providence-Ripley aquifer) after 50 yr of projected increased withdrawals.

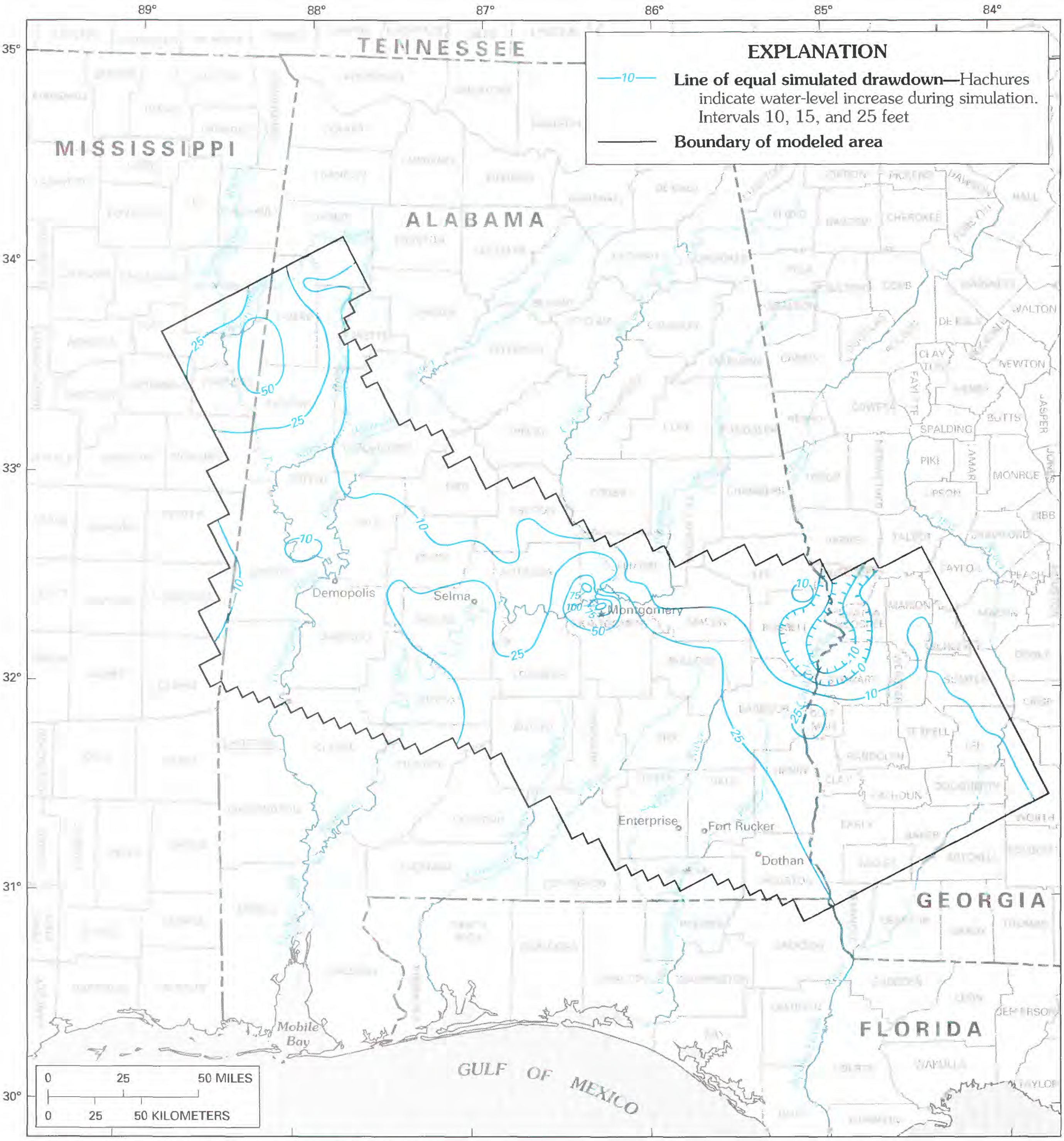
posed mainly of sand aquifers in rocks of Cretaceous and Tertiary ages in Alabama, Georgia, Mississippi, and South Carolina. Metamorphic, igneous, and sedimentary rocks of Paleozoic and early Mesozoic age underlie and form the base of the aquifer system. The Southeastern Coastal Plain study was accomplished by dividing the region into four subregions that roughly coincide with State boundaries, but using hydrologic boundaries as limits of the subregional studies. The Alabama subregional study encompasses about 46,500 mi^2.



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 53.—Simulated drawdown in model layer A5 (Eutaw aquifer) after 50 yr of projected increased withdrawals.

The Southeastern Coastal Plain aquifer system within the Alabama subregional study area is composed of sediments that can be divided into a series of aquifers and confining units. The sediments are unconsolidated sand, clay, and gravel, and consolidated or semiconsolidated limestone and chalk. Sediments underlying shallow aquifers not directly studied by this project (coastal lowlands aquifer system and Upper Floridan aquifer) were divided into five aquifers. Upper Cretaceous rocks make up three aquifers: (1) the Tuscaloosa aquifer,



Base modified from U.S. Geological Survey digital data, 1:2,000,000, 1972

FIGURE 54.—Simulated drawdown in model layer A6 (Tuscaloosa aquifer) after 50 yr of projected increased withdrawals.

comprising the Coker and Gordo Formations, (2) the Eutaw aquifer, comprising the Eutaw and McShan Formations, (3) and the Providence-Ripley aquifer, comprising the Providence Sand and the Ripley and Blufftown Formations. The Gordo and Selma confining units that separate the aquifers in Cretaceous rocks are composed of marine clay (Gordo) or chalk (Selma); the chalk has the lowest leakance values within the subregion's aquifer system. Tertiary rocks make up the two remaining aquifers: (1) the Nanafalia-Clayton aquifer, comprising

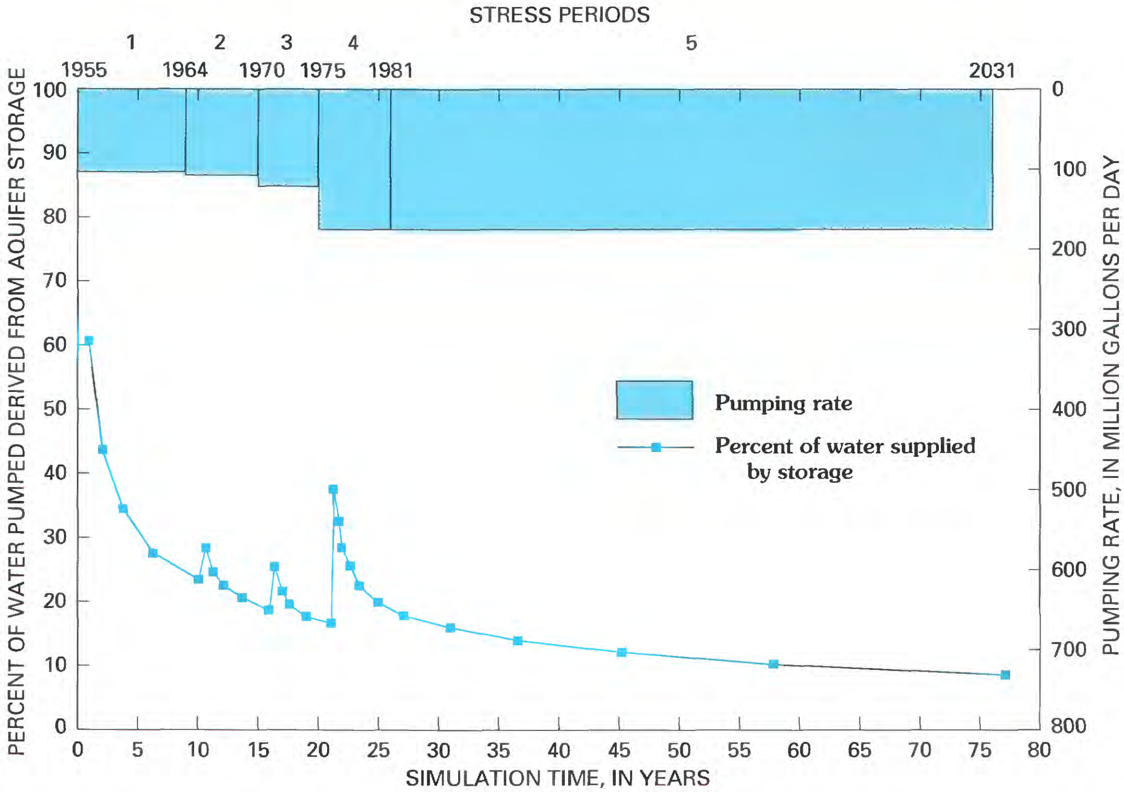


FIGURE 55.—Percentage of pumpage derived from ground-water storage, 1955–2031, with no increase in pumping for the period 1982–2031.

the Clayton, Naheola, and Nanafalia Formations and the Baker Hill Formation, the updip equivalent of the Nanafalia Formation, plus the lower part of the Tuscahoma Formation, and (2) the Lisbon aquifer, comprising the upper part of the Tuscahoma Formation plus the Bashi, Hatchetigbee, Tallahatta, and Lisbon Formations, the Gosport Sand, and part of the Moodys Branch Formation. The Tuscahoma and Yazoo confining units separating the aquifers in Tertiary rocks are composed predominantly of clay.

The philosophy adopted in modeling the aquifers in the Southeastern Coastal Plain was that the ground-water flow could be analyzed at three levels or scales: local, intermediate, and regional. It was reasoned that, because only the major drains could be properly simulated with the large grid spacings used in the flow model, not all the recharge that supplies the local and intermediate flow systems could be included in the simulation. Therefore, only a regional component of ground-water discharge to regional drains, which was determined by estimates of base flow in large rivers during the 1954 drought, was considered during the simulation.

The finite-difference grid for the Alabama subregional model consisted of 68 rows and 60 columns, with each cell 4 mi on a side and representing an area of 16 mi^2. The flow system was modeled in six layers vertically—five active aquifer layers and an overlying source-sink layer (coastal lowlands aquifer system and Upper Floridan aquifer). Simulated vertical flow between adjacent aquifer layers was controlled by areally varying vertical leakage in intervening layers. The model inputs varied during model calibration were transmissivity, leakance, recharge, and riverbed conductance. The model was calibrated to match water-level measurements made before significant ground-water withdrawals began. The criterion for an acceptable head match was set at 25 ft. A minimum 1-day streamflow in regional drains observed during a 1954 drought was also used for comparison of base flow during calibration.

The overall match between observed and simulated predevelopment heads for all layers was 81 percent (365 of 451) within the 25-ft criterion. Success in matching base flow was mixed. The potential error in estimating regional base flow from observations of flow made during the 1954 drought may be large, and this may account for the disparities between the calculated and simulated base flows. The ground-water flow patterns in the aquifers indicated by the simulated potentiometric surfaces agree well with conceptual flow patterns developed for the outcrop and near-outcrop areas prior to the modeling effort. However, simulation has indicated a different flow pattern for downdip, confined areas. Specifically,

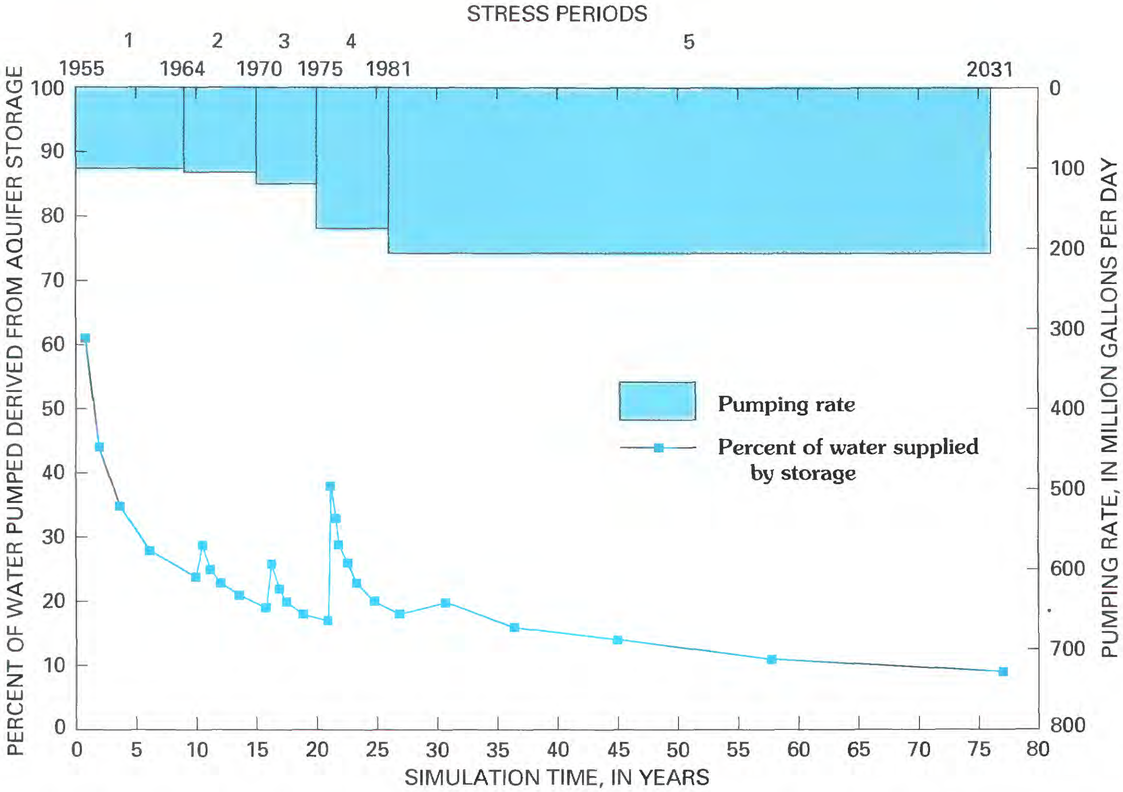


FIGURE 56.—Percentage of pumpage derived from ground-water storage, 1955–2031, with an 18-percent overall increase in pumping for the period 1982–2031.

simulation indicates a large amount of water flowing into the study area from both the east and west sides of the model. A previous model of the Eutaw and Tuscaloosa aquifers in western and central Alabama had simulated the same flow patterns as this subregional model. However, the pattern had not been recognized regionally until the current modeling effort.

The approach in the transient modeling was to attempt to duplicate the response to pumpage from wells during each of three pumping periods between 1964 and 1981: 1964 through 1970, 1971 through 1975, and 1976 through 1981. The calibration procedure involved approximating rates of drawdown in observation wells for each pumping period. Success at approximating drawdown rates in continuous-record wells was good. However, in areas for which only intermittent measurements were available, results were mixed. Therefore, the approach to calibration was modified, duplicating drawdown rates when possible, but at least approximating the total drawdown for areas in which only intermittent measurements had been made.

Of 6 wells with continuous record and 34 intermittent-measurement wells that had more than 5 ft of drawdown, simulated and observed drawdown rates matched within 1 ft in 13 wells, and within 2 ft in 3 others. General patterns of total drawdown defined by 86 observation wells were simulated reasonably well, although in a few areas observed drawdowns were less than drawdowns predicted by the model.

As simulated, four sources of water contributed to the pumpage—capture of flow that would normally discharge to rivers, induced flow from outside the borders of the model, capture of water from the source-sink layer, and water derived from storage within the aquifer. Pumpage from the aquifer system was simulated at 175 Mgal/d; 52 percent (91 Mgal/d) was derived from discharge to rivers, 19 percent (33 Mgal/d) was flow induced from outside the model borders, 10 percent (17 Mgal/d) was captured from water formerly lost to the source-sink layer, and 19 percent (34 Mgal/d) was obtained from storage. The relatively large contribution from storage indicates that the simulated aquifer system had not reached a new equilibrium by the end of 1981.

Additional simulations were made to determine the length of time needed to reestablish equilibrium conditions in the aquifer system. At the 1981 pumping rates, the final pumping period was extended for 50 yr, from 1982 to 2031 (total simulation time 77 yr). Those simulations indicate that drawdowns that may be expected after an additional 50 yr at pumping centers in Alabama are 3 ft at Dothan, 3 ft in the Fort Rucker-Enterprise area, 3 ft at Montgomery, 2 ft at Selma, and 2 ft at

Demopolis. The volume of water derived from storage in the final time step was 8 percent. This value indicates that, at the 1981 pumping rates, the aquifer system is approaching equilibrium conditions, with water levels changing at a rate of less than 0.5 ft per 20 yr.

Another simulation was performed in which withdrawal rates were increased by 37.5 percent over the 1981 pumping rate at the large pumping centers in Montgomery, Selma, and Dothan, and by 10 percent over the 1981 pumping rate at all other Alabama pumping centers (a model-wide increase of 18 percent) and was again run an additional 50 yr. Drawdowns at some pumping centers increased markedly under this greater withdrawal rate. The predicted additional nodal drawdowns at Dothan, Montgomery, Fort Rucker-Enterprise, Selma, and Demopolis after 50 yr at the increased pumping rates were 37, 33, 17, 12, and 4 ft, respectively. Under this pumping scheme, the volume of water derived from storage during the final time step was 9 percent of the total pumping, and, again, water levels were changing at a rate of less than 0.5 ft per 20 yr for the final time step, which indicates that the system was approaching equilibrium conditions.

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