REGIONAL HYDROLOGY AND SIMULATION OF DEEP GROUND-WATER FLOW IN THE SOUTHEASTERN COASTAL PLAIN AQUIFER SYSTEM IN MISSISSIPPI, ALABAMA, GEORGIA, AND SOUTH CAROLINA

REGIONAL AQUIFER-SYSTEM ANALYSIS

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Regional Hydrology and Simulation of Deep Ground-Water Flow in the Southeastern Coastal Plain Aquifer System in Mississippi, Alabama, Georgia, and South Carolina

By RENÉ A. BARKER *and* MARIBETH PERNIK

REGIONAL AQUIFER-SYSTEM ANALYSIS—SOUTHEASTERN COASTAL PLAIN

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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.

Gordon P. Eaton

Gordon P. Eaton
Director

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CONVERSION FACTORS AND ALTITUDE DATUM

Multiply inch-pound units	By	To obtain metric units
foot (ft)	0.3048	meter
foot per mile (ft/mi)	0.1894	meter per kilometer
foot squared (ft^2)	0.0929	meter squared
square foot per square mile (ft^2/mi^2)	0.0359	square meter per square kilometer
cubic foot (ft^3)	0.0283	cubic meter
cubic feet per second per square mile [$(ft^3/s)/mi^2$]	0.0109	cubic meters per second per square kilometer
gallon per minute (gal/min)	0.0631	liter per second
gallon (gal)	3.785	liter
million gallons per day (Mgal/d)	0.0438	cubic meter per second
inch (in)	0.2540	centimeter
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer

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.

REGIONAL HYDROLOGY AND SIMULATION OF DEEP GROUND-WATER FLOW IN THE SOUTHEASTERN COASTAL PLAIN AQUIFER SYSTEM IN MISSISSIPPI, ALABAMA, GEORGIA, AND SOUTH CAROLINA

By RENÉ A. BARKER and MARIBETH PERNIK

ABSTRACT

The Southeastern Coastal Plain aquifer system underlies about 120,000 square miles of the Southeastern United States, including most of the Coastal Plain of Mississippi, Alabama, Georgia, and South Carolina and small contiguous parts of Tennessee, Florida, and North Carolina. The system is composed predominantly of unconsolidated to semiconsolidated clastic rocks that dip gently seaward within a wedge of Cretaceous and Tertiary sediments that thickens from a featheredge along the inner margin of the Coastal Plain to more than 10,000 feet near the Gulf of Mexico and Atlantic Ocean. The system is underlain updip by dense, relatively impermeable rocks that form the base of the Coastal Plain. In addition to being hydraulically interconnected with carbonate rocks of the overlying Floridan aquifer system, the Southeastern Coastal Plain aquifer system merges with clastic aquifers of the Mississippi embayment and coastal lowlands aquifer systems on the west and with the Northern Atlantic Coastal Plain aquifer system on the northeast.

Ground-water conditions in the study area result from a homoclinal coastal plain setting and a beveled outcrop area of hummocky relief that is subjected to a humid climate and drained by an extensive surface-water network. Most recharge occurs in the interstream parts of the outcrop area under predominantly unconfined conditions; some recharge occurs in updip, confined parts of aquifers in areas of eastern Alabama, Georgia, and South Carolina as downward leakage from the Upper Floridan aquifer. Most discharge from the aquifer system occurs from the updip, unconfined parts of the system as base flow to streams in or near the outcrop area; smaller amounts of middip and downdip discharge occur from confined parts of the aquifer system as diffuse upward leakage to the Floridan aquifer system, or to its clastic, surficial aquifer equivalent in South Carolina. West to east increases in infiltration, recharge, and base flow result from the generally coarse texture of the outcropping sediments of predominantly fluvial origin in much of eastern Alabama, Georgia, and South Carolina, compared with the predominance of chalk, clay, and shale of marine origin in Mississippi and western Alabama.

Ground-water flow is much more dynamic in the outcrop area than in the subcrop area. Water that infiltrates the shallow, unconfined parts of the outcrop area cannot as readily penetrate the deeper, confined parts of the ground-water flow system. Consequently, a considerable amount of the updip recharge flows laterally to nearby streams; less than 1 percent of the precipitation that falls in the outcrop area ultimately reaches the buried parts of the aquifer system. Ground-water conditions in updip areas are dominated by shallow recharge that quickly discharges along short flow paths to topographically low parts of the land surface. Ground-water flow is relatively sluggish in downdip areas, where it is characterized by long, complex flow paths; most downdip flow culminates as diffuse upward leakage into the overlying Floridan aquifer system.

To obtain a more complete understanding of the deep, confined parts of the ground-water flow system, a digital-computer model was constructed that employs the technique of finite-difference approximation. The model, based on regional hydrogeologic definition, was calibrated to simulate both predevelopment (pre-1900) and transient (1900–85) flow conditions. The model uses three layers to simulate the Black Warrior River aquifer (bottom layer), the Chattahoochee River aquifer (middle), and the Pearl River aquifer (top). A fourth (source-or-sink) layer of constant-head nodes is used atop these layers to convey the effects of the Upper Floridan aquifer and its clastic, surficial aquifer equivalent in South Carolina. The Claiborne-Wilcox aquifers of the Mississippi embayment aquifer system are simulated with constant-head nodes as a western extension of the Pearl River aquifer in Mississippi and southwestern Alabama. The Chickasawhay River aquifer of Mississippi and southwestern Alabama was excluded from the simulation of the Southeastern Coastal Plain aquifer system. The regional-scale model accounts for the circulation of freshwater downgradient from small streams in the outcrop area by simulating the deep, predominantly confined ground-water regime and its interaction with major streams.

The distribution and hydraulic properties of the aquifers and confining units are controlled by lithologic and structural patterns within the sediments. Transmissivity patterns in the Black Warrior River aquifer reflect the characteristics of a sedimentary sequence that deepens from east to west and grades from mostly coarse-grained, nonmarine deposits near the inner margin of the Coastal Plain to finer grained, marginal-marine or marine sediments toward the southern limits of the modeled area. Model-derived transmissivity values for the Black Warrior River aquifer range from about 0.00001 to about 0.2 feet squared per second and average about 0.06 feet squared per second. Reflecting the effects of areal differences in both sediment thickness

and lithology, model-derived transmissivity values for the Chattahoochee River aquifer range from about 0.00001 to about 0.6 feet squared per second and average about 0.1 feet squared per second. Owing to widespread marine environments during deposition of the Pearl River aquifer, there is not as much variation in transmissivity in this aquifer as in the Chattahoochee River and Black Warrior River aquifers. Model-derived transmissivity values for the Pearl River aquifer range from about 0.001 to about 0.6 feet squared per second and average about 0.07 feet squared per second.

Model-derived storage coefficient values for the aquifer system range from 1×10^{-4} to 1×10^{-1} and average about 4×10^{-3}. The largest values of storage coefficient apply to updip parts of the Pearl River aquifer, where they reflect unconfined (water-table) or semiconfined conditions within generally coarse-grained, reasonably well sorted nonmarine sediments. The smallest values of storage coefficient are associated with the deepest, downdip parts of the Chattahoochee River and Black Warrior River aquifers, where the intermixing of sands with marine silts and clays confines the flow system.

Model-derived values of confining unit leakance range from about 5×10^{-16} to about 5×10^{-9} per second. As a consequence of clay and silt deposits that generally increase in thickness away from the inner margin of the Coastal Plain, the largest values of leakance occur updip and the smallest values occur downdip. Leakance values are substantially smaller in the Black Warrior River confining unit than in the Pearl River confining unit.

The steady-state, predevelopment (pre-1900) water budget for the Southeastern Coastal Plain aquifer system can be summarized as follows:

Precipitation	=51 inches per year
Overland flow	=12 inches per year
Evapotranspiration	=32 inches per year
Ground-water discharge from shallow flow regime to small streams	=6.4 inches per year
Ground-water discharge from deep flow regime to major streams	=0.5 inch per year
Downward leakage from updip parts of Floridan aquifer system	=less than 0.1 inch per year
Diffuse upward leakage to downdip parts of Floridan aquifer system	=less than 0.2 inch per year

Results of the regional model indicate that under predevelopment conditions, recharge to the deep ground-water flow regime totaled 1,990 cubic feet per second, or about 0.6 inch per year over an outcrop area of about 46,500 square miles. Approximately 1,720 cubic feet per second (about 0.5 inch per year) discharged as base flow to the major streams. The difference between the simulated base flow to major streams and the simulated outcrop recharge is 270 cubic feet per second (about 0.1 inch per year). This residual, called deep seepage, migrated downgradient to recharge the confined subcrop area. The deep seepage was joined by downward leakage, which is simulated to have come from updip parts of the Floridan and Mississippi embayment aquifer systems and totaled 165 cubic feet per second (less than 0.1 inch per year). An additional 5 cubic feet per second entered laterally from the Mississippi embayment aquifer system. A total of 440 cubic feet per second (less than 0.2 inch per year) is simulated to have discharged from the subcrop of the Southeastern Coastal Plain aquifer system. Of that amount, 410 cubic feet per second was diffuse upward leakage to the Floridan aquifer system, the surficial aquifer in South Carolina, and the Mississippi embayment aquifer system; the remaining 30 cubic feet per second discharged laterally to the Floridan aquifer system.

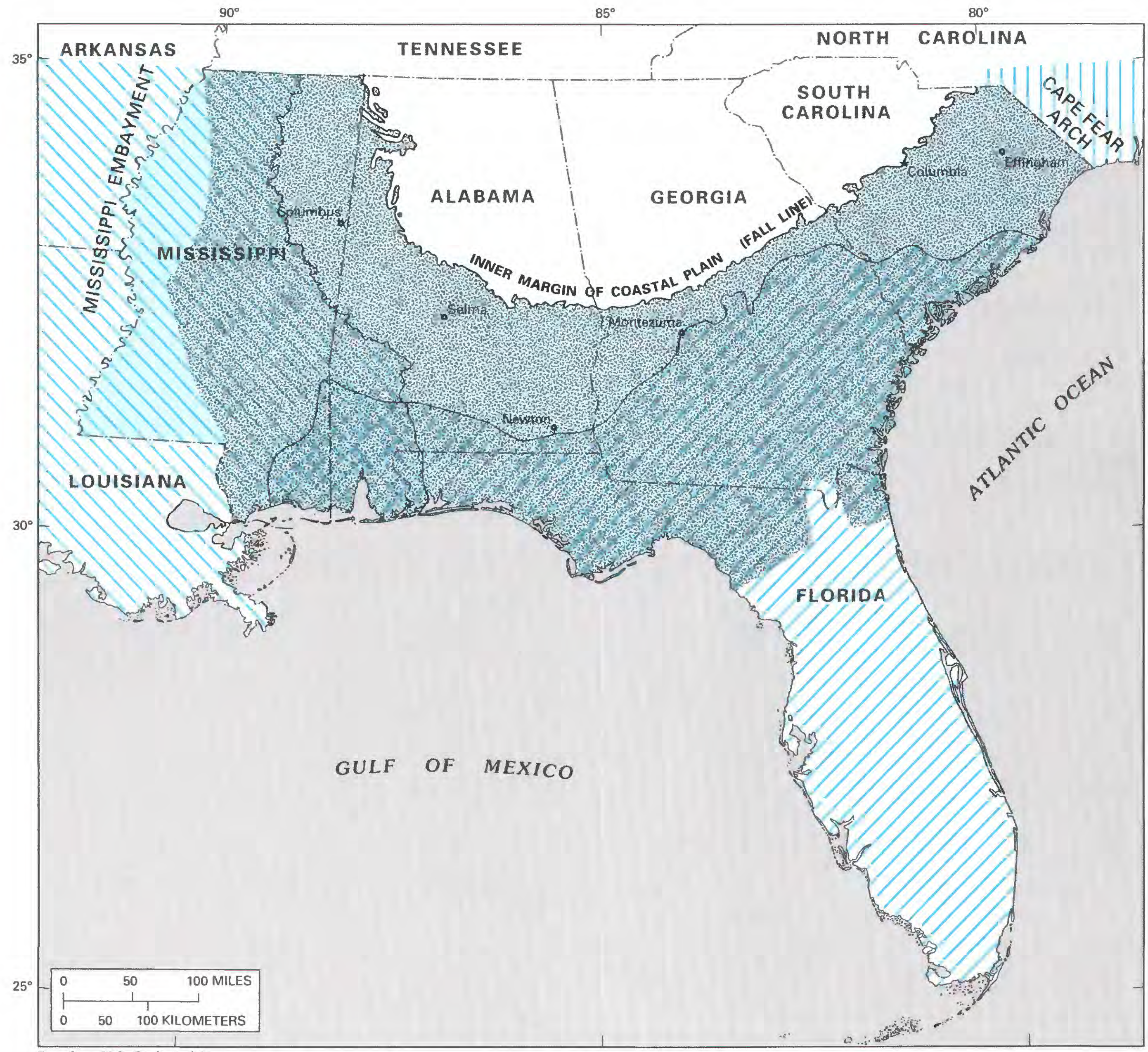
The results of simulation also indicate that the Southeastern Coastal Plain aquifer system has adjusted to ground-water development through a combination of increased inflow, decreased outflow, and a reduction in the amount of water in storage. Ground-water pumpage began about 1900 and increased to about 765 cubic feet per second by 1985, causing head declines that in places exceed 150 feet. The reduction of ground water in storage is simulated to have averaged 235 cubic feet per second during 1981–85. Base flow is simulated to have decreased from 1,720 cubic feet per second (in 1900) to 1,375 cubic feet per second (during 1981–85), representing a reduction in ground-water discharge of 345 cubic feet per second. Between 1900 and 1985, the inflow from all adjacent aquifer systems is simulated to have increased by 85 cubic feet per second, while outflow to these aquifer systems decreased by 100 cubic feet per second. The net effect of all flow adjustments with respect to adjacent aquifer systems is equivalent to a source of 185 cubic feet per second, of which 175 cubic feet per second was contributed by the highly permeable Floridan aquifer system.

INTRODUCTION

In 1978, the U.S. Geological Survey began a nationwide program to study a number of the regional aquifers that provide a significant part of the country's water supply. This program, called the Regional Aquifer-System Analysis (RASA), has been discussed in detail by Miller (1992). The general objectives of each RASA study are (1) to describe the present-day and predevelopment ground-water flow systems, (2) to analyze changes between predevelopment and present-day systems, (3) to integrate the results of previous studies of discrete aspects of the system, or of local areas within it, and (4) to provide some capability for evaluating the hydrologic effects that future ground-water development will have on the system. The Southeastern Coastal Plain aquifer system was one of the regional aquifer systems chosen for study.

AREA OF STUDY

The area of investigation is in the southeastern part of the United States. It includes most of the Coastal Plain of Mississippi, Alabama, Georgia, and South Carolina as well as small contiguous parts of Tennessee, Florida, and North Carolina. The Southeastern Coastal Plain aquifer system consists predominantly of clastic sediments of Cretaceous and Tertiary age (pl. 1) that crop out in an arcuate band that roughly parallels the inner margin of the Coastal Plain. Whereas the study area totals approximately 120,000 square miles (mi^2), the outcrop area spans about 60,000 mi^2. The aquifer system extends from the eastern flank of the Mississippi embayment in Mississippi and Tennessee to the southwestern flank of the Cape Fear Arch in North Carolina (fig. 1). The fresh ground-water flow system extends nearly to the Mississippi River, the Gulf of Mexico, and the Atlantic Ocean. The Southeastern Coastal Plain aquifer system borders



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

EXPLANATION

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

- Southeastern Coastal Plain
- Study area
- Floridan
- Gulf Coast—Mississippi embayment and coastal lowlands
- Northern Atlantic Coastal Plain

FIGURE 1.—Location of study area and areal relation among regional aquifer systems in the southeastern part of the United States (modified from Barker, 1986).

four other regional aquifer systems: the Northern Atlantic Coastal Plain aquifer system on the northeast, the Mississippi embayment and coastal lowlands aquifer systems on the west, and the Floridan aquifer system on the south.

The clastic sediments that make up the Southeastern Coastal Plain aquifer system have been separated regionally into seven major hydrogeologic units, some of which are hydraulically interconnected with the interfingering and partly overlying Floridan aquifer system. The composition, texture, and bedding character of the major units vary from place to place. Aquifers of the Southeastern Coastal Plain aquifer system are within massive to thinly bedded strata composed mostly of fine- to coarse-grained quartz and feldspathic sand, with minor amounts of limestone. Chalk, clay, shale, and mudstone form the confining units that separate the major aquifers. Locally, the major aquifers have been subdivided into smaller aquifers of subregional extent.

PURPOSE AND SCOPE

As one of several chapters of Professional Paper 1410 describing different aspects of the Southeastern Coastal Plain regional aquifer system, this report (chapter C) summarizes the regional hydrology and discusses a computer analysis of the deep ground-water flow system. Regional ground-water conditions are evaluated by integrating what is known about the system through previous investigations with the results of recent conceptualization, hydrogeologic-framework definition, and computer simulation. This report (1) summarizes the relation between the hydrology of the Southeastern Coastal Plain aquifer system and the four adjoining regional aquifer systems, (2) explores the influence of topography, geology, and climate on the ground-water flow system, (3) discusses the major elements of the water cycle and presents an overall water budget, and (4) describes a computer-model simulation of the deep, regional flow system under predevelopment and 1900–85 conditions.

APPROACH

A basic incentive for the RASA program was a need for a more complete understanding of the natural, undeveloped status of the regional aquifer systems in the Nation, and of the changes that have resulted from efforts to develop the ground-water resources. It was recognized from the outset that computer modeling would be a key to the success of the RASA program, especially for areas—such as the Southeastern Coastal Plain—for which basic data and prior knowledge of the hydrology were sparse. Under the RASA program, hydrologic analysis could be conducted over large areas and across political boundaries that would have limited the extent of previous ground-water studies. The ability of the computer model to simulate hydrologic conditions where such conditions were not well understood was largely responsible for making it possible to address the entire Southeastern Coastal Plain aquifer system during a single investigation.

Owing to the relatively large study area and the generally sparse nature of ground-water data for most of the area, the approach to improving the understanding of the aquifer system was to (1) synthesize existing hydrogeologic descriptions into a tentative conceptual model of the Southeastern Coastal Plain aquifer system, (2) construct a digital-computer model of the deepest, most extensive part of the ground-water flow system, and (3) use this model to help:

- assess the distributions of recharge, transmissivity, leakance, storage coefficient, and streambed conductivity;
- delineate the distributions of hydraulic head and ground-water flow;
- quantify the exchange of freshwater between the Southeastern Coastal Plain aquifer system and adjacent regional aquifer systems;
- evaluate the hydrologic changes that have resulted from ground-water development; and
- integrate the results of the simulation with other hydrogeologic information to refine the conceptual model and derive a general water budget for the study area.

PREVIOUS WORK AND ACKNOWLEDGMENTS

Since the turn of the century, a multitude of reports, articles, maps, and data tabulations have resulted from hydrologic, geologic, and water-quality investigations in the study area. However, most work prior to the RASA study covered only small areas and was limited by county or State boundaries, or was confined to a particular aquifer or formation of the aquifer system within a particular county, group of counties, or State. The intent of this RASA study was to integrate all pertinent data from all reliable sources into a data base for the entire study area.

Reports providing hydrogeologic interpretation that aided construction of the computer model or derivation of the water budget include those by Siple (1957), Knowles and others (1963), Boswell (1963), Callahan (1964), Swindel and others (1964), Siple (1967), and Wasson (1980). Particularly useful hydrologic reports include those by Clarke and others (1983, 1984), Aucott and Speiran (1984a, 1985a, 1985b), Williams, DeJarnette, and Planert (1986a, 1986b), Williams, Planert, and DeJarnette

(1986a, 1986b, 1986c), Aucott and others (1987), M.J. Mallory (U.S. Geological Survey, written commun., 1990), and Faye and Mayer (1990). Previously published explanations of the ground-water flow system that were at least partly derived through computer-model simulation include those by Gardner (1981) for the Gordo and Eutaw aquifers in west-central Alabama, Kernodle (1981) for the Cretaceous aquifer system in the vicinity of Lee County, Miss., and Bush (1982) and Krause (1982) for all or parts of the Tertiary limestone (Floridan) aquifer in Florida, Georgia, South Carolina, and Alabama. During the present study, preliminary computer-model analyses were made for predevelopment conditions in all or parts of the Southeastern Coastal Plain aquifer system by Barker (1986), Aucott (1988), and M.J. Mallory (U.S. Geological Survey, written commun., 1990).

An overview of the ground-water resources of the South Atlantic-Gulf Coastal region of the United States by Cederstrom and others (1979) provided background at the beginning of the RASA study. A description of how the hydrology is controlled by structural and geomorphic conditions on homoclinal flanks by LeGrand and Pettyjohn (1981) greatly enhanced an understanding of the geohydrology of the Southeastern Coastal Plain. Precipitation and runoff maps from the U.S. Geological Survey's National Water Summary (1985, 1986) provided key data for the water budget of the Southeastern Coastal Plain aquifer system. The regional potentiometric surface of the Tertiary limestone (Floridan) aquifer system presented by Johnston and others (1981) was used to establish a source-or-sink boundary condition in the computer model of the Southeastern Coastal Plain aquifer system, where the two regional aquifer systems merge.

The present study produced eight reports of regional scale that were important to the design and calibration of the computer model of the Southeastern Coastal Plain aquifer system. A description of the hydrogeologic framework by Renken (1984) was the template for defining aquifer layers and confining layers in the model. The delineation of the pre-Cretaceous rocks underlying the Southeastern Coastal Plain by Wait and Davis (1986) defined the base of the aquifer system as well as the depth of computer-model simulation. Maps of freshwater-saltwater contacts by Lee and others (1986) and Strickland and Mahon (1986) greatly aided the positioning of peripheral no-flow boundaries in the model. Base-flow analyses by Stricker (1983) helped bracket rates of recharge used to construct the water budget. Potentiometric maps by Stricker and others (1985a, 1985b, 1985c) provided virtually the only hydraulic head data with which to check the validity of the steady-state calibration.

REGIONAL HYDROGEOLOGIC SETTING

The Southeastern Coastal Plain aquifer system merges with the clastic aquifers of the Mississippi embayment and coastal lowlands aquifer systems on the west and with the Northern Atlantic Coastal Plain aquifer system on the northeast; it is partly overlain by carbonate rocks of the Floridan aquifer system. The Southeastern Coastal Plain aquifer system is everywhere underlain by low-permeability consolidated rocks (fig. 2, pl. 1). Owing to its position relative to the base of Coastal Plain sediments and adjacent aquifer systems, the Southeastern Coastal Plain aquifer system is potentially affected by conditions outside its boundaries. An overview of the regional hydrogeologic units that make up the Southeastern Coastal Plain aquifer system, as well as the likelihood of hydraulic interaction among these units and between this system and contiguous aquifer systems, is discussed below.

BASE OF THE SOUTHEASTERN COASTAL PLAIN AQUIFER SYSTEM

The Cretaceous and younger clastic rocks of the Coastal Plain are underlain by an assortment of igneous, metamorphic, and indurated sedimentary rocks of Paleozoic and early Mesozoic age that, collectively, form the base of the Coastal Plain (Renken, 1984). These generally dense, relatively impermeable rocks are in some places extensions of the Appalachian Mountains and rocks of the Piedmont physiographic province. In other places, they are graben-fill red beds and igneous intrusions that reflect an ancient rift system. The configuration of the upper surface of pre-Cretaceous rocks is the result largely of erosional forces. The surface slopes regionally toward the Atlantic Ocean and Gulf of Mexico at generally less than 75 feet per mile (ft/mi). Local anomalies result from structural features that may disrupt the distribution of overlying clastic rocks.

Geologic structures that significantly affect the configuration of the base of the Southeastern Coastal Plain are shown in figure 3. Depositional basins that received thick accumulations of sediments include the southeast Georgia embayment, the southwest Georgia embayment, and the Mississippi embayment. Coastal plain rocks thin, and some stratigraphic units are absent, over structural highs such as the Cape Fear Arch and the Peninsular Arch. The Pickens-Gilbertown fault system is responsible for pronounced downward displacement of overlying sediments.

The Mississippi embayment forms a broad, southward-plunging trough whose axis corresponds roughly to the present-day course of the Mississippi River. This depositional basin contains marginal-marine to marine depos-

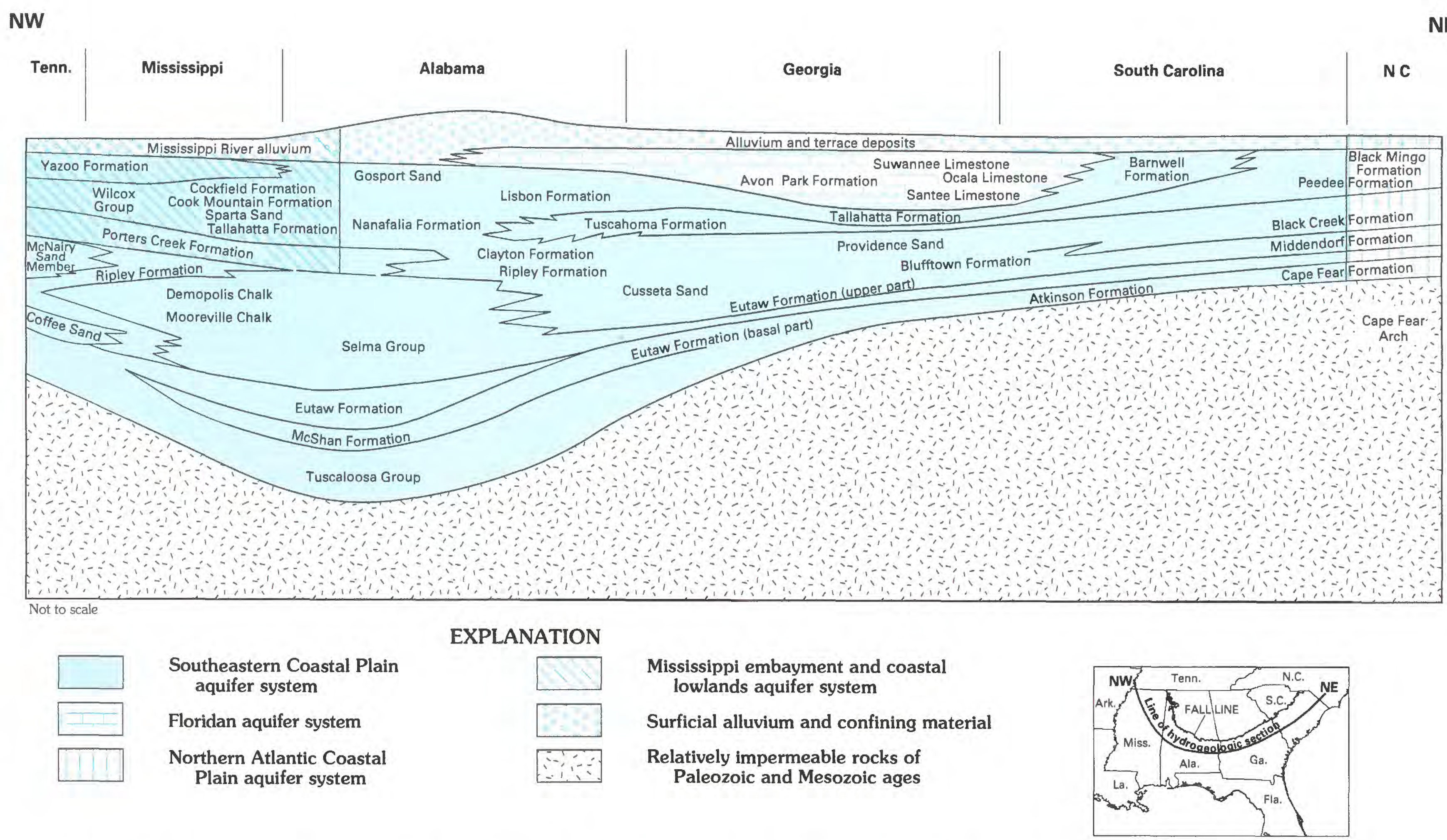


FIGURE 2.—Vertical relation among regional aquifer systems in the southeastern part of the United States with respect to selected local rock-stratigraphic units.

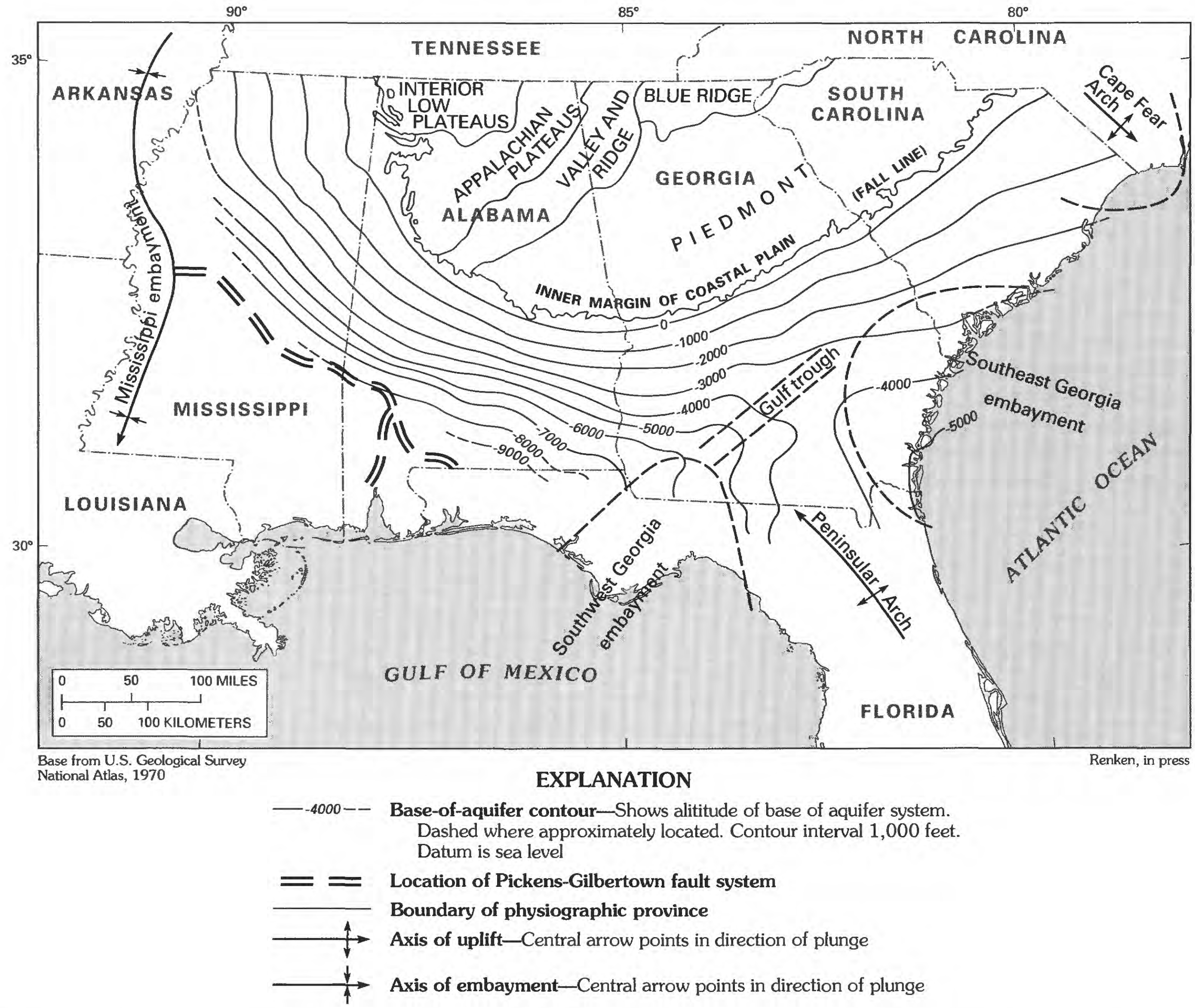


FIGURE 3.—Physiographic provinces and structural features on the base of the Southeastern Coastal Plain aquifer system (modified from Renken, in press)

its of Jurassic to Holocene age. Strata of the Mississippi embayment have a synclinal symmetry about the river, in contrast to the coastward dip of the Coastal Plain sediments in Alabama, Georgia, and South Carolina. Because surface-water drainage and the direction of ground-water flow in the embayment area are controlled largely by structural aspects of the embayment, the hydrology within the embayment is generally distinct from that of adjoining areas.

The southeast and southwest Georgia embayments appear as depressions in the base of the Coastal Plain and are separated by the Peninsular Arch of northern Florida and south-central Georgia. Cretaceous rocks are relatively thin atop the Peninsular Arch, suggesting that it was an area of relatively high relief during Cretaceous deposition in the adjacent embayments. Tertiary sediments have a relatively uniform thickness in both embayments and over the intervening arch; thus, these structures do not appear to significantly affect the hydrology of the Tertiary sediments.

The Cape Fear Arch forms a wide hump in the base of the Coastal Plain and plunges toward the southeast, roughly paralleling the Cape Fear River in North Carolina. Tectonic uplift associated with this feature has caused a significant reduction in the accumulation of sediments above it. Many rock units are missing atop the arch, in contrast to thick accumulations of Cretaceous and Tertiary rocks to the northeast and southwest, suggesting that the arch was well above sea level during much of its post-Jurassic history. These differences have made it logical to use the arch to separate the Southeastern Coastal Plain aquifer system from the Northern Atlantic Coastal Plain aquifer system.

The Pickens-Gilbertown fault system (fig. 3) is a 300-mile (mi)-long network of numerous, apparently complex, grabens extending through central Mississippi and southwestern Alabama. Owing to its depth and the associated scarcity of data, the influence of this structure on the regional geohydrology is largely speculative. However, it is reasonable to assume that any displacement of permeable strata due to the faulting would likely impede the downdip migration of ground water. The downdip limit of freshwater, herein defined as water containing less than 10,000 milligrams per liter (mg/L) dissolved solids, in Cretaceous clastic sediments occurs just updip from the fault system. Consequently, the present position of the freshwater-saltwater interface within the Cretaceous rocks of Mississippi and Alabama may result from a buildup of dissolved solids—with limited opportunity for flushing—due in part to a structurally controlled stagnation of ground water near the Pickens-Gilbertown fault system.

In addition to illustrating the configuration of the pre-Cretaceous rocks that underlie the Southeastern Coastal Plain, Wait and Davis (1986) integrated pertinent data from more than 50 publications about the geology and hydrology of these rocks. They concluded that

> The permeability of these rocks is extremely low and little exchange of water occurs upward to the overlying Cretaceous aquifers. Porosity values are in the range of 10 percent or less in some of the rocks, indicating little water is stored in them. The transmissivity values are extremely low, indicating very slow movement of water in these rocks. The small volume of water moving at a very slow rate has a long residence time and is usually mineralized to a greater degree than water in the more permeable overlying sediments.

It appears that the hydrology of the pre-Cretaceous rocks has little influence on conditions within the overlying clastic regional aquifer system. Accordingly, it is assumed that the base of the Coastal Plain functions as the lower hydraulic boundary of the Southeastern Coastal Plain aquifer system.

SOUTHEASTERN COASTAL PLAIN AQUIFER SYSTEM

The Southeastern Coastal Plain aquifer system is made up of unconsolidated to semiconsolidated sedimentary rocks of Cretaceous and Tertiary age. The predominantly clastic strata of this system crop out in a series of irregular, arcuate bands that bend around the southernmost exposures of the Piedmont, Valley and Ridge, Appalachian Plateaus, and Interior Low Plateaus physiographic provinces (fig. 3). The aquifer system in many places directly overlies the base of the Coastal Plain, which consists largely of subsurface extensions of the generally dense, relatively impermeable rocks underlying these provinces. The trace of the surface contact between the clastic strata and the underlying crystalline or consolidated sedimentary rocks is called the inner margin of the Coastal Plain, or the Fall Line. The Southeastern Coastal Plain aquifer system thickens seaward, away from the Fall Line, forming a wedge of strata that dip (tracking west to east) toward the Mississippi River and the coastlines of the Gulf of Mexico and the Atlantic Ocean. The aquifer system thickens from east to west; coastline accumulations range from less than 1,500 feet (ft) near the South Carolina–North Carolina State line to more than 10,000 ft in Alabama and Mississippi. Regional dips range from generally less than 20 ft/mi in South Carolina to more than 75 ft/mi in downdip parts of western Alabama and eastern Mississippi.

The Coastal Plain sediments reflect the effects of a complex history of shoreline fluctuations resulting from tectonic and erosional forces during Jurassic through Holocene time. The rocks deposited and preserved in any given locality depend on the relative elevations of the landmasses, stream channels, and sea levels during this time. Depositional conditions varied among marginal marine, shallow-shelf marine, shoreline, and nonmarine. Although fluvial-deltaic sands and (or) gravels generally were deposited near the Fall Line and grade downdip into marine silts and marls, structural and erosional variations in the source areas, as well as differential subsidence and erosion in the depositional areas, created local depositional environments that varied considerably along both strike and dip. The lithology of a single depositional unit can vary greatly over a few tens of miles, reflecting, for example, fluvial processes in one area, deltaic activity in another, and shallow marine conditions nearby. Although successively older rocks crop out toward the Fall Line in Mississippi, Alabama, and western Georgia, erosional remnants of Lower Cretaceous nonmarine deposits are overlapped near the Fall Line by Upper Cretaceous and Tertiary marine rocks in eastern Georgia and western South Carolina.

The occurrence, movement, and availability of ground water in the Southeastern Coastal Plain aquifer system are controlled greatly by the depositional, tectonic, and diagenetic history of the rocks that compose it. Given the variability in the geology, it is understandable that the hydrology, too, is highly variable. The distributions of recharge, discharge, hydraulic head, transmissivity, confining unit leakance, and storage coefficient are dependent on the distribution of lithology, which is in turn dependent on sediment texture, degree of sorting, and type of bedding.

An understanding of the three-dimensional distribution of rock type, geologic structure, and hydrologic characteristics of the rocks was prerequisite to developing a computer model for simulating the regional groundwater flow system. The hydrogeologic framework provided by Renken (1984) was the key to differentiating between regional aquifers and confining units for the purposes of computer modeling. The regional aquifers include deposits of gravel, sand, and minor amounts of limestone; the regional confining units are composed mostly of clay, chalk, mudstone, and shale. From top to bottom, the four regional aquifers are designated the (1) Chickasawhay River, (2) Pearl River, (3) Chattahoochee River, and (4) Black Warrior River aquifers (Miller and Renken, 1988). The relation between these regional aquifer designations and the stratigraphic nomenclature is shown in figure 4. The outcrop pattern of the four regional aquifers is shown in figure 5.

The Chickasawhay River aquifer occurs in southwest Alabama and southern Mississippi. Consisting of clastic and carbonate rocks of Oligocene and Miocene age, this unit extends toward the southwest into Louisiana and is equivalent to the coastal lowlands aquifer system in southwestern Alabama and southern Mississippi. As a result of a facies change to limestone, the Chickasawhay River aquifer merges with the upper part of the Floridan aquifer system in eastern Alabama and northern Florida. The Chickasawhay River aquifer is underlain by the Pearl River confining unit, which consists primarily of clays of the Yazoo Formation of Eocene age (pl. 1).

The Pearl River aquifer (fig. 4) is a relatively thick rock sequence containing mostly sand but also minor amounts of sandstone, gravel, and transitional carbonate rocks. Most of this aquifer is in rocks that range from late Eocene to Paleocene in age. Locally, in South Carolina, it is composed of the upper part of the Peedee Formation (fig. 2) of Late Cretaceous age. Rocks of the Pearl River aquifer were deposited for the most part under marine conditions, except in parts of Mississippi where they originated under conditions ranging from deltaic to fluvial. Rocks of the Pearl River aquifer grade toward the west and southwest into rocks that contain the Mississippi embayment aquifer system. In southern Georgia and parts of southern Alabama and southwest South Carolina, the rocks of the Pearl River aquifer undergo a facies change into carbonate rocks of the Floridan aquifer system. The division in these areas between the Southeastern Coastal Plain and Floridan aquifer systems is based on a lithologic gradation; it does not represent a clearcut break in either geologic or hydrologic conditions. Generally there is hydraulic connection across the rather arbitrarily drawn boundary that represents both the base of the Floridan aquifer system and the top of the Southeastern Coastal Plain aquifer system.

The Pearl River aquifer is separated from the underlying Chattahoochee River aquifer (fig. 4) in eastern Alabama and over most of Georgia and South Carolina by the Chattahoochee River confining unit, which is composed predominantly of Paleocene clay. In middip and downdip areas of Georgia and South Carolina, the Chattahoochee River confining unit effectively limits groundwater flow between these two aquifers. However, the confining unit is locally absent in updip areas of Georgia and South Carolina where the Chattahoochee and Pearl River aquifers are in direct physical as well as hydraulic contact. From eastern Alabama, the Chattahoochee confining unit extends nearly to the Alabama River in west-central Alabama where, through a facies change, clay of the Tuscahoma Formation (fig. 2) is replaced by sand of the Nanafalia Formation (which composes the lower part of the Pearl River aquifer). Between the Alabama River and northern Mississippi, the Pearl River aquifer is separated from the Black Warrior River aquifer, the basal aquifer in the system, by thick accumulations of clay and chalk, collectively called the Black Warrior River confining unit. The Chattahoochee River confining unit is largely equivalent to the Tuscahoma Formation in Georgia and eastern Alabama, and to the lower part of the Black Mingo Formation and middle part of the Peedee Formation in South Carolina (fig. 2). The Black Warrior River confining unit consists mostly of chalk of the Selma Group but also includes clays of the Porters Creek Formation, the middle part of the Eutaw Formation, and the upper part of the Cape Fear Formation.

The Chattahoochee River aquifer (fig. 4) is continuous in the subsurface from western Alabama into North Carolina, where it merges with aquifers that are part of the Northern Atlantic Coastal Plain aquifer system. In outcrop, through northeastern Georgia and southwestern South Carolina, the Chattahoochee River aquifer is overlapped in places by the Pearl River aquifer. Sediments of the Chattahoochee River aquifer were deposited mostly in deltaic to shallow marine environments; however, fluvial deposits make up part of the unit in South Carolina. Sandy, glauconitic limestone of the Clayton Formation of Paleocene age makes up part of the aquifer in western Georgia and eastern Alabama. From central Alabama, the Chattahoochee River aquifer grades westward by facies change into calcareous shale and chalk that are part of the Selma Group (fig. 2) and

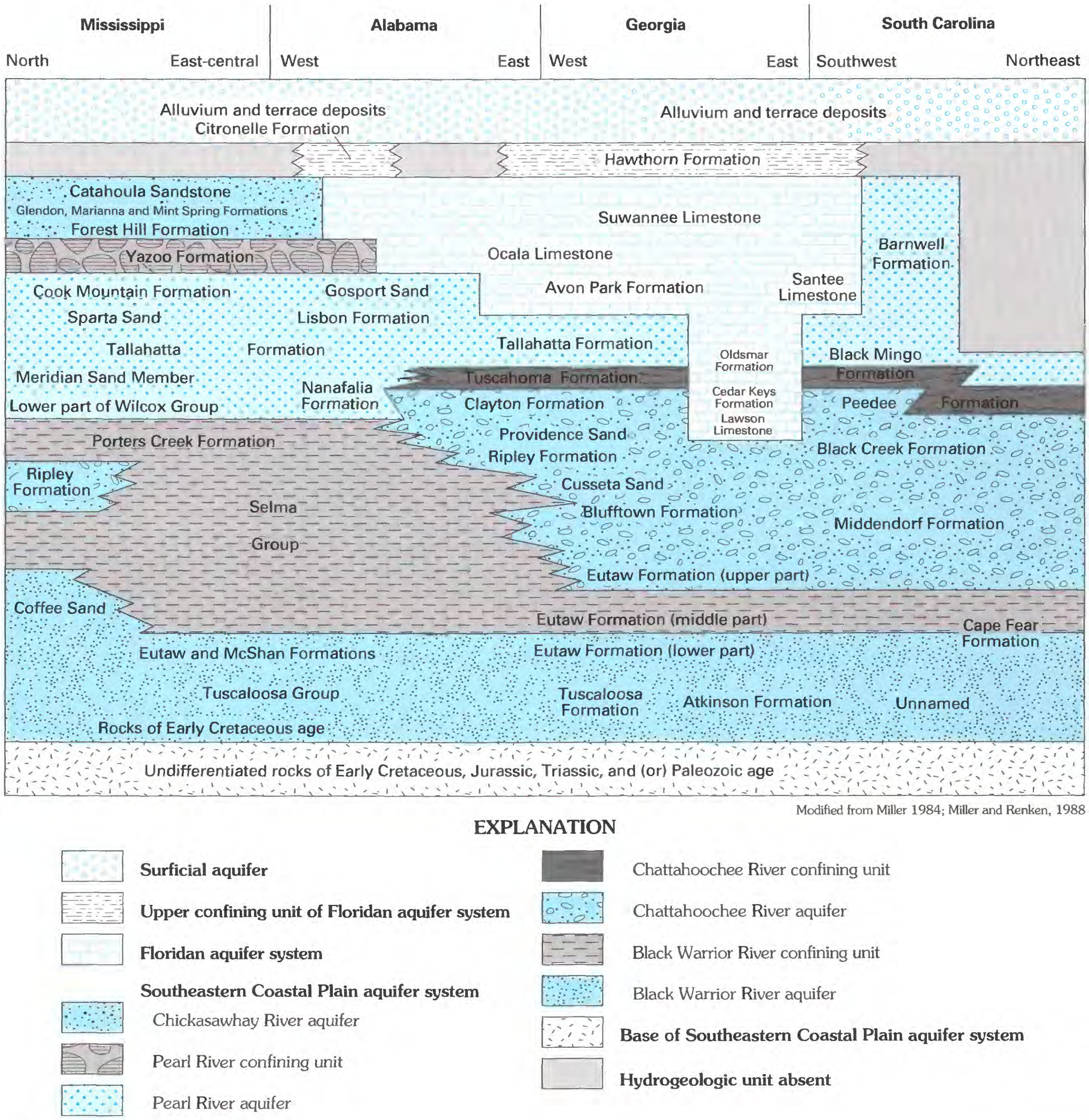


FIGURE 4.—Relation among regional hydrogeologic units and selected rock-stratigraphic nomenclature (modified from Miller and Renken, 1988).

are included in the Black Warrior River confining unit. Because of their stratigraphic position, permeable terrigenous sediments of the Ripley Formation in northern Mississippi (the Ripley aquifer) are included as a western appendage of the Chattahoochee River aquifer, even though the water-bearing parts of the Ripley are not physically or hydraulically connected to the main body of the aquifer in eastern Alabama, Georgia, and South Carolina. The Chattahoochee River aquifer is separated from the underlying Black Warrior River aquifer by clay and chalk of the intervening Black Warrior River confining unit.

FIGURE 5.—Outcrop or shallow subcrop areas of regional aquifers and confining units (modified from Miller and Renken, 1988).

The Black Warrior River aquifer (fig. 4) is the basal and most extensive aquifer in the Southeastern Coastal Plain aquifer system. This aquifer extends in the subsurface from the Mississippi-Tennessee border eastward into North Carolina, where it merges with rocks of the Northern Atlantic Coastal Plain aquifer system. However, the unit crops out only in Mississippi, Alabama, and a small part of western Georgia. In Mississippi and western Alabama, the Black Warrior River aquifer consists mostly of Upper Cretaceous sands of fluvial and deltaic origin that belong to the Coffee Sand, the Eutaw and McShan Formations, and the Tuscaloosa Group; underlying, hydraulically connected Lower Cretaceous sands are also included in this aquifer in the western part of the study area. Marine sand and sandy clay make up part of the aquifer in southern Alabama and southwestern Georgia. Toward the east, in the subsurface of east-central Georgia and South Carolina, the Black Warrior River aquifer is composed of the relatively thin, mostly nonmarine, Atkinson and Cape Fear Formations (fig. 2).

The Black Warrior River aquifer is underlain by relatively dense rocks of Paleozoic to Jurassic age. The small amount of water in these rocks is thought to have

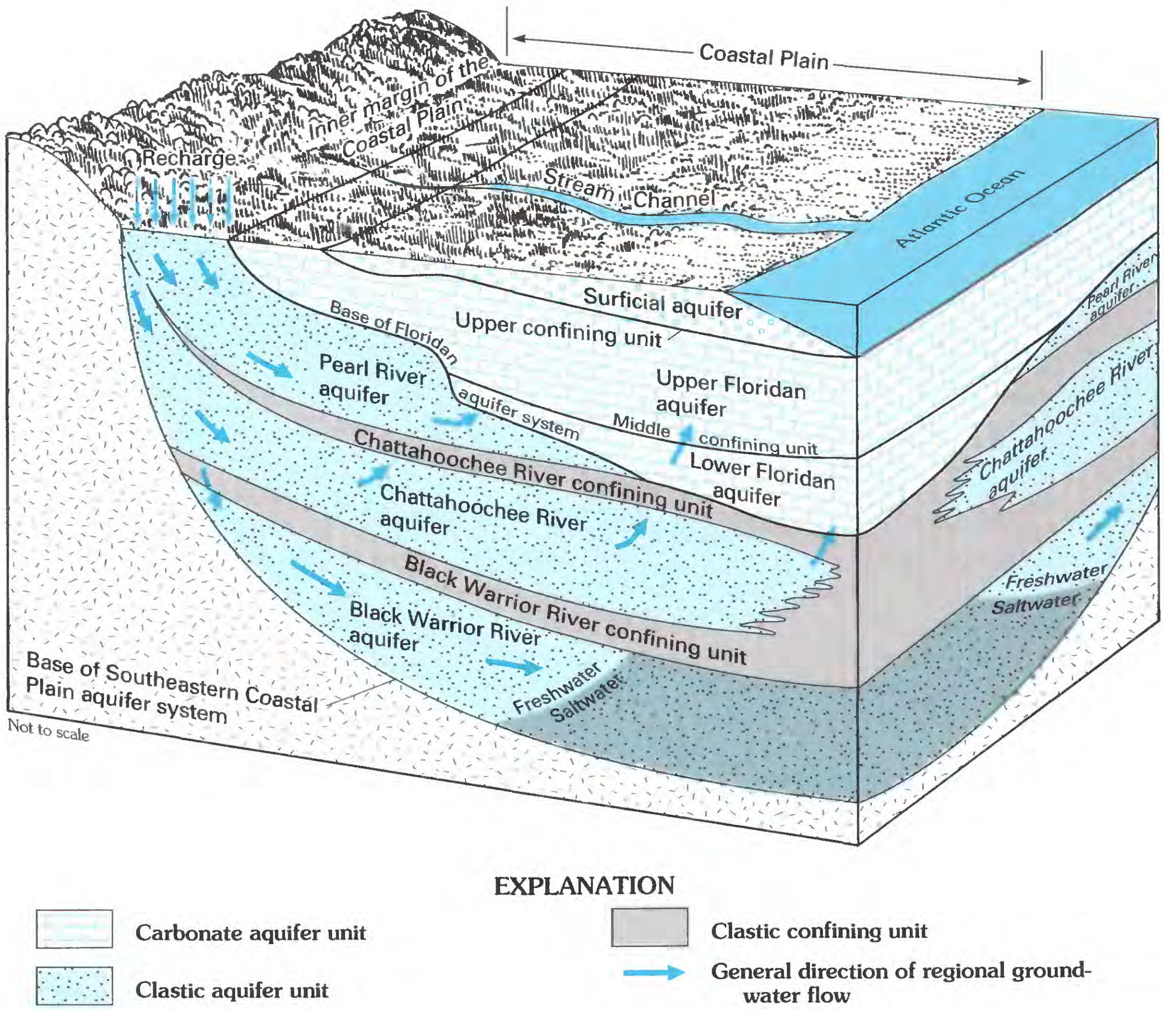


FIGURE 6.—Hydrogeologic relation between the Floridan aquifer system and the Southeastern Coastal Plain aquifer system in east-central Georgia (modified from Barker, 1986).

a dissolved-solids concentration greater than 10,000 mg/L everywhere except, perhaps, in the outcrop area or just downdip from the Fall Line. Because ground-water circulation within these rocks is extremely sluggish, and presumably does not significantly affect the freshwater flow system within the overlying clastic sediments, the upper surface of this group of relatively impermeable rocks is delineated as the base of the Southeastern Coastal Plain aquifer system (Wait and Davis, 1986).

FLORIDAN AQUIFER SYSTEM

The clastic rocks of the Southeastern Coastal Plain regional aquifer system grade laterally and vertically in western South Carolina, south Georgia, and southeastern Alabama into carbonate rocks of the Floridan aquifer system (figs. 2, 4, 6). Occupying a total area of about 100,000 mi^2, the highly productive Floridan aquifer system overlies about 40,000 mi^2 of the Southeastern Coastal Plain aquifer system. The Floridan is overlain in much of its area by generally low-permeability clastic rocks of predominantly Miocene age that, taken together, constitute an upper confining unit (Miller, 1986). This confining unit is in turn overlain by deposits of Pliocene to Holocene age that are mostly sand but include gravel, sandy limestone, and limestone in places and are referred to as the "surficial aquifer." This shallow aquifer is for the most part unconfined and in many places is important as a source of recharge (through downward leakage) to the underlying Floridan aquifer system. Although predominantly limestone, the Floridan aquifer system in places is composed largely of dolomite and contains minor amounts of clay, sand, gypsum, anhydrite, and marl. The system includes rocks of late Paleocene to early Miocene age that generally contain highly permeable zones either consisting of very porous coquina or grainstone or riddled with solution cavities or fractures (Bush and Johnston, 1988, p. C7).

The high degree of permeability common in much of the Floridan in the lateral direction is less common in the vertical direction. According to Miller (1986, p. B40), the degree of vertical hydraulic connection depends largely on the texture and mineralogy of the rocks that make up the system. In and near the outcrop area, the system is composed of one vertically continuous permeable unit. Farther downdip, less permeable rocks of subregional extent were used by Miller (1986, p. B45) to separate the system into an Upper and a Lower Floridan aquifer. The less permeable rocks that separate these aquifers are collectively referred to as the "middle confining unit." The middle confining unit is present over about two-thirds of the area underlain by the Floridan. It varies from being very leaky to virtually nonleaky, depending on the local lithology of the unit. In areas where there is little hydraulic connection across the middle confining unit, significant differences can exist between the Upper and Lower Floridan aquifers with respect to their hydraulic properties, water chemistry, and rates of ground-water flow.

Miller (1982) defined the base of the Floridan aquifer system as coinciding with the top of rock whose permeability is significantly lower than that of the limestone above. He emphasized that his delineation generally does not correspond to formation or time-stratigraphic tops, but rather to a permeability contrast within rocks whose age and lithology may vary considerably. Miller (1986, p. B40) described the base as consisting of either low-permeability clastic rocks or evaporite deposits that everywhere separate the high-permeability carbonate rocks of Tertiary age from the deeper clastic rocks of predominantly Cretaceous age. Accordingly, Renken (1984) adopted Miller's base of the Floridan aquifer system as the top of the underlying Southeastern Coastal Plain aquifer system. Consequently, the division between the Floridan aquifer system and the Southeastern Coastal Plain aquifer system is based on a facies transition between clastic and carbonate rocks.

The Floridan aquifer system differs hydraulically from the underlying Southeastern Coastal Plain aquifer system. Based on simulation, Bush and Johnston (1986, p. 20) estimated that recharge to the Floridan averages about 4.4 inches per year (in/yr); recharge to the deep flow regime of the clastic rocks of the Southeastern Coastal Plain system averages less than 1 in/yr. While transmissivity in the Southeastern Coastal Plain aquifer system averages less than 10,000 feet squared per day (ft^2/d), transmissivity in the Floridan averages more than 250,000 ft^2/d and in places exceeds 1,000,000 ft^2/d. The hydraulic contrasts between the Floridan and the Southeastern Coastal Plain aquifer systems determine the relative importance of the ground-water flow between the two aquifer systems.

In updip areas, ground water generally leaks from the Floridan (carbonate) aquifer system to the underlying Southeastern Coastal Plain system; in downdip areas, where the head gradient reverses, flow is generally from the Southeastern Coastal Plain system to the overlying Floridan. Because the mechanics of recharge and discharge are considerably more dynamic and the transmissivity values are significantly greater in the Floridan aquifer system, the exchange of water between the two systems is not important to the regional hydrology of the Floridan. However, this exchange substantially affects the water budget of the Southeastern Coastal Plain aquifer system because of its relatively sluggish flow regime, much lower transmissivity values, and comparatively smaller recharge rates. (See the discussion of water budget in the section "Model Results: Simulation of Deep Ground-Water Flow.")

NORTHERN ATLANTIC COASTAL PLAIN AQUIFER SYSTEM

Sediments of the Southeastern Coastal Plain regional aquifer system merge in the area of the Cape Fear Arch of North Carolina with those of the Northern Atlantic Coastal Plain aquifer system (fig. 2). The Northern Atlantic Coastal Plain aquifer system (Meisler, 1980a) exists within a seaward-thickening wedge of predominantly unconsolidated Jurassic to Holocene rocks that underlies about 50,000 mi^2 of the Atlantic Coastal Plain from the South Carolina–North Carolina State line to Long Island, N.Y. From a featheredge at the Fall Line, this sedimentary sequence thickens to a maximum thickness of about 10,000 ft near Cape Hatteras, N.C. Atop the Cape Fear Arch, the sedimentary section is less than 1,500 ft thick. The Northern Atlantic Coastal Plain aquifer system extends downdip from the Fall Line to a transition zone between freshwater and saltwater near the coastline. The freshwater-saltwater transition zone, defined as the area where the chloride concentration equals 10,000 mg/L, is assumed to represent a downdip no-flow boundary (Meisler and others, 1986). From primarily fluvial to fluvio-deltaic (nonmarine) sediments at the base of the aquifer system, the Cretaceous section grades upward through generally marginal-marine and marine deposits into Tertiary deposits of predominantly marine origin (Meisler and others, 1986). The uppermost, Pleistocene and Holocene part of the section consists mainly of marine, terrace, alluvial, and dune deposits—in addition to glacial drift on Long Island.

The Northern Atlantic Coastal Plain sediments were divided for simulation purposes into 10 regional aquifers, which consist principally of sand, gravel, or limestone, and 9 intervening confining units, which are composed predominantly of clay and silt. According to Meisler and others (1986), the regional aquifers coincide with local aquifers in some areas, comprise several local aquifers in others and constitute only part of an aquifer in still other areas. No single regional aquifer exists everywhere in the system. A single regional aquifer may include rocks of different ages in different places. The relation between the regional aquifer nomenclature used by the Northern Atlantic Coastal Plain RASA and that used by the Southeastern Coastal Plain RASA is shown in table 1.

Ground-water flow between the Northern Atlantic Coastal Plain and Southeastern Coastal Plain aquifer systems is minimal. These clastic aquifer systems merge over the Cape Fear Arch, where permeable rocks are considerably thinner than they are toward the northeast and southwest. A ground-water divide occurs along this broad arch, with an orientation similar to that of the axis of the structure. Because the arch plunges from the northwest toward the southeast, the preferential direction of ground-water flow is perpendicular to the southwest-to-northeast strike of most rocks in the Atlantic Coastal Plain; therefore, the potential for ground-water flow between the two regional aquifer systems is limited. Meisler and others (1986) noted the possibility of flow from South Carolina toward North Carolina in the deeper parts of the aquifer system. However, the amount of flow is negligible because regional hydraulic gradients are very small and transmissivity in the lower two (Chattahoochee River and Black Warrior River) aquifers of the Southeastern Coastal Plain aquifer system in South Carolina is much smaller than the average transmissivity of aquifers in either regional aquifer system. The simulated rates of ground-water flux between the Northern Atlantic and Southeastern Coastal Plain regional aquifer systems are discussed in relation to the water budget in the section "Model Results: Simulation of Deep Ground-Water Flow."

GULF COAST AQUIFER SYSTEMS

The Gulf Coast RASA study (Grubb, 1987) identified three regional aquifer systems: the Mississippi embayment aquifer system, the Texas coastal uplands aquifer system, and the coastal lowlands aquifer system. These aquifer systems occupy about 230,000 mi^2 of the Gulf Coastal Plain in Alabama, Arkansas, Florida, Illinois, Kentucky, Louisiana, Mississippi, Missouri, Tennessee, and Texas. These systems dip toward the Gulf of Mexico and thicken from less than a hundred feet near their updip limits to thousands of feet near their downdip limits. The Mississippi embayment and Texas coastal uplands aquifer systems consist principally of Eocene sediments; the coastal lowlands aquifer systems consists mainly of Miocene and younger deposits.

Sediments of the Mississippi embayment and coastal lowlands aquifer systems merge with those of the Southeastern Coastal Plain aquifer system (fig. 2). The relation between the regional aquifer nomenclature used by the Gulf Coast RASA and that used by the Southeastern Coastal Plain RASA is shown in table 1. The coastal lowlands aquifer system consists of upper Oligocene through Holocene deposits above the uppermost massive clay of the Vicksburg Group or the Jackson and Vicksburg Groups where they are undifferentiated. The eastern limit of this system has been placed at the Alabama and Escambia Rivers in Alabama and through the panhandle of northwestern Florida. The coastal lowlands aquifer system overlies extreme downdip parts of the Pearl River aquifer of the Southeastern Coastal Plain aquifer system in southeastern Mississippi and southwestern Alabama, where the bottom part of the coastal lowlands aquifer system is equivalent to the Chickasawhay River aquifer. Collectively, the Claiborne-Wilcox aquifers of the Mississippi embayment aquifer system are equivalent to the Pearl River regional aquifer of the Southeastern Coastal Plain aquifer system. The water-bearing sands of the McNairy Sand member of the Ripley Formation in northern Mississippi and the time-equivalent McNairy Sand in southwestern Tennessee make up the western appendage of the Chattahoochee River aquifer of the Southeastern Coastal Plain aquifer system (pl. 1; figs. 2, 4). Water-bearing sands of the Ripley Formation south of the Mississippi-Tennessee State line constitute the Ripley aquifer of northern Mississippi. The water-bearing sands of the McNairy Sand in Missouri, Tennessee, and Kentucky and the equivalent Cretaceous Nacatoch Formation in Arkansas (Brahana and Mesko, 1988) are herein referred to as the "McNairy-Nacatoch aquifer" (table 1).

Results of computer models developed by Brahana and Mesko (1988) and M.J. Mallory (1993) indicate that the Mississippi embayment aquifer system is hydraulically isolated from underlying parts of the Southeastern Coastal Plain aquifer system by the intervening clays and shales of the Midway and Chattahoochee River confining units (table 1). Therefore, the potential for significant ground-water exchange between the Gulf Coast regional aquifers and the Southeastern Coastal Plain aquifer system is limited to horizontal flow within the aquifers in Tertiary sediments above the Midway and Chattahoochee River confining units. Ground-water flow near the easternmost edges of the Mississippi embayment and coastal lowlands aquifer systems is dominated by steep hydraulic gradients toward the Tombigbee, Alabama, and Escambia Rivers; this is especially true of the unconfined flow, which is directly breached by these

TABLE 1.—*Relation among hydrogeologic units designated by the Southeastern Coastal Plain, Gulf Coast, Floridan, and Northern Atlantic Coastal Plain Regional Aquifer-System Analysis (RASA) studies*

[Modified from Miller and Renken, 1988. Sources: a, Grubb (1987), Hosman and Weiss (1991), and Weiss (1992); b, Miller (1986); c, Henry Trapp, Jr., U.S. Geological Survey, written commun.); d, confining units between aquifers not shown; e, stratigraphic position varies. RASA, Regional Aquifer-System Analysis]

SOUTHEASTERN COASTAL PLAIN RASA		GULF COAST RASA [a]			FLORIDAN RASA [b]		NORTHERN ATLANTIC COASTAL PLAIN RASA [c]	
		Coastal lowlands aquifer system [d]	Permeable zone A	(Holocene-upper Pleistocene deposits)		Surficial aquifer	Northern Atlantic Coastal Plain aquifer system [d]	Surficial aquifer
			Permeable zone B	(lower Pleistocene–upper Pliocene despoits)				Upper Chesapeake aquifer
Southeastern Coastal Plain aquifer system	Chickasawhay River aquifer		Permeable zone C	(lower Pliocene–upper Miocene deposits)		Upper confining unit [e]		Lower Chesapeake aquifer
			Permeable zone D	(middle Miocene deposits)				
			Permeable zone E	(lower Miocene—upper Oligocene deposits)	Floridan aquifer system	Upper Floridan aquifer [e]		Castle Hayne–Piney Point aquifer
	Pearl River confining unit	Vicksburg–Jackson confining unit						
	Pearl River aquifer	Mississippi embayment aquifer system [d]	Upper Claiborne aquifer					
			Middle Claiborne aquifer			Middle confining unit [e]		
			Lower Claiborne–upper Wilcox aquifer			Lower Floridan aquifer [e]		Beaufort-Aquia aquifer
			Middle Wilcox aquifer					
			Lower Wilcox aquifer					Brightseat–upper Potomac aquifer (upper part) [e]
	Chattahoochee River confining unit	Midway confining unit				[e]		
	Chattahoochee River aquifer	McNairy-Nacatoch aquifer				Lower confining unit		Peedee–Severn aquifer
		Not studied (permeable rocks mostly contain saline water)						Black Creek–Matawan aquifer
								Magothy aquifer
	Black Warrior River confining unit							Brightseat-upper Potomac aquifer (lower part) [e]
	Black Warrior River aquifer							Middle Potomac aquifer
								Lower Potomac aquifer

regional drains. Owing to the nearly impermeable Midway confining unit and the proximity of the RASA boundaries to major rivers, the amount of ground-water flow between the Gulf Coast regional aquifers and the Southeastern Coastal Plain aquifer system is inconsequential to the overall water budget of either. (See the discussion of water budget in the section "Model Results: Simulation of Deep Ground-Water Flow.")

HYDROLOGIC ENVIRONMENT

As background to the development of the computer model, the hydrology of the study area is summarized below. The hydrologic cycle, ground-water flow system, and water budget are discussed so that the reader might better understand the design, calibration, uses, and limitations of the model.

HYDROLOGIC CYCLE

Natural conditions and events that control the occurrence, availability, and development of water resources in the Southeastern Coastal Plain are linked through the endless succession of hydrologic phenomena known as the hydrologic cycle. The hydrologic cycle incorporates numerous processes. Those most important to the Southeastern Coastal Plain aquifer system are (1) precipitation (including rain, snow, sleet, and hail), (2) runoff (both overland flow and base flow), (3) evapotranspiration (evaporation and transpiration), and (4) infiltration of ground water (to the saturated part of the aquifer system). These processes are discussed below.

PRECIPITATION

The primary source of freshwater in the Southeastern Coastal Plain aquifer system is precipitation that falls on the outcrop area, predominantly as rain but also as minor amounts of sleet, snow, and hail. Most precipitation is evaporated from the land surface, is transpired by vegetation, or moves directly to nearby streams as overland flow. Depending on the amount, duration, and intensity of the precipitation, as well as on the nature of the terrain, soil, and hydraulic gradient, 0 to 75 percent of the precipitation infiltrates the land surface; some of the infiltrated water may eventually recharge the ground-water system. Although most of the water in the shallow, unconfined parts of the Coastal Plain sediments fell from the atmosphere during the last hundred or thousand years, most of the confined water probably fell between 10,000 and 100,000 years ago. The age, or residence time, of the confined water is much greater than that of the unconfined water because the flow paths between the recharge and discharge areas are much longer and deeper, and the hydraulic gradients are smaller, in the confined part of the system. Most ground water satisfying domestic, municipal, industrial, and agricultural needs comes from relatively shallow sources, so its availability is substantially affected by the temporal and spatial variability of precipitation.

The historical distribution of precipitation in the study area is shown in figure 7. The contours in this figure depict the average annual precipitation across Mississippi, Alabama, Georgia, and South Carolina during 1951–80. Also shown are bar graphs of annual precipitation and 7-year moving average curves (insets) for the period of record at four meteorological stations in the outcrop area of the Southeastern Coastal Plain aquifer system. The 1951–80 data appear to be reasonably compatible with the long-term (period-of-record) data, as the differences between the long-term average annual precipitation and the 1951–80 average annual precipitation are less than 2 in/yr (5 percent) at three of the four stations represented in figure 7.

Precipitation generally increases from east to west and from north to south; it decreases with distance from the Gulf of Mexico and from the Atlantic Ocean up to the Fall Line, where this pattern reverses owing to orographic effects over the higher elevations of the Piedmont and mountains. Measured as a combination of rain, snow, and ice, precipitation in the study area ranges from about 48 to about 68 in/yr and averages about 55 in/yr. In the outcrop areas of the four regional aquifers making up the Southeastern Coastal Plain aquifer system, precipitation averages about 51 in/yr.

Departures from the long-term average rates of precipitation have affected the hydrologic cycle within the study area. The 7-year moving average graphs show that since the late 1800's, the Southeastern Coastal Plain has cycled through three or four extended periods when precipitation was much greater than average and a similar number of intervals when precipitation was much less than average. The extremely dry conditions occurred during the early 1900's, the 1930's, and the early 1950's; they correspond to major droughts throughout the conterminous United States. During these prolonged periods of much-less-than-average precipitation, surface-water runoff, evapotranspiration, and ground-water infiltration decreased; consequently, ground-water levels declined as the demand for ground water increased. During the extended periods of much-greater-than-average precipitation, ground-water levels recovered and withdrawals of ground water decreased while runoff, evapotranspiration, and infiltration increased.

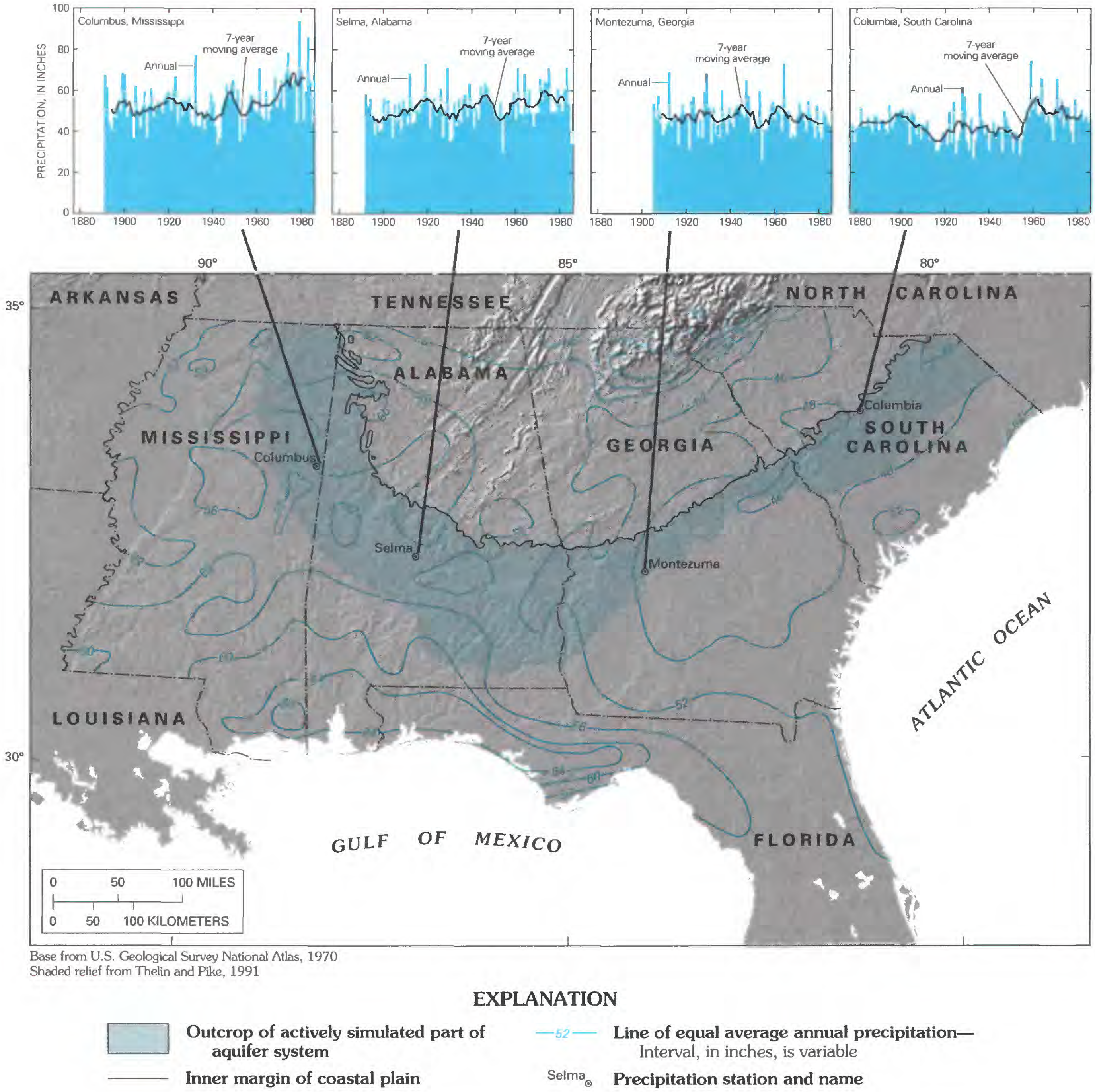
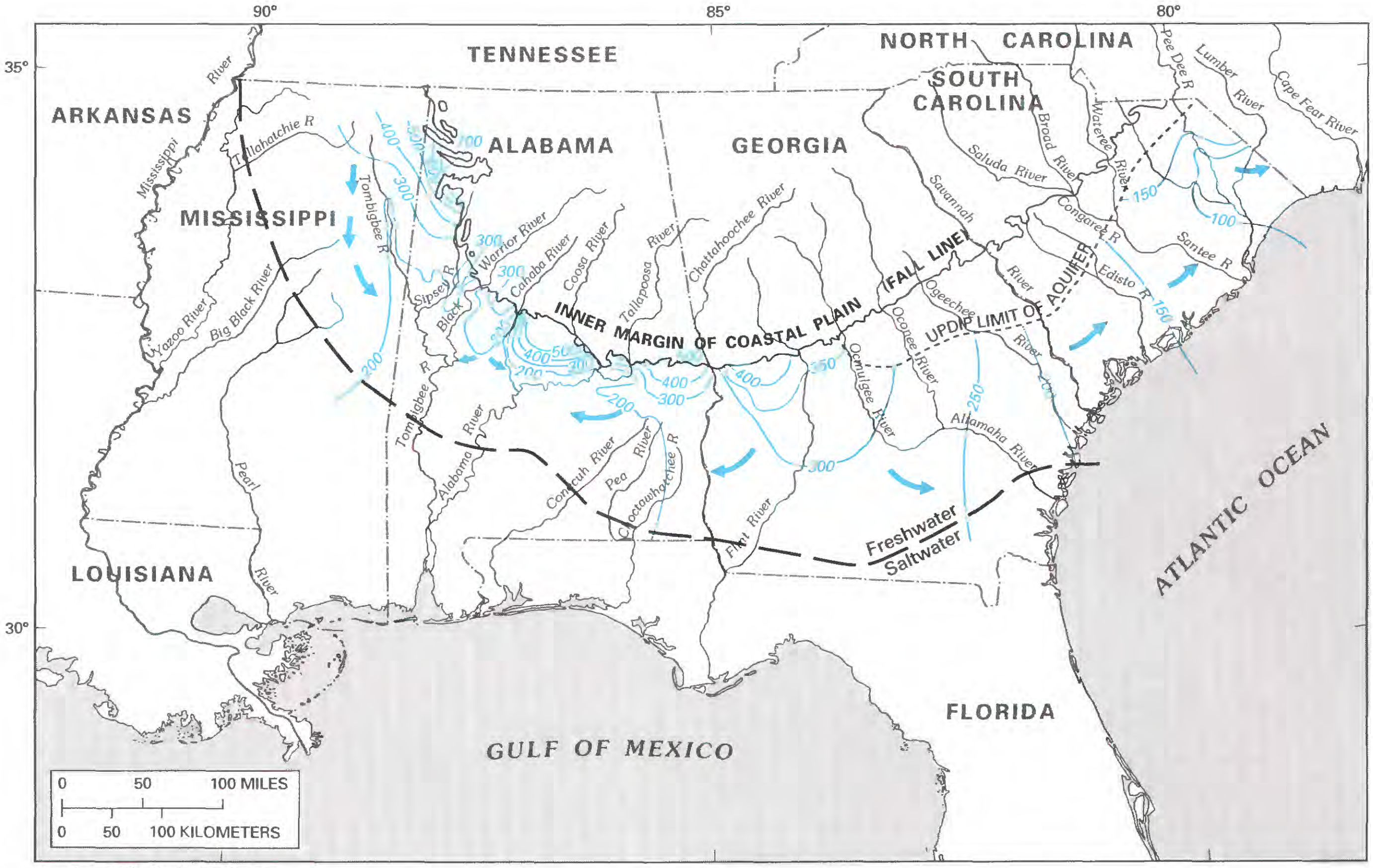


FIGURE 7.—Areal distribution of average annual 1951–80 precipitation, long-term annual precipitation, and 7-year moving average precipitation at selected stations (insets) in the outcrop area of the Southeastern Coastal Plain aquifer system. (Contoured data modified from U.S. Geological Survey, 1986.)

RUNOFF

Runoff is the second largest element of discharge in the Southeastern Coastal Plain (after evapotranspiration). It averages about one-third the amount of precipitation. Runoff has two components: (1) overland flow, or water that flows directly over the ground as surface runoff, and (2) base flow, or water that discharges from the ground-water system into some part of the surface-drainage network. Overland flow is generally most important where the terrain is steep, the soil texture is fine, and there is little plant cover, such as in the barren outcrop areas of low permeability in central Mississippi



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

EXPLANATION

—100— **Potentiometric contour**—Shows altitude at which water level would have stood in tightly cased wells [pre-1950]. Contour interval 50 and 100 feet. Datum is sea level

— — **Freshwater–saltwater interface**—Downdip extent of ground water having less than 10,000 milligrams per liter dissolved solids

→ **General direction of ground-water flow**

FIGURE 8.—Location of major streams, approximate location of the freshwater-saltwater interface, and generalized potentiometric surface of the Black Warrior River aquifer (modified from Barker, 1986, Stricker and others, 1985c, and Strickland and Mahon, 1986).

and western Alabama. Base flow is controlled largely by the underlying geology, the degree of stream entrenchment, and the head relations between ground-water levels and water levels in the surface drains. Shallow headwater streams receive base flow from locally occurring, principally unconfined aquifers. The major, more deeply entrenched streams—such as the Tombigbee, Alabama, Chattahoochee, Flint, Savannah, Edisto, and Pee Dee Rivers—receive base flow from the deep, principally confined aquifers, as shown by potentiometric surface maps (fig. 8). Although over the long term the shallow streams drain off a significant amount of ground water, many dry up during extended periods of little precipitation. Because the major streams tap flow paths deeper in the regional ground-water flow regime, they are less affected by either droughts or periods of above-average rainfall.

Results of base flow analyses by Aucott and others (1986, p. 26) and Faye and Mayer (1990) indicate that discharge from the deep ground-water flow regime to major streams in Georgia and South Carolina averages about 1 in/yr, or less than 0.1 cubic feet per second per square mile [$(ft^3/s)/mi^2$]. Using 1954 drought data for the Coastal Plain of Georgia, Callahan (1964, p. 12) determined that base flow to "minor streams" (presumably from the shallow ground-water regime) ranges from about 5 to 40 in/yr, or 0.4 to about 3.0 $(ft^3/s)/mi^2$. Results of hydrograph separation for 26 small streams in the outcrop of the Southeastern Coastal Plain aquifer system (Stricker, 1983) indicate that base flow from the shallow

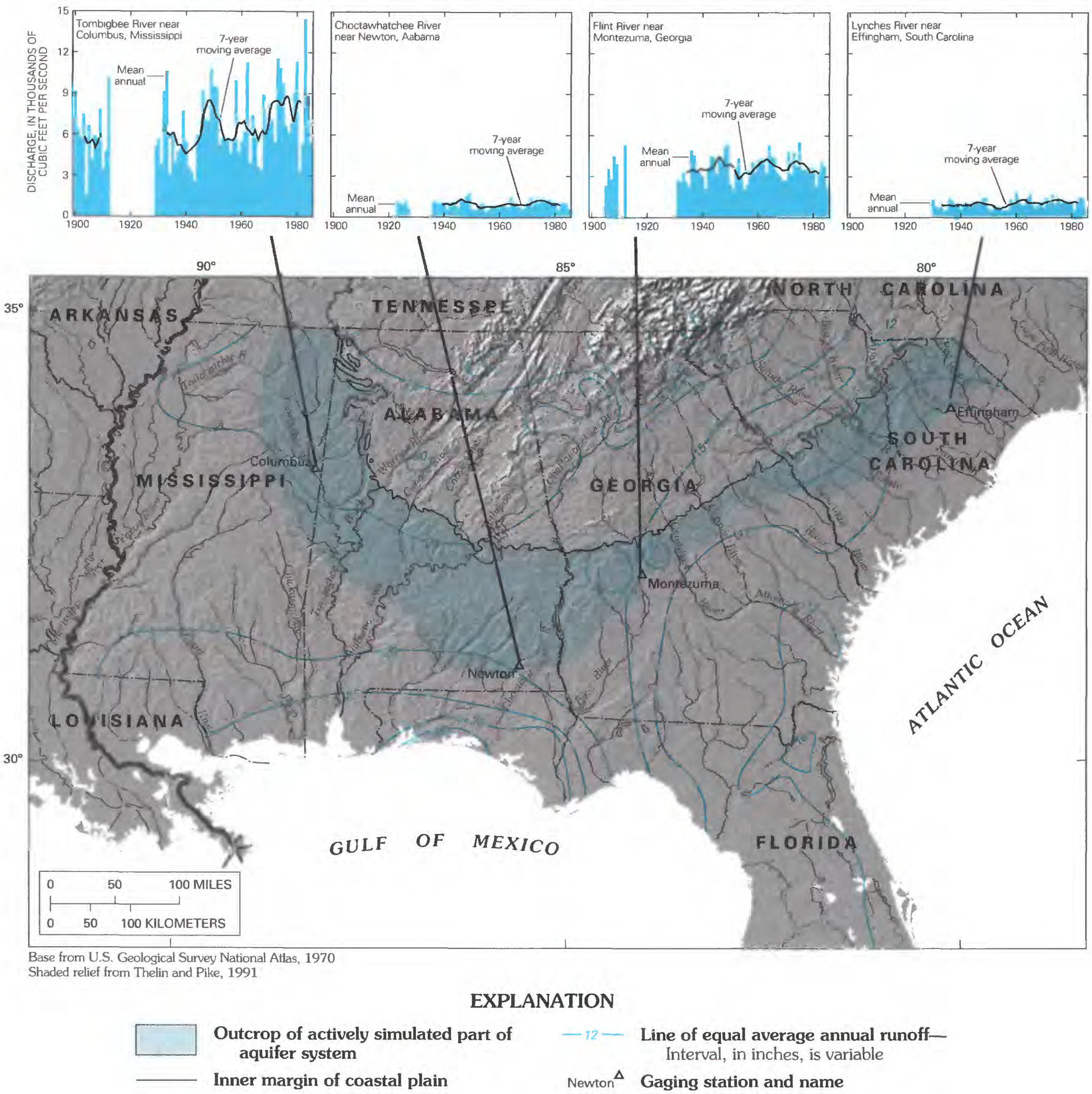


FIGURE 9.—Areal distribution of average annual 1951–80 runoff, long-term annual runoff, and 7-year moving average runoff at selected stations (insets) in the outcrop area of the Southeastern Coastal Plain aquifer system. (Contoured data from Gebert and others, 1987.)

ground-water regime in the study area averages about 6.5 in/yr, or 0.5 $(ft^3/s)/mi^2$.

Historical distributions of runoff in the study area are shown in figure 9. The contours illustrate the average annual runoff across Mississippi, Alabama, Georgia, and South Carolina during 1951–80. A set of bar graphs and 7-year moving average curves (insets) depict period-of-record distributions of annual runoff at representative stations on four major rivers in the outcrop of the Southeastern Coastal Plain aquifer system, all of which drain the regional ground-water flow regime.

The areal pattern of runoff (fig. 9) is similar to that of precipitation (fig. 7). Runoff generally increases from east to west, and in southern Mississippi and southern

Alabama it decreases with distance from the Gulf of Mexico. Within the study area, runoff ranges from about 12 in/yr along the coastline of Georgia and South Carolina to about 28 in/yr in southern Mississippi and averages about 17 in/yr. Within the outcrop area of the Southeastern Coastal Plain aquifer system, runoff generally increases from downdip to updip and averages about 19 in/yr.

Comparison of figures 7 and 9 shows that runoff is a smaller percentage of precipitation in the eastern part of the outcrop area than it is in the western part. The difference between precipitation and runoff in the outcrop area is less in Mississippi and Alabama than it is in Georgia and South Carolina (McGuinness, 1963). This circumstance occurs despite the fact that precipitation in the outcrop area of Mississippi and Alabama is nearly 6 in/yr more than it is in Georgia and South Carolina. Evapotranspiration appears to be relatively uniform across the outcrop area (Hamon, 1961; Geraghty and others, 1973). It appears, then, that infiltration (roughly the difference between precipitation and the sum of overland flow, detention storage, and evaporation) is greater in the outcrop area of eastern Alabama, Georgia, and South Carolina than it is in Mississippi and western Alabama. This is consistent with Stricker's (1983) results indicating that base flow (and, therefore, infiltration and ground-water recharge) increases from west to east across the outcrop area of the Southeastern Coastal Plain aquifer system. While averaging less than 5 in/yr for 13 stream reaches in Mississippi and western Alabama, Stricker's estimates of base flow average nearly 9 in/yr for the 19 analyzed stream reaches in eastern Alabama, Georgia, and South Carolina.

The suggested west-to-east increase in infiltration, ground-water recharge, and base flow would most likely result from the generally coarser texture of the sediments of predominantly fluvial origin that crop out across much of eastern Alabama, Georgia, and South Carolina (Renken, in press). A much higher percentage of the total outcrop area in western Alabama and Mississippi is composed of fine-grained materials such as the chalk, clay, and shale of the Selma Group and Midwayan sediments. The indication that infiltration and base flow are less in the western part of the study area, coupled with the fact that precipitation (fig. 7) and total runoff (fig. 9) are significantly larger here, suggests further that the ratio of overland flow to base flow is greater in the western part of the study area than in the eastern part. Again, this is to be expected owing to the wide expanses of fine-grained rocks of marine origin that crop out in Mississippi and western Alabama. The low hydraulic conductivity of these clays and chalks limits the opportunity for infiltration of precipitation to the water table, while enhancing the conditions for surface runoff.

EVAPOTRANSPIRATION

The greatest losses of water from the Southeastern Coastal Plain result from evaporation and transpiration, which together consume about 60 percent of the water from precipitation. It is difficult to assess independently the rates of evaporation and transpiration and separately evaluate their effects. For the study of the Southeastern Coastal Plain aquifer system, it was most efficient to consider evaporation and transpiration as the single process known as evapotranspiration. Evapotranspiration, as discussed here, is assumed to incorporate all discharge from the land surface and shallow subsurface of the outcrop area and transpiration by vegetation that is not accounted for in the water budget as either overland flow or infiltration.

Areal delineations of evapotranspiration for the Southeastern Coastal Plain are rare. Although regional maps of actual evapotranspiration are not available, two published maps of potential evapotranspiration cover the study area. Geraghty and others (1973, pl. 13) provided a map of potential evapotranspiration that is used by the U.S. Forest Service (G.E. Dissmeyer, U.S. Department of Agriculture, written commun., 1985) to estimate evapotranspiration losses from the oak-hickory-pine forests of the Southeastern United States. The map indicates that potential evapotranspiration ranges from about 36 in/yr across the northern edge of the study area to about 40 in/yr along the southern limits of the area. Hamon (1961, fig. 6) provided a map of potential evapotranspiration covering the eastern part of the United States that is based on his equation, which represents potential evapotranspiration as a function of daytime hours, absolute humidity, and mean air temperature. Because Hamon's methodology yields results that compare very well with the results of Lowry and Johnson (1942), Thornthwaite (1948), and Penman (1956) for specific areas of the United States where comparisons are possible, Hamon's map (fig. 10) was adopted to help estimate evapotranspiration losses from the area underlain by the Southeastern Coastal Plain aquifer system.

As shown in figure 10, potential evapotranspiration ranges from about 36 in/yr across the northeastern part of Mississippi to about 42 in/yr near the southeastern tip of Georgia. Potential evapotranspiration averages about 38 in/yr over the outcrop of the Southeastern Coastal Plain aquifer system. These estimates of potential evapotranspiration cannot be inserted directly into a water budget, because they are presumably on the high side of long-term average rates of actual evapotranspiration; they first must be corrected downward. To obtain estimates of actual evapotranspiration for Florida and parts of Georgia and Alabama, Bush and Johnston (1988) applied a nomogram by Holdridge (1967) to their estimates of potential evapotranspiration for the Floridan

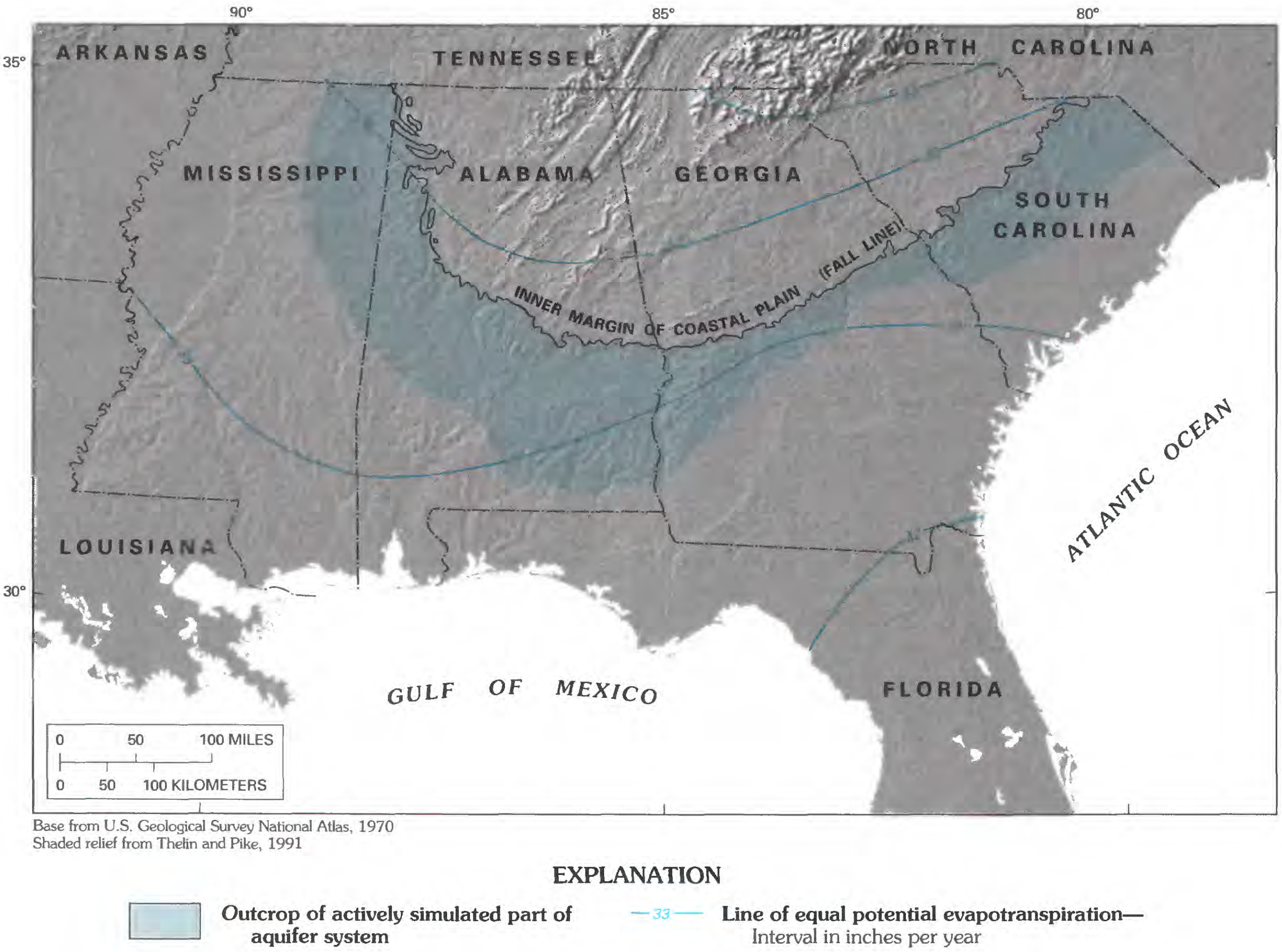


FIGURE 10.—Areal distribution of long-term average potential evapotranspiration (modified from Hamon, 1961).

aquifer system. Based in part on a "life zone" biochemical classification system and in part on empirical observation, the Holdridge nomogram assumes that mean annual precipitation and potential evapotranspiration are known. To obtain estimates of average annual actual evapotranspiration for the water budget of the Southeastern Coastal Plain aquifer system (see section on "Conceptualization of the Ground-Water Flow System"), the Holdridge nomogram was applied to the distribution of potential evapotranspiration depicted in figure 10. Results indicate that actual evapotranspiration probably ranges from about 30 to 35 in/yr across the study area and averages about 32 in/yr over the outcrop areas of the regional aquifers of the Southeastern Coastal Plain aquifer system. The average rate of actual evapotranspiration computed for the Southeastern Coastal Plain regional aquifer system in southern Alabama and Georgia (about 34 in/yr) compares very well with the average of the rates published by Bush and Johnston (1988, pl. 9) for these areas (about 33 in/yr).

INFILTRATION

The amount of recharge to an aquifer is limited by the amount of infiltration, which in turn is limited by the difference between precipitation and overland flow (ignoring the effects of surface storage and evaporation). Infiltration and overland flow are inversely related (fig. 11). The ratio of infiltration to overland flow decreases as the rate of precipitation exceeds the infiltration capacity of the soil.

The areal distribution of recharge to the regional ground-water flow regime of the Southeastern Coastal Plain was analyzed by model simulation. An areawide, long-term rate of recharge to the local ground-water regime was calculated for the water budget from what is known about precipitation, total runoff, evapotranspira-

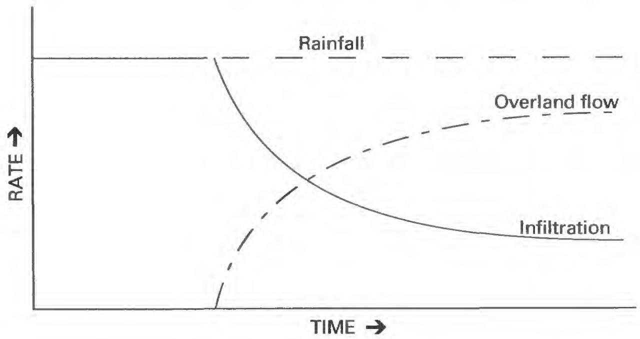


FIGURE 11.—Relation between rates of infiltration and overland flow, assuming constant rate of rainfall (modified from Freeze, 1979, p. 214).

tion, and base flow. Although these estimates of recharge are not based on direct observation of infiltration or recharge, they are consistent with the pattern of recharge potential.

Figure 12 illustrates the relative potential for recharge to result from the infiltration of precipitation in the outcrop area of the Southeastern Coastal Plain aquifer system. Because recharge is a direct residual of infiltration, the regional patterns reflected by this map were used to estimate the recharge rates for input to the simulation model of the aquifer system. Figure 12 shows three broad categories of recharge potential, which were inferred from the (1) outcrop lithology from Renken (in press), (2) soil type from the U.S. Department of Agriculture Soil Conservation Service (1967), and (3) baseflow distribution from Stricker (1983), Aucott and others (1986), and Faye and Mayer (1990).

Owing to the layered nature of the soils and clay accumulations common to the outcrop area of the Southeastern Coastal Plain aquifer system, much of the area has a low or intermediate recharge potential. Most areas of low recharge potential are in Mississippi and western Alabama; the clay-dominated soils in these areas are weathering products of extremely fine-grained rocks of marine origin, including the Porters Creek Formation of Paleocene age (Lowe, 1933) and the Prairie Bluff, Demopolis, and Mooreville Chalks of Late Cretaceous age. Most areas of intermediate recharge potential are within outcrops of the Tuscaloosa Group (Mississippi, Alabama, and Georgia), which were described by Stephenson and Monroe (1940), Cooke (1943), and Carlston (1944) and the Middendorf Formation (South Carolina), which was described under "Tuscaloosa Formation" by Cooke (1936). The areas of highest recharge potential are dominated by sandy soils that are rich in quartz and other minerals highly resistant to weathering. These particularly clean, porous sands are derivatives of fluvial deposits making up the Clayton Formation, the Providence Sand, the Ripley Formation, and the Cusseta Sand in eastern Alabama and Georgia (Cooke, 1943; Carlston, 1944), as well as the Barnwell Formation in South Carolina (Cooke, 1936).

CONCEPTUALIZATION OF THE GROUND-WATER FLOW SYSTEM

The Southeastern Coastal Plain aquifer system is underlain updip by dense, relatively impermeable rocks and downdip by a vast saltwater system. Most of the recharge to the aquifer system occurs in the interstream parts of the outcrop area under predominantly unconfined conditions; considerably less water enters updip, confined areas of eastern Alabama, Georgia, and South Carolina through downward leakage from the Floridan aquifer system, or from lateral clastic equivalents of the Floridan. Most discharge occurs from the updip, unconfined parts of the system as base flow to topographically low parts of the outcrop area. Smaller amounts of middip and downdip discharge occur from confined parts of the system as diffuse upward leakage to the Floridan aquifer or its equivalents. The rate of ground-water intake at any given place in the Southeastern Coastal Plain aquifer system is limited by the capacity of the system to transmit the water received.

Because topographic relief and sediment grain size within the Southeastern Coastal Plain aquifer system generally decrease downdip from the outcrop area (Renken, in press), and because dissolved-solids concentrations in the ground water increase with both depth below land surface and distance from the updip recharge areas (Lee, 1993), hydraulic gradients and conductivities decrease from updip to downdip while water density increases. Consequently, the energy available to circulate freshwater decreases as a function of depth below land surface and distance from recharge areas; ground-water-flow velocities decrease substantially between the shallow outcrop area (updip) and the deep freshwater-saltwater interface (downdip). The downdip vertical leakage is diffuse and sluggish compared with the updip discharge as base flow to surface-water bodies. Shallow, updip parts of the aquifer system transmit more water than deeper, downdip parts because the updip discharge under atmospheric conditions toward surface streams is less restricted than the downdip discharge that takes place under confined conditions as leakage through thick confining beds. Water that infiltrates the relatively high-energy environment of the shallow outcrop area cannot as readily penetrate deeper, less dynamic parts of the confined flow system; consequently, a considerable amount of the interstream recharge discharges updip as base flow to streams within relatively short distances of where it enters. The effects of the interconnected factors

FIGURE 12.—Recharge potential in outcrop areas of aquifers and confining units simulated in the model.

of climate, geology, and topography on the Southeastern Coastal Plain aquifer system are discussed below.

CLIMATIC EFFECTS

Despite seasonal and geographical variations, the humid climate of the Southeastern United States generally ensures an abundance of precipitation for the study area (fig. 7). Most precipitation results from cyclonic, convectional, or orographic circulation of moisture-laden air from the Gulf of Mexico and, to a lesser extent, the Atlantic Ocean. The wettest seasons are winter, early spring, and midsummer; the driest periods are generally May through June and October through November. Most of the moisture that results in recharge to the groundwater system falls during the winter months as warm, moist air from the Gulf of Mexico is pushed northeastwardly by cyclonic forces over a relatively cool Coastal Plain, producing typically mild but prolonged storms over wide areas. The most intense rainfall occurs during the hot summer months as the result of convectional or thunderstorm activity that is comparatively turbulent and spotty in areal extent. The nonuniform nature of summer precipitation occasionally causes concern for farmers and officials responsible for public and recreational water supplies. Precipitation increases north of the topographic break (at the Fall Line, fig. 10) between the relatively low lying Coastal Plain and the Piedmont, Valley and Ridge, and Appalachian Plateaus physiographic provinces—over which the cooling and condensation of airborne moisture is enhanced by orographic lifting of the airmasses.

Recharge to the Southeastern Coastal Plain aquifer system starts with precipitation in the outcrop area. As a result of warm summer temperatures and brisk air movement in the region, potential evapotranspiration demand is relatively high, ranging from about 36 to 42 in/yr (fig. 10). The difference between the long-term average rates of precipitation (51 in/yr) and estimated evapotranspiration (32 in/yr) is about 19 in/yr. Of this 19 in/yr, approximately 12 in/yr runs off the land surface as overland flow. Although evapotranspiration and overland flow reduce the supply of water from precipitation by about 85 percent, an average of 7 inches (in) remains on an annual basis to become recharge.

Callahan (1964, p. 8) pointed out that little recharge occurs to aquifers in the Southeastern Coastal Plain during the warm growing season, because most of the precipitation is used to renew soil moisture, is transpired by vegetation, or is evaporated from the land surface. Hydrographs of ground-water levels indicate that except for some of the shallowest unconfined aquifers, nearly all recharge takes place during the cool nongrowing season, when transpiration and evaporation losses are at a minimum (Callahan, 1964, p. 8). Because precipitation generally exceeds evapotranspiration during the period November through March (the nongrowing season), the opportunity for recharge in the Southeastern Coastal Plain peaks during this 5-month period. Although recharge varies seasonally, it occurs with sufficient regularity that aquifer-to-stream gradients are seldom reversed by natural events.

The net effect of precipitation, evapotranspiration, and overland flow in the outcrop area provides a high recharge potential that, in combination with the decreasing hydraulic conductivity and topographic relief in a downdip direction, results in a water-table configuration that resembles the hummocky, rolling terrain it underlies. Although the water table in the stream valleys is typically at or within a few feet of land surface, levels are seldom more than a few tens of feet below land surface in the interstream areas of the outcrop. Where and when recharge to the outcrop area exceeds the capacity of the downgradient ground-water system to transmit water, the excess discharges to nearby surface-water bodies. The relatively high water levels that are sustained by frequent recharge events cause most watercourses in the Coastal Plain to be perennial, fed by gaining reaches the year around. Normally, only the shallowest upland tributaries lose water to the ground-water system or become dry during lulls in the supply of precipitation.

GEOLOGIC CONTROLS

The amount of ground water flowing through downdip, confined parts of the Southeastern Coastal Plain aquifer system is less than that flowing through the updip, unconfined parts. Most of the potential energy within the aquifer system is dissipated between areas of recharge and discharge across the zones of least conductivity (Freeze and Witherspoon, 1967). The dynamics of confined ground-water flow deep within the aquifer system are substantially less than that of the shallow, unconfined flow because the downdip resistance to discharge through thick confining beds is greater than the updip resistance to base flow through thinner, more permeable streambeds.

LeGrand and Pettyjohn (1981) have suggested that the interaction between ground water and surface water in a Coastal Plain setting depends on the region's climate, topography, and geologic structure and on the distribution of permeability. These authors suggest that the opportunity for ground-water discharge decreases rapidly with increasing distance from the outcrop area and, in particular, increasing distance from the stream channels. They point out that stream channels in a coastal plain environment are generally of the consequent type. In high-rainfall regions, consequent streams drain both aquifers and confining units, owing to relatively high ground-water levels in interstream areas and a pattern of stream entrenchment that is generally perpendicular to the strike of the bedding. Noting that flow lines usually converge toward major rivers near the downdip margins of aquifer outcrops, LeGrand and Pettyjohn suggest that the greatest accretions of base flow occur near downdip parts of aquifer outcrop areas, where the outcrops are widest and river-bottom elevations are lowest.

To explain the concentration of ground-water discharge near the downdip edges of aquifer outcrop areas, LeGrand and Pettyjohn (1981) discussed the implication of erosional notches that breach the updip edges of overlying confining units where they are crossed by consequent rivers (fig. 13). The authors contend that resistance to upward leakage is short circuited within these V-shaped "artesian-water gaps," causing ground-water discharge to be enhanced locally. They point out that these incisions into a beveled outcrop area result from river downcutting, and suggest that they provide "the last downdip place where ground water can be discharged readily" from a sloping Coastal Plain aquifer system. As an example of an artesian-water gap, LeGrand and Pettyjohn (1981) referred to an area along the Savannah River downstream from Augusta, Ga.,

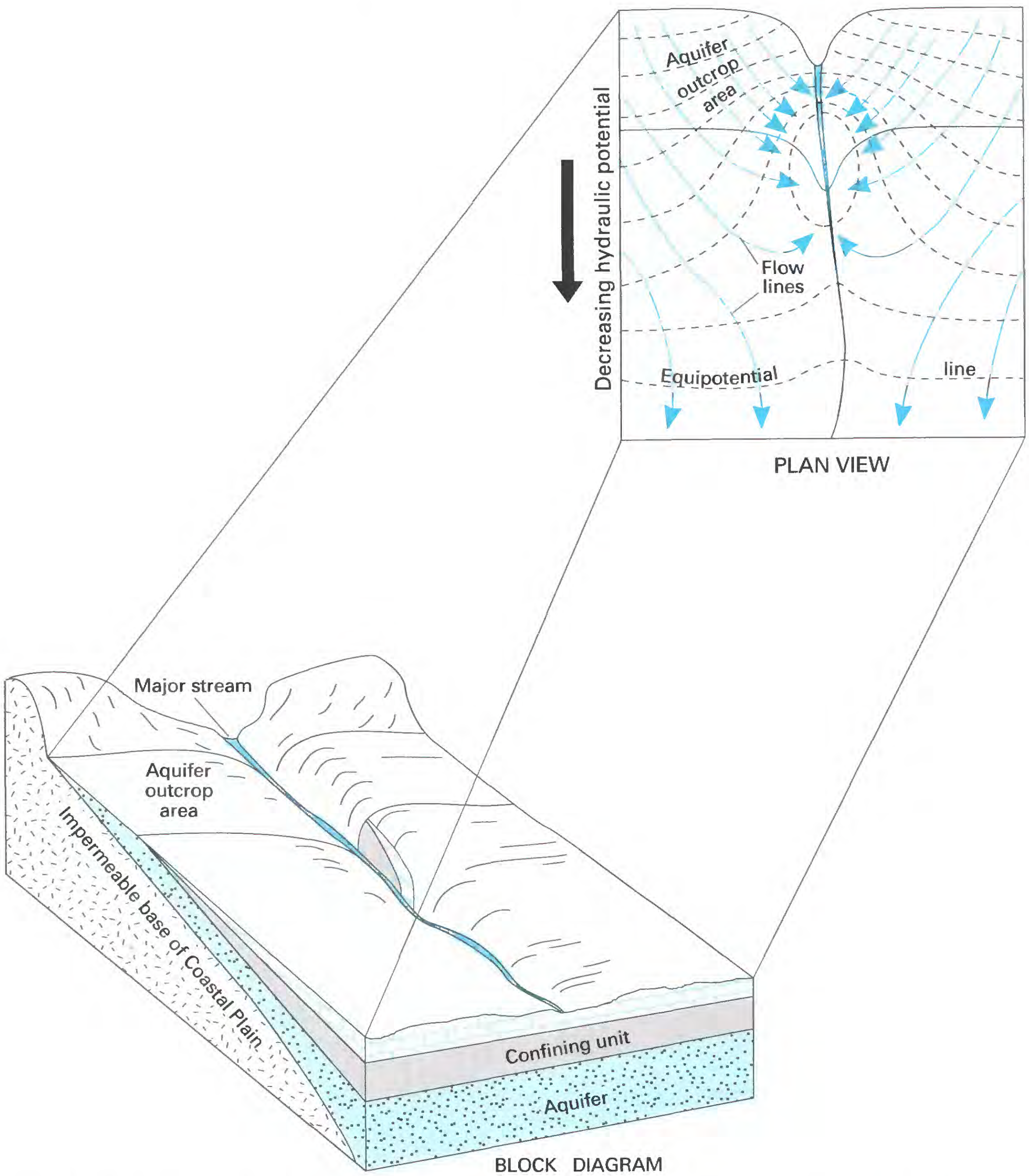


FIGURE 13.—How erosional notches breach updip edges of shallow confining units and enhance discharge from underlying coastal plain aquifers (modified from Heath, 1984).

that spans the outcrop of the "Cretaceous sand aquifer" (Chattahoochee River aquifer of this report) in addition to other "younger Coastal Plain sand, clay, and limestone deposits" (Pearl River aquifer and Floridan aquifer system of this report).

Using October 1954 water-level measurements, Siple (1960) compiled a potentiometric map for the outcrop area of the "principal sand aquifer of Late Cretaceous age" (Chattahoochee River aquifer of this report) adjacent to the Savannah River in Georgia and South Carolina. Because his potentiometric map is dominated by a pronounced depression that straddles the river and is bounded by closed contours on the downriver side, Siple inferred that ground water discharges very readily into the river. Based on estimates of average transmissivity and gradient between the interstream and stream areas, Siple calculated that about 260 cubic feet per second (ft^3/s) discharged from the aquifer through roughly 30 mi of river channel. His calculation of ground-water discharge is substantiated by his independent computations of streamflow gain, indicating a base-flow accretion of about 300 ft^3/s during water year 1954–55. Siple (1960) attributed the predominance of ground-water discharge near the Savannah River to a pattern of aquifer confinement that decreases toward the river. While acknowledging that one would ordinarily expect the overlying clay to limit ground-water discharge, he proposed that the river had cut through the confining clays into the aquifer, probably sometime during the Pleistocene epoch, when the sea level was at least 200 ft lower than at present and the river had adjusted to the change in gradient by downcutting in its upper reaches.

The confining unit that separates the Chattahoochee River aquifer from the overlying Pearl River aquifer near the Savannah River was mapped by Renken (in press) as the Chattahoochee River confining unit (fig. 4). The map of the thickness of this unit (Renken, in press) indicates that the updip extent of the unit is farther south and (or) is thinner near the river than away from the river; this observation is consistent with the configuration suggested by Siple (1960).

Siple believed the conditions for enhanced ground-water discharge that he identified along the Savannah River to be characteristic of hydrologic conditions on other large streams where they cross the outcrops of confining beds in the Coastal Plain. Maps of confining unit thicknesses (Renken, in press) indicate patterns of postdepositional scouring, similar to that along the Savannah River, where these confining units are traversed in their outcrop areas by major rivers, such as the Congaree, Oconee, Ocmulgee, Flint, and Chattahoochee. Maps by Renken of confining unit thickness do not indicate such a pattern west of eastern Alabama, however.

Results of a computer-model study by Gardner (1981) indicate comparatively moderate rates of ground-water leakage to the major rivers from the Eutaw and Gordo sediments in western Alabama. The Eutaw and Gordo sediments make up the upper part of the Black Warrior River aquifer (pl. 1). Gardner concluded that large troughlike depressions in the potentiometric surfaces of the Eutaw and Gordo aquifers are caused by upward leakage from the aquifers through their confining units to the valleys of the Alabama, Black Warrior, and Tombigbee Rivers. Simulated steady-state rates of aquifer discharge to these rivers ranged from 70 to 100 ft^3/s and averaged about 85 ft^3/s. Using 1942–44 streamflow records, Gardner calculated that streamflow increased about 10 percent across a 20-mi reach of the Tombigbee River and deduced that about 6 to 7 percent of the total flow was attributable to ground-water inflow. Because the data (Gardner, 1981, table 4) indicate an average increase in streamflow of about 50 ft^3/s, perhaps about 30 ft^3/s represents base-flow accretion. Acknowledging that the streamflow error may be 5 to 10 percent, Gardner was cautious about comparing his simulated rates of discharge directly with the base-flow data. Nevertheless, his simulated losses to reaches of the Alabama, Black Warrior, and Tombigbee Rivers (totaling as much as 100 ft^3/s) appear to be consistent with the deduced base flow (of about 30 ft^3/s) to a relatively short reach of the Tombigbee River, alone. Gardner's (1981) modeling analysis was limited to west-central Alabama, where the rivers flow over thick sequences of clay and chalk that make up the Black Warrior River and Chattahoochee River confining units (fig. 5).

Because the confining units are thicker and dip more steeply in western Alabama and in Mississippi, the effects of stream entrenchment are less significant there than in Georgia and South Carolina. As a result, the rates of ground-water discharge to rivers in Mississippi and western Alabama are less than those in the eastern part of the Southeastern Coastal Plain. Callahan (1964, p. 12) stated that perhaps the largest base flows to streams occur in western Georgia and eastern Alabama, where the aquifers are exposed over wider areas, are not overlapped by younger rocks, and are therefore in a position to absorb larger amounts of rainfall. Callahan's observation is consistent with the east-to-west decrease in infiltration potential (fig. 12), but east-to-west differences in the geologic control on upward leakage from confined parts of the system are also very important. Although upward leakage to stream channels is the most important avenue for discharge in Mississippi and western Alabama, the predominance of fine-grained sediment in the streambeds severely limits the rate of leakage to the Alabama, Black Warrior, and Tombigbee Rivers.

Geologic control is also responsible for a significant contrast in tributary contributions to the Tombigbee River in Mississippi; a marked difference exists in low flow between the tributaries draining the eastern side of the Tombigbee basin and those draining the western side (Boswell, 1963). The difference in ground-water discharge to the Tombigbee River occurs because the course of the river is controlled by the strike of the bedding, and the river is located between the outcrop of a nearly impermeable confining sequence on the west and the outcrop of a productive group of aquifers on the east. The western tributaries are underlain by rocks that make up the Black Warrior River confining unit. Tributaries on the east side of the Tombigbee River drain the outcrop of the Eutaw and McShan Formations in addition to the Tuscaloosa Group (all of which constitute the Black Warrior River aquifer). These strata include layers of sand and gravel, many of which are exceptionally permeable. While ground water on the east side of the Tombigbee River is released at a fairly uniform rate, sustaining the base flow of streams, the western basins are dominated by rapid overland runoff and flash flooding during periods of heavy precipitation and by little or no base flow during dry periods, according to Boswell (1963).

Regional potentiometric maps by Stricker and others (1985a, 1985b, 1985c) indicate a consistent link between the major rivers atop the Southeastern Coastal Plain aquifer system and potentiometric "sinks" that are limited, for the most part, to the outcrop area of the system. The equipotential lines on these maps form depressions adjacent to the valleys of the major rivers where they cross the updip parts of the regional aquifer units. These elongate potentiometric lows result from the loss of hydraulic head in the aquifer, as the deeply entrenched rivers pick up ground water from the outcrop area. The relative absence of potentiometric relief in the deeper, downdip parts of the system (fig. 8) emphasizes that the outcrop area is the most efficient place for ground-water discharge. The water discharged per unit area is much greater in the outcrop area, where the permeability is relatively high and the gradients between the interstream recharge areas and the topographically low discharge areas are relatively large. The downdip, confined ground water flows over long, circuitous routes and eventually discharges upward through wedges of confining material that thicken and become less permeable toward the Atlantic Ocean and Gulf of Mexico.

TOPOGRAPHIC INFLUENCES

The Southeastern Coastal Plain aquifer system is a hydrologic continuum encompassing wide-ranging conditions. To understand the differences between the shallow conditions updip and the deep conditions downdip, it is helpful to segregate the system into its various components. Numerous researchers have studied the tendency for a ground-water system to be hydraulically segmented. Most of these investigators separate the whole system into two or three zones or subsystems based on the observation that circulation patterns differ from place to place within a large ground-water basin. Although circulation patterns are controlled by a combination of climatic, geologic, and topographic factors, researchers seem to agree that the nature of ground-water motion varies fundamentally with basin geometry, permeability, and the shape of the water table. Aquifer systems are often subdivided on the basis of depth—or the ratio of basin depth to some measure of basin extent, such as length or width. As a general rule, the more pronounced the surface topography, the greater the water-table relief, and—as a consequence—the more segmented the system's flow dynamics. Contrasts between updip and downdip dynamics in the Southeastern Coastal Plain aquifer system are related to topographic differences between the updip, outcrop area and the downdip, subcrop area.

Using analytical models, Toth (1963) explored a wide range of topographic effects on ground-water motion by solving for fluid potential, or hydraulic head, under isotropic conditions within hypothetical basins. He theorized that three distinctly different types of ground-water systems occur: local, intermediate, and regional (fig. 14). From the results of Toth (1963), we know that the greater the topographic relief, the greater the opportunity for local flow systems to exist. The more pronounced the relief of the water table, the deeper the local flow systems extend. Therefore, the shallower the basin, the greater the chances that only a local system prevails. As the ratio of basin depth to basin width increases, so do the chances for intermediate, or intermediate and regional, systems to develop. Where the combination of a slight regional dip (of bedding) and negligible local topography occurs, the potential is great that a regional flow system exists.

The rate of ground-water movement is directly influenced by hydraulic conductivity and gradient. In a coastal plain environment, the ground-water gradient is determined largely by (1) the elevation at which recharge occurs in the outcrop (interstream) areas, (2) the elevation at which discharge occurs in the most topographically depressed areas downgradient, and (3) the ease with which water travels between the areas of recharge and discharge. The velocity of ground-water flow is, therefore, dependent on the amount of topographic relief in the outcrop area and on the distribution of fine-grained, low-permeability sediments within the subcrop area. Because the thickness of clay deposits increases downdip from the outcrop area, the opportunity for discharge from the wedge-shaped Coastal Plain system should decrease as a function of depth below land surface as well as distance from the outcrop area. This idealization is supported with respect to the Southeastern Coastal Plain by the results of Lee (1993).

Lee (1993) used the technique of carbon-14 dating to compute both the age and the velocity of water in the Southeastern Coastal Plain aquifer system. Lee found that while the residence time of ground water in the system increases with depth and length of the flow path, the velocity of ground-water flow decreases with depth of confinement and distance from the outcrop. According to Lee, average velocities within the aquifer system decrease from about 15 feet per year (ft/yr) in the relatively shallow Pearl River aquifer, to about 10 ft/yr

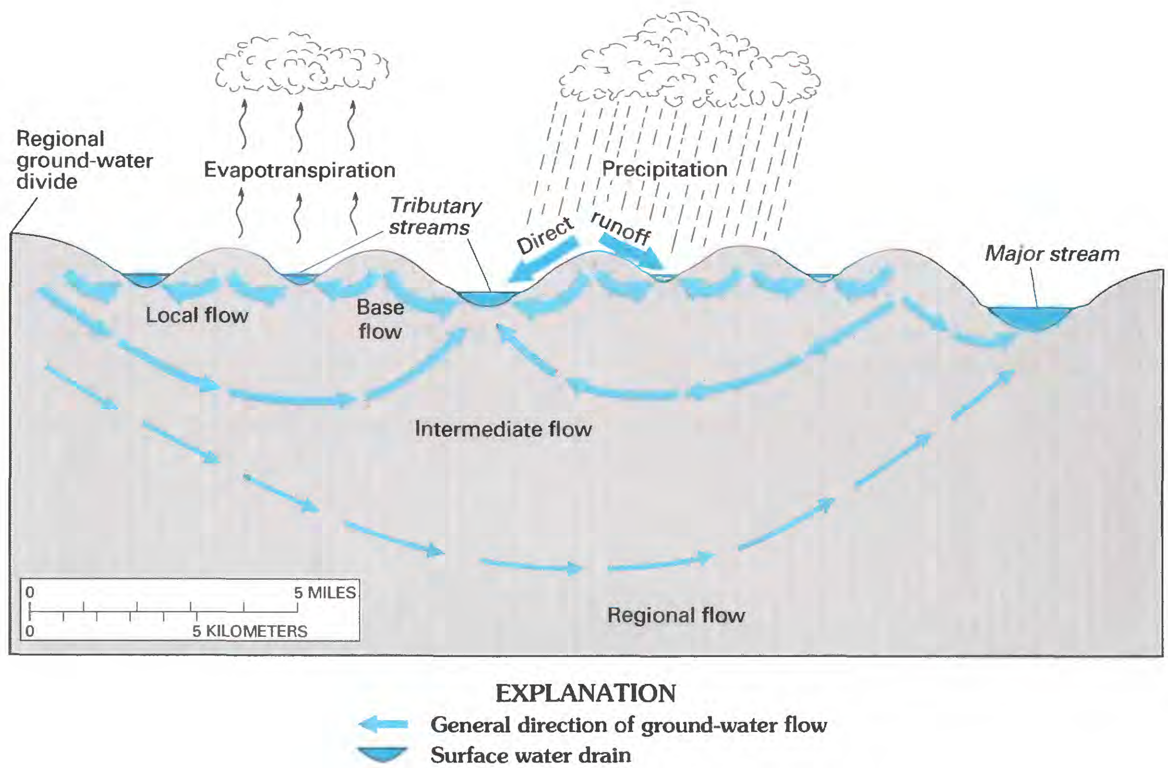


FIGURE 14.—Relation among local, intermediate, and regional patterns of ground-water flow and surface-water drainage (modified from Toth, 1963).

in the deeper Chattahoochee River aquifer, to less than 3 ft/yr in the basal Black Warrior River aquifer.

Freeze and Witherspoon (1967) investigated the effect of basin stratigraphy on ground-water flow by splitting a hypothetical basin into two or more layers; the permeability was held constant within each layer but differed between adjacent layers by at least one order of magnitude. Through computer simulation, they tested the effects of layer thickness, permeability contrasts, and water-table relief on the hydraulic segmentation of the flow system. Their results suggest that differences in the flow patterns between shallow and deep parts of a layered system depend primarily on the amount of variation in the water-table surface within the uppermost layer. Although the amount of flow through the system depends on layer thickness, as well as on permeability and hydraulic gradient, thickness alone has little effect on the nature of the flow pattern. Regardless of the permeability contrasts among layers, the existence of significantly different flow patterns between the uppermost and deeper layers seems predicated upon there being a "hummocky" water-table configuration within the upper layer. Freeze and Witherspoon (1967) concluded that hummocky water-table configurations are conducive to the formation of small subbasins within major basins; in such cases, the concept of a total basin yield is misleading and each component basin must be considered separately. It appears that relatively little topographic variation at the top of a layered flow system is required to induce patterns of ground-water motion that resemble the local, intermediate, and regional components of Toth (1963).

In light of the research by Toth (1963), Freeze and Witherspoon (1967), and others, it is conceivable that the relatively short flow paths and steep gradients in the updip parts of the Southeastern Coastal Plain aquifer system are caused more by the hummocky land surface in the outcrop area than by variations in permeability. Accordingly, the longer and flatter flow paths common to the downdip area result mostly from the (1) comparatively gentle overlying land-surface topography, (2) downdip decrease in the hydraulic conductivity of aquifers, and (3) downdip increase in the thickness of confining units.

WATER BUDGET

Ground-water circulation differs from place to place within the Southeastern Coastal Plain aquifer system as a function of depth below land surface and distance from the outcrop area. These differences were used to subdi-

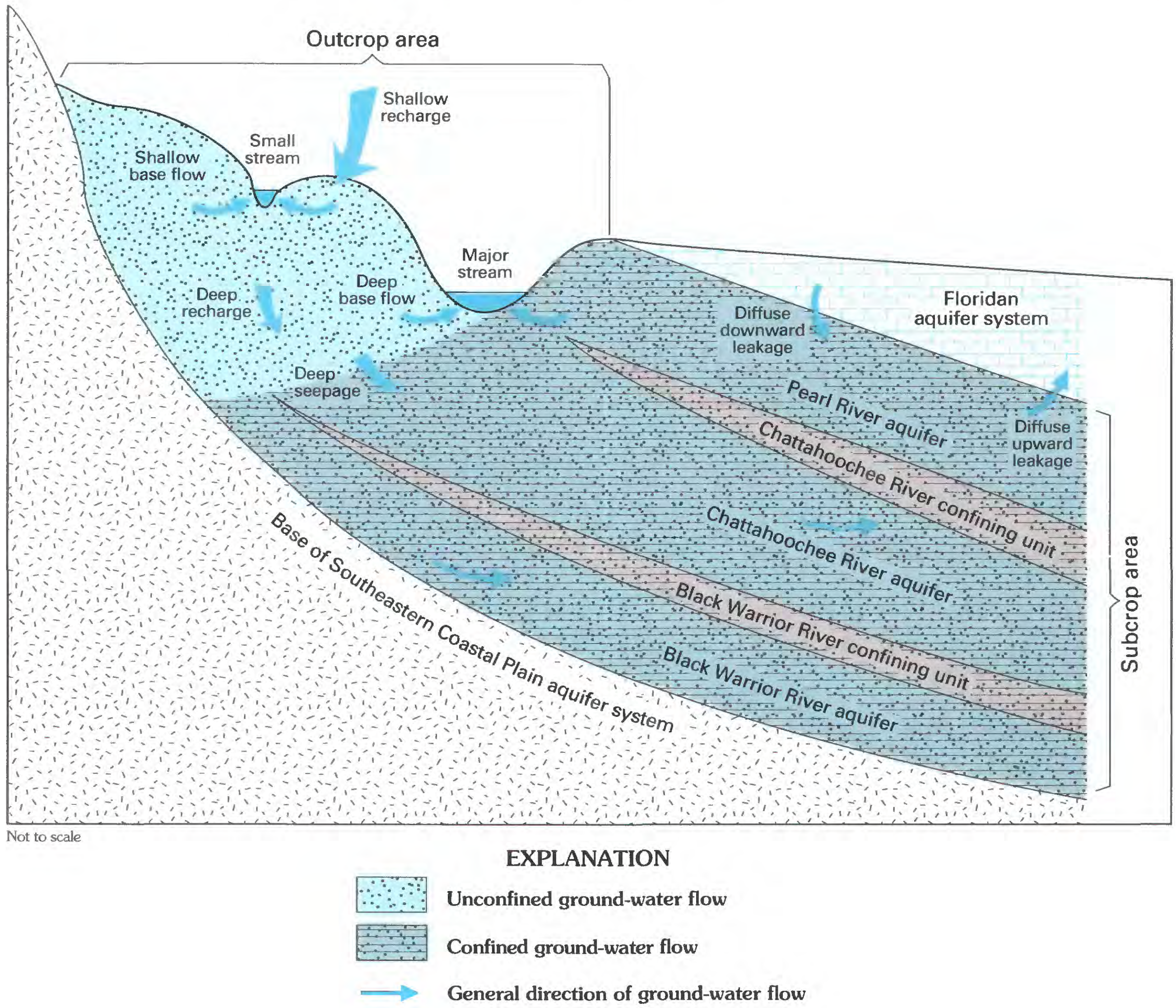
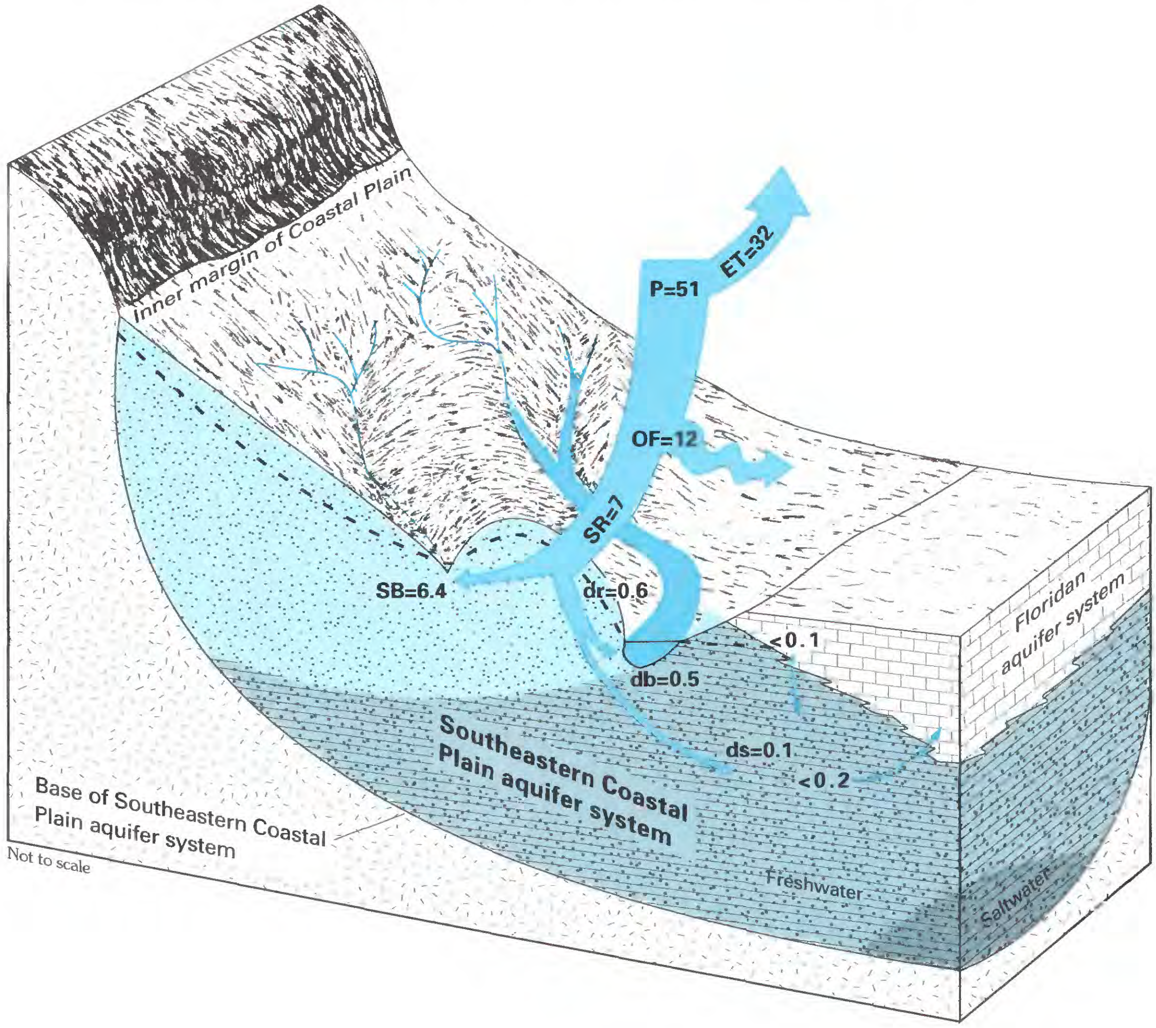


FIGURE 15.—Relation between the outcrop and subcrop areas and the shallow, unconfined and deep, confined components of ground-water flow.

vide the system for the purposes of modeling and estimating the water budget. Because the aquifer system is too complex to be modeled in its entirety, it was necessary to isolate the less extensive, shallow flow paths and exclude them from the simulation. Consequently, the water budget—which was derived partly through simulation—represents the aquifer system as two subsystems: (1) a "shallow" (nonsimulated) flow regime that is limited to the outcrop area, is under predominantly unconfined conditions, and is drained by relatively small streams, and (2) a "deep" (simulated) flow regime that exists everywhere in the subcrop and in the deepest parts of the outcrop, is under predominantly confined conditions, and is drained by major rivers in the Coastal Plain (fig. 8).

It is impractical to attempt to rigorously quantify the hydrologic subdivision of the Southeastern Coastal Plain aquifer system. It is clear from the literature that there is no consistent set of criteria upon which to base the subdivision of a hydrologic system. Nevertheless, figure 15 and equations 1 and 2 (below) are included to aid understanding of the relation between the conceptualization of the shallow and deep subdivisions of the flow system and the water budget presented in figure 16. The shallow flow regime described herein relates to the "local"—and probably some of the "intermediate"—circulation that was described by Toth (1963). As used herein, the deep flow regime includes all of Toth's "regional" flow, plus the part of his "intermediate" flow that occurs under confined conditions.



EXPLANATION

Aquifer in outcrop area—Under predominantly unconfined conditions

Aquifer in subcrop area—Under predominantly confined conditions

Major stream—Interacts primarily with regional flow system

Small stream—Interacts primarily with local to intermediate flow system

- - - **Hydraulic head**

Schematic depiction of average annual water budget—Based on published data and results of steady-state simulation. Numbers are flow rates, in inches per year

Nonsimulated:

P=Precipitation

ET=Evapotranspiration

OF=Overland flow

SR=Shallow (total) **recharge**—The percolation of precipitation below land surface that is lost neither to overland flow nor evapotranspiration

SB=Shallow base flow (to small streams)

Simulated:

dr=Deep recharge—Component of shallow (total) recharge that percolates below the hydraulic influence of small streams

db=Deep base flow (to major streams)

ds=Deep seepage—Component of total recharge that percolates below level of major streams

FIGURE 16.—Simulated and nonsimulated components of the hydrologic cycle and the long-term water budget under predevelopment conditions for the Southeastern Coastal Plain aquifer system (modified from Barker, 1986).

For the purpose of constructing the computer model and compiling the water budget, recharge was divided into three categories, or levels, according to depth and characteristics of the flow system. *Shallow recharge* occurs when precipitation infiltrates the water table in unconfined parts of the outcrop area. Shallow recharge is especially susceptible to the effects of gravity and the relatively unrestricted avenues of shallow discharge. As shown in figure 16, about 90 percent of the shallow recharge becomes *shallow base flow* to streams, creeks, and lakes—leaving about 10 percent of the recharge to percolate deeper. *Deep recharge*, the residual of shallow recharge and shallow base flow, is water that percolates from the unconfined toward confined parts of the flow system. Most deep recharge occurs in the interstream areas of major drainages. Similar to the way shallow recharge is depleted by shallow base flow, most of the deep recharge becomes *deep base flow* to major rivers draining the lower elevations of the outcrop area. The difference between deep recharge and deep base flow is herein termed *deep seepage*. Too deep and too far downdip to interact significantly with even the most deeply entrenched rivers in the Coastal Plain, deep seepage occurs only under confined conditions. Normally, deep seepage passes from interstream areas of the outcrop to subcrop areas, where it remains until it discharges as diffuse upward leakage. In places, however, deep seepage migrates from subcrop to outcrop areas—along sweeping, arcuate, nearly horizontal flow paths—where it may discharge to major rivers near the downdip edges of the outcrop area (fig. 13). Discharge resulting from this kind of circulation is most prevalent in the deeper aquifers of Mississippi and western Alabama (M.J. Mallory, U.S. Geological Survey, written commun., 1990), although it also exists near the Chattahoochee and Savannah Rivers in eastern Alabama, Georgia, and South Carolina.

Shallow recharge to the outcrop area can be expressed mathematically as

$$SR=P-OF-ET, \quad (1)$$

where

SR = shallow recharge (in inches per year);
P = precipitation (in inches per year);
OF = overland flow (in inches per year); and
ET = evapotranspiration (in inches per year).

Equation 1 can be modified to show the conceptual difference between the simulated recharge to the deep, confined part of the Southeastern Coastal Plain aquifer system and the shallow recharge to the nonsimulated part. Thus

$$dr=P-OF-ET-SB \quad (2)$$

or

$$dr=SR-SB,$$

where

dr = deep recharge of ground water below the level of small streams (in inches per year); and
SB = shallow base flow to small streams that drain unconfined parts of the aquifer system (in inches per year).

The relatively small component of the deep recharge that does not discharge as deep base flow to the major rivers can be expressed as

$$ds=dr-db, \quad (3)$$

where

ds = deep seepage of ground water below the reach of major rivers (in inches per year); and
db = deep base flow to major rivers that drain mostly confined parts of the aquifer system (in inches per year).

A schematic water budget depicting predevelopment conditions in the Southeastern Coastal Plain aquifer system is shown in figure 16. The rates of precipitation and runoff (overland flow plus base flow) were planimetered across the outcrop area of the aquifer system from the mean annual contour maps of these data for the period 1951–80 (figs. 7, 9). The rate of evapotranspiration was planimetered across the outcrop area from a contour map of long-term-average potential evapotranspiration by Hamon (1961, fig. 6) and adjusted for estimates of field evapotranspiration with a nomogram prepared by Holdridge (1967). The rate of base flow to the shallow surface water network was calculated from estimated rates of base flow to 26 small streams in the outcrop area (Stricker, 1983, table 1). The rates of recharge to the deep, confined parts of the aquifer system and of discharge to major rivers and overlying Tertiary rocks were simulated with the digital model described below.

Although it was necessary to draw from a combination of sources to construct the water budget, care was taken to ensure that the data were as compatible as possible. The indicated rates are rounded according to the relative amount of control on the potential accuracy of the data. The numbers that result directly from long-term average distributions of contoured data (precipitation, evapotranspiration, overland flow, and shallow recharge) are rounded to the nearest inch. The model-derived numbers (deep recharge, deep base flow, and deep seepage—as well as the discharge and recharge to and from the overlying Floridan aquifer system) are rounded to the nearest tenth of an inch.

The evapotranspiration rate (32 in/yr) is equal to the difference between the planimetered rates of precipitation (51 in/yr) and runoff (19 in/yr). The shallow base flow

(6.4 in/yr) is the area-weighted average of base flow from small basins in the outcrop area (Stricker, 1983). The rate of overland flow (12 in/yr) results from subtracting the sum of the shallow base flow (6.4 in/yr) and the simulated deep base flow (0.6 in/yr) from the planimetered total runoff of 19 in/yr. The rate of evapotranspiration (32 in/yr) is the planimetered rate of potential evapotranspiration (Hamon, 1961), corrected for actual evapotranspiration in the outcrop area. The shallow recharge (7 in/yr) results from the difference between precipitation (51 in/yr) and the sum of evapotranspiration and overland flow (44 in/yr). The shallow recharge equals the sum of the base flow to nonsimulated streams (6.4 in/yr) and the simulated recharge (0.6 in/yr) to the simulated deep ground-water flow regime.

Through simulation, the ground-water flow model quantifies (1) the entry of deep recharge (*dr*), at a rate of about 0.6 in/yr, (2) the loss of deep base flow (*db*), at about 0.5 in/yr, and (3) the residual of these two—called deep seepage (*ds*)—at about 0.1 in/yr. The deep seepage is joined by downward leakage of less than 0.1 in/yr, from updip parts of the Floridan aquifer system and its clastic equivalents. The deep seepage and downward leakage discharge as diffuse upward leakage, totaling less than 0.2 in/yr. The ratio of the estimated shallow recharge (7 in/yr) to the simulated deep recharge (0.6 in/yr) is approximately 10 to 1. This ratio emphasizes the sharp contrast between the relatively large amount of water circulating within the dynamic, shallow flow regime and the relatively small amount circulating within the less vigorous, deeper parts of the confined flow regime.

Figure 16 depicts the long-term water budget for the Southeastern Coastal Plain aquifer system as it is conceptualized to have existed during predevelopment. Considered over the entire Coastal Plain, adjustments within the system in response to ground-water development probably have not caused substantial departures from this conceptualization. Changes since 1900 that have resulted from the pumpage of ground water from the deep ground-water flow regime are discussed in the section "Model Results."

SIMULATION OF GROUND-WATER FLOW

BACKGROUND

The computer model described in this report was developed to enhance the understanding of deep ground-water flow in the Southeastern Coastal Plain aquifer system. The model simulates only the deep ground-water flow regime (fig. 16). The coarseness of the finite-difference grid prevents the model from simulating relatively small-scale conditions of the aquifer system; most of the recharge to the shallow ground-water flow regime discharges as base flow to small streams within drainage basins that occupy less area than the 64 mi^2 represented by a single grid block of the simulation model. By accounting for the net effect of the shallow activity on the deep flow regime, the regional model simulates ground-water conditions in sediments below the level of ground-water flow to small streams that drain the shallow flow regime. The extent of the shallow flow regime is determined largely by topography and drainage. By virtue of the tilted, wedge-shaped configuration of the aquifers and confining units (fig. 15), the shallow flow regime in the Southeastern Coastal Plain exists only in the outcrop area; in addition to including the deeper, confined parts of the outcrop area, the deep ground-water flow regime includes all of the subcrop area. Therefore, the model fully accounts for any ground water that infiltrates below the level of the hydraulic influence of small streams near the outcrop belt and seeps laterally or leaks vertically into subcropping parts of the aquifer system.

The Southeastern Coastal Plain aquifer system was modeled using the U.S. Geological Survey's modular three-dimensional finite-difference ground-water flow model code described by McDonald and Harbaugh (1984). Options in the McDonald-Harbaugh code were used to simulate recharge (to the aquifer outcrop areas), stream-aquifer interaction, and well discharge. The strongly implicit numerical procedure (SIP) was used to solve the finite-difference formulation of the ground-water flow equations.

Once calibrated, the model was used to refine the hydrologic conceptualization and improve the data base for the deep ground-water flow regime of the Southeastern Coastal Plain aquifer system by helping to determine

1. the distributions of ground-water flow and hydraulic head;
2. the amount of ground-water exchange with adjacent regional aquifer systems;
3. a water budget that covers both the outcrop area and the subcropping part of the aquifer system; and
4. the changes in ground-water flow, hydraulic head, and the water budget as a result of ground-water withdrawals since 1900.

The regional model is a realistic, albeit simplified, representation of the Southeastern Coastal Plain aquifer system. By helping to circumvent gaps in the basic data and to compile water budgets, model results have enhanced the understanding of the ground-water flow system. Assuming the model is used in conjunction with field observations and sound hydrologic reasoning, the regional model is a viable learning tool that should provide a basis for future models of finer resolution and greater capability.

MODEL DESIGN

Because the regional hydrogeologic framework (Renken, in press) differentiates between rock sequences that are predominantly aquifer units and those that are predominantly confining units, it served as a template for model layering. Important characteristics of this framework are described below.

HYDROGEOLOGIC FRAMEWORK

The hydrogeologic framework subdivides the aquifer system into aquifer and confining units primarily on the basis of permeability contrasts. The determination of relative permeability was based largely on detailed analyses of lithologic, paleontologic, and geophysical log data from nearly 1,000 oil, gas, and water wells in the region. The aquifer units were extended horizontally and separated vertically from intervening confining units mainly on the basis of the (1) hydraulic interconnection, (2) hydraulic head, and (3) physical continuity of the various strata making up the aquifer system. Because of the regional scale of the study and the need to generalize from site-specific data, it is unavoidable that the aquifer units in places include confining strata, and that the confining units in places include rocks permeable enough to be important sources of water locally.

The complexly interbedded strata that constitute the Southeastern Coastal Plain aquifer system contain numerous aquifers and confining beds. Permeable sequences that appear to be more hydraulically connected than hydraulically isolated are combined into regional aquifer units. The degree of interconnection was judged primarily from the compatibility in hydraulic head among the locally occurring aquifers. Similarly, sequences of confining beds that appear to be physically continuous, or that have the same hydraulic effect, are grouped into regional confining units. The head differences between discrete aquifers within a regional aquifer unit generally are less than the head differences between adjacent regional aquifer units. Regional confining units generally separate the regional aquifer units; however, where an intervening confining unit pinches out, two regional aquifer units may merge. Similarily, two regional confining units may converge where the intervening regional aquifer unit is missing.

The regional aquifers and confining units were subdivided principally on the basis of their hydraulic attributes. Although the physical limits of these units may locally parallel the boundaries of stratigraphic intervals, the hydrogeologic and stratigraphic boundaries do not everywhere coincide. The discord between the regional aquifer and confining units and the stratigraphic units, such as formations, is especially prevalent where there are facies transitions resulting from different depositional environments that existed during a given interval of geologic time. The top or bottom of a regional hydrogeologic unit may cut through a stratigraphic boundary, and stratigraphically equivalent strata may be part of a regional aquifer unit in one place and part of a regional confining unit elsewhere.

As shown in figure 4, the hydrogeologic framework of the Southeastern Coastal Plain aquifer system includes four regional aquifers and three regional confining units (Renken, 1984). Although it may be possible to subdivide them further at a local scale, each of the regional hydrogeologic units is, for the most part, a collection of aquifers or confining units that respond regionally as an hydrologic entity. The regional aquifers consist principally of coarse to fine sand, but locally they may include small amounts of gravel or limestone. The regional confining units are mostly clay, mudstone, or shale; however, thick sequences of chalk make up the Black Warrior River and Chattahoochee River regional confining units in western Alabama and Mississippi.

LAYERING SCHEME

The hydrogeologic framework of the Southeastern Coastal Plain aquifer system is illustrated by numerous diagrams, cross sections, and structure contour and thickness maps that graphically describe the spatial distribution and physical attributes of the regional aquifers and confining units (Renken, in press). To ensure that the regional model would properly represent the aquifer system, the layering scheme for the model was linked to the hydrogeologic framework described by Renken. The model uses a source-or-sink layer at the top (SS) and three aquifer layers (A2, A3, and A4) which, for the most part, are simulated actively. The source-or-sink layer and aquifer layers are separated by three confining layers (C1, C2, and C3). The vertical sequencing of the model layers from west to east (*A–A′*) and from north to south (*B–B′*) is shown in figure 17. The relation between the model and the hydrogeologic units is shown in figure 18. Figures 19–24 show the correspondence between the hydrogeologic framework of Renken (in press) and the layering scheme of the model.

While most of the Southeastern Coastal Plain aquifer system is simulated actively by the model, certain parts of the hydrogeologic framework are either incorporated as boundary conditions or are not included in the model. The model simulates only those parts of the aquifer system that were of principal interest to the Southeastern Coastal Plain RASA study, and the active model layers are linked by boundary conditions to only those parts of the framework that are hydraulically connected to these major areas of interest. The rocks of primary importance to the study are the Cretaceous sediments

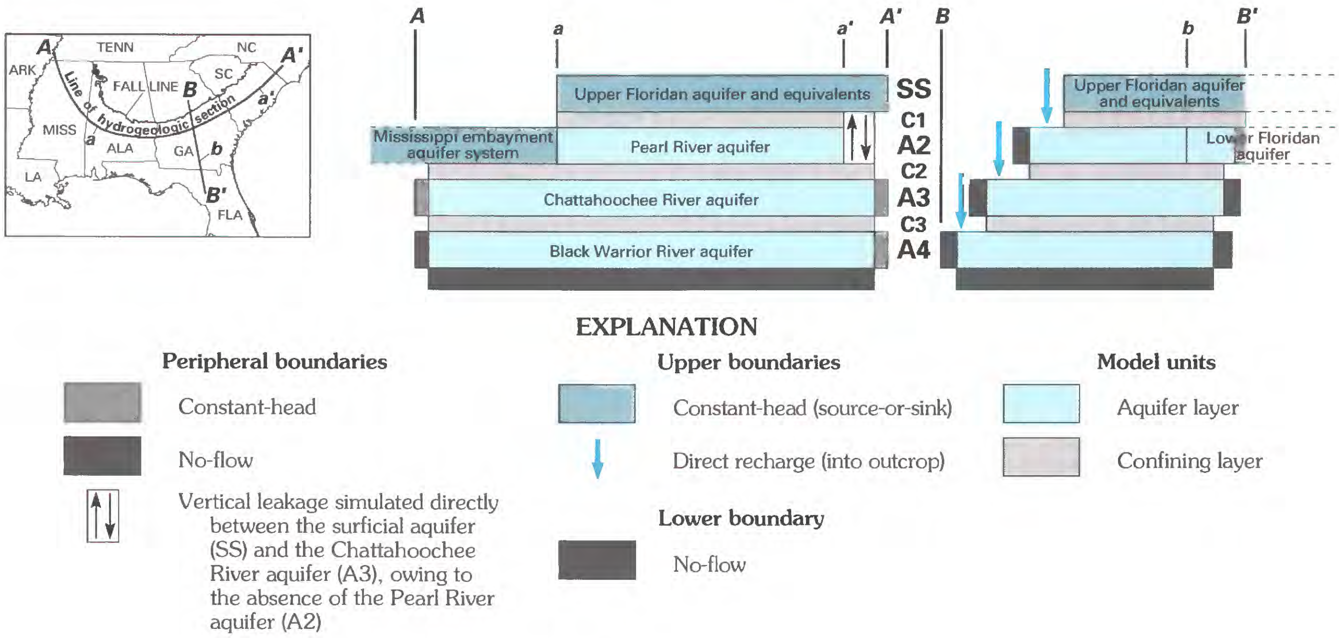


FIGURE 17.—Vertical relation between simulated aquifer and confining units and boundary conditions in the model (modified from Barker, 1986).

that are included in the Chattahoochee River and Black Warrior River aquifers; these aquifers are simulated entirely, as model layers A3 and A4, respectively. The aquifers in Cretaceous rocks represented by model layer A3 are hydraulically connected to the eastern two-thirds of the Pearl River aquifer and locally to the Floridan aquifer system (Miller, 1986). Thick accumulations of chalk and clay within the Black Warrior River confining unit prevent significant flow between the aquifers in Cretaceous rocks of interest and the overlying Pearl River and Chickasawhay River aquifers in Mississippi and southwestern Alabama. The western third of the Pearl River aquifer—equivalent to the Claiborne-Wilcox aquifers of the Mississippi embayment aquifer system of Mississippi and southwestern Alabama (Grubb, 1986a)—were studied and modeled actively by the Gulf Coast RASA study (table 1). Therefore, the Pearl River aquifer was modeled as a source-or-sink boundary condition west of the Alabama River, and the Chickasawhay River aquifer was excluded from the model of the Southeastern Coastal Plain aquifer system.

The eastern two-thirds of the Pearl River aquifer lies between the Chattahoochee River and the Floridan aquifer system. The Floridan aquifer system was modeled previously by the Floridan RASA study (Bush and Johnston, 1988). Because the Upper Floridan aquifer is a very transmissive sequence of limestone within which the flow dynamics have been persistent over time, it was incorporated as a source-or-sink boundary condition within the SS model layer. A shallow water-table aquifer in South Carolina, mapped as the "surficial aquifer" by Renken (1984), is included as an eastern extension of the SS layer because the water table is hydraulically continuous with the potentiometric surface of the Upper Floridan aquifer. The "surficial aquifer" (SS) directly overlies the Chattahoochee River aquifer (A3) where the normally intervening Pearl River aquifer (A2) is absent. Therefore, the model was configured to simulate vertical leakage directly between layers SS and A3 (fig. 17), which is analogous to the way this leakage occurs in nature. In southeastern and south-central Georgia, the clastic rocks of the Pearl River aquifer grade by facies change into carbonate rocks of the Lower Floridan aquifer (fig. 6). To most expeditiously simulate the hydraulic continuity between the sandy facies of the Pearl River aquifer and the carbonate facies of the Lower Floridan aquifer, model layer A2 combines the Lower Floridan aquifer and the Pearl River aquifer where they respond together as a single hydraulic unit characterized by predominantly lateral flow in eastern Alabama, Georgia, and western South Carolina (figs. 19–22).

FINITE-DIFFERENCE GRID AND SUBREGIONAL MODEL COORDINATION

Simulation of the Southeastern Coastal Plain aquifer system was accomplished with one regional model that spans the entire study area and four models of subregional extent that approximately cover, individually, Mississippi, Alabama, Georgia, and South Carolina

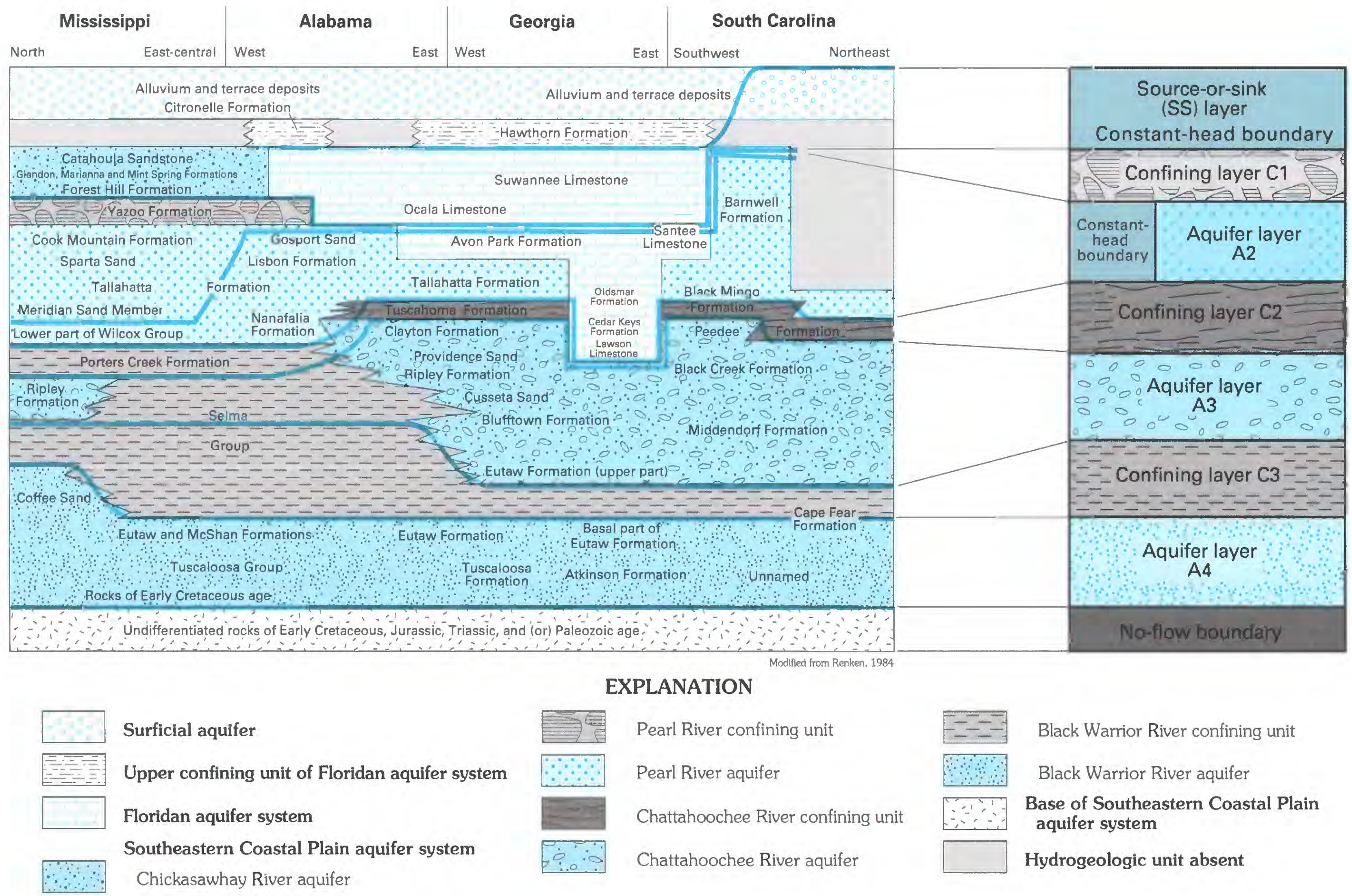


FIGURE 18.—Relation among selected rock-stratigraphic units, the regional hydrogeologic units, and the aquifers and confining units simulated in the model (modified from Miller and Renken, 1988).

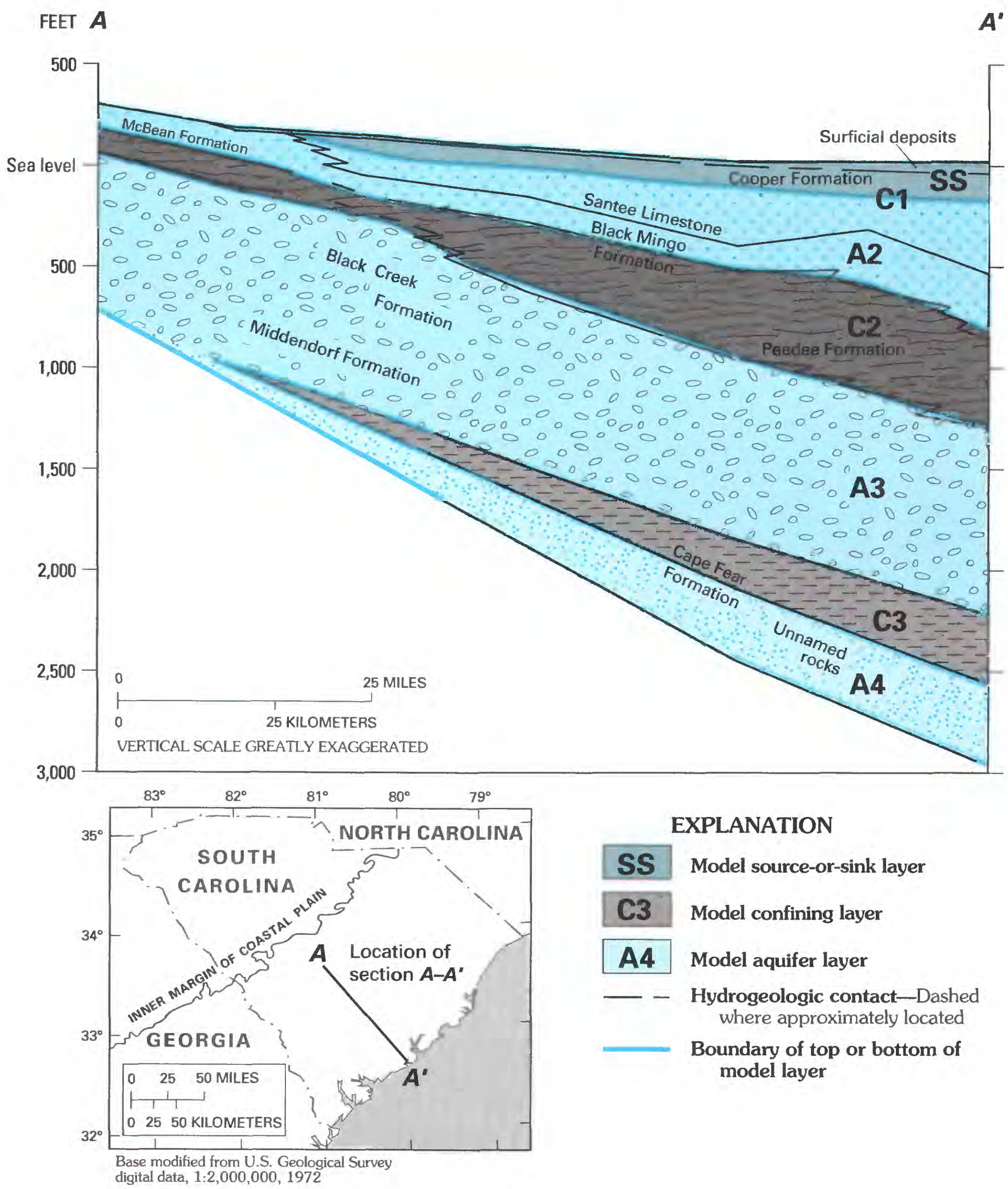


FIGURE 19.—Relation between model units and selected rock-stratigraphic units along section between Orangeburg and Charleston Counties, S.C. The hydrogeologic section (*A–A'*) is from Renken (in press).

(fig. 25). An intermeshing finite-difference grid system was used to coordinate the entry and calibration of model data. The regional grid has 60 rows and 93 columns; each of the 5,580 node blocks measures 8 mi on a side and covers 64 mi^2 (fig. 26). The subregional grids are meshed with the regional grid such that four subregional blocks fit within one regional block. The subregional models are thus based on node blocks that are 4 mi on a side and cover 16 mi^2.

The models adhere to the regional hydrogeologic framework of Renken (in press) and share a common data base in overlapping areas of coverage. The relations among the regional and subregional model units and the local aquifers and confining units are shown in figure 27.

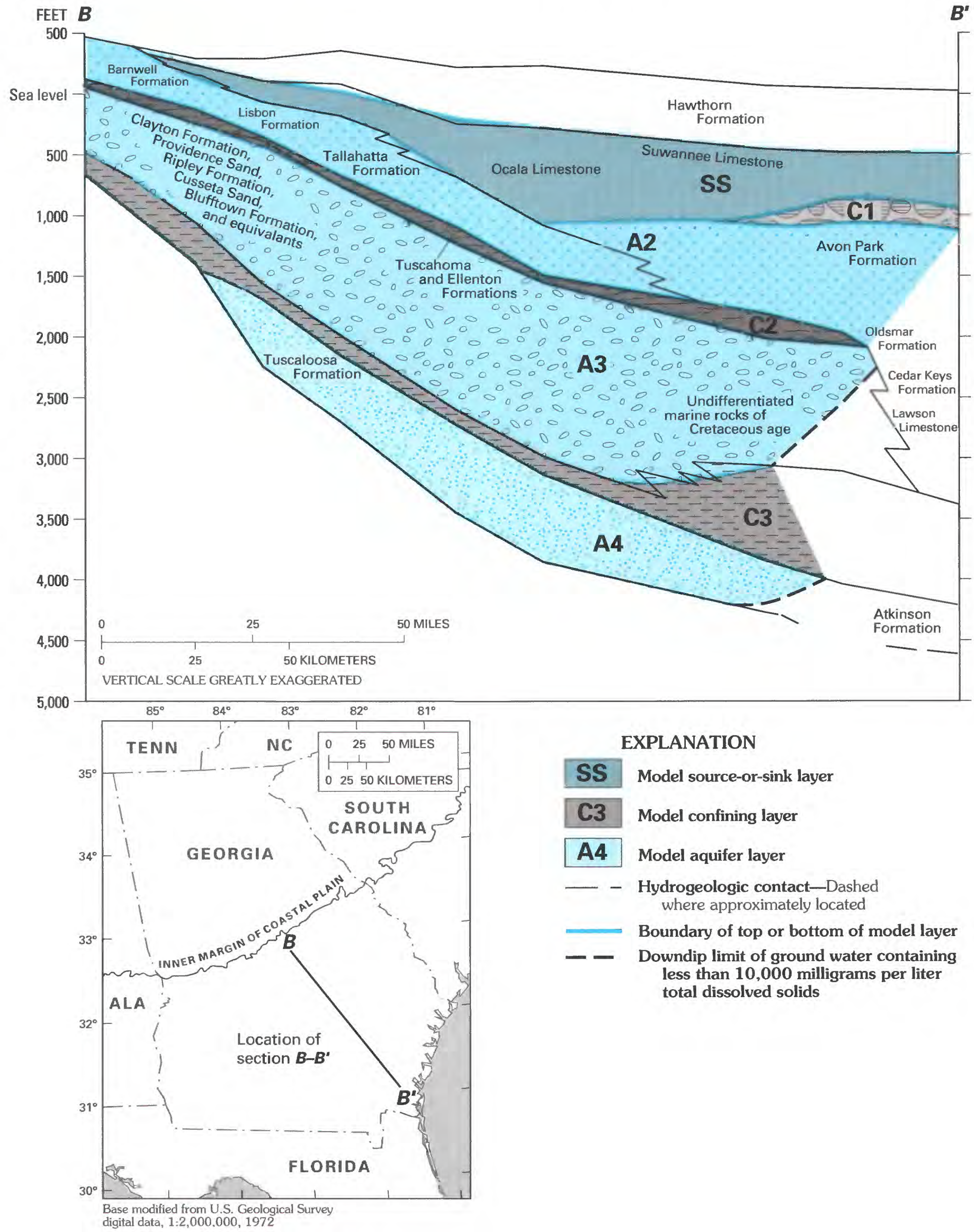


FIGURE 20.—Relation between model units and selected rock-stratigraphic units along section between Wilkinson and Glynn Counties, Ga. The hydrogeologic section (*B-B'*) is from Renken (in press).

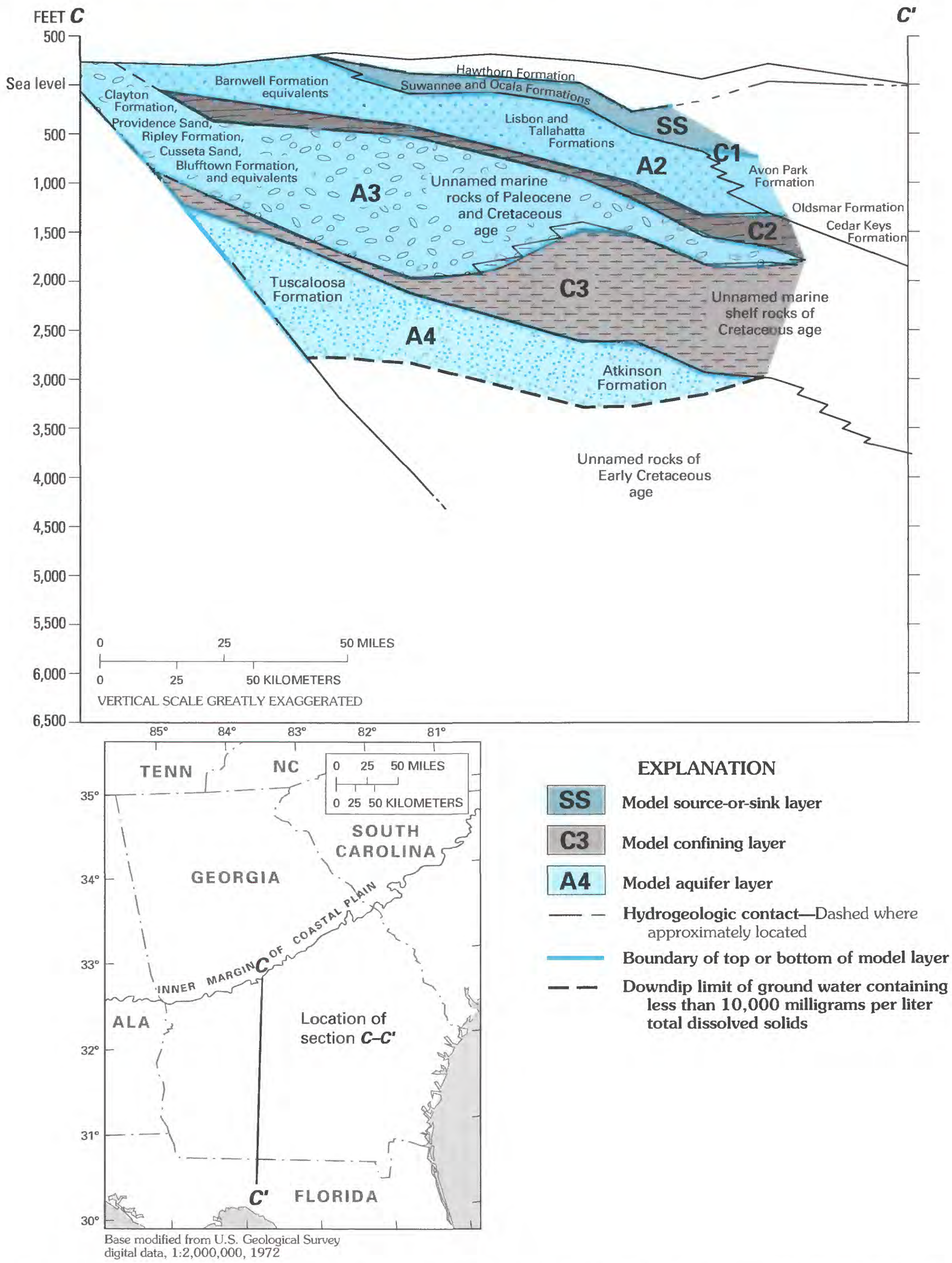


FIGURE 21.—Relation between model units and selected rock-stratigraphic units along section between Bibb County, Ga., and Jefferson County, Fla. The hydrogeologic section (*C–C'*) is from Renken (in press).

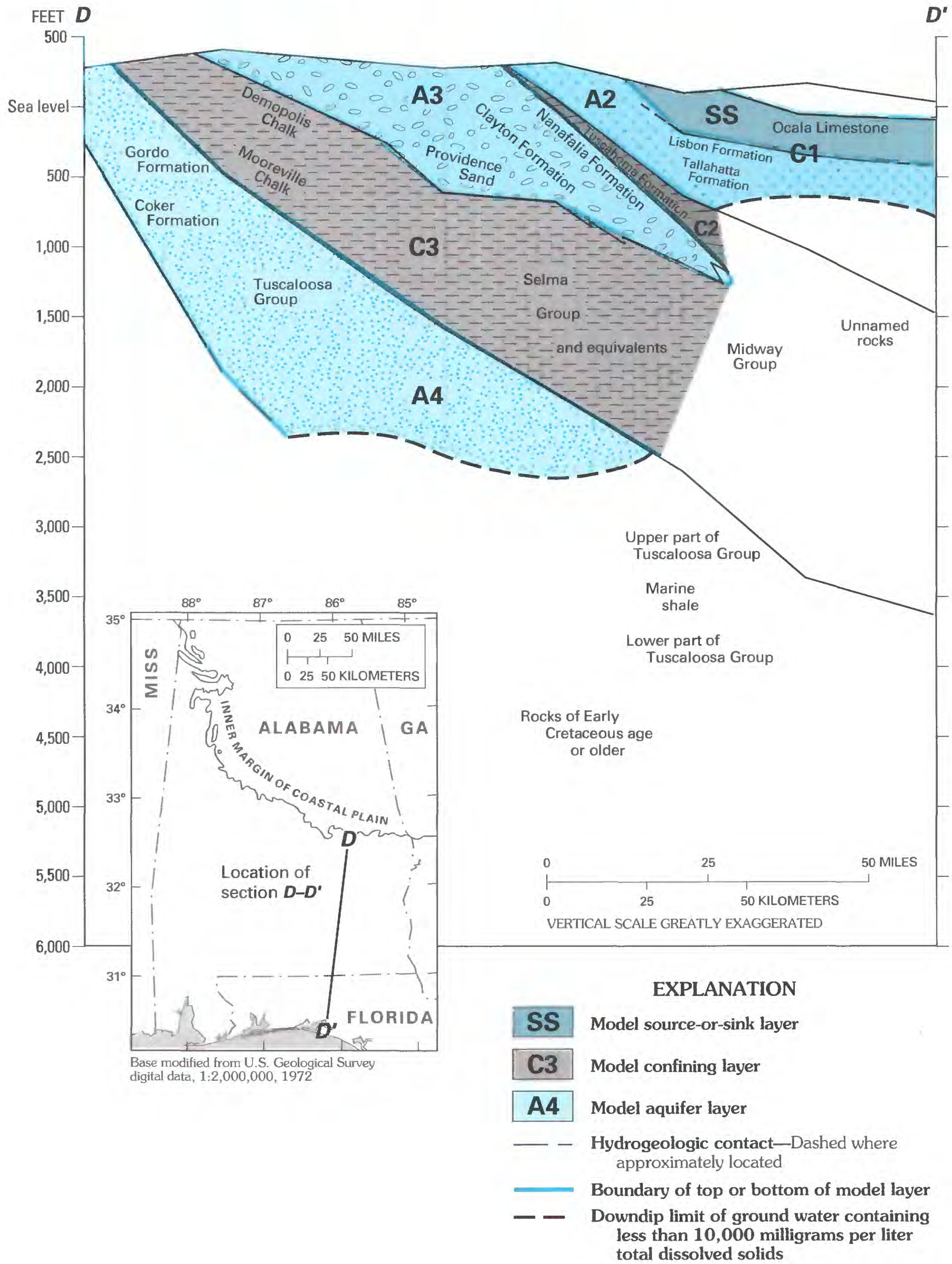


FIGURE 22.—Relation between model units and selected rock-stratigraphic units along section between Macon County, Ala., and Walton County, Fla. The hydrogeologic section (*D–D'*) is from Renken (in press).

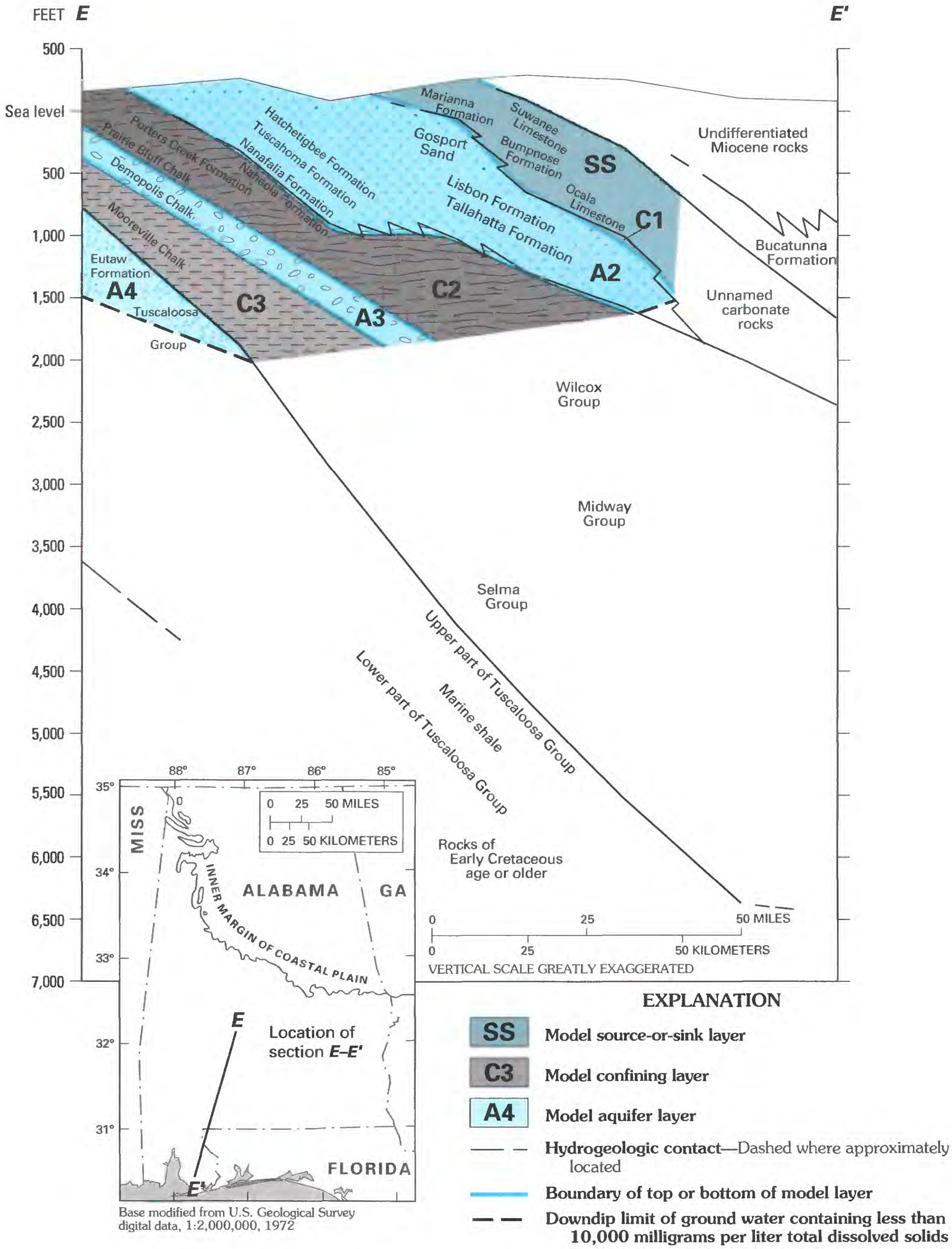


FIGURE 23.—Relation between model units and selected rock-stratigraphic units along section between Dallas and Baldwin Counties, Ala. The hydrogeologic section (*E–E'*) is from Renken (in press).

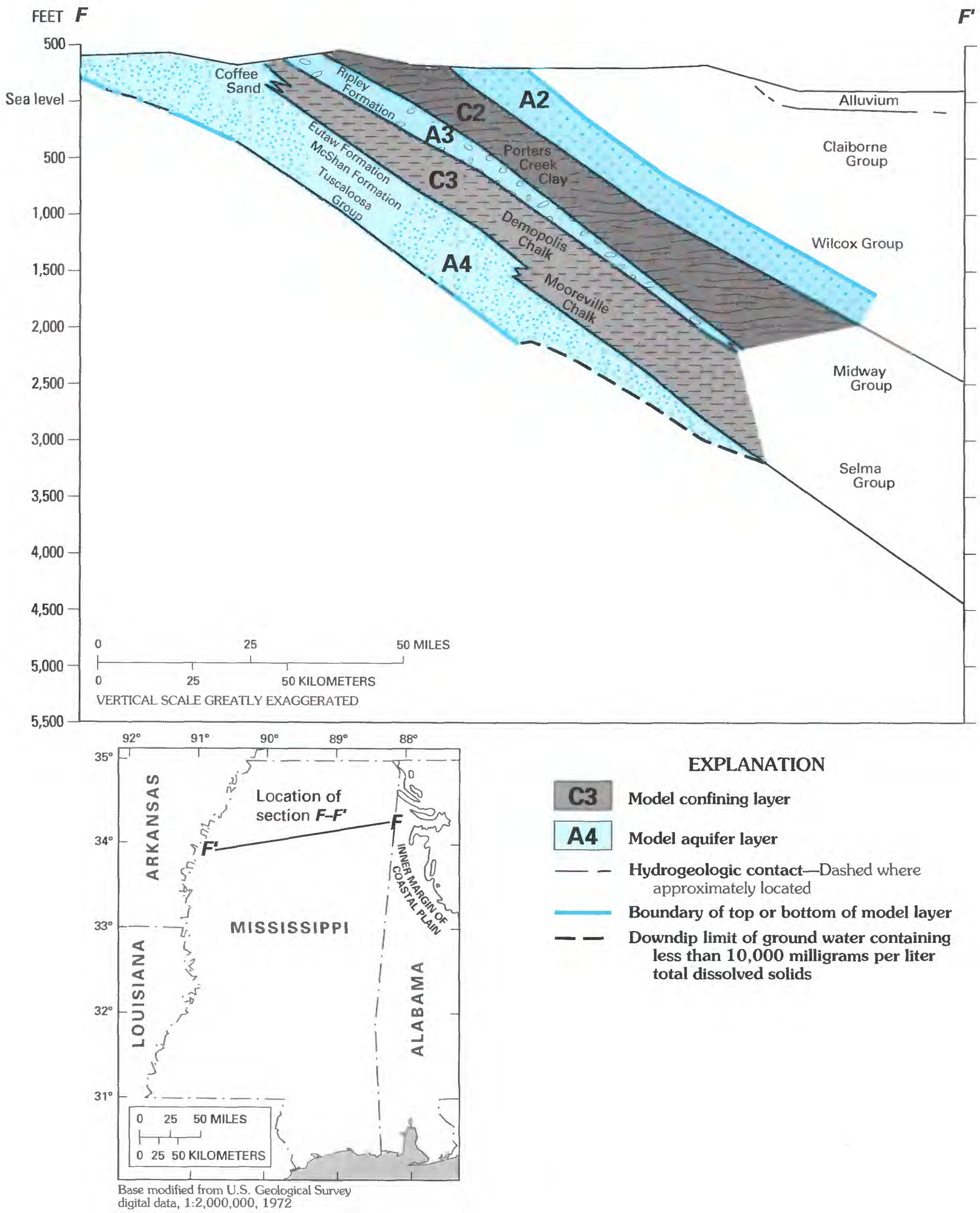


FIGURE 24.—Relation between model units and selected rock-stratigraphic units along section between Itawamba and Bolivar Counties, Miss. The hydrogeologic section (*F–F'*) is from Renken (in press).

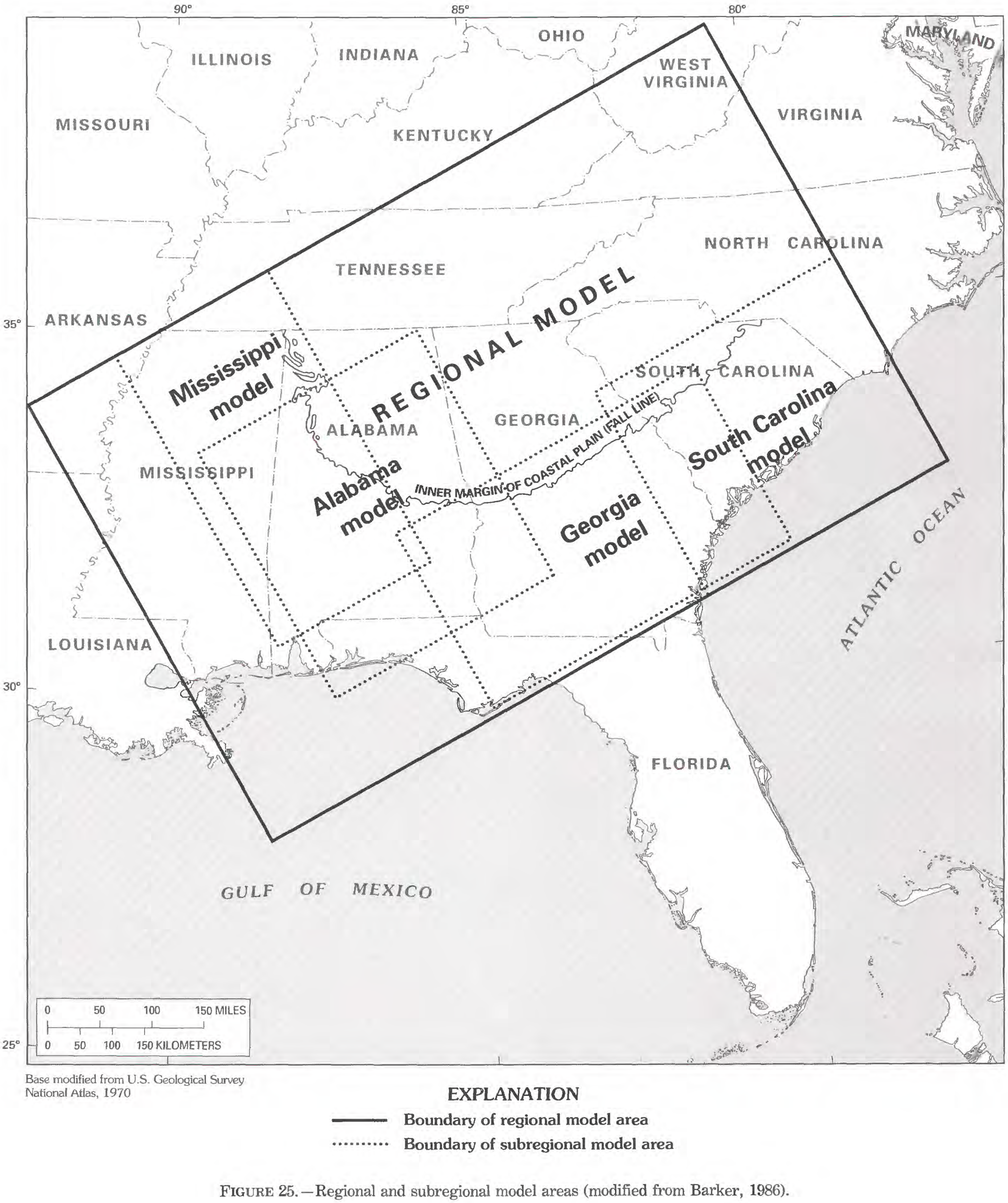
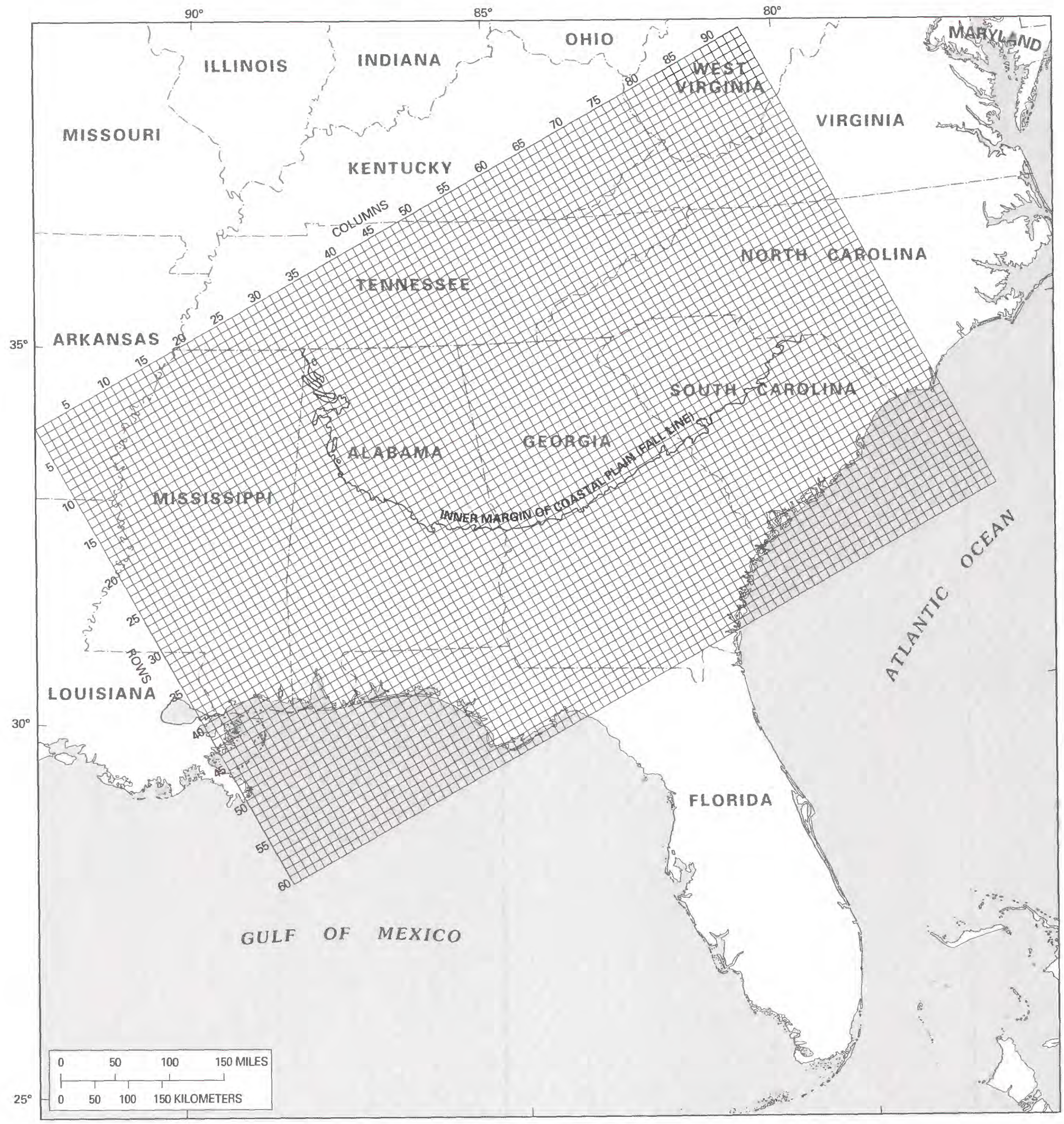


FIGURE 25.—Regional and subregional model areas (modified from Barker, 1986).



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

FIGURE 26.—Finite-difference grid for the regional model.

Regional model layer	Regional hydrogeologic unit	Mississippi model unit	Alabama model unit	Georgia model unit[2]	South Carolina model unit
SS			Upper Floridan aquifer	Upper Floridan aquifer	Surficial aquifer
C1	Pearl River confining unit		Yazoo confining unit	Lisbon-McBean confining unit	Unnamed confining unit
A2	Pearl River aquifer	Lower Wilcox aquifer	Lisbon aquifer	Lower Floridan aquifer Barnwell aquifer Tallahatta-Gordon aquifer Tuscahoma-Black Mingo aquifer	Floridan aquifer system[1] Tertiary sand aquifer
C2	Chattahoochee River confining unit	Unnamed confining unit	Tuscahoma confining unit	Tuscahoma-Fishburne confining unit	Unnamed confining unit
A3	Chattahoochee River aquifer	Ripley aquifer	Nanafalia-Clayton aquifer	Clayton-Ellenton aquifer	Black Creek aquifer
			Prairie Bluff Chalk confining unit	Providence Sand-Peedee confining unit	
			Providence-Ripley aquifer	Providence Sand-Peedee aquifer	
				Ripley-Black Creek confining unit	Unnamed confining unit
				Eutaw-Midville aquifer	Middendorf aquifer
C3	Black Warrior River confining unit	Unnamed confining unit	Selma Chalk confining unit	Eutaw-Cape Fear confining unit	Unnamed confining unit
A4	Black Warrior River aquifer	Eutaw-McShan aquifer	Eutaw aquifer	Upper Atkinson-Upper Tuscaloosa aquifer Lower Atkinson-Lower Tuscaloosa aquifer	Cape Fear aquifer
		Unnamed confining unit	Gordo confining unit		
		Tuscaloosa aquifer system	Tuscaloosa aquifer		

[1]Upper Floridan aquifer absent in South Carolina

[2]Georgia nomenclature partly based on names of laterally equivalent formations that extend into modeled areas in adjacent states

EXPLANATION

Simulated as aquifer layer in subregional model

Simulated as confining layer in subregional model

FIGURE 27.—Relation among the regional and subregional model units and the subregional hydrogeologic nomenclature (adapted from Miller and Renken, 1988).

TABLE 2.—*Input data sets for the regional finite-difference model of the Southeastern Coastal Plain aquifer system*

[*, transient model only]

Description of
Finite-difference grid
Aquifer and confining layers
Active model nodes
* Time steps
Specification of
Boundary conditions
Initial conditions of hydraulic head
Aquifer transmissivity
Confining-unit leakance
Stream-stage elevation
Streambed conductance
Outcrop recharge
*Aquifer storage coefficient
*Municipal, industrial, and irrigation pumpage

To a great extent, the data base of the regional model (table 2) is a composite of data that were compiled at the subregional level and calibrated concurrently in the subregional models. Although the regional model is less detailed than the subregional models with respect to the vertical and horizontal resolution of hydrologic conditions, only the regional model could check boundary conditions for the subregional models and simulate a water budget for the entire study area.

The regional simulation is discussed herein. The State-based models and results are described in chapters of this Professional Paper as follows: Mississippi (Mallory, 1993, chapter G), Alabama (Planert and others, 1993, chapter H), Georgia (Faye and Mayer, in press, chapter F), and South Carolina (W.R. Aucott, in press, chapter E).

BOUNDARY CONDITIONS

Boundary conditions are one of the most important inputs to a simulation model. Owing to data voids and limitations on time and financial resources, the boundary specifications are necessarily generalized. In addition to distinguishing between simulated and nonsimulated areas, the boundary specifications dictate the hydraulic head or ground-water flow (flux) that occurs at the chosen limits of active simulation, which are generally outside the major areas of interest to the study. Because the boundary specifications represent observed or inferred conditions at the limits of the aquifer system, the simulated conditions for the interior parts of the flow system probably are reasonably free of boundary error. This assessment assumes that the model results are used in conjunction with other sources of information and are tempered with the understanding that the model is a learning tool for regional application, rather than a management tool with local application.

Two kinds of boundary conditions are used: specified flux and specified head (fig. 28). Whereas the specified-flux boundaries designate zero (no) flow everywhere they are used, the specified-head boundaries allow simulated flow across them to vary on a node-by-node basis. All no-flow and specified-head boundaries are constant with time.

NO-FLOW BOUNDARIES

No-flow boundaries are used in the model where the circulation of fresh ground water appears to be negligible. No-flow conditions are assumed (1) on the seaward side of the freshwater-saltwater interface, (2) at the updip, downdip, and basal limits of significant permeability, and (3) along a path parallel to the regional flow in the lowermost active layer (A4) near the Mississippi-Tennessee State line (fig. 17).

Whereas hydraulic conductivities of the regional aquifers range from about 1×10^{-6} to about 1×10^{-3} feet per second (ft/s), and perhaps average about 1×10^{-4} ft/s, those in the underlying crystalline rocks appear to be less than about 5×10^{-9} ft/s (Wait and Davis, 1986). Owing to this degree of contrast, no-flow boundaries are used for the base and updip fringes of the aquifer system, along the Fall Line. No-flow boundaries are, likewise, employed along the downdip limits of significant ground-water circulation in model layers A3 and A2, where Renken (in press) has mapped the downdip limits of permeability for the Chattahoochee River and Pearl River aquifers. Owing to a widespread transition from a sandy facies (updip) to clay, marl, shale, or chalk (downdip), the downdip limits of permeability generally correspond to rocks having a hydraulic conductivity of less than about 1×10^{-9} ft/s.

A map showing the distribution of saline water in Cretaceous rocks (Lee and others, 1986) greatly aided the placement of no-flow boundaries in model layer A2 across southern Mississippi and Alabama. Contours on this map depicting the altitude of the shallowest rocks containing water with greater than 10,000 mg/L dissolved solids were used in combination with maps of Renken (in press) describing the spatial distribution of the Pearl River aquifer to approximate the position of the freshwater-saltwater interface in this unit.

The specification of no-flow conditions in the downdip parts of model layer A4 was based on a map of the freshwater-saltwater interface within the Black Warrior River aquifer (Strickland and Mahon, 1986). Because water-quality analyses for the Black Warrior River aquifer are rare, Strickland and Mahon used geophysical logs from more than 150 oil, gas, and water wells in Mississippi, Alabama, and Georgia to calculate dissolved-

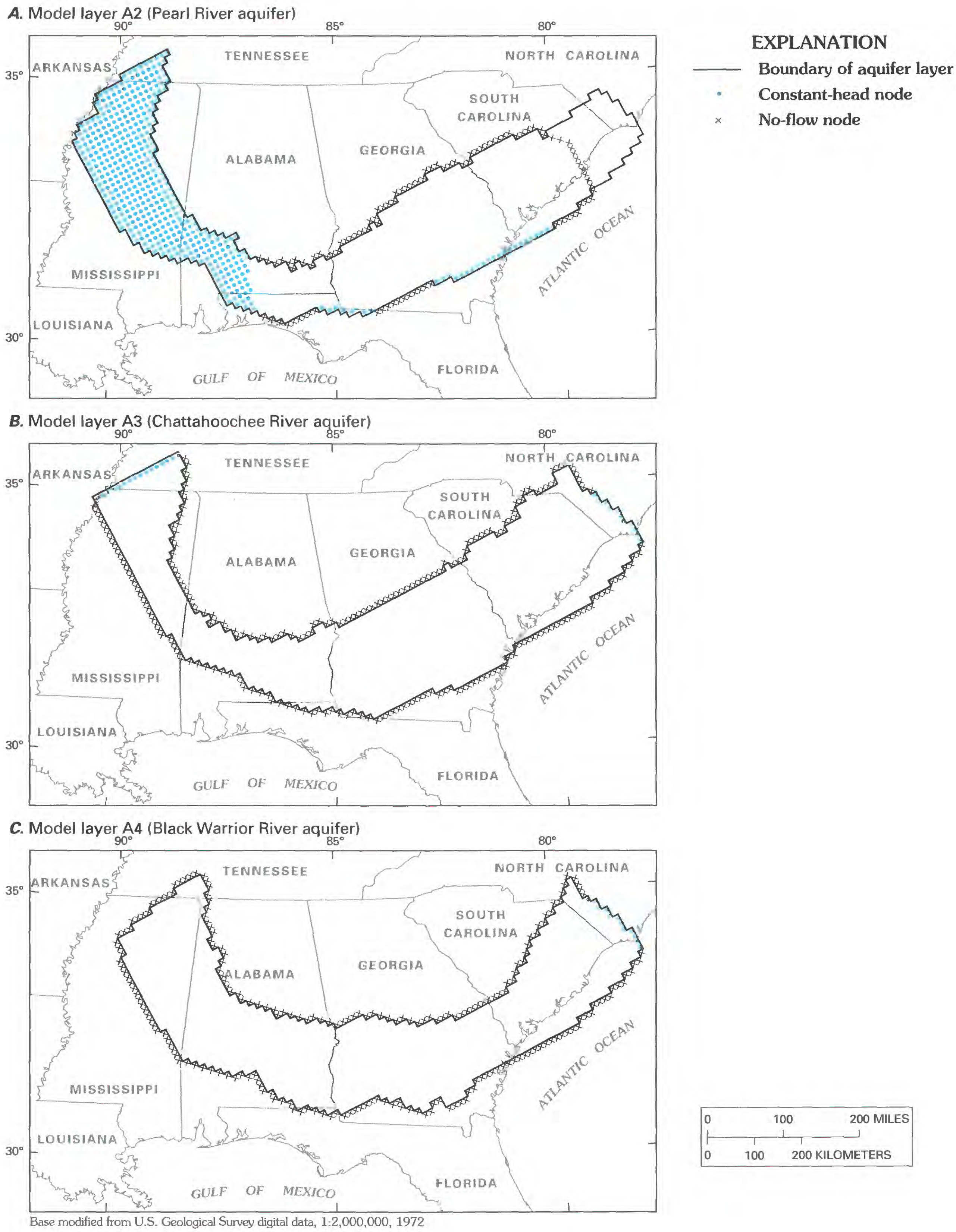


FIGURE 28.—Areal distribution of boundary conditions in active layers of the regional model.

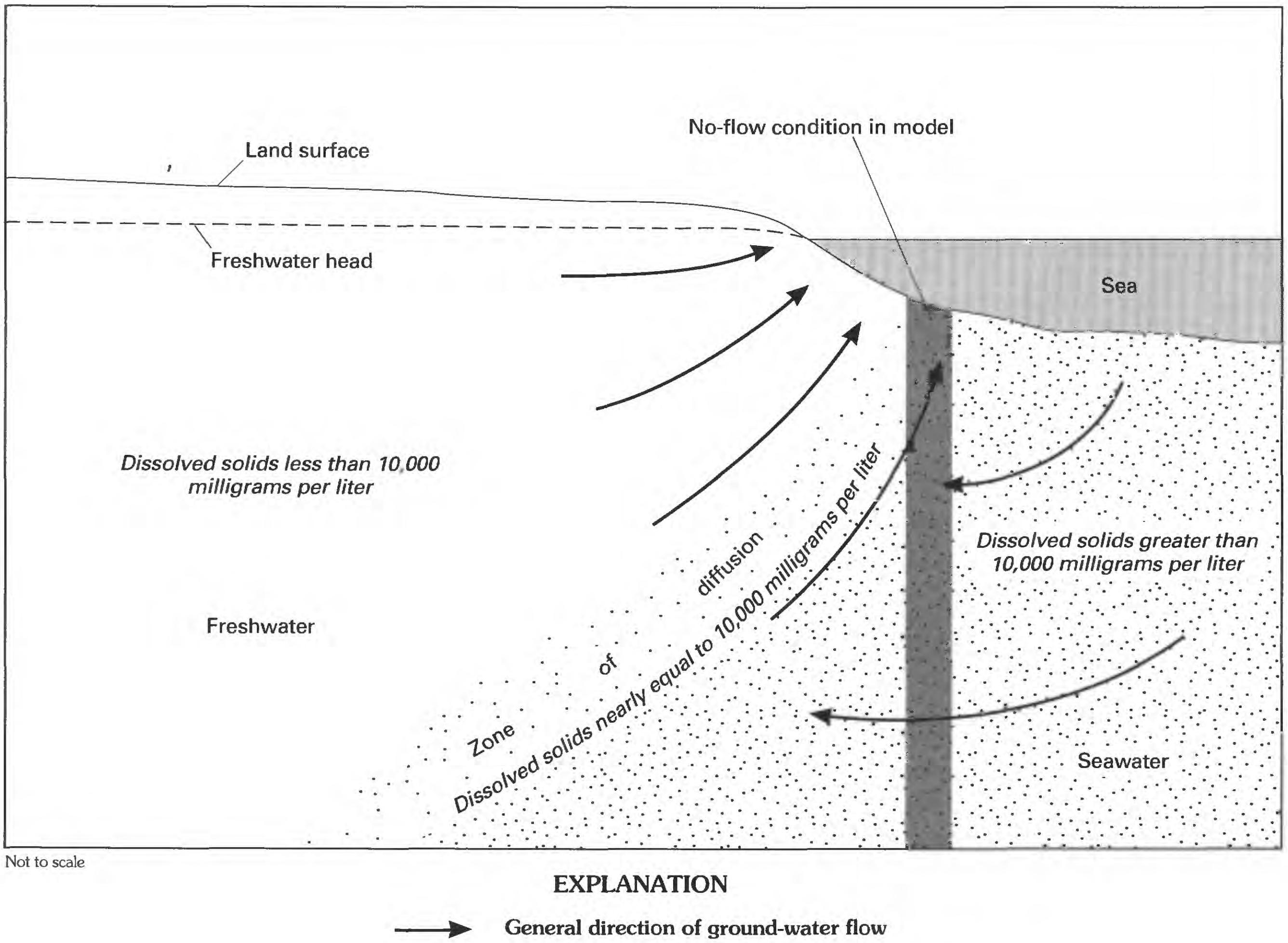


FIGURE 29.—Conceptual relation among concentration of dissolved solids, position of the freshwater-saltwater interface, and placement of the no-flow boundary condition in the model (modified from Cooper and others, 1964).

solids concentrations of formation water from the spontaneous potential curves of electric logs (Wyllie, 1963). They used a dissolved-solids concentration of 10,000 mg/L to delineate the downdip limit of significant freshwater flow, assuming this concentration indicates the threshold of a transition zone between the circulation of freshwater on one side and saltwater on the other. This assumption is supported by the postulation by Ghyben (1889) and Herzberg (1901) that freshwater and saltwater are at equilibrium across a relatively narrow, essentially immobile membrane.

According to Cooper and others (1964), saltwater and freshwater blend across an interval of mechanical dispersion and chemical diffusion. Cooper and others explain that the dispersion of salts is produced by a reciprocative motion of the saltwater front in coastal aquifers that induces a flow of saltwater from the floor of the sea into a zone of diffusion and back to the sea (fig. 29). Rather than being a relatively simple, sharp interface such as that postulated by Ghyben and Herzberg, the shape and location of the freshwater-saltwater transition zone is, perhaps, more realistically a function of permeability and the relative amounts of upgradient recharge and discharge. Because transmissivity values are relatively small near the interface, it is believed that the amount of error associated with the simplification of these conditions in the model is, likewise, small.

As illustrated in figure 29, the placement of the downdip boundary assumes that the vertical components of freshwater flow on the landward side of the boundary (at dissolved-solids concentrations of less than about 10,000 mg/L) outweigh the components of dispersion and diffusion on the seaward side. This assumption is supported by the conclusion of Cooper and others (1964) that the head loss accompanying the landward migration of saltwater tends to lessen the extent to which the saltwater occupies the aquifer. Hubbert's (1940) explanation for moving freshwater on one side of the interface and static saline water on the other also supports the use of no-flow boundaries to simulate conditions near the freshwater-saltwater interface. Because saline water is denser than freshwater, it seeks a lower, less mobile position. If the saline water is prevented from migrating inland, then the fresher ground water should be fairly well contained on the freshwater side of the interface, and there should be no significant net gains or losses of ground water across the interface.

Owing to a lack of offshore well data, the location of the interface off the coast of Georgia and South Carolina is largely unknown. For this reason, the no-flow boundaries used in the model to represent the presumed conditions along much of the Atlantic coastline were extrapolated from Cederstrom and others (1979), Johnston and others (1982), and Bush (1982). Cederstrom and others (1979) deduced that the downdip limit of freshwater in the Tuscaloosa Formation (Black Warrior River aquifer) is either just offshore or coincident with the coastline of South Carolina and is inland everywhere in Georgia. Using the Hubbert (1940) interface equation, Bush (1982) calculated that freshwater flow in the Upper Floridan aquifer extends about 50 to 80 mi off the coast of Georgia. Offshore sampling by Johnston and others (1982) confirmed the existence of a freshwater-saltwater interface within sediments of the Floridan aquifer system at a location 55 mi off the Georgia-Florida coastline.

Constant-Head Boundaries

Where specification of a no-flow boundary was inappropriate, specified heads were used to control the hydraulic gradient—and, thus, the inflow or outflow—near the limits of the simulated aquifer system. Because the specified-head values do not change with simulated time, these boundaries are more appropriately called constant-head boundaries (fig. 28). Constant heads are used in the model where water levels in the aquifer system have remained reasonably stable over time and it was expeditious to limit the areal extent of simulation but neither realistic nor practical to truncate the simulation with no-flow boundaries. Constant heads were used where (1) the limits of the modeled area could not be aligned perpendicularly to equipotential lines on the potentiometric surfaces, or (2) it was advantageous to incorporate the results of recent RASA simulations of adjoining regional aquifer systems, rather than attempt to duplicate them.

Constant heads are used where the first condition discussed above applies to model layer A3 near the Mississippi-Tennessee State line and to layers A4 and A3 near the South Carolina-North Carolina State line (fig. 28). In addition to placing these boundaries where the effect of water-level change due to ground-water development is minimal, they were configured to minimize the amount of simulated cross-boundary flow. The constant-head data used in model layer A3 near the Mississippi-Tennessee State line were acquired from J.V. Brahana and T.O. Mesko (U.S. Geological Survey, written commun., 1986); locally, these data relate to the McNairy-Nacatoch aquifer of southwestern Tennessee and the Ripley aquifer of northern Mississippi (fig. 2). As computed by the model, layer A3 receives about 5 ft^3/s from the northwest—into northern Mississippi, out of southwestern Tennessee. The head data for model layers A4 and A3 near the South Carolina-North Carolina State line were derived from potentiometric maps by Aucott and Speiran (1984a) for sediments that are locally known as the Cape Fear Formation (A4) and the Black Creek and Middendorf aquifers (A3). Model layers A4 and A3, together, discharge about 5 ft^3/s of ground water across this boundary toward the northeast, into North Carolina.

As explained previously, model layer A2 includes both clastic rocks of the Pearl River aquifer and carbonate rocks of the Lower Floridan aquifer (fig. 18). In the extreme southern parts of Georgia, model layer A2 represents only the Lower Floridan aquifer, which continues southward for hundreds of miles. Because the hydraulic condition of the Lower Floridan aquifer is thought to have remained essentially static in this area and the aquifer was previously simulated in its entirety by Bush and Johnston (1988), this unit is simulated in the model of the Southeastern Coastal Plain aquifer system as a constant-head boundary near the Georgia-Florida State line. The constant heads representing the Lower Floridan in model layer A2 (fig. 28) were derived from the model of the Floridan aquifer system to effectively couple the simulation of conditions within the sequence of carbonate rocks that is common to both models. The rates of simulated flow across this and all other constant-head boundaries were monitored closely during model calibration and compared with observed hydraulic gradient and estimated transmissivity data to ensure that observed conditions were being simulated realistically.

In addition to the constant-head boundary along the periphery of the study area, constant-head nodes provide source-or-sink conditions atop actively simulated parts of the Southeastern Coastal Plain aquifer system. These source-or-sink configurations represent adjacent parts of adjoining regional aquifer systems that were studied concurrently or studied previously by other RASA studies.

The rocks composing the Chattahoochee River aquifer undergo a substantial reduction in grain size and permeability between the sandy, relatively permeable facies of South Carolina and Georgia and the finer grained sediments of central Alabama and northern Mississippi

(Renken, in press). As a result, the aquifers in Cretaceous rocks beneath most of Mississippi and western Alabama are separated from the overlying aquifers in Tertiary rocks by more than 500 ft of fine-grained material, mapped as the Black Warrior River confining unit (fig. 4). This substantial thickness of shale, clay, or chalk—locally assigned to the Porters Creek Formation and (or) the Selma Group—inhibits hydraulic interaction between the Chickasawhay River and Pearl River aquifers of Tertiary age and the Chattahoochee River and Black Warrior River aquifers of Cretaceous age.

Because the Cretaceous rocks are of primary interest to the Southeastern Coastal Plain RASA, and because the Tertiary rocks making up the Pearl River aquifer in Mississippi and western Alabama were modeled actively by the Gulf Coast RASA (Williamson and others, 1990), constant heads are used west of the Alabama River in model layer A2 to represent the lowermost parts of the Pearl River aquifer (lower Wilcox aquifer of local usage). This source-or-sink condition is based on a potentiometric map of predevelopment conditions in the Pearl River aquifer (Stricker and others, 1985a). A map of comparatively recent potentiometric conditions in the lower Wilcox aquifer of Mississippi suggests little or no water-level change in or near the outcrop area (Darden, 1986). Downdip, water-level declines have occurred in the lower Wilcox aquifer, but they are west of the downdip limits of the underlying Chattahoochee River and Black Warrior River aquifers (model layers A3 and A4). The hydraulic effects of the thick sequence of shale, clay, and chalk that constitutes the western parts of the Chattahoochee River and Black Warrior River confining units (fig. 18) are simulated in the model using extremely low leakance values in confining units C2 and C3 west of the Alabama River. This input reflects the poor hydraulic connection between aquifers in Tertiary sediments (model layer A2) and in Cretaceous sediments (layers A3 and A4) in western Alabama and Mississippi. As a result, the sum of the simulated vertical leakage (either up or down) is less than 10 ft^3/s between the Mississippi embayment aquifer system (studied by the Gulf Coast RASA) and the Southeastern Coastal Plain aquifer system.

Compared with widespread, long-term water-level declines in the Southeastern Coastal Plain aquifer system, water levels in the Upper Floridan aquifer have been relatively stable. Because hydraulic conditions in the Upper Floridan appear unlikely to change significantly in the near future, water levels from this aquifer were used as a source-or-sink boundary condition atop the model of the Southeastern Coastal Plain aquifer system. Studied previously under the RASA program, the Upper Floridan aquifer has been mapped by Miller (1986) and modeled by Bush and Johnston (1988). These investigators characterize flow in the Upper Floridan aquifer as "dynamic" and "vigorous." Pumping from the highly permeable Upper Floridan aquifer has not, for the most part, significantly affected the general characteristics of the natural flow system of the Floridan or of the underlying Southeastern Coastal Plain aquifer system. Response to changes in the distribution of pumping typically dissipates within days or weeks in most areas. The Floridan aquifer system, according to Bush and Johnston (1986, p. 22), is considered to be approximately at equilibrium except during short periods following sustained increases in pumpage. An exception to the general stability of the Upper Floridan aquifer occurs near Savannah, Ga., where pumping began in the 1880's and hydraulic head declines of more than 100 ft have resulted from large withdrawals of ground water. The Upper Floridan aquifer in the Savannah area is separated from the underlying Chattahoochee River aquifer by more than 200 ft of calcareous and evaporitic rocks of low permeability (Miller, 1982; Renken, in press). Therefore, neither the ground-water withdrawals from the Floridan nor the resulting water-level declines are expected to have affected the hydrology of the underlying aquifers in Cretaceous sediments (J.A. Miller and R.A. Renken, U.S. Geological Survey, pers. commun., 1982). Thus, the predevelopment potentiometric surface for the Upper Floridan aquifer (Bush and Johnston, 1988, pl. 4) was adapted for model layer SS as a source-or-sink boundary condition in the regional model of the Southeastern Coastal Plain aquifer system (fig. 17). Where the carbonate rocks of the Upper Floridan aquifer pinch out in South Carolina, hydraulic heads from a laterally equivalent, clastic, water-table aquifer (W.R. Aucott, U.S. Geological Survey, written commun., 1990) are used to extend the source-or-sink boundary condition to the eastern limit of the study area.

MODEL CALIBRATION

STRATEGY

Before the results, or output data, of the computer model of the Southeastern Coastal Plain aquifer system could be deemed acceptable, the input data (table 2) had to be calibrated. Calibration was largely a trial-and-error process in which the input data were modified in response to shortcomings in the model, as determined by the importance of differences between the simulated conditions and the observed (or inferred) conditions. The basic goal of calibration was to obtain a model that could simulate actual hydrologic conditions within acceptable limits of error. The data used as calibration standards were based on field observations, as well as on the previously discussed conceptual model of the ground-water flow system. The results of calibration were used to reevaluate and improve the conceptual model of the system, as well as compile a water budget for the deep ground-water flow regime.

Numerous sets of input data were required to simulate the Southeastern Coastal Plain aquifer system. The accuracy of those data determined the reliability of the simulated conditions. In turn, the accuracy of the input data was strongly influenced by the availability and validity of control data, which for the most part consisted of field observations made during previous hydrogeologic investigations. Owing to the general sparseness of control data, especially for middip and downdip areas, the accuracy of the initial input data was highly variable. Numerous generalizations and assumptions were necessary to fulfill the input needs of the model. More than half the nodal values of input data had to be extrapolated from more-or-less qualitative sources, such as the conceptual model, without the aid of specific field data.

The overall strategy of model calibration was to (1) delineate a plausible set of boundary conditions, (2) initiate simulation using preliminary estimates of recharge, transmissivity, leakance, riverbed conductance, and storage coefficient, and (3) refine the original parameter estimates by trial-and-error simulation until the model's output data satisfied the calibration criteria. The calibrated model generally reflects calibration priorities, which were defined by the scale of the study, the density and integrity of the control data, and the objectives of the simulation. Ordinarily, it might have been advantageous to use a parameter estimation technique (Cooley, 1977); however, the lack of an areally balanced distribution of control data precluded this. A large number of interrelated factors affected the output of the model, causing the calibration criteria—or the standards of calibration—to be highly subjective. In consideration of the regional perspective of the study and the coarse scale of the model, the calibration tolerances with respect to head match were expanded where the control data were sparse or nonexistent to minimize the time spent attempting to simulate parts of the flow regime that will not be properly defined until more data become available. It is not possible to know how good or how poor the calibration is for areas for which few or no control data are available. However, it is important to point out that the model may not simulate conditions for these parts of the deep flow regime as well as it simulates conditions for areas for which there is control. Although it was expected that the simulated hydraulic heads should generally match observed counterparts by ±50 ft, the head mismatches may exceed 50 ft in areas of the system that are not well defined.

Konikow (1978) suggested that the amount of adjustment in any model input parameter should generally be directly proportional to the uncertainty of its value. Accordingly, most of the calibration effort was spent refining the input parameters that were based on the least amount of control data, especially where the output data of the model were reasonably sensitive to changes in those parameters. Because the model was not expected to represent the ground-water flow system exactly, relatively simple areawide and time-constant distributions of model input data were used until such distributions were proved inadequate or insufficiently detailed. Rather than speculate on the spatial or temporal variation of a given parameter where the control data were too sparse to define it properly, it was generally more expedient to assume a blanket distribution initially and let the model results determine the need for refinement.

Despite the shortcomings of the control data, the simplistic extrapolations of data input, and the trial-and-error approach to calibration, an attempt was made throughout the process of model development to achieve a physically meaningful characterization of the flow system. Just as it was unwise to overlook important details in areas where the control data were plentiful, it was considered equally imprudent to exclude from the simulation areas that were not well understood owing to data deficiencies. It was hoped that calibration would help fill the data gaps. In addition to simplifying the initial data input, an effort was made to keep the ensuing adjustments within the limits of sound hydrologic judgment and geologic principles. The difficulty of obtaining an acceptable level of calibration generally increases with the complexity of the simulation model. While attempting to avoid the pitfalls of an overly complex model, a course for calibration was sought that would result in an improved data base as well as a model capable of enhancing the understanding of the regional part of the ground-water flow system.

The model was calibrated for both steady-state and transient conditions. The steady-state model was developed first, and provided the initial conditions for the transient simulations. The transient model, a functional extension of the steady-state model, represents the stresses of pumpage and considers the effects of time and ground-water storage. Although the steady-state and transient versions of the model contain much of the same program logic and input data, the individual versions depict different sets of conditions, thus requiring adjustment of different data sets during the calibration process. The transmissivity, leakance, riverbed conductance, and recharge data were calibrated during steady-state runs; the storage coefficient data were calibrated during transient runs. Because both the boundary conditions and the pumpage were considered known components, neither of these data sets was modified as part of the calibration process. The validity of transient results proved to be very dependent on the distribution of recharge and vertical leakage, which were functions of the steady-state calibration. Consequently, during the latter stages of calibration, steady-state runs were alternated with transient runs, and adjustments were made to the appropriate data set(s) in the appropriate model so that the calibration of each model appeared to improve from each change.

The steady-state model was calibrated to simulate conditions in the deep ground-water flow regime prior to about 1900, before the beginning of significant pumpage. Before 1900, the aquifer system was for the most part in its natural, predeveloped state. In such a state, recharge was approximately equal to discharge, water levels were essentially stable, and there were no significant changes of ground water in storage.

The steady-state model explicitly depicts a state of hydraulic equilibrium. Solution of the steady-state flow equation requires that recharge equal discharge and that the boundary conditions and stresses do not change with time. The steady-state condition is a relatively simple mathematical concept, one that (owing to the complexity of physical systems) may never exist in the real system. The condition can be approximated for simulation purposes, however, if the historic observations on which the model is based represent the average of actual conditions over a long period of time. If the hydrologic system undergoes uniform and cyclic changes (such as seasonal fluctuations in precipitation and evapotranspiration), then the average of a resulting hydrologic response (such as the decline and recovery of hydraulic head) can define a steady-state condition for modeling purposes. In general, the longer the period of hydrologic observation and the greater the number of control points, the better the definition of steady-state conditions used for calibration.

Owing to a rather sketchy definition of steady-state conditions on which to base calibration of the steady-state model of the Southeastern Coastal Plain aquifer system, it was assumed that the model simulates the average of long-term, equilibrium conditions that are inferred to have existed prior to about 1900. The steady-state model was calibrated principally against potentiometric surfaces drawn, independently of the modeling exercise, to represent average predevelopment conditions, based on water-level observations dating from about 1900 to about 1950 (Stricker and others, 1985a, 1985b, 1985c). The scarcity of contemporaneous head data necessitated the use of data from such a long time span. Data specific to the late 1800's and early 1900's are virtually nonexistent. Although it was desirable to use only control data that are consistent with the predevelopment status of the aquifer system, some of the data used to construct the potentiometric maps probably do not represent equilibrium, or steady-state conditions. Some observations doubtless were affected by pumping, while others probably were influenced by abnormal natural conditions such as drought or flooding. For most downdip areas, potentiometric data for calibration purposes had to be extrapolated from Stricker and others (1985a, 1985b, 1985c), using only the conceptual model as a guide. In consideration of the deficiencies in and discrepancies among the control data, the steady-state model could not be expected to everywhere match the potentiometric surfaces constructed from the available water-level data. Where the potentiometric data were inferred from a rough conceptualization of the flow system, the output from the calibrated model may be more accurate than the original extrapolations.

The transient model was calibrated to simulate the response of the deep ground-water flow regime to the withdrawal of ground water through industrial, irrigation, and public supply wells. Between about 1900 and 1985, the calibration period for the transient model, water levels in some areas have declined more than 150 ft. As a result of the pumpage and the ensuing water-level decline, the distribution of ground-water flow has changed significantly in some places. The transient model was calibrated primarily against hydrographs drawn from water-level measurements made since the early 1900's on the premise that if the model could be calibrated to replicate long-term patterns of head change, then it would inherently simulate the important changes in the distribution of flow.

GOODNESS OF FIT

Goodness-of-fit tests were made at the conclusion of each calibration run to monitor the progress of calibration and to check the model's response to parameter adjustment. Statistical accounts of the difference between the observed (or inferred) and simulated data were used as a guide in determining the direction and magnitude of subsequent changes in the model input. The distribution of the differences between simulated and observed hydraulic heads provided the basis for eventually accepting the model as being calibrated. Whereas visual differences between actual and simulated hydrographs proved to be the best means of monitoring the transient calibration, the statistical tests were more useful for evaluating the steady-state calibration.

Most of the goodness-of-fit compilations that were used to monitor model development were based on two statistical algorithms, known as the mean absolute head residual (*MAHR*) and the root mean squared error (*RMSE*). The compilations were based on a node-by-node assessment of simulated versus observed (or inferred) hydraulic head over each aquifer layer in the model. The compilations were made as follows:

$$MAHR=\frac{\sum_{i=1}^{N} \mid h^s-h^o \mid}{N} \qquad (4)$$

and

$$RMSE=\sqrt{\frac{\sum_{i=1}^{N}[h^s-h^o]^2}{N}}, \qquad (5)$$

where

N = number of active nodes in aquifer layer;
h^s = simulated head; and
h^o = observed (or inferred) head.

The mean absolute head residual (*MAHR*) is the average of the absolute differences between the model-simulated and observed (or inferred) hydraulic heads at all nonboundary nodes making up an aquifer layer in the model. The root mean squared error (*RMSE*) is a measure of the spread, or disparity, between the observed (or inferred) and simulated hydraulic heads. The RMSE is analogous to the standard deviation of the disparity (Sippl and Sippl, 1981), and is sometimes termed the "standard error of the estimate" for modeling applications (Konikow, 1978). Assuming a normal distribution of the disparity between simulated and observed values, roughly two-thirds of the sampled model nodes should have a simulated head that differs from the observed (or inferred) head by no more than the indicated *RMSE*. The *RMSE* provides a more stringent test of calibration than does the *MAHR* because the *RMSE* places more emphasis on larger deviations, owing to the squaring process in equation 5.

The distribution of the *MAHR* and the *RMSE* in the calibrated steady-state model of the Southeastern Coastal Plain aquifer system is shown in table 3. The best fit of hydraulic head is for aquifer layer A2 (*MAHR* of 17 ft), and the worst is for layer A3 (*MAHR* of 39 ft). Aquifer layer A4 shows a *MAHR* of 28 feet. Over the entire model, the *MAHR* is 30 ft and the *RMSE* is 43 ft.

The results of the steady-state calibration appear to be least "good" for model layer A3 (Chattahoochee River aquifer). This model layer consists of several aquifers of varying extent that were simulated as a single aquifer layer in the model. Composite head data were used to construct the potentiometric map for this model layer. If enough data were available to separate the various contributing aquifers into a greater number of model layers, the actual vertical head distribution within the

TABLE 3.—*Distribution of the mean absolute head residual and root mean square error in the calibrated steady-state model of the Southeastern Coastal Plain aquifer system*

Model Layer	Mean absolute head residual, in feet	Root mean square error, in feet
A2	17.4	24.0
A3	39.0	49.7
A4	28.0	42.8
Overall	30.1	43.4

Chattahoochee River aquifer perhaps could have been simulated more accurately.

A better comparison of simulated versus observed heads results for model layer A2 (Pearl River aquifer) because the control data were substantially better and the simulated heads are closely tied to the source-or-sink boundary condition representing the Upper Floridan aquifer. Much of model layer A2 consists of the Lower Floridan aquifer; water-level data for the Upper and Lower Floridan aquifers indicate a high degree of head correspondence between these aquifers (Bush and Johnston, 1988).

Although model layer A4 represents an aquifer deeper than the one simulated as layer A3 and is supported by less control data, the calibration appears better for layer A4. This probably is because the lateral head gradients are generally less in layer A4 than in A3, owing to the fact that the aquifer represented by layer A4 has a limited outcrop area and interacts to a much lesser extent with the surface-water network and the shallow ground-water flow regime. The relatively coarse model grid blocks of 8 mi on a side limit the ability of the model to simulate the steeper head gradients common to outcrop areas, especially near major stream channels.

Although the transient calibration was not evaluated on the basis of *MAHR* and *RMSE* calculations, inspection of the simulated versus observed hydrograph data (pls. 5–7) suggests that the transient model, for the most part, simulates head data that compare to within about 25 ft of those measured between 1900 and 1985. The greatest deviations generally occur near the most heavily pumped areas and near the end of the calibration period. For the most part, the model simulates water levels that are shallower than those measured. The most significant discrepancies probably are due to the limited ability of the model to simulate the steeper hydraulic gradients that occur near pumping wells. This problem is analogous to the shortcomings of the model with respect to simulating aquifer-to-river gradients. The limited resolution of the model stems directly from its relatively large time steps and grid blocks. The model-simulated hydraulic heads represent relatively long term conditions over 64-mi^2 areas, whereas the field-measured heads may include short-term, local effects of pumping. Owing to these considerations, as well as to the fact that the model is intended to be no more than a learning tool, the discrepancies between observed and simulated conditions are not considered large enough to significantly affect the intended application of the model. Achieving a significantly better calibration would require a model with smaller grid blocks and more accurate input data.

PARAMETER DEFINITION

Transmissivity

The definition of the regional distribution of transmissivity within the Southeastern Coastal Plain aquifer system has evolved through various stages, beginning with the compilation of site-specific data for individual wells or pumping centers and ending with the calibration of a data base for simulation. Prior to the RASA program, transmissivity values had not been mapped on a regional scale. Certainly, no effort had been made to evaluate the distribution of transmissivity with respect to the hydrogeologic-framework units delineated by Renken (1984). The current understanding of the regional distribution of transmissivity is the result of much fieldwork during past investigations, as well as of considerable data compilation, extrapolation, and trial-and-error simulation during the study described here.

First, aquifer-test results, specific-capacity observations, and aquifer-diffusivity calculations were sorted for pertinent information and checked or analyzed as necessary. The representative data are for the most part documented in the following publications: Newcome (1971), Stricker (1983), Faye and McFadden (1986), and Aucott and Newcome (1986). Second, the original transmissivity calculations were corrected for the effects of partial penetration with respect to the regional aquifer units of the hydrogeologic framework (Renken, 1984). Third, a tentative distribution of transmissivity was interpolated or extrapolated for all simulated parts of the aquifer system. Finally, regional patterns of transmissivity were defined on the basis of model calibration, with the initial input data being adjusted to minimize the differences between model output and observed (or deduced) field conditions.

Where aquifer-test data were available, they were given the most weight in assigning the initial, as well as the final (calibrated), values of model transmissivity. Most of the aquifer-test data from multiple-well tests were analyzed using either the Theis (1935) nonsteady equation for nonleaky confined aquifers or the Hantush-Jacob (Hantush and Jacob, 1955) method for leaky confined aquifers. Data from single-well tests generally were analyzed using the modified nonequilibrium formula of Cooper and Jacob (1946) or the Theis recovery method (Ferris and others, 1962, p. 100). As a rule, aquifer tests involving multiple wells provided more reliable data than did single-well tests.

Specific-capacity data were also used to estimate initial transmissivity input for the model. The derivations of transmissivity from specific-capacity data were based on the modified nonequilibrium formula as applied by Faye and McFadden (1986, p. 15), and on the Theis equation (Ferris and others, 1962, p. 91). Results from the Theis equation were corrected with the Jacob (1950) modification of the Theis (1935) nonsteady equation to account for the effects of the storage and time. The principal relations are

$$T = \underbrace{\left(\frac{Q}{s}\right)}_{\text{specific capacity}} 2.3 \log\left(\frac{r^e}{r^w}\right) \tag{6}$$

where

$$r^e = \sqrt{\frac{2.25\ T'\ t}{S'}} \quad \text{and} \tag{7}$$

T = transmissivity (in feet squared per second);
Q = discharge of pumped well (in cubic feet per second);
s = drawdown at pumped well (in feet);
r^e = distance from pumped well to equilibrium head (in feet);
r^w = effective well radius (in feet);
T' = estimate of areal transmissivity (in feet squared per second);
t = time of pumping (in seconds); and
S' = estimate of areal storage coefficient (dimensionless).

This procedure, which has been explained in detail by Bedinger and Emmett (1963), requires areal (order of magnitude) estimates of transmissivity (T') and storage coefficient (S') as well as reasonable estimates of effective well radius (r^w) and time of pumping (t). The computed transmissivity (T) is relatively insensitive to the assumed values of T', S', t, and r^w because the log is taken of the (r^e/r^w) term before it is multiplied by specific capacity. For the Southeastern Coastal Plain aquifer system, areal estimates of 0.1 foot squared per second (ft^2/s) and 0.0005 were assumed for transmissivity and storage coefficient, respectively. When specific values of well radius (r^w) or time (t) were not known, values of 1 ft and 1 day, respectively, were assumed.

Nearly 225 transmissivity values were available from aquifer-test data for the simulated parts of the aquifer system. Of these values, 25 percent were for Georgia, 35 percent each were for Mississippi and South Carolina, and 5 percent were for Alabama. More than half the aquifer-test data were for model layer A3 in Georgia and South Carolina. Although Mississippi had 75 transmissivity values for model layer A4 alone, Alabama, Georgia, and South Carolina combined had fewer than 10 values for layer A4.

The need for additional transmissivity data was partially met with the estimation of transmissivity from specific-capacity data. Of the nearly 300 specific-capacity values judged to be representative, about 45 percent came from Alabama, about 40 percent from Georgia, and about 15 percent from South Carolina. As with the aquifer-test data, the majority (70 percent) of the specific-capacity data were for model layer A3.

Thirteen additional estimates of transmissivity were obtained for modeling purposes from calculations of hydraulic diffusivity (Stricker, 1983, p. 12). The diffusivity data were calculated from the slope of base-flow recession curves and basin geometry using the method of Rorabaugh (1960). The diffusivity values yielded three additional estimates of transmissivity for model layer A2, six for layer A3, and four for layer A4.

After all the transmissivity data from all sources were consolidated, the precalibration definition of transmissivity probably was best for model layer A3. The least amount of control was for layer A4 in Georgia and South Carolina; only five values of transmissivity were available prior to calibration of the model. Although only five values of transmissivity were available for model layer A3 in Mississippi, the definition of transmissivity there was not significantly hindered because the extent of the Chattahoochee River aquifer (simulated as model layer A3) in that State is limited to the water-bearing part of the Ripley Formation of northeastern Mississippi.

Except for the basinwide estimates of transmissivity from the diffusivity data, most of the individual transmissivity values are representative of relatively shallow depths, over relatively small areas. The individual transmissivity values are correlative, for the most part, with specific sediments opposite specific well screens and are, therefore, not necessarily good indicators of the total productivity of the regional aquifer units. The general condition of partially penetrating wells precluded direct use of transmissivity values derived from either the aquifer-test or specific-capacity data. Because the regional aquifer units generally are thicker than the screened-interval lengths, the transmissivity values from the aquifer-test or specific-capacity data generally were smaller than required to adequately represent transmissivity of the entire aquifer thickness.

In an attempt to provide the model with initial values of transmissivity that were consistent with the hydrogeologic framework units described by Renken (1984), the original computations of transmissivity were adjusted upward, similar to the method used by McClymonds and Franke (1972) for an analogous situation in Long Island, N.Y. Assuming that length of the well screen is a reasonable measure of thickness of the aquifer contributing to well yield, the average hydraulic conductivity values of the contributing zones were determined by dividing the original transmissivity values by the lengths of the well screens at the appropriate wells. Transmissivity values were then recomputed by two slightly different procedures. In the first case, the average hydraulic conductivity values were multiplied by the total thickness of the regional aquifer unit near the wells under consideration. In the second, the average conductivity values were multiplied by only the sand thickness in that aquifer. Because the most permeable zones generally are screened, the modified transmissivity values from the first procedure were judged to overestimate the actual values of transmissivity. Although the second procedure provides transmissivity values that may underestimate actual conditions, these values appeared more realistic than the uncorrected transmissivity values that were based solely on the length of well screen. Because the sand thickness in a regional aquifer unit generally is less than the total aquifer thickness but greater than the length of the well screen, the second procedure was judged to provide the most appropriate distribution of transmissivity values with which to begin simulation. These initial input values of transmissivity were refined through model calibration.

The final, calibrated distribution of transmissivity in the model of the Southeastern Coastal Plain aquifer system is shown in figures 30 and 31. Differences between the initial input data and the final, calibrated transmissivity data are relatively minor regionally, and generally are less than ±50 percent locally. Many of the most significant differences result from calibration improvements that were identified in one or more of the subregional models and later incorporated in the data base of the regional model. The calibrated transmissivity values are closely correlated with the depositional environments of the Southeastern Coastal Plain aquifer system.

The Coastal Plain sediments originated in environments ranging from nonmarine, through marginal marine, to marine (Renken, in press). In general, the nonmarine deposits are products of relatively dynamic, high-energy processes, whereas the marine sediments reflect the influence of relatively tranquil, low-energy conditions. The bulk of the nonmarine rocks originated under fluvial or fluvial-deltaic conditions. Most marginal-marine sediments were deposited in deltaic, estuarine, tidal-flat, or barrier-island settings. Whereas the nonmarine and marginal-marine deposits are relatively

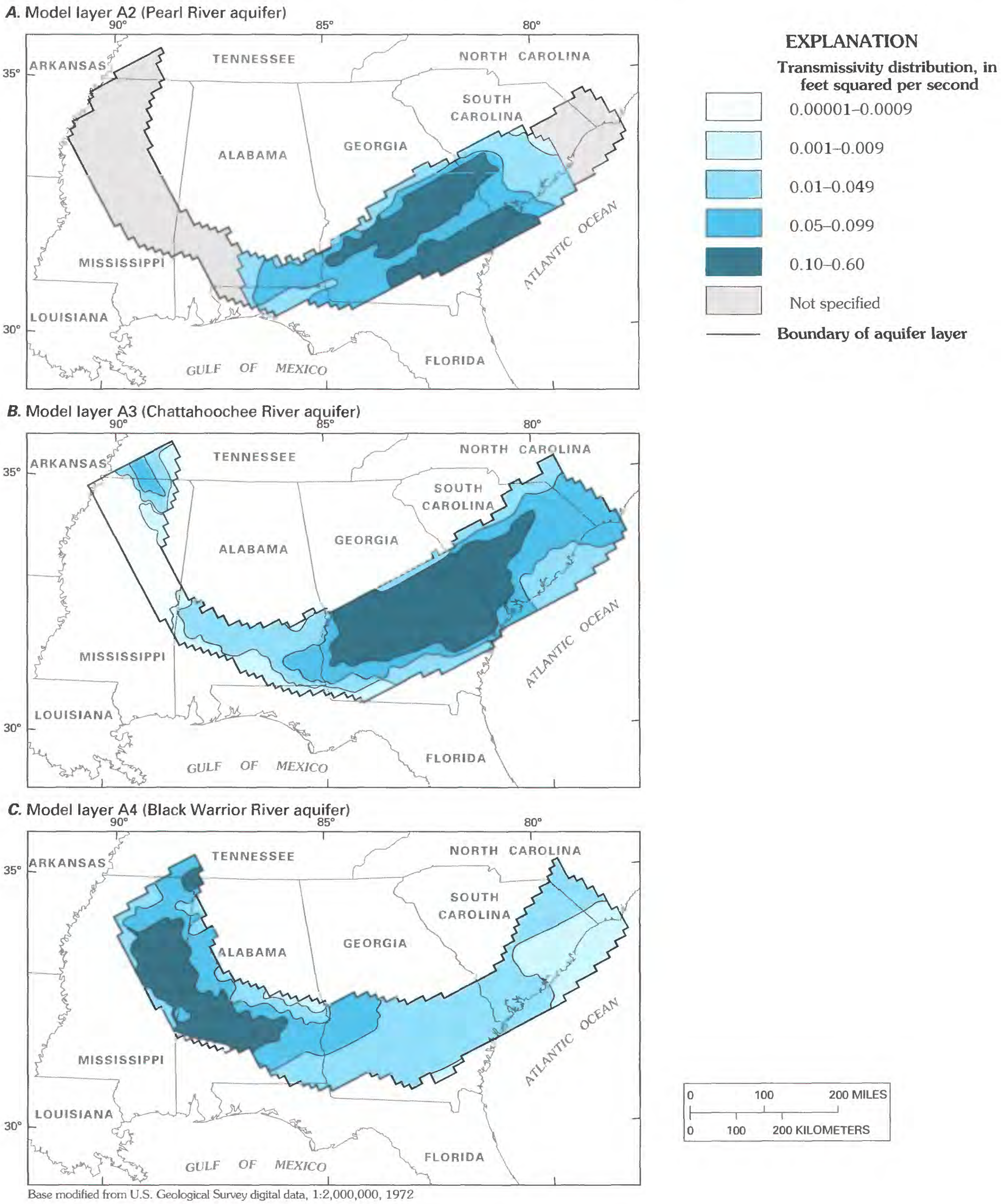


FIGURE 30.—Areal distribution of calibrated values of transmissivity in the model (modified from Pernik, 1987).

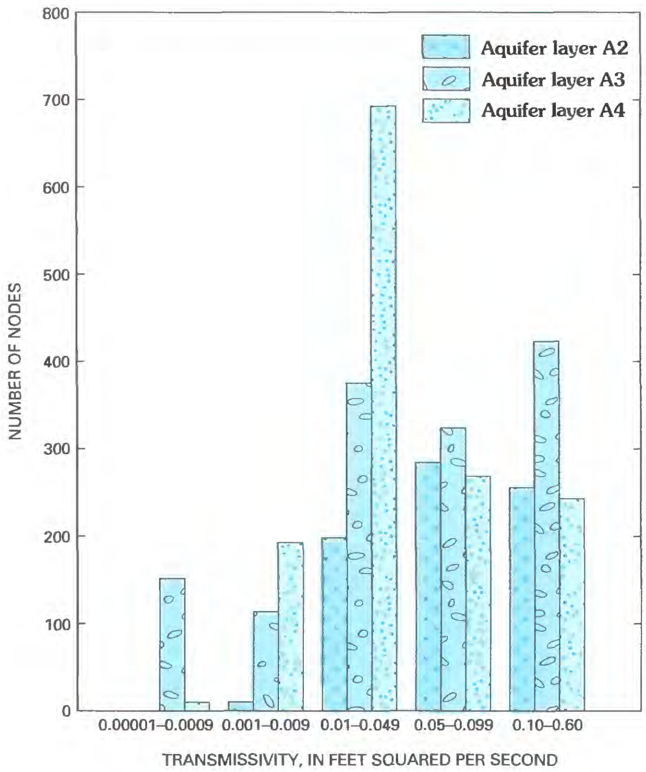


FIGURE 31.—Frequency distribution of calibrated values of transmissivity in the model.

localized, owing to often-changing depositional conditions, marine rocks are generally widespread and more lithologically uniform, owing to the more persistent nature of a fully submerged depositional environment. While the nonmarine clastic rocks are typically coarse grained and well sorted, their marine counterparts are generally fine grained and poorly sorted. Of the clastic deposits, the nonmarine sediments are the most permeable, the marine sediments are the least permeable, and the marginal-marine sediments are of intermediate and more variable permeability.

Figures 30*C* and 31 illustrate the distribution of transmissivity in model layer A4, representing the Black Warrior River aquifer, the most extensive clastic aquifer in the Southeastern Coastal Plain aquifer system (fig. 18). Calibrated transmissivity values range from about 0.00001 to about 0.2 ft^2/s and average about 0.06 ft^2/s. The transmissivity patterns in layer A4 reflect the characteristics of a sedimentary sequence that deepens from east to west and grades from mostly coarse-grained, nonmarine deposits near the inner margin of the Coastal Plain to finer grained, marginal-marine or marine sediments toward the southern limits of the modeled area. The largest transmissivity values are in Mississippi and Alabama, where the influences of fluvial-deltaic deposition are predominant and the accumulated thickness of sediment are the greatest. Although the Black Warrior River aquifer is not more than 500 ft thick in South Carolina, it is greater than 500 ft thick almost everywhere in east-central Mississippi, Alabama, and western Georgia. The largest transmissivity values in the western part of the study area result in part from the predominance of quartz sand and cherty gravel deposits of the Tuscaloosa Group and in part from thicker accumulations caused by subsidence that was contemporaneous with deposition within the Mississippi embayment. Overall, the smallest transmissivity values are in southeastern South Carolina, where the aquifer materials are much thinner and generally less permeable than in areas nearer the Mississippi embayment. The sediments in South Carolina are relatively thin owing to the influence of the Cape Fear Arch, a structural high during much of Woodbinian and Eaglefordian (Late Cretaceous) time, when deposition was occurring in the west. The Black Warrior River aquifer sediments are typically less permeable in South Carolina because the nonmarine and marginal-marine deposits there tend to be less well sorted and contain more interbedded clay and mudstone than deposits in the west. The transmissivity values in most of Georgia are in the intermediate range, which reflects a transitional depositional environment situated between a subsiding Mississippi embayment toward the northwest and a persistently high Cape Fear Arch toward the northeast. The relatively large values of transmissivity in western Georgia reflect the influence of predominantly nonmarine and shallow-marine conditions, whereas the smaller transmissivity values in southeastern Georgia and southern South Carolina suggest fully marine origins.

Model layer A3 (fig. 18) represents the Chattahoochee River aquifer, a diverse collection of sediments that were deposited under marine to nonmarine conditions during Late Cretaceous (Austinian) to late Paleocene (Sabinian) time. The highly variable lithology, texture, and bedding character of this aquifer result from a wide range of depositional environments. Although the areas of greatest permeability in the aquifer are composed predominantly of coarse-grained quartz sand that is interbedded in places with gravel, there are local occurrences of limestone (Clayton Formation) in the upper part of the unit in Georgia that are especially permeable owing to well-developed solution channels. The Chattahoochee River aquifer is continuous from North Carolina to central Alabama but is absent in western Alabama and east-central Mississippi. The aquifer is present, however, in north-central Mississippi, where it consists of the Ripley Formation, a glauconitic quartz sand of marine origin. The absence of water-bearing sediments between eastern Alabama and north-central Mississippi is due to massive thicknesses of chalk, shale, and clay of the Selma Group. Consequently, extremely small transmissivity values are used in model layer A3 to represent the hydraulic discontinuity through western Alabama and east-central Mississippi.

The calibrated transmissivity values for model layer A3 (fig. 30*B*) range from about 0.00001 to about 0.6 ft^2/s; the majority exceed 0.01 ft^2/s (fig. 31). Over the entire layer, transmissivity averages about 0.1 ft^2/s. The wide-ranging transmissivity values reflect the effects of areal differences in sediment thickness and lithology. The largest transmissivity values (0.1–0.6 ft^2/s) are in Georgia and west-central South Carolina. The largest values correspond to the area of greatest sediment accumulation; aquifer thicknesses range from more than 1,500 ft in central Georgia to about 1,000 ft in western South Carolina (Renken, in press). The moderately large transmissivity values (0.05–0.099 ft^2/s) in southeastern Alabama and southern Georgia are due to the largely terrestrial nature of the aquifer materials (Providence Sand, Ripley Formation, Cusseta Sand, and Blufftown Formation) in addition to the presence of highly permeable, sandy limestone of the Clayton Formation. Although the Clayton Formation (largely a calcareous marine sequence) is connected hydraulically to the underlying clastic sediments of the Chattahoochee River aquifer, it is separated from the overlying Floridan aquifer system (also limestone) by fine sand and clay of the Porters Creek Formation, which constitutes an effective confining unit. Despite a diminishing thickness of carbonate rock in the Chattahoochee River aquifer east of Georgia, a band of relatively large transmissivity values (0.05–0.6 ft^2/s) persists through middip South Carolina, owing mainly to the large permeability associated with massively bedded, fluvial-deltaic quartz sand and gravel deposits of the highly productive Middendorf Formation, which is included in the aquifer there. The lobate pattern of moderately large transmissivity values (0.05–0.099 ft^2/s) in northern Mississippi and southwestern Tennessee (fig. 30*B*) corresponds to marine sand of the Ripley Formation, which was deposited on the eastern flank of the Mississippi embayment during Late Cretaceous (Navarroan) time. The transmissivity values in model layer A3 are the largest almost everywhere in middip areas and decrease downdip. This can be attributed to the gradation from nonmarine or marginal-marine clastic strata in updip areas to calcareous shale, chalk, and clay of a shelfal, marine depositional environment downdip (Renken, in press).

The Pearl River aquifer, represented by model layer A2 (fig. 18), is predominantly a clastic sequence of unconsolidated to poorly consolidated sediments that were deposited for the most part under marine conditions except in western Alabama and Mississippi, where the sediments grade into sand formations of fluvial origin. Middip sediment thicknesses within the study area range from less than 200 ft in South Carolina to more than 1,000 ft in western Alabama and most of Mississippi. In western Alabama and Mississippi, the Pearl River aquifer grades in a downdip direction from porous sand, sandstone, gravel, and minor limestone to low-permeability clay, shale, mudstone, chalk, and chalky limestone near the downdip limit of the fresh ground-water flow regime. In southwestern South Carolina and southern Georgia, sands of the Pearl River aquifer grade easterly into limestone and dolomite that are part of the more permeable Floridan aquifer system. In much of southern Alabama and southeastern Georgia and South Carolina, the Pearl River aquifer is immediately overlain by, and hydraulically interconnected to, the Floridan aquifer system. The carbonate and clastic rocks merge hydraulically to form one more-or-less continuous conduit for ground-water flow. Because model layer A2 combines the effects of the Pearl River aquifer and the Lower Floridan aquifer where they are hydraulically continuous, the transmissivity data for this layer are influenced largely by the character of the generally more permeable Lower Floridan aquifer.

Calibrated transmissivity values for model layer A2 (fig. 30*A*) range from about 0.001 to about 0.6 ft^2/s and average about 0.07 ft^2/s. Figures 30A and 31 show that there is not as much variation in transmissivity within layer A2 as there is within layers A3 and A4. Part of this uniformity results from widespread marine conditions during much of Paleocene and Eocene time, when much of the limestone in the Lower Floridan aquifer was deposited in Georgia and South Carolina, the areas of largest transmissivity values in figure 30*A* (0.10–0.60 ft^2/s). The most permeable clastic rocks represented on the transmissivity map for model layer A2 are in updip parts of the Tallahatta and Barnwell Formations in Alabama, Georgia, and South Carolina (figs. 19–22). The areas of smallest transmissivity values in figure 30*A* (less than 0.05 ft^2/s) are in southeastern Alabama and western South Carolina (where the carbonate sequence is less than 200 ft thick) and in updip Georgia and South Carolina, where the nonmarine or marginal-marine clastic sequence pinches out against older, marine sediments of the Chattahoochee River confining unit. The transmissivity values of 0.05 to 0.099 ft^2/s in a band trending southwest-to-northeast across southern Georgia correspond to a structural depression known as the Gulf Trough (fig. 3). Miller (1986) notes that within this grabenlike feature the Floridan aquifer system is relatively thin and the limestone is less permeable. The reduced thickness and permeability combine to cause reduced transmissivity, with a coincident steepening of the hydraulic gradient, as shown on a potentiometric

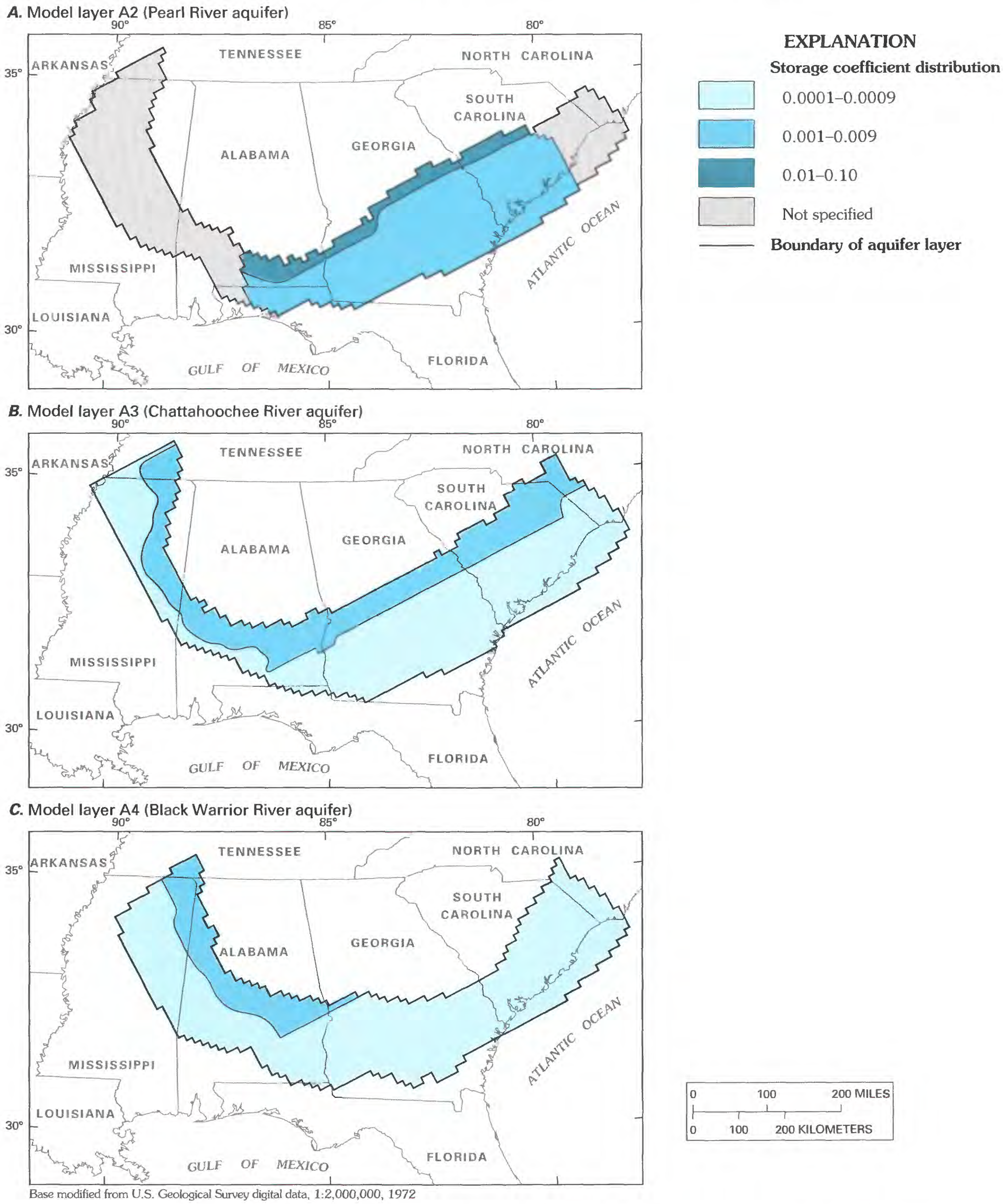


FIGURE 32.—Areal distribution of calibrated values of storage coefficient in the model (modified from Pernik, 1987).

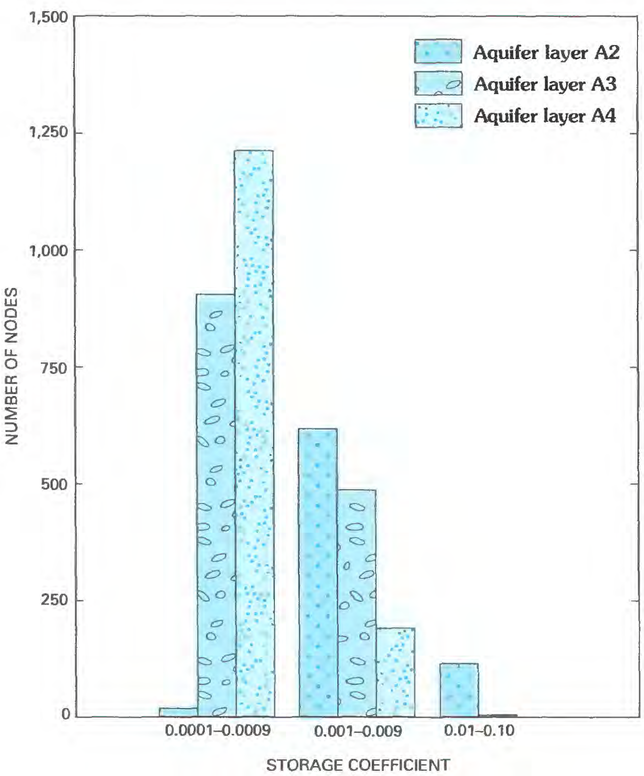


FIGURE 33.—Frequency distribution of calibrated values of storage coefficient in the model.

map of the Upper Floridan aquifer (Bush and Johnston, 1988, pl. 4).

Because the model does not actively simulate conditions in the Pearl River aquifer west of the Alabama River, the distribution of transmissivity in western Alabama and Mississippi is not discussed here. However, hydraulic conductivity data are available from Williamson and others (1990, fig. 20, table 4) for the Wilcox and Claiborne aquifers of the Mississippi embayment aquifer system (Grubb, 1986a).

STORAGE COEFFICIENT

Storage coefficient, as used herein, includes both the storage coefficient of a confined aquifer and the specific yield of an unconfined aquifer. Storage coefficient is the only input parameter that was calibrated strictly on the basis of output from the transient model; the results of calibration are shown in figures 32 and 33. Prior to construction of the computer model, storage coefficient was known for fewer than 100 sites in the aquifer system. Most of the individual storage-coefficient data were provided through use of the Theis (1935) type-curve or the Cooper-Jacob (Cooper and Jacob, 1946) straight-line approximation for nonleaky confined aquifers and are listed in various reports, including those by Newcome (1971), Knowles and others (1963), Faye and McFadden (1986), and Aucott and Newcome (1986). Owing to the limited number of the values of storage coefficient with which to estimate the model input data, average values were applied uniformly throughout each model layer at the beginning of the calibration process. The initial input values were 5×10^{-4} for model layer A4, 5×10^{-3} for layer A3, and 5×10^{-2} for layer A2.

The model input values of storage coefficient were adjusted by trial and error during the latter stages of calibration, with the purpose of simulating hydrograph trends that resembled those observed (pls. 5–7). The degree of adjustment was dictated by the apparent improvement in the calibration of the model and the sensitivity of the model output to changes in the input data. Although the simulated and observed hydrograph data differ in many cases, most of the discrepancies cannot be remedied through adjustment of the storage-coefficient data alone. The model is not very sensitive to changes within one order of magnitude of the original input values of storage coefficient (Pernik, 1987). As a result, the configuration of the calibrated storage-coefficient data does not differ substantially from the uniform pattern used initially. The calibrated values of storage coefficient range from 1×10^{-4} to 1×10^{-1} and average about 4×10^{-3}. In updip, unconfined to semiconfined areas, the calibrated values average about 1×10^{-2}. In middip and downdip, confined areas, the calibrated values average about 1×10^{-3}.

Lowering storage-coefficient values below those thought to provide the best overall representation of field conditions produces simulated water levels that in some areas appear to improve the comparisons of simulated and observed hydrograph data (pls. 5–7). Values small enough to significantly reduce the divergence between simulated and observed hydrograph data, however, are not compatible with those calculated from aquifer-test data and therefore are not considered realistic; nor are they considered appropriate to represent regional aspects of Coastal Plain hydrogeology. It is believed that the largest hydrograph mismatches evident in plates 5, 6, and 7 are due to errors stemming from the input of pumpage (which was not a calibrated parameter) and the coarse model grid, rather than to input values of storage coefficient.

The major difference between the uniform distribution of average data used initially and the calibrated values illustrated in figure 32 is that the calibrated distributions grade from larger values near the inner margin of the Coastal Plain to smaller values near the downdip margins of the simulated areas. As the frequency diagram (histogram) in figure 33 shows, relatively few nodes have large values of storage coefficient. The nodes with the largest values—regardless of the layer—are those representing the outcrop area, where the aquifers are largely unconfined and the sediments are generally coarser grained and better sorted (owing to nonmarine, perhaps fluvial-deltaic origins) than those downdip (owing to marginal-marine and marine depositional environments).

The deep ground-water flow regime of the Southeastern Coastal Plain aquifer system is for the most part confined; however, in the shallow, updip areas of the outcrops, especially near major rivers, even the deepest circulation may be unconfined, or confined to a lesser degree than in the subcrop. Because of this likelihood and the fact that the results of calibration tend to agree, it is believed that the gradational pattern of storage-coefficient values depicted in figure 32 is reasonable. The largest values are updip, in the shallowest parts of aquifer system, where they reflect semiconfined conditions within generally coarse-grained, reasonably well sorted, largely nonmarine sediments. Although the overall porosity may be greater in the relatively fine-grained marine sediments, the middip and downdip intermixing of sands with marine silts and clays inhibits the capacity of subcropping parts of the aquifer system to release stored water. The larger storage-coefficient values associated with the updip, semiconfined aquifers make them more efficient in terms of water supply and development, compared with the completely confined downdip aquifers.

Leakance

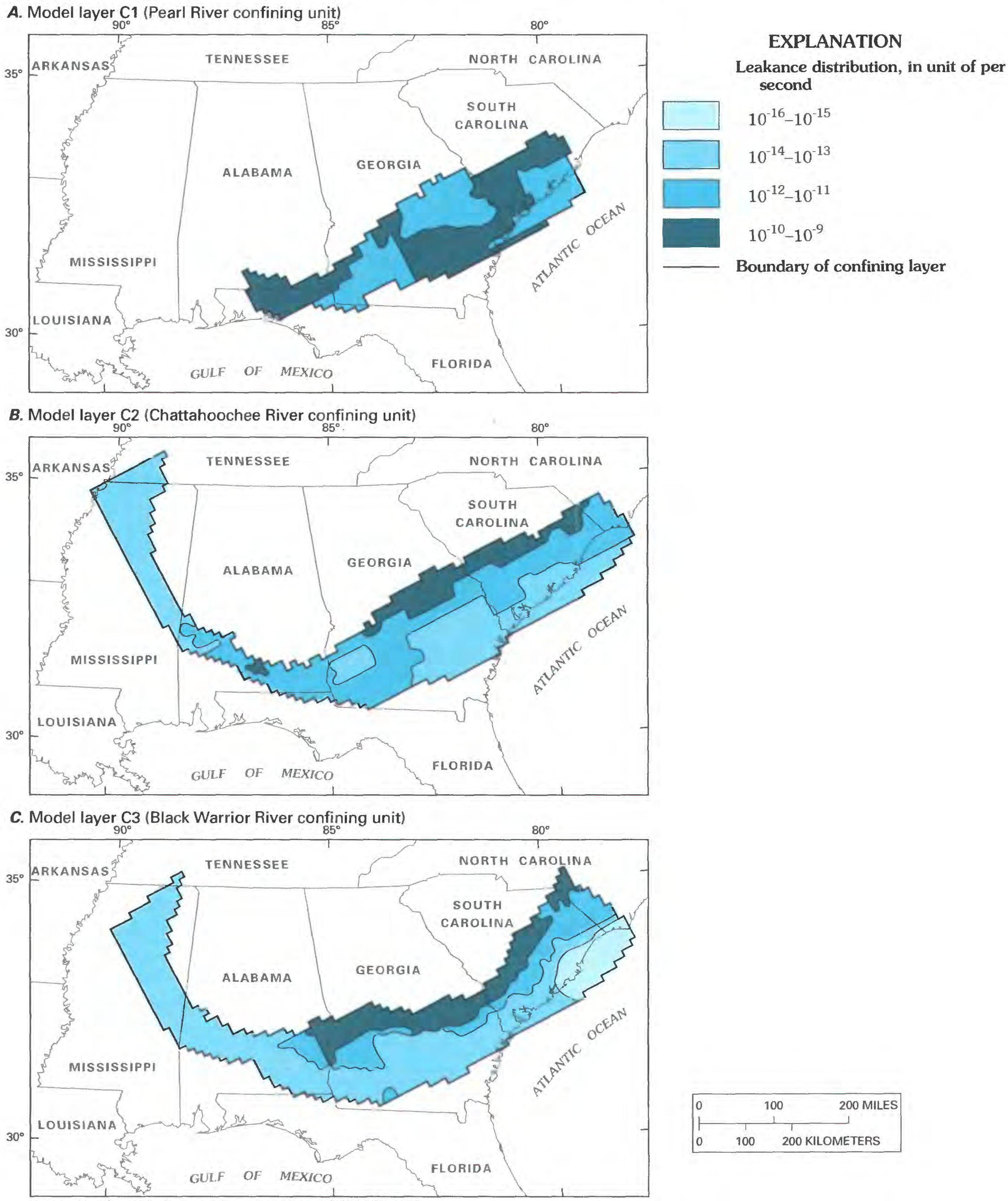
Leakance data are provided to the model of the Southeastern Coastal Plain aquifer system to simulate conditions of vertical ground-water flow across each of the three simulated confining units. Although good estimates of confining-unit thicknesses were available from the delineation of the hydrogeologic framework (Renken, in press), virtually no quantitative information was available for the vertical hydraulic conductivity of the confining materials within the study area. Consequently, the initial input of leakance was based on the mapped confining-unit thicknesses and an assumed vertical hydraulic conductivity of 1×10^{-10} ft/s, which was suggested by a list of clay hydraulic conductivities compiled by Bredehoeft and Hanshaw (1968). Although the initial estimates of leakance were appropriate at the beginning of model development, they had to be calibrated through trial-and-error simulation to provide a reasonably accurate distribution of vertical leakage and head gradient in the model.

The results of calibration with respect to confining-unit leakance are shown in figures 34 and 35. The leakance values generally are smaller for model confining layer C3 (equivalent to the Black Warrior River confining unit) and larger for confining layer C1 (equivalent to the Pearl River confining unit). The values range from about 5×10^{-16} per second in a deeply buried downdip area of South Carolina, where the confining sediments are essentially impermeable owing to their clay content, to about 5×10^{-9} per second in shallow, updip areas of eastern Alabama, Georgia, and South Carolina, where the sediments are relatively leaky because of their coarse, sandy nature. The largest values of leakance are updip and the smallest values are downdip, in accordance with a general increase in fine-grained sediment away from the inner margin of the Coastal Plain. The bands of relatively small leakance (less than 1×10^{-12} per second) in the deeper, downdip areas of South Carolina and Georgia can be attributed to moderately thick accumulations of calcareous clay, shale, mudstone, or marl of predominantly marginal-marine and marine origin. The relatively small values in Mississippi and Alabama correspond to the thick fine-grained facies of the Porters Creek Formation and the Prairie Bluff, Demopolis, and Mooreville Chalks (pl. 1).

The larger leakance values (greater than 1×10^{-11} per second) that rim the updip margins of confining layers C3 and C2 in Georgia and South Carolina are associated with relatively leaky, highly oxidized, clastic deposits that are relatively free of clay and other fine-grained material. Much of the recharge to the Black Warrior River aquifer (model layer A4) in eastern Georgia and South Carolina occurs as downward leakage through sandy, updip margins of the Black Warrior River and Chattahoochee River confining units. Because the model of the Southeastern Coastal Plain aquifer system includes the Lower Floridan aquifer as the upper part of layer A2 over much of eastern Alabama, Georgia, and western South Carolina (figs. 19–22), confining layer C1 corresponds in part to the middle confining unit of the Floridan aquifer system (Miller, 1986). The leakance values assigned to confining layer C1 were for the most part transferred directly from the model of the Floridan aquifer system that was calibrated earlier (Bush and Johnston, 1988, p. C25–C27). Leakage coefficients for the middle confining unit of the Floridan aquifer system were assigned initially on the basis of the presumed relation between lithology and the degree of confinement (Miller, 1986, p. 55–63). Consisting primarily of calcareous sand and sandy clay over much of southeastern Georgia and southwestern Alabama, the middle confining unit is relatively leaky where it overlies the Pearl River aquifer; thus, fairly large leakance values are shown in such places for confining layer C1 (fig. 34*A*).

Streambed Conductance

Owing to the relatively large grid-block dimensions of 8 mi on a side (fig. 26), the model simulates the effects of



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

FIGURE 34.—Areal distribution of calibrated values of confining-unit leakance in the model (modified from Pernik, 1987).

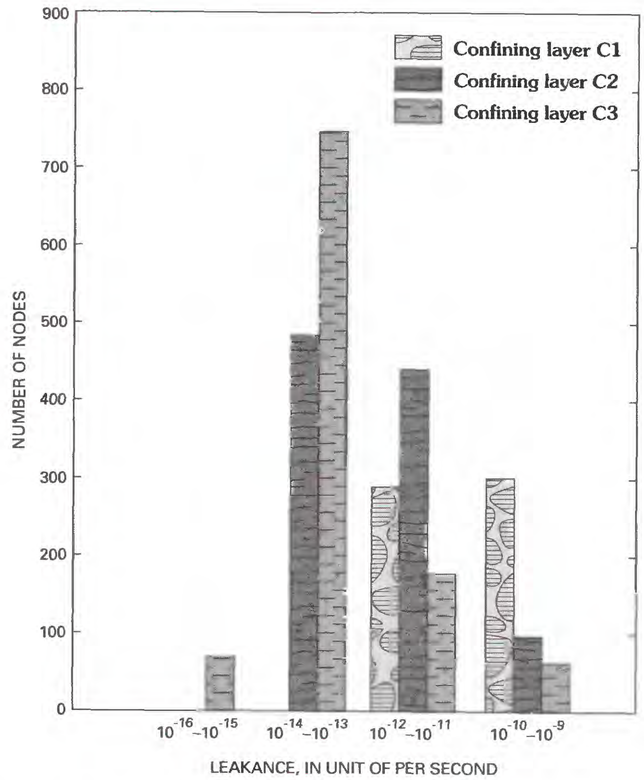


FIGURE 35.—Frequency distribution of calibrated values of confining-unit leakance in the model.

base flow to only the major streams in the Coastal Plain (fig. 14). For the purpose of model design, major streams (fig. 8) were defined as those that receive base flow from the deep flow regime (fig. 15). Major streams in the Coastal Plain can affect ground-water flow patterns as much as 50 or 100 mi away laterally, and perhaps 1,000 ft vertically. Figure 36 shows the nodal location of stream reaches within the Southeastern Coastal Plain for which interaction between the regional ground-water flow regime and major surface-water systems was simulated. The simulation of stream-aquifer leakage by the model is controlled most directly by an input parameter called streambed conductance, or *CRIV* (eq. 9).

The model employs Darcy's law in the vertical direction to simulate leakage through a reach of streambed (McDonald and Harbaugh, 1984). The relation between the simulated leakage (model output) and the associated field conditions (model input) for each grid block of the model can be expressed as

$$QRIV = K\,L\,W\,(HRIV - HAQ)/M, \tag{8}$$

where

$QRIV$ = leakage through a reach of streambed (in cubic feet per second);
K = vertical hydraulic conductivity of the streambed (in feet per second);
L = length of the streambed reach (in feet);
W = width of the streambed (in feet);
$HRIV$ = head on the river side of the streambed (in feet);
HAQ = head on the aquifer side of the streambed (in feet); and
M = thickness of the streambed (in feet).

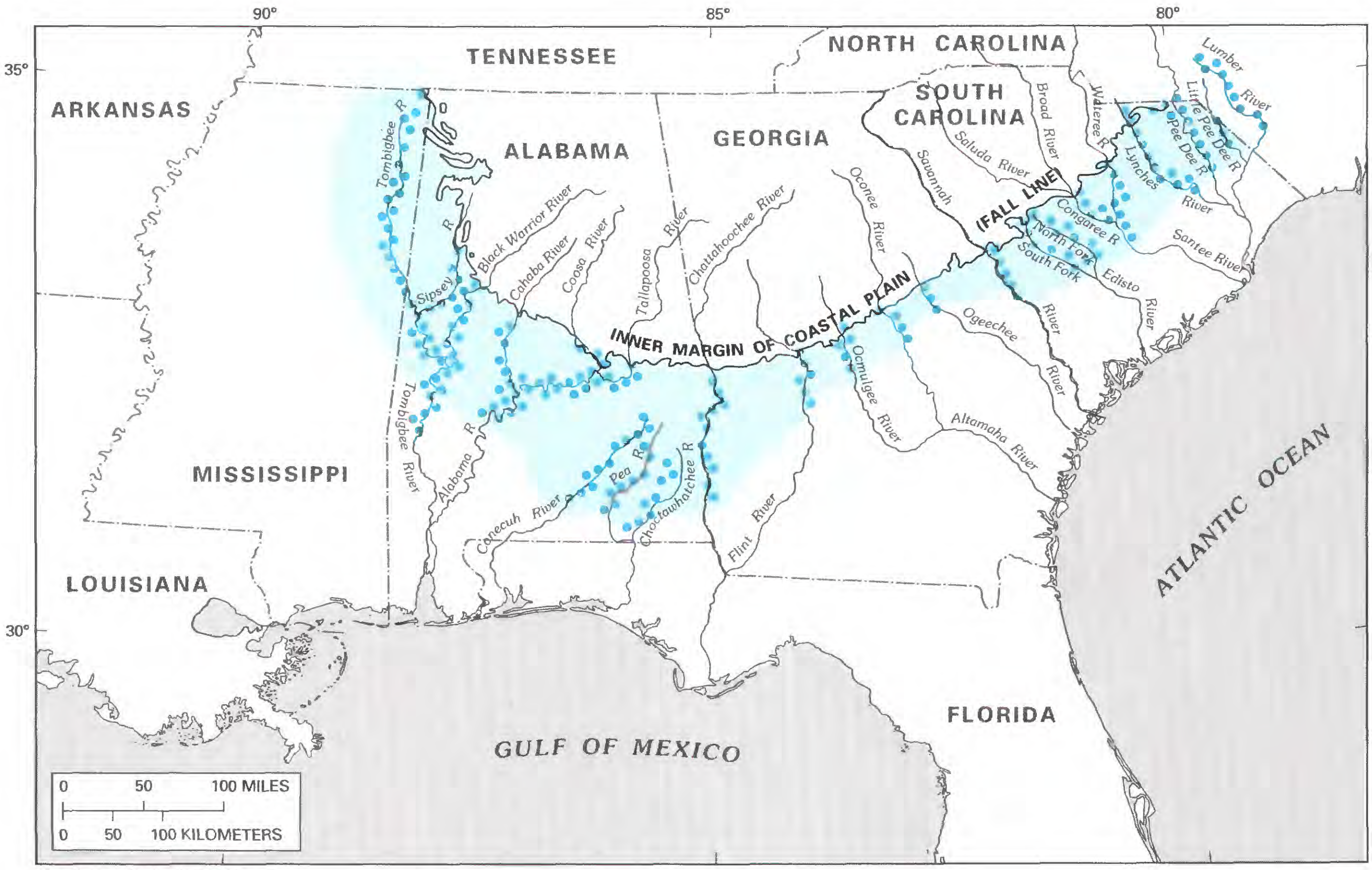
The numerical difference between the head in the stream (*HRIV*) and the head in the aquifer (*HAQ*) determines the hydraulic gradient across the streambed thickness (*M*) and the direction of flow between a given stream reach and the adjacent aquifer. A positive head gradient provides leakage into the aquifer from the stream; a negative gradient results in aquifer discharge to the stream, or base flow. Results of simulation indicate that under predevelopment conditions, the aquifer system discharged about 1,720 ft^3/s of ground water to major streams, while receiving less than 10 ft^3/s. The head on the aquifer side of the streambed (*HAQ*) is the head simulated by the model for aquifer material included in a particular grid block. This simulated aquifer head can vary during a model run, depending on the net effect of recharge into, and discharge out of, that grid block from one time step to another. The head on the river side of the streambed (*HRIV*) represents the average elevation of stream stage within the grid block and remains constant for any given stream reach throughout a model run. Values of *HRIV* for the model were estimated from surface-water records maintained by U.S. Geological Survey offices in Mississippi, Alabama, Georgia, and South Carolina; in a few cases, these stream elevations were scaled from U.S. Geological Survey 1:24,000 topographic maps. For simulation of long-term conditions, the estimated values of *HRIV* were assumed to be constant with time.

Although it was possible to estimate the lengths (*L*) and widths (*W*) of simulated stream reaches, it was not practical to assess the individual effects of thickness (*M*) and hydraulic conductivity (*K*) and apply the results consistently on an independent basis. Therefore, the effects of *K*, *M*, *W*, and *L* were considered in combination and incorporated into a single conductance parameter (*CRIV*), where

$$CRIV = (K\,L\,W)/M. \tag{9}$$

Input values of *CRIV* (in feet squared per second) were calibrated through an iterative, trial-and-error process to reflect the net effect of streambed geometry and permeability on stream-aquifer leakage.

Calibrated streambed conductances (fig. 37, table 4) range from about 0.005 to nearly 1.0 ft^2/s at individual river nodes and average about 0.2 ft^2/s across the model area. Assuming an average streambed thickness (*M*) of



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

EXPLANATION

Outcrop area of actively simulated part of aquifer system

Stream node in model

FIGURE 36.—Location of stream nodes and outcrop area of the actively simulated part of the Southeastern Coastal Plain aquifer system (modified from Barker, 1986).

about 10 ft and an average streambed area (*LW*) of 10,000,000 ft^2 per grid block, the conductance data suggest that the vertical hydraulic conductivity (*K*) values of the simulated streambeds average about 10^{-7} ft/s. The value appears reasonable, considering the infrequently scoured, silt-laden channel bottoms characteristic of the relatively low gradient, meandering streams in the Southeastern Coastal Plain. In comparison, lateral hydraulic conductivity values of adjacent aquifers probably average between 10^{-4} and 10^{-5} ft/s, or about two to three orders of magnitude greater than the average vertical conductivity indicated for streambeds. Streambed leakance is equal to streambed conductance (*CRIV*) divided by streambed area (*LW*). The average streambed leakance (*K/M*) of about 10^{-8} per second is about 10 times greater than the calibrated leakance values for the most permeable parts of the confining units in the Southeastern Coastal Plain aquifer system (fig. 34).

RECHARGE

Because the regional model of the Southeastern Coastal Plain aquifer system simulates only the deep flow regime, the input recharge is necessarily less than the total recharge to the entire aquifer system (Barker, 1986). Simulated recharge to the deep flow regime averages about 0.6 in/yr, which is less than one-tenth of the estimated total recharge of about 7 in/yr (fig. 16). The difference between the two is equal to ground-water discharge to relatively shallow surface drainages, which is not simulated within the coarse grids of the regional model. As previously explained, the deep flow regime is the predominantly confined, less dynamic part of the flow system—including all of the subcrop area and the part of the outcrop area that discharges to the major streams. Conceptually, simulated recharge equals precipitation minus the sum of evapotranspiration and all runoff that is not simulated. Because the model is limited to simulating

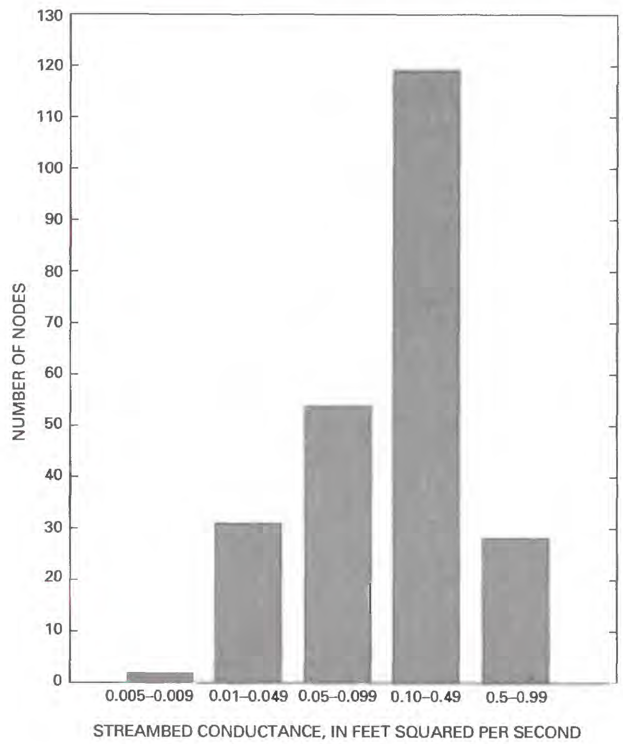


FIGURE 37.—Frequency distribution of calibrated values of streambed conductance in the model.

TABLE 4.—*Average streambed conductance values for simulated reaches of the major streams in the model of the Southeastern Coastal Plain aquifer system*

[Dashes indicate reach of stream not simulated]

River	Average conductance in model layer, in feet squared per second		
	A2	A3	A4
Tombigbee	—	0.05	0.09
Sipsey	—	—	.06
Black Warrior	—	—	.12
Alabama	—	.24	.14
Cahaba	—	.06	.08
Coosa	—	—	.27
Tallapoosa	—	—	.03
Pea	0.04	.12	—
Conecuh	.12	.27	—
Choctawhatchee	.15	.04	—
Chattahoochee	.12	.27	.25
Flint	.10	.64	—
Ocmulgee	.30	.08	—
Oconee	.18	.04	—
Ogeechee	.04	.20	—
Savannah	.53	.33	—
South Fork Edisto	.34	.68	—
North Fork Edisto	.30	.32	—
Congaree	.18	.39	—
Wateree	.10	.24	—
Lynches	—	.45	—
Pee Dee	—	.38	—
Little Pee Dee	—	.50	—
Lumber	—	.39	—

the base flow of only major streams (fig. 36), the input recharge represents only the water that discharges either to these streams or from downgradient parts of the deep, confined flow regime. The original understanding of the deep flow regime was largely conceptual; the definition of recharge to this regime is, likewise, conceptual. The rates of input recharge were fashioned initially from the map of infiltration potential (fig. 12) and were refined later through trial-and-error simulation.

The calibrated distribution of recharge is shown in figure 38. Simulated recharge averages about 0.6 in/yr over the nearly 46,500 mi^2 of actively simulated outcrop area. Figure 39 illustrates the frequency distribution of the simulated recharge. Model layer A4 receives the least amount of recharge, averaging about 0.4 in/yr across the outcrop area of the Black Warrior River aquifer, which covers about 12,000 mi^2. Model layer A3 (Chattahoochee River aquifer) receives about 0.7 in/yr across its outcrop area of about 18,500 mi^2. Model layer A2 is recharged in the actively simulated part of the outcrop area of the Pearl River aquifer (about 16,000 mi^2) at rates averaging about 0.6 in/yr.

As explained earlier, the steady-state (predevelopment) and transient models were calibrated in an iterative fashion; as a result, the calibrated rates of outcrop recharge should represent long-term average conditions, rather than specifically those of predevelopment or of any time since. Recharge to the deep ground-water flow regime from the outcrop area may have increased since 1900 (predevelopment) because the gradient between the deep and shallow regimes may have increased owing to water-level declines caused by pumpage from the deep confined aquifers. If such an increase in recharge has occurred, the calibrated recharge rates may be somewhat larger than those that actually existed under predevelopment conditions and somewhat smaller than those under present (1985) conditions. The calibrated rates of recharge (fig. 38), however, are believed to be the most appropriate set of time-constant values for simulation between 1900 and 1985, owing to the iterative nature of the calibration and the fact that the effects of the pumpage on water-level declines in the interstream (recharge) areas of the outcrop area appears to be relatively insignificant or localized.

There are important similarities among the simulated rates of recharge (fig. 38), the infiltration potential of the outcrop areas (fig. 12), and the lithology of the outcropping aquifer and confining units (Renken, in press). The least amount of recharge (less than 1 in/yr) occurs in Mississippi and western Alabama, where clay-rich soils associated with the outcrop areas of the Prairie Bluff, Demopolis, and Mooreville Chalks dominate much of the

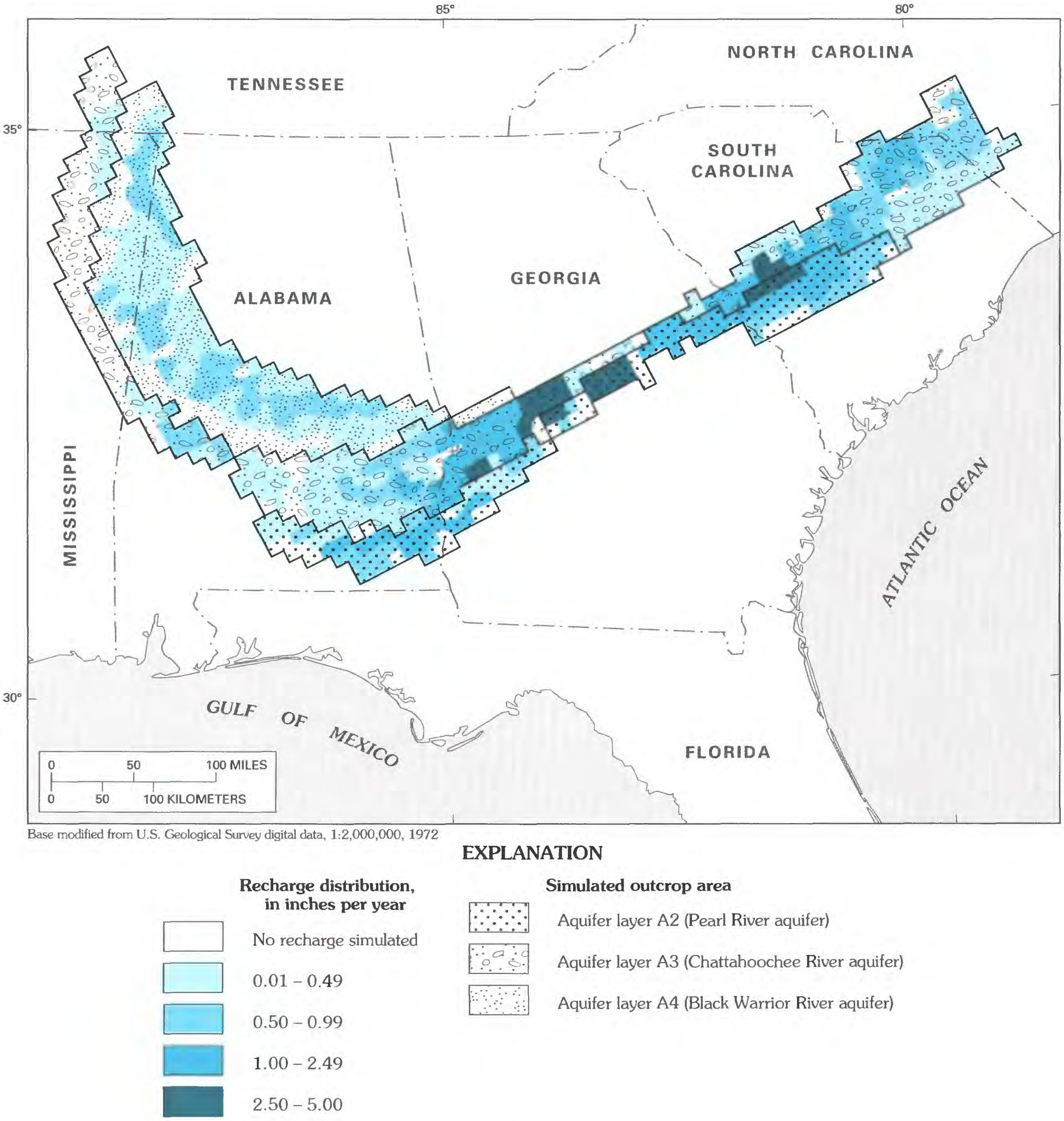


FIGURE 38.—Areal distribution of calibrated values of recharge in the model (modified from Pernik, 1987).

area and significantly limit infiltration. The underlying marine strata of chalk, shale, and clay additionally limit subsurface permeabilities, forcing water to discharge at or near the land surface as overland flow or shallow base flow. Recharge to the deep flow regime averages about 0.25 in/yr in Mississippi and western Alabama. The areas of highest recharge (greater than 1 in/yr) correspond to the sandy soils present from southeastern Alabama, across the outcrop area of Georgia, to western South Carolina. Both infiltration potential and permeability are significantly greater in eastern Alabama, Georgia, and western South Carolina than in western Alabama and

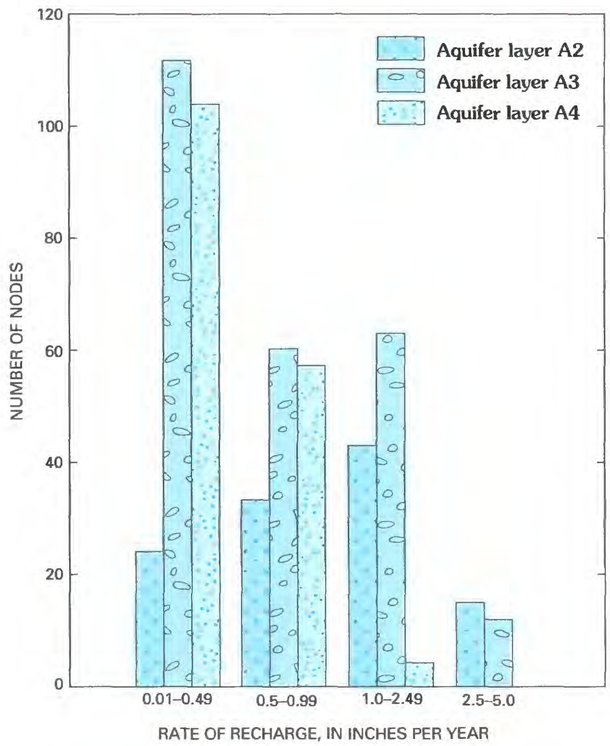


FIGURE 39.—Frequency distribution of calibrated values of recharge in the model.

Mississippi. Recharge to the deep flow regime averages about 1 in/yr in the eastern part of the study area.

The simulated rates of recharge to the deep, confined part of the Southeastern Coastal Plain aquifer system are consistent with those simulated in models of the adjacent regional aquifer systems. Brahana and Mesko (1988) reported that simulated rates of "deep recharge" average about 0.3 in/yr for both the McNairy-Nacatoch and lower Wilcox aquifers, which are correlative to all or parts of the Chattahoochee River and Pearl River aquifers, respectively, in northeastern Mississippi (table 1). Meisler and others (1986) reported simulated rates of "deep percolation" to confined aquifers of the Northern Atlantic Coastal Plain aquifer system ranging between 0 and 1 in/yr across northeastern South Carolina and eastern North Carolina. For the Floridan aquifer system, Bush and Johnston (1988) reported simulated rates of recharge ranging from 0 to about 20 in/yr and averaging less than 5 in/yr across eastern Alabama, Georgia, and western South Carolina. The larger recharge rates for the Floridan aquifer system are compatible with transmissivity values in the outcrop area of the Floridan that exceed those of the Southeastern Coastal Plain aquifer system by an order of 10 or more. The greatest rates of recharge (10 to 20 in/yr) to the Floridan system are in karstic limestone terrains where there is little or no overland runoff; therefore, they are not comparable to parts of the Atlantic-Gulf Coastal Plain underlain by clastic sediments (R.H. Johnston, U.S. Geological Survey, written commun., 1990).

MODEL RESULTS: SIMULATION OF DEEP GROUND-WATER FLOW

PREDEVELOPMENT (PRE-1900) CONDITIONS

Plates 2, 3, and 4 illustrate simulated distributions of hydraulic head under predevelopment (pre-1900) conditions in the deep, confined parts of three regional aquifers of the Southeastern Coastal Plain aquifer system. In addition to replicating key aspects of hand-drawn potentiometric surfaces (Stricker and others, 1985a, 1985b, 1985c), the simulated potentiometric surface maps cover downdip parts of the deep flow regime and, thus, augment the understanding of the system in areas for which few field data exist. Except for the simulated potentiometric surfaces, information on hydraulic head is sparse and discontinuous for the deeper, downdip parts of the system.

The simulated water budget for the deep flow regime under predevelopment, or steady-state, conditions is shown in figure 40. The water budget shows the relative importance of vertical and lateral aspects of inflow and outflow. While the sum of vertical inflow is 165 ft^3/s, vertical outflow totals 410 ft^3/s; lateral inflow and outflow total only 10 ft^3/s and 35 ft^3/s, respectively.

The simulated interstream potentiometric highs on plates 2, 3, and 4 correspond to recharge mounds in the outcrop area of the Southeastern Coastal Plain aquifer system. The simulated potentiometric surface for model layer A4 (pl. 4) shows the significance of downward leakage from layer A3 (Chattahoochee River aquifer) near the updip fringes of layer A4 (Black Warrior River aquifer) in eastern Georgia and South Carolina. Because the Black Warrior River aquifer does not crop out there, downward leakage from the overlying Chattahoochee River aquifer is the sole source of recharge to this part of the aquifer. The simulated potentiometric surfaces for model layers A2, A3, and A4 (pls. 2–4) show the effects of ground-water discharge to the major rivers. Simulated base flow is concentrated near the downdip limits of the outcrop areas and decreases with distance from the outcrop and depth to the aquifer top below land surface. This simulated pattern of stream-aquifer interaction corresponds well with conceptualization by LeGrand and Pettyjohn (1981), who referred to erosional notches (fig. 13) that breach updip edges of the confining units and enhance ground-water discharge to deeply entrenched, consequent streams such as the Chattahoochee, Savannah, and Congaree Rivers of the Southeastern Coastal Plain. The concentration of base flow near these artesian-

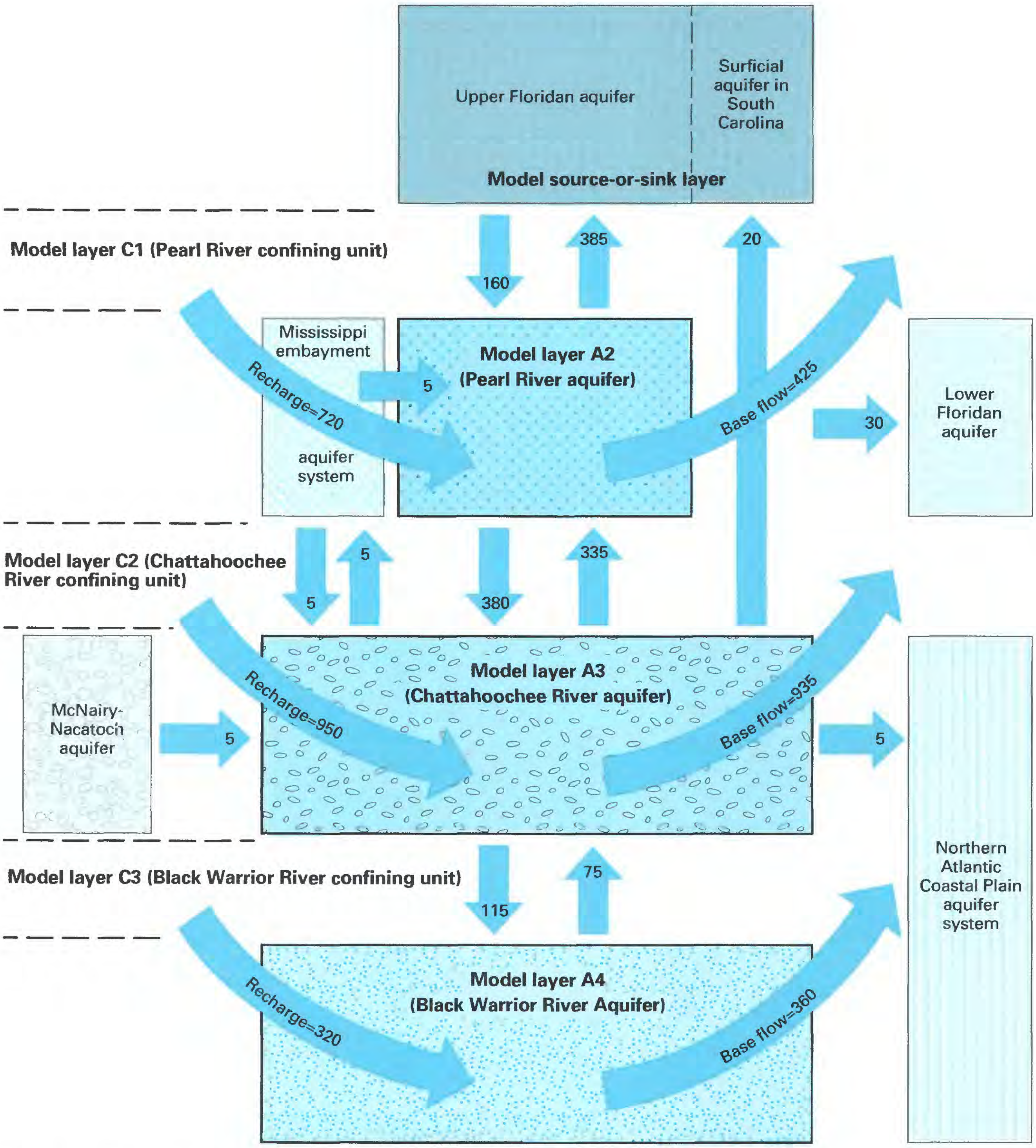


FIGURE 40.—Simulated water budget of predevelopment (pre-1900) conditions in the Southeastern Coastal Plain aquifer system. (In units of cubic feet per second, rounded to the nearest 5 cubic feet per second.)

water gaps, as described by LeGrand and Pettyjohn, is reflected on the simulated potentiometric maps by the closed, or sharply bent (concave) contours in the outcrop areas along the major streams.

Figure 41 shows the distribution of simulated base flow to the major streams in the Southeastern Coastal Plain. Although there are few specific data on streamflow gains from the deep ground-water flow regime with

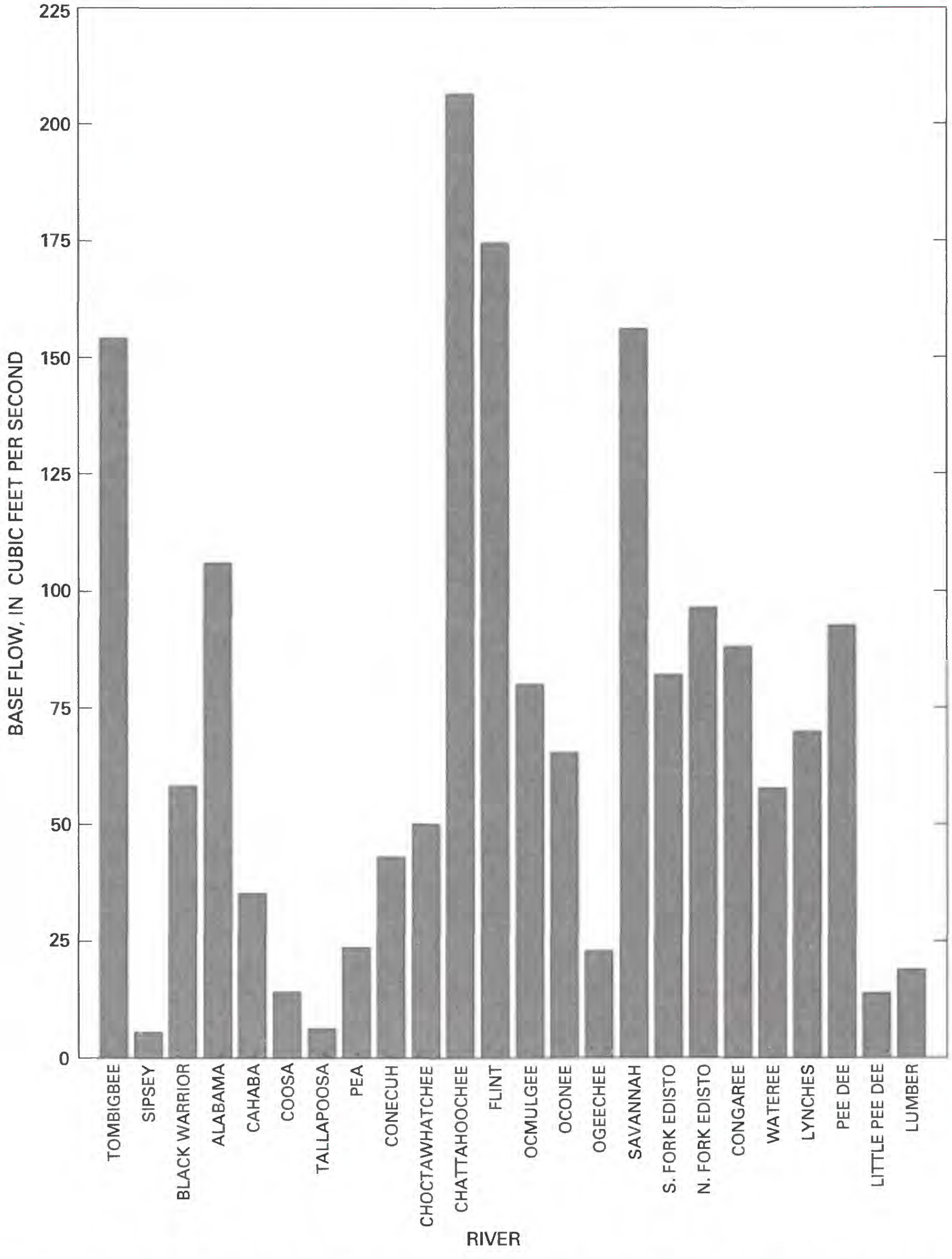


FIGURE 41.—Rates of simulated base flow from the deep ground-water flow regime to major streams in the Southeastern Coastal Plain under predevelopment (pre-1900) conditions.

which to evaluate the accuracy of the simulated base flow, the simulated values appear to be consistent with the results of low-flow analyses by Siple (1960), Callahan (1964), Gardner (1981), Stricker (1983), Aucott and others (1986), and Faye and Mayer (1990). Where it is possible from the analyses to differentiate between discharge from the shallow (local) ground-water flow regime and discharge from the deep (regional) flow regime, the simulated base flow typically approximates- the discharge from the deep flow regime to within ±25 percent. Owing to the coarse grid of the model, it seems appropriate that simulated base flow typically is less than estimates of total base flow, which include discharge from the shallow ground-water flow regime.

Model results indicate that 1,720 ft^3/s (or about 86 percent of the simulated deep recharge of 1,990 ft^3/s) discharged under predevelopment conditions as base flow to the major streams in the outcrop area of the aquifer system. The difference between the simulated-deep recharge and the simulated base flow to the major rivers is 270 ft^3/s. This residual is termed "deep seepage" (fig. 15). Although the limited resolution of the finite-difference grid (fig. 26) prevents the model from simulating the shallow flow regime of the outcrop area, the model fully accounts for water entering the subcropping part of the aquifer system. Simulation suggests, therefore, that under predevelopment conditions about 270 ft^3/s of water flowed from the outcrop area and infiltrated the subcropping, confined part of the aquifer system as deep seepage.

Although simulated hydraulic gradients for updip parts of the aquifer system are relatively steep and indicate flow paths that, for the most part, parallel the dip of the bedding, the downdip gradients—especially in the deeper parts of the flow system—suggest relatively gentle, long flow paths that in places cut arcuately across the regional dip. These patterns indicate that deep seepage migrates circuitously within the confined part of the aquifer system toward remote outlets. Because deep, downdip parts of the aquifer system are buffered from the outcrop surface drainage network, diffuse upward leakage (fig. 42) is essentially the only way deep seepage exits the confined parts of the aquifer system.

Owing to vertical gradients and confining-unit leakances that generally decrease with depth and distance from the outcrop area, the simulated rates of vertical discharge decrease from updip to downdip (fig. 43) and taper to insignificant amounts near the mapped limits of freshwater circulation (Lee and others, 1986; Strickland and Mahon, 1986). This is consistent with the belief that ground-water flow is essentially stagnant near the deepest downdip parts of the freshwater flow system.

Figure 40 indicates that the amount of ground-water exchange between the Southeastern Coastal Plain aquifer system and the adjacent regional aquifer systems is small, except in the case of the Floridan aquifer system. Simulated lateral and vertical inflow from the Mississippi embayment aquifer system totals 10 ft^3/s; the simulated outflow (all vertical) to that aquifer system is 5 ft^3/s. Lateral inflow from the McNairy-Nacatoch aquifer of southwestern Tennessee is simulated to have been about 5 ft^3/s under predevelopment conditions. Simulated discharge to the Northern Atlantic Coastal Plain aquifer system is 5 ft^3/s. Simulated predevelopment inflow to the Southeastern Coastal Plain aquifer system from the Floridan aquifer system is 160 ft^3/s, all of which is downward leakage from the Upper Floridan aquifer. Simulated outflow to the Floridan aquifer system (including the hydraulically equivalent surficial aquifer in South Carolina) totals 435 ft^3/s, of which 30 ft^3/s is lateral discharge within the Lower Floridan aquifer across the southern, constant-head boundaries in model layer A2. The balance, 405 ft^3/s, represents diffuse upward leakage; of that amount, 20 ft^3/s is simulated to leak vertically from model layer A3 (Chattahoochee River aquifer) directly to the unconfined surficial aquifer in eastern South Carolina, where the normally intervening Pearl River aquifer is not present (figs. 17, 18). Flow to the surficial aquifer is included in the water budget (fig. 40) with that to the Floridan aquifer system because the water table in the former is continuous with the hydraulic heads in the Upper Floridan aquifer. As previously explained, the water-table heads were included in the model as an eastern extension of the source-or-sink boundary condition that represents the Upper Floridan aquifer in eastern Alabama, Georgia, and western South Carolina.

POSTDEVELOPMENT (1900–85) CONDITIONS

Ground-water pumpage from the Southeastern Coastal Plain aquifer system began about 1900. Widespread water-level declines have resulted from the pumpage; locally, declines have exceeded 150 ft and have raised concern among farmers, water managers, and public officials. Owing to the decline, the water budget and flow patterns of the aquifer system have changed. The calibrated transient model simulates the effects of pumpage on ground-water levels, flow patterns, and the water budget.

To evaluate these effects with the computer model, the period of ground-water withdrawal, 1900–85, was divided into 10 pumping periods (fig. 44). The pumpage data used in the regional model were extracted directly from pumpage records for Mississippi, Alabama, Georgia, and South Carolina that were compiled, respectively, by Mallory (1993), Planert and others (1993), Faye and Mayer (in press), and W.R. Aucott (in press). The areal distribution of the simulated pumpage during the 1981–85 period is shown in figure 45. In addition to calibrating the transient model to approximate hydrograph trends, the capacity of the model to simulate present-day conditions was evaluated against potentiometric surfaces constructed from recent water-level data by Wasson (1980), Darden (1984, 1985a, 1985b), Aucott and Speiran (1984b), Williams, DeJarnette, and Planert (1986b), and Williams, Planert, and DeJarnette (1986a, 1986c).

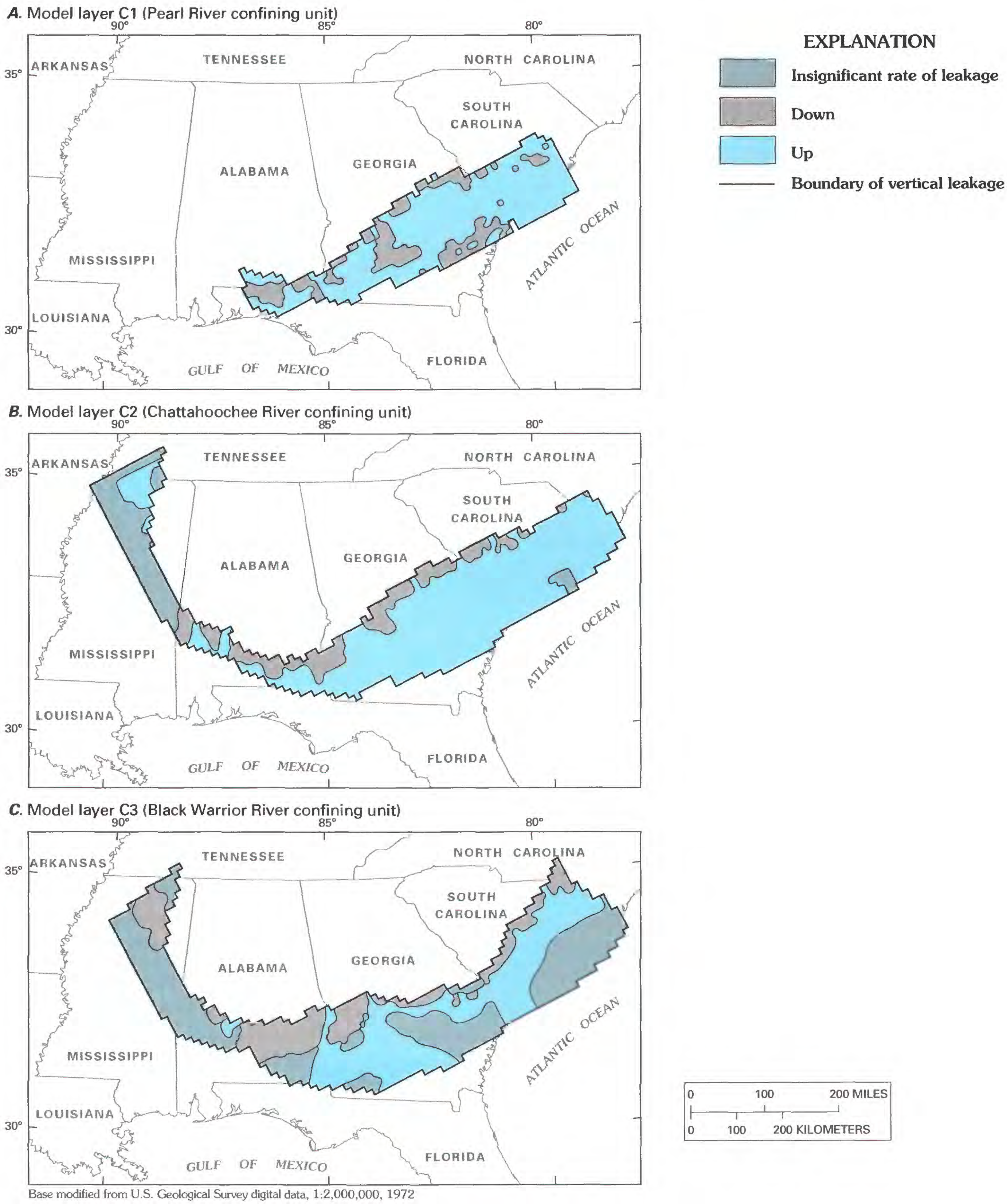


FIGURE 42.—Areal distribution and direction of simulated vertical leakage through confining units under predevelopment (pre-1900) conditions.

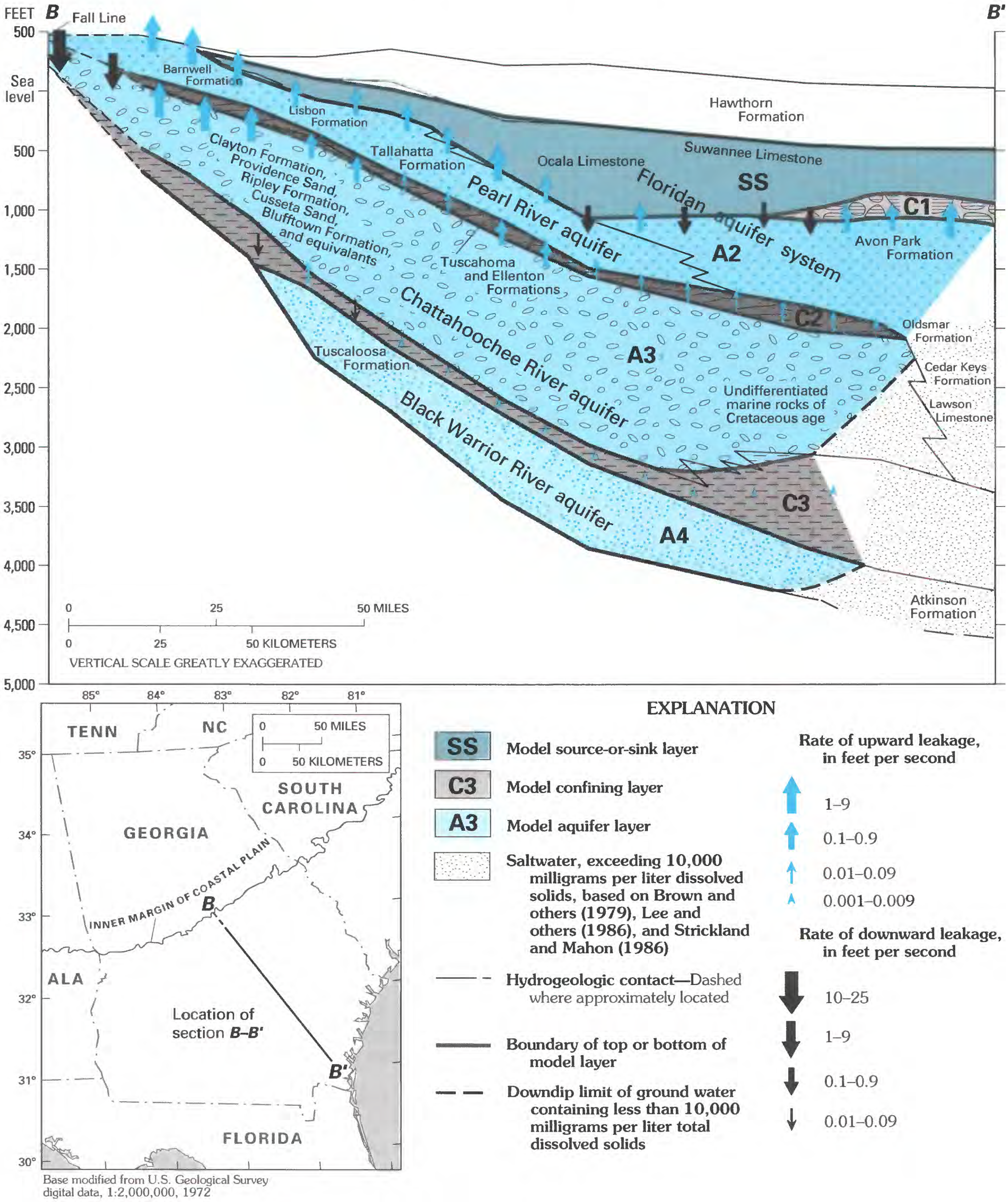


FIGURE 43.—Simulated rates of vertical leakage under predevelopment (pre-1900) conditions along a line through southeastern Georgia (hydrogeologic section modified from Renken, in press).

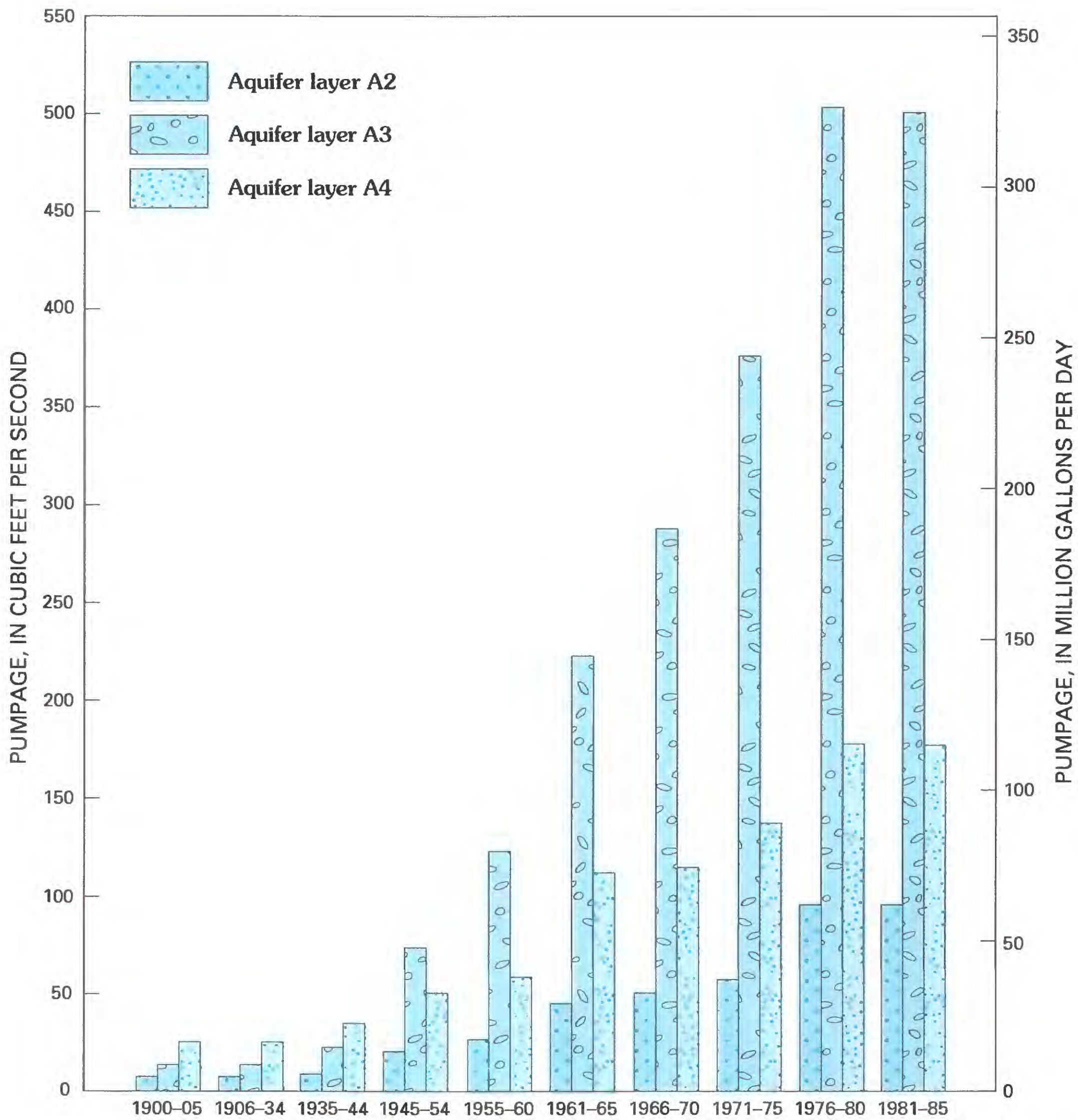


FIGURE 44.—Time distribution of simulated pumpage from the Southeastern Coastal Plain aquifer system, 1900–85.

Plates 5, 6, and 7 show the simulated decline in hydraulic head during 1900–85, and plates 8, 9, and 10 show the simulated 1985 potentiometric surfaces, for model layers A2, A3, and A4, respectively. The simulated conditions in model layer A2 relate most directly to the proximity of pumping from the Pearl River aquifer (fig. 45*A*). Most of the pumpage, and therefore most of the simulated decline, has occurred in updip parts of Georgia and western South Carolina. The simulated conditions in layers A3 and A4 suggest that the most significant changes in hydraulic gradient and flow direction have occurred in subcropping parts of the Chattahoochee River and Black Warrior River aquifers. Although heads have declined at least 5 or 10 ft throughout most of the study area, the largest cones of depression have developed in middip areas of the Chattahoochee River and Black Warrior River aquifers; this pattern of decline—simulated in model layers A3 and A4—reflects the effect of storage coefficient values being generally larger in shallow, updip areas where conditions range from unconfined to semiconfined, and the fact that captured base flow in the outcrop area more effectively compensates for pumpage than does downdip change in the rates and directions of vertical leakage.

Despite there being virtually no pumpage from the Black Warrior River aquifer (model layer A4) in eastern Georgia and South Carolina (fig. 45*C*), up to 30 ft of water-level decline is simulated to have occurred here

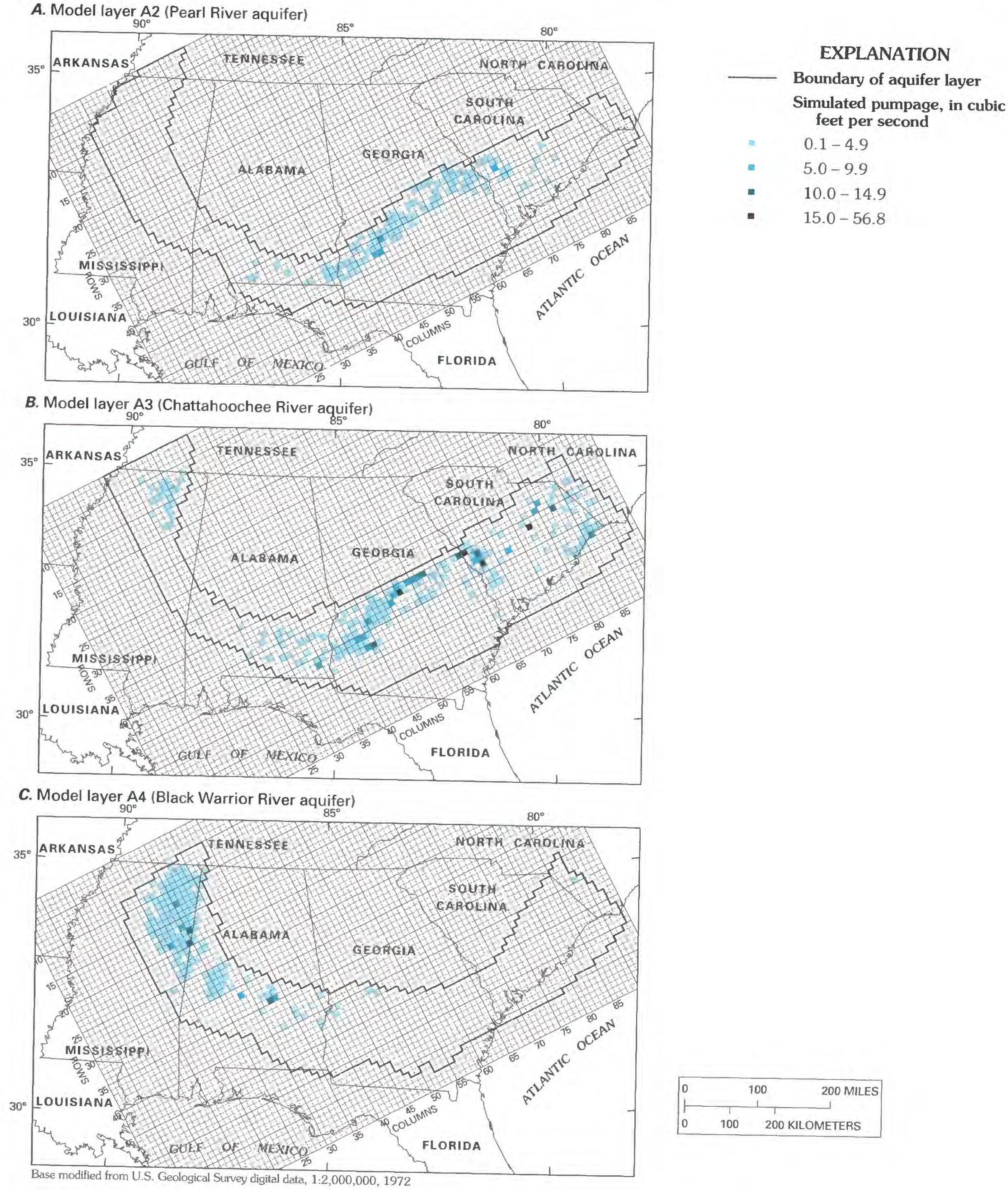


FIGURE 45.—Areal distribution of simulated pumpage from the Southeastern Coastal Plain aquifer system, 1981–85.

(pl. 7). This decline is due apparently to the combination of (1) no direct recharge to this part of the Black Warrior River aquifer (fig. 38), (2) relatively small values of transmissivity in this part of the Black Warrior River aquifer (fig. 30), (3) large rates of pumpage from the overlying Chattahoochee River aquifer (fig. 45*B*), and (4) large leakance values in updip parts of the intervening Black Warrior River confining unit (fig. 34). Model results indicate that pumpage from the overlying Chattahoochee River aquifer has removed water from the Black Warrior River aquifer through upward leakage across the Black Warrior River confining unit.

Plates 5, 6, and 7 also show comparisons of simulated and observed hydrographs. Owing to the limited resolution of the model, the simulated head decline is not as much as the measured decline near pumping wells in the areas of greatest change. However, the most important areal and temporal patterns of decline since 1900 are reasonably well reproduced by the model. The overall compatibility of the simulated and observed potentiometric data suggest that the model can be used to evaluate regional changes in the water budget.

Comparing the simulated water budget for the period 1981–85 (fig. 46) with that for predevelopment (fig. 40) suggests that the aquifer system has adjusted to development through a combination of increased inflow, decreased outflow, and a reduction of the amount of water in storage. The major changes in the simulated water budget between 1900 and 1985 are summarized in figure 47. Simulated ground-water withdrawal during 1981–85 was 765 ft^3/s. Although the loss in ground-water storage during the period was substantial, averaging 235 ft^3/s, other adjustments in the flow system accounted for an additional 530 ft^3/s. Simulated downward leakage from the Upper Floridan aquifer and its hydraulic equivalents in South Carolina increased nearly 30 percent, from about 160 ft^3/s to about 240 ft^3/s. Vertical outflow to the Upper Floridan aquifer decreased about 95 ft^3/s, or about 20 percent. (The totals for the Upper Floridan aquifer include flow to and (or) from the clastic water-table aquifer in South Carolina, which is hydraulically equivalent to the Upper Floridan aquifer.) Reduction of ground water discharging to streams as base flow accounted for an additional 345 ft^3/s. Simulated lateral inflow increased, and outflow decreased, each only about 5 ft^3/s.

Figure 48 shows the areal distribution of simulated vertical flow across the regional confining units for the 1981–85 period. Comparison with figure 42 suggests that most of the induced downward leakage occurs updip, while most of the decrease in upward leakage occurs in the middip and downdip parts of the aquifer system.

Although simulated recharge to the outcrop area does not change with time during transient simulation, actual recharge to the deep flow regime through the outcrop may have increased during 1900–85, in response to increased gradients between shallow water-table aquifers and the deep pumped aquifers. However, field evidence indicates that hydraulic heads in the outcrop area were affected significantly by pumping in only a few localized places. Where heads in the deep aquifers of the outcrop area decreased in response to pumping, recharge to the deep aquifers undoubtedly increased. However, with the large grid blocks (64 mi^2) of the regional model, it was impractical to simulate the increases in recharge to the small areas of the outcrop affected by pumping. Nevertheless, the overall results of the model indicate that most of the flow adjustments in response to pumping during 1981–85 were adequately simulated with respect to the effect on the deep flow regime.

Figures 46, 47, and 49 show that the simulated loss of water from storage was 235 ft^3/s during 1981–85. A change in storage—in the model, as well as in the aquifer—results directly from head change. Mathematically, a change in storage represents the rate of head change, multiplied by the area over which the change occurs, multiplied by the applicable storage coefficient. Over the modeled system, the storage coefficient averages about 4×10^{-3}. The system underlies about 120,000 mi^2. These data, together with the simulated change in storage of 235 ft^3/s, indicate that during 1981–85 heads declined an average of about 0.5 ft/yr over the entire system. Although the hydrographs on plates 5, 6, and 7 suggest that within the major pumping centers head declines have recently averaged between 2 and 5 ft/yr, the average decline over the entire system was probably about 0.5 ft/yr. In other words, the simulated loss of ground water in storage (235 ft^3/s) appears to be consistent with observed conditions during 1981–85.

It should be recognized that to avoid serious, irreparable dewatering, the system must equilibrate with pumpage without continuing to release water from storage. However, this cannot occur as long as pumpage continues to exceed the capacity of the system to capture additional recharge or to reduce natural discharge. Owing to the coarse scale of the regional model, it was not used to test the response of the system to future pumping.

Base flow from the deep ground-water regime is simulated to have decreased from about 1,720 ft^3/s prior to 1900 to about 1,375 ft^3/s during 1981–85, which represents a reduction in the discharge of deep base flow totaling about 345 ft^3/s. This decrease in base flow indicates a reduction in hydraulic gradients between regional aquifers and major streams, owing to water-level decline caused by pumpage. Although field data are not available to check the simulated decreases in the discharge of deep base flow, the total simulated decrease

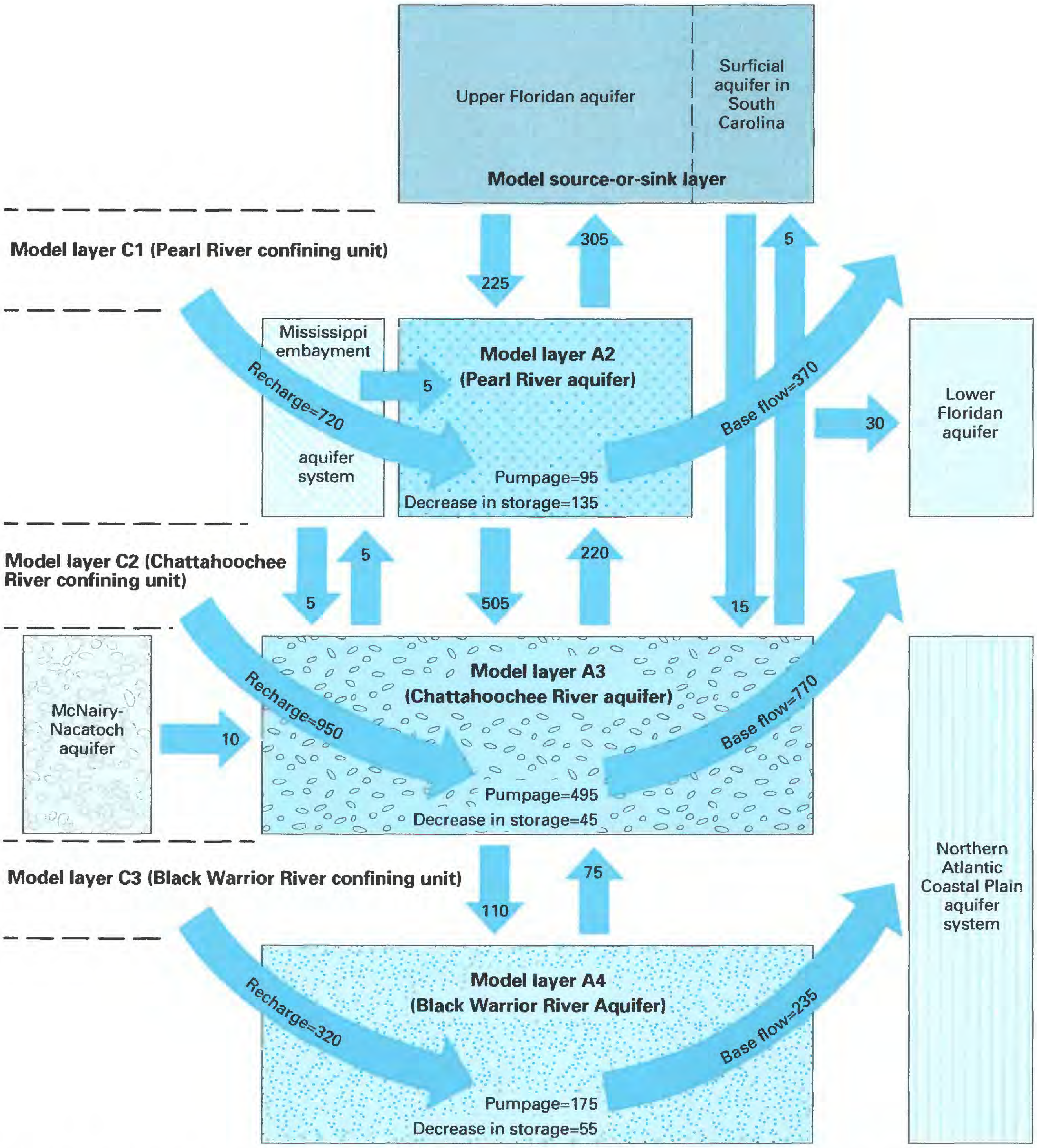


FIGURE 46.—Simulated water budget of 1981–85 conditions in the Southeastern Coastal Plain aquifer system. (In units of cubic feet per second, rounded to the nearest 5 cubic feet per second.)

in base flow (345 ft^3/s) is insignificant compared with the average discharge of each major stream in the study area. Of the decreases simulated for individual streams, the decreases exceed 25 ft^3/s at only three streams: the Tombigbee River (reduction of 79 ft^3/s), the Flint River (30 ft^3/s), and the Savannah River (36 ft^3/s). To compensate for errors in the calibration, the transient model may simulate too much, or too little, of the water that discharged to streams under predevelopment conditions. The simulated reduction of base flow appears to be reasonable, however, considering (1) the observed head decline in wells adjacent to the major rivers and (2) the relatively minor amount of simulated decreases in base flow compared with observed conditions of low flow.

Simulated changes in ground-water flow to and from the Southeastern Coastal Plain aquifer system are minor with regard to the McNairy-Nacatoch aquifer, the Mississippi embayment aquifer system, and the Northern Atlantic Coastal Plain aquifer system. Figures 40, 46, and 49 show that lateral inflow from the McNairy-Nacatoch aquifer (to model layer A3) increased 5 ft^3/s, while lateral outflow to the Northern Atlantic Coastal Plain aquifer system (from model layer A3) decreased 5 ft^3/s. Overall, inflow from all adjacent regional aquifers increased by 85 ft^3/s during 1900–85, while outflow decreased by 100 ft^3/s. The effect is equivalent to a source of 185 ft^3/s, with 175 ft^3/s of this coming from the Floridan aquifer system and the hydraulically equivalent surficial aquifer in South Carolina.

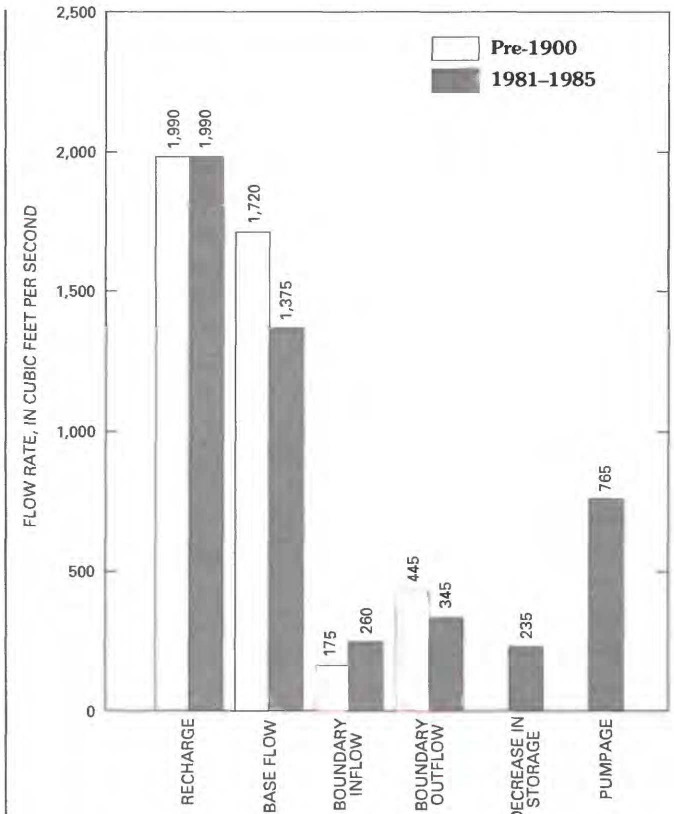


FIGURE 47.—Simulated predevelopment (pre-1900) and average 1981–85 water-budget elements.

MODEL SENSITIVITY

To help assess the importance of the uncertainty associated with the definition of parameters used in the model of the Southeastern Coastal Plain aquifer system, a sensitivity analysis was conducted on those parameters that were routinely adjusted during the calibration process. The sensitivity testing was made on four data sets calibrated in the steady-state model: outcrop recharge, aquifer transmissivity, streambed conductance, and confining-unit leakance. The sensitivity of the steady-state (predevelopment) model was also tested with respect to changes in the location of no-flow boundaries and to variation in the altitude of constant-head boundaries. A single data set, storage coefficient, was tested in the transient (1900–85) model. Sensitivity was measured by varying the model input through increments both greater than and less than the calibrated value of each parameter and observing the resultant change in simulated hydraulic head and base flow. Each parameter was tested independently of the others. The results of the sensitivity testing illustrate how the model performs using alternative distributions of input, as opposed to the calibrated input. The interested reader is referred to Pernik (1987) for a detailed description of the sensitivity analysis.

The steady-state model of predevelopment conditions is most sensitive to changes in the rates of recharge in the outcrop area. After recharge, the model is most sensitive to changes in transmissivity, especially in the interstream parts of the outcrop area, and is more affected by changes in updip transmissivity values than by changes in downdip transmissivity values. Near streams, the model is more sensitive to changes in streambed conductance than to changes in either leakance or transmissivity. Changes in leakance values have comparatively little effect on simulated base flow, but they more significantly affect simulated hydraulic heads, especially where the confining units are relatively leaky. The model is relatively insensitive to changes in the location of downdip no-flow boundaries and to moderate changes (±10 ft) in the altitude of constant-head boundaries. The transient model is less sensitive to increases in storage coefficient than to decreases.

The results of the sensitivity analysis show that the calibrated values of model input are, for the most part, internally consistent and within the range of reasonable possibilities. The simulated response to departures from the calibrated input suggests that the capacity of the model to simulate field conditions deteriorates as the departures increase. During the compilation of model input, an effort was made to keep the individual parameters realistic; this effort provided a consistency within the input data that became obvious during sensitivity

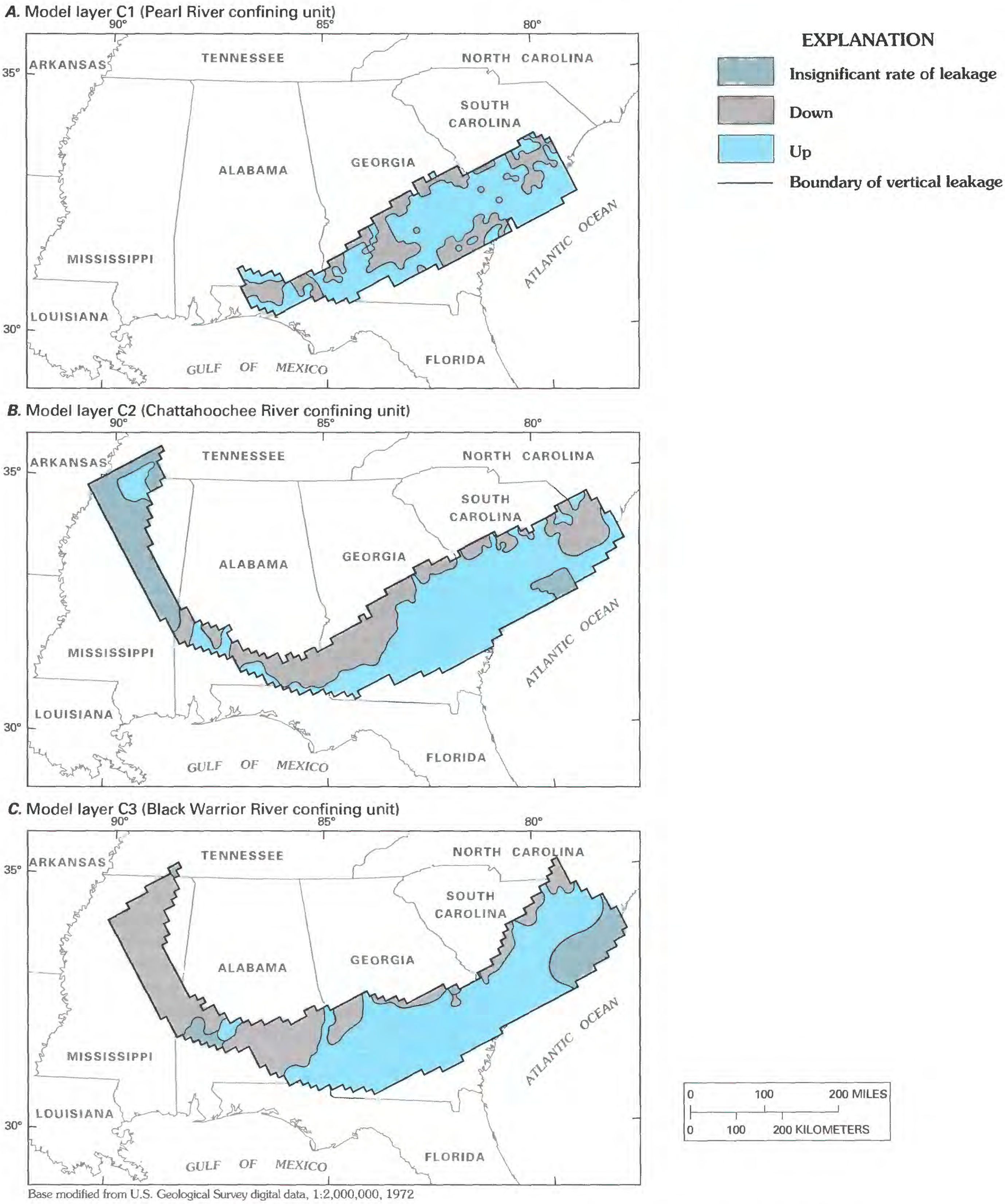


FIGURE 48.—Areal distribution and direction of simulated vertical leakage through confining units under 1981–85 conditions.

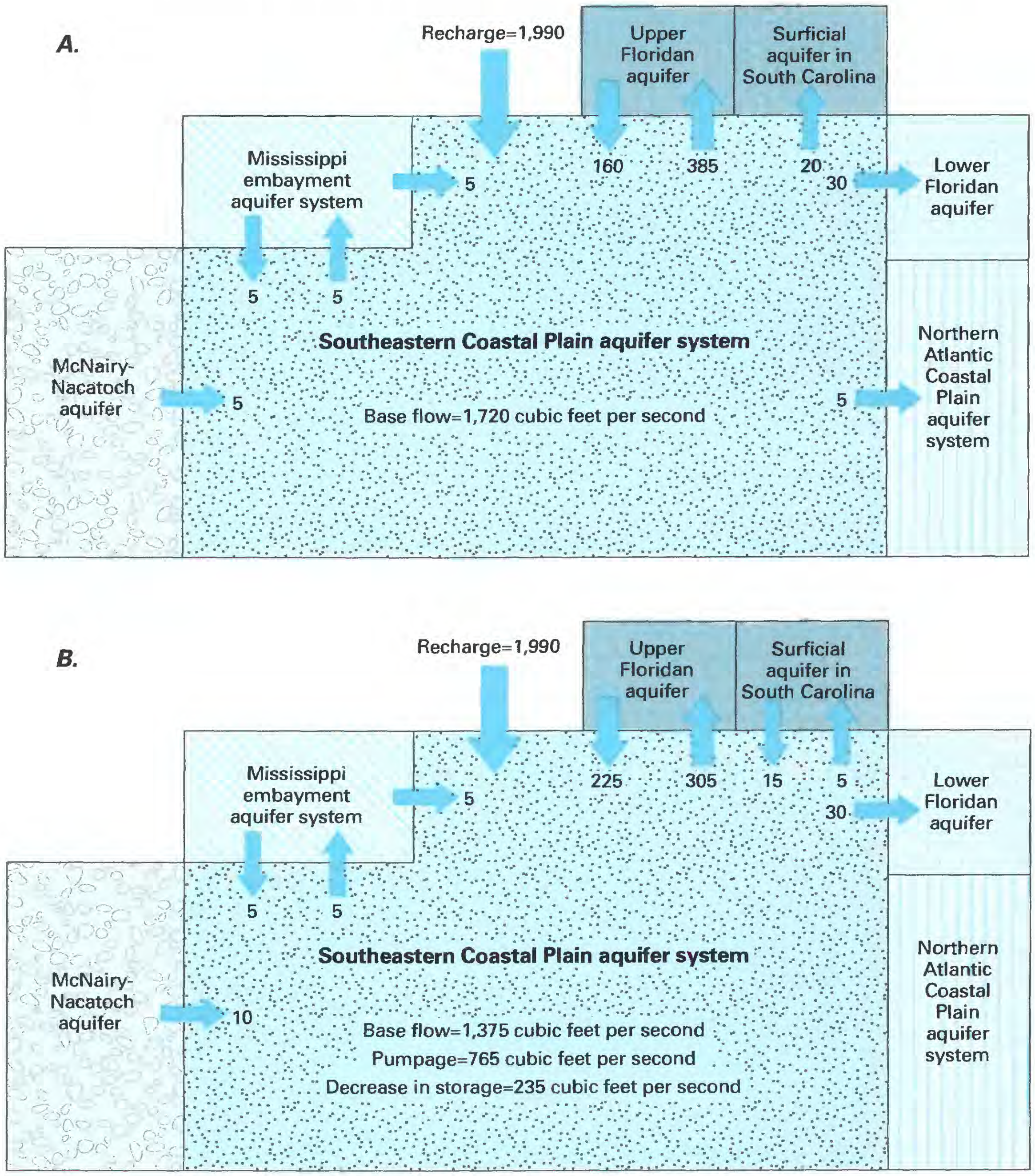


FIGURE 49.—Simulated rates of ground-water flow between the Southeastern Coastal Plain aquifer system and adjacent regional aquifers and aquifer systems, under (*A*) predevelopment (pre-1900) and (*B*) 1981–85 conditions (modified from Miller, 1992). (Units are rounded to the nearest 5 cubic feet per second.)

testing. For example, sensitivity tests showed that transmissivity must be altered unrealistically to compensate for the effects of small departures from the calibrated rates of recharge. The model appears to be only moderately sensitive to what is believed to be the maximum possibility of error in the downdip boundary conditions. This is due partly to the fact that the model is inherently insensitive to change of any kind where the transmissivity is small, which it is in most downdip parts of the aquifer system.

SUMMARY

The Southeastern Coastal Plain aquifer system underlies about 120,000 mi^2 of the Coastal Plain of the Southeastern United States in Mississippi, Alabama, Georgia, and South Carolina and small contiguous parts of Tennessee, Florida, and North Carolina. The aquifer system is composed predominantly of unconsolidated to semiconsolidated clastic rocks of Cretaceous and Tertiary age. The distribution and hydraulic properties of the aquifers and confining units are controlled by lithologic and structural patterns within rocks that dip and thicken coastward. The freshwater part of the aquifer system thickens from a featheredge along the inner margin of the Coastal Plain (Fall Line) to more than 1,000 ft on the landward side of a freshwater-saltwater interface. The freshwater is underlain updip by dense, relatively impermeable pre-Cretaceous rocks and downdip by Lower Cretaceous strata containing sluggishly moving saline water. The freshwater-saltwater interface corresponds generally with the downdip limits of permeability in Mississippi, Alabama, and Georgia and is generally less than 75 mi off the coast of South Carolina.

The total thickness of the Cretaceous and Tertiary rocks that make up the aquifer system ranges from less than 1,500 ft near the South Carolina-North Carolina State line to more than 10,000 ft in Alabama and Mississippi. Regional dips of the rocks range from generally less than 20 ft/mi in South Carolina to more than 75 ft/mi in downdip parts of western Alabama and eastern Mississippi. The aquifer system crops out in an arcuate pattern just downdip from the Fall Line. In addition to being hydraulically interconnected with carbonate rocks of the overlying Floridan aquifer system, the Southeastern Coastal Plain aquifer system merges with clastic aquifers of the Mississippi embayment and coastal lowlands aquifer systems on the west and with the Northern Atlantic Coastal Plain aquifer system on the northeast.

Ground-water conditions in the Southeastern Coastal Plain aquifer system result from a homoclinal Coastal Plain setting and a beveled outcrop area of hummocky relief that is subjected to a humid climate and is drained by an extensive surface-water network. Most recharge to the freshwater system occurs in the interstream parts of the outcrop area under predominantly unconfined conditions. Less water enters updip, confined areas of eastern Alabama, Georgia, and South Carolina as downward leakage from the Upper Floridan aquifer or its clastic (surficial aquifer) equivalent in South Carolina. Most discharge from the aquifer system occurs in the updip, unconfined parts of the system as base flow to streams in the outcrop area; smaller amounts of middip and downdip discharge occur in confined parts of the aquifer system as diffuse upward leakage to the Floridan aquifer system or its clastic equivalent, the surficial aquifer in South Carolina.

Topographic relief and sediment grain size generally decrease with distance from the Fall Line, and the concentration of dissolved solids in the ground water tends to increase with distance from the areas of recharge, which are generally in the outcrop area. As a result, hydraulic gradients and hydraulic conductivity values generally decrease from updip to downdip, and water density increases with distance from the outcrop area and with depth below land surface. Therefore, the energy available to circulate freshwater decreases as a function of depth below land surface and distance from the outcrop area.

Because water that infiltrates the dynamic flow regime of the shallow outcrop area cannot as readily penetrate the deeper, less dynamic parts of the regional flow system, a considerable amount of the interstream, updip recharge discharges directly to nearby streams. In contrast, middip flow is characterized by relatively long, nearly horizontal flow paths. Downdip flow typically is limited to sluggish diffuse upward leakage.

South of the Fall Line, precipitation increases from north to south and from east to west and averages about 51 in/yr across the outcrop area. Runoff (including base flow) generally increases from east to west; in southern Mississippi and Alabama, it decreases with distance from the Gulf of Mexico. Generally decreasing from updip to downdip, runoff in the outcrop area averages about 19 in/yr. Evapotranspiration averages about 32 in/yr over the outcrop area. Through a combination of overland flow and base flow, gaining streams in the Coastal Plain receive more than one-third of the water falling on the outcrop as precipitation; evapotranspiration removes practically all the remainder. Less than 1 percent of the water falling as precipitation in the outcrop area percolates from the unconfined parts of the outcrop area to the deep, confined parts of the aquifer system.

Runoff is greater in the western part of the outcrop area than in the eastern part. Owing to greater amounts of overland flow in the west, the residual between precipitation and runoff generally is less in Mississippi and Alabama than in Georgia and South Carolina. Although precipitation in Mississippi and Alabama is nearly 6 in/yr more than in Georgia and South Carolina, infiltration is greater in the outcrop area of eastern Alabama, Georgia, and South Carolina than in Mississippi and western Alabama. Accordingly, both recharge and base flow increase from west to east across the outcrop area. The west-to-east increases in infiltration, ground-water recharge, and base flow result from the generally coarser texture of the sediments of predominantly fluvial origin that crop out across much of eastern Alabama, Georgia, and South Carolina, compared with

the predominance of chalk, clay, and shale that crop out in Mississippi and western Alabama.

For modeling purposes, the Southeastern Coastal Plain aquifer system was subdivided conceptually into two flow regimes: (1) a shallow ground-water flow regime, limited to the outcrop area, that is under predominantly unconfined conditions and is drained by relatively small streams, and (2) a deep flow regime, existing everywhere in the subcrop and in the deepest parts of the outcrop, that is under predominantly confined conditions and is drained by the major rivers. The regional model was designed to simulate the deep, predominantly confined flow regime and its interaction with the major streams; the model does not directly simulate the shallow flow regime nor the discharge to small streams. By simulating only a fraction of the total recharge, the model simulates the net effect of the shallow conditions on the deep flow regime. The model thus accounts for all freshwater below the level of the shallow flow regime and all freshwater in the subcrop area.

To expedite model development, three categories of recharge and two levels of ground-water discharge were considered. Shallow recharge occurs when precipitation infiltrates the water table in unconfined parts of the outcrop area. About 90 percent of the shallow recharge becomes shallow base flow to streams, creeks, and lakes in the outcrop area—leaving about 10 percent to percolate deeper. Deep recharge, the residual of shallow recharge and shallow base flow, is water that percolates from unconfined to confined parts of the aquifer system. Most deep recharge occurs in the interstream areas of the major drainages. Most of the deep recharge eventually discharges as deep base flow to major rivers, which drain lower parts of the outcrop area. The difference between deep recharge and deep base flow is deep seepage. Deep seepage is the small fraction of deep recharge and the smaller fraction of shallow recharge that seeps below the most deeply entrenched streams in the Coastal Plain and recharges only the deepest parts of the aquifer system. The deep seepage eventually discharges as diffuse upward leakage, lateral outflow, or well pumpage.

Aquifers and confining units of the hydrogeologic framework were translated into layers of the finite-difference model. The model layers represent aquifers of gravel, sand, and minor amounts of limestone and confining units of clay, chalk, mudstone, or shale. A layer of constant-head nodes was used atop three actively simulated layers in the model to convey the effects of the Upper Floridan aquifer and a shallow water-table aquifer (the surficial aquifer in South Carolina) where they overlie the Southeastern Coastal Plain aquifer system. The three actively simulated layers represent the (1) Black Warrior River aquifer (at the bottom), (2) Chattahoochee River aquifer (middle), and (3) Pearl River aquifer (top).

The Black Warrior River aquifer is the basal and most extensive aquifer in the regional aquifer system. Although the aquifer extends through the subsurface from near the Mississippi-Tennessee border into North Carolina, it crops out only in Mississippi, Alabama, and a small part of western Georgia. In Mississippi and western Alabama, the aquifer consists mostly of Upper Cretaceous sands of fluvial and deltaic origin, but locally it contains hydraulically connected Lower Cretaceous sands. Sand and sandy clay of marine origin make up part of the aquifer in southern Alabama and southwestern Georgia. Toward the east, from east-central Georgia through South Carolina, the Black Warrior River aquifer consists of relatively thin, mostly nonmarine clastic rocks.

The Chattahoochee River aquifer is a thick, diverse group of rocks deposited under a wide range of conditions during Late Cretaceous to Paleocene time. Although the areas of highest permeability in this aquifer consist predominantly of coarse-grained quartz sand that in places is interbedded with gravel, there are local occurrences of limestone (Clayton Formation) near the upper part of the aquifer in Georgia that are especially permeable, owing to well-developed solution channels within the limestone. The Chattahoochee River aquifer is continuous from North Carolina to western Alabama. Although it is absent in east-central Mississippi, a westward extension is present in north-central Mississippi, where it is known locally as the Ripley Formation.

The Pearl River aquifer is predominantly a clastic sequence of unconsolidated to poorly consolidated sediments that for the most part were deposited under marine conditions, except in western Alabama and Mississippi, where the aquifer grades into a thick fluvial sequence. In western Alabama and Mississippi, the aquifer grades in a downdip direction from porous sand, sandstone, gravel, and minor occurrences of limestone to low-permeability clay, shale, mudstone, chalk, and chalky limestone near the fringes of the freshwater flow system. In southwestern South Carolina and southern Georgia, the aquifer grades into stratigraphically equivalent limestone and dolomite that make up the lower part of the highly permeable Floridan aquifer system. Likewise, in much of central and eastern Alabama, Georgia, and South Carolina, the Pearl River aquifer is immediately overlain by, and hydraulically connected to, the Floridan aquifer system. Although the limestone of the Lower Floridan aquifer is in places more permeable than the clastic strata, the carbonates and clastics merge hydraulically to form one more-or-less continuous hydrologic unit across eastern Alabama, Georgia, and South Carolina.

Transmissivity patterns in the Black Warrior River aquifer reflect the characteristics of a sedimentary sequence that deepens from east to west and grades from mostly coarse-grained, nonmarine deposits near the inner margin of the Coastal Plain to finer grained, marginal-marine or marine sediments toward the southern limits of the modeled area. Model-derived transmissivity values range from about 0.00001 to about 0.2 ft^2/s and average about 0.06 ft^2/s. The largest transmissivity values are in Mississippi and Alabama, where fluvial-deltaic sediments (making up the Eutaw, McShan, Gordo, and Coker Formations) are the thickest. The smallest transmissivity values are in east-central Georgia and South Carolina, where deposits (of the Atkinson and Cape Fear Formations) are much thinner and less permeable. Transmissivity values in most of Georgia range from about 0.01 to about 0.1 ft^2/s, reflecting a transitional depositional environment situated between a subsiding Mississippi embayment on the west and a persistently high Cape Fear Arch toward the northeast.

The model-derived transmissivity values for the Chattahoochee River aquifer range from about 0.00001 to about 0.6 ft^2/s and average about 0.1 ft^2/s. The wide-ranging transmissivity values reflect the effects of areal differences in both sediment thickness and lithology. The largest transmissivity values (0.1–0.6 ft^2/s) are in Georgia and west-central South Carolina. The largest values correspond to the area of greatest sediment accumulation, where thicknesses range from about 1,000 ft in western South Carolina to more than 1,500 ft in central Georgia. Moderately large transmissivity values (0.05–0.099 ft^2/s) in southeastern Alabama and Georgia are due to the thick accumulation of coarse clastic sediments that compose the Providence Sand, Ripley Formation, Cusseta Sand, and Blufftown Formation, in addition to the overlying, highly permeable sandy limestone of the Clayton Formation. Although the calcareous Clayton Formation, largely of marine origin, is relatively well connected hydraulically to underlying clastic sediments of the Chattahoochee River aquifer, it is separated from the overlying Floridan aquifer system (also limestone) by fine-grained sand and clay of the Porters Creek Formation. Despite a diminishing thickness of carbonate sediments in the Chattahoochee River aquifer east of Georgia, a band of relatively large transmissivity values (0.05–0.6 ft^2/s) extends through middip South Carolina, owing mainly to the large permeability associated with massively bedded, fluvial-deltaic quartz sand and gravel deposits of the highly productive Middendorf Formation. Moderately large transmissivity values (0.05–0.099 ft^2/s) in northern Mississippi and southwestern Tennessee correspond to marine sand of the Ripley Formation.

Model-derived transmissivity values for the Pearl River aquifer range from about 0.001 to about 0.6 ft^2/s and average about 0.07 ft^2/s. There is not as much variation in transmissivity within the Pearl River aquifer as in the Chattahoochee River and Black Warrior River aquifers. Part of this uniformity results from the predominantly marine character of the sediments in the Pearl River aquifer. The most permeable clastic sediments belong to updip parts of the Tallahatta and Barnwell Formations in Alabama, Georgia, and South Carolina. The areas of smallest transmissivity values (less than 0.05 ft^2/s) are in southeastern Alabama and western South Carolina, where the sequence is thinnest, and in updip Georgia and South Carolina, where the nonmarine or marginal-marine clastic sequence pinches out against older, marine sediments of the Chattahoochee River confining unit. Transmissivity values of 0.05 to 0.099 ft^2/s in a band trending southwest to northeast across southern Georgia correspond to a structural depression known as the Gulf Trough.

Storage coefficients derived from model calibration range from 1×10^{-4} to 1×10^{-1} and average about 4×10^{-3}. Storage coefficients in shallow, updip parts of the aquifer system reflect unconfined to semiconfined conditions and average about 1×10^{-2}. In middip and downdip areas, an interfingering of fluvial-deltaic sands with marine silts and clays inhibits the capacity of subcropping parts of the aquifer system to release stored water; storage coefficients there average about 1×10^{-3}.

Leakance values are generally the smallest in the Black Warrior River confining unit and largest in the Pearl River confining unit. The values range from about 5×10^{-16} per second in a deeply buried downdip area of South Carolina, where the confining sediments are virtually impermeable owing to their clayey consistency, to about 5×10^{-9} per second in shallow, updip areas of eastern Alabama, Georgia, and South Carolina, where the sediments are relatively leaky because of their sandy nature. Overall, the largest values of leakance are updip and the smallest values are downdip; this is in accordance with clay and silt accumulations that generally increase away from the inner margin of the Coastal Plain, causing permeabilities to decrease, owing to a general decrease in grain size and degree of sorting. Bands of relatively small leakance (less than 1×10^{-12} per second) in the deeper, downdip parts of South Carolina and Georgia can be attributed to moderately thick accumulations of calcareous clay, shale, mudstone, or marl of predominantly marginal-marine and marine origin. Relatively small values of leakance in Mississippi and Alabama correspond to thick chalk, shale, and clay facies of the Black Warrior River confining unit. Relatively large leakance values (greater than 1×10^{-11} per second) near the updip margins of the Black Warrior and Chattahoochee River confining units in Georgia and South Carolina correspond to highly oxidized, relatively coarse-grained deposits of sand. Because the Black Warrior River aquifer does not crop out in eastern Georgia and South Carolina, this aquifer is recharged there primarily by downward leakage through these updip sequences of sand.

Calibrated streambed conductances in the model range from about 0.005 to nearly 1.0 ft^2/s and average about 0.2 ft^2/s, suggesting that the vertical hydraulic conductivity values of the simulated streambeds average about 10^{-7} ft/s. In comparison, lateral hydraulic conductivity values for adjacent aquifers probably average between 10^{-4} and 10^{-5} ft/s, or about two to three orders of magnitude greater than the average vertical conductivity indicated for streambeds. The average streambed leakance of about 10^{-8} per second is about 10 times greater than calibrated leakance values for the most permeable parts of the confining units in the aquifer system.

Model results indicate that under predevelopment conditions, 1,720 ft^3/s of water (about 0.5 in/yr)—or 86 percent of the deep recharge of 1,990 ft^3/s (about 0.6 in/yr)—discharged as base flow to major streams in the outcrop area. The difference between the simulated base flow to the major rivers and the simulated deep recharge is 270 ft^3/s (about 0.1 in/yr). This residual, which represents the net effect of precipitation, evapotranspiration, and runoff in the outcrop area, migrated downgradient into the confined subcrop area as deep seepage. The deep seepage was joined by downward leakage, which is simulated to have come from updip parts of the Floridan and Mississippi embayment aquifer systems and totaled 165 ft^3/s (less than 0.1 in/yr). An additional 5 ft^3/s entered laterally from the Mississippi embayment aquifer system. A total of 440 ft^3/s (less than 0.2 in/yr) is simulated to have discharged from the subcrop of the Southeastern Coastal Plain aquifer system, of which 410 ft^3/s is diffuse upward leakage to the Floridan aquifer system, the surficial aquifer in South Carolina, and the Mississippi embayment aquifer system. Simulation indicates that under predevelopment conditions 30 ft^3/s discharged laterally to the Floridan aquifer system.

Recharge to the deep ground-water flow regime is substantially greater in eastern Alabama, Georgia, and western South Carolina than in western Alabama and Mississippi. The smallest rates of simulated deep recharge (less than 0.25 in/yr) are in Mississippi and western Alabama, where clay soils and low-permeability rocks in the subsurface limit infiltration. The areas of greatest simulated deep recharge (more than 1 in/yr) correspond to the sandy soils in southeastern Alabama, Georgia, and western South Carolina.

A total of 410 ft^3/s of water is simulated to have discharged vertically from the deep, confined part of the aquifer system under predevelopment conditions. This is nearly 2.5 times the amount (165 ft^3/s) that is simulated to have entered the deep flow regime vertically from overlying shallow aquifers. Owing to vertical gradients and confining-unit leakances that generally decrease in a downdip direction, the simulated rates of vertical discharge decrease from updip to downdip and reduce to negligible rates near the downdip limits of freshwater circulation.

Under predevelopment conditions, the rate of ground-water exchange with all adjacent regional aquifer systems—except the Floridan aquifer system—was negligible. The simulated predevelopment inflow from the Floridan aquifer system is 160 ft^3/s, all of which is downward leakage from the Upper Floridan aquifer. The simulated predevelopment outflow to the Floridan aquifer system (and the hydraulically equivalent clastic rocks of the surficial aquifer in South Carolina) totals 435 ft^3/s, of which 405 ft^3/s is diffuse upward leakage.

Model simulation indicates that the system has adjusted to ground-water development through a combination of increased inflow, decreased outflow, and decreased water in storage. Ground-water pumpage began about 1900 and increased to about 765 ft^3/s by 1985, causing head declines that in places exceed 150 ft. The resulting cones of depression and flow lines indicate that the most important changes in hydraulic head and flow direction have occurred in confined, middip areas, although heads have declined to some extent over most of the study area. This pattern of change results from generally larger storage coefficients updip (owing to coarser grained sediments and semiconfined conditions), coupled with a reduction in base flow; this combination has more effectively compensated for the effects of pumpage than have changes in the downdip vertical leakage. The simulated changes in storage, averaging 235 ft^3/s during 1981–85, are consistent with observed head declines, which during this period averaged between 2 and 5 ft/yr in the major pumping centers and about 0.5 ft/yr over the entire aquifer system. Base flow is simulated to have decreased from about 1,720 ft^3/s (pre-1900) to about 1,375 ft^3/s (during 1981–85). Most of the induced downward leakage is simulated to have occurred updip, while the reduced upward leakage occurs in middip and downdip areas.

Between 1900 and 1985, the inflow from all adjacent regional aquifer systems is simulated to have increased by 85 ft^3/s, while outflow to these aquifer systems declined by 100 ft^3/s. The simulated downward leakage from the Upper Floridan aquifer (and its clastic, surficial aquifer equivalent in South Carolina) increased nearly 30 percent, from about 160 ft^3/s to about 240 ft^3/s. Vertical outflow to the Upper Floridan aquifer decreased about 95 ft^3/s (about 20 percent). The net effect of all adjustments with respect to adjacent regional aquifer systems is equivalent to a source of 185 ft^3/s, of which 175 ft^3/s was contributed by the highly permeable Floridan aquifer system.

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