# THE PREDEVELOPMENT GROUND-WATER FLOW SYSTEM AND HYDROLOGIC CHARACTERISTICS OF THE COASTAL PLAIN AQUIFERS OF SOUTH CAROLINA

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## CONTENTS


| | Page |
|---|---|
| Abstract | 1 |
| Introduction | 1 |
| Geohydrologic framework | 2 |
| Coastal Plain aquifer units | 2 |
| Surficial aquifer | 2 |
| Floridan aquifer system | 2 |
| Tertiary sand aquifer | 7 |
| Black Creek aquifer | 7 |
| Middendorf aquifer | 7 |
| Cape Fear aquifer | 8 |
| Confining units | 8 |
| Conceptual model of the ground-water flow system prior to development | 9 |
| Modeling the ground-water flow system | 16 |
| Model introduction | 16 |
| Model boundaries | 16 |
| Model inputs | 21 |
| Model calibration | 21 |
| Sensitivity testing | 31 |
| Model-derived hydrologic characteristics | 38 |
| Summary and conclusions | 63 |
| Selected references | 64 |


## ILLUSTRATIONS


| | Page |
|---|---|
| Figure 1. Map of the location of the study area | 3 |
| 2. Map showing the relation between overlapping Regional Aquifer System Analysis models in South Carolina, Georgia, and North Carolina | 4 |
| 3. Generalized geohydrologic section A-A' | 4 |
| 4. Generalized geohydrologic section B-B' | 5 |
| 5. Map of generalized Coastal Plain aquifer outcrops | 5 |
| 6–9. Maps of the potentiometric surface of the: | |
| 6. Floridan aquifer system and the Tertiary sand aquifer prior to development | 10 |
| 7. Black Creek aquifer prior to development | 11 |
| 8. Middendorf aquifer prior to development | 12 |
| 9. Cape Fear aquifer prior to development | 13 |
| 10. Generalized depiction of a stratified ground-water flow system | 14 |
| 11. Map of model grid and specified boundary conditions | 17 |
| 12. Schematic geohydrologic section | 19 |

## ILLUSTRATIONS—Continued


Page

13. Conceptualization of model layers . . . . . . . . . . . . . . . . . . . . . . . . . . . 20

14–17. Maps of the observed and simulated water levels for the:
 14. Floridan aquifer system and the Tertiary sand aquifer . . . . . . . . . . . . . . . . 23
 15. Black Creek aquifer . . . . . . . . . . . . . . . . . . . . . . . . . . . . 25
 16. Middendorf aquifer . . . . . . . . . . . . . . . . . . . . . . . . . . . . 27
 17. Cape Fear aquifer . . . . . . . . . . . . . . . . . . . . . . . . . . . . 29

18–22. Graphs showing relation between changes in magnitude of input parameters and average absolute head residual per grid block for the:
 18. Floridan aquifer system and the Tertiary sand aquifer . . . . . . . . . . . . . . . . 32
 19. Black Creek aquifer . . . . . . . . . . . . . . . . . . . . . . . . . . . . 33
 20. Middendorf aquifer . . . . . . . . . . . . . . . . . . . . . . . . . . . . 34
 21. Cape Fear aquifer . . . . . . . . . . . . . . . . . . . . . . . . . . . . 35
 22. All aquifers . . . . . . . . . . . . . . . . . . . . . . . . . . . . 36

23. Graph showing relation between changes in magnitude of input parameters and total river flow . . . . . . . . . . . . . . . . . . . . . . . . . . . . 37

24–27. Maps of simulated transmissivity of the:
 24. Floridan aquifer system and the Tertiary sand aquifer . . . . . . . . . . . . . . . . 39
 25. Black Creek aquifer . . . . . . . . . . . . . . . . . . . . . . . . . . . . 41
 26. Middendorf aquifer . . . . . . . . . . . . . . . . . . . . . . . . . . . . 43
 27. Cape Fear aquifer . . . . . . . . . . . . . . . . . . . . . . . . . . . . 45

28–31. Maps of simulated leakage coefficient of the confining unit between the:
 28. Surficial aquifer and underlying units . . . . . . . . . . . . . . . . . . . . . . . . 49
 29. Floridan aquifer system/Tertiary sand aquifer and the Black Creek aquifer . . . . 51
 30. Black Creek aquifer and the Middendorf aquifer . . . . . . . . . . . . . . . . . 53
 31. Middendorf aquifer and the Cape Fear aquifer . . . . . . . . . . . . . . . . . . . 55

32. Map of simulated average recharge to and discharge from the deep regional ground-water flow system . . . . . . . . . . . . . . . . . . . . . . . . . . . . 59

33. Diagram of the model-simulated water budget for the deep flow system . . . . . . . . . 61


## TABLES


Page

Table 1. Generalized geohydrologic correlation table . . . . . . . . . . . . . . . . . . . . . . . . 6

2. Mean absolute differences and standard deviations calculated from simulated water levels and water levels contoured from field measurements . . . . . . . . . . . . . . . . 22

3. Comparison of observed and simulated flows from aquifers to rivers . . . . . . . . . . . . 31

4. Summary of model inflow and outflow . . . . . . . . . . . . . . . . . . . . . . . . 62

# THE PREDEVELOPMENT GROUND-WATER FLOW SYSTEM AND HYDROLOGIC CHARACTERISTICS OF THE COASTAL PLAIN AQUIFERS OF SOUTH CAROLINA

By Walter R. Aucott


## ABSTRACT

*A quasi-three-dimensional digital ground-water flow model was constructed as a part of the Regional Aquifer System Analysis program of the U.S. Geological Survey to describe flow in the Coastal Plain aquifers of South Carolina and parts of Georgia and North Carolina prior to development. The finite difference model simulated deep regional ground-water flow, which does not include local shallow flow systems, and consists of 5 layers and a uniform square 48 by 63 grid of 4 miles on a side.*

*Simulations indicate that total recharge and discharge in the deep ground- water flow system was 825 cubic feet per second prior to development. Simulated direct recharge in outcrop areas is 789 cubic feet per second. The remainder of total recharge is from leakage from overyling source-sink beds (15 cubic feet per second) and inflow across boundaries (21 cubic feet per second). The principal discharge from the ground-water flow system is to the large upper Coastal Plain rivers (735 cubic feet per second). Because smaller rivers are a part of shallow flow systems not the deep regional flow system, they were not considered in this simulation. The remaining discharge is by upward leakage to the overlying source-sink beds (64 cubic feet per second) and by outflow across lateral boundaries (26 cubic feet per second).*

*Model-simulated transmissivities range from less than 1,000 feet squared per day near the updip limit of most aquifers in the study area to a high of about 30,000 feet squared per day in the Middendorf aquifer in the Savannah River Plant area. Vertical hydraulic conductivities of confining units ranged from about $2x10^{-7}$ feet per day for the confining unit between the Cape Fear and Middendorf aquifers in the eastern part of the lower Coastal Plain to $5x10^{-2}$ feet per day for most of the confining units near their updip limits.*


## INTRODUCTION

The U.S. Geological Survey is conducting a series of investigations of regional aquifers throughout the United States as a part of the Regional Aquifer Systems Analysis (RASA) program. These studies provide a comprehensive understanding of ground-water availability throughout the Nation. The flow systems of the Coastal Plain aquifers in South Carolina are being studied as a part of this program.

This report describes the ground-water flow system of the Coastal Plain aquifers of South Carolina prior to development. A ground-water flow model was developed to aid in understanding and describing the flow system of the Coastal Plain aquifers. This model, constructed for predevelopment steady-state conditions, is discussed in detail in this report.

An understanding of the predevelopment flow system will aid in understanding the present-day flow system, the changes that have occurred because of manmade stress, and the effective development of the ground-water resources. Effective development is important because of the increasing use of ground water and because of the dependence of many users on ground water.

The study area encompasses the Coastal Plain of South Carolina and adjacent areas in Georgia, North Carolina, and offshore. The Coastal Plain of South Carolina covers about 20,000 square miles in the southeastern two-thirds of the State. It has been subdivided into the upper Coastal Plain and the lower Coastal Plain (fig. 1) on the basis of ground-water flow system characteristics and aquifer discharge to streams.

Previous investigations of most of the South Carolina Coastal Plain ground-water flow system have been either of a local or rather general nature. Ground-water flow models by Counts and Krause (1976), Bush (1982), Krause (1982) and Randolph and Krause (1984) have described flow in the Floridan aquifer system, composed of carbonate rock of Tertiary age present in southwestern South Carolina. The availability of detailed geohydrologic frameworks (Colquhoun and others, 1983; Renken, 1984; Aucott, Davis, and Speiran, 1986), comprehensive statewide potentiometric maps (Aucott and Speiran, 1985b), and a statewide evaluation of aquifer parameters (Aucott and Newcome, 1986) now enables more detailed and comprehensive descriptions of the hydrologic system to be made in South Carolina. Ground-water flow models have also been constructed for the Coastal Plain of Georgia and North Carolina as well as a regional model for the Southeastern United States in conjunction with the RASA program of the U.S. Geological Survey (fig. 2).

## GEOHYDROLOGIC FRAMEWORK

The Coastal Plain province is underlain by a wedge of unconsolidated to poorly consolidated sand, clay, and limestone sediments of Late Cretaceous and younger ages deposited on consolidated pre-Cretaceous metamorphic, igneous, and sedimentary rocks. The sedimentary wedge thickens from the Fall Line toward the present-day shoreline, and can be divided into a series of aquifers and confining units based on the relative permeability, areal extent, and lithologic continuity of the sediments.

The aquifers consist of layers of sand or high-permeability limestone separated by confining layers of clay, silt, or low-permeability limestone. Water generally moves laterally within each of the aquifers. The confining units inhibit, but do not prevent, the vertical movement of water between aquifers.

A regional framework for the aquifers of the Southeastern Coastal Plain has been developed during previous work on the carbonate Floridan aquifer system (Miller, 1986) and in preliminary work on the sand aquifers (Renken, 1984). The regional framework has been modified in South Carolina by subdividing some of the regional aquifers into a framework that better represents the hydrology of the aquifers in the study area and takes into account differences in data density and scale (Aucott, Davis, and Speiran, 1987). More detailed discussions of the geohydrology of the Coastal Plain aquifers can be found in the above reports.

Six water-bearing units comprise the Coastal Plain aquifer system of South Carolina. From youngest to oldest these are the surficial aquifer, Floridan aquifer system, Tertiary sand aquifer, Black Creek aquifer, Middendorf aquifer, and Cape Fear aquifer. These aquifers are generally associated with a geologic formation or group of formations as indicated in table 1. This association is general because formational descriptions are frequently local in scope and because an aquifer may contain parts of other formations. Generalized geohydrologic sections (figs. 3 and 4) and a generalized outcrop map (fig. 5) show the relations of the aquifers and the confining units that separate them.

### Coastal Plain Aquifer Units

#### Surficial Aquifer

The surficial aquifer consists of marine terrace deposits. These sediments are generally less than 40 feet thick and consist primarily of sand, shell, and clay that were deposited in a series of transgressions and regressions of the sea during the Pleistocene Epoch (Siple, 1946). The surficial aquifer is present throughout the lower Coastal Plain and contains water under unconfined conditions. It overlies the Floridan aquifer system in the western part of the lower Coastal Plain and the Black Creek aquifer in the eastern part of the lower Coastal Plain.

#### Floridan Aquifer System

The term Floridan aquifer system was applied by Miller (1986) to rocks previously called the Tertiary limestone aquifer and the principal artesian aquifer in parts of South Carolina and Georgia. The Floridan aquifer system in South Carolina generally consists of white to creamy-yellow limestone of late to middle Eocene age. The sediments comprising this system are parts of the Ocala Limestone, where present, and the underlying Santee Limestone. The Floridan aquifer system as defined by Miller (1986) throughout the Southeastern United States has

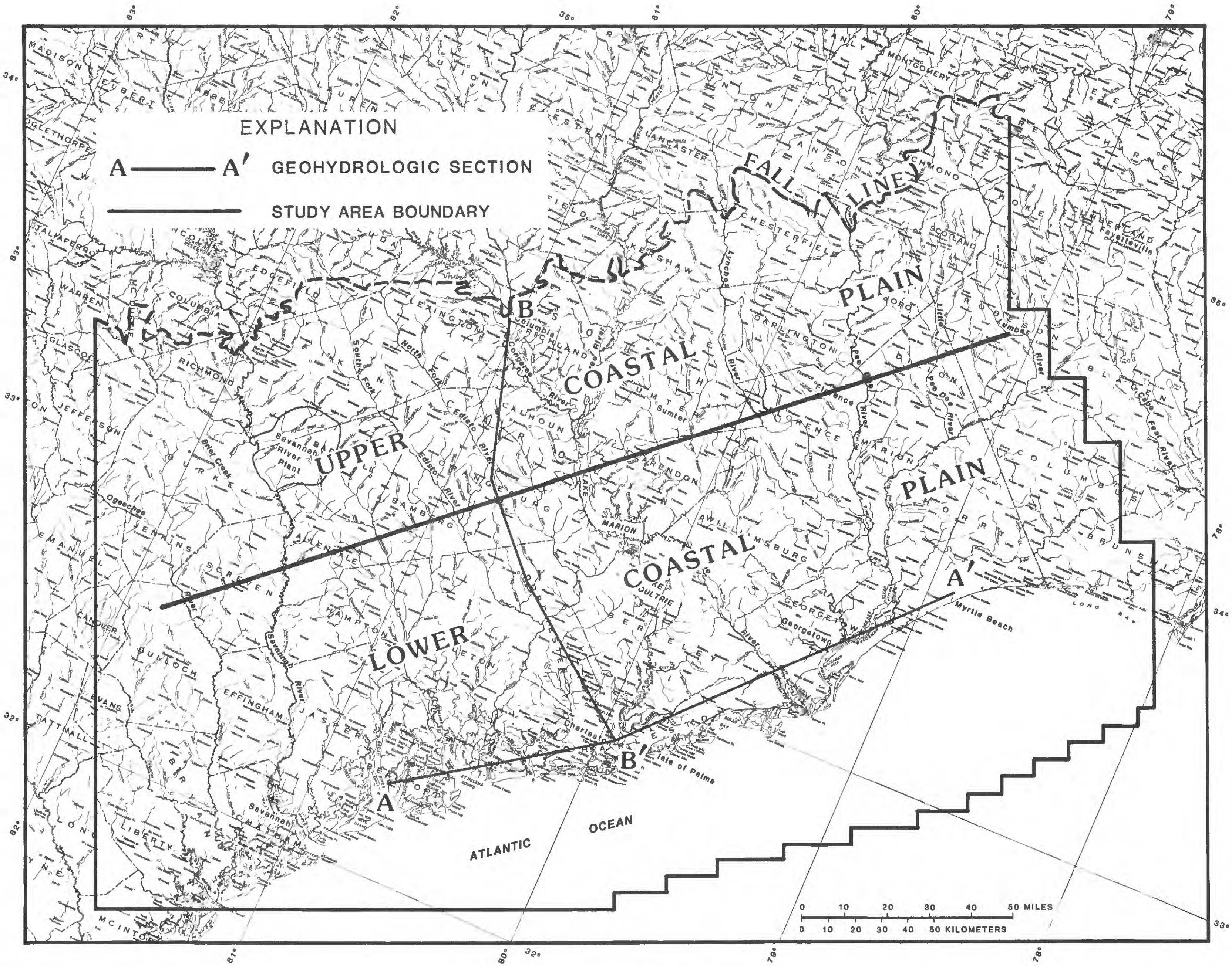


Figure 1.—Location of the study area.

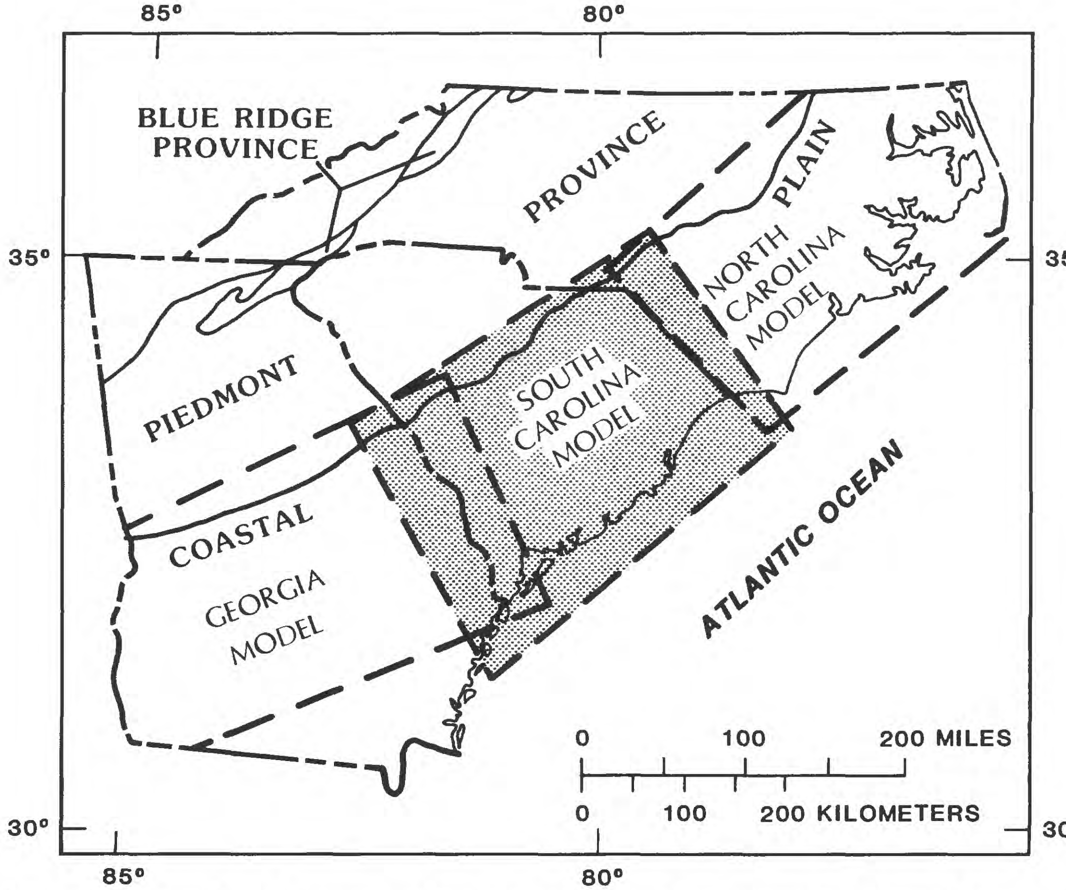


Figure 2. – The relation between overlapping Regional Aquifer System Analysis models in South Carolina, Georgia, and North Carolina.

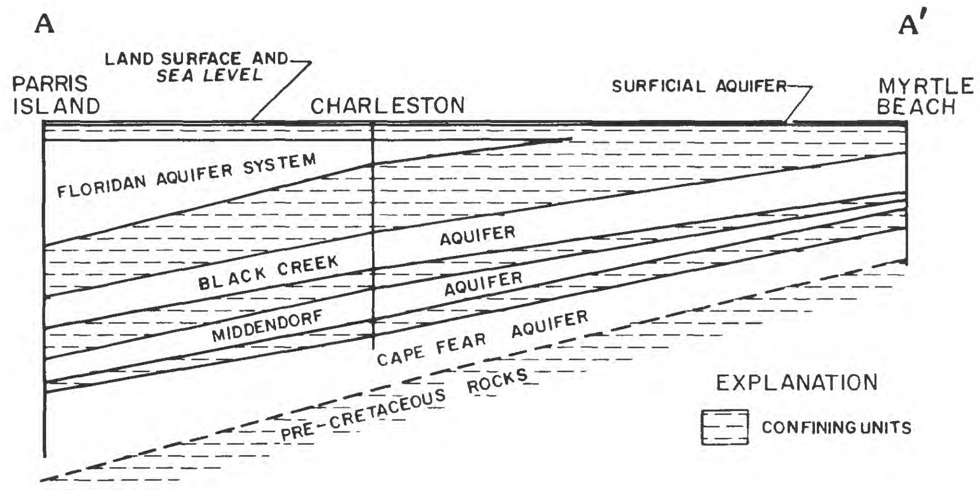


Figure 3. – Generalized geohydrologic section A–A'. (Modified from Aucott and Speiran, 1985a.)

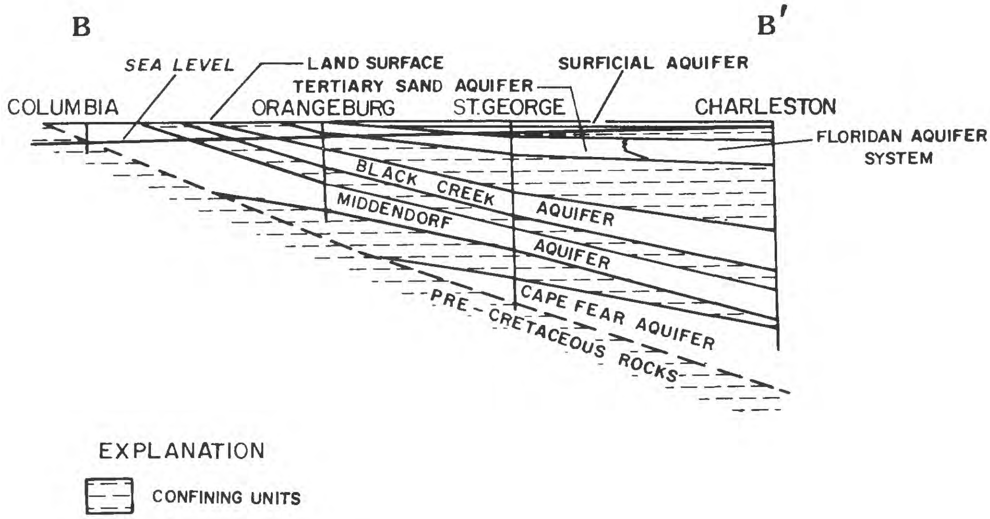


Figure 4. – Generalized geohydrologic section B–B'. (Modified from Aucott and Speiran, 1985a.)

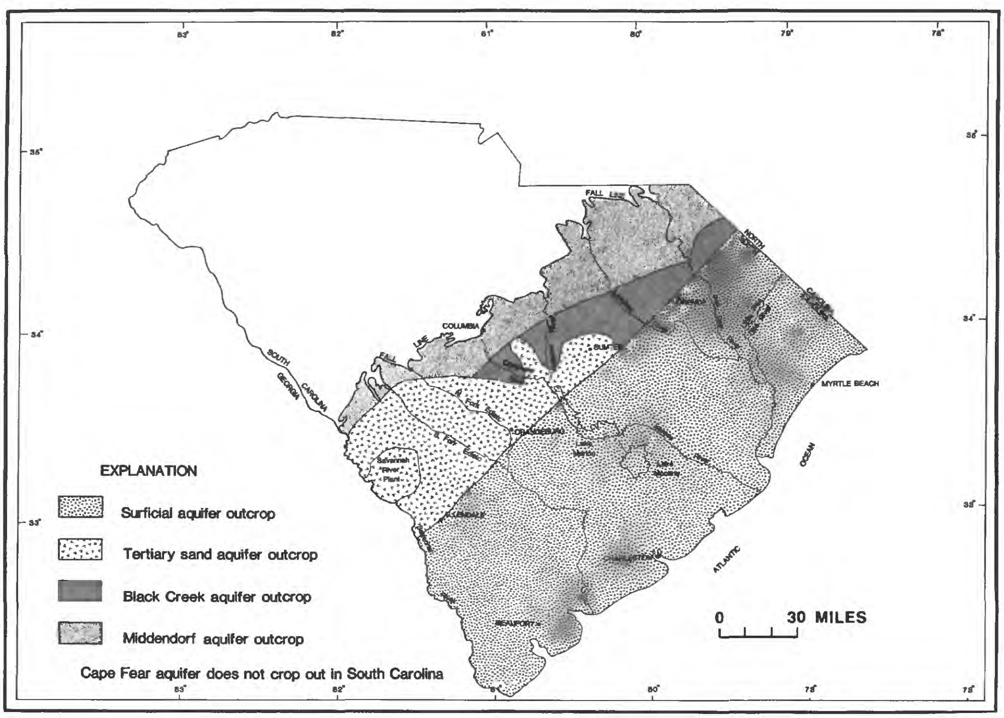


Figure 5. – Generalized Coastal Plain aquifer outcrops. (Modified from Colquhoun and others, 1983; and Cooke, 1936.)

Table 1.—Generalized geohydrologic correlation chart

[Modified from Siple, 1959]

| Aquifer | System | Geologic formation[1] | Description |
|---|---|---|---|
| Surficial aquifer | Quaternary | Coastal terrace deposits | Sand and clay, reddish-brown, orange and white. |
| Floridan aquifer system[2] (downdip) | Tertiary | Cooper Group (lower part) | Limestone and marl, gray to white, silty to sandy, phosphatic. |
| | | Ocala Limestone | Limestone, white to cream, calcitized, fossiliferous, glauconitic |
| | | Santee Limestone | Limestone, white to creamy yellow, fossiliferous, glauconitic; interbedded in part with gray to yellow sandstone. |
| Tertiary sand aquifer (updip) | Tertiary | Barnwell Formation | Sand, red to brown, fine- to coarse-grained massive. |
| | | McBean Formation | Sand, green to yellow, fine-grained, galuconitic; gray-green glauconitic marl. |
| | | Congaree Formation | Sand and sandstone, yellowish-brown to green, fine- to coarse-grained, quartzose, glauconitic; dark green to gray clay. |
| | | Black Mingo Formation (upper part) | Shale, gray, sandy; black sandy limestone, may be carbonaceous and fossiliferous in places. |
| Black Creek aquifer | Cretaceous | Black Creek Formation | Sand, gray to white, quartzose, calcareous, micaceous, phos-- phatic, glauconitic; dark gray to black thinly laminated clay containing nodules of pyrite and marcasite and fragments of lignite. Locally may include parts of overlying Peedee Formation. |
| Middendorf aquifer | Cretaceous | Middendorf Formation | Sand, light-gray, fine- to coarse-grained, micaceous, glauconitic, and in part calcareous; green, purple, and maroon clay; greenish-gray micaceous silty sandstone. |
| Cape Fear aquifer | Cretaceous | Cape Fear Formation | Clay, reddish-brown, gray to green; yellow to white fine- to coarse-grained sand with traces of mica. |

[1]These are geologic formations that are generally associated with a given aquifer. However, a given aquifer may not consist of the same formations in all areas, and locally an aquifer may include parts of additional formations not listed.

[2]Carbonate equivalent of the Tertiary sand aquifer.

been expanded and redefined by Aucott, Davis, and Speiran (1987) in South Carolina to include the permeable parts of the Santee Limestone in outcrop and in the subsurface. The Floridan aquifer system as redefined extends over the southwestern quarter of the Coastal Plain of South Carolina.

The hydraulic characteristics of the Floridan aquifer system result from both primary and secondary porosity. Primary porosity is high in the upper part of the Ocala Limestone as a result of the presence of coquina. Secondary porosity that results from the dissolution of calcium carbonate is locally present but is not as significant in the Floridan aquifer system in South Carolina as in the cavernous limestones found in some parts of Florida. The hydraulic conductivity of the Santee Limestone is generally lower than that of the Ocala because the clay content of the Santee is greater. The net result is that the transmissivity of the Floridan aquifer system is higher in the southern part of the State where the Ocala Limestone is present than in areas where it is not present (Hayes, 1979).

### Tertiary Sand Aquifer

The Tertiary sand aquifer is the clastic facies equivalent of the Floridan aquifer system and extends from near the Fall Line to the sand-limestone interface in the vicinity of the updip limit of the Floridan aquifer system. The Tertiary sand aquifer previously has been informally designated as aquifer A2 (Renken, 1984; Aucott and Speiran, 1985b). Sediments comprising the Tertiary sand aquifer include the Barnwell, McBean, and the Congaree Formations and the upper part of the Black Mingo Formation. Sediments from these Eocene and upper Paleocene formations have been considered together because they act hydraulically as a single aquifer on a regional scale. This is indicated by the general lack of a significant vertical hydraulic gradient between these formations except in some areas adjacent to Georgia and near the Fall Line. The Tertiary sand aquifer and the Floridan aquifer system can be treated as a single hydrologic unit in the study area because there are no regionally significant water-level differences between them and there is little evidence of an intervening confining unit (Aucott and Speiran, 1985b).

### Black Creek Aquifer

The Black Creek aquifer consists mostly of sediments of the Black Creek Formation and its equivalents but may locally include sediments that are part of the overlying Peedee Formation or the underlying Middendorf Formation. Sediments comprising the Black Creek Formation are principally thin laminated layers of dark gray to black clay and gray micaceous sand. The Black Creek aquifer is the uppermost regional aquifer consisting of sediments of Cretaceous age. This aquifer has been informally called aquifer A3a2 (Renken, 1984; Aucott and Speiran, 1985b). The updip limit of the Black Creek aquifer is located in the upper Coastal Plain and generally parallels the Fall Line. This aquifer crops out in the eastern part and subcrops in the western part of its updip limit.

In the eastern part of the Coastal Plain, the hydraulic conductivity of the Black Creek aquifer is relatively uniform. As a result, the transmissivity becomes greater as the aquifer thickens toward the coast, then remains fairly constant where the aquifer thickness is generally constant (Aucott and Newcome, 1986). In the western part of the upper Coastal Plain, transmissivities are relatively high because of the coarse sand and low clay content of the aquifer there. In the southern part of the study area, the hydraulic conductivity and transmissivity of the aquifer appear to be much lower than in the upper Coastal Plain or along the coast to the east because of the higher clay content (Aucott, Davis, and Speiran, 1987).

### Middendorf Aquifer

The Middendorf aquifer consists mostly of sediments of the Middendorf Formation but may locally include sediments of the overlying Black Creek Formation or the underlying Cape Fear Formation. This aquifer has previously been referred to as all or part of the "Tuscaloosa aquifer" (Siple, 1967; Park, 1980) and informally as aquifer A3a3 (Renken, 1984; Aucott and Speiran, 1985b).

In both outcrop and subsurface in the upper Coastal Plain, sediments of this aquifer are primarily light-gray, white-to-buff sand frequently interfingered with lenses of white, pink, or purple clay that were deposited in an upper delta-plain environment. In the lower Coastal Plain, the sediments of the Middendorf aquifer are lithologically similar to those of the Black Creek aquifer. The Middendorf aquifer exists throughout the Coastal Plain of South Carolina (fig. 5) and crops out along the Fall Line except locally in the western part of the Coastal Plain.

The transmissivity of the Middendorf aquifer varies in a pattern of bands that are approximately parallel to the Fall Line. The hydraulic conductivity of the Middendorf sediments in the upper Coastal Plain remains relatively constant for some distance from the Fall Line. Because the aquifer thickens away from the Fall Line and toward the coast, the transmissivity generally increases coastward. Transmissivity of the Middendorf aquifer is greatest in a band approximately parallel to the Fall Line in the lower part of the upper Coastal Plain (Siple, 1957). In this band, aquifer thickness remains constant but hydraulic conductivity and transmissivity are greater on the west side of the band because the aquifer contains coarse sand and little clay there.

In the lower Coastal Plain the hydraulic conductivity of the Middendorf aquifer generally decreases toward the coast as the percentage of clay in the aquifer increases. This results in a general decrease in transmissivity despite a small increase in thickness in most areas. Despite this decrease the transmissivity of the Middendorf aquifer near the coast is as great or greater than that of adjacent aquifers. This is indicated by lithologic data, aquifer test data, and by the fresher quality of the water, which indicates more complete flushing of the Middendorf aquifer than adjacent aquifers (Aucott and Newcome, 1986; Aucott and Speiran, 1986).

### Cape Fear Aquifer

The Cape Fear aquifer consists of part of the Cape Fear Formation and is the basal aquifer in the Coastal Plain aquifer system of South Carolina. It has informally been referred to previously as aquifer A4 (Renken, 1984; Aucott and Speiran, 1985b), as the lower part of the Middendorf aquifer (Colquhoun and others, 1983), or as the Middendorf aquifer (Zack, 1977). It consists predominantly of sand, silt, and gravel separated by relatively thick silt and clay layers. The extent of this aquifer has not been well defined in the Coastal Plain of South Carolina, but it probably occurs only in the lower Coastal Plain and the eastern part of the upper Coastal Plain and is not known to crop out in South Carolina. The extent of the Cape Fear aquifer is rather poorly defined because few wells penetrate it, because it is deeper and has poorer aquifer characteristics than the overlying Middendorf aquifer, and because it contains poor quality water in the lower Coastal Plain.

## Confining Units

Much less is generally known about the hydraulic characteristics of the confining units than is known about the aquifers of the Coastal Plain. Vertical movement of water within the system is controlled by the confining units, which are the hydrologic units of lowest permeability. Vertical movement is usually controlled by the least permeable layer in a confining unit which is typically a tight marine clay in the Coastal Plain sediments of South Carolina. All of the confining units identified allow limited movement of water through them.

The confining unit between the surficial aquifer and underlying aquifers is not comprised of a single formation. As such, its characteristics probably vary significantly. Where the surficial aquifer is underlain by the Cooper Group, in the central and eastern part of the lower Coastal Plain (Colquhoun and others, 1983), this confining unit can be expected to be very effective in inhibiting the vertical movement of water through it. Even within the coastal terrace deposits, many discrete layers of clayey material occur in the surficial aquifer locally creating artesian conditions in the surficial aquifer at shallow depths below the water table.

The confining unit that seems to have the greatest effect on the Coastal Plain flow system consists of clayey Paleocene sediments. This confining unit is located below the Floridan aquifer system and the Tertiary sand aquifer and above the Black Creek and Middendorf aquifers and is typically 40 to 100 feet thick.

The effectiveness of a confining unit in inhibiting vertical flow can be an important factor in differences in the flow systems of the aquifers immediately above and below it and in water level and water quality differences between these aquifers. The Paleocene clay is so effective in inhibiting the vertical movement of water in the southern part of the Coastal Plain that the flow system in the aquifers of Tertiary age and the flow system of the Black Creek aquifer differ more than the flow systems of any other adjacent aquifers (Aucott and Speiran, 1985a). Where this unit thins and becomes sandier northwestward toward its updip limit, it becomes less effective in inhibiting the vertical movement of water through it.

The confining unit between the Black Creek aquifer and the Middendorf aquifer, which consists primarily of sandy clay in the lower part of the Black Creek Formation, does not appear to be as effective in inhibiting vertical flow as are other confining units in the system. Under unstressed conditions, the flow systems of the Black Creek and Middendorf aquifers appear to be quite similar (Aucott and Speiran, 1985b). However, under stressed conditions significant water-level differences exist. Significant water-quality differences between these aquifers also occur in some areas.

The confining unit between the Middendorf and Cape Fear aquifers is very effective in separating the flow systems of these aquifers in the eastern part of the study area. This separation is suggested by important differences in water quality and water levels between these aquifers (Aucott and Speiran, 1985b, 1986). In the western part of the lower Coastal Plain and eastern part of the upper Coastal Plain, the effectiveness of this confining unit is questionable because limited data indicate that water-quality differences appear to exist but water-level differences appear to be minor.

## CONCEPTUAL MODEL OF THE GROUND-WATER FLOW SYSTEM PRIOR TO DEVELOPMENT

The ground-water flow system of the Coastal Plain aquifers in the study area can be best described with the aid of potentiometric maps of the aquifers. Figures 6, 7, 8, and 9 are potentiometric maps with flow lines for the combined Floridan aquifer system and the Tertiary sand aquifer, the Black Creek aquifer, the Middendorf aquifer, and the Cape Fear aquifer, respectively. These maps were developed by Aucott and Speiran (1985b) for South Carolina and extended into Georgia and North Carolina by using data from R. Faye (U.S. Geological Survey, written commun., 1984) and U.S. Geological Survey and North Carolina Department of Natural and Economic Resources file data. These maps depict the predevelopment potentiometric surfaces which are defined as the long-term average potentiometric surfaces that existed under natural conditions, prior to manmade stress on the aquifers. A potentiometric map of the surficial aquifer is not presented because of the localized nature of its flow system.

The potentiometric map of the Cape Fear aquifer (fig. 9) has been extended from Aucott and Speiran (1985b, sheet 5) into the eastern part of the upper Coastal Plain. The little data that are available for the Cape Fear aquifer in this area indicates that the hydraulic connection between the Cape Fear and Middendorf aquifers is relatively good in the eastern part of the upper Coastal Plain. The data also indicate that there is a small downward head gradient in interstream areas and probably a small upward gradient in stream valleys. The potentiometric surface of the Cape Fear aquifer in this area was derived mostly from the potentiometric surface of the Middendorf aquifer which has much more data.

The major source of recharge to the Coastal Plain aquifers is precipitation in their outcrop areas (fig. 5). Recharge in the updip interstream areas results in potentiometric highs such as those in the Tertiary sand aquifer, the Black Creek aquifer, and the Middendorf aquifer near their updip limits (figs. 6, 7, and 8).

Leakage through confining units between aquifers also is an important mechanism for recharge in the flow system. Downward leakage in the upper Coastal Plain (for example in some parts of the Savannah River Plant) provides an important source of recharge to the Black Creek aquifer and to the Middendorf aquifer. This recharge occurs because of the relatively high permeability of the confining units and a downward potentiometric gradient. In the western part of the upper Coastal Plain of South Carolina, downward leakage from the Tertiary sand aquifer is the principal source of recharge to the Black Creek and Middendorf aquifers. This is because the Black Creek and Middendorf aquifers crop out primarily in stream valleys, which are areas of aquifer discharge.

Discharge from the Tertiary sand aquifer, the Black Creek aquifer, and the Middendorf aquifer is primarily to rivers in the vicinity of the aquifer outcrops in the upper Coastal Plain. Upstream bending of the potentiometric contours in the vicinity of the Savannah River and other major rivers (figs. 6, 7, and 8) indicates discharge from the aquifers to the rivers. Discharge to smaller streams has a corresponding effect on the potentiometric surface in the upper Coastal Plain, but is not explicitly shown because of the map scale and data density.

In the lower Coastal Plain, the principal discharge from the Black Creek, the Middendorf, and the Cape Fear aquifers is by diffuse upward leakage to overlying aquifers. Flow quantities from upward leakage are small because of the low

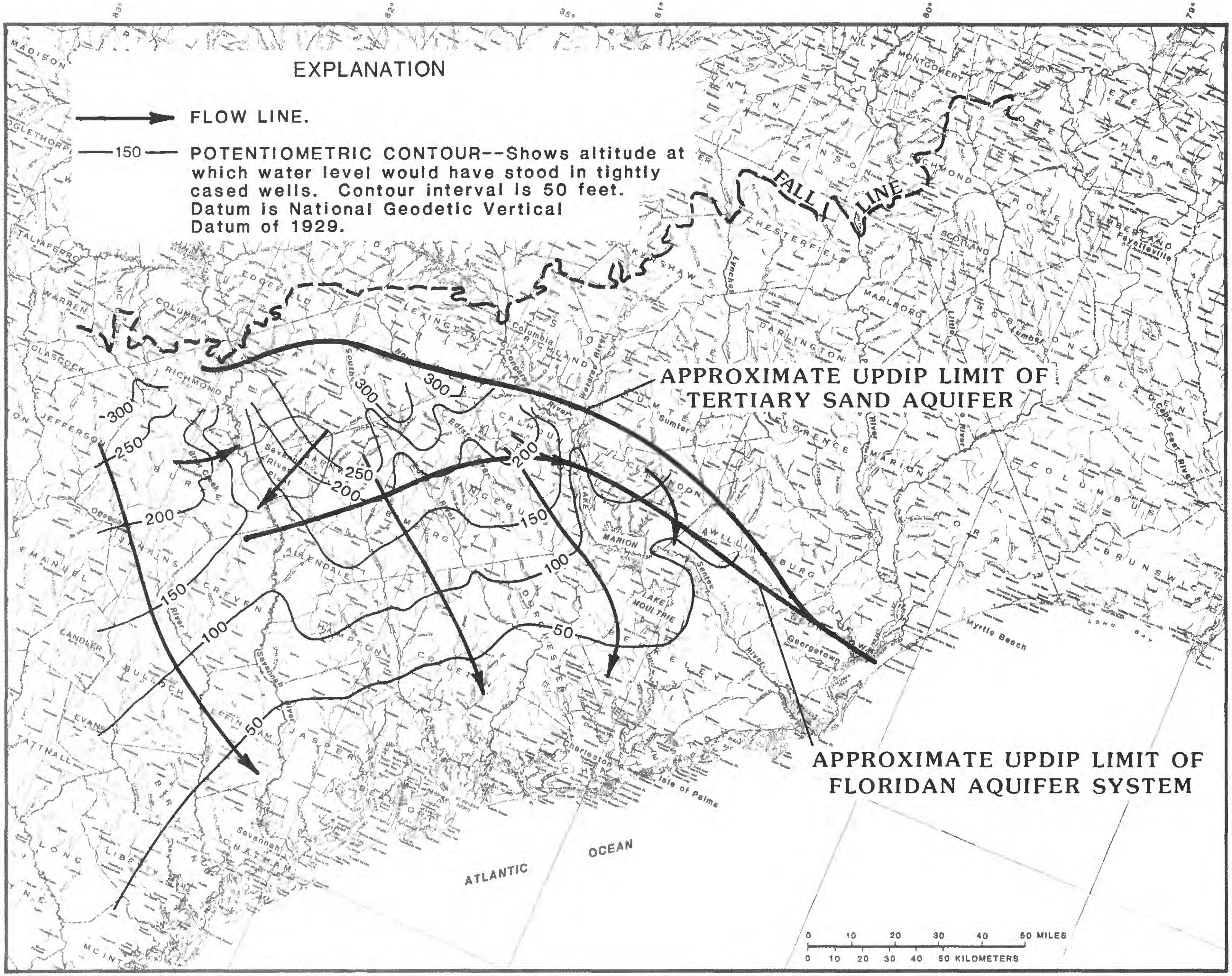


Figure 6.—The potentiometric surface of the Floridan aquifer system and the Tertiary sand aquifer prior to development. (Modified from Aucott and Speiran, 1985b.)

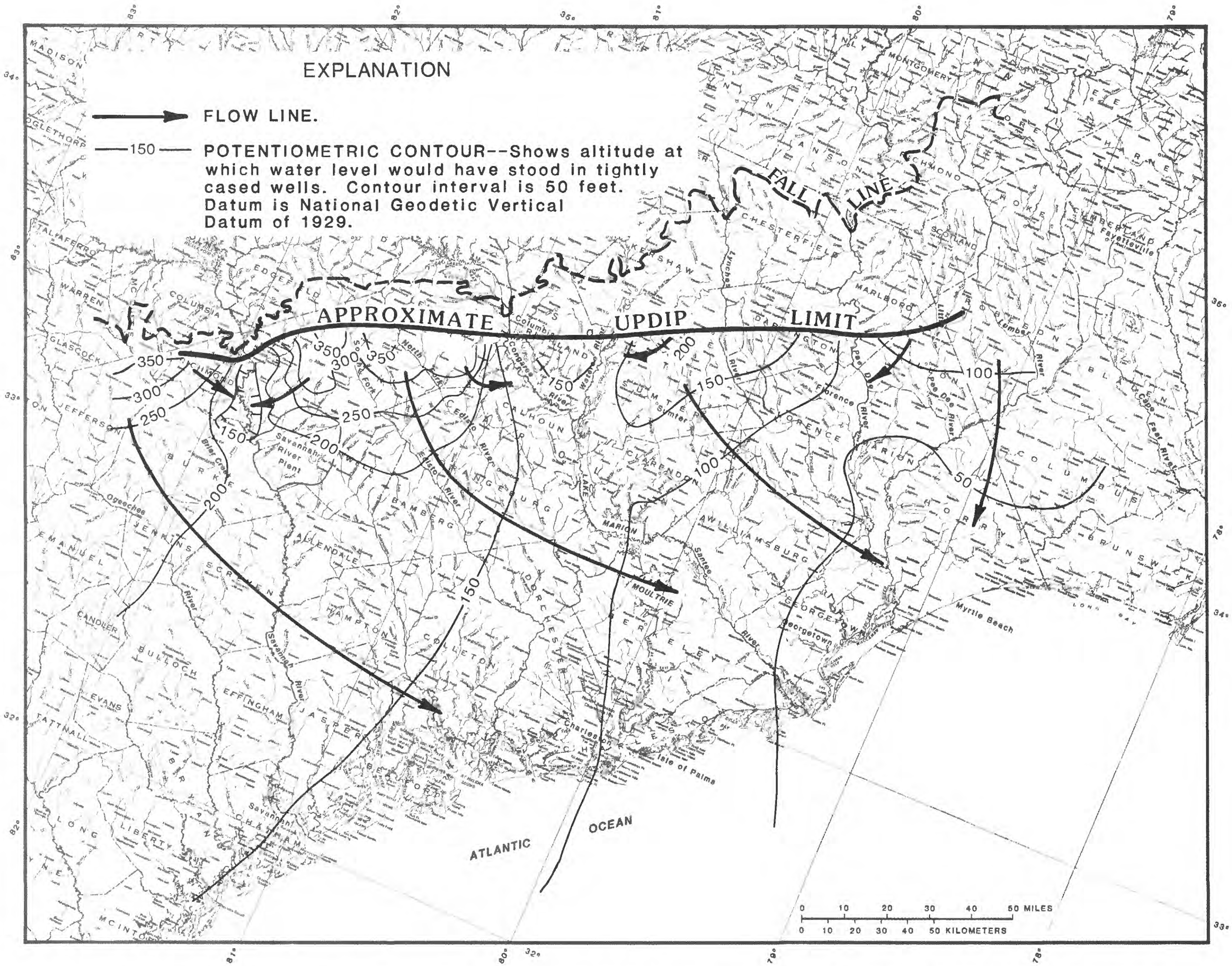


Figure 7.—The potentiometric surface of the Black Creek aquifer prior to development. (Modified from Aucott and Speiran, 1985b.)

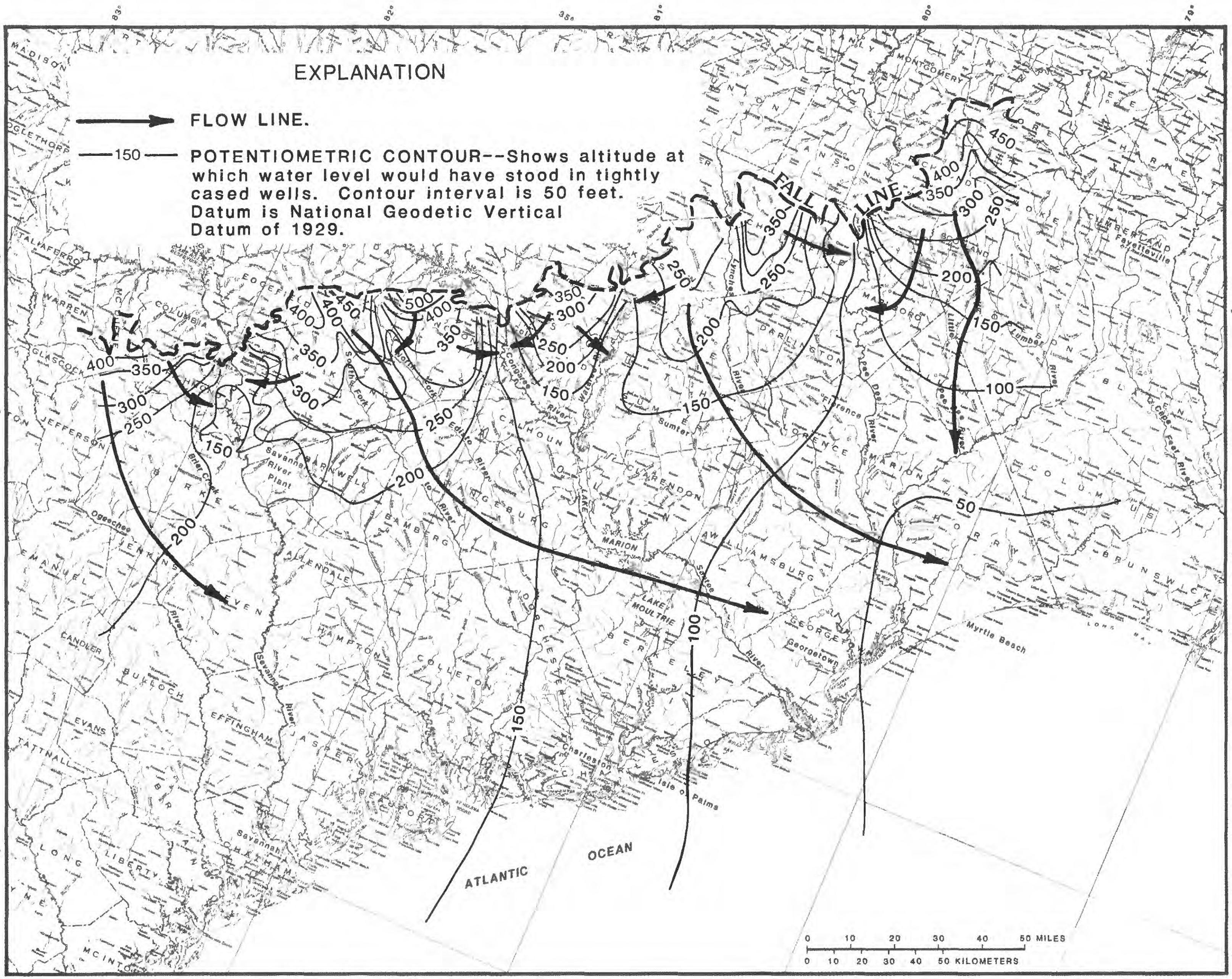


Figure 8. – The potentiometric surface of the Middendorf aquifer prior to development. (Modified from Aucott and Speiran, 1985b.)

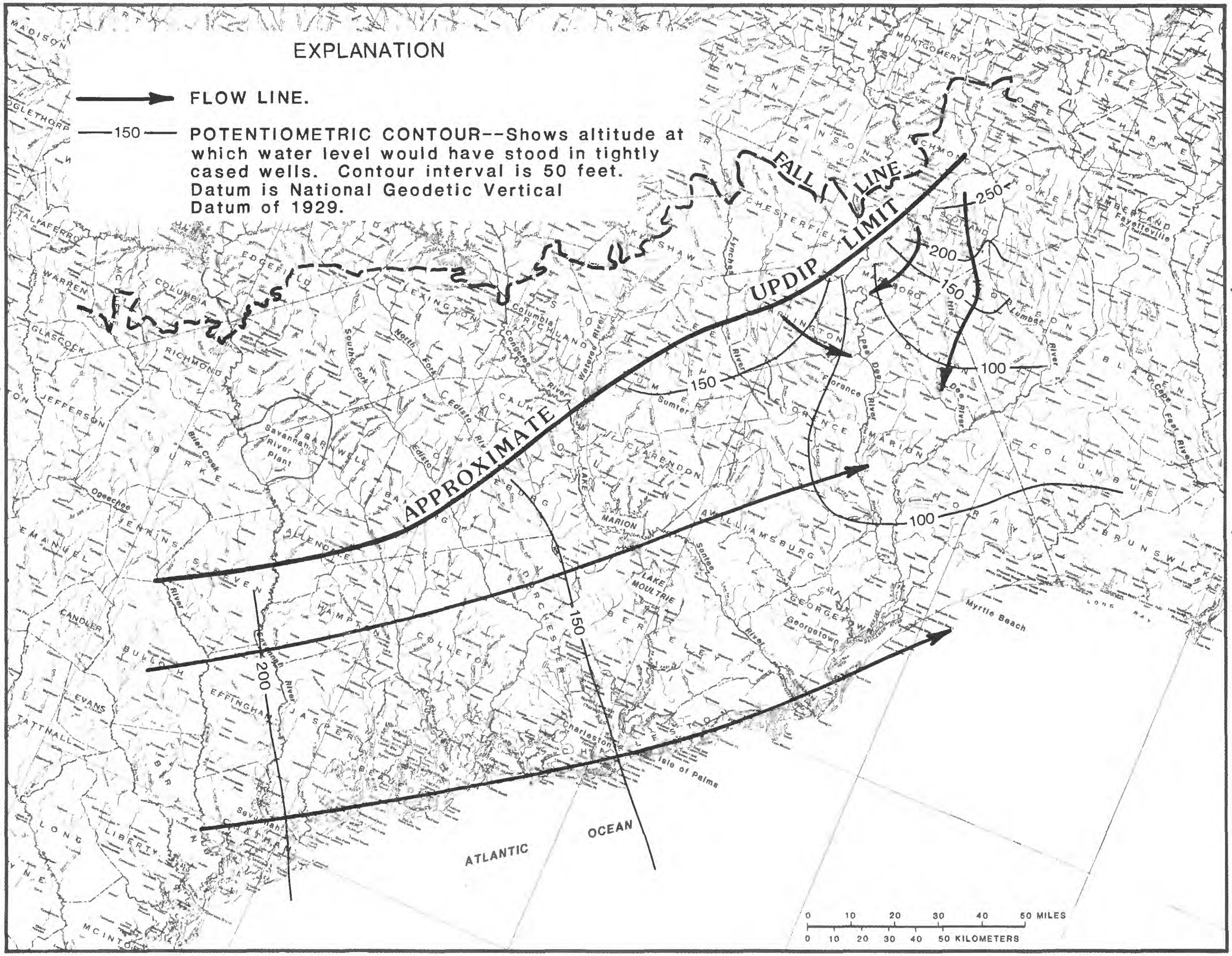


Figure 9.—The potentiometric surface of the Cape Fear aquifer prior to development. (Modified from Aucott and Speiran, 1985b.)

permeability of the confining unit overlying the Black Creek aquifer. Water that is discharged by upward leakage eventually reaches the surficial aquifer or the Atlantic Ocean. If discharged to the surficial aquifer, it eventually leaves the ground-water system by way of evapotranspiration and discharge to surface-water bodies.

The ground-water flow system in the upper Coastal Plain can be generalized in cross section as shown in figure 10. Water enters the system as recharge in topographically high areas between rivers and lakes, flows down the potentiometric gradient, and discharges to rivers, lakes, and swamps. Depending on a number of factors, such as topography, aquifer thickness, and transmissivity, a stratified flow system such as described by Toth (1963), Freeze and Witherspoon (1966), and Winter (1976), and depicted in figure 10 may occur. Such a flow system consists of a shallow flow system and a deep flow system. Water moving in the shallow flow system is quite different in flow path, velocity, and areas of discharge than water in the deep flow system.

The shallow flow system is characterized by relatively short flow paths. Typically, much of the water in a given ground-water system moves through the shallow flow system at relatively high velocities and discharges to surface-water bodies located near the recharge areas. This system is close to land surface and may be relatively thin.

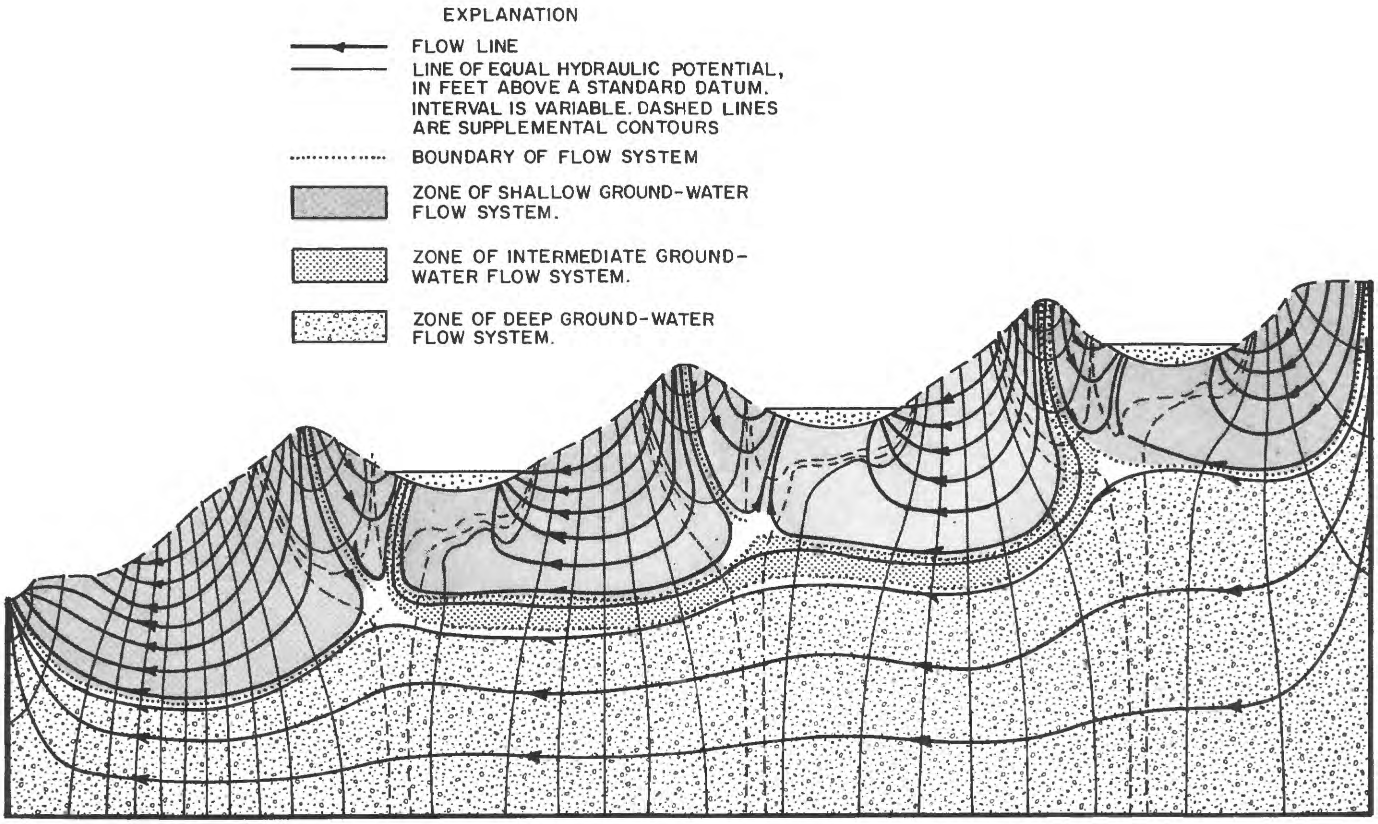


Figure 10.—Generalized depiction of a stratified ground-water flow system (Winter, 1976).

As a result, variations in recharge over time can have a considerable effect on the amount of water flowing in the shallow flow system.

In contrast to the shallow system, the deep flow system is more typically characterized by much longer flow paths and lower velocities. Because the time of travel from sources of recharge to areas of discharge is longer for the deep flow system than for the shallow flow system, the deep flow system is less affected by short-term environmental factors such as seasonal variations in recharge. The deep flow system, as referred to for the Coastal Plain aquifers, is defined as that part of the flow system that discharges to regional river drains or flows downgradient to the lower Coastal Plain.

Intermediate flow systems probably also occur in the upper Coastal Plain between the shallow and deep flow systems (fig. 10). These flow systems have characteristics intermediate between those of the shallow and deep systems, and can be considered either shallow or deep.

Many factors that affect ground-water discharge to streams also determine whether the ground-water flow system will be stratified. Some of these factors include topography; aquifer characteristics such as hydraulic conductivity, specific yield (storage coefficient), and aquifer thickness; the quantitative and temporal distribution of recharge; and conditions affecting discharge such as stream altitude and incisement and streambed hydraulic conductivity. In the upper Coastal Plain of South Carolina these factors combine to create a stratified flow system, similar to that shown in figure 10.

Ground water flows from recharge areas to discharge areas. In the upper Coastal Plain, most of the ground water in the Tertiary sand aquifer, the Black Creek aquifer, and the Middendorf aquifer flows from interstream recharge areas to rivers and small streams where the water is discharged. These flow paths are comparatively short. In the lower Coastal Plain, flow paths within each of these aquifers are much longer and the hydraulic gradients in the horizontal direction are lower than those in the upper Coastal Plain (figs. 6, 7, and 8). Ground water flows downgradient from the upper Coastal Plain to the lower Coastal Plain where it discharges as diffuse upward leakage to overlying aquifers.

In the lower Coastal Plain, the direction of flow in the Floridan aquifer system and the Tertiary sand aquifer is generally perpendicular to the coast, and, to a lesser degree, toward the major rivers (fig. 6). This differs markedly from the flow paths of the Black Creek and Middendorf aquifers. Water in these aquifers flows from the recharge areas toward the coast, then turns gradually eastward until it is moving nearly parallel to the coast (figs. 7 and 8). Because of the reduced horizontal hydraulic gradient and generally smaller transmissivities in the lower Coastal Plain, flow in the Black Creek and Middendorf aquifers is more sluggish there than in the upper Coastal Plain.

The direction of ground-water flow in the Floridan aquifer system and the Tertiary sand aquifer in the lower Coastal Plain is approximately perpendicular to flow in the Black Creek, Middendorf, and Cape Fear aquifers. This difference in flow directions in the lower Coastal Plain is probably a result of differences in the hydraulic continuity between surface discharge points and each aquifer. The Floridan aquifer system and the Tertiary sand aquifer, throughout much of their extent in the study area, have relatively good hydraulic contact with the surficial aquifer and with rivers. The Black Creek and Middendorf aquifers in the lower Coastal Plain, in contrast, are in much more effective hydraulic contact with the surficial aquifer and streams in eastern South Carolina and southeastern North Carolina than in southwestern South Carolina (Aucott and Speiran, 1985b).

The distinctive flow pattern in the Cretaceous aquifers in the lower Coastal Plain is due to three factors. First, confining units above the Black Creek aquifer are more effective in inhibiting upward leakage in southwestern South Carolina where clayey sediments of Paleocene age exist than in eastern South Carolina where the confining unit consists of more permeable sediments of the Peedee Formation. Secondly, because the dip of the Coastal Plain sediments in southeastern North Carolina is substantially less than the dip in southwestern South Carolina, the aquifers are closer to the land surface and in better hydraulic contact with the rivers farther downdip in the east than the west. Finally, the Cape Fear River and to a lesser extent the Pee Dee River are lower in altitude farther upstream than rivers to the west. These river drains of lower altitude enable a lower potentiometric surface to occur in the Cretaceous aquifers in the east. These three factors combine to provide a more effective discharge area in the eastern part of the study area than in the southwest for the Black Creek, Middendorf, and Cape Fear aquifers. This east-to-west imbalance in discharge in the lower Coastal Plain causes a major alteration in the flow direction from perpendicular to the coast to nearly parallel to the coast toward the primary discharge area in southeastern North Carolina.

Ground-water flow in the Cape Fear aquifer differs from that in other aquifers. In western Georgia, the Cape Fear aquifer is recharged from precipitation in its outcrop and by downward leakage downdip of its outcrop. In the eastern upper Coastal Plain part of the study area, it is recharged by downward leakage from the Middendorf aquifer. As mentioned previously, discharge from the Cape Fear aquifer occurs as diffuse upward leakage to the Middendorf aquifer in the lower Coastal Plain, especially in southeastern North Carolina. This results in a very long flow path from western Georgia along the South Carolina coast into southeastern North Carolina. Because of the relatively low transmissivity of the Cape Fear aquifer in much of South Carolina, flow within it is rather sluggish.

## MODELING THE GROUND-WATER FLOW SYSTEM

### Model Introduction

A ground-water flow model was constructed to describe the areal distribution of aquifer parameters and ground-water flow in the Coastal Plain aquifer system. The U.S. Geological Survey's three-dimensional finite-difference modular flow model (McDonald and Harbaugh, 1984) was used for the simulation. The strongly implicit numerical procedure was used to solve the ground-water flow equations. Detailed information on the model and solution technique can be found in the report cited above and in Trescott (1975) and Trescott and others (1976).

The predevelopment flow system of the Coastal Plain aquifers of South Carolina was simulated with a steady-state, quasi-three-dimensional approach. The Coastal Plain of South Carolina and adjacent areas of Georgia and North Carolina were divided into grid blocks 4 miles on a side as shown in figure 11. This discretization yielded a three-dimensional network of 5 layers, 48 rows, and 63 columns. It was assumed in the model design that flow within aquifers was predominantly horizontal. Because horizontal flow in confining layers is negligible compared to that in aquifers, flow within confining units was assumed to be principally vertical. It was also assumed that the aquifers and confining units are heterogeneous and isotropic. Each of the 5 aquifers previously defined (fig. 12) was assigned a model layer, numbered 1 to 5 from land surface down (fig. 13). Flow across confining units was represented in the model as flow between model layers.

The model of the Coastal Plain aquifer system only simulates the deep flow system as previously described. This includes all of the lower Coastal Plain and the part of the flow system in the upper Coastal Plain that directly flows to the large rivers (Savannah, North and South Forks of the Edisto, Congaree, Wateree, Lynches, Pee Dee and Lumber Rivers) of the upper Coastal Plain or that flows downgradient to the lower Coastal Plain. It was not possible to simulate the shallow flow system in the upper Coastal Plain because of its complexities and the large grid size necessitated by the regional scale of this investigation.

### Model Boundaries

Selection of model boundaries, using hydrologic boundaries wherever possible, is an important element in an accurate simulation of a ground-water flow system. The model boundaries used for this simulation are depicted in figures 11 and 13. Two types of boundaries are used in this model. The first type, a no-flow boundary, prevents the movement of water across it in the simulation but allows the hydraulic head, or aquifer water level, to vary. The second type, a constant-head boundary, fixes the hydraulic head or water level at a specified value but allows the movement of water across the boundary. In



page 19 follows

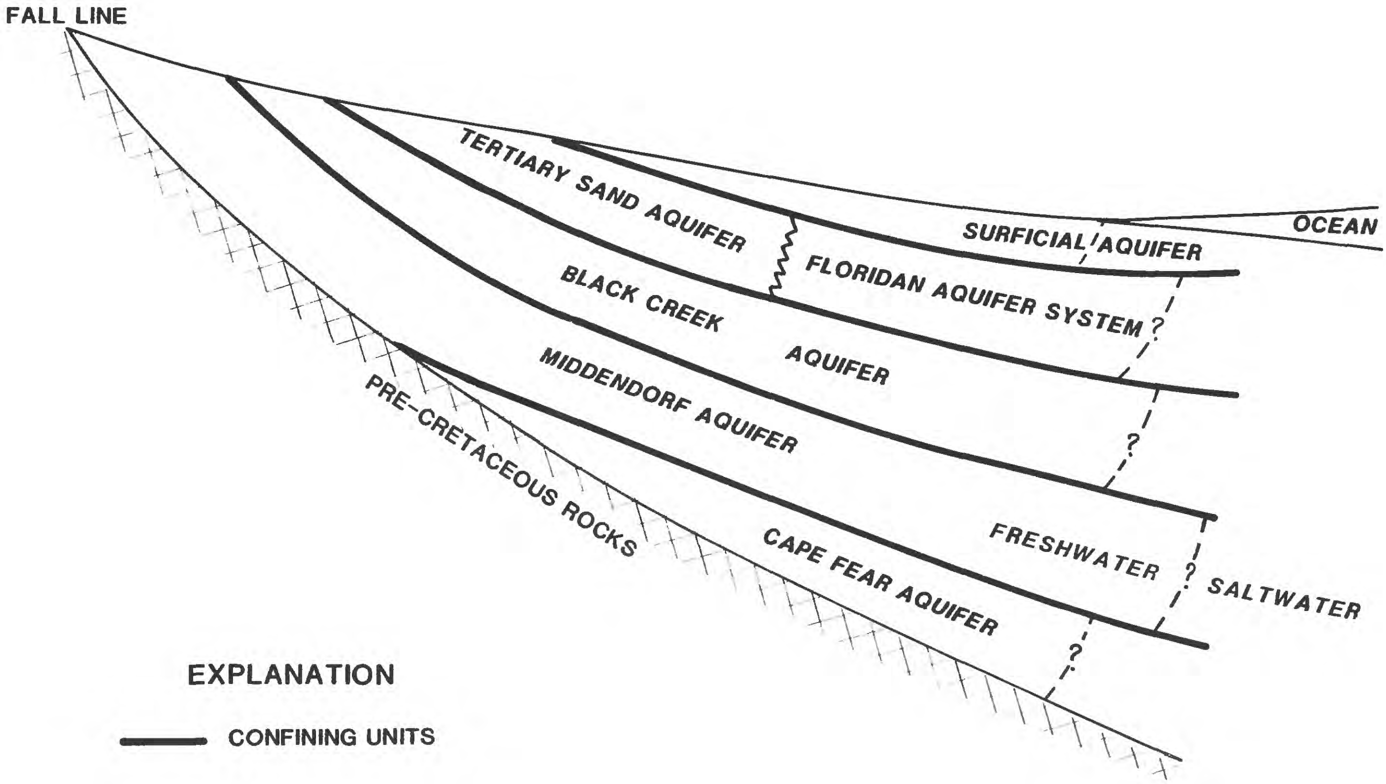


Figure 12.—Schematic geohydrologic section.

general, a no-flow boundary is useful where little or no water flows across a particular boundary, and a constant-head boundary is useful where water flows across the boundary but the head is known and remains constant through the interval of time of the simulation. All areas within the model boundaries are being actively simulated. Water levels and flows are computed in active areas of the simulation, while either water levels or flows or both are not being computed in inactive areas, but are controlled, in constant head or in no flow cells.

Overlying the clastic aquifers are the surficial aquifer in the east and the part of the Floridan aquifer system previously modeled by Bush (1982), Krause (1982) and Randolph and Krause (1984) in the southwest. The surficial aquifer is simulated with constant heads and acts as a source or sink layer. A source-sink enables the simulation of water movement into (source) or out of (sink) the actively simulated area. The part of the Floridan aquifer system in South Carolina modeled by Bush (1982) and Krause (1982) is simulated with constant heads while the remainder of the Floridan aquifer system and its updip equivalent, the Tertiary sand aquifer, are simulated actively (figs. 11 and 13).

Underlying the Coastal Plain aquifer system are very low permeability pre-Cretaceous rocks. As a result the flow of water within these rocks and between these rocks and the Coastal Plain aquifers is considerably less than flow within the Coastal Plain aquifers. The interface between Coastal Plain sediments and the underlying pre-Cretaceous rocks can thus be simulated as a no-flow boundary.

Except for the surficial aquifer, the updip limit of all aquifers is simulated with a no-flow boundary. The updip limit of each unit represents the apex of the wedge of sediments of a particular aquifer. Updip of this limit, the aquifer does not

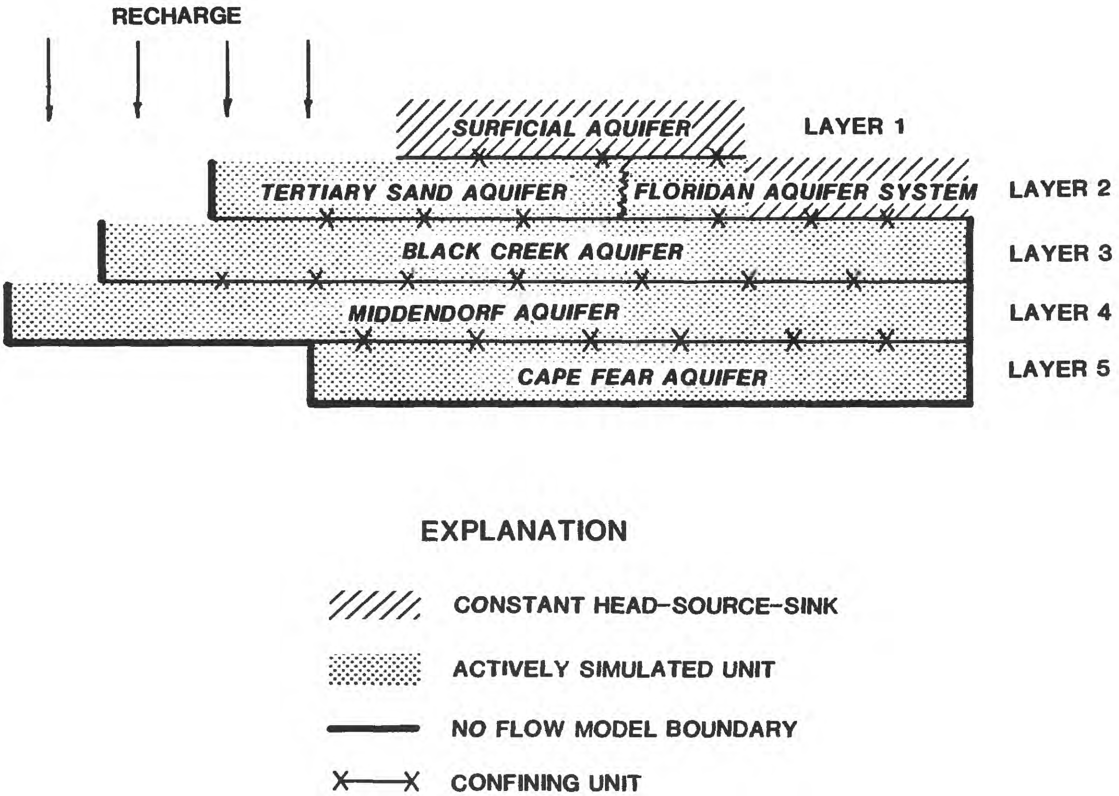


Figure 13.—Conceptualization of model layers.

exist, consequently no flow occurs in a layer across this boundary (figs. 11 and 12).

The downdip limit of all units is the saltwater-freshwater interface. Only the freshwater flow system is simulated by this model. Although the saltwater-freshwater interface in nature is gradational, it is considered in this report to represent a sharp boundary as defined by a line representing a concentration of 10,000 mg/L dissolved solids as sodium chloride in the water within a given aquifer. As a result of steady-state conditions being present in the flow system prior to development, the saltwater-freshwater interface is considered to be stationary. It can be assumed that there is no flow across the interface, although some circulation does exist between the freshwater and saltwater flow systems (Glover, 1964; G Bennett, U.S. Geological Survey, written commun., 1979). This interface is thus simulated as a no-flow boundary except for the surficial aquifer and the parts of the Floridan aquifer system that are simulated with constant heads. The position of this interface, which exists nearly everywhere under the Atlantic Ocean, cannot be accurately determined using existing data. Its approximate position was estimated using data from Georgia and North Carolina and projecting a line between these data (Lee and others, 1985).

Unlike the boundaries previously discussed, the lateral boundaries to the northeast and the southwest do not necessarily represent definite hydrologic boundaries. Although potentiometric divides and rivers may provide reasonable boundary conditions for one or two aquifers in limited areas, good lateral hydrologic boundaries of regional extent are generally not available within the Coastal Plain. The northeast and southwest boundaries are set outside of South Carolina to coincide with a potentiometric divide in North Carolina and at a position far enough to the southwest in Georgia so that boundary effects did not significantly affect the simulation within the Coastal Plain of South Carolina. The lateral boundaries are simulated as constant heads except for the upper Coastal Plain part of the boundary in North Carolina which is simulated as no-flow because of the potentiometric divide that exists there.

### Model Inputs

Prior to model calibration, initial estimates of aquifer and confining unit characteristics were estimated and translated into model input. Calibration parameters were determined and put into a format so that they could be compared to the model output. The calibration parameters were not adjusted during the simulation but were used for comparison between simulated values and field-measured values to determine the adequacy of the model calibration. The calibration parameters for this simulation are the predevelopment water levels for the aquifers in all actively simulated areas and base flows for selected upper Coastal Plain river reaches. Predevelopment water levels were determined by using potentiometric maps by Aucott and Speiran (1985b) (figs. 6, 7, 8, and 9). Base flows used were determined from low-flow calculations by Aucott, Meadows, and Patterson (1986).

Input data included defining boundary conditions, transmissivities for each aquifer, leakage coefficients (vertical hydraulic conductivity divided by thickness) of each confining unit, recharge, river stage altitudes, and streambed conductances. All parameters were fully described throughout the areal extent of each hydrogeologic unit.

Heads for areas simulated as constant head were estimated differently from those areas that were simulated actively. Surficial aquifer heads were estimated using topographic maps and general estimates of depth to water using the few measurements available. Heads for the part of the Floridan aquifer system modeled by Bush (1982) and Krause (1982) were derived by using simulated heads from Krause (1982) as modified by more recent field data in Georgia by R. Faye (written commun., 1984)

Transmissivities were estimated for each aquifer using data from aquifer tests and specific capacity tests (Aucott and Newcome, 1986). River stage altitudes were obtained from U.S. Geological Survey gaging stations where they exist and otherwise estimated from topographic quadrangles with 5- to 20-foot contour intervals. Vertical hydraulic conductivities were initially estimated as being uniform for each confining unit throughout its area of active simulation from published estimates, preliminary modeling work and laboratory permeabilities on samples of two test wells drilled during this study (J. Cahill, written commun., 1985). The recharge distribution was initially estimated by projecting base flow rates throughout the outcrop areas using base flow work by Stricker (1983). Streambed conductances were initially selected from the results of preliminary modeling.

### Model Calibration

The objective of calibrating this ground-water flow model was to obtain a simulation tool that accurately describes the ground-water flow system under steady-state conditions prior to development. The result of this calibration is a model that improves the conceptualization of the flow system as well as an improved knowledge of the distribution of the input parameters.

Model calibration was accomplished by the trial and error adjustment of model inputs. After each adjustment, the simulated water levels and river flows were compared to those derived from field measurements to evaluate the progress attained in that adjustment. After many trial and error adjustments, the criteria for calibration were achieved.

Calibration of the South Carolina Coastal Plain ground-water flow model was achieved by the adjustment of all input parameters except river altitudes. Those that were adjusted the most were recharge, leakage coefficients of confining units, aquifer transmissivity (mostly near the updip limit of each aquifer) and streambed conductance. These parameters were adjusted the most because the initial estimates of them were relatively poor and because the simulation was sensitive to adjustments of these parameters. Initial estimates of other parameters such as aquifer transmissivity (away from the updip limit) were better, resulting in a much smaller range of reasonable adjustments.

The accuracy of the calibration achieved can best be measured by a comparison of simulated calibration parameters to their field-measured counterparts. Water levels were the major calibration parameter. A statistical comparison of the mean absolute differences between simulated water levels and contoured data from field measurements is shown in table 2. An areal comparison of the manually contoured and simulated potentiometric surfaces is shown for each active model layer in figures 14 through 17.

Although the manually drawn contours were not duplicated exactly by the model, the general shape of the potentiometric surfaces and the indicated directions of flow match very well. Considering that (1) a 25-foot contour interval was used for contoured field data and (2) most of the water-level surfaces in the upper Coastal Plain are extremely variable because of the effects of rivers, the total mean differential of 8.9 feet and standard deviation of 9.1 feet suggest that calibration is sufficient to satisfy the intended uses of the model. The maximum difference is 86 feet. The maximum differences occurred in the Middendorf aquifer near the Fall Line where the areal changes in water levels were greatest. Calibration was also compared at cells where actual water levels were made. This was considered to be a less reliable method of measuring the accuracy of calibration because of the uneven distribution of data points and the extreme variability of the potentiometric surface, with respect to grid size, in the upper Coastal Plain where most of the field measurements were located.

One simulation problem has remained unresolved through calibration. Although the difference between field contoured and simulated water levels was not great, the proper flow direction in the downdip eastern part of the Cape Fear aquifer could not be simulated using reasonable values for the transmissivity of the Cape Fear and leakage coefficient of the overlying confining unit (fig. 17). The reason for this is not known, but the scarcity of existing data for these units may be a contributing factor. In any case, problems with this unit appear to have little effect on the other aquifers of the Coastal Plain aquifer system. This is because of the relatively minor role and relative isolation of the Cape Fear aquifer from the total flow system.

A second, but less accurate, check on calibration was a comparison of simulated river base flow with base flow calculated from field measurements. Table 3 lists river flows determined from base-flow calculations and from simulated rates of aquifer to river discharge for those river reaches

Table 2.—*Mean absolute differences and standard deviations calculated from simulated water levels and water levels contoured from field measurements*

| Layer | Mean absolute error (feet) | Standard deviation (feet) | Maximum error (feet) |
|---|---|---|---|
| 2 | 9.4 | 9.6 | 76 |
| 3 | 7.0 | 7.0 | 56 |
| 4 | 8.8 | 9.8 | 86 |
| 5 | 11.2 | 9.5 | 66 |
| Total | 8.9 | 9.1 | 86 |



page 31 follows

Table 3.—*Comparison of observed and simulated flows from aquifers to rivers*
[$ft^3/s$, cubic feet per second]

| River reach and abbreviated station number | Observed: from base flow analysis[1] ($ft^3/s$) | Model simulated ($ft^3/s$) |
|---|---|---|
| Savannah River between Augusta (1970) and Millhaven (1975). | 154 | 134 |
| South Fork Edisto River between Montmorenci (1725) and Denmark (1730). | 45 | 42 |
| North Fork Edisto River between Steedman and Orangeburg (1735). | 100 | 79 |
| Lynches River between Jefferson (1313) and Bishopville (1315). | [2]110 | 41 |
| Pee Dee River between Rockingham (1290) and Peedee (1310). | 101 | 70 |
| Lumber River between Derby (1329.43) and Pembroke (1336.4). | 29 | 28 |

1From Aucott, Meadows, and Patterson (1986).

[2]Poor estimate that resulted primarily from insufficient tributary data.

for which base-flow calculations were possible. As described in Aucott, Meadows, and Patterson (1986), the probable error in the base-flow calculations is relatively high and the flows calculated from low-flow data are probably greater than should be expected. The base-flow calculations may include some effects of the shallow flow system, close to the large rivers, in addition to the deep flow system, although the model simulates discharge from only the deep flow system to the large rivers. Given the uncertainty of these values, the similarity between the calculated and simulated aquifer-to-river discharge can only be judged in a general sense. With the exception of the Lynches River, all of the simulated aquifer-to-river discharges are reasonable approximations of calculated base-flows but are less, as expected. The divergence of simulated and calculated base flows of the Lynches River is due to a relatively poor computation from low flow data resulting from insufficient flow data on tributaries.

### Sensitivity Testing

Testing was performed to determine the sensitivity of the model to changes in various input parameters. Given the complexity of the model and the large number of parameters, layers, and nodes, complete testing of all the possible combinations was not practical. The final procedure used to test sensitivity was to vary one parameter by a given multiple for every node in every layer of the model and observe the effects on simulated heads and the water budget. This was done repeatedly until enough data were assembled to construct figures 18 through 23, which show the sensitivity of the model solution to variations in each input parameter. Head residuals are the differences between simulated water levels and water levels contoured from field data. It was found from preliminary sensitivity testing that variations in a parameter in one layer resulted in smaller changes in the difference between observed and simulated water levels than

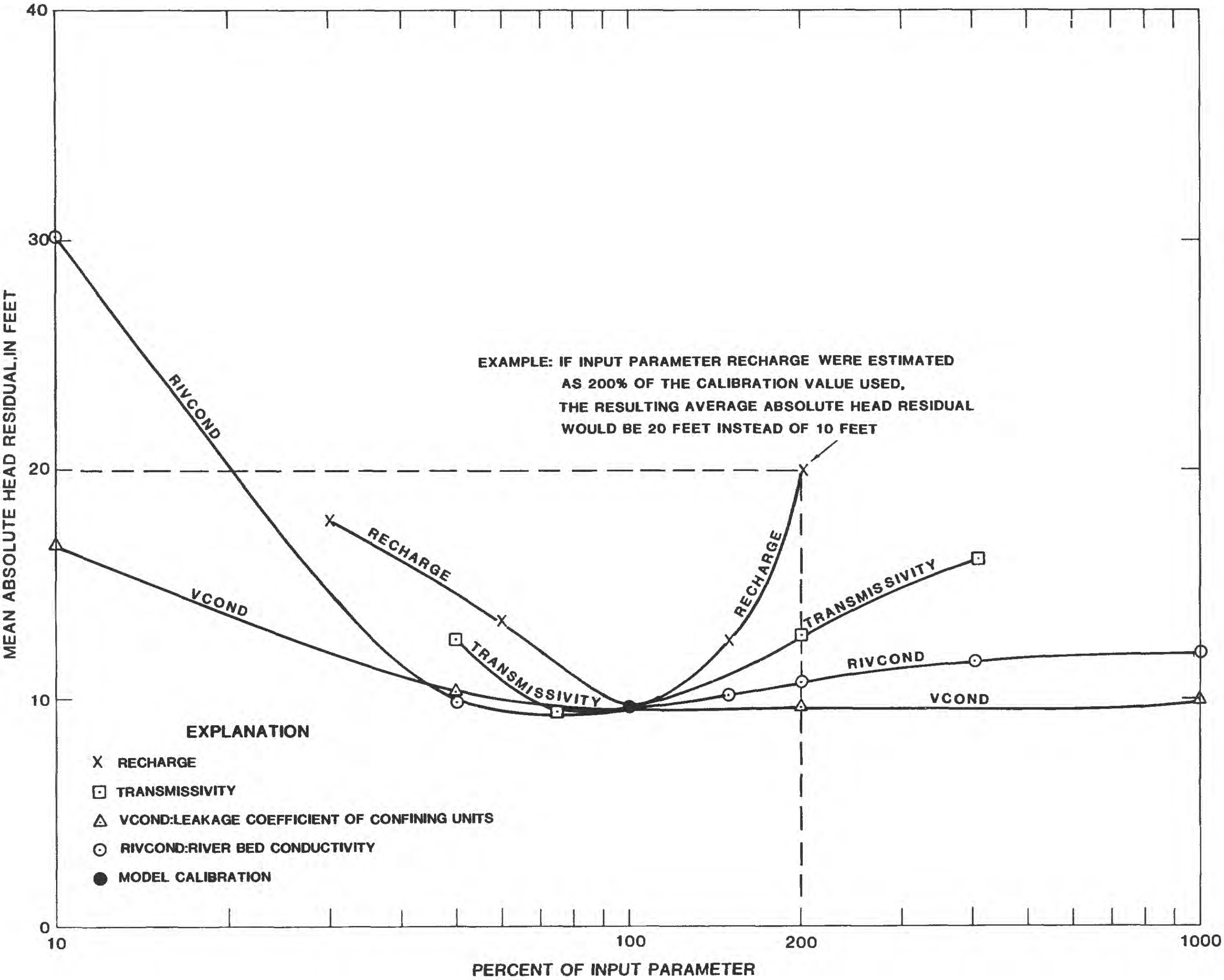


Figure 18.—Relation between changes in magnitude of input parameters and average absolute head residual per grid block for the Floridan aquifer system and the Tertiary sand aquifer.

occurred if the variation was applied throughout the model. This should also apply generally to variations made over only part of a layer, versus changes made over the entire layer.

Because the sensitivity testing was of a general nature, the conclusions resulting from this testing must be general as well. The calibration in most layers appears to be most sensitive to increases in recharge. Changes in transmissivity of the aquifers, leakage coefficeint of confining units, and decreases in riverbed conductance also affected the model solution significantly. Most layers were less sensitive to increases in riverbed conductance and decreases in recharge. Changes in the position of the no-flow boundary used to

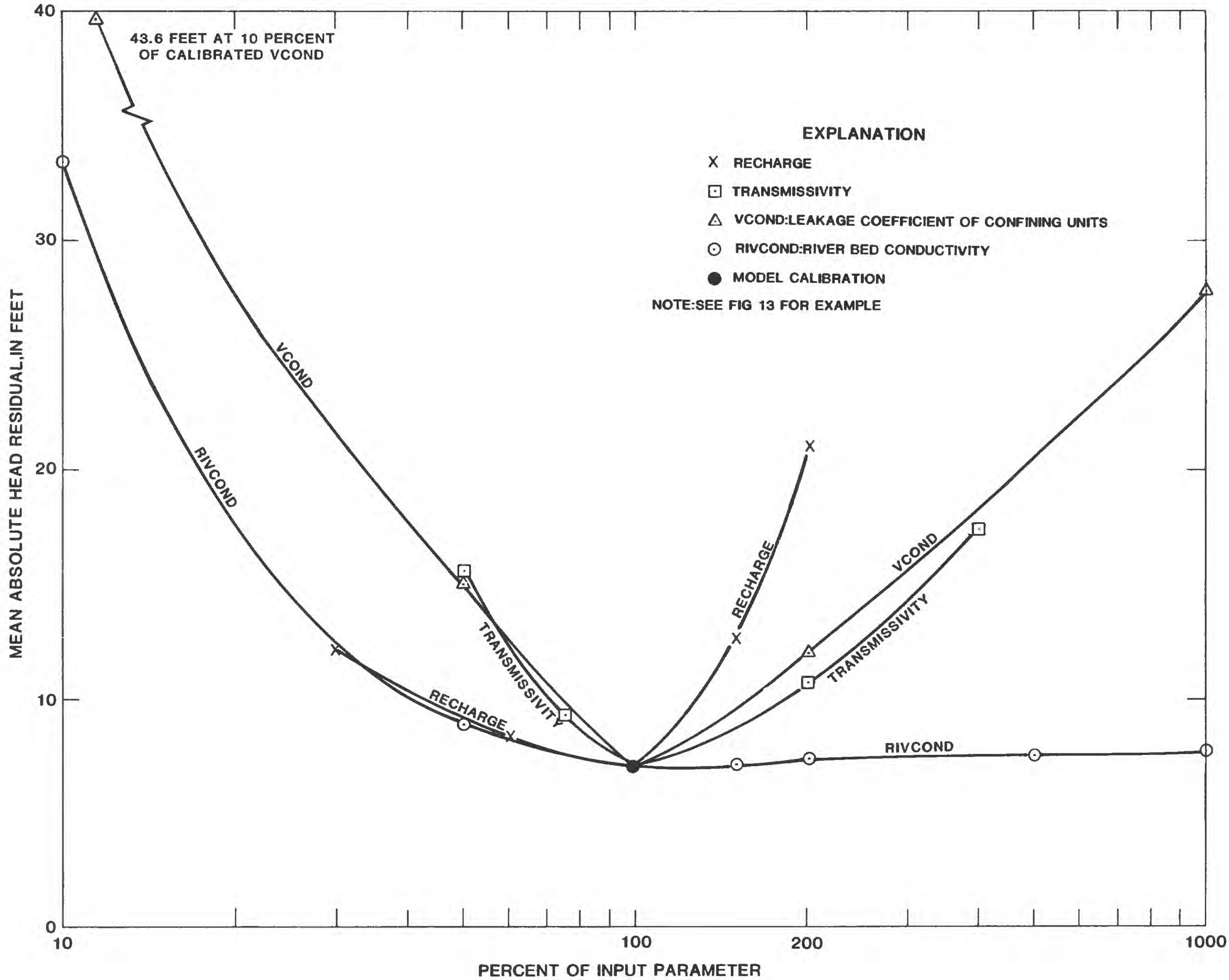


Figure 19.—Relation between changes in magnitude of input parameters and average absolute head residual per grid block for the Black Creek aquifer.

simulate the saltwater-freshwater interface indicate that the model solution is relatively insensitive to the location of this boundary.

The calibration of the Middendorf aquifer was quite sensitive to decreases of transmissivity (fig 20). The Middendorf aquifer has the largest area of outcrop where direct recharge is applied. The close relation between aquifer transmissivity and recharge in the extreme updip part of the flow system results in the Middendorf aquifer being very sensitive to decreases in transmissivity.

The sensitivity analysis demonstrated that an increase in Cape Fear aquifer transmissivity would result in an improvement in the model

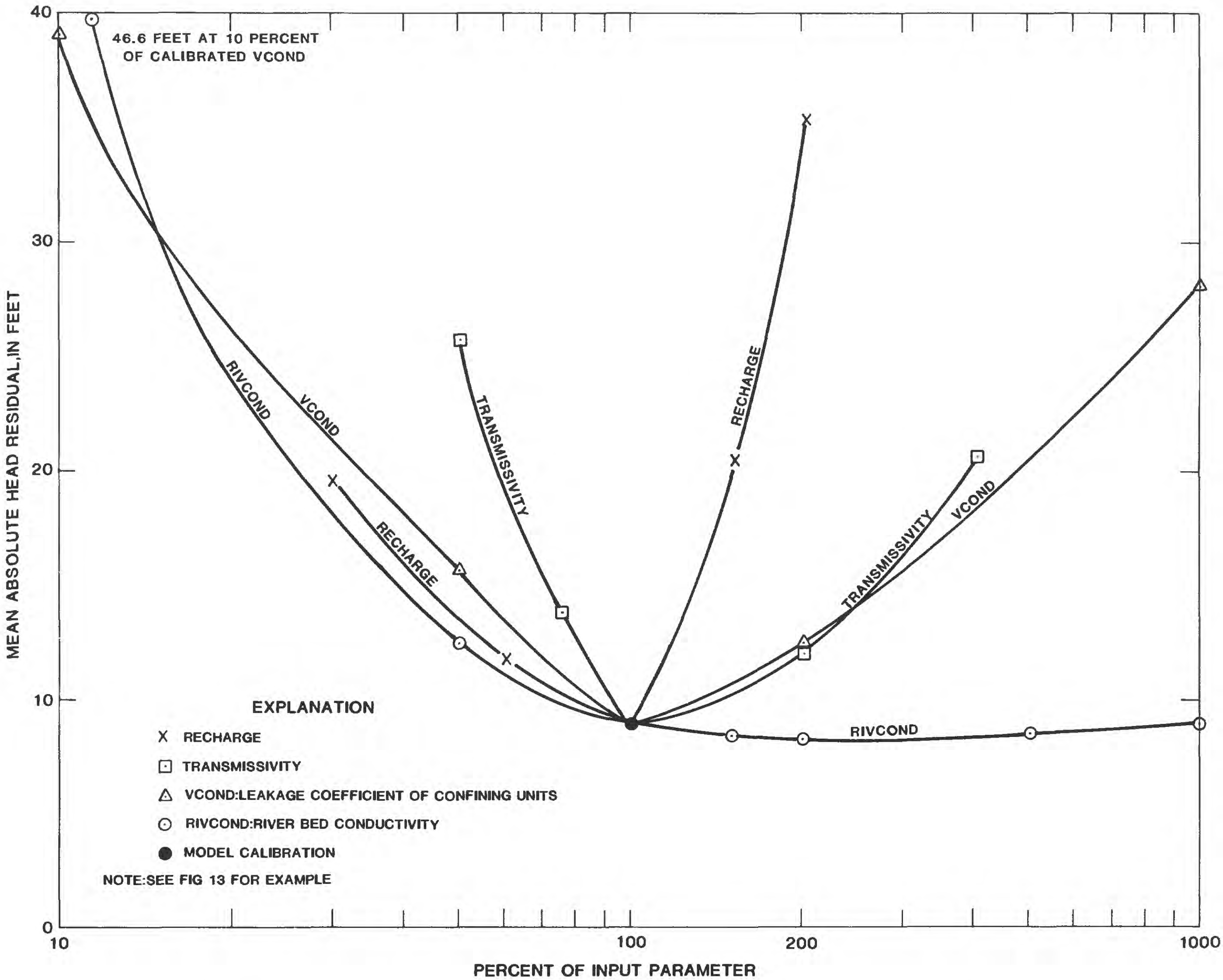


Figure 20.—Relation between changes in magnitude of input parameters and average absolute head residual per grid block for the Middendorf aquifer.

calibration as measured by the difference between field contoured and simulated potentiometric surfaces. The existing simulated transmissivity of the Cape Fear aquifer is already considered to be a realistic maximum value in most areas considering available hydraulic and lithologic information. This is particularly the case in the downdip eastern part of the Cape Fear aquifer which is the area presenting the greatest problem in model calibration. More investigation is needed to determine if a greater transmissivity than that being used in the simulation is actually justified by the field data. Given the data currently available, the calibration presented is the most reasonable.

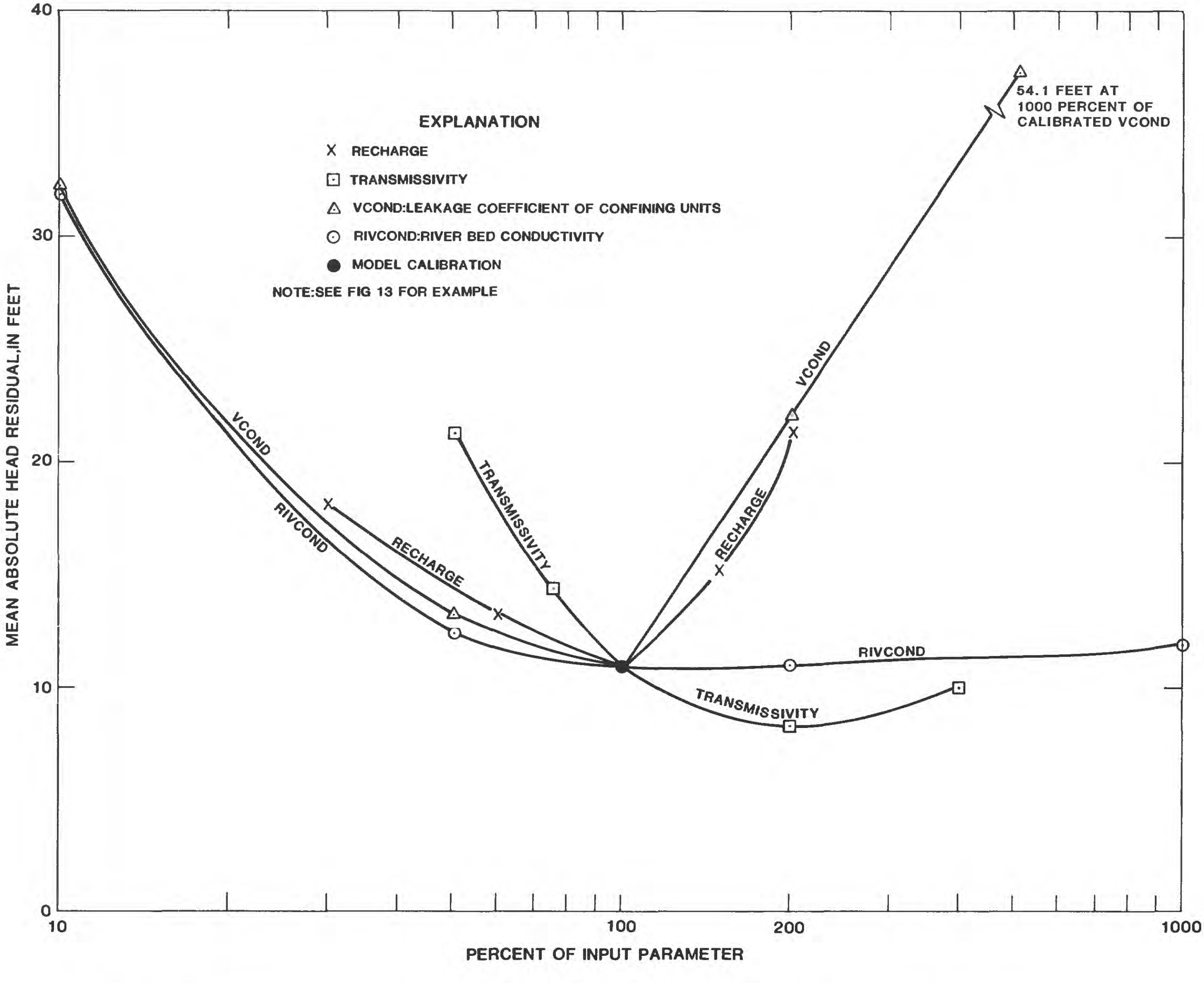


Figure 21.—Relation between changes in magnitude of input parameters and average absolute head residual per grid block for the Cape Fear aquifer.

Because the model calibration is very sensitive to two input parameters (leakage coefficient and recharge) that are largely derived by model calibration, the confidence in the calibration is greater than had the calibration been relatively insensitive to these parameters. The results of the model do not represent a unique solution in the description of the flow system. The results of sensitivity testing, however, give some indication of the limits within which the parameters may vary and still maintain a reasonable calibration.

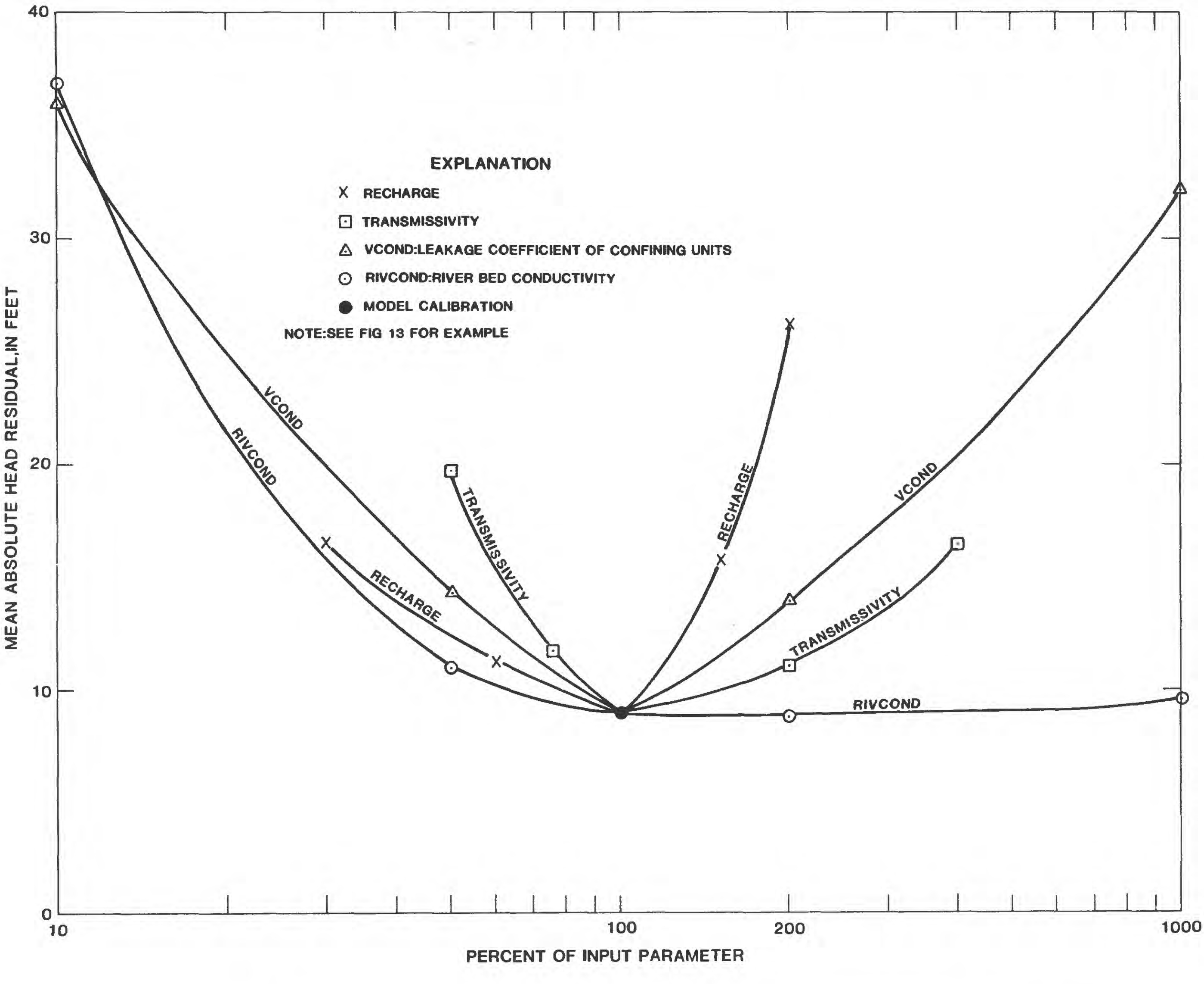


Figure 22.—Relation between changes in magnitude of input parameters and average absolute head residual per grid block for all aquifers.

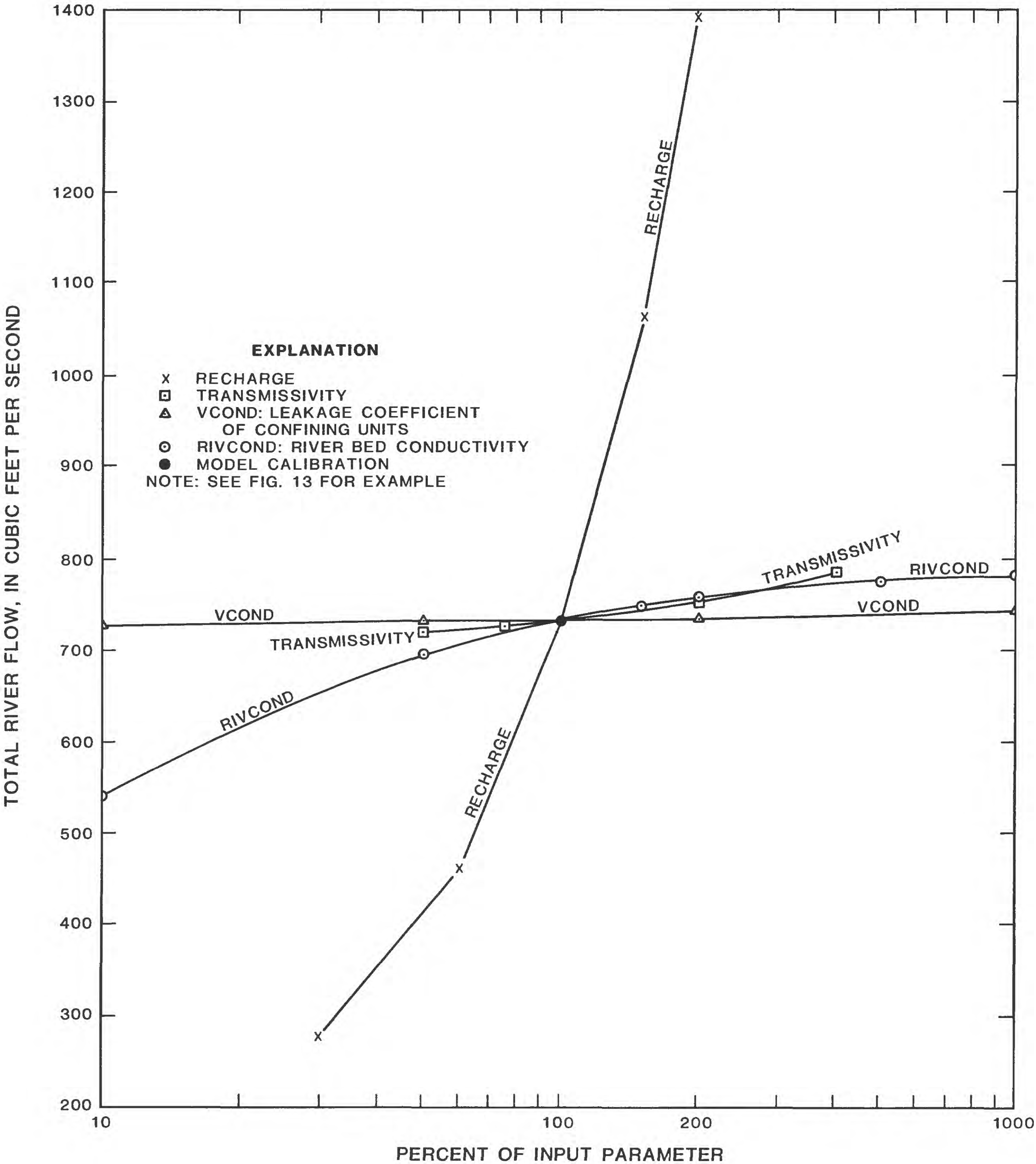


Figure 23.—Relation between changes in magnitude of input parameters and total river flow.

## MODEL-DERIVED HYDROLOGIC CHARACTERISTICS

The model results are presented as maps of the calibrated parameters: transmissivity (figs. 24 through 27) and leakage coefficient (figs. 28 through 31). Recharge to or discharge from the deep ground-water flow system is depicted in figure 32. A simulated water budget is shown in figure 33. Future investigations involving transient simulations may require adjustment of these parameters to satisfy calibration criteria specific to transient simulation.

The transmissivity distributions from the calibrated model are little changed from the initial estimates of transmissivity. The principal change is higher transmissivity values near the updip limit of the Tertiary sand aquifer, the Black Creek aquifer, and especially the Middendorf aquifer. The reliability of initial transmissivity estimates in the extreme updip parts of these units was low because few reliable aquifer tests were available there (Aucott and Newcome, 1986). The changes made resulted in transmissivities that seem reasonable in comparison to transmissivities immediately downdip where more reliable information exists.

Calibrated transmissivities range from about 700 to 11,000 $ft^2/d$ for the Tertiary sand aquifer and the actively simulated part of the Floridan aquifer system; from about 400 to 11,000 $ft^2/d$ for the Black Creek aquifer; from about 300 to 30,000 $ft^2/d$ for the Middendorf aquifer; and from about 1,100 to 3,600 $ft^2/d$ for the Cape Fear aquifer. Transmissivities are lowest near the updip limit of each unit because of the greatly reduced aquifer thickness there. Transmissivities are not presented for the surficial aquifer and for the part of the Floridan aquifer system simulated with constant heads because transmissivities in these inactive parts of the model are not part of the calibration process.

The transmissivity of the Tertiary sand aquifer and Floridan aquifer system is lowest near their updip limits (fig. 24). The transmissivity increases rapidly from northwest to southeast along the Savannah River through the Tertiary sand aquifer and across the limestone facies change into the Floridan aquifer system as described by Krause (1982) (fig. 24). This occurs because of the coarse-grained material present in the Tertiary sand aquifer in this area. Increases in transmissivity from northeast to southwest, for example from Berkeley County to Beaufort County, are much less rapid. This is because the Floridan aquifer system is composed of the moderately permeable Santee Limestone in Berkeley, Charleston, and Dorchester Counties (Park, 1985; Aucott and Newcome, 1986). The transmissivity greatly increases in Beaufort and Jasper Counties where the much more permeable Ocala Limestone comprises the upper part of the Floridan aquifer system there (Hayes, 1979; Krause, 1982). The transmissivity of these units was not substantially altered during calibration with the exception of increases near the updip limits as previously discussed and increases in the Barnwell County, S.C.,-Burke County, Ga., area that reflected data acquired during model calibration.

Black Creek aquifer transmissivity varies generally as a band parallel to its updip limit (fig. 25). Near the updip limit, the transmissivity is less than 2,000 $ft^2$/day. Moving toward the coast, it increases because of increased aquifer thickness. The coarser sediments found in this aquifer in the western part of the study area result in greater transmissivities there than to the east. These coarser materials become much finer toward the coast. This results in a significant decrease in transmissivity to the west near the coast. Throughout most of the eastern part of the study area the sediments are fine sand laminated with clay. Because the aquifer thickness in the east remains relatively constant toward the coast, the transmissivity is consistently between 2,000 and 5,000 $ft^2$/day in much of the eastern part of the study area. The major changes during calibration were an increase of transmissivity within 20 miles of the updip limit and increases in most of the eastern part of the study area. The revised transmissivities in the east correlate well with the greater transmissivity estimates for the Black Creek aquifer derived from aquifer test and specific capacity data.

The pattern of transmissivities in the Middendorf aquifer is somewhat similar to the banded pattern found in the Black Creek aquifer (fig. 26). In the Middendorf, transmissivities increase from



page 47 follows

a minimum at the Fall Line, which is the updip limit for most of the Middendorf aquifer, and reach a maximum about one-third of the way toward the coast. As in the Black Creek, the transmissivity in the Middendorf aquifer is greater in the west than in the east primarily as a result of coarser sediments in the west. Transmissivities decrease toward the coast but are generally greater than or equal to that found in the Black Creek aquifer. Data near the coast and near the updip limit are much sparser than elsewhere. Adjustments in Middendorf aquifer transmissivity during model calibration were confined to areas near the Fall Line, where they were increased over initial estimates except in the east (Richmond and Scotland Counties, N.C., and northern Marlboro County, S.C.) where aquifer transmissivities were reduced. Calibrated transmissivities in this area may vary from actual values because of boundary and data density problems.

Transmissivities in the Cape Fear aquifer are less than 2,000 $ft^2$/day nearly everywhere in the study area (fig. 27). The sparse hydraulic and lithologic data available suggest that they are even poorer than that derived from the model calibration. Transmissivities in the Cape Fear aquifer were increased during calibration to what was considered a reasonable upper limit given the available data. As mentioned earlier, sensitivity testing indicates that the calibration could be further improved with greater increases. Because this is not supported by the existing data, the transmissivities presented are considered as best estimates until new data become available.

Leakage coefficients were adjusted extensively during calibration because they were one of the parameters initially estimated with the least confidence, and because the system was sensitive to its changes. Calibrated leakage coefficient values range from about $2x10^{-9}$ to $5x10^{-4}$ (ft/d)/ft ($d^{-1}$). Assuming an average thickness of the confining units of 100 feet, vertical hydraulic conductivity values range from about $2x10^{-7}$ to $5x10^{-2}$ ft/d. In general, the leakage coefficient is greatest near the updip limit of most units, where confining units typically consist of coarser sediments and are thinnest, and lowest toward the coast, where most units have undergone a facies change to finer sediments. For example, all confining units are relatively tight at Charleston: VCond values there for all confining beds are within a $2x10^{-9}$ to $1x10^{-8}$ (ft/d)/ft range. Although the relative distributions of leakage coefficients seem reasonable when compared to geologic information (Colquhoun and others, 1983; Aucott, Davis, and Speiran, 1987) few data are available to provide quantitative verification for these values.

The east-west variation of leakage coefficients along the coast for the confining unit overlying the Black Creek aquifer is evident in the model results. Figure 29 shows that the confining unit overlying the Black Creek aquifer in the western part of the lower Coastal Plain, the Black Mingo Formation, is much tighter than the Peedee Formation which comprises most of the confining unit overlying the Black Creek aquifer in the eastern part of the study area along the coast (fig. 28). This model result supports the general hypothesis that the direction of flow in the lower Coastal Plain in the Cretaceous aquifers is at least in part a result of the east-west difference in this confining unit.

Another interesting variation is again from east to west along the coast in the confining unit between the Black Creek and Middendorf aquifers (fig. 30). Final model results show this confining unit as significantly tighter in the west than in the east. This corresponds well with the limited available data that indicate water-level and water-quality differences between the Black Creek and Middendorf aquifers in the west that are much more significant than any differences found to the east (Aucott and Speiran, 1985a).

Streambed conductance, defined by McDonald and Harbaugh (1984) as the streambed hydraulic conductivity multiplied by the streambed area divided by the streambed thickness, ranged from about 10,000 to 100,000 $ft^2$/d in the model. The conductances of most streambeds were greater near the Fall Line than downstream. On the basis of an assumed average stream length per grid block of 4 miles, an average stream width of 500 feet, and an average thickness of 10 feet, the resulting range in streambed hydraulic conductivity is 1 to 10 ft/d.

These values appear to be reasonable compared to other published values from



page 57 follows

ground-water flow models (Weeks and others, 1965; Moore and Jenkins, 1966; Walton and others, 1967; MacNish and Barker, 1975; Barker and others, 1983). The average streambed conductivity from the above references is $2.5x10^{-5}$ ft/s (2 ft/d) which is within the range of values used in this model. As a result, the streambed conductivities in this model, which were derived largely from model calibration, are reasonable when compared to other situations. Because the model calibration is relatively insensitive to increases in streambed conductivity (fig. 22), the certainty of these values is somewhat reduced.

The average recharge to or discharge from the deep regional ground-water flow system is shown in figure 32. The rates of recharge and discharge shown in figure 32 represent the deep flow system as previously discussed, not the total flow system. Efforts to compute total recharge would require inclusion of the shallow flow system which is beyond the scope of this report.

Recharge to the deep flow system varies from 0 to 4 in/yr in the study area. It occurs primarily in updip interstream areas. Recharge rates are greatest in the western part of the upper Coastal Plain. This is supported by the generally greater base flows of small streams in this area (Bloxham, 1976; Stricker, 1983). Lower rates of recharge occur in the eastern part of the upper Coastal Plain as well as in small isolated parts of the lower Coastal Plain. Recharge to the deep flow system is possible in the lower Coastal Plain only in areas of relatively high land surface altitude such as northern Horry and Columbus Counties and northern Berkeley County. The leveed Lake Moultrie in Berkeley County also acts as a source of recharge in addition to adjacent topographically high areas.

Recharge was adjusted significantly from original values. The major change involved increasing recharge in the western part of the model and decreasing it in the east. This is generally supported by base flow calculations (Bloxham, 1976; Stricker, 1983; Aucott, Meadows, and Patterson, 1986). Because the simulation is so sensitive to changes in recharge, numerous changes were also made near the Fall Line to improve the calibration.

Discharge from the deep flow system is primarily to large streams in the upper Coastal Plain. It varies from 0 to 19 in/yr. Discharge is generally greatest from large streams in the western part of the upper Coastal Plain. Diffuse upward leakage to the surficial aquifer in most of the lower Coastal Plain and to the Atlantic Ocean also occurs as a result of discharge from the deep flow system. These rates are less than 1 in/yr everywhere.

No independent verification by field measurement was possible for recharge. Although heads and river flows were used as calibration criteria, recharge values should be considered to be one of the least reliable of the calibrated parameters. Discharge rates are somewhat more reliable despite the coarseness of the calculations for aquifer-river discharge (Aucott, Meadows, and Patterson, 1986).

The simulated water budget is summarized in figure 33. This water budget represents flow in the deep flow system only. The five model layers are represented by the five boxes in the center of the figure. A general explanation is shown at the bottom of the figure. Arrows pointing into each box represent flow into that layer. Arrows pointing out represent flow out of that layer. Recharge to each layer is represented by the number to the extreme left of each box. Discharge to rivers in the upper Coastal Plain from each layer is represented by the number to the right of each box. Numbers between the boxes represent flow between the aquifers and through confining units. Values at the corners of each box represent lateral flow across model boundaries (from the upper right-hand corner clockwise) toward or from the northeast, southeast, southwest and northwest.

Simulated recharge to the ground-water flow system includes 789 $ft^3/s$ of direct recharge and 8 $ft^3/s$ of downward leakage from the overlying surficial aquifer. Simulated discharge is 735 $ft^3/s$ to rivers in the upper Coastal Plain and 50 $ft^3/s$ to the overlying surficial aquifer in the lower Coastal Plain. Leakage from and to the source-sink part of the Floridan aquifer system is 7 $ft^3/s$ and 14 $ft^3/s$, respectively. Lateral flow across all boundaries in all active layers is 21 $ft^3/s$ into the modeled area and 26 $ft^3/s$ out of the modeled area. This results



page 61 follows

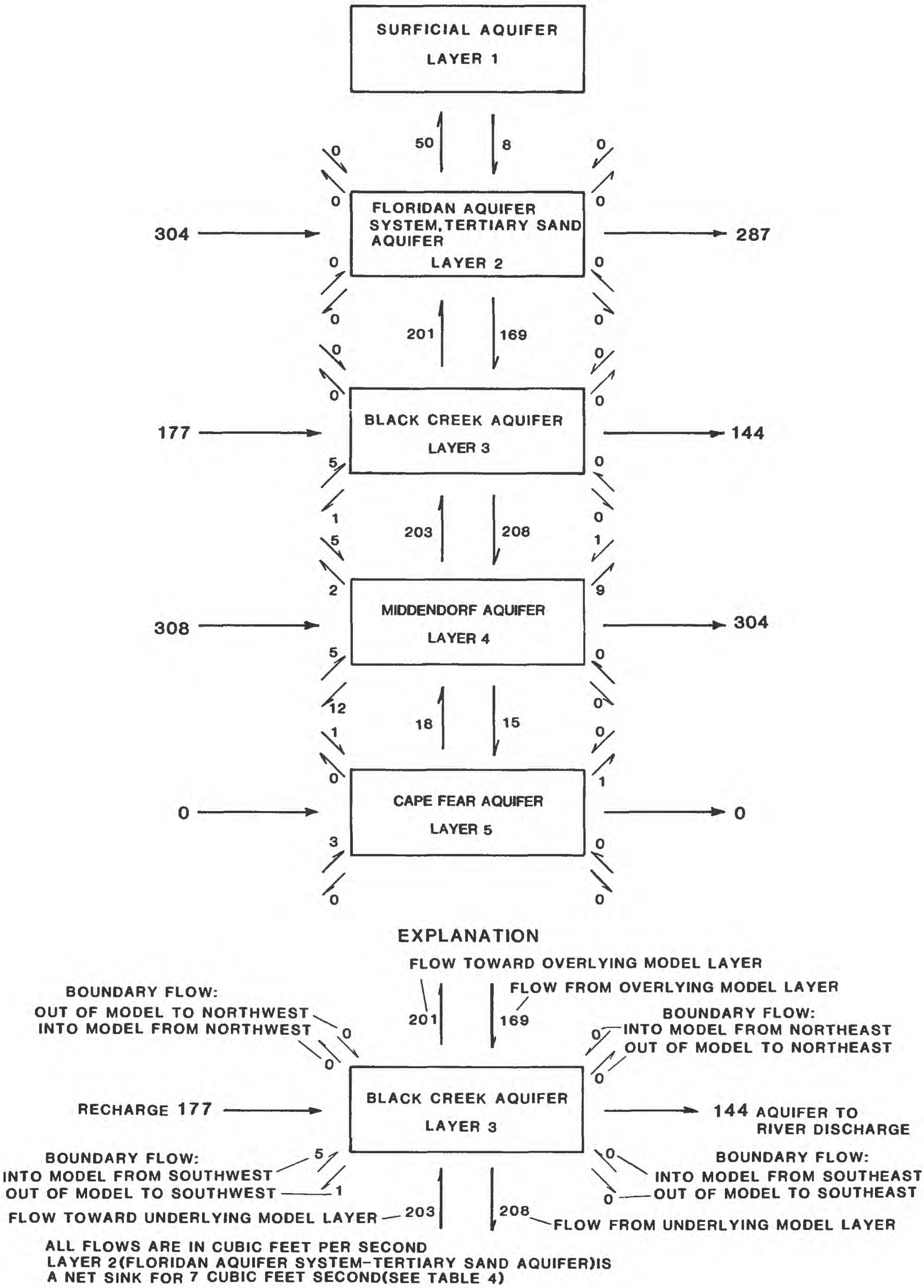


Figure 33. – Model-simulated water budget for the deep flow system.

in a total inflow and outflow to the deep ground-water flow system of the Coastal Plain aquifers of 825 $ft^3/s$ prior to development (table 4).

Flow between aquifers is quantitatively important to the operation of the total flow system (fig. 33). The Cape Fear aquifer, through which little water flows, is not a significant part of the flow system. Much of the downward leakage from the source-sink parts of the surficial aquifer and Floridan aquifer system occur near their updip limit in the upper Coastal Plain. Upward leakage to the sources and sinks is less per square mile but occurs over a large area throughout the lower Coastal Plain.

The total flow across the lateral boundaries of 21 $ft^3/s$ into and 26 $ft^3/s$ out of the active model area is of minor importance in comparison to the 825 $ft^3/s$ in the overall flow system. The lateral boundaries selected are thus a good choice in that they do not have a significant regional influence on the flow system. The relatively insignificant effect of these boundaries confirms that their selection was reasonable. Boundary flows, however, can be important in the functioning of parts of the flow system.

The total simulated inflow and outflow (825 $ft^3/s$) represents only flow in the deep flow system. Flow in the shallow system such as that to small streams in the upper Coastal Plain cannot be simulated at the scale of this model. Total flow in the entire ground-water flow system (shallow and deep) is much greater than that simulated for the deep system alone. A rough relation between the total ground-water flow in the system and deep flow as computed by the model can be seen by the relation of total base flow versus deep flow discharging to major streams in the upper Coastal Plain. Average total base flow of six small streams in the upper Coastal Plain of South Carolina was calculated to be 0.9 $ft^3/s/mi^2$ by Stricker (1983). Deep flow discharging to six large streams in the upper Coastal Plain of South Carolina and North Carolina was calculated to be 0.11 $ft^3/s/mi^2$ by Aucott, Meadows, and Patterson, (1986). These figures clearly show that the deep flow simulated by the model represent only a small part of that in the entire flow system.

Table 4.—*Summary of model simulated inflow and outflow*

[All flows are in cubic feet per second]

| | Inflow | | Outflow |
|---|---|---|---|
| 789 | Recharge | 735 | Rivers |
| | Source-sink | | Source-sink |
| 8 | From surficial aquifer | 50 | To surficial aquifer |
| 7 | From Floridan aquifer system | 14 | To Floridan aquifer system |
| 21 | Lateral boundaries | 26 | Lateral boundaries |
| 825 | Total | 825 | Total |

## SUMMARY AND CONCLUSIONS

A wedge of sand, silt, clay, and limestone sediments is present beneath the Coastal Plain of South Carolina. These sediments have been subdivided into six regional aquifer units: the surficial aquifer, the Floridan aquifer system, the Tertiary sand aquifer, the Black Creek aquifer, the Middendorf aquifer, and the Cape Fear aquifer. Intervening confining units separate the aquifers, except for the Floridan aquifer system and the Tertiary sand aquifer which act together as a single hydrologic unit.

The sources of recharge to the Coastal Plain aquifers are precipitation in the aquifer outcrop areas and leakage from overlying and underlying aquifers. Prior to development, discharge from the aquifers was to streams, to overlying and underlying aquifers, and to the ocean. The ground-water flow system in all aquifers was dominated in the upper Coastal Plain by flow toward rivers and streams.

In the lower Coastal Plain, ground-water flow patterns in individual aquifers were quite different. Ground water in the Floridan aquifer system and the Tertiary sand aquifer flowed generally perpendicular to the coast in the lower Coastal Plain. Flow in the lower Coastal Plain part of the Cretaceous aquifers was to the east, almost parallel to the coast and toward southeastern North Carolina. The reason for this eastward flow is the existence of more effective discharge areas in the east. This results from three factors: (1) confining units are less effective to the east; (2) the Cretaceous aquifers in the east are closer to the land surface and, thus, in relatively good hydraulic contact with the rivers farther into the lower Coastal Plain; and (3) rivers in the east are lower in altitude farther upstream.

The ground-water flow system of the Coastal Plain aquifers can be divided into shallow and deep flow systems. Because the scale of this effort involved the study of regional flow systems and the complexities involved with shallow flow systems, only the deep flow system was described.

A ground-water flow model, which has an evenly spaced grid mesh of 4 miles, was constructed to simulate the predevelopment flow system. Boundaries are no-flow at the updip limit of the units, at the base of the Coastal Plain sediments, and at the saltwater-freshwater interface. The lateral boundaries in Georgia and North Carolina are predominantly constant head. The surficial aquifer and parts of the Floridan aquifer system previously modeled in other studies are considered as source-sink layers and simulated with constant heads.

The model was calibrated with aquifer water levels and streamflow data. The mean absolute difference between simulated water levels and water levels contoured from field data is 8.9 feet. Because of the irregular potentiometric surface and 25-foot contour interval, the match between simulated and observed water levels is considered to be acceptable. The general shape of the simulated potentiometric surfaces and directions of ground-water flow also matched well with those derived from field data. Simulated aquifer-to-stream flows generally were within the probable error of the base-flow calculations. Because of the multiple variables in the solution of the flow equation, this solution, while reasonable, cannot be considered unique.

Testing was performed to determine the sensitivity of the model to changes in various parameters. The simulation appears to be most sensitive to increases in recharge, changes in aquifer transmissivity in aquifer outcrop areas, and changes in the vertical hydraulic conductivity of confining units. The model was least sensitive to increases in streambed conductance, decreases in recharge, changes in transmissivity outside outcrop areas, and changes in the position of the saltwater-freshwater interface.

The output of the ground-water flow model is the water budget, streambed conductances and the distributions of recharge, transmissivity, and leakage coefficients of the confining units. Simulations indicate that total recharge and discharge in the deep ground-water flow system is 825 $ft^3/s$. Simulated direct recharge in outcrop areas is 789 $ft^3/s$. The remainder of total recharge is from leakage from overlying source-sink beds (15 $ft^3/s$) and inflow across boundaries (21 $ft^3/s$). Discharge to the upper Coastal Plain rivers is 735 $ft^3/s$, whereas the remaining discharge is by upward leakage to the overlying source-sink beds

($64\ ft^3/s$) and by outflow across lateral boundaries ($26\ ft^3/s$).

Simulated transmissivities differ little from original estimates because the initial estimates from aquifer test and specific-capacity data were generally adequate and because of the relative insensitivity of the model to changes in transmissivity, except in the extreme updip part of the flow system. Aquifer transmissivities of all aquifers ranged from less than 1,000 to 30,000 $ft^2/d$. Leakage coefficients of the confining units were adjusted considerably from initial estimates because there were insufficient data to establish realistic initial estimates and because the model is relatively sensitive to changes in this parameter. On the basis of an assumed average confining unit thickness of 100 feet, vertical hydraulic conductivities of the confining units ranged from $2x10^{-7}$ to $5x10^{-2}$ ft/d.

This model was calibrated under predevelopment steady-state conditions. No model verification was performed. Subsequent work that uses transient simulations of conditions from predevelopment to November 1982 is planned to improve and verify this predevelopment model under stressed conditions.

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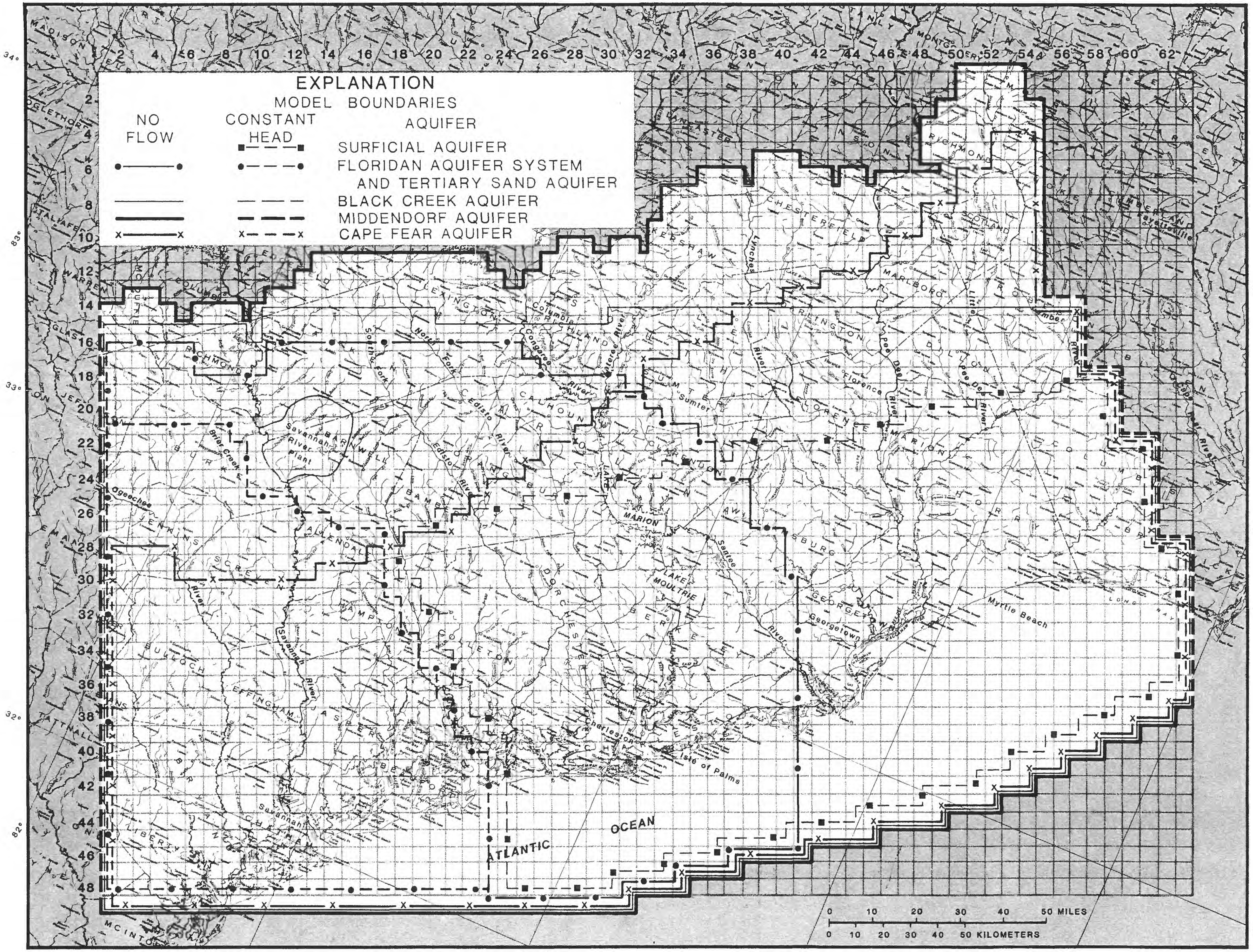


Figure 11.—Model grid and specified boundary conditions.

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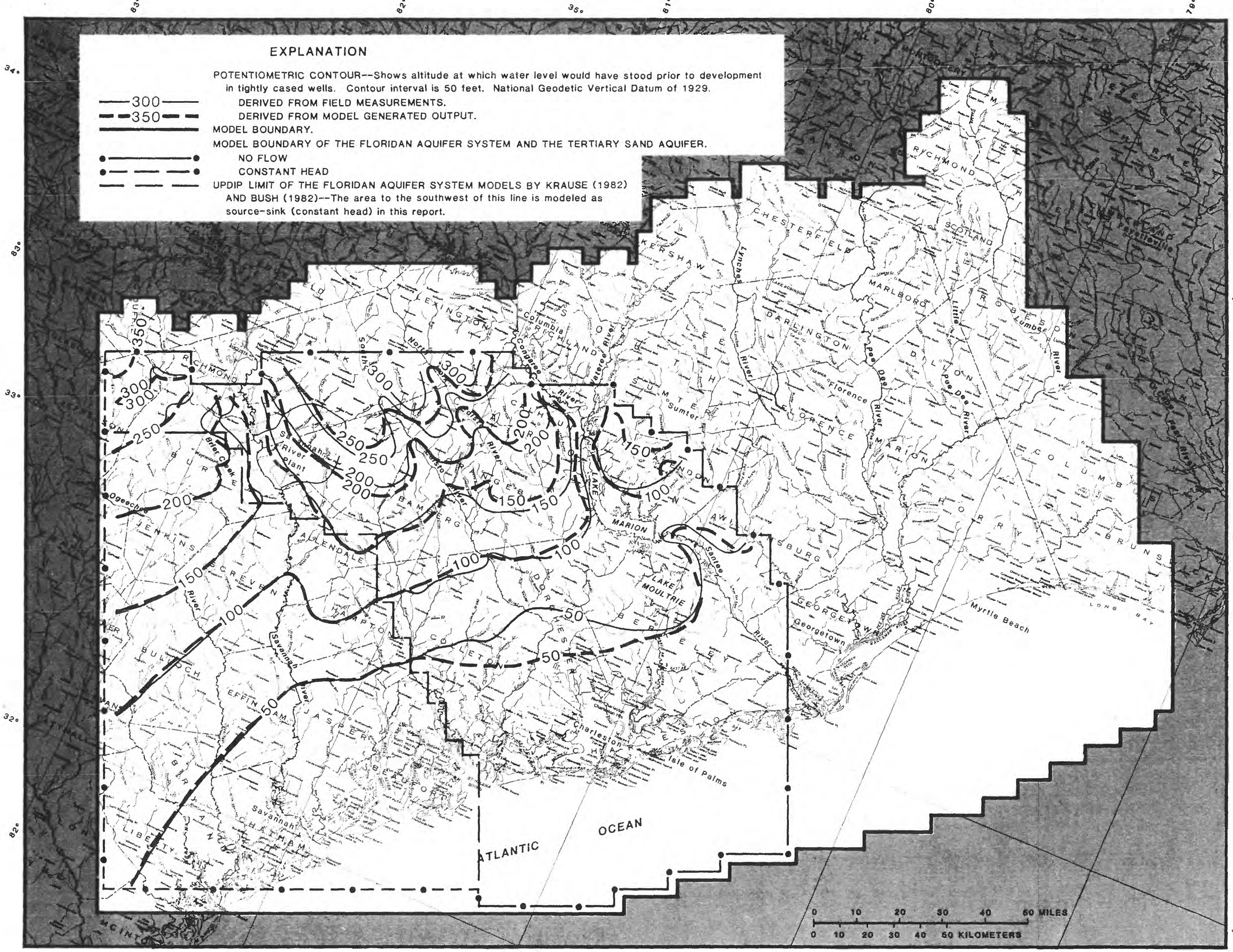


Figure 14.—Observed and simulated water levels for the Floridan aquifer system and the Tertiary sand aquifer.

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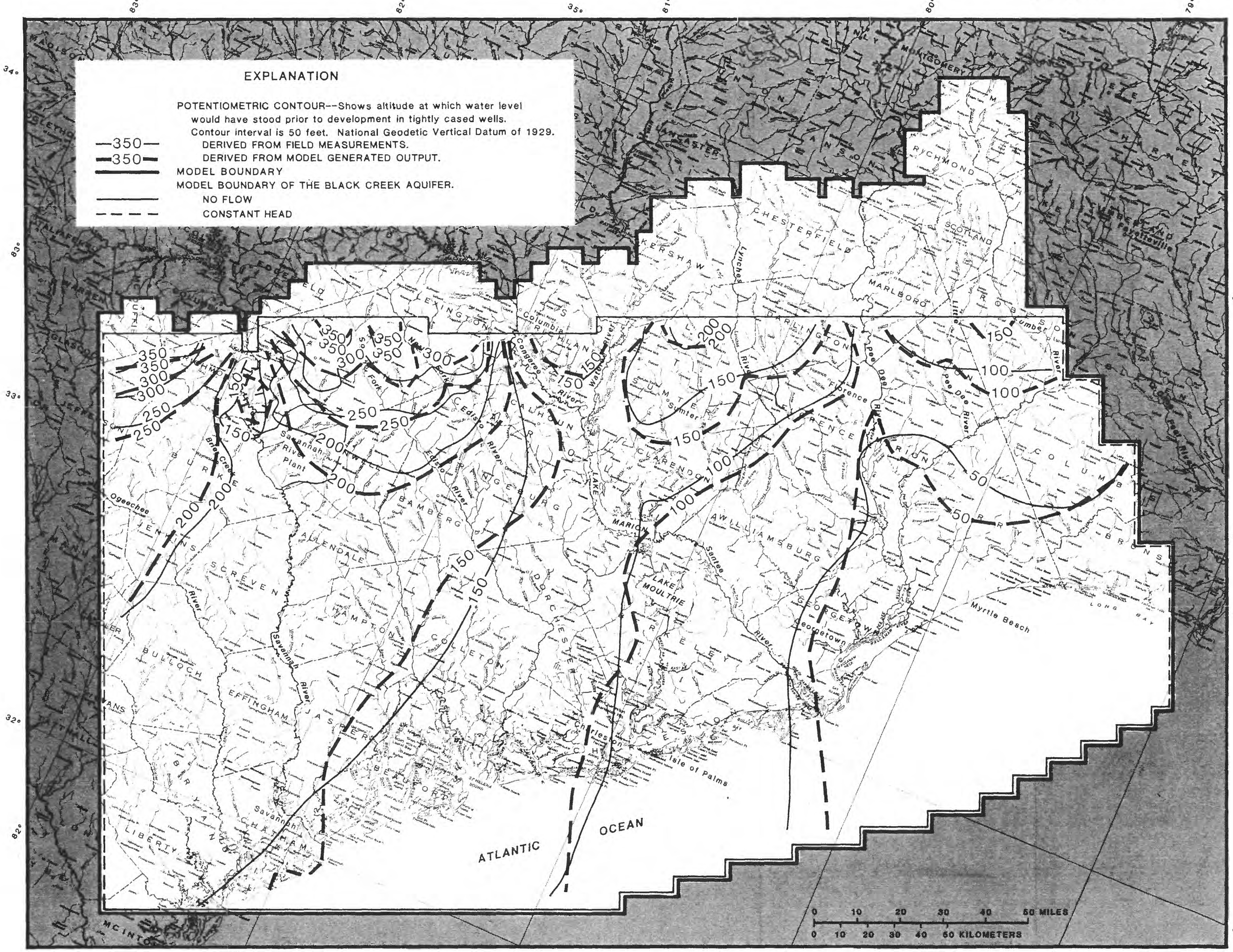


Figure 15.—Observed and simulated water levels for the Black Creek aquifer.

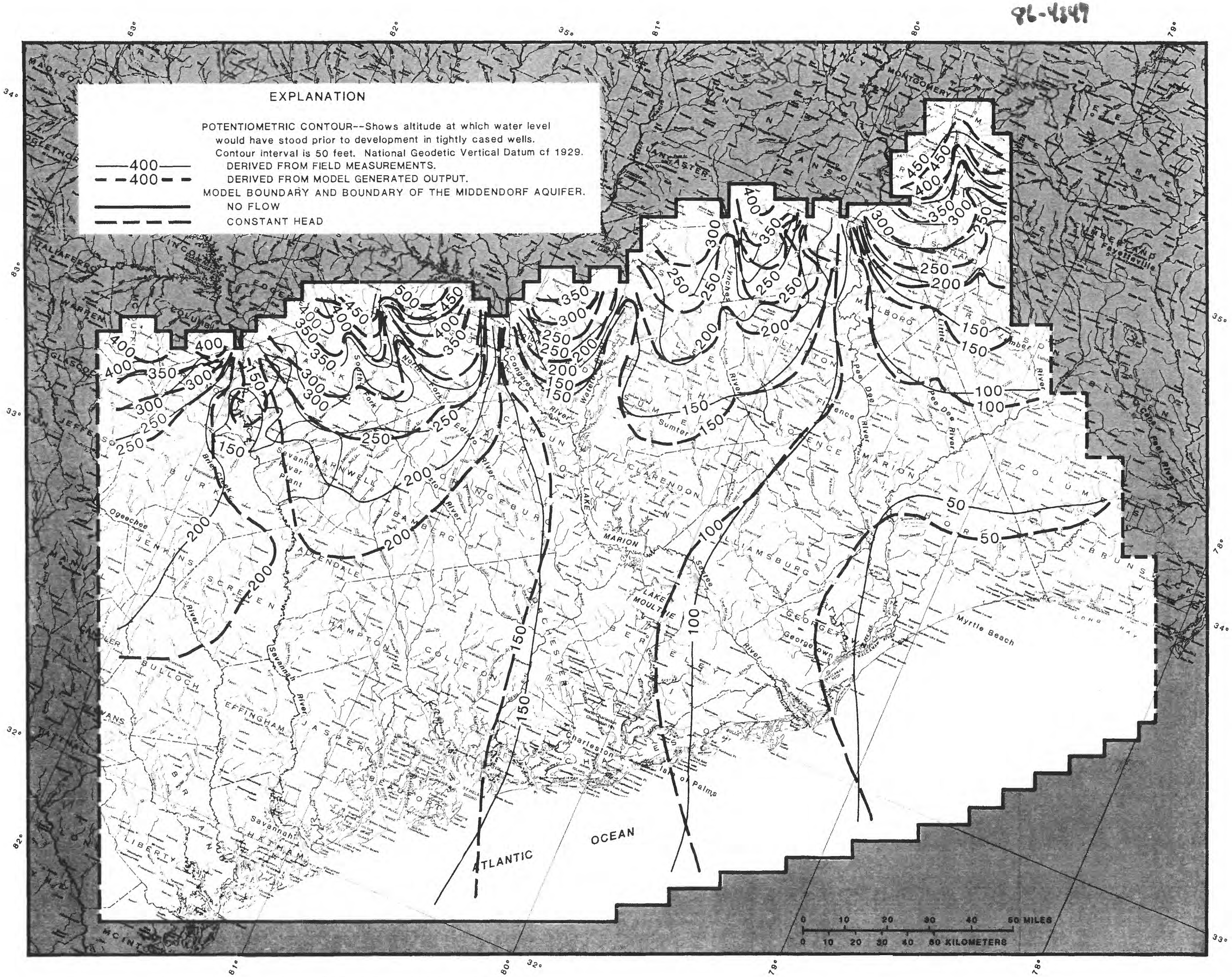


Figure 16.—Observed and simulated water levels for the Middendorf aquifer.

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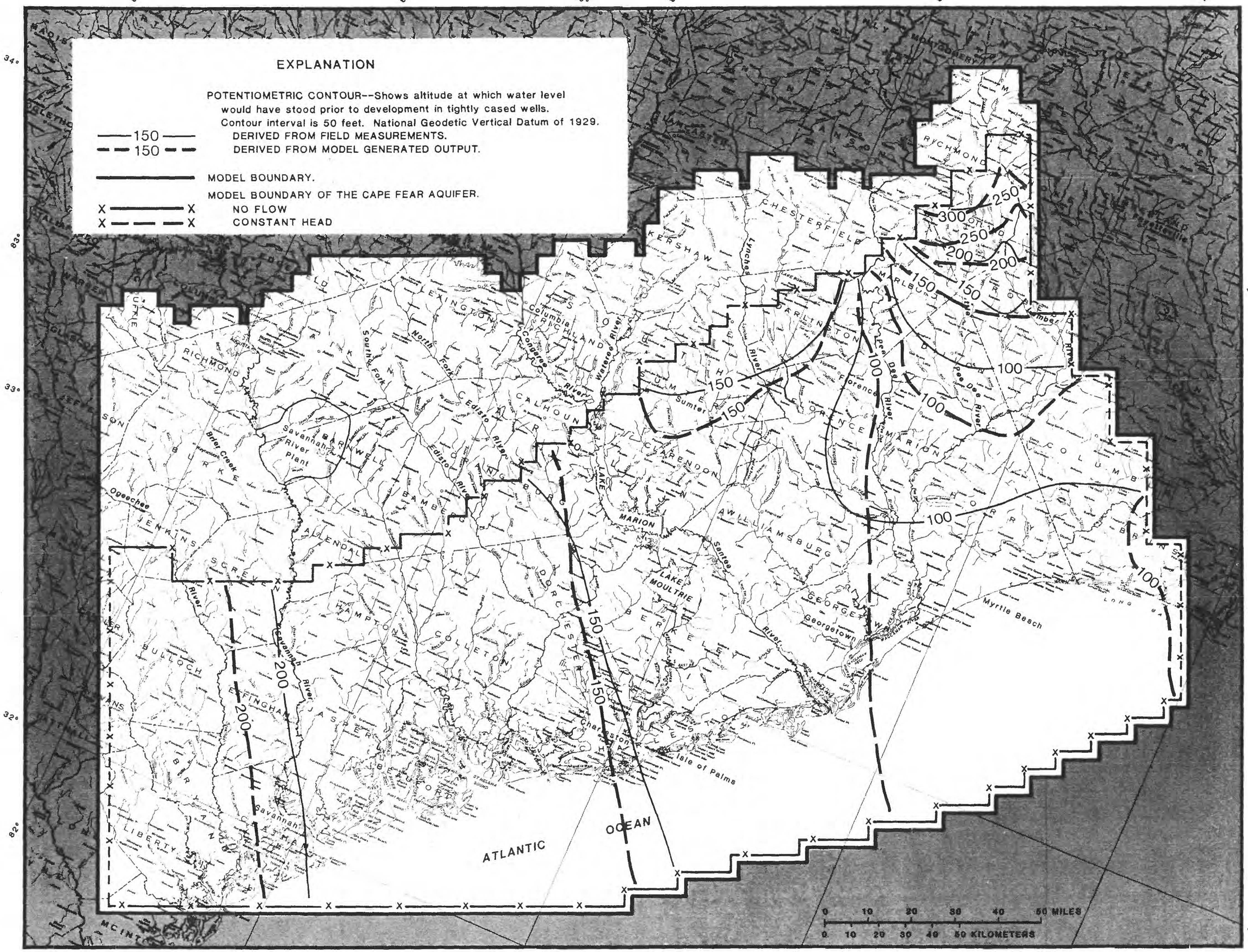


Figure 17.—Observed and simulated water levels for the Cape Fear aquifer.

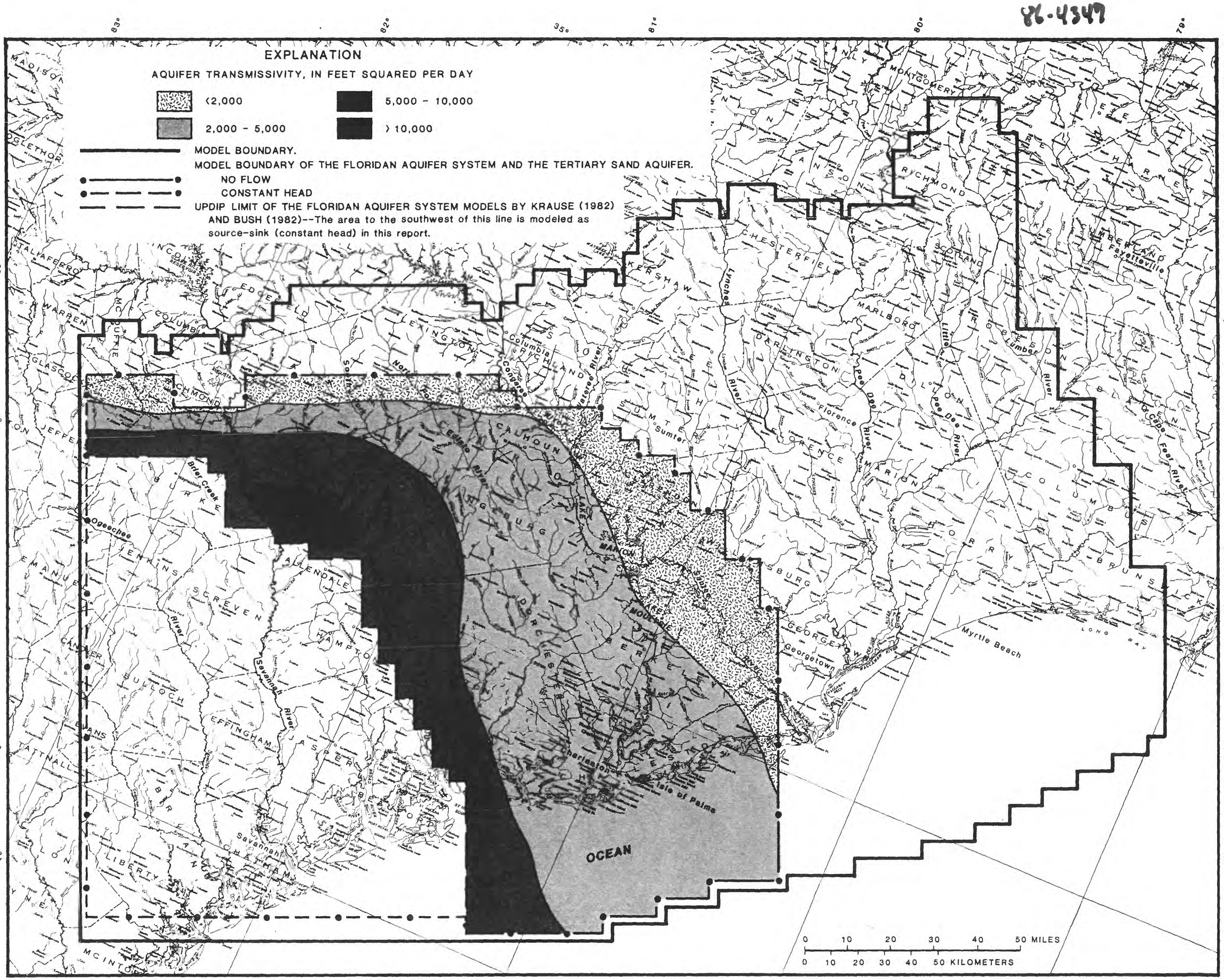


Figure 24.—Simulated transmissivity of the Floridan aquifer system and the Tertiary sand aquifer.

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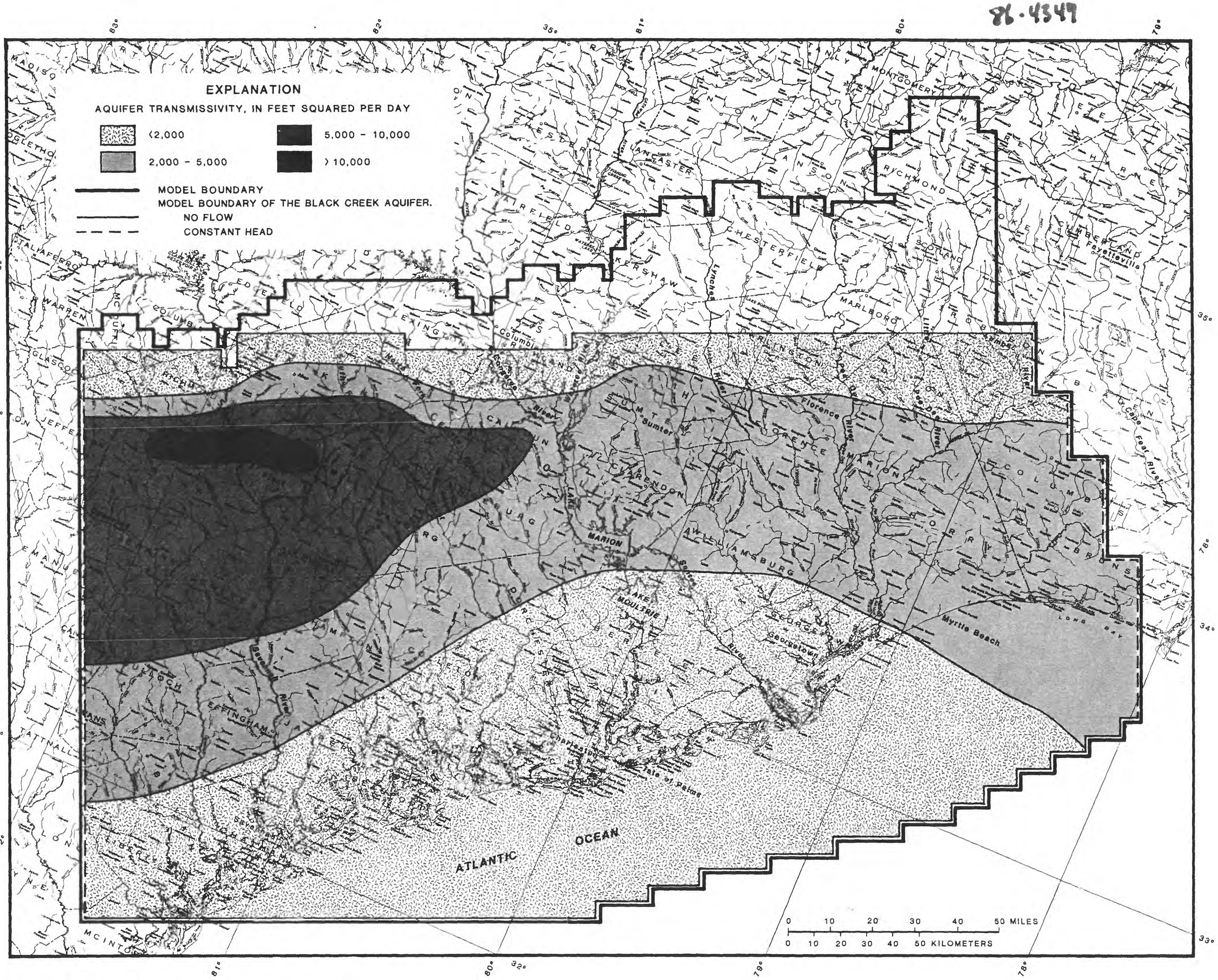


Figure 25.—Simulated transmissivity of the Black Creek aquifer.

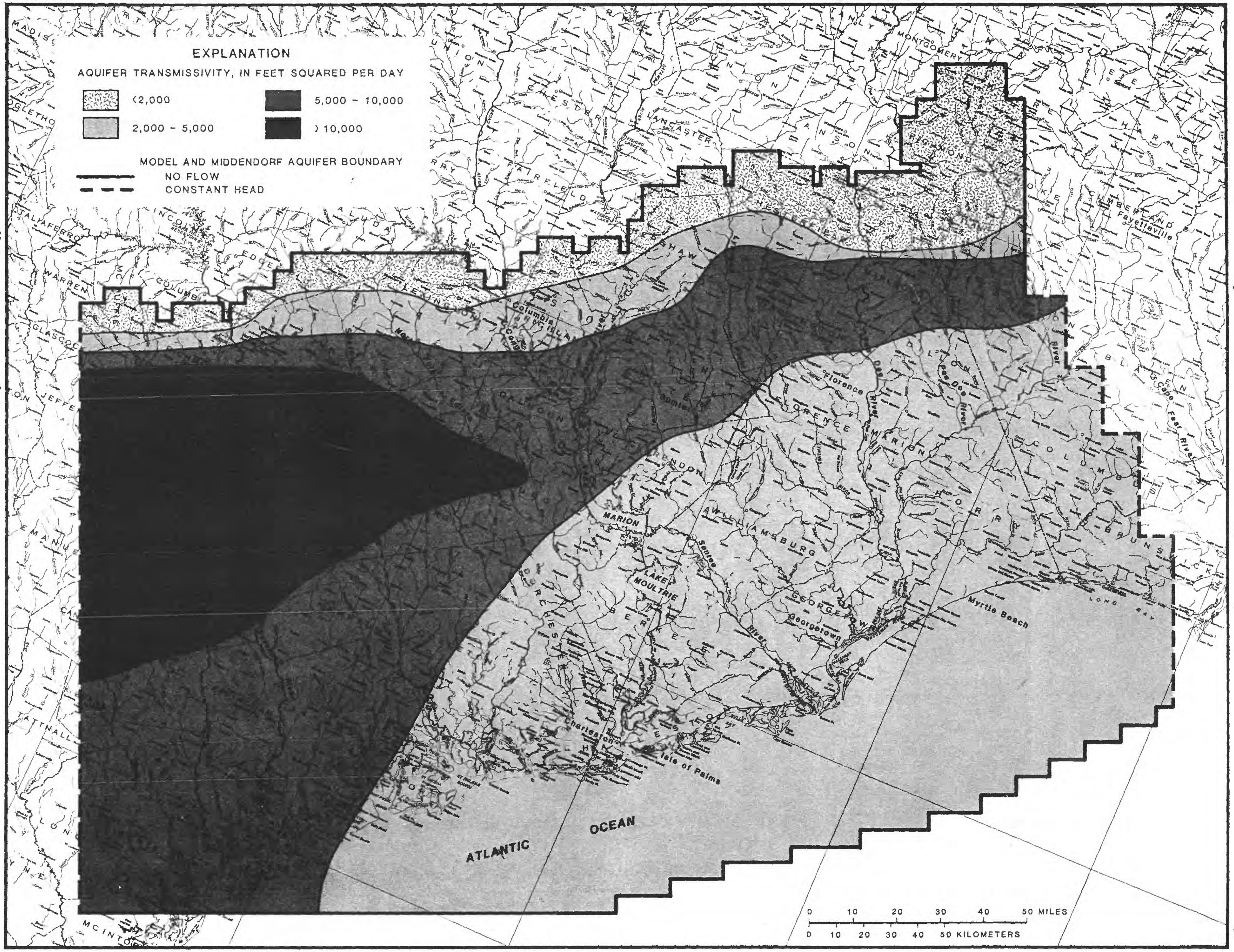


Figure 26. – Simulated transmissivity of the Middendorf aquifer.

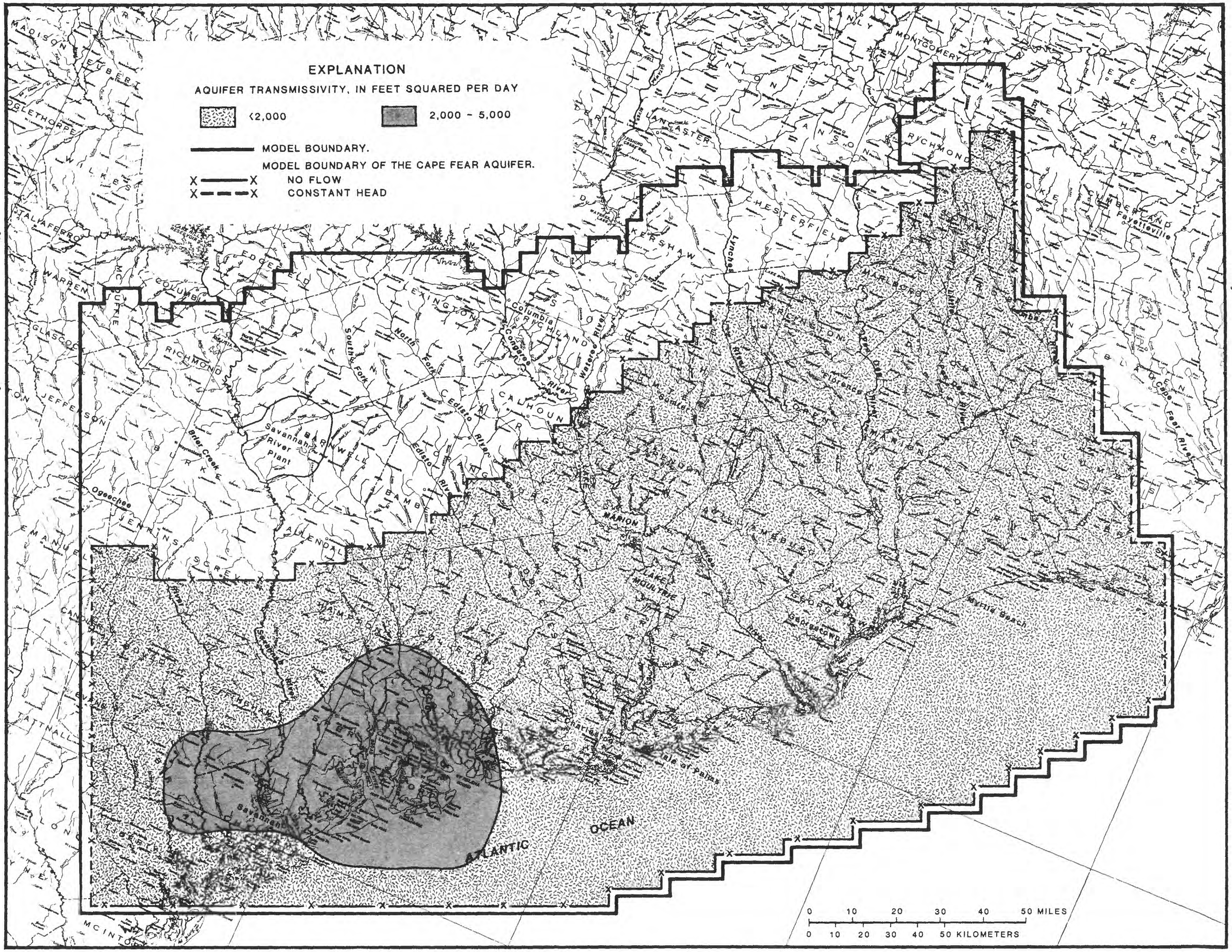


Figure 27.—Simulated transmissivity of the Cape Fear aquifer.

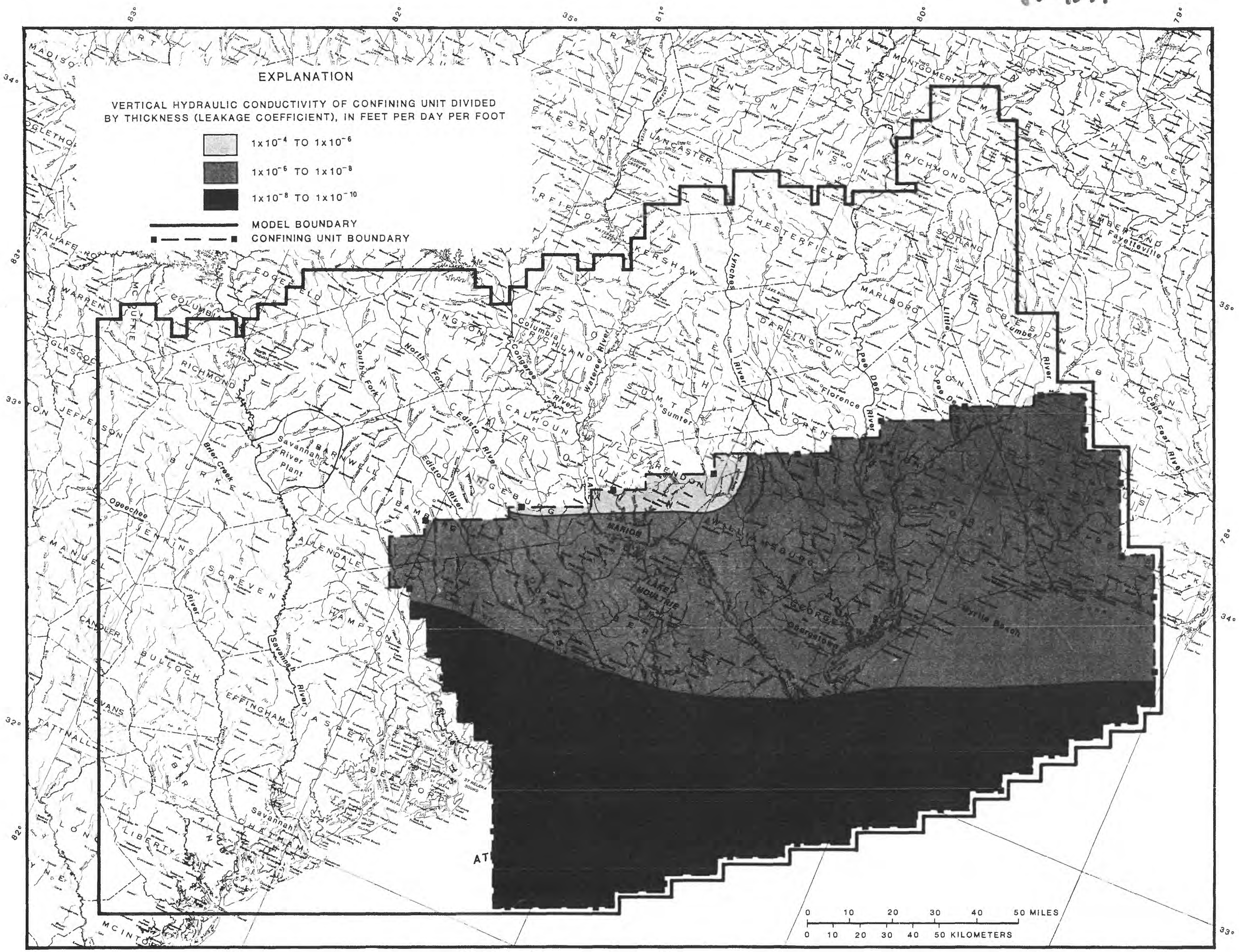


Figure 28. – Simulated leakage coefficient of the confining unit between the surficial aquifer and underlying units.

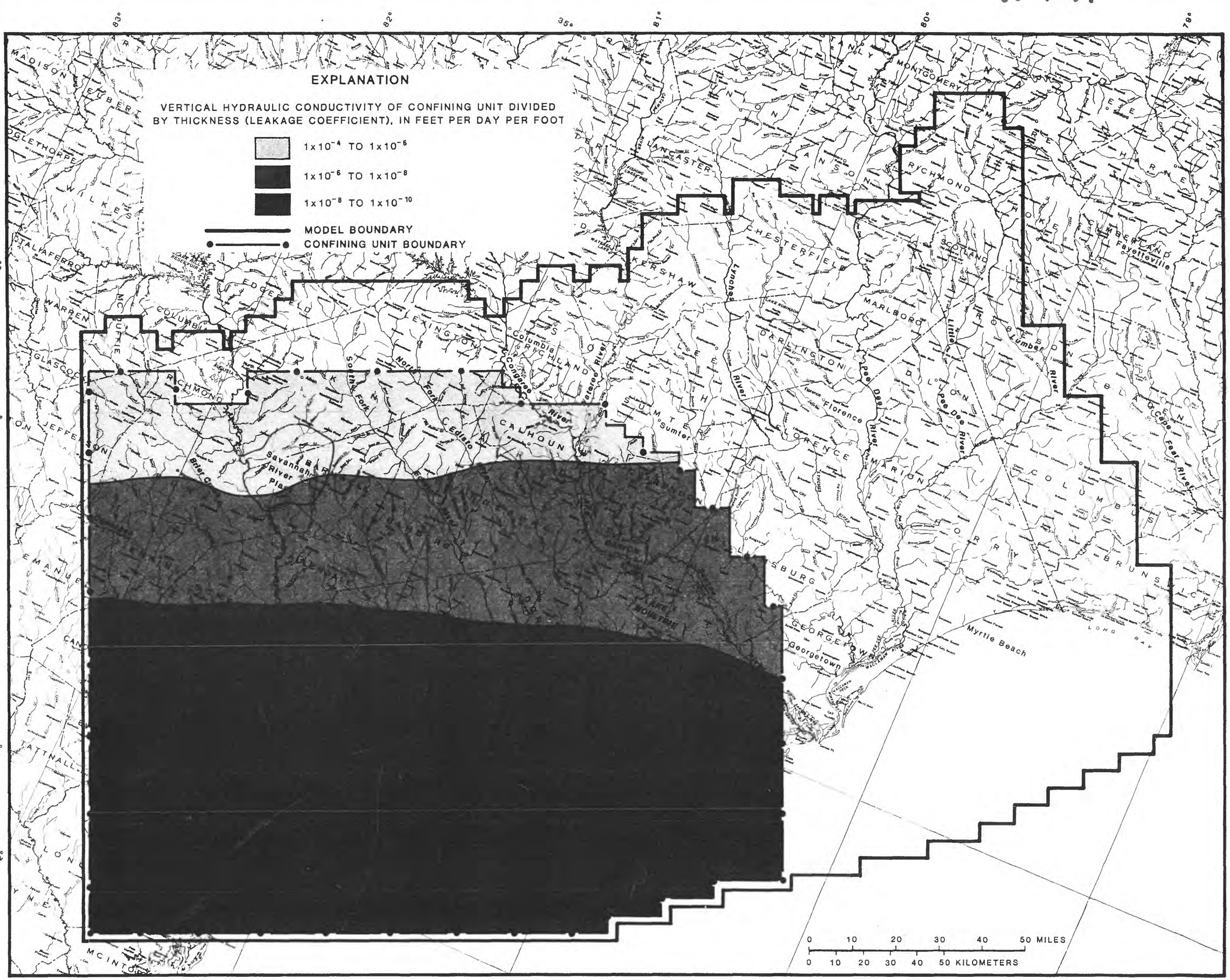


Figure 29. – Simulated leakage coefficient of the confining unit between the Floridan aquifer system/Tertiary sand aquifer and the Black Creek aquifer.

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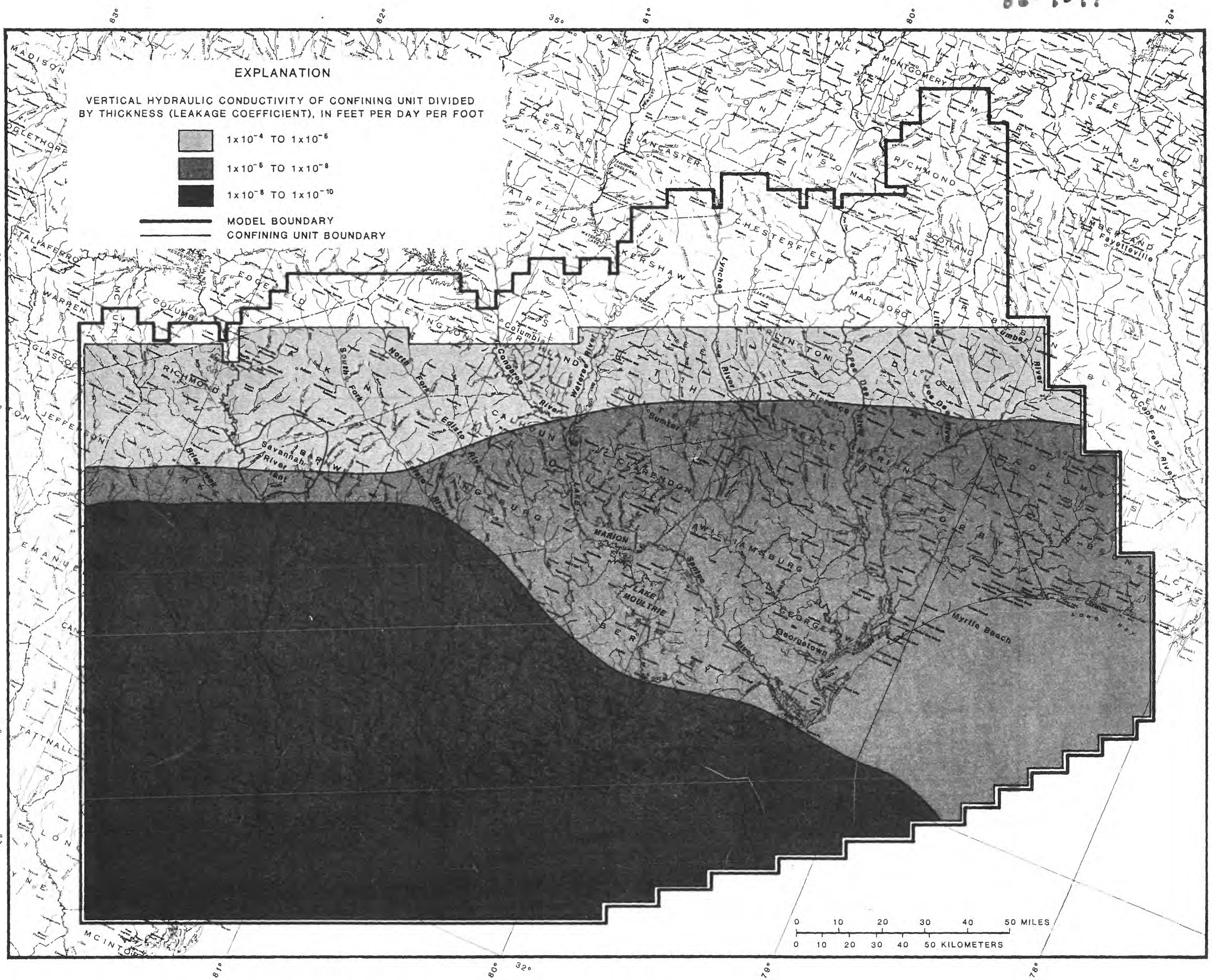


Figure 30. – Simulated leakage coefficient of the confining unit between the Black Creek aquifer and the Middendorf aquifer.

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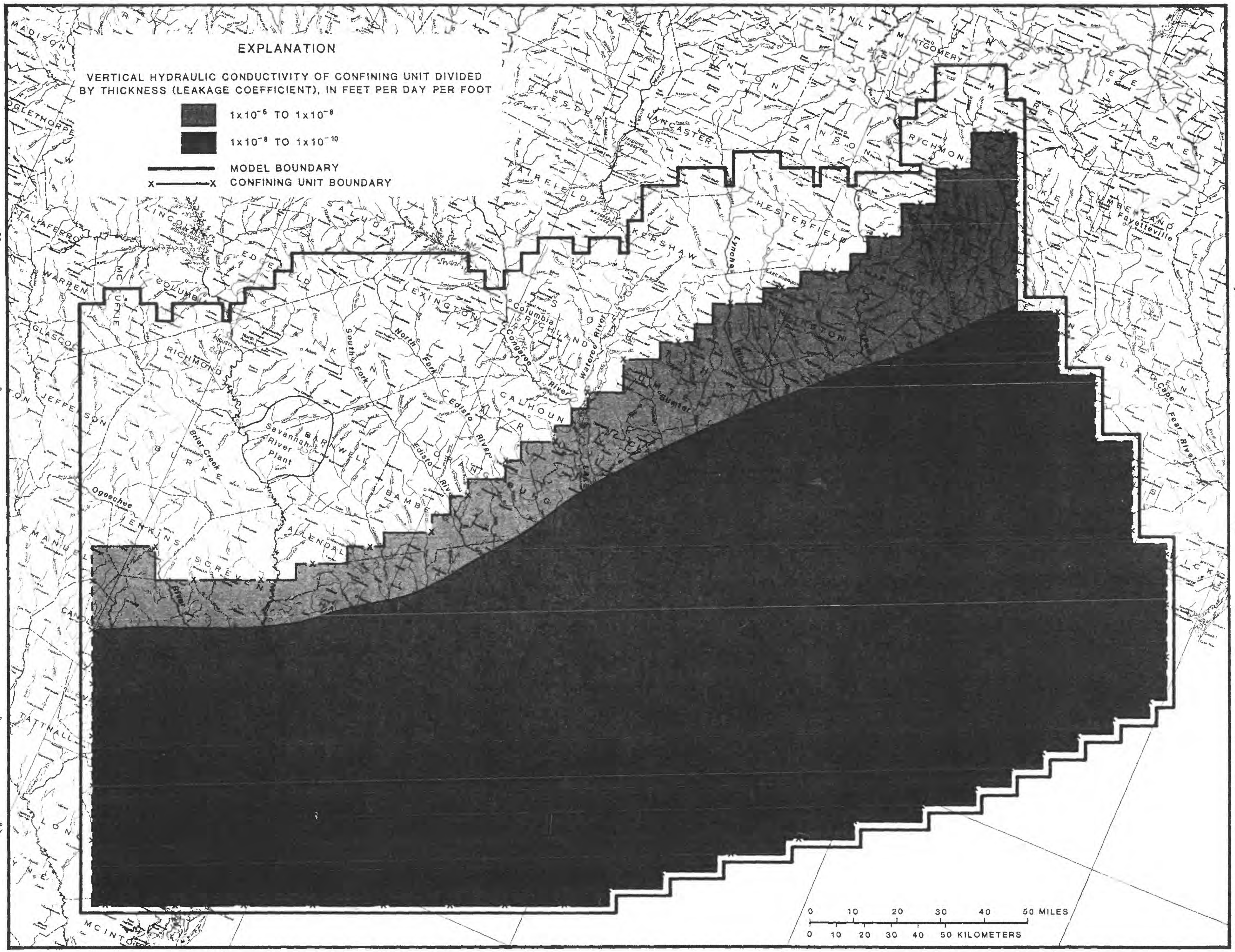


Figure 31.—Simulated leakage coefficient of the confining unit between the Middendorf aquifer and the Cape Fear aquifer.

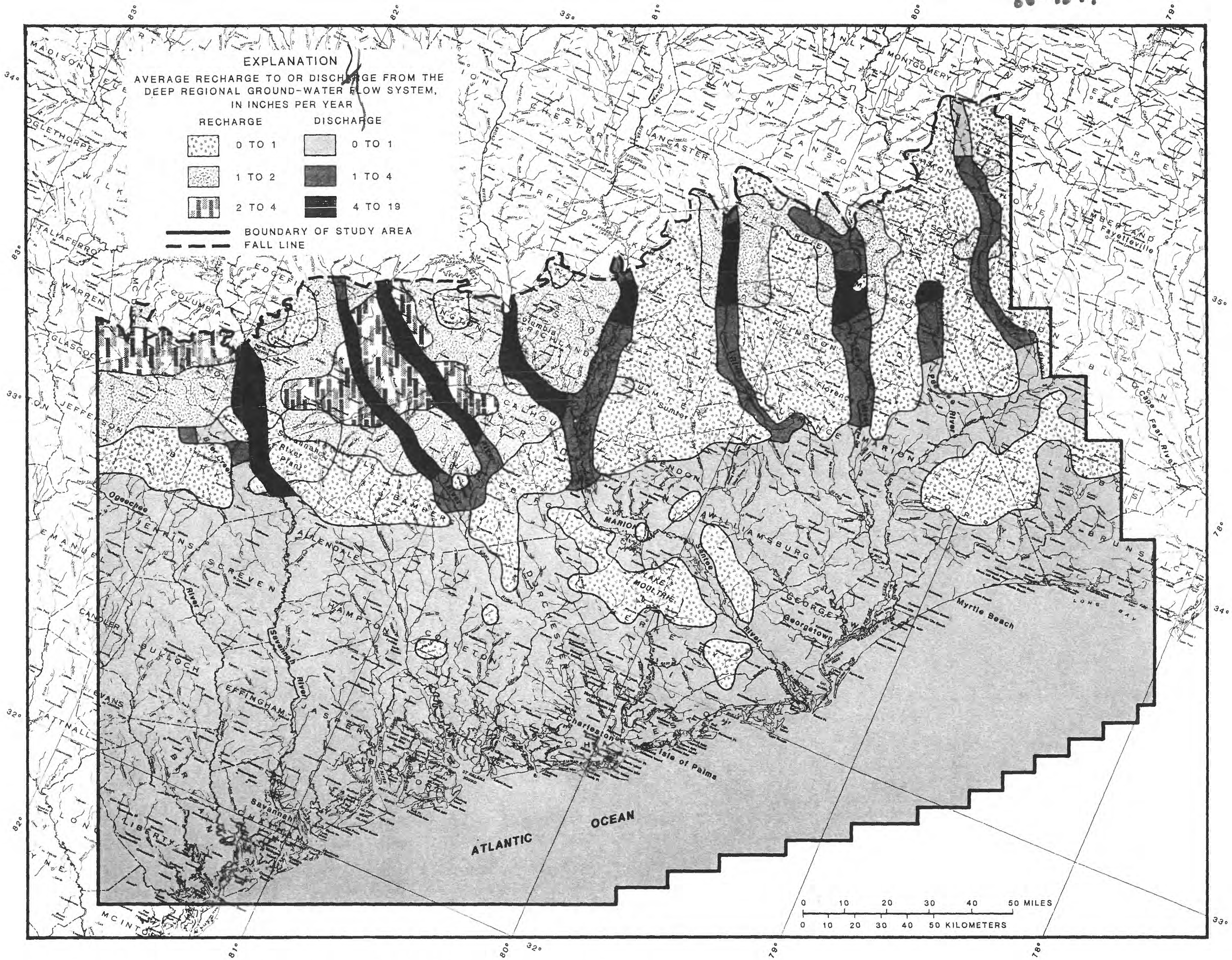


Figure 32.—Simulated average recharge to and discharge from the deep regional ground-water flow system.