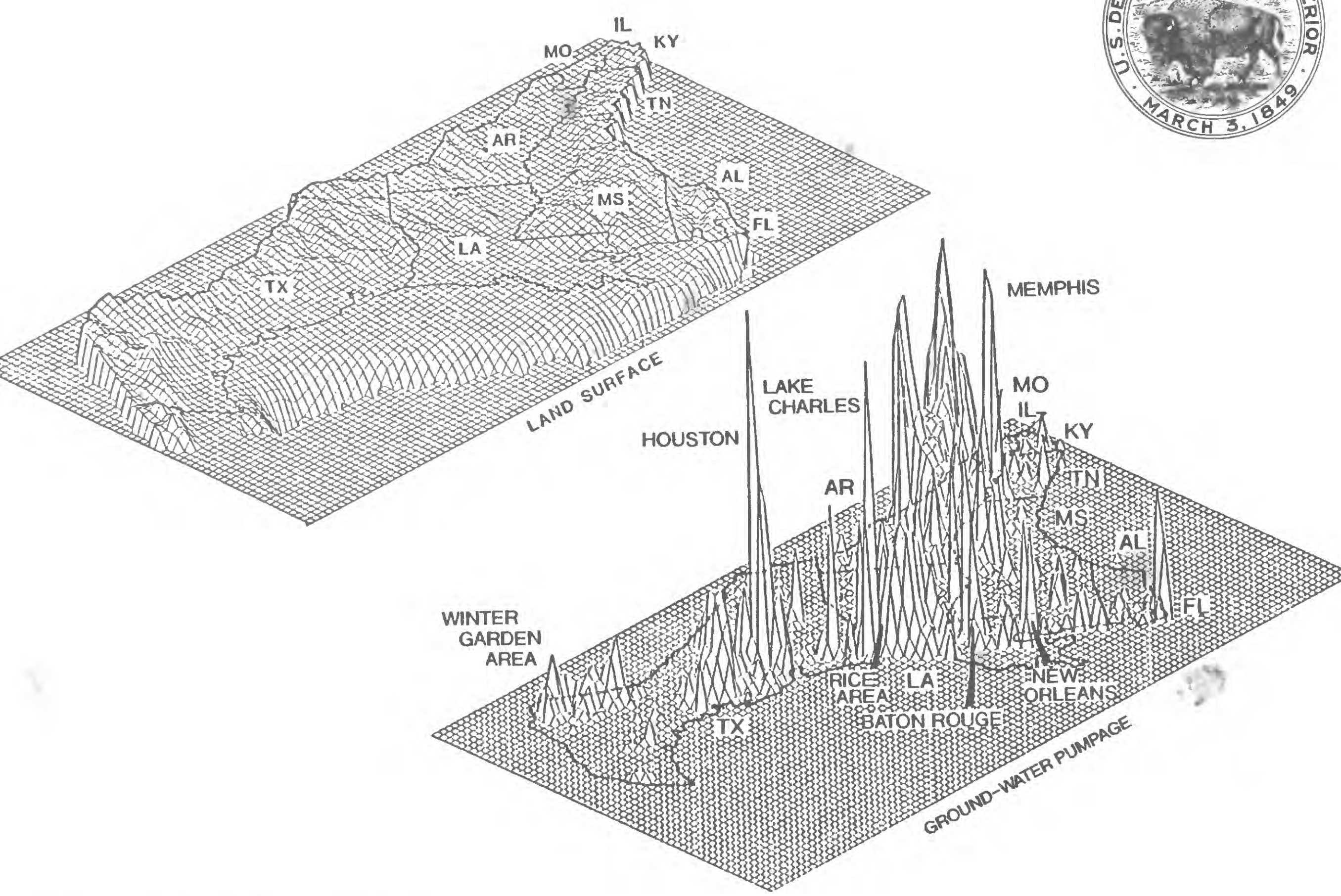
GROUND-WATER FLOW IN THE GULF COAST AQUIFER SYSTEMS, SOUTH CENTRAL UNITED STATES--A PRELIMINARY ANALYSIS

U.S. GEOLOGICAL SURVEY
Water-Resources Investigations Report 89-4071



A Contribution of the
Regional Aquifer-Systems Analysis
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By Alex K. Williamson, Hayes F. Grubb, and Jonathan S. Weiss

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Austin, Texas
1990

DEPARTMENT OF THE INTERIOR

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ABBREVIATIONS AND CONVERSION FACTORS

Factors for converting inch-pound units to International System of Units (SI) and abbreviations of units

Multiply inch-pound units	By	To obtain SI units
inch (in.)	25.40	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer
million acre-feet	1,233	cubic hectometer
inch per year (in/yr)	25.4	millimeter per year
million gallons per day (Mgal/d)	0.04381	cubic meter per second
billion gallons per day (Bgal/d)	43.81	cubic meter per second

Chemical concentrations, water density and temperature are given in metric units. Chemical concentration is given in milligrams per liter (mg/L). Milligrams per liter is a unit expressing the concentration of chemical constituents in solution as weight (milligrams) of solute per unit volume (liter) of water. For concentrations of dissolved solids less than 7,000 mg/L, the numerical value is the same as for concentrations in parts per million. Water density is given in grams per cubic centimeter (g/cm^3). Water temperature is given in degree Celsius (°C), which can be converted to degrees Fahrenheit (°F) by the following equation:

$$°F = 1.8(°C) + 32$$

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

GROUND-WATER FLOW IN THE GULF COAST AQUIFER SYSTEMS SOUTH-CENTRAL UNITED STATES--A PRELIMINARY ANALYSIS

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ABSTRACT

A major objective of the Gulf Coast Regional Aquifer-System Analysis is to use digital models of regional ground-water flow systems to develop better understanding and to enable better management of the resource. Modeling is the best available tool to synthesize most of the known information about the aquifer systems and to test hypotheses about the relative importance of the components of the systems. The 290,000-square-mile study area in the Gulf of Mexico Coastal Plain includes the Mississippi embayment, the Gulf Coastal Plain of Texas, and offshore areas beneath the Continental Shelf that are underlain by deposits of Tertiary and younger age which contain fresh and saline water. A 10-layer, finite-difference, variable-density model, with blocks 10 miles on a side, was used to simulate ground-water flow before development (about 1900) and in 1980, assuming steady-state conditions for both, and transient conditions from 1935-1980.

Preliminary results indicate that the major factors controlling predevelopment regional flow are the topography and the outcrop pattern and geometry of aquifers and confining units. Geologic structure and the distribution of precipitation were less significant factors. Regional recharge areas include the hills east of the Mississippi River Alluvial Plain and areas in a band parallel to and about 75 miles inland from the coast where the distance between major rivers is largest. Major regional discharge areas are the low-lying, nearly flat Mississippi River Alluvial Plain, the coastal lowlands, and major river valleys.

The density of saline water in the deeper parts of the aquifer system probably has a substantial effect on regional ground-water flow and this effect extends even into the freshwater part of the system. Variable water density resulting from temperature and salinity variations may be a significant driving force that transports saline water from salt formations great distances in many directions, including updip.

The distribution and rates of regional recharge and discharge have been substantially changed by development. Ground-water pumpage in 1980 was about five times the value of predevelopment regional recharge. About 80 percent of the pumpage was being supplied from increased regional recharge, with lesser amounts from decreased natural discharge and aquifer storage. Simulation of 1980 conditions, assuming no change in storage in the aquifer system, resulted in a reasonable match of simulated and measured heads. Results also indicate that resistance to vertical flow caused by many fine-grained beds within the permeable zones can be as important as resistance caused by regional confining units.

Manuscript approved for publication May 5, 1989

INTRODUCTION

The Regional Aquifer-System Analysis (RASA) program was started in 1978 by the U.S. Geological Survey after a congressional mandate to develop quantitative appraisals of the major ground-water systems of the United States. Major objectives of the RASA program are to analyze and develop an understanding of the ground-water flow system on a regional scale, and to develop predictive capabilities that will contribute to effective management of the system (Bennett, 1979). The use of computer simulation of ground-water flow in the aquifer system under natural, undisturbed conditions, and conditions affected by human activities, is a major tool of the RASA studies to reach this objective.

Natural hydrologic boundaries, rather than political boundaries, have been used to determine areas to be studied. The Gulf Coast Regional Aquifer-System Analysis (Gulf Coast RASA) study area encompasses approximately 230,000 mi^2 onshore in parts of Alabama, Arkansas, Florida, Illinois, Kentucky, Mississippi, Missouri, Tennessee, Texas, and all of Louisiana. The aquifer system (and the study area) extends offshore beneath the Gulf of Mexico to include an additional 60,000 mi^2 and truncates at the edge of the Continental Shelf.

The Gulf Coast Regional Aquifer-System Analysis study is limited to coastal plain sediments of mostly Cenozoic age except for the northernmost part of the area, where it includes Upper Cretaceous rocks. The thickness of the aquifer system increases toward the Gulf of Mexico in a general wedge shape and has a thickness of more than 17,000 ft near the coastline of southeastern Louisiana. The shallower parts of the aquifers contain freshwater, but the deeper and offshore parts contain mostly highly mineralized water. The Gulf Coast RASA study area and its relation to adjacent RASA study areas is shown in figure 1. The plan of study for the Gulf Coast RASA was described by Grubb (1984, 1985, and 1987).

Ground water is important in the study area even though rainfall and surface water are relatively abundant. Nearly 10 billion gallons per day of ground water are withdrawn from the aquifers; mostly for irrigation. Municipal and industrial withdrawals are also very important uses of ground water in the region. The effect of ground-water withdrawals has already become regional in scope. Water-level declines, with resultant increased pumping costs, have spread across local and state political boundaries beyond the immediate vicinity of the irrigated areas, cities, and industrial areas where the water is pumped. Land subsidence and salt-water intrusion are other significant effects of ground-water use in the region.

Three aquifer systems have been delineated in the Gulf Coast RASA study area: 1) the Mississippi embayment aquifer system; 2) the Texas coastal uplands aquifer system; and 3) the coastal lowlands aquifer system (fig. 2) (Grubb, 1984). The delineation of these systems is based upon the geologic structure and sedimentation pattern, separation by regional confining units, and the presence of more than two regional aquifers in each system. The topography and outcrop pattern which control the regional ground-water flow in the Mississippi embayment have been substantially affected by the synclinal structure of the embayment. Two regional confining units, the Midway and the Vicksburg-Jackson, were referred to as the coastal uplands confining system and the coastal lowlands confining system by Grubb (1984). Although all three aquifer systems are mixtures of fine- and coarse-grained sediments, the fine-grained beds in the coastal lowlands aquifer system are generally thinner, dispersed vertically throughout the aquifers, and not areally as extensive.

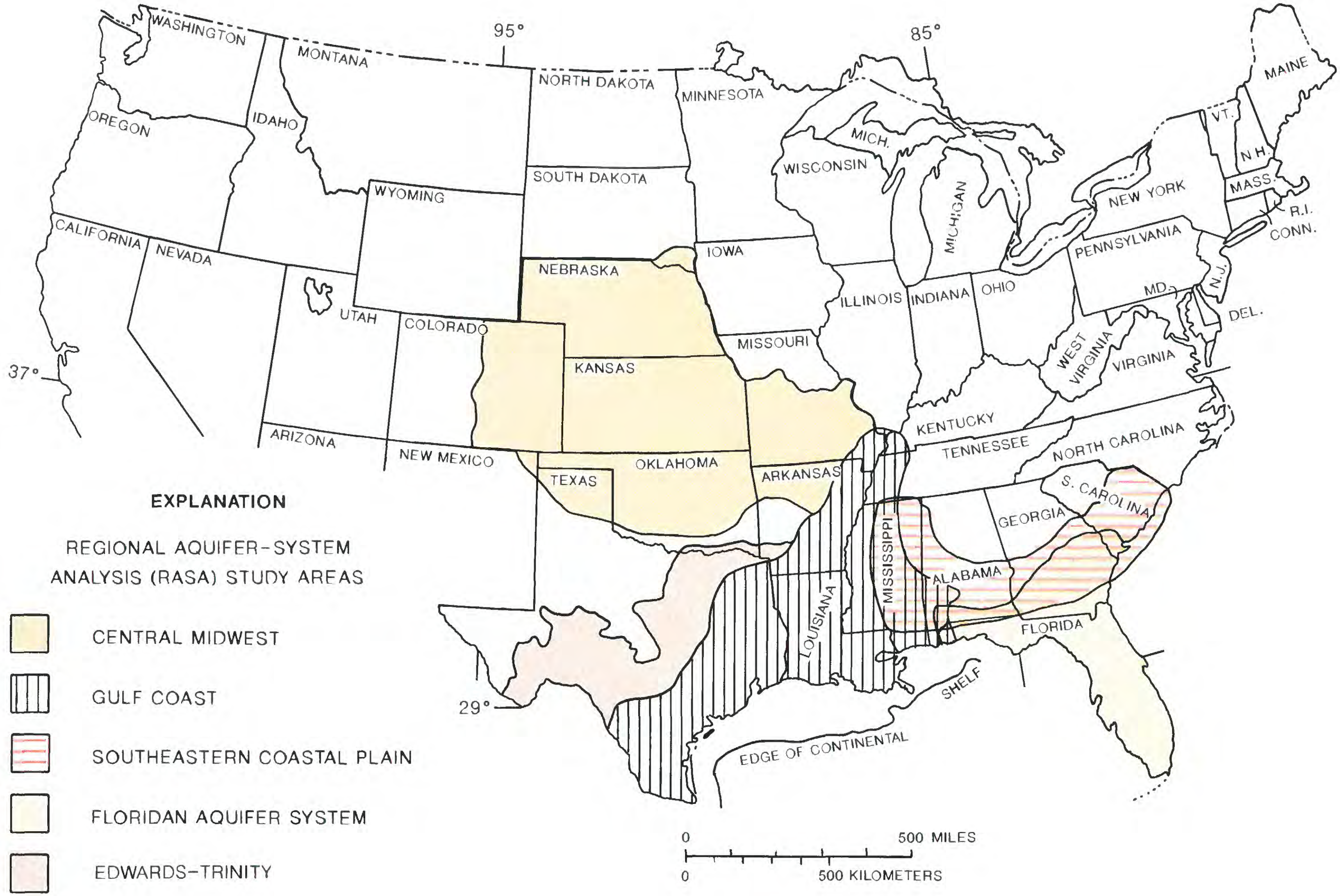
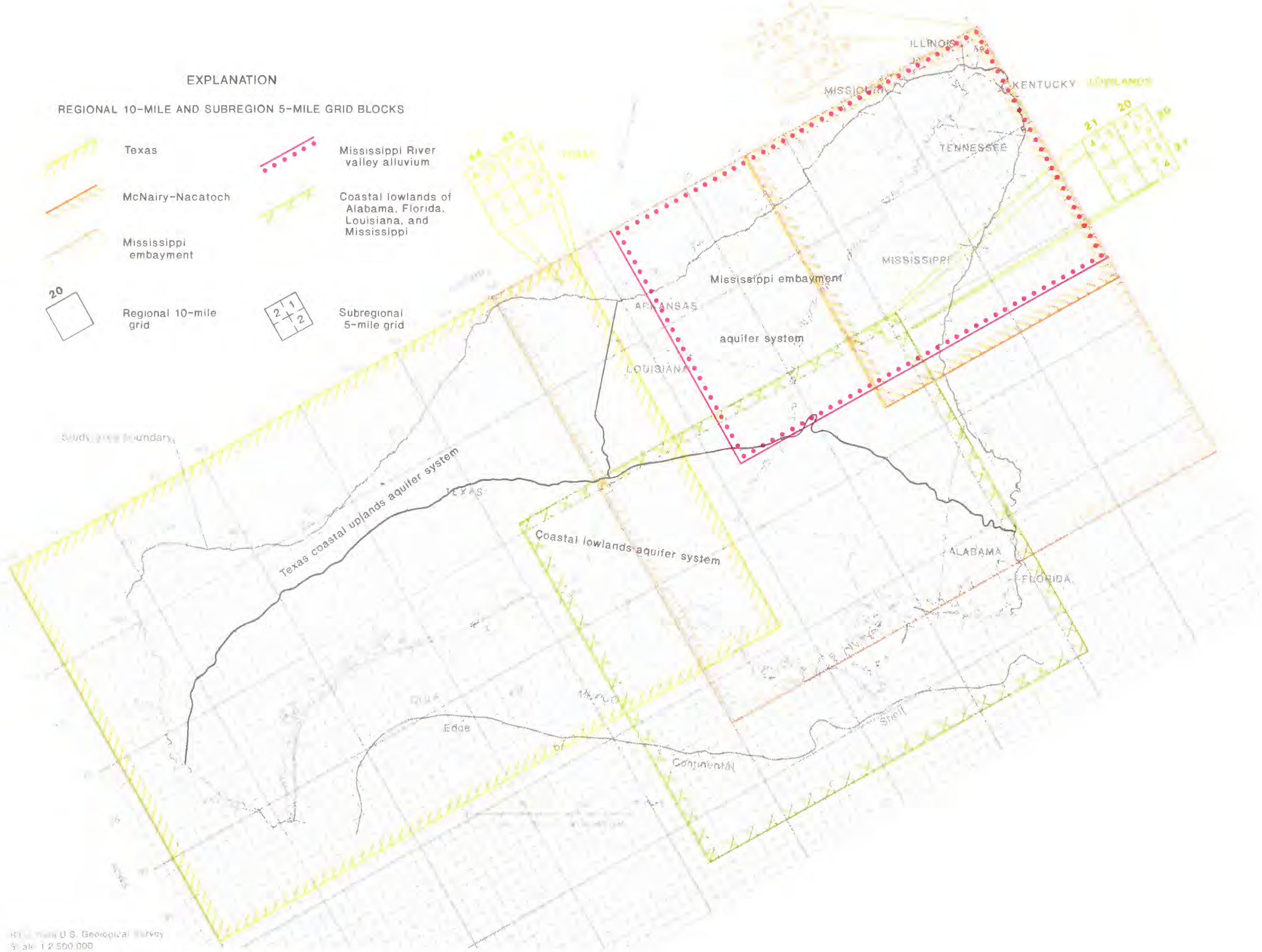


Figure 1.--Location of Gulf Coast Regional Aquifer-System Analysis study area and adjacent Regional Aquifer-System Analysis study areas.



Base from U.S. Geological Survey
Scale 1:2,500,000

Figure 2.--Regional and sub-regional model areas.

Purpose and Scope

This report describes the methods and data sources used to make a model of the ground-water flow system, and preliminary results from several simulations. It also describes the concept of the flow system and the transformation of that concept into discrete values. It also includes the methods used, data sources, and values of the initial estimates of characteristics of the aquifer systems. The entire gulf coast aquifer system except for the partially isolated Cretaceous sediments will be simulated. Konikow (1978) reported that the most significant calibration changes to a model are made during the first few runs, therefore the preliminary results are usually indicative of results obtained after more detailed calibrations. Therefore, for simplicity, simulations have been primarily limited to steady-state (that is, no change in flow rates with time). The predevelopment flow system (which is assumed to be steady-state) is described and some of the major changes that have occurred due to water development are discussed. The model will be further refined as data becomes available.

Approach

The reasons for using a computer model to simulate regional ground-water flow are described in this section. In addition, the relation of the Gulf Coast RASA to adjacent RASA studies and to more detailed subregional analysis of ground-water flow that were undertaken as part of the Gulf Coast RASA are discussed.

Why Use a Model?

A digital model is the best available tool to: 1) Integrate most of the known information about the ground-water flow in an aquifer system, 2) test hypotheses about the interaction of processes in the aquifer system, and 3) compare characteristics of the system from one area or aquifer to another. A digital aquifer model is the only feasible way to estimate the various inflow and outflow rates for an aquifer system under predevelopment and development conditions. The goal of this study is to develop a better understanding of the flow system, and define the controls on it, not to just build a model and make predictions about future developments. To accomplish this, regional flow paths must be determined, and characteristics of the system compared from one area or aquifer to another. The human influences on flow need to be explained. The effects of future development on ground-water flow and related problems need to be studied.

Adjacent Regional Aquifer-System Analysis Studies

The relation of the Gulf Coast RASA study area to four adjacent aquifer systems that have been or are being studied under the RASA program is shown in figure 1.

Edwards-Trinity

The Edwards-Trinity RASA includes carbonate aquifers in Cretaceous sediments which underlie the gulf coast aquifers in Texas and adjacent states (Bush, 1986). The volumes of vertical flow to or from the overlying Tertiary aquifers and confining units are thought to be small and to have a small effect on flow in the lower part of the aquifer system due to the great thickness of fine-grained marine sediments of the Midway Group.

Central Midwest

The Central Midwest RASA includes the Paleozoic rocks west of the northern tip of the Gulf Coast RASA study area (Jorgensen and Signor, 1981). These Paleozoic rocks extend beneath the Upper Cretaceous sediments in the Mississippi embayment, but are not used for water supply because they are overlain by other aquifers. The Paleozoic rocks are generally less permeable than the Upper Cretaceous sediments but some water may flow upward from the Paleozoic rocks into the Upper Cretaceous sediments.

Southeastern Coastal Plain

The Southeastern Coastal Plain RASA includes Cretaceous and Tertiary sediments east of the Gulf Coast RASA study area (Renken, 1984). The study areas overlap in western Alabama and eastern Mississippi; however, the Southeastern Coastal Plain RASA study is restricted to aquifers in Cretaceous sediments which dip beneath the Midway Group (Paleocene) clays at the base of the gulf coast aquifer systems. The Midway Group becomes more permeable to the east due to facies changes and yields water in Georgia and east-central Alabama, but westward towards the gulf coast aquifer systems it is predominantly composed of thick marine clays.

Floridan

The Floridan RASA includes the highly permeable carbonate aquifers in sediments of Tertiary age in Florida, Alabama, Georgia, and South Carolina (Johnston and Bush, 1988). The boundary between the Floridan and the Gulf Coast RASA study area occurs as a facies change from the more permeable carbonate rocks to the less permeable marine clays of the Jackson and Vicksburg Groups which together constitute a regional confining unit in the Gulf Coast RASA study area.

Regional and Subregional Models

The regional model includes all of the aquifers over the entire study area, except for the underlying McNairy-Nacotoch aquifer in the Cretaceous McNairy and Nacotoch Sands in the northernmost part of the area. Subregional models are being used to investigate parts of the aquifer systems (fig. 2) with more detail. The subregional models, except for Texas, simulate only the freshwater part of the aquifer system under study, whereas the regional model simulates flow in the entire thickness of Tertiary and younger sediments down to the top of the geopressured zone. The subregional models are described separately:

McNairy-Nacotoch aquifer (Brahana and Mesko, 1988)
Mississippi River Valley alluvial aquifer (Ackerman, 1989)
Mississippi embayment aquifer system (Arthur and Taylor, 1989)
Coastal lowlands aquifer system of Alabama, Florida, Louisiana,
 and Mississippi (Martin and Whiteman, 1989)
Gulf coast aquifers in Texas (Ryder, 1988)

Previous Investigations

Studies of ground water in the Gulf Coastal Plain of the United States were published early in this century (Hill, 1901, and Fuller, 1904). Hill recognized three artesian-well systems in that part of Texas covered by the current study and noted that the Coastal Plain of Texas, "constitutes one of the most productive artesian regions in America, if not in the world * * * (Hill, 1901, p. 398). A lengthy discussion of the three artesian-well systems was presented along with a detailed lithologic description of samples from a well located at Galveston, Texas, that was 3,070 ft deep (Hill, 1901, p. 402-405). The report by Fuller (1904) is a compilation of a short description of the rocks in all of the States within the Gulf Coast RASA study area except Illinois, Louisiana, and Texas. The discussion of Coastal Plain sediments was limited to one or two sentences or paragraphs depending upon the proportion of the State that is underlain only by older rocks; the bulk of the report consists of well tables with notes on individual wells. A longer report was released the same year (Harris, 1904) describing the rocks and ground water of southern Louisiana, perhaps the first full scale interpretative report of ground-water hydrology in the Gulf Coast RASA study area.

A compilation by M.L. Fuller was published in 1905 which contained the most extensive discussion of ground-water hydrology in Coastal Plain sediments for all of the States within the study area except Texas and Louisiana. In Fuller's report results are presented by different individuals for the following States: Fuller, Florida; Glenn, Tennessee and Kentucky; Johnson, Mississippi; Purdue, Arkansas; Shepard, Missouri; Smith, Alabama; and Veatch, Louisiana and southern Arkansas. By the middle of the second decade of the century detailed reports of the ground water in Coastal Plain sediments for every State in the study area had been published (Harris, 1904, Crider and Johnson, 1906, Glenn, 1906, Veatch, 1906, Shepard, 1907, Smith, 1907, Taylor, 1907, Matson and Sanford, 1913, Stephenson and Crider, 1916). Extensive tables of data on wells, surficial geologic maps, and detailed discussion, county-by-county, of the occurrence of ground water is typical of these reports.

A report released in 1916 as a separate volume and as part of a volume entitled "Contributions to the hydrology of the United States 1915" was devoted to the ground-water resources of LaSalle and McMullen Counties, Texas (Deussen and Dole). In this report, the first devoted to an intensive study of the ground-water resources in the Gulf Coast RASA study area, the authors pointed out that, "Close spacing of wells results in interference among them * * *. Thus as development of ground-water supplies increased over the next two decades, so did the intensive studies of counties and local areas with large ground-water withdrawals or special problems. For example, several reports dealing with the water resources of individual Texas counties were released in the mid to late 1930's (Lonsdale, 1935, Livingston and Bridges, 1936, Sayre, 1937, and Lonsdale and Day, 1937). A report on the water supply of Memphis, Tennessee (Wells, 1932), was the first in the Gulf Coast RASA study area dealing exclusively with the ground-water resources in a small area. Reports continued to include data tables and descriptive materials, but they also began to become more quantitative and the authors attempted to relate cause and effect. Wells (1932, p. 31-32) presented a profile of the pressure surface and the location of pumping in the lower Wilcox aquifer at Memphis for 1898, 1902, 1914, and 1928. White, Rose, and Guyton (1944) presented tables of drawdown in the Houston District, Texas, and used the results of aquifer tests to develop curves showing the expected drawdown with distance from a pumped well for three different time periods as predicted by the Theis method.

By the mid 1970's reports documenting intensive studies of the ground-water resources had been published for most of the counties in the study area. Also many studies of special ground-water problem areas were documented, for example, salt water encroachment in the Houston-Galveston area (Winslow, and others, 1957), effects of faults on the ground-water flow system (Whiteman, 1979, and Gabrysch, 1984), land-surface subsidence due to pumping ground water (Winslow and Doyel, 1954, and Winslow and Wood, 1959, Gabrysch, 1969, 1977, and 1982, Gabrysch and Bonnett, 1975, and Whiteman, 1980).

A few studies of a regional scale were documented in some early reports such as those of the Texas Coastal Plain by Deussen (1914, 1924). Much later a study of the ground-water resources of the Mississippi embayment was documented in a series of reports (Cushing and others, 1970, Boswell and others, 1965, 1968, Hosman and others, 1968). A multistate study of the Claiborne Group from eastern Mississippi to the Rio Grande in Texas was conducted by Payne (1968, 1970, 1972, 1975). A series of ground-water appraisals for the major river basins of the study area were published during the 1970's (West and Broadhurst, 1975, Baker and Wall, 1976, Bedinger and Sniegocki, 1976, Zurawski, 1978, and Cederstrom and others, 1979, Terry and others, 1979).

The interest in quantitative analysis of ground-water flow, primarily related to areas of intensive ground-water pumpage led to the early application of simulation techniques to the analysis of ground-water flow in the study area. Wood and Gabrysch (1965) used an analog model to simulate the flow of ground water in the Houston District, Texas. During the 1970's several applications of simulation techniques had been applied to a variety of problems across the study area. Reed (1972) applied an analog model to simulate the decline in water levels in the areally extensive Sparta Sand in the Mississippi embayment. The application of an analog model to simulate stream-aquifer interactions was given by Broom and Reed (1973) for the Bayou Bartholomew alluvium in Arkansas. Jorgensen (1973, 1975), and Jorgensen and Gabrysch (1974) reported on several applications of analog models to ground-water flow problems in the Houston area, Texas. Jorgensen (1975), in a follow-up study to the earlier work of Wood and Gabrysch (1965) of the Houston area, Texas, used an analog model to simulate ground-water flow in a heavily pumped, multi-layer aquifer system with vertical movement of water between layers. The area studied by Jorgensen (1975) was about 9,100 mi^2 which is almost twice as large as the area studied by Wood and Gabrysch (1965).

By the mid 1970's, studies began to use digital models to simulate ground-water flow. Garza (1974) reports on the use of a digital model to simulate ground-water flow and water levels in an alluvial aquifer and adjacent Carrizo-Wilcox aquifers near a reservoir on the Trinity River in eastern Texas. Klemt and others (1976) report on a study of the Carrizo aquifer in south Texas in which they used a digital model to simulate the effects of extensive ground-water pumpage for irrigation on water-levels in an area of about 12,000 mi^2.

By the end of the decade of the 1970's, digital models were being applied to most types of ground-water flow problems in the study area. Reed and Broom (1979) simulated stream-aquifer interactions of the Bayou Bartholomew alluvium in Arkansas. Meyer and Carr (1979) used a digital model to simulate ground-water flow in the heavily pumped Houston area, Texas. The digital model used in the Houston area simulated flow in two pumped zones, the release of water from the inelastic compaction of clays, and land-surface subsidence. The 27,000 mi^2 area studied was about three times larger than the area of the analog model study reported by Jorgensen in 1975. Jorgensen (1981) gave a more detailed description of the studies involving ground-water flow simulation in Houston.

The number of areas and the type of problems to which simulation techniques were applied continued to increase in the decade of the 1980's, and digital models replaced analog models in large-scale applications. The use of simulation techniques within the study area are discussed below under five categories: shallow alluvial aquifers; multi-layer aquifers with pumpage from more than one layer and vertical flow between layers; effects of pumpage on water-levels and salinity in a single layer; effects of development of other resources on the ground-water system; and the use of optimization models for managing ground-water withdrawals.

Parts of the shallow Mississippi River Valley alluvial aquifer were the subject of two separate studies. Broom and Lyford (1981) simulated the effects of large scale pumping on water-levels and stream flow in the Cache and St. Francis River basins in northeastern Arkansas. Sumner and Wasson (1984a, 1984b) simulated the effects of large ground-water withdrawals on water levels in northwestern Mississippi. The effects of underlying aquifers were not simulated in either of these studies, although regionally extensive aquifers of Tertiary age are present throughout the areas underlain by the Mississippi River Valley alluvial aquifer.

The principal areas with multilayer aquifers that have pumpage from more than one layer and that have been the subject of ground-water flow simulation are the Houston area, Texas and the Baton Rouge area, Louisiana. Carr and others (1985) report on the use of a multilayer model to simulate the effects of ground-water withdrawals on water-levels and land-surface subsidence for most of the Texas Gulf Coastal Plain that is underlain by the Chicot and Evangeline aquifers. The effects of ground-water withdrawals on water levels in three intensively pumped zones in the Baton Rouge area, Louisiana, were reported by Torak and Whiteman (1982) and Huntzinger and others (1985). The effects of saline water contained in sediments located gulfward of these two areas were not evaluated in the flow models. Brahana (1982) reported on the use of simulation techniques in the Memphis area, Tennessee, where pumpage occurs from more than one layer although his study was limited to the most heavily pumped aquifer, the Memphis Sand. Baker (1986) reported on the use of simulation techniques in a study of the Jasper aquifer in east Texas.

The presence of slightly saline to saline water in sediments adjacent to or gulfward of heavily pumped areas causes concern regarding the movement of saline water into the pumped areas and subsequent deterioration of the quality of water withdrawn. Two studies have been conducted within the study area in which simulation techniques were used to address the problem of water-level declines and movement of saline water due to ground-water pumping. Groschen (1985) reported that the movement of saline water from the overlying Chicot aquifer downward into the pumped Evangeline aquifer has not created a regional-scale degradation in water quality over an area of about 5,000 mi^2 southwest of Corpus Christi, Texas. Trudeau and Buono (1985) report on the effects of proposed increased pumpage on water-levels and salinity in the Sparta aquifer near West Monroe, Louisiana.

The evaluation of effects of development of other resources on the ground-water flow system using simulation techniques is illustrated by three studies in Texas. Fogg and others (1983) report on the results of a multi-layer model study of the Wilcox Group in the vicinity of the Oakwood salt dome, located in east Texas. The study was the result of a proposal to use the salt dome for the isolation of radioactive waste. Henry and others (1982) report the results of simulating ground-water flow along a fault zone in the vicinity of uranium solution mining in sediments of Miocene age, located in south Texas. Charbeneau and Wright (1983) reported the results of simulating effects of open-pit lignite mining and in place gasification of lignite on ground-water flow in Eocene sediments in east-central Texas.

Peralta and co-workers have proposed the use of optimization models to regulate pumpage such that a minimum saturated thickness is maintained (Peralta and Peralta, 1984, Peralta and Killian, 1985, Peralta and others, 1986, and Yazdanian and Peralta, 1986). Many of their suggested applications for these optimization models have used the Grand Prairie area of Arkansas (for example Yazdanian and Peralta, 1986), where the Mississippi River Valley alluvial aquifer is intensively pumped and has been used as a source of water for the irrigation of rice for most of this century.

The primary purpose of most of the studies cited above has been to address a specific problem, and only secondarily to develop an understanding of the aquifer under study and how it interacted with the larger regional ground-water flow system. The design and use of the model described herein, as noted earlier, is directed toward the goal of increased understanding of the regional ground-water flow system.

Setting

The study area is a gently sloping coastal plain in a mostly humid subtropical climate, underlain by thick, mostly clastic sediments deposited in a gulfward offlapping sequence. Much agricultural and industrial development has occurred in the area, requiring substantial ground- and surface-water use.

Topography

The topography has a major effect on ground-water flow, because the water-table is a subdued replica of the configuration of the land-surface altitude (Freeze and Witherspoon, 1967). The land-surface altitude in the study area varies from sea level to over 800 ft (fig. 3). The dominant feature of the topography is the flat, low-lying Mississippi River Alluvial Plain (Fenneman, 1938). The topography of the Mississippi embayment is asymmetrical in that the alluvial plain lies to the west side of the embayment and the topographically higher hills are mostly to the east side. The river itself, which is an important feature of the hydrologic system, generally traverses the east side of the alluvial plain. Toward the coast, the general slope down toward the Gulf is intersected by large stream valleys which are generally perpendicular to the coastline.

Climate

The warm, relatively humid climate of the area is very favorable for agriculture. Mean annual rainfall in the area varies from less than 24 in. near the Mexico border to more than 60 in. along the gulf coast of Louisiana, Mississippi, and Alabama (fig. 4), and averages about 48 in/yr. The rainfall is fairly evenly distributed throughout the year, though the spring is usually wetter and the summer and fall are usually drier, except in the western part of the area where most of the rain falls in the spring and summer (fig. 6). The mean annual temperature, ranging from less than 16 degrees Celsius in the north to more than 21 degrees Celsius in the south, and the potential evapotranspiration, are more consistent across the area.

In part of the area, part of the time, the actual evapotranspiration is limited to the amount of rainfall, which is less than the potential evapotranspiration. Pan evaporation, which is generally greater, but closely related to potential evapotranspiration, is shown in figure 5. In the western part of the area, in Texas, potential evapotranspiration generally exceeds the rainfall, especially in the summer and fall, so most of the rainfall returns to the atmosphere in a short time (fig. 6). In wetter times or areas when rainfall exceeds potential evapotranspiration, the remainder either infiltrates directly into the ground or runs off into streams and is available to recharge the aquifers. In the eastern part of the area, rainfall substantially exceeds potential evapotranspiration, providing abundant surface-water runoff. The part of rainfall that recharges the aquifers is usually only a small proportion of the precipitation in excess of actual evapotranspiration, so it is difficult to estimate from a water budget approach.

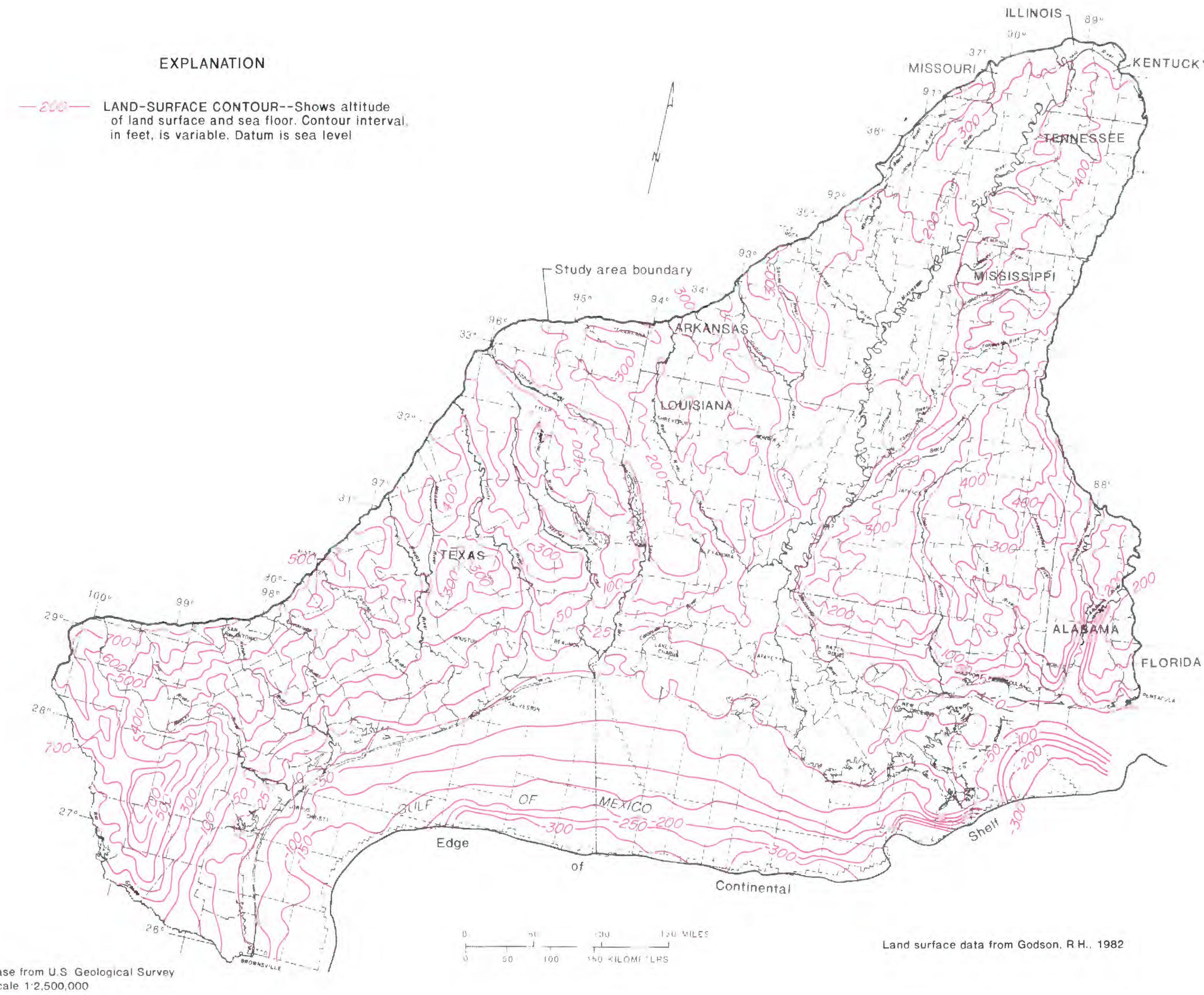


Figure 3.--Generalized average land surface altitude and bathymetry.

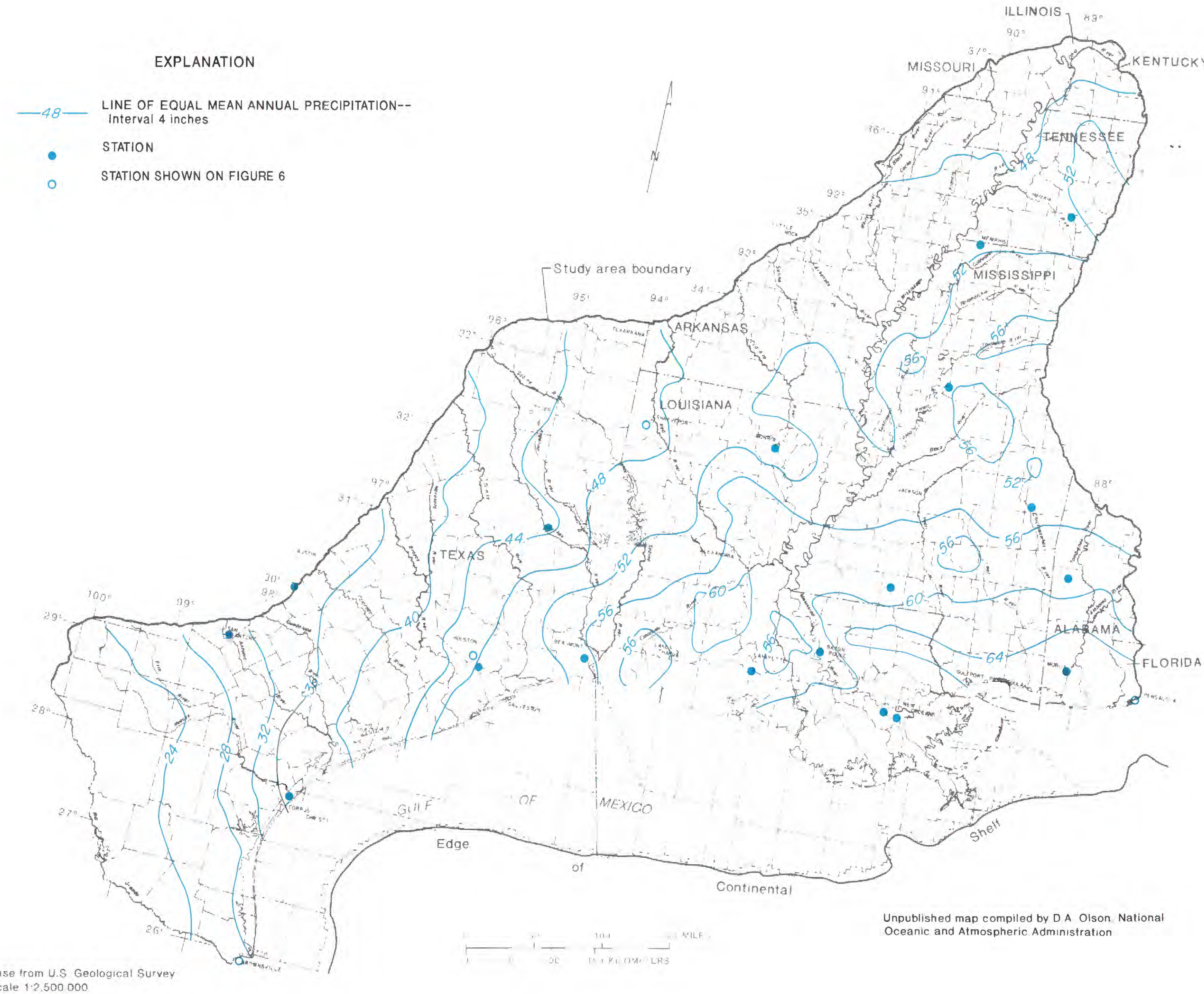


Figure 4.--Mean annual precipitation, 1951-80.

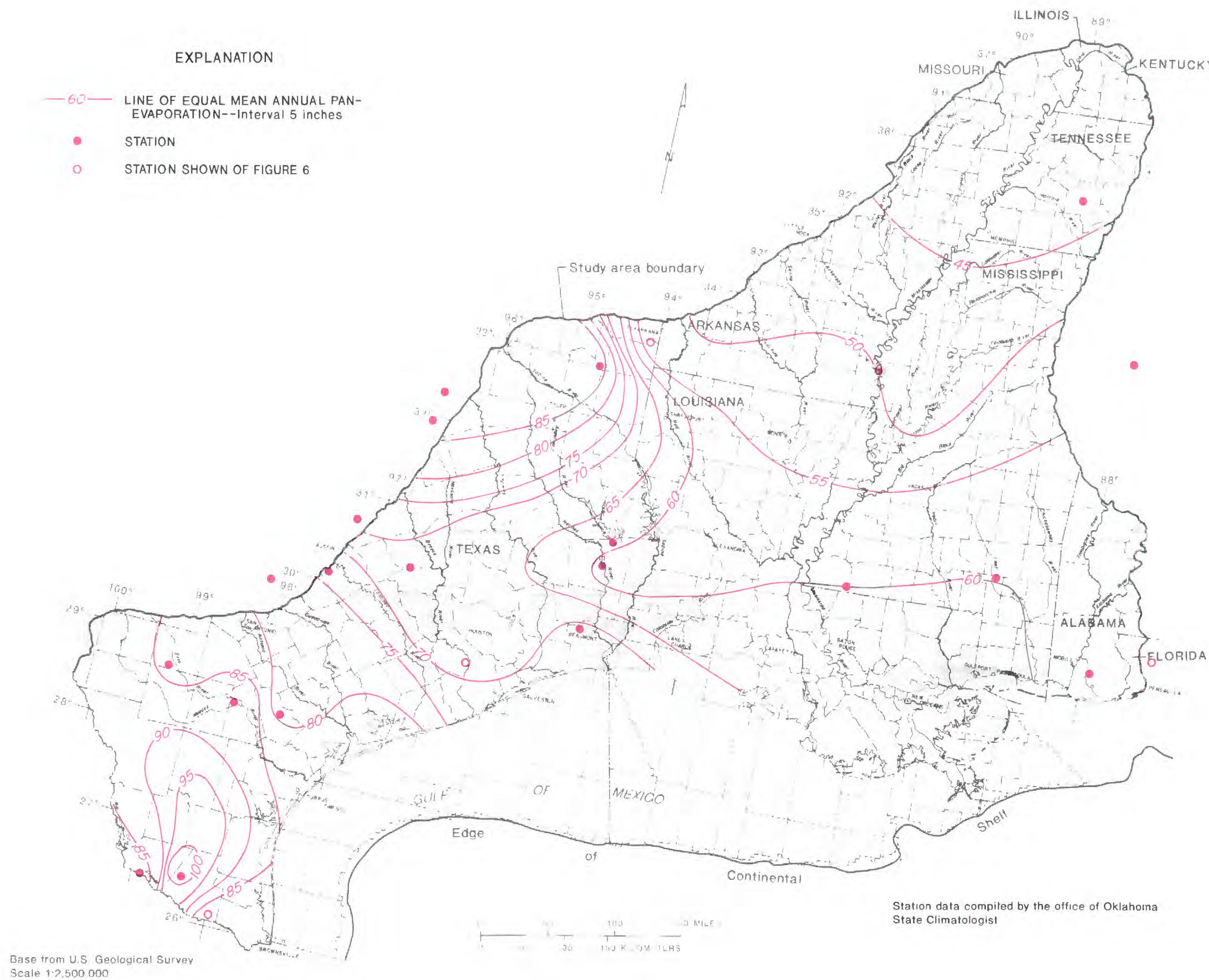


Figure 5.--Mean annual pan evaporation, 1951-80.

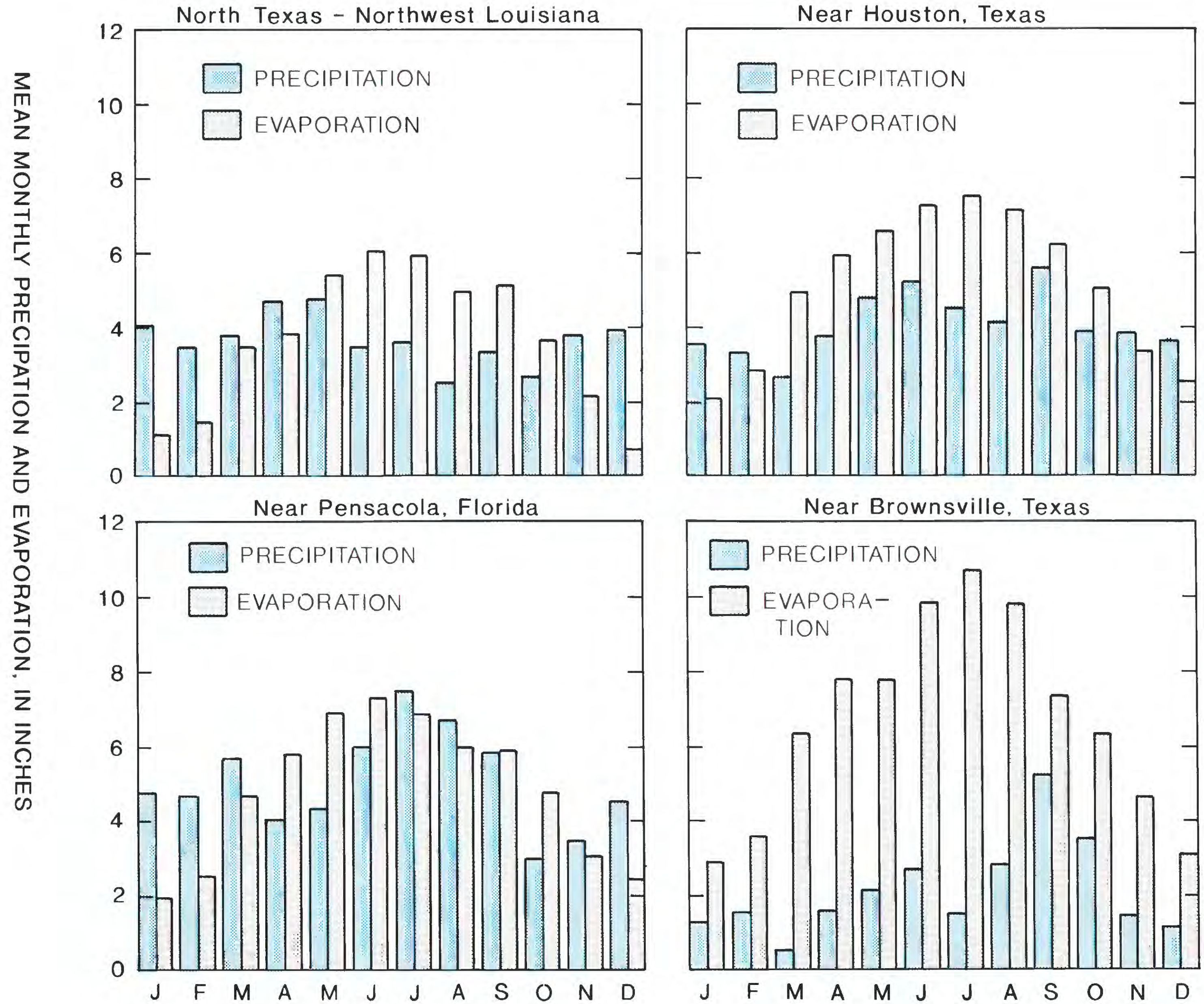


Figure 6.--Mean monthly precipitation and pan evaporation at four sites, 1951-80.

Surface Water

Many major rivers flow through the study area and some originate in the area. The mean annual unit runoff (fig. 7) varies from less than 1 in/yr in the southwestern part of the area to more than 20 in/yr in the northern and eastern parts and averages about 15 in/yr. Table 1 lists the major rivers with their mean annual flow and the storage capacity of major reservoirs within the study area or near its boundary. Streamflow rates are generally so large (table 1) that the portion moving into or out of the aquifers is not detectable because it is generally less than 1 percent of the streamflow. The unregulated mean annual flow of the major rivers that flow through the study area is 467 million acre-ft/yr (416,000 Mgal/d) (table 1). The combined conservation pool storage in major reservoirs in or near the study area is larger than 30 million acre-ft. However, the surface-water storage is very small relative to the amount, over 2,000 million acre-ft, of freshwater in storage in the top 200 ft of the aquifer system (assuming an aquifer outcrop area of 180,000 mi^2, a depth-to-water of 25 ft, and a specific yield of 0.10). Nevertheless, surface water provides about one-half of the water used in the study area (excluding nonconsumptive uses associated with hydroelectric power generation).

Table 1.--*Mean annual flow of major rivers and storage of major reservoirs in or near study area*

River at gaging station	Contributing drainage area (square miles)	Unregulated mean annual flow (millions of acre-feet per year)	Dam and lake or reservoir name	Year built	Conservation capacity (millions of acre-feet)
Rio Grande near Falcon, Texas	229,900	0.19	International Amistad	1969	3.505
			International Falcon	1954	2.767
Nueces near Three Rivers, Texas	9,937	0.599	Choke Canyon	1982	0.697
Guadalupe near Victoria, Texas	3,766	1.178	Canyon	1964	0.382
Colorado near Bay City, Texas	30,840	1.682	Mansfield Dam/ Lake Travis	1940	1.172
Brazos near Rosharon, Texas	35,770	5.587	Robertson Dam/Lake Navasota	1978	0.225
			Lake Somerville	1962	0.160
San Jacinto near Sheldon, Texas	2,879	0.869	Lake Houston	1954	0.147
Trinity near Rosmayer, Texas	17,190	5.184	Haggset Dam (Cedar Creek Reservoir)	1966	0.679
			Lake Livingston	1969	1.788
Neches near Evadale, Texas	7,951	4.507	Sam Rayburn	1965	2.853
Sabine near Ruliff, Texas	9,329	6.102	Iron Bridge Dam/ Lake Tawakoni	1960	0.936
			Toledo Bend	1964	4.477
Red near Alexandria, Louisiana	61,564	22.42	Farrell's Bridge Dam/ Lake O' the Pines	1959	0.255
			Lake Texarkana (Wright Patman Lake)	1957	0.145

Table 1.--*Mean annual flow of major rivers and storage of major reservoirs in or near study area--Continued*

River at gaging station	Contributing drainage area (square miles)	Unregulated mean annual flow (millions of acre-feet per year)	Dam and lake or reservoir name	Year built	Conservation capacity (millions of acre-feet)
Ouachita at Camden, Arkansas	5,357	5.47	Lake Ouachita	1952	2.768
			Lake DeGray	1969	0.882
Arkansas at Little Rock, Arkansas	158,030	29.10	----	----	----
White near Clarendon, Arkansas	25,560	21.61	Bull Shoals	1951	5.408
St. Francis near Riverfront, Arkansas	6,475	3.86	Wappapelo	1941	0.625
Mississippi near Thebes, Illinois	713,200	143.00	----	----	----
Ohio near Metropolis, Illinois	203,000	194.75	----	----	----
Obion at Obion, Tennessee	1,852	1.963	----	----	----
Hatchie near Bolivar, Tennessee	1,480	1.745	----	----	----
Coldwater near Arkabutla, Mississippi	1,000	0.965	Arkabutla Lake	1941	0.0315
Tallahatchie near Swan Lake, Mississippi	5,130	5.447	Sardis Lake	1939	0.108
Big Black near Bovina, Mississippi	2,810	2.578	----	----	----
Pearl near Jackson, Mississippi	3,100	2.850	Ross Barnett	1962	0.31
Chickasaway near Leaksville, Mississippi	2,680	2.739	----	----	----
Tombigbee near Coffeeville, Alabama	18,417	2.228	----	----	----
Totals 1/	1,557,000	466.6			30.320

1/ totals may not agree due to rounding

Geology

The sediments of the gulf coast aquifer systems were deposited mostly during Cenozoic time, but also include Upper Cretaceous sediments which are being evaluated for their degree of hydraulic interaction with the Cenozoic units. Deposition occurred in fluvial, deltaic, or shallow-marine environments, resulting in an interbedded sequence of sand, silt, and clay, with some gravel, lignite, and limestone. Changes in land-surface elevation (caused by basin subsidence) and changes in sea level caused

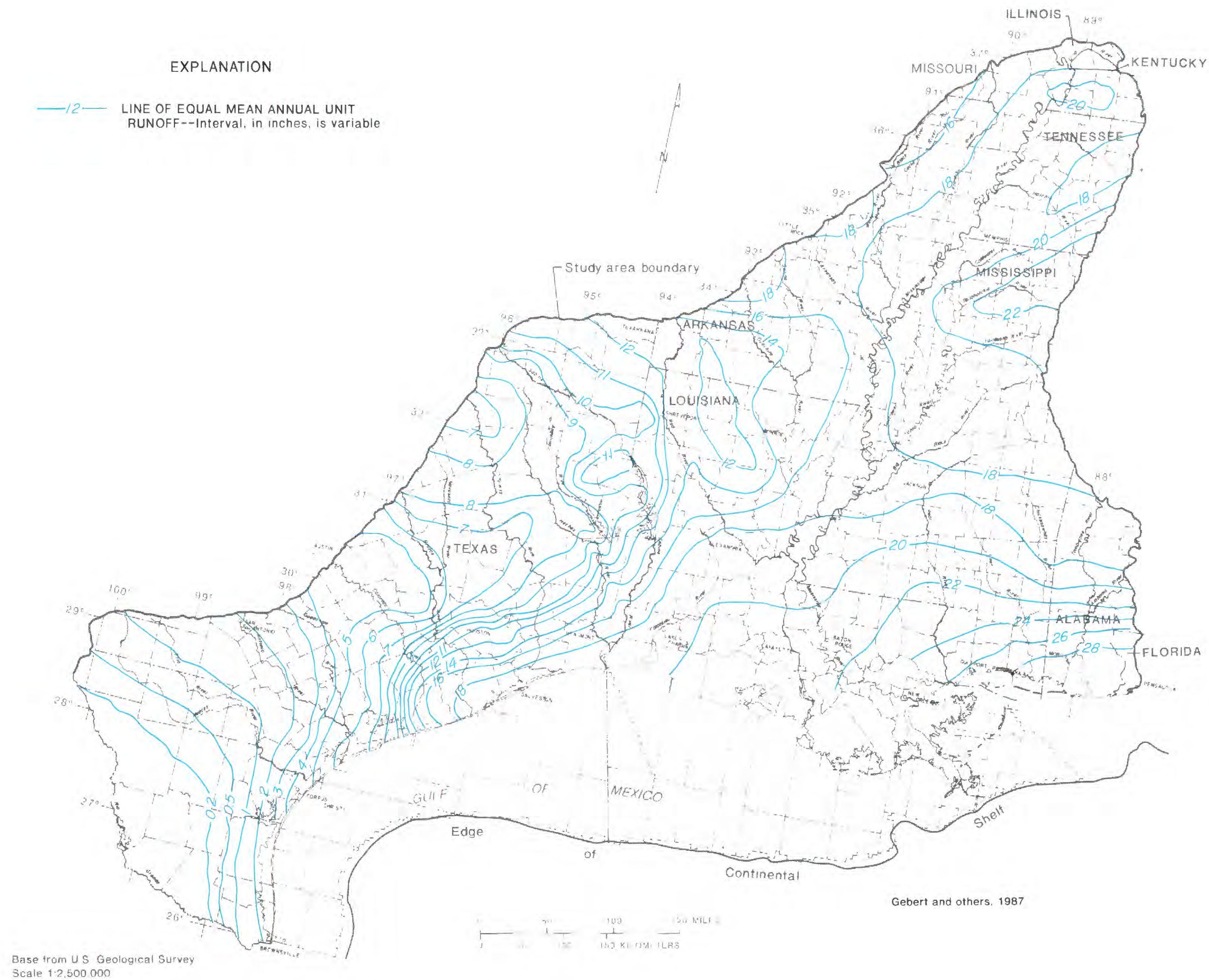


Figure 7.--Mean annual unit runoff, 1951-80.

transgressions and regressions of the sea. This caused cyclical sedimentation, alternating from predominantly continental to predominantly marine environments of deposition. Sedimentary units are exposed in roughly parallel bands that are younger progressively gulfward in a typical offlap sequence. The shifting of facies, both laterally and vertically, resulted in a complex interbedding of sediment types. In general, the more clastic continental deposits have higher permeabilities and comprise the aquifers, and the marine deposits have lower permeabilities and comprise the confining units.

The Mississippi, east Texas, and Rio Grande embayments are the major structural features of the study area (fig. 8), and largely control the pattern and thickness of sedimentation (fig. 9). The base of the gulf coast aquifer systems shown in figure 9 is controlled in the north by the top of the Midway confining unit and in the southern part of the area by the transition zone into geopressured sediments. These major structural features were present prior to Cenozoic deposition, and continued to develop as the basins subsided to accommodate the increased sediment buildup. Except where affected by local uplift, the general pattern of sedimentation is one of increasing thickness in a gulfward, downdip direction. Uplifted features which affected the deposition patterns are the Sabine uplift, San Marcos arch, Monroe uplift, Pascola arch, Jackson dome, LaSalle arch, Wiggins uplift and Hancock arch. Downwarp features associated with greater sediment buildup are the Desha basin, East Texas basin, Houston embayment, Rio Grande embayment, and Terrebonne embayment.

Faults are common throughout the area, although their effect on regional ground-water movement is poorly known. In general, fault throws are not great enough to entirely offset hydrologic units described in this report, although individual beds could be offset. Much of the faulting occurs in zones which contain grabens and horsts. Faults zones near the perimeter of the Mississippi embayment and the Gulf Coast geosyncline are the New Madrid fault zone, the Balcones-Mexia fault zone, and the Pickens-Gilbertown fault zone. The New Madrid fault zone is within the Mississippi embayment and may provide a pathway for ground-water flow from underlying Paleozoic rocks (Brahana 1987, and Brahana and Mesko, 1988). Numerous growth faults, which occur during deposition, exist farther gulfward. Whiteman (1979) describes the significance of the Baton Rouge fault zone as a hydraulic barrier, and it is possible that other growth faults act similarly.

Geopressured zones (fig. 9) have fluid pressures which are significantly greater than the normal hydrostatic pressure (Jones, 1969). The most probable cause for the development of these abnormally high fluid pressures is: the escape of fluids is restricted during sediment compaction, thus causing pressure buildup and undercompaction of sediments (Fertl, 1976, p. 16). The high pressures could not exist unless flow out of these zones was small, and thus the boundaries of these zones are considered flow barriers. However, these barriers are not completely impermeable, if sedimentation slows and given enough time, the abnormally high pressures will eventually reduce to near normal hydrostatic pressures.

Salt domes occur throughout the gulf coast aquifer systems, particularly in belts near and roughly parallel to the present-day coastline. The source of the salt domes is the deep-seated Louann Salt of Jurassic age, which has risen as diapirs that penetrate varying thicknesses of Cenozoic strata. The structural effects of the domes are relatively localized. However, the domes can have a significant effect on water quality due to dissolution of salt by ground water. The highly mineralized ground water may affect regional flow by creating areas in the subsurface with high fluid densities and high viscosities.

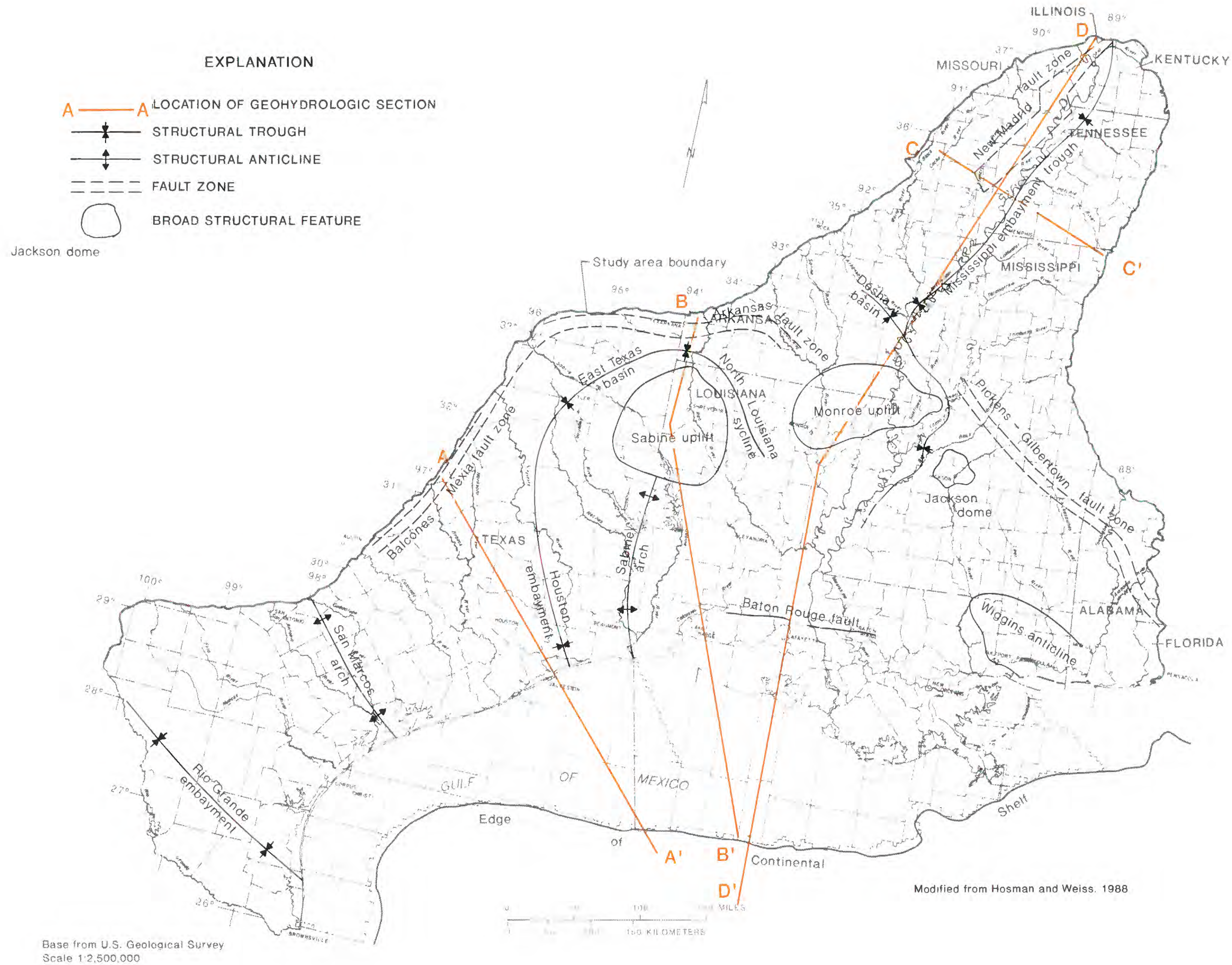


Figure 8.--Generalized structural features and location of geohydrologic sections.

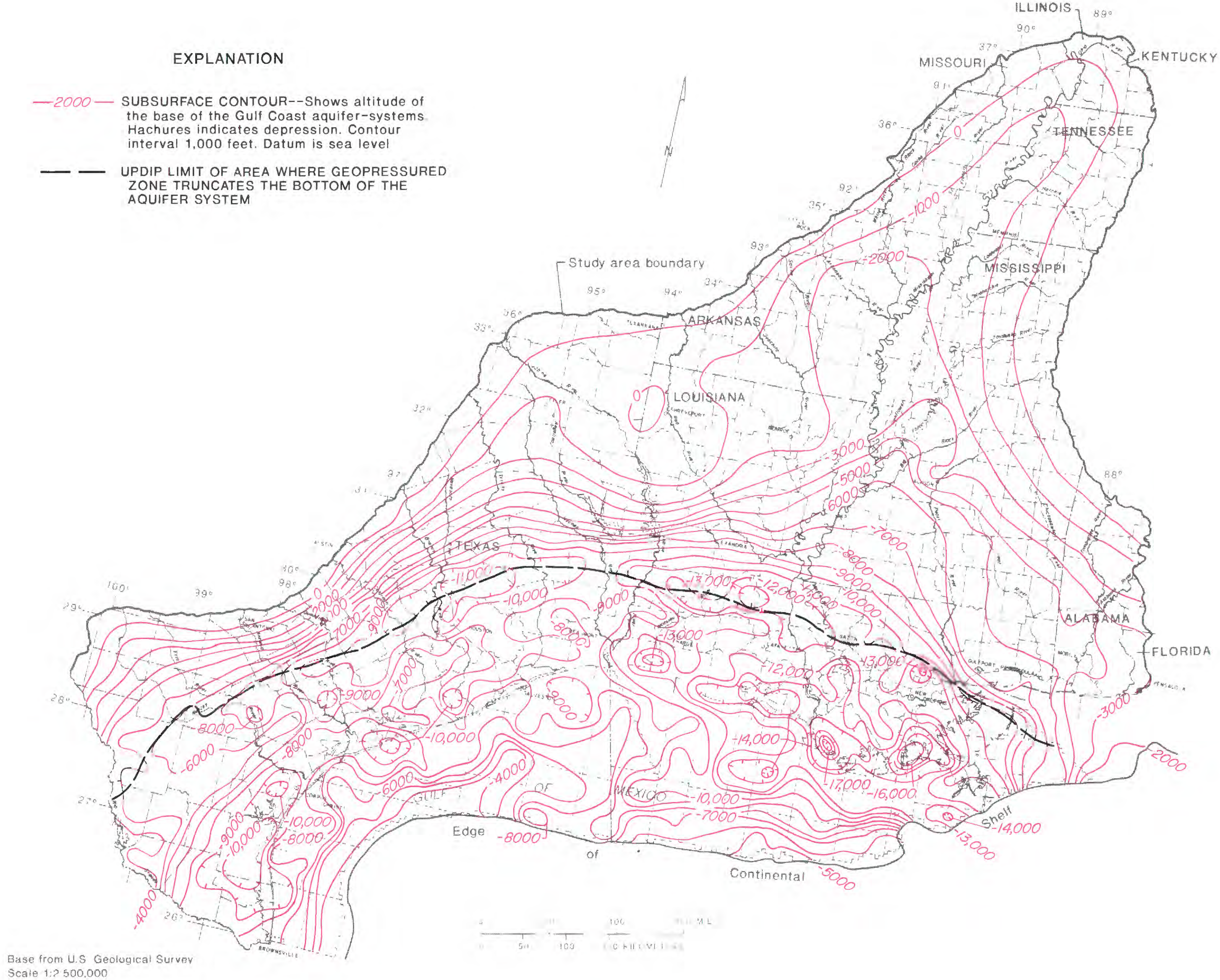


Figure 9.--Altitude of the base of the Gulf Coast aquifer systems and area of occurrence geopressure.

Geohydrologic Units

The gulf coast aquifer systems were subdivided into a series of geohydrologic units for analysis in the Gulf Coast RASA study. Weiss and Williamson (1985) describe the methods used for subdividing the thick sequence of sediments. The methods utilize a combination of lithologic and hydraulic information. Hosman and Weiss (1988) provide a detailed description of the units of the Mississippi embayment and Texas coastal uplands aquifer systems, including the McNairy-Nacotoch aquifer. J.S. Weiss (U.S. Geological Survey, written commun., 1986) provides a detailed description of the units of the coastal lowlands aquifer system. The relations of the geologic units, previously described geohydrologic units, and layers used in the regional flow model of the three aquifer systems are shown in table 2. The relation of the various geohydrologic units (equivalent to model layers) are shown in cross-sections (fig. 10 and plates 2-3). The sections were made from values of layer thicknesses for blocks spaced 5 mi apart which will be described in following sections. The areas where each of the units crop out at land surface and where they subcrop under the Mississippi River Valley alluvial aquifer are shown in figure 11.

The permeable zones are numbered sequentially up from layer 2 stratigraphically lowest to layer 11 at the top (table 2). Confining units are numbered sequentially up from layer 12 to layer 17. Layer 1 is reserved for potential future use to simulate flow in the McNairy-Nacotoch aquifer in the northern Mississippi embayment aquifer system. Layer 11 is used to represent both the Mississippi River Valley alluvial aquifer of the Mississippi embayment aquifer system and permeable zone A (Holocene-upper Pleistocene deposits) of the coastal lowlands aquifer system. The two parts of layer 11 are connected horizontally across a narrow band in central Louisiana.

The names assigned to the geohydrologic units of the gulf coast aquifer systems (table 2) use rock-stratigraphic terms for aquifers and confining units for the Mississippi embayment and Texas coastal uplands aquifer systems, and zones lettered A through E from youngest to oldest in the coastal lowlands aquifer system. The aquifer and confining unit names were assigned based on the stratigraphic name applied to most of the sediments within the hydrologic unit. The advantage of using these names is that they can be applied across the entire study area with a minimum of confusion with currently named rock units, and they imply a vertical ordering of the units.

Three criteria were used to subdivide the aquifer systems into geohydrologic units:

1. Borehole geophysical logs were used to map regionally significant confining units (layers 12-17), leaving aquifers in between (layers 5, 6 and parts of 7 and 8).

2. Logs were also used to identify large hydraulic conductivity contrasts between adjacent water-bearing zones not separated by regional confining units (layers 2 and 4 that are massive sands and layer 3 that is composed of thin, complexly interbedded coarse- and fine-grained deposits).

3. In the coastal lowlands aquifer system (layers 7-11) where criteria 1 and 2 did not apply, hydraulic head data at different depths and the patterns of pumping depths were used to subdivide the section into geohydrologic units so that the minimum vertical head change occurred within a unit.

Table 2.--*Relation of geologic units, previously defined geohydrologic units, and layers used in regional flow model.*

[Note: correlations shown here are generalized. Exact relations vary widely from place to place.]

Mississippi embayment and Texas coastal uplands aquifer systems

Geologic Unit			Geohydrologic units defined by previous studies	Gulf Coast Regional Aquifer-System Analysis	
System	Series	Group		Model layer number	Geohydrologic units
Quaternary	Pleistocene and Holocene		Mississippi River Valley alluvial aquifer (Boswell and others, 1968)	11	Mississippi River Valley alluvial aquifer *
Tertiary	Eocene and Oligocene	Jackson and Vicksburg	Jackson and Vicksburg Groups	15	Vicksburg-Jackson confining unit 1/
Tertiary	Eocene	Claiborne	Cockfield aquifer system (Payne, 1970) Cockfield Formation (Hosman and others, 1968)	6	Upper Claiborne aquifer
Tertiary	Eocene	Claiborne	Cook Mountain Formation	14	Middle Claiborne confining unit
Tertiary	Eocene	Claiborne	Sparta hydraulic system (Payne, 1968) Sparta Sand (Hosman and others, 1968) Memphis aquifer (Hosman and others, 1968) (layers 4 and 5)	5	Middle Claiborne aquifer
Tertiary	Eocene	Claiborne	Cane River Formation**	13	Lower Claiborne confining unit
Tertiary	Eocene	Claiborne / Wilcox	Carrizo and Meridian Sand aquifer (Payne, 1975) Carrizo Sand and Meridian-upper Wilcox aquifer (Hosman and others, 1968)	4	Lower Claiborne-upper Wilcox aquifer
Tertiary	Eocene	Wilcox	Wilcox Group (Hosman and others, 1968)	3	Middle Wilcox aquifer
Tertiary	Paleocene	Wilcox	Lower Wilcox aquifer (Hosman and others, 1968)	2	Lower Wilcox aquifer*
Tertiary	Paleocene	Midway	Midway Group	12	Midway confining unit 1/

* Not present in the Texas coastal uplands aquifer system.

** Not present north of latitude 35° north.

Table 2.--*Relation of geologic units, previously defined geohydrologic units, and layers used in regional flow model--Continued.*

Coastal lowlands aquifer system

Geologic Unit			Geohydrologic units defined by previous studies	Gulf Coast Regional Aquifer-System Analysis	
System	Series	Group		Model layer number	Geohydrologic units
Quaternary	Pleistocene and Holocene		Upper Chicot aquifer (Jorgensen, 1975) Chicot aquifer (Meyer and Carr, 1979)	11	Permeable zone A (Holocene-upper Pleistocene deposits)
				10	Permeable zone B (Lower Pleistocene-upper Pliocene deposits)
Tertiary	Pliocene		Evangeline aquifer (Whitfield, 1975) (Meyer and Carr, 1979)	9	Permeable zone C (Lower Pliocene-upper Miocene deposits)
	Miocene		'2,000-foot' sand of the Baton Rouge area (Torak and Whiteman, 1982) Jasper aquifer (Whitfield, 1975)	17	Zone D confining unit
				8	Permeable zone D (Middle Miocene deposits)
				16	Zone E confining unit
	Eocene and Oligocene	Jackson and Vicksburg		7	Permeable zone E (Lower Miocene-upper Oligocene deposits)
				15	Vicksburg-Jackson confining unit 1/

1/ The Midway confining unit was referred to as the coastal uplands confining unit and the Vicksburg-Jackson confining unit was referred to as the coastal lowlands confining unit by Grubb, p. 11, (1984).

Due to the similarity in ages and deposition of the sediments of the Texas coastal uplands and Mississippi embayment aquifer systems, they were subdivided into similar hydrologic units. The lowermost unit of the Mississippi embayment aquifer system is the lower Wilcox aquifer (layer 2), which is a sandy interval at or near the bottom of the Wilcox Group. The unit is restricted to the Mississippi embayment aquifer system and does not extend into the Texas coastal uplands aquifer system. The sediments immediately above the Midway confining unit in the Texas coastal uplands aquifer system, and above the lower Wilcox aquifer in the Mississippi embayment aquifer system, are composed predominantly of interbedded coarse- and fine-grained beds. The coarse-grained beds have varying degrees of lateral hydraulic connection, and therefore have a relatively low effective horizontal permeability. The horizontal component of flow was assumed significant enough to consider these sediments collectively as one water-bearing unit because of the large overall thickness. The unit was extended across

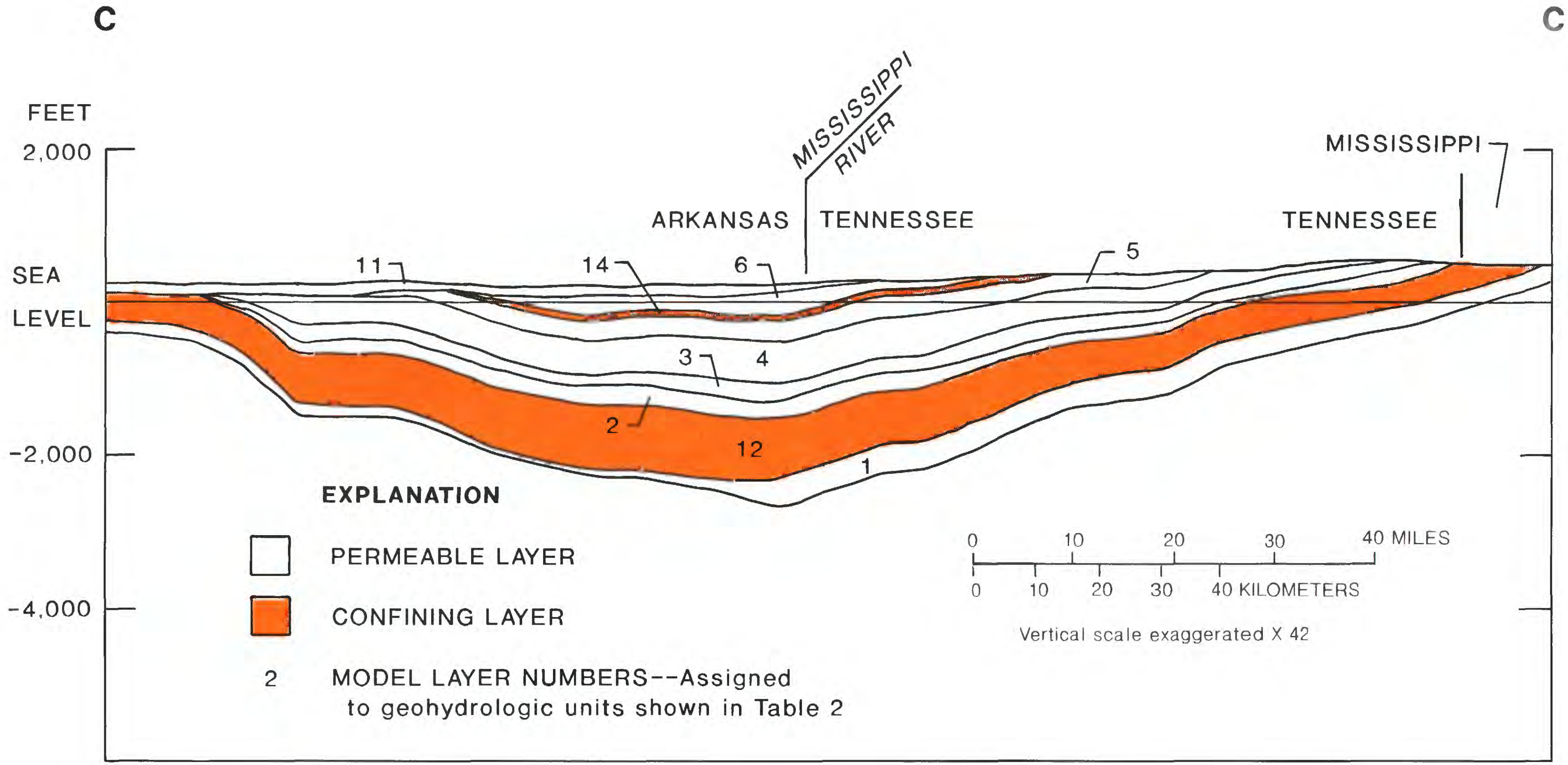


Figure 10.--Section C showing generalized geohydrologic units.

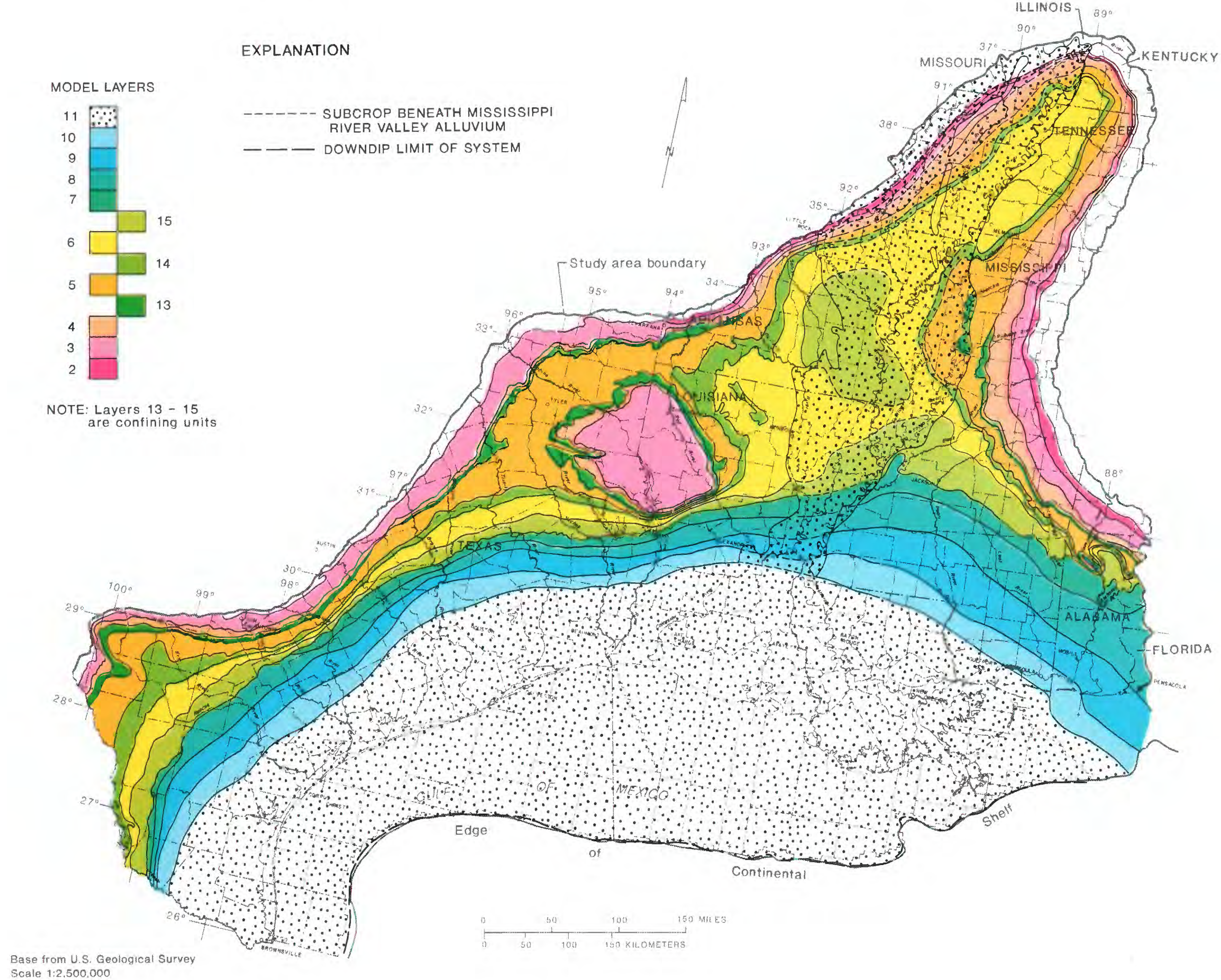


Figure 11.--Outcrop and subcrop of geohydrologic units.

areas of relatively low sand percentage in southeastern Mississippi, southern Alabama, and in the northern Mississippi embayment. Horizontal flow might be minimal in these areas, although they were included in the unit to preserve horizontal continuity. The middle Wilcox aquifer exists without overlying or underlying confining units, being recognized by the contrast in relative permeabilities. Although recognized as a permeable zone, the clays within the middle Wilcox aquifer are the major restriction to vertical flow between overlying and underlying units.

The lower Claiborne-upper Wilcox aquifer (layer 4) is a sandier interval consisting of the top of the Wilcox Group and the bottom of the Claiborne Group, which directly overlies the middle Wilcox aquifer (layer 3). The upper surface of this unit is recognized by the occurrence of a massive clay, identified as the lower Claiborne confining unit (layer 13), which separates the lower Claiborne-upper Wilcox aquifer from the middle Claiborne aquifer. The lower Claiborne confining unit does not extend into the northern Mississippi embayment where layers 4 and 5 together comprise the Memphis aquifer. The lower Claiborne-upper Wilcox aquifer (layer 4) is not present near the Texas-Louisiana State line, due to the structural influence of the Sabine uplift. The middle Claiborne aquifer (layer 5) is overlain by the middle Claiborne confining unit (layer 14), which is in turn overlain by the upper Claiborne aquifer (layer 6). The middle and upper Claiborne aquifers, and the middle Claiborne confining unit are not present over the Sabine uplift near the Texas-Louisiana State line. The upper Claiborne aquifer directly overlies the middle Claiborne aquifer in a small area of northeast Louisiana, where the middle Claiborne confining unit does not exist. The upper Claiborne aquifer is the uppermost unit of the Texas coastal uplands and parts of the Mississippi embayment aquifer systems, and is separated from the coastal lowlands aquifer system by the Vicksburg-Jackson confining unit.

The coastal lowlands aquifer system was subdivided into five permeable zones and two confining units. The lowermost zone (zone E, table 2) is primarily composed of lower Miocene-upper Oligocene deposits. The sand beds of zone E pinch out updip towards the surface in an area of south Texas, although these beds were represented in the model as if they existed with a very small thickness (10 ft) to preserve hydraulic continuity. Permeable zone E is, in most places, separated from the overlying permeable zone (zone D, table 2, composed of middle Miocene deposits) by the zone E confining unit (table 2). Permeable zone D is locally separated from the overlying permeable zone of lower Pliocene-upper Miocene deposits (zone C) by the zone D confining unit. Because neither of the confining units are identifiable at the surface, and the zone D confining unit is not recognized in much of the area, the permeable zones can exist without intervening confining units. All of the sediments above the zone D confining unit lack any regionally extensive clay bodies, and were subdivided into permeable zones A, B, and C without intervening confining units (table 2).

The methods used to subdivide the sediments of the coastal lowlands aquifer system, where confining units are not regionally extensive, was based on recognition of vertical differences in the hydraulic head due to resistance to vertical flow and ground-water pumpage. Subdivisions based on these head differences were defined to minimize variations in hydraulic head within a permeable zone. The zones were extended to areas where the hydraulic-head gradient is unknown by keeping the zone a constant proportion of the total system thickness, thereby avoiding abrupt discontinuities in thickness and thus preserving horizontal hydrologic continuity. In ascending order, the units above zone C (lower Pliocene-upper Miocene deposits) are zone B (lower Pleistocene-upper Pliocene deposits), and zone A (Holocene-upper Pleistocene deposits).

The zone of geopressured sediments truncates the water-bearing zones from the bottom. Geopressured zones encroach shallower units as they occur higher in the section in a gulfward direction.

Ground-water Pumpage

Nearly 10 Bgal/d of ground water were pumped from the gulf coast aquifer systems in 1980 (Mesko and others, in press). Irrigated agriculture uses the largest volume (about three-fourths) of ground water in the study area (Grubb, 1984), and also represents the fastest growth in ground-water pumpage (fig. 12) during the past two decades. Much of the increase in agricultural pumpage has occurred since 1970, especially in Arkansas. The major areas of irrigation are the Mississippi River Alluvial Plain, southwestern Louisiana, and south Texas. Figure 13 shows the distribution of ground-water pumpage in 1980. Note that localized pumpage is generally used for public supply and industry whereas areally high values probably indicate irrigated areas. Over 60 percent of the public supply and industrial pumpage is withdrawn from the coastal lowlands aquifer system. Industrial pumpage at many locations has actually declined since 1970 (fig. 12), mainly in response to the large drawdowns, and in some cases land subsidence and other adverse effects, which have occurred. About three-fourths of the total pumpage is from layer 11, the shallowest layer over much of the study area. The remainder of the pumpage is generally withdrawn from the shallower zones in an area.

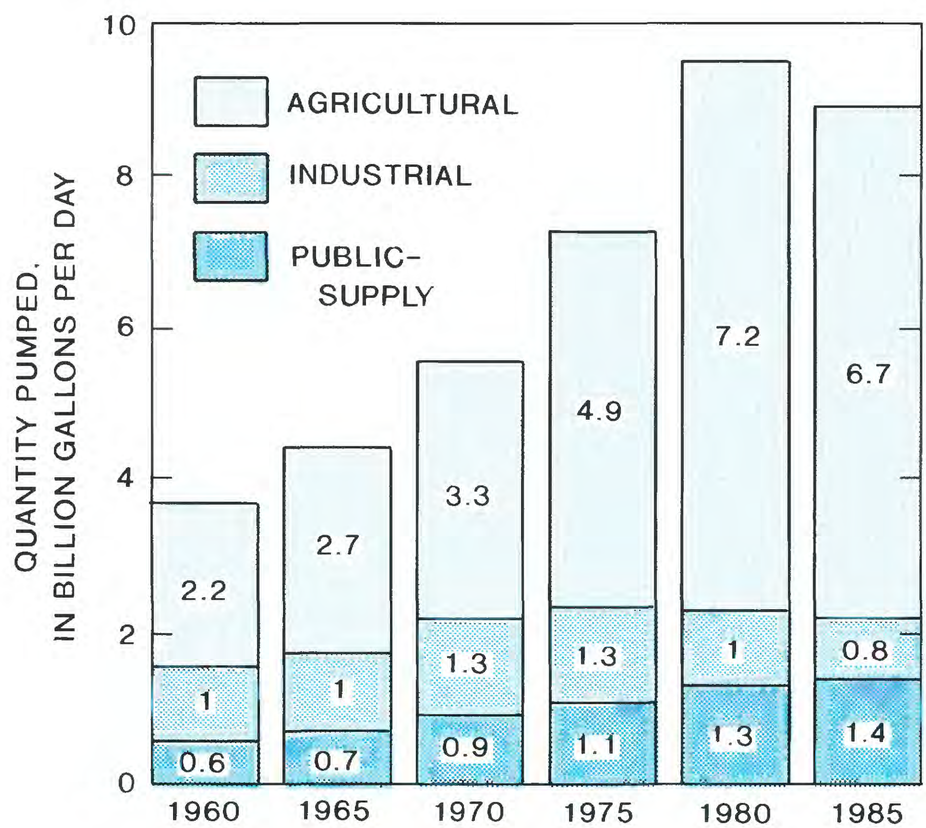


Figure 12.--Ground-water pumpage by use, 1960-85.

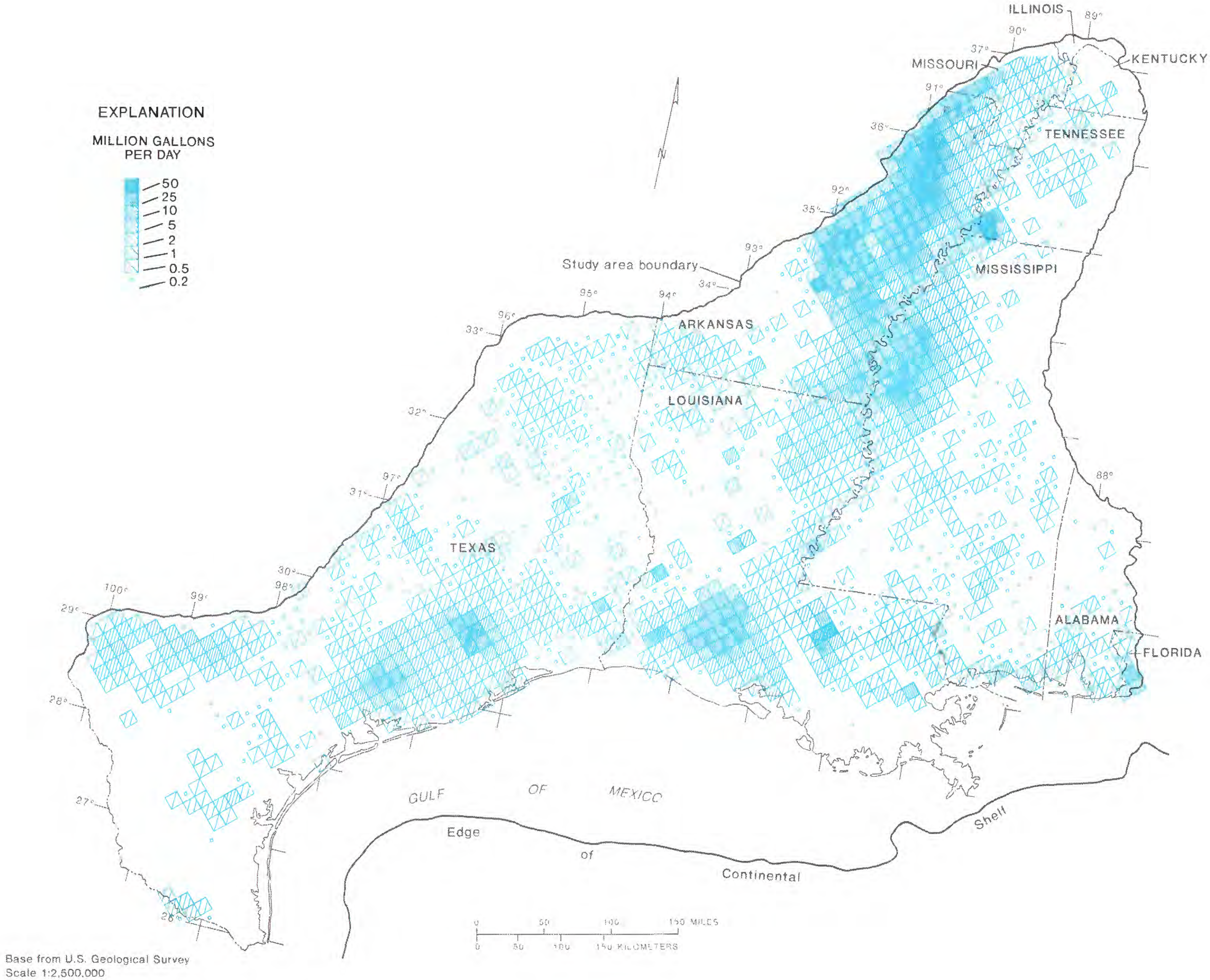


Figure 13a.--Ground-water pumpage from the Gulf Coast aquifer systems, 1980.

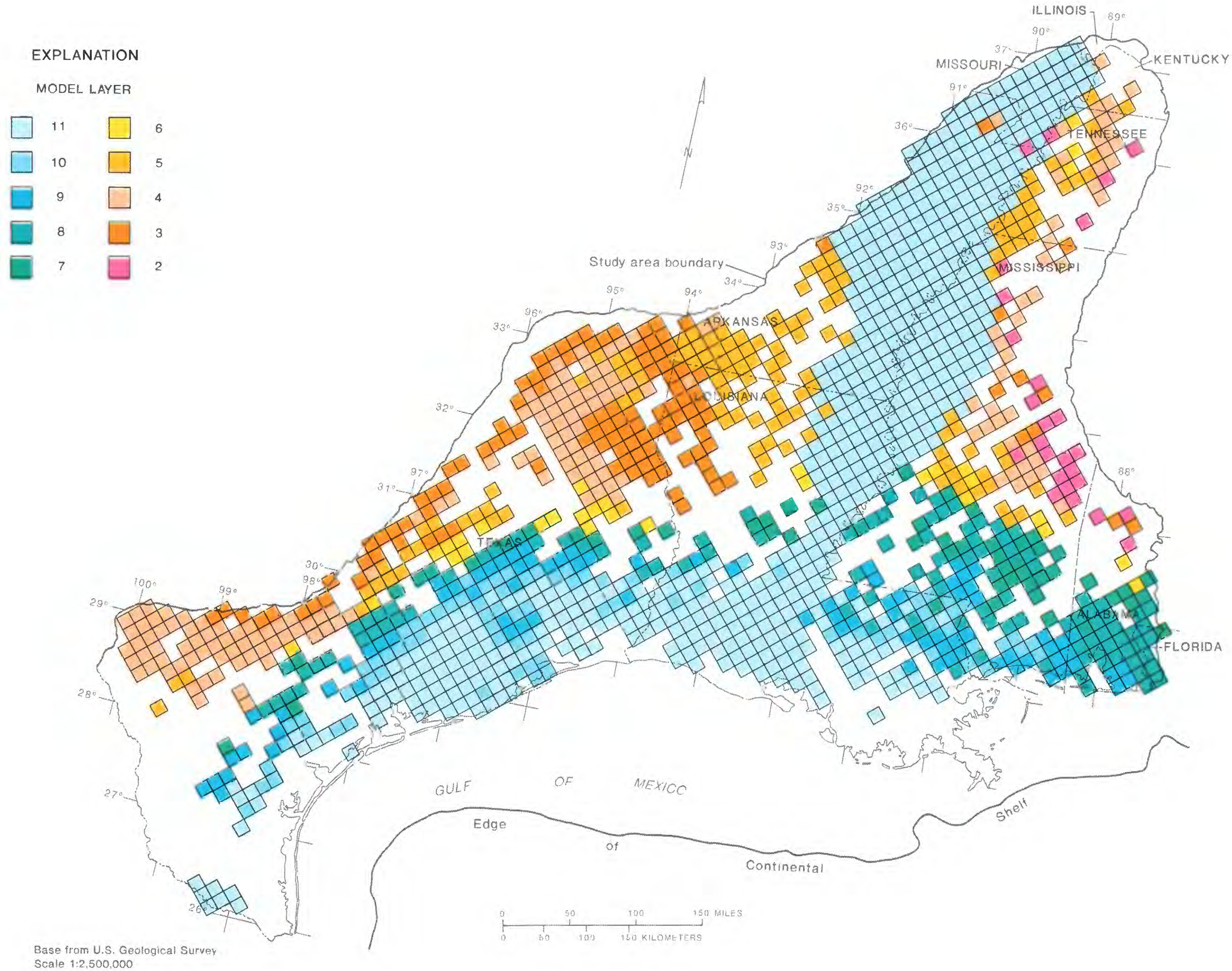


Figure 13b.--Main producing model layer, 1980.

APPROACH TO SIMULATION

The ground-water flow system in the Gulf Coast RASA study area is too complicated to permit simulation of every characteristic, feature, and process. The characteristics and processes selected for this analysis and the features of the model used for the simulation of ground-water flow are discussed in the following sections. The boundaries of the aquifer system and the treatment of recharge and discharge at the land surface will be discussed following a general discussion of flow concepts and how they apply in the study area at this regional scale of analysis. The numerical model is then described followed by an overview of the discretization process.

Flow Concepts

Aquifers are recharged in areas of higher topography either directly by infiltrating rainfall, or by seepage from streams and lakes. Even though vertical hydraulic conductivity (K) is much smaller than horizontal K, the main component of regional flow in the recharge areas could be vertically downward, due to the geometry of the system and its head gradients. Water is discharged from the aquifer system in areas of relatively low topography to streams, lakes, the ocean, and to the soil where some of it is evapotranspired. The main component of regional flow in discharge areas could be vertically upward to the surface. In between recharge areas and discharge areas, the main component of flow is probably horizontal. Development has largely modified the amounts, locations, and directions of these flows mainly due to the introduction of wells which discharge water from specific depths in the aquifer for use at the land surface.

These flow conditions occur at various scales, from small valleys where there is recharge on the hill and discharge to a creek in the valley, to regional flow patterns (fig. 14). The very small scale flow paths tend to have higher flow rates, because the flow paths are shorter and the gradients are steeper. These will not be shown or discussed in this report because these flow paths usually occur within a single model block. The net of recharge and discharge for a model block is the recharge minus the discharge which occurs within the block. It can have a positive sign (net recharge) or negative sign (net discharge).

The amount of recharge to the regional ground-water flow system is probably not related to the average annual rainfall over most of the study area because of the relatively humid climate. It is more likely that the amount of rainfall which recharges the regional flow system is limited by the capacity of the flow system to carry ground water away from recharge areas. Where the flow capacity is limited by the small regional hydraulic gradients or by the resistance to flow of the aquifer system, most of the rainfall which potentially would recharge the regional aquifer systems is discharged to local surface-water bodies such as creeks and streams or through evapotranspiration. In the western part of the area where the climate is much drier, recharge to the regional ground-water flow system probably is limited by the smaller rainfall amounts, most of which falls in the hot summer months when it quickly evaporates or is transpired by vegetation.

The way that the geohydrologic units (table 2) are represented in the model is shown in figure 15. Where interior layers pinch out, the adjacent layers are directly connected with no extra resistance to flow other than what is inherent in their hydraulic conductivities and thicknesses. This includes the Mississippi River Valley alluvial aquifer (layer 11) where it overlies the Mississippi embayment aquifer system (layers 2 through 6).

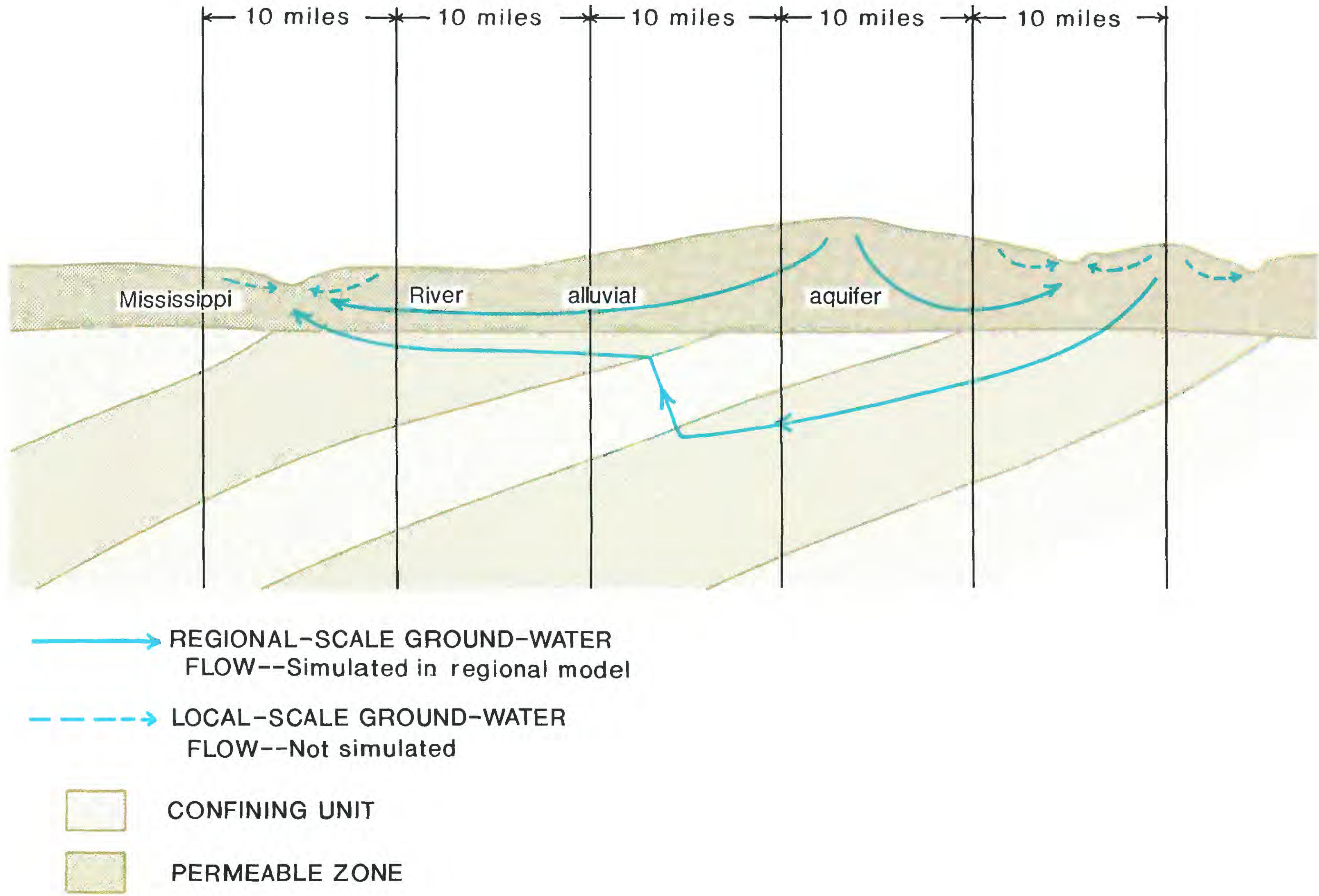


Figure 14.--Diagram showing local- and regional-scale ground-water flows. (Modified from R.H. Johnston, written communication, 1989)

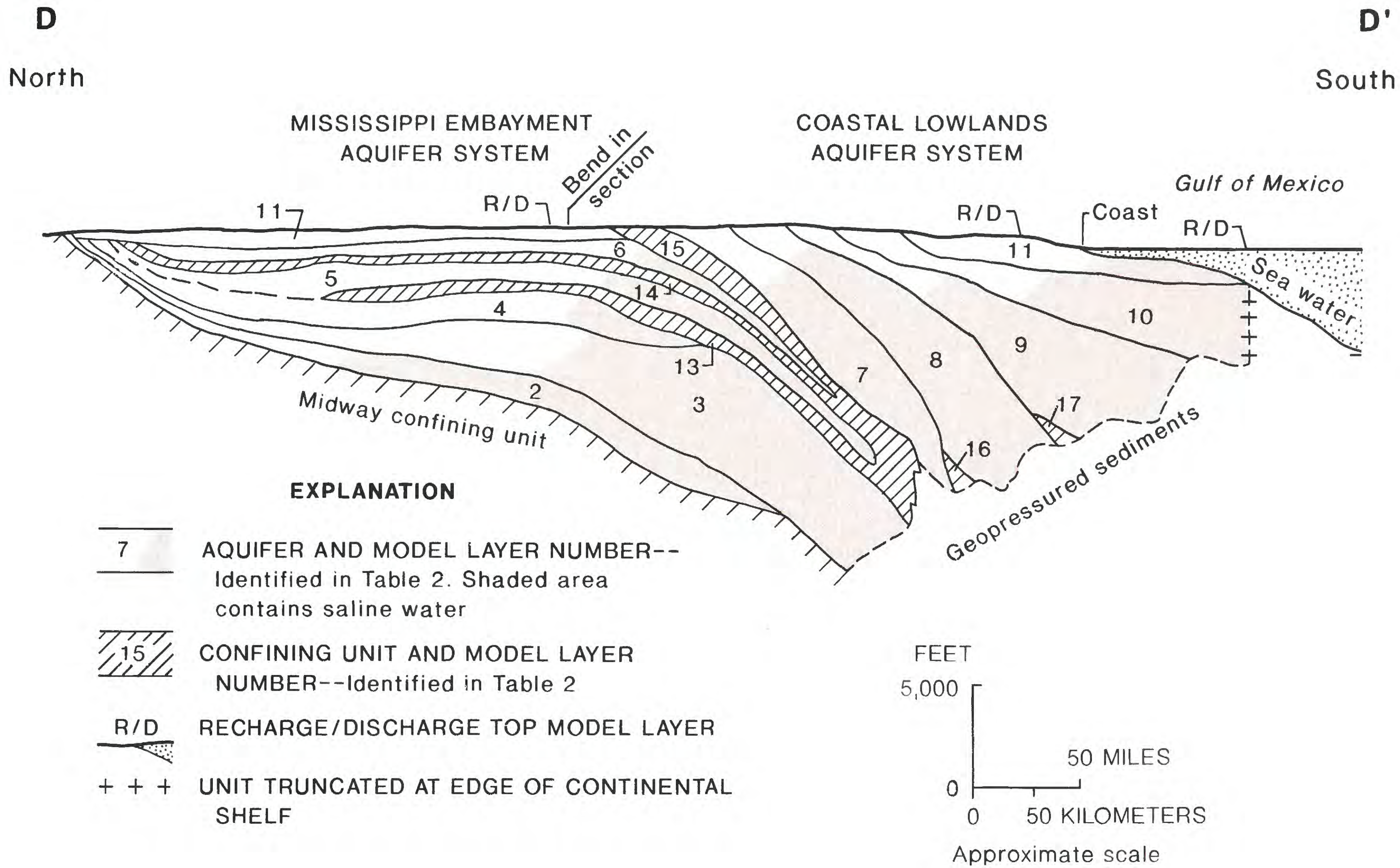


Figure 15a.--Sketch showing vertical relations of the aquifer systems.

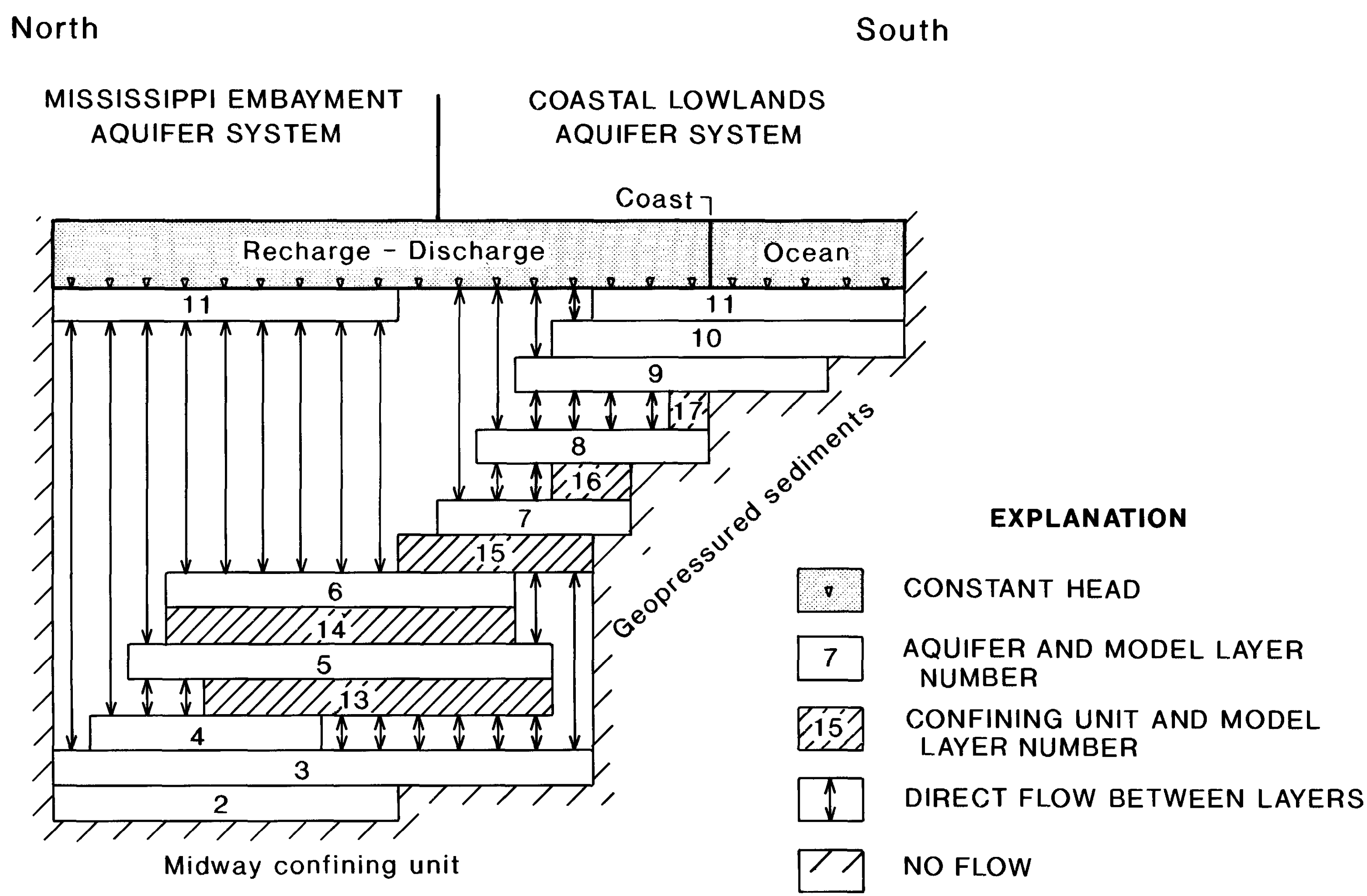


Figure 15b.-- Diagram showing idealized representation of the aquifer systems in the regional flow model.

There are two types of resistance to vertical flow: across regionally identifiable confining units, and within aquifers. The aquifers contain many fine-grained beds which may have considerable areal extent locally, yet are not traceable over multi-county areas. Vertical resistance to flow may be proportional to the net thickness of clay beds, whether they have regional lateral extent, or not. Regional clay units, where they exist, may cause most of the resistance to vertical flow probably because of minimal hydraulic conductivity and the largest thickness of fine-grained deposits.

Some vertical flow occurs between aquifer systems across the Vicksburg-Jackson confining unit, thus it is not an impermeable barrier to flow as might be suggested from its thickness and lithology. The downdip hydraulic gradient which continues underneath the overlying clay unit, indicates horizontal flow which must discharge up through the confining unit. This situation also exists below the Midway confining unit at the base of the aquifer systems. However, the effect of this upward leakage on the underlying flow system is much greater than on the overlying flow system because it represents a greater proportion of the flow in the lower system. These concepts are similar in nature to those discussed by Bredehoeft and others (1983) concerning the regional confining units above the Cretaceous Dakota aquifer in South Dakota.

The horizontal hydraulic conductivity of sand beds is many times higher than the conductivity of fine-grained beds, therefore, the fine-grained beds can be ignored in the calculation of transmissivity of the aquifers. The transmissivity of the aquifer layers is therefore considered to be equal to the sum of the thicknesses of the sand beds in a layer times the effective hydraulic conductivity of the sand beds.

The natural geothermal increase in temperature with depth causes a corresponding decrease in kinematic viscosity which, in turn, causes an increase in hydraulic conductivity (K). This effect can change the viscosity by a factor of several times (Rouse, 1950, p. 1,011) from near land surface where the water temperature is about 20 degrees Celsius to depths of more than 10,000 ft where the temperature can be more than 150 degrees Celsius. This increase in K is partially negated by an increase in density of water downdip due to increased salinity which causes a slight increase in viscosity and corresponding decrease in hydraulic conductivity. The estimation of K is further compensated for by the decrease in porosity and permeability due to compaction through increasing pressure with depth. For this preliminary analysis, it is assumed that these factors cancel each other out within the limits of our accuracy in estimating conductivity.

The presence of saline water at depth throughout most of the gulf coast aquifer systems affects the ground-water flow in the aquifers. Freshwater tends to "float" up over water of high salinity and density, whereas heavier, saline water moves downward and under fresher waters if possible. The flow downdip in the saline part of the system is probably reduced somewhat by this effect although the simulations presented in this report will not assume that the brine is stagnant. It is unknown whether the deep brine system is in equilibrium with other large changes such as sea level changes and natural sediment compaction. It has probably not reached equilibrium conditions yet, due to the long flow paths and the relatively low conductivities.

For the purpose of this study, density is assumed constant in time so that the effect of solute transport on ground-water flow can be ignored. This assumption is valid because in the short time scales of a few decades, the volume of water that can move into an adjacent model block at maximum flow rates will not significantly affect the average density of the water in the large volumes of aquifer represented by the model blocks.

Large differences in water quality on opposite sides of confining units or large clay beds can introduce osmotic gradients. These osmotic gradients induce flow from the fresher side of the confining unit toward the more saline side in order to equalize the concentrations. These gradients might cause significant amounts of flow under certain conditions. However, the estimation of the rate of this flow is wholly dependent on estimating osmotic conductivities. The values of osmotic conductivity can vary many orders of magnitude and the characteristics of the confining units and water chemistry, which control the values, are largely unknown. Therefore, this effect has been ignored except that it will be considered as a possible source of systematic error in flow rates in the saline part of the aquifers.

Boundaries

The gulf coast aquifer systems have boundaries below them and downdip, along the strike at both ends, updip at the outcrop and above them with the surface hydrologic system, that is, weather, lakes, streams, and ocean. The conditions at each of these boundaries must be approximated in the simulations.

Base of Flow System

The base of the flow system is represented in the regional flow model at the top of the Midway confining unit, or the top of the zone of geopressure, whichever is shallowest (figs. 9 and 15). As stated above, the effect of the small amount of flow leaking up through the Midway confining unit, which is composed of marine clays several hundred feet thick, is thought to be insignificant compared to the volume of flow in the aquifer above it. A detailed investigation into this possibility is being done as part of the analysis of McNairy-Nacotoch aquifer, which directly underlies the Midway confining unit in the northern Mississippi embayment (Brahana, 1987, and Brahana and Mesko, 1988, p. 38). The McNairy-Nacotoch aquifer in this area is composed predominantly of sand, and is distinctly different from the Upper Cretaceous limestone in the southern Mississippi embayment (approximately south of the 35th parallel) and below the Texas coastal uplands aquifer system. In these areas, the Midway confining unit has been assumed to be completely effective as a no-flow boundary.

An exception to the basal Midway confining unit being the base of the flow system exists in the extreme northwestern part of the Mississippi embayment. The Mississippi River Valley alluvial aquifer extends beyond the subcrop of the Midway confining unit in an area of a few hundred square miles and directly overlies the McNairy-Nacotoch aquifer and Paleozoic rocks of the Central Midwest aquifer system. It is possible that this occurrence, in addition to the McNairy-Nacotoch aquifer directly overlying the Paleozoic sediments, allows a transfer of water between the Central Midwest aquifer system and the Mississippi embayment aquifer system.

Downdip and gulfward in the gulf coast aquifer systems, the top of the zone of geopressure is above the top of the Midway confining unit. Geopressured zones exist in the transition from a predominantly sand facies to a predominantly clay facies. The occurrence of overpressured deposits is evidence that hydraulic conductivities are very small, otherwise fluid expulsion would not be restricted, and pressures would not build up. Therefore, the base of the flow system is assumed to be the top of the geopressured zone where it occurs above the top of the Midway confining unit. However, very small amounts of fluid probably do migrate upwards from geopressured zones, particularly along planes of growth faults, thus eventually relieving the overpressure.

Lateral Boundaries

On the eastern edge of the study area near the Gulf of Mexico, the lithology of most of the aquifers undergoes a facies change. Permeable zones in the coastal lowlands aquifer system pinch out. The Vicksburg-Jackson confining unit gradually grades into the more permeable limestone beds of the Floridan aquifer system to the east. Much of the Mississippi embayment aquifer system becomes calcareous with greatly reduced hydraulic conductivities. Most of the coastal lowlands aquifer system outcrop curves to the south and offshore. All of these factors tend to restrict flow across the eastern boundary of this study area. This boundary was treated as a no-flow boundary.

The only data available on the gulf coast aquifer systems in Mexico was topography and a surficial geology map (Direccion General de Geografica del Territorio Nacional, 1984). The approach was to use speculative data at distances from any area of interest, rather than try to force an arbitrary hydrologic boundary at the international boundary. Data from several geophysical logs on the United States side of the border were duplicated and transposed to a location in Mexico in about the equivalent stratigraphic position as shown by the surficial geology map and used to represent the units in that location.

Surface Recharge and Discharge

Neither estimates of the value of the net recharge and discharge nor its components are available over most of the study area. Most of the recharge and discharge occurs where the ground-water system interacts with sources or sinks of water at or near the land surface. A constant head boundary condition was used for the layer above the top aquifer layer, so that the hydraulic head does not vary in time in this source-sink layer. The flow into or out of the top layer will be calculated by the model as the product of (1) the difference in head between the uppermost active aquifer block in the appropriate layer and the constant-head value above it, and (2) the conductance to vertical flow between the two blocks. This conductance is based on the harmonic mean conductance of the top one-half of the aquifer block and the bottom half of the constant-head block. The constant head layer conductance is uniform all over the study area, for simplicity, and simulates resistance to water entering or leaving the aquifer through features such as stream beds, the soil zone, lake bottoms, etc. This constant-head layer accounts for ground-water and surface-water interaction from all surface sources and sinks such as rainfall which has infiltrated into the soil, seepage into or out of lakes and streams, and evapotranspiration from ground water.

This method works relatively well for estimating net recharge and discharge for steady-state conditions, especially when there are no alternative methods that will work at this scale in this type of a system. Other methods can frequently estimate net recharge and discharge values with the wrong signs, that is, recharge where there should be discharge, or vice versa. This is because, in a relatively humid climate with large streams crossing the aquifer system, the magnitude of all surface sources and sinks of water is so large compared to the ground-water component (involved in the regional flow system), that the regional ground-water component is usually of the same order of magnitude or smaller than the error in estimating surface-water gains or losses. However, under transient conditions, the method chosen in this study can overestimate the volume of recharge unless it is limited to a maximum reasonable amount.

Numerical Model

The computer program being used to simulate variable density ground-water flow in three dimensions is that of Kuiper (1983, 1985). It was chosen for this study because it can handle variable-density flow in a straightforward manner that has been tested against analytical solutions and because it is well documented. The model uses an integrated finite-difference grid in which the thickness of the layers is specified separately from their hydraulic conductivities. Vertical hydraulic conductivity is specified for active aquifer layers as well as for inactive confining units. Inactive confining units have no storage, horizontal conductivity, nor horizontal flow. As stated previously, the water density, although variable in space, is assumed to be approximately constant in time. The program has three types of solution algorithms: the strongly implicit procedure (SIP); successive over-relaxation (SOR); and preconditioned conjugate gradient (PCG, which has eight sub-types). The PCG method achieved a solution to some simulations not solvable with the other methods. The PCG method was also faster in some simulations as has been documented previously (Kuiper, 1981 and 1987), especially in linear, steady-state problems.

Model Discretization

Numerical simulation requires that the aquifer system be discretized into blocks over which the characteristics of the aquifer or confining unit materials are averaged. The immense areal extent of the gulf coast aquifer systems required that the regional model horizontal grid spacing be large--10 mi on a side and vertical discretization be limited to 10 aquifer (active) layers. Even with these large blocks, there are 5,916 total blocks per layer (102 rows by 58 columns) times 10 active layers which equals 59,160 blocks, although only about one-fourth of these are active due to the dip of the system. The horizontal grid is oriented approximately 45 degrees from north (plate 1 and fig. 2). This orientation was chosen mainly for convenience, minimizing the size of the matrices required to fit the entire areal extent of the aquifer systems, including the offshore portions. Note that although figure 2 has been reduced to fit page size, plate 1 shows the model grid at the same scale as the other maps in this report.

Even with these large blocks, the regional features affecting flow in the aquifer systems are still generally preserved. Probably the most difficult aspect of the aquifer systems to simulate with this large grid spacing is the pattern of the outcrop of each layer. The sub-regional models use 5-mile horizontal grid spacing for more detail, while keeping the same vertical discretization. All data sets were prepared for model blocks on a 5-mile grid spacing suitable for the sub-regional modeling. Values for the regional model were calculated from the 5-mile arrays by calculating the mean of the values in the four 5-mile blocks which corresponded to a 10-mile block, or, as in the case of pumpage, the sum.

Many of the aquifer characteristics, such as thickness, sand percentage, temperature, concentrations of dissolved solids, etc., were derived from a computerized file of 989 geophysical logs (Wilson and Hosman, 1988). The logs were chosen to show regional trends with a nearly uniform distribution of both the onshore (895 logs) and offshore (94 logs) parts of the aquifer system (fig. 16).

Two mutually exclusive computational approaches to discretizing aquifer geometry could have been used. In a system with dipping beds, both have advantages and drawbacks. The first approach involves interpolating altitudes of the tops of aquifers and then subtracting adjacent tops to get thicknesses, or isopach maps. This approach could easily result in small areas of negative thicknesses (obviously erroneous) where

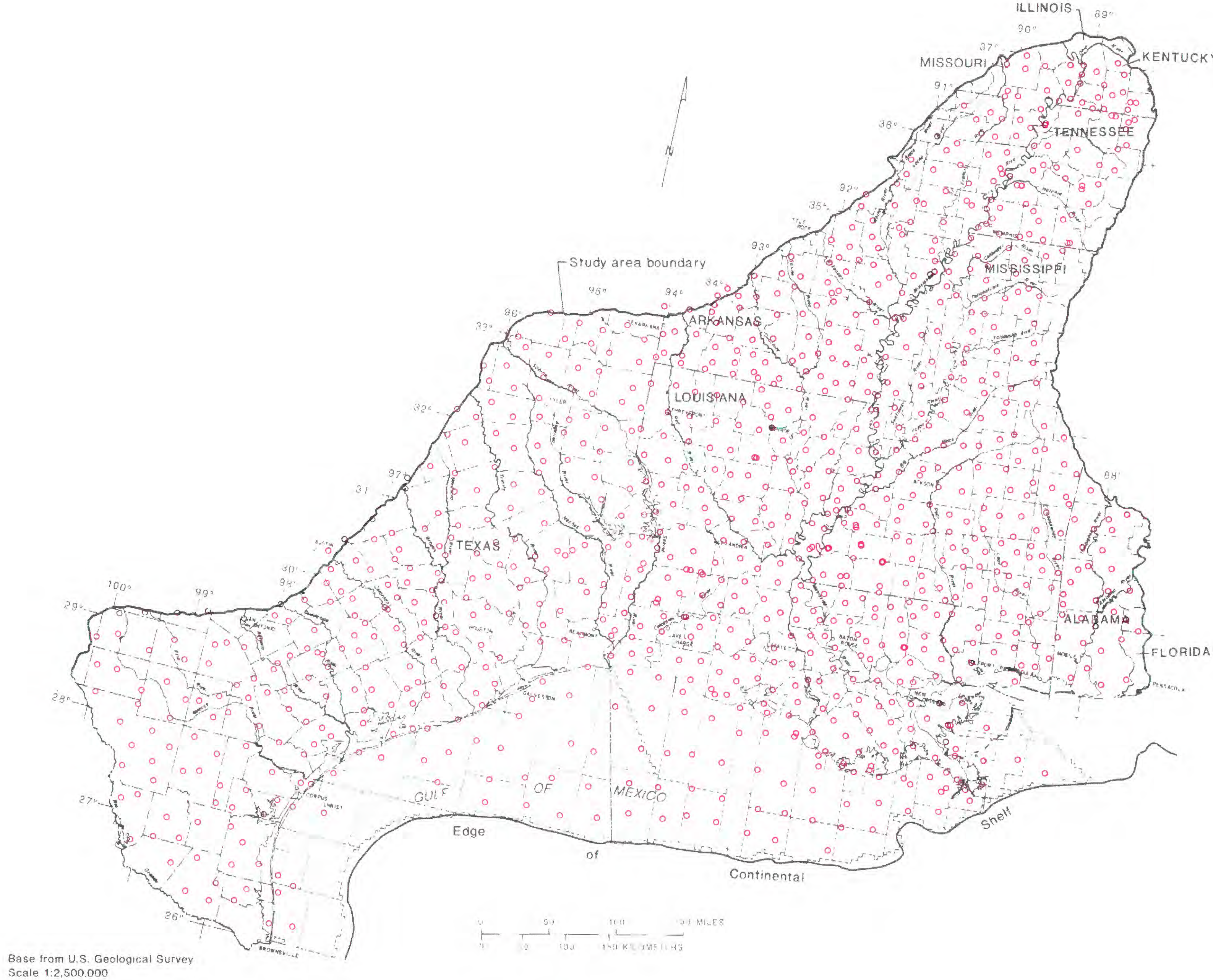


Figure 16.--Geophysical well-log locations. (Modified from Wilson and Hosman, 1988)

the layers are relatively thin and the structure is dipping. The second approach involves interpolating thicknesses and cumulatively subtracting them from the land-surface altitude to estimate altitude of aquifer tops. The interpolation error can increase cumulatively as depth increases. This method also has the drawback of producing top maps that look a little odd in that they tend to have little benches around each control point. Both methods yield equal results near the control points (wells), but differ in between and beyond the control points. The second method was chosen for this study assuming that simulation is most dependent on the thickness of the aquifers, rather than the absolute altitudes, and that the largest errors occurring at the deepest depths would have the least effect on the simulation.

Spacial interpolation of point data to model grid-block locations was done using an inverse distance-squared weighted averaging procedure of the Surface II Graphics System 1/ (Sampson, 1978). Universal kriging, which makes optimum use of the spacial autocorrelation of the data was tried but did not change the resulting layer thicknesses significantly. Kriging does have a slight advantage because it decreases the effect of flattening the surfaces (making small benches) around control points. For some data sets, such as the altitude of the top of some layer or bottom of the system, which have a general slope which extends beyond the limit of data, extrapolation was found to be improved substantially by using kriging or a two-phase gridding technique of the Surface II software package. The estimates from kriging were very insensitive to the value picked for the slope of the semi-variogram. In the two-phase technique, first, slopes of the rate of change of the data are calculated from each neighboring data point toward the block to be estimated, then an inverse-squared weighting is used to average the values calculated in phase one.

Although Surface II has options to select four different types of neighborhood selection procedures, the quadrant procedure, where the two nearest points from each quadrant are selected, was generally used. An error analysis option of Surface II calculates the differences between the original point data value and values interpolated from the four adjacent gridded values to the location of the data point. The standard deviation of these differences, an estimate of the error of the interpolation procedure, was nearly always less than 10 percent of the mean value of the point data. The arrays used in the simulation were used directly from Surface II. Array data which had been extrapolated beyond the areal extent of a layer as defined by the discretized outcrop and subcrop arrays (plate 4) was set to zero. Use of these computer programs to directly generate all model arrays allowed the luxury of frequently updating the model data sets as more data became available or as errors were discovered. Even a total restructuring of the definition of aquifer layers in the coastal lowlands aquifer system was completed in mid-project.

AQUIFER-SYSTEM CHARACTERISTICS

The characteristics of the gulf coast aquifer systems which affect the simulation of ground-water flow are described in this section. The methods used to estimate the values of these characteristics are described as well as the methods used to regionalize the results so that they are appropriate and significant for this scale of analysis. Most of these characteristics remain constant over long periods of time, whereas some of them, like hydraulic head, pumpage, and to some extent density, must be described at specific times.

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

Location of Outcrops, Subcrops, and Extent of Model Layers

The outcropping (or subcropping under the Mississippi River Valley alluvial aquifer) layer was mapped using available geologic maps and the surficial layer coded on the 989 geophysical well logs (fig. 16) (Hosman and Weiss, 1988). The outcrop (or subcrop) band was discretized to a 5-mile model grid by specifying the stratigraphically highest layer that covered at least one-half of the 5-mile block (plate 4). Arrays of the extent of layers on 5- and 10-mile model grids were made by computer program from the layer outcrop file by assuming that all layers stratigraphically below the outcropping (or subcropping) layer could exist. If any one of the four 5-mile model blocks existed, then the 10-mile block was assumed to exist. This was done to preserve compatibility and comparability between the regional and the sub-regional models, as much as possible.

Model Layer Thickness and Sand Percentage

Layer thickness was calculated from the geophysical well-log data using the procedures outlined above to subdivide the aquifer system into model layers. The sand percentage of the layer was calculated by summing the thickness of all sand beds and dividing by the total layer thickness. Composite thickness and sand percentage of the coastal lowlands, Mississippi embayment, and the Texas coastal uplands aquifer systems are shown in figures 17 and 18. Thickness and sand percentage of the model layers are summarized in table 3 and the thickness of each layer and the downdip extent of freshwater (<10,000 mg/L dissolved solids) is shown in figure 19. Layers 8, 9, and 10 have the largest average thickness as well as the largest average net sand thickness, while layer 3 has the largest areal extent. The lowest average sand percentage occurs in layers 7 and 9. Somewhat surprising is that layer 3 has the largest average net sand thickness in the Texas coastal uplands and Mississippi embayment aquifer systems. The highest sand percentage occurs in layers 4, 2, and 11, respectively. The distribution of layer thickness and sand percentage is described fully elsewhere (Grubb, 1986 and 1987, Hosman and Weiss, 1988, and J.S. Weiss, written commun., 1986).

Horizontal Hydraulic Conductivity and Transmissivity

Assuming that nearly all of the significant horizontal flow occurs in sand beds, the transmissivity of each model layer was calculated as the product of layer thickness, sand percentage, and the effective hydraulic conductivity of the sand beds. The value of each component was extrapolated to the block locations from the point data first and then the transmissivity was calculated. Since the layer thickness is entered into the computer program of Kuiper (1985) separately from the conductivity, the conductivity was entered as the product of the sand percentage and the effective hydraulic conductivity. One constant value of effective hydraulic conductivity was used for each layer in these preliminary simulations.

The effective horizontal hydraulic conductivity on a regional scale must be less than the mean conductivity determined from aquifer tests because: (1) The presence of clay beds within the aquifer disrupts the horizontal continuity of the sand beds, and (2) hydraulic conductivity varies between beds. If a given flow path went through a group of beds consisting of very low conductivity clay beds and higher conductivity sand beds with equal lengths of the flowpath in each bed (analogous to series flow in electrical circuits), the overall effective conductivity would be equal to the harmonic mean (inverse of the mean of the inverses of the values) of the conductivities of the individual beds. If the flow could go through any of a group of beds (analogous to parallel flows in electrical circuits), all with equal flow areas, the effective

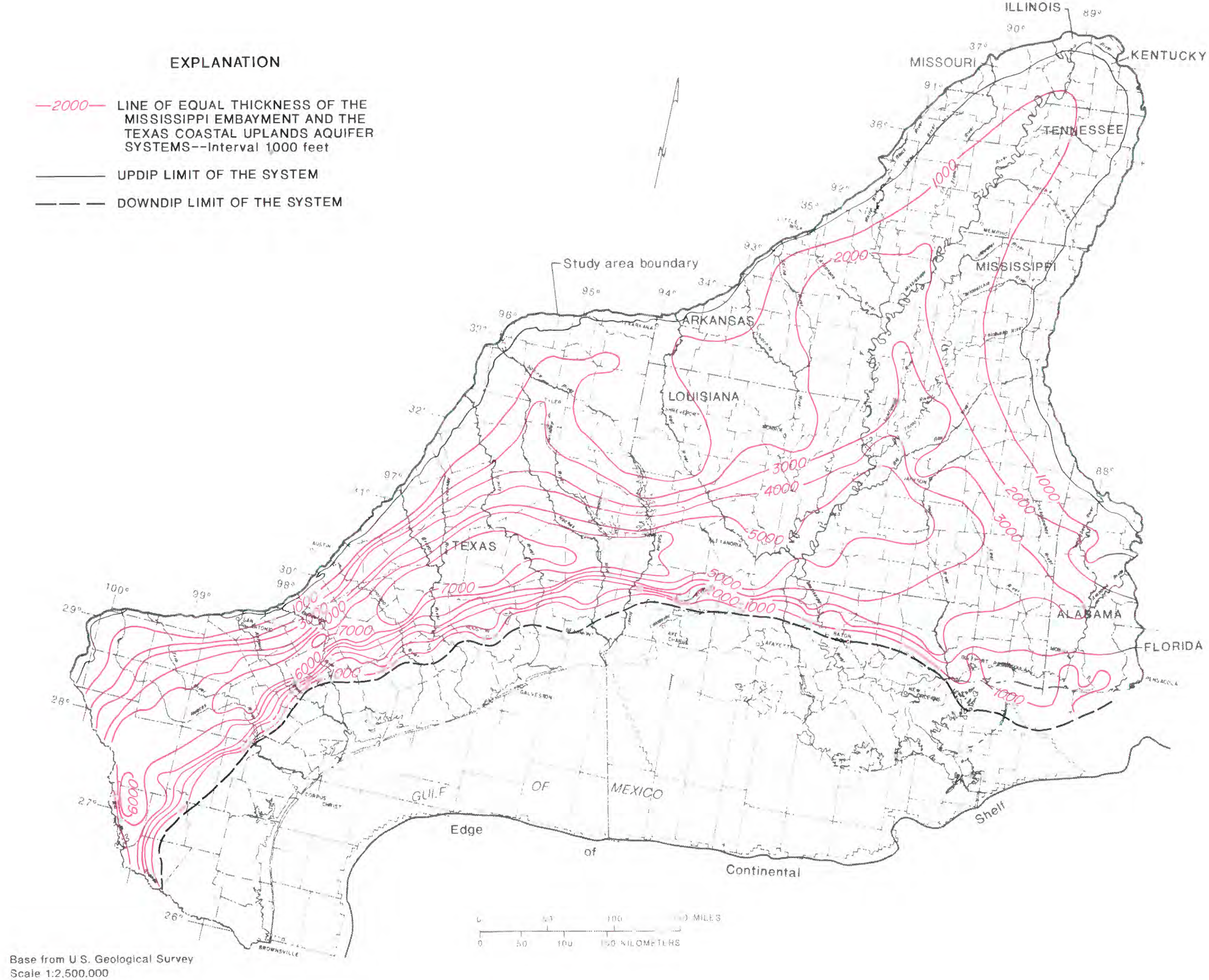


Figure 17a.--Thickness of the Mississippi embayment, Texas coastal uplands aquifer systems.

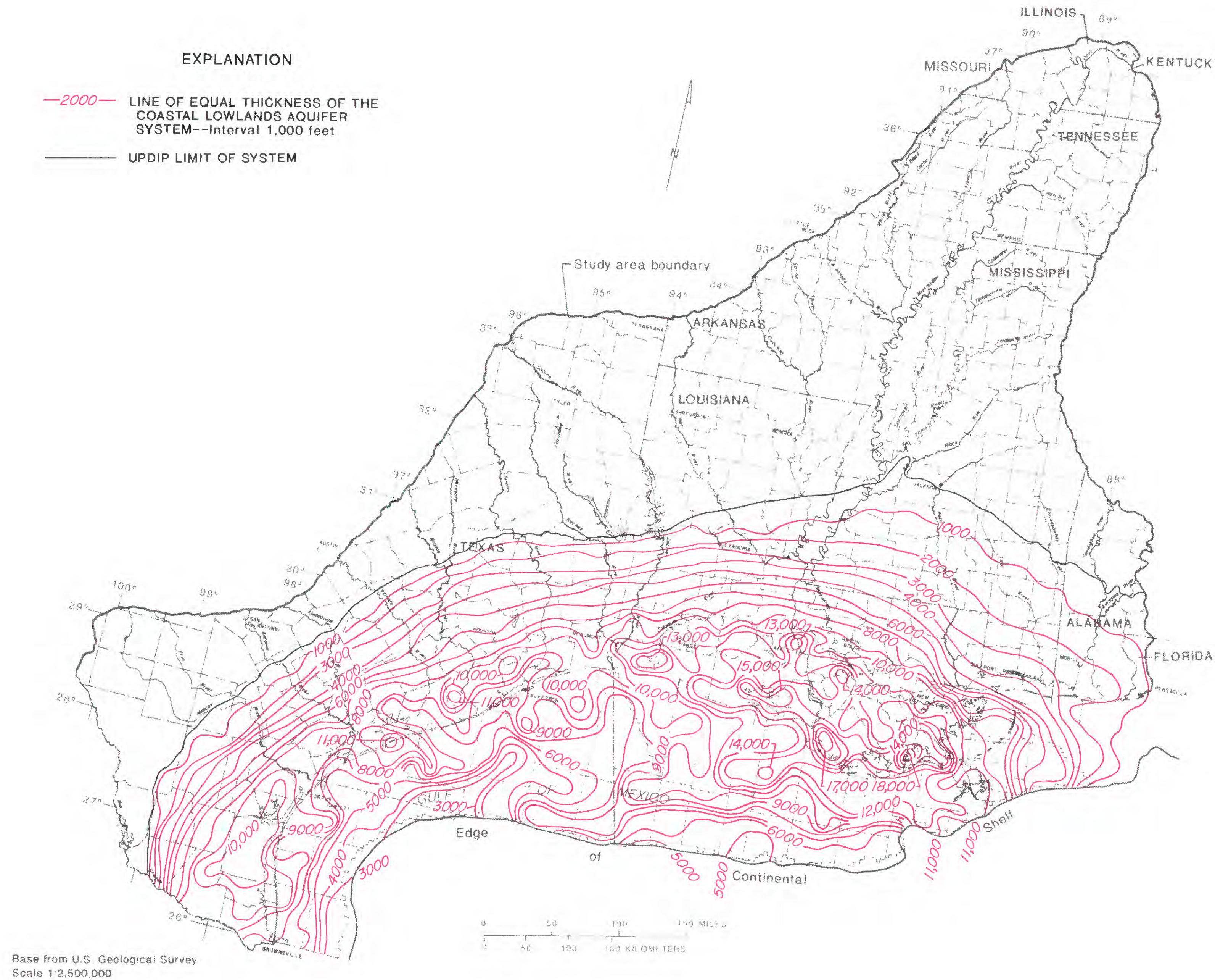


Figure 17b.--Thickness of the coastal lowlands aquifer systems.

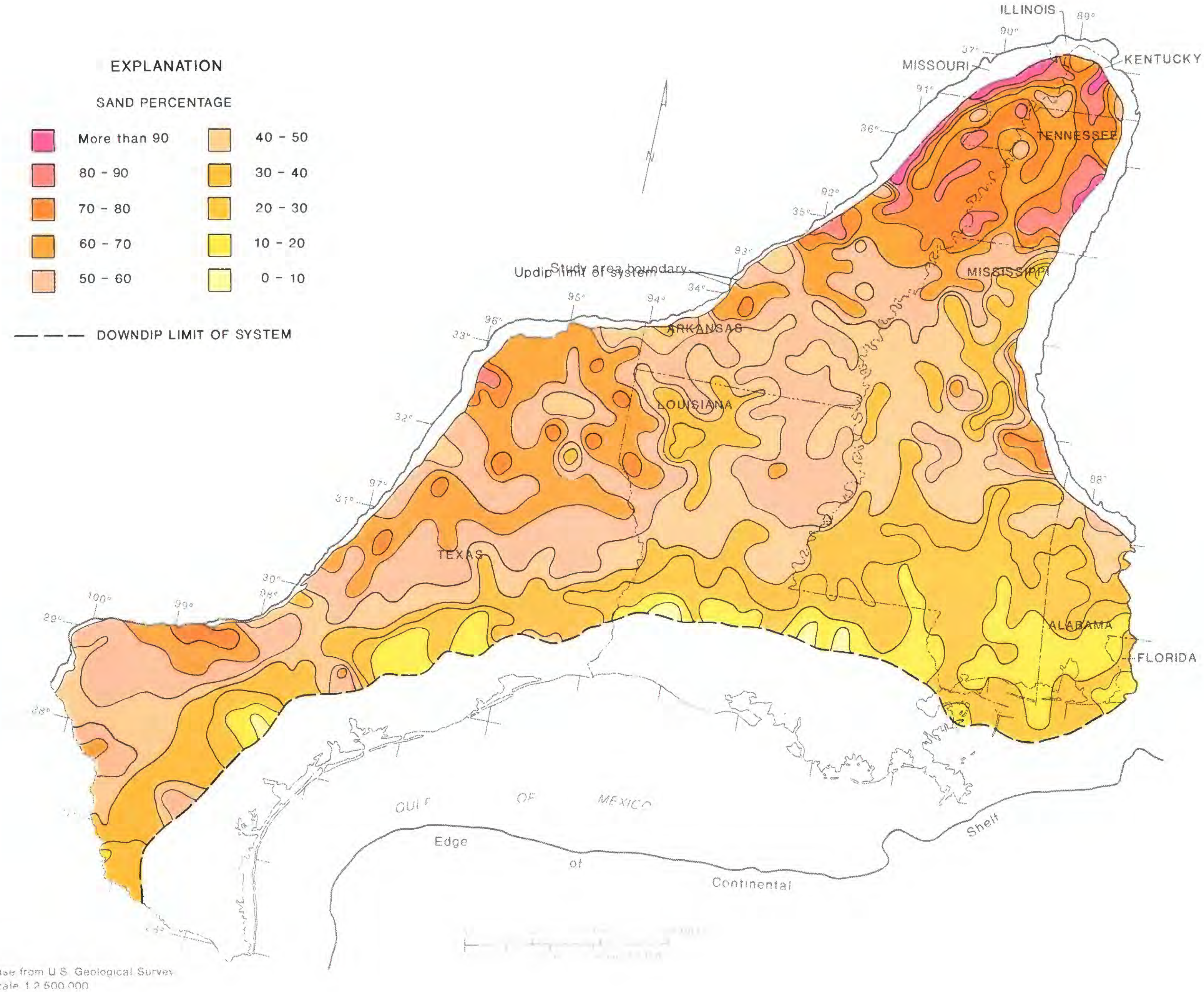


Figure 18a.--Sand percentage of the Mississippi embayment, Texas coastal aquifer systems.

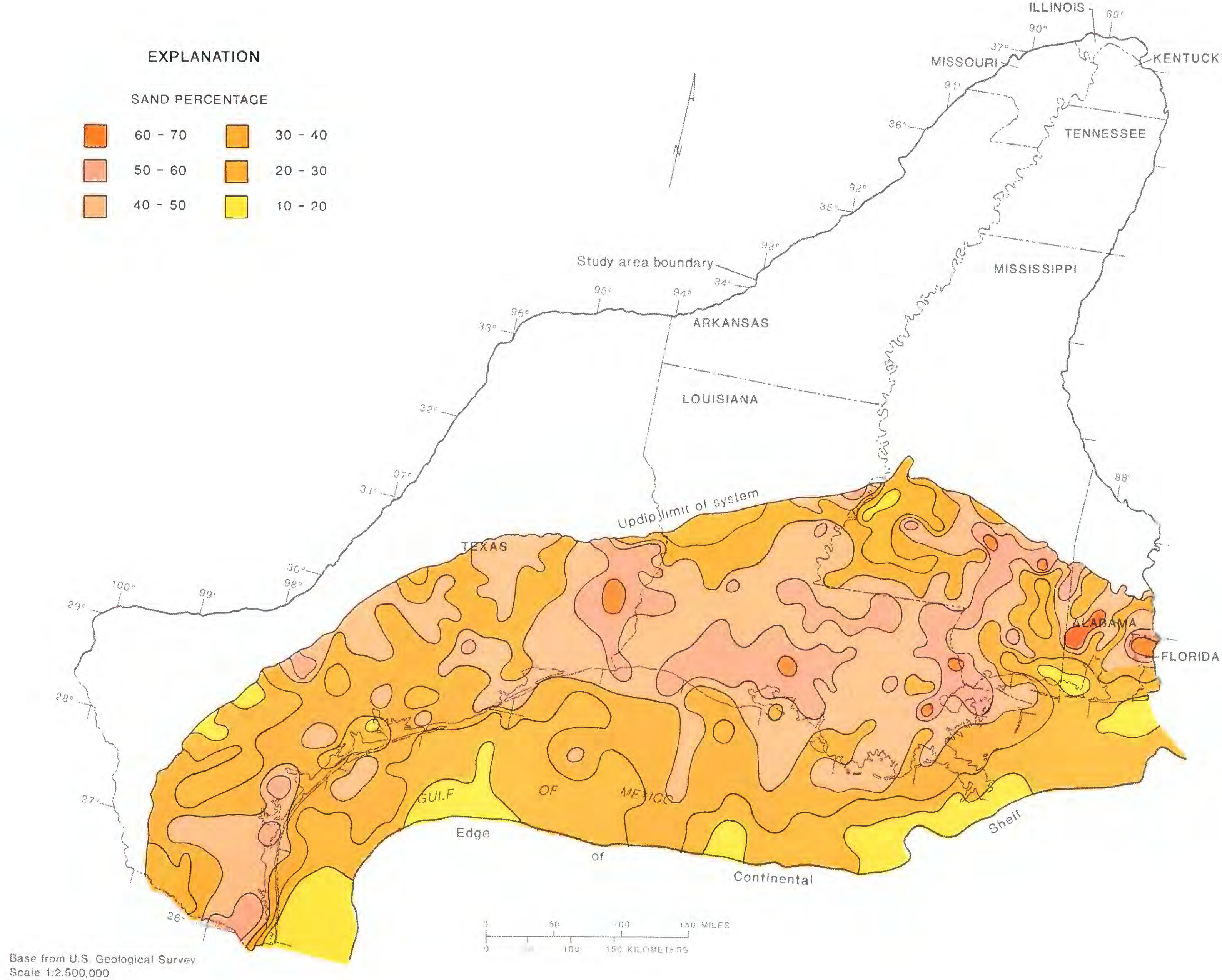


Figure 18b.--Sand percentage of the coastal lowlands aquifer systems.

Table 3.--*Summary of thickness and sand percentage of model layers*

[Some numbers may not agree due to rounding]

Layer number	Area (1,000 square miles)	Thickness (feet)				Sand percentage			Mean aggregate thickness (feet)	
		Number of logs	Mean	Standard deviation	Maximum	Number of logs	Mean	Standard deviation	Sand	Fine-grained
11	150	286	560	340	1,220	279	56	23	310	250
10	130	334	1,890	1,600	5,650	334	42	17	790	1,100
9	140	379	1,960	1,500	6,330	379	38	15	740	1,200
17	28	52	470	470	1,960	52	9	5	42	430
8	120	408	1,760	1,300	7,580	407	41	15	720	1,000
16	45	115	520	600	3,990	115	5	3	26	490
7	90	331	1,340	970	4,450	331	37	13	500	840
15	92	361	580	710	6,580	361	4	4	23	560
6	90	321	500	470	2,490	318	51	16	260	240
14	92	390	360	520	3,440	362	5	6	18	340
5	140	461	470	340	2,530	456	47	19	220	250
13	110	420	230	190	1,240	420	5	7	12	220
4	110	403	260	210	1,300	400	74	19	190	68
3	170	631	1,250	1,200	5,210	625	41	17	510	740
2	100	423	300	250	1,290	421	63	18	190	110
12	170	575	850	500	3,770	575	1	4	9	840

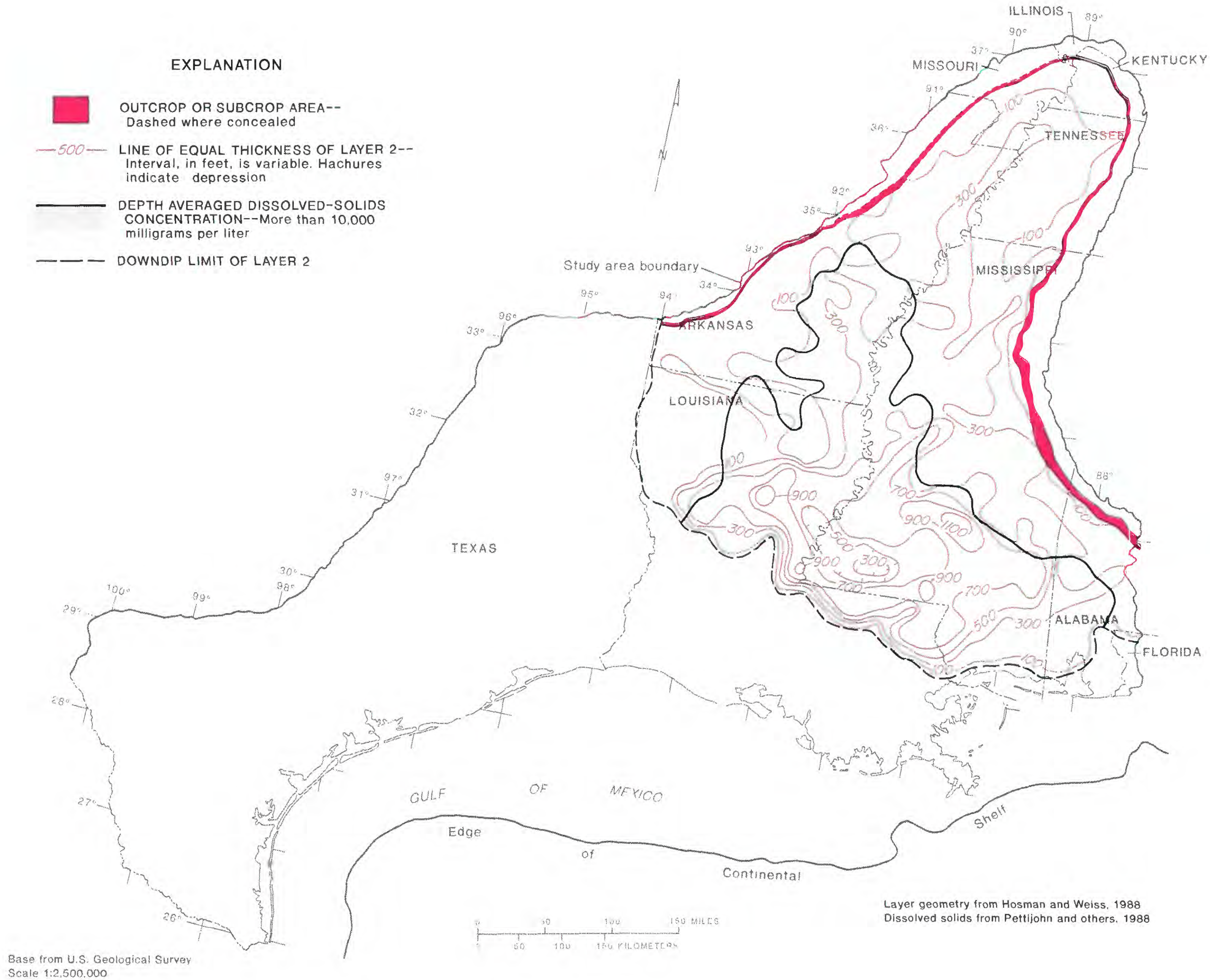


Figure 19a.--Thickness and extent of model layer 2 (lower Wilcox aquifer).

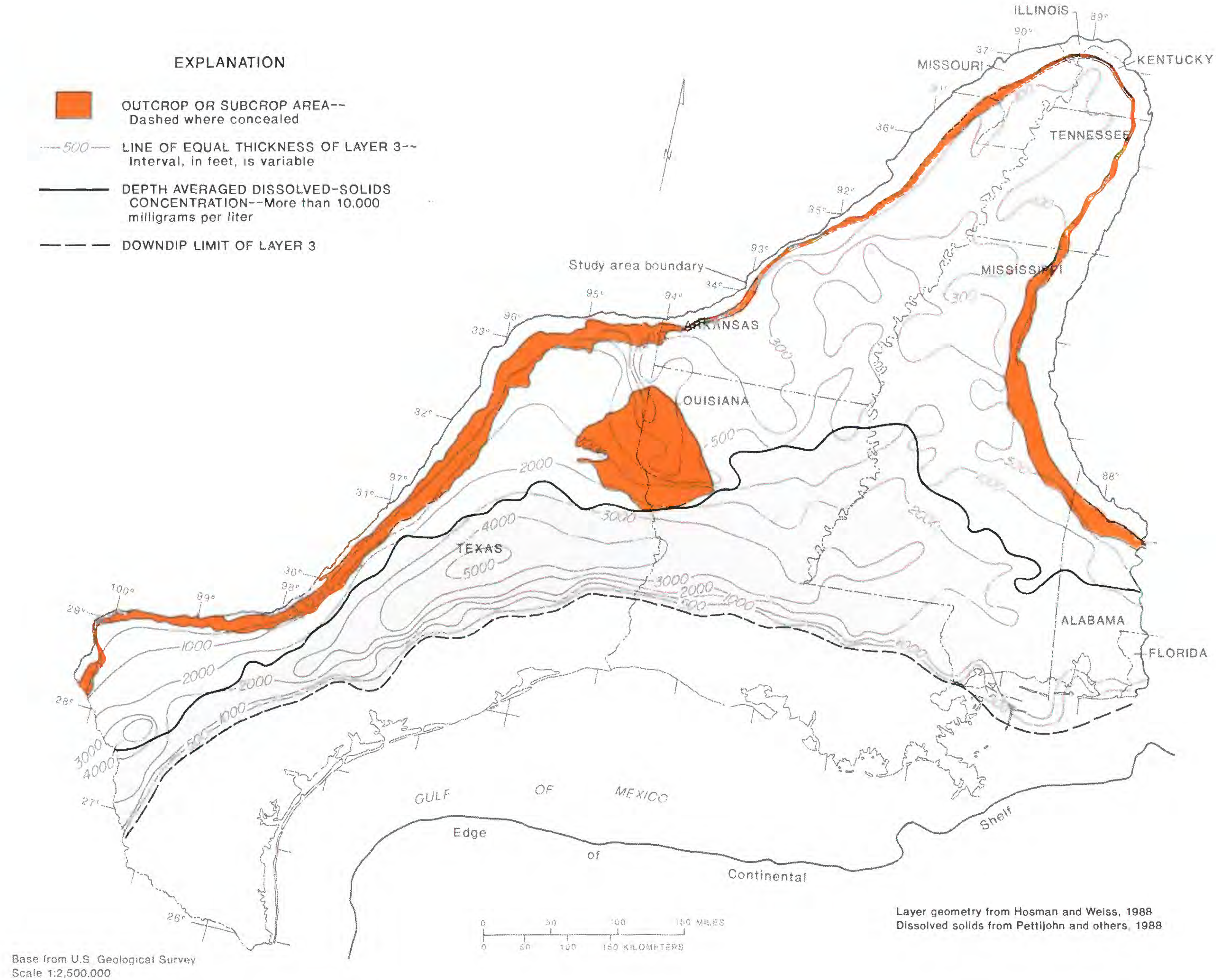


Figure 19b.--Thickness and extent of model layer 3 (middle Wilcox aquifer).

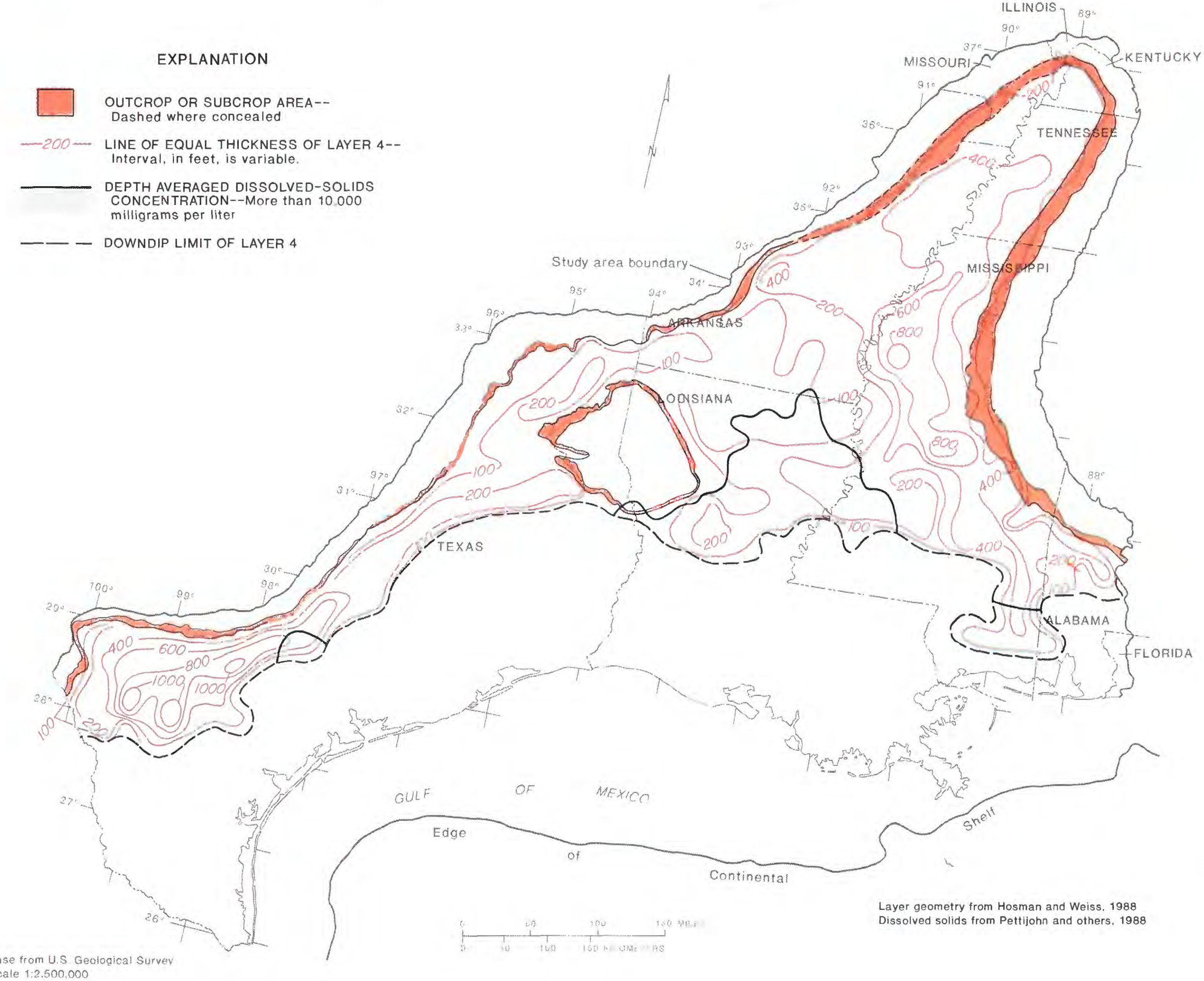


Figure 19c.--Thickness and extent of model layer 4 (lower Claiborne-upper Wilcox aquifer).

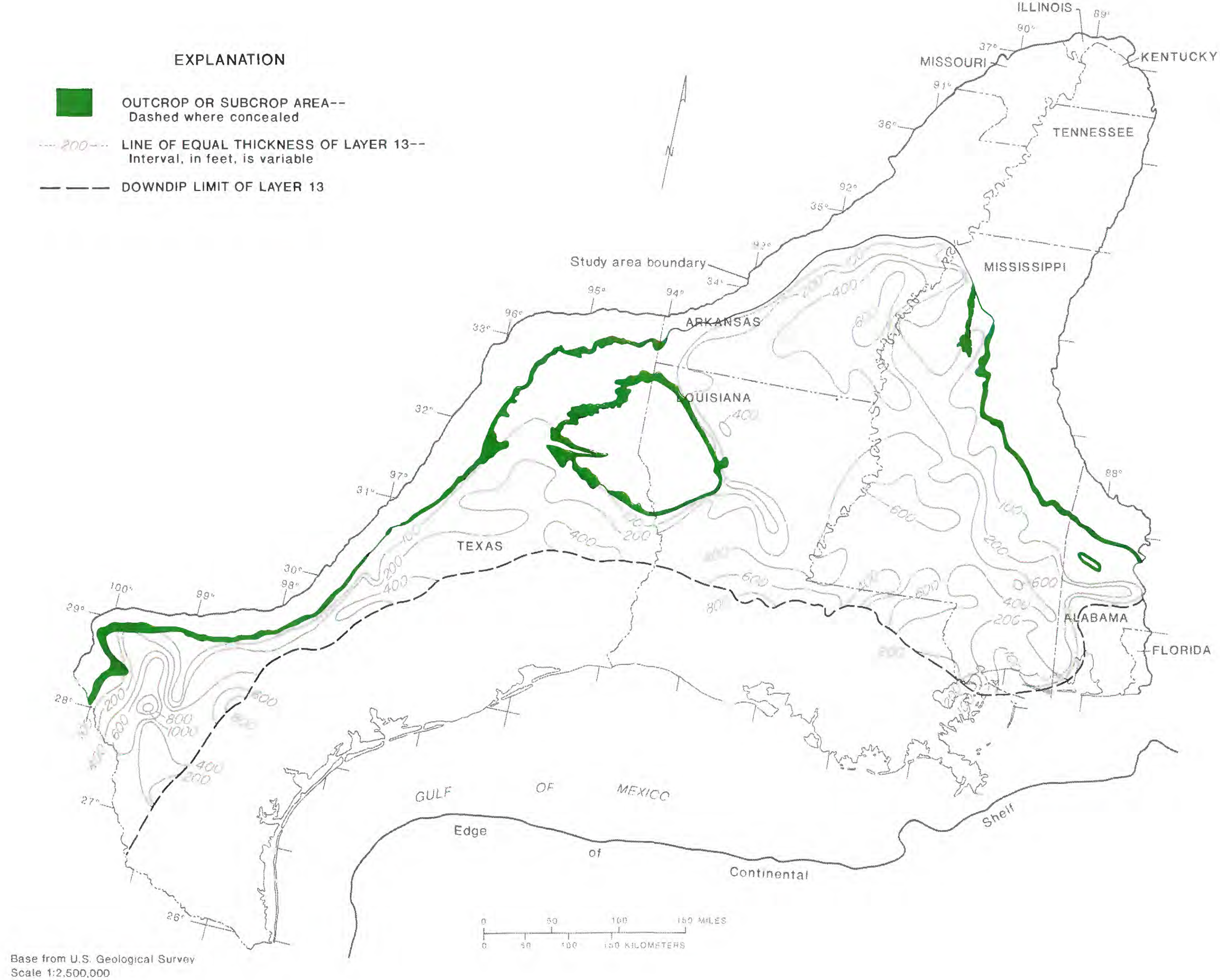


Figure 19d.--Thickness and extent of model layer 13 (lower Claiborne confining unit).

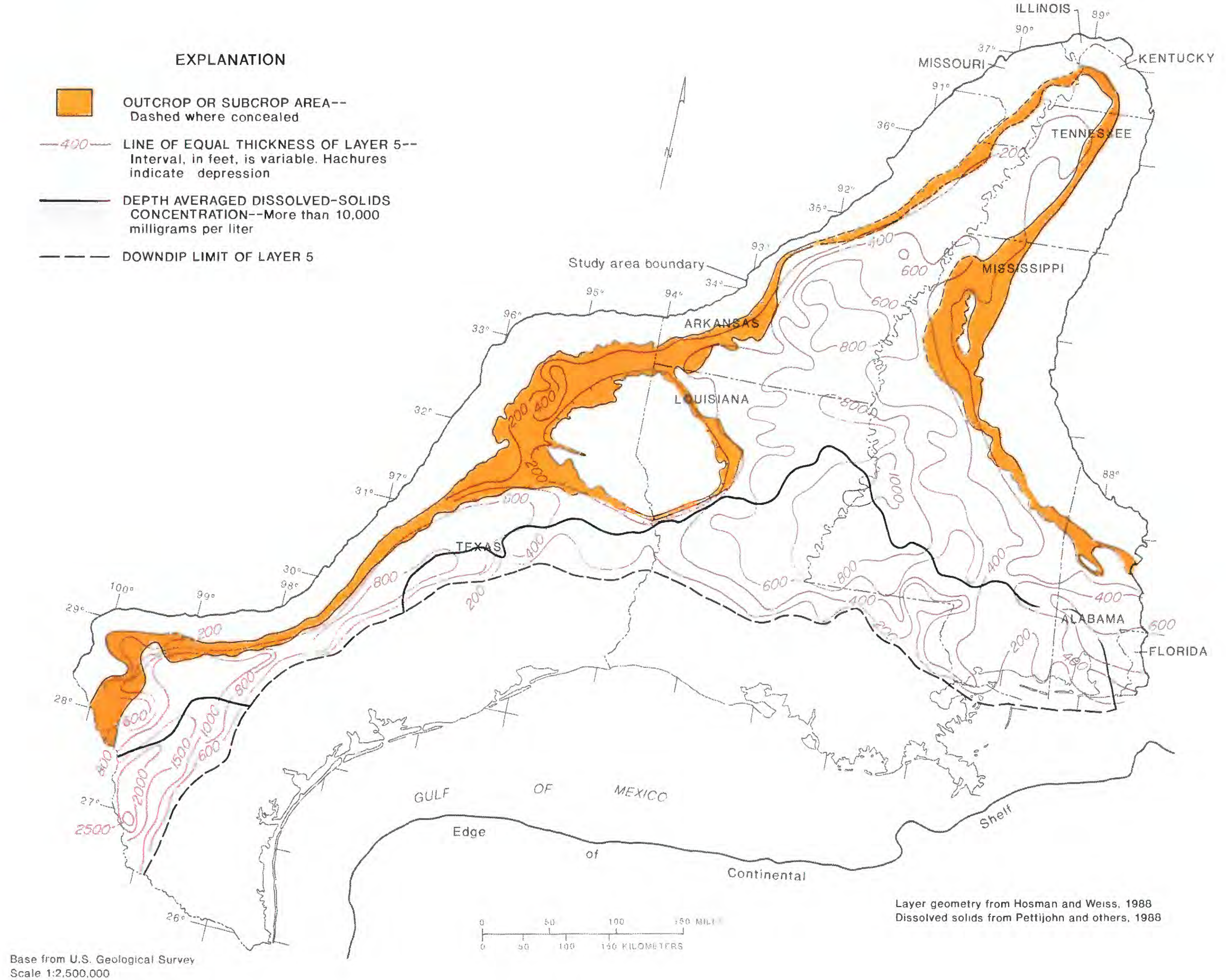


Figure 19e.--Thickness and extent of model layer 5 (middle Claiborne aquifer).

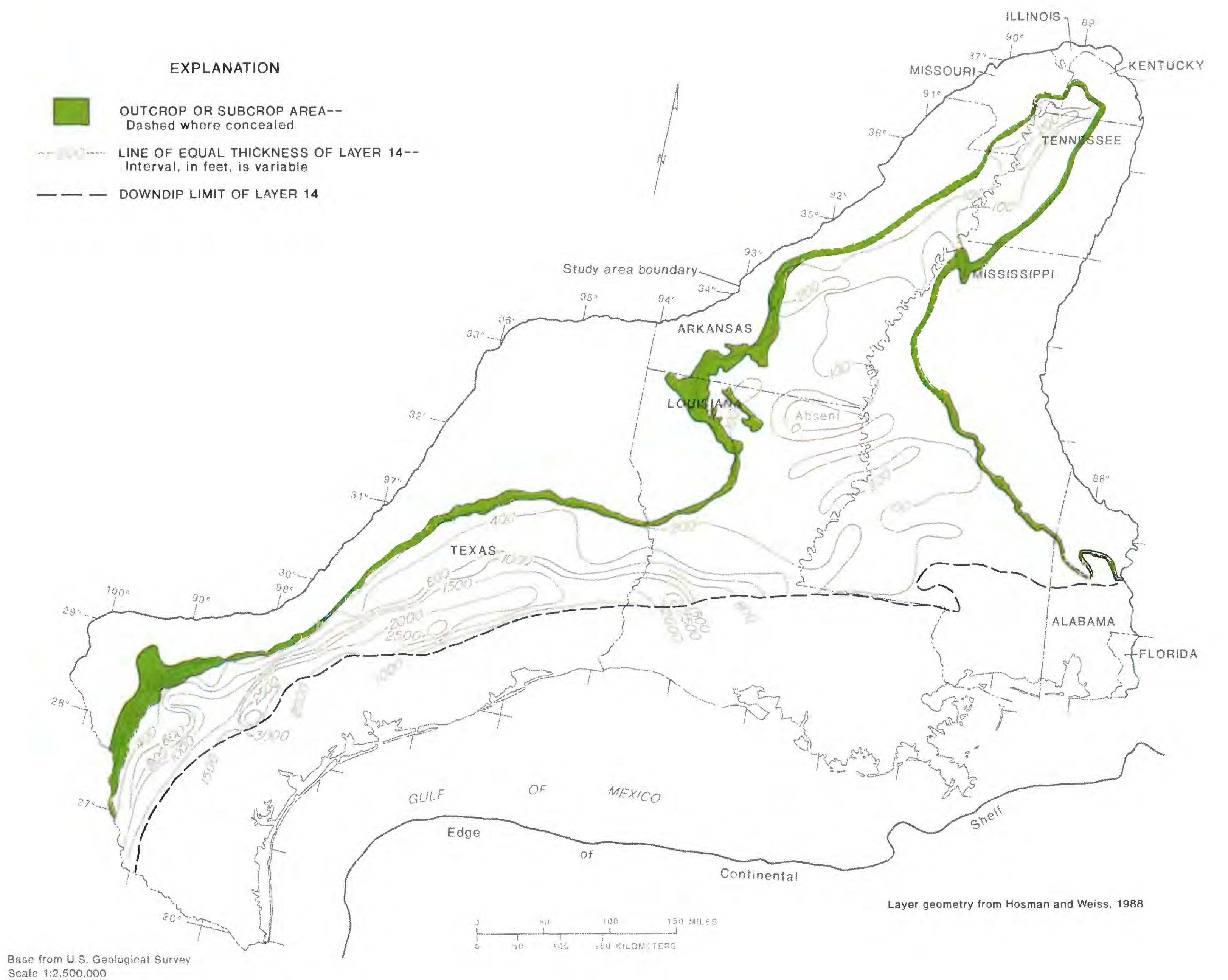


Figure 19f.--Thickness and extent of model layer 14 (middle Claiborne confining unit).

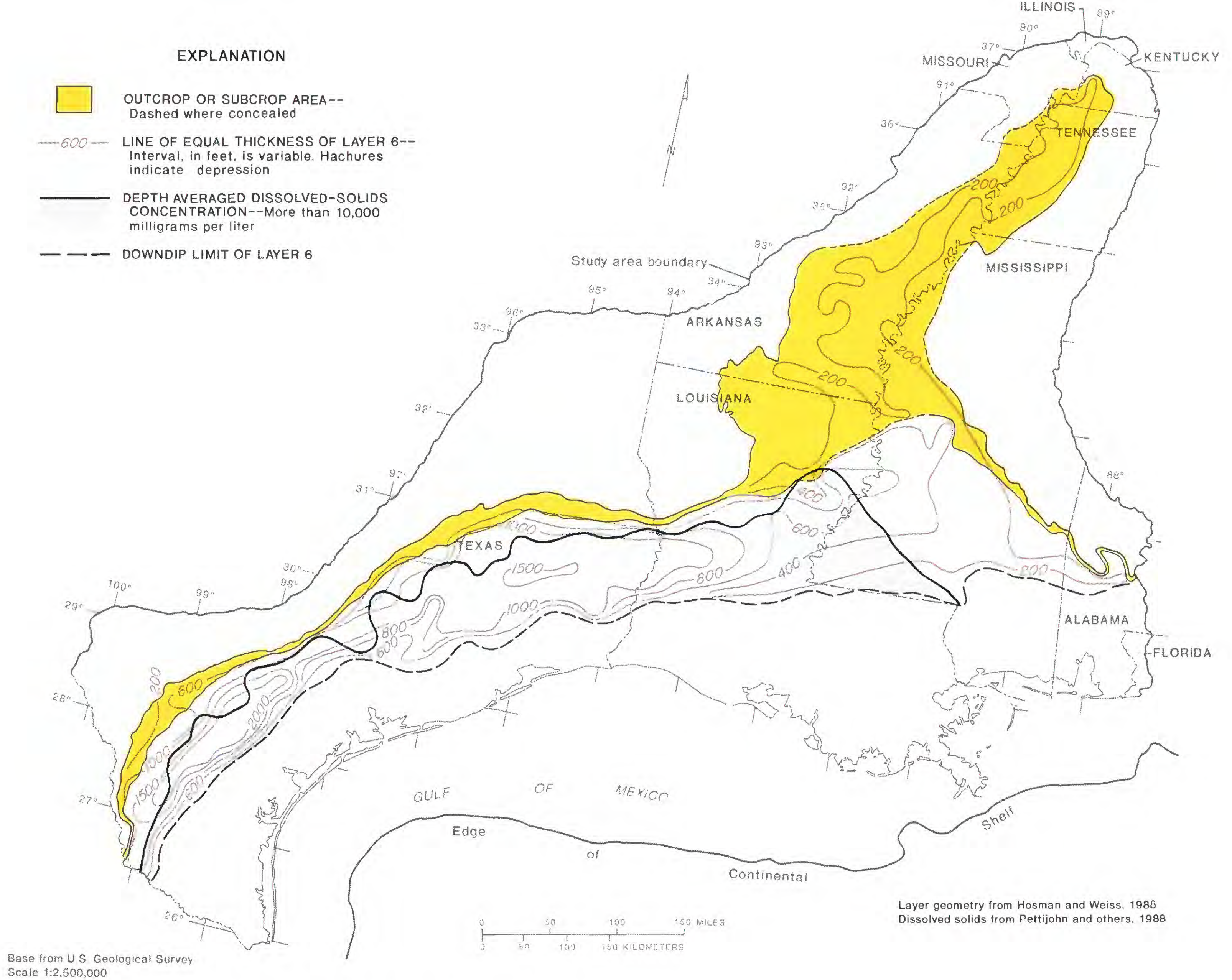


Figure 19g.--Thickness and extent of model layer 6 (middle Claiborne aquifer).

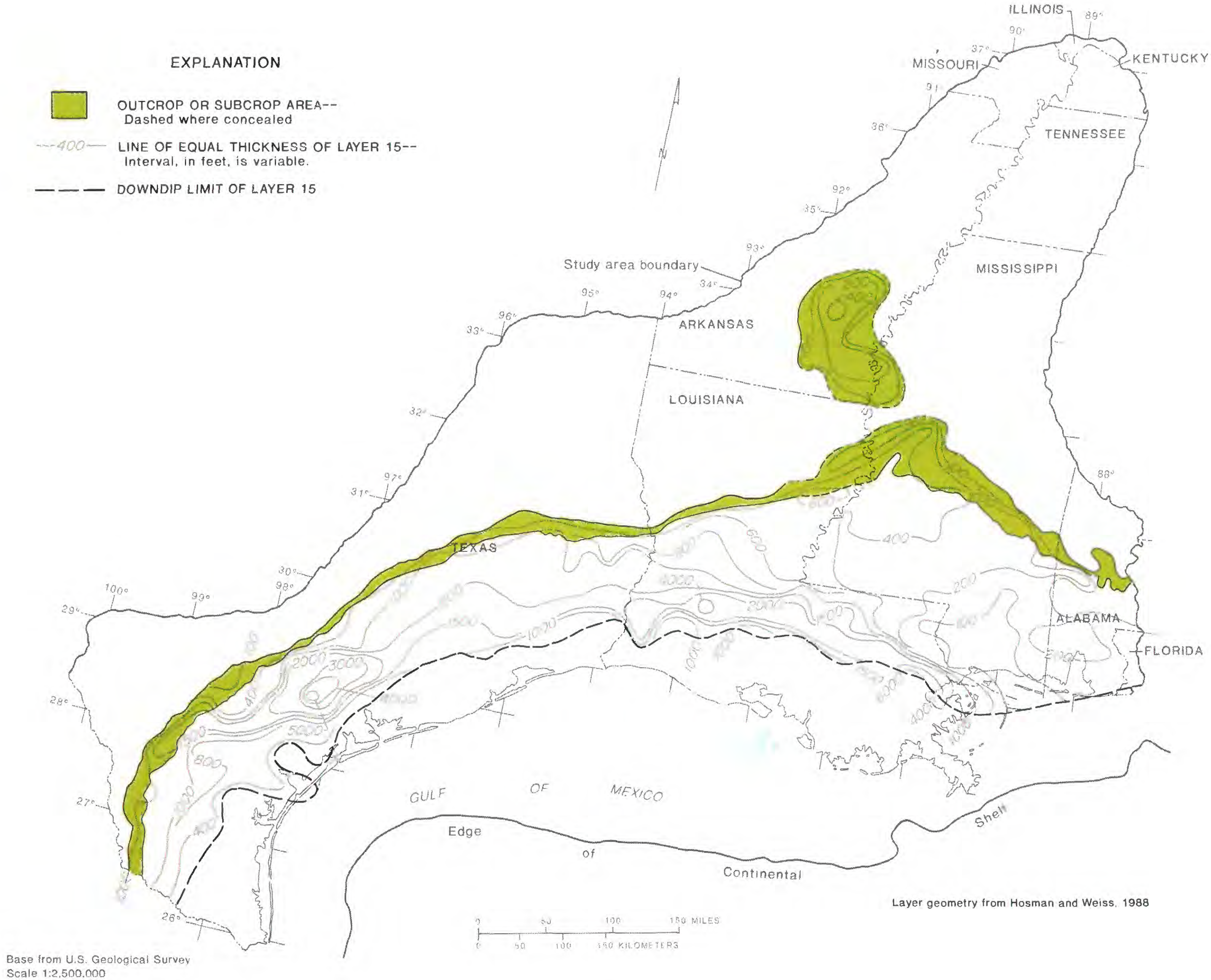


Figure 19h.--Thickness and extent of model layer 15 (Vicksburg-Jackson confining unit).

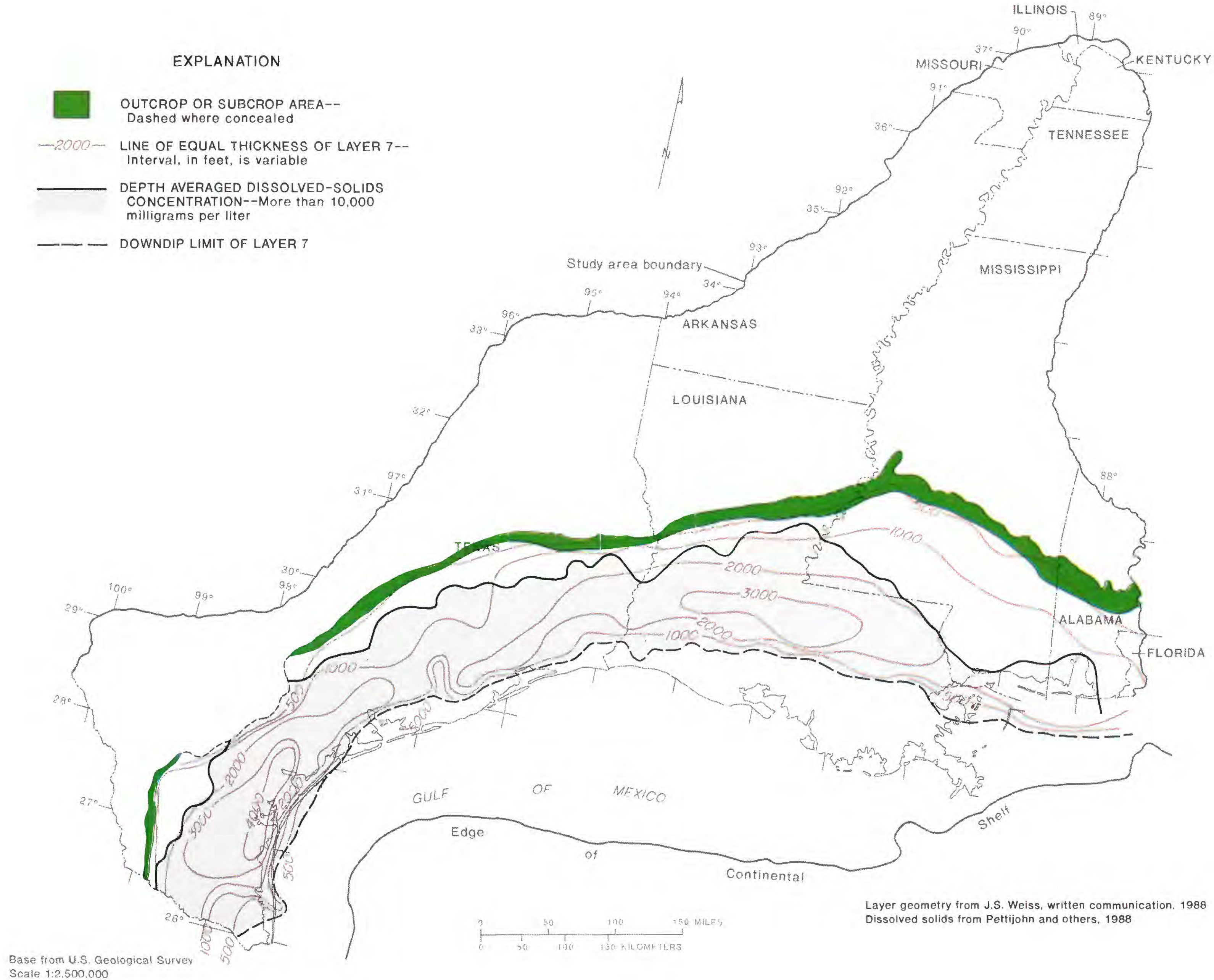


Figure 19i.--Thickness and extent of model layer 7 (permeable zone E, lower Miocene-upper Oligocene deposits).

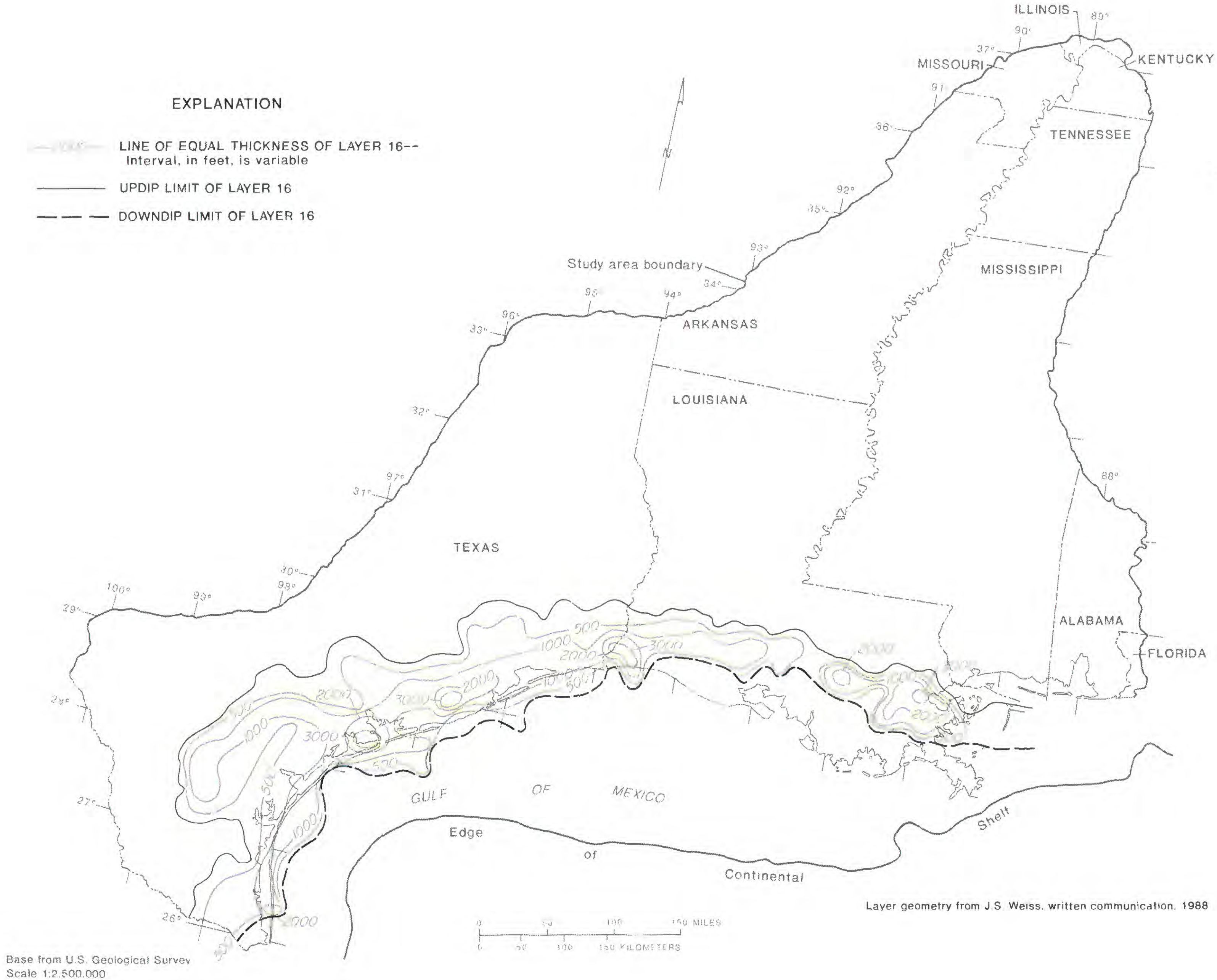


Figure 19j.--Thickness and extent of model layer 16 (zone E confining unit).

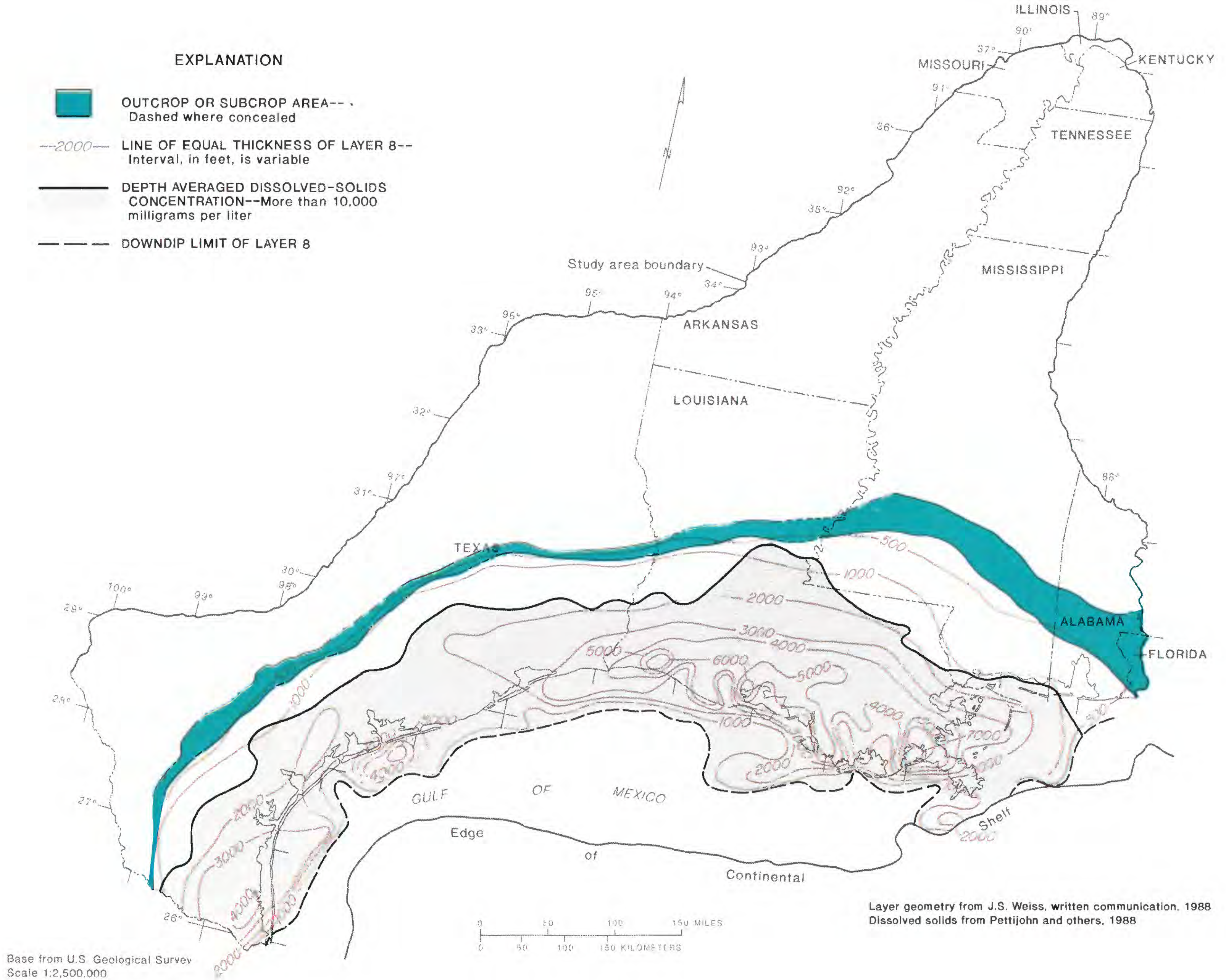


Figure 19k.--Thickness and extent of model layer 8 (permeable zone D, middle Miocene deposits).

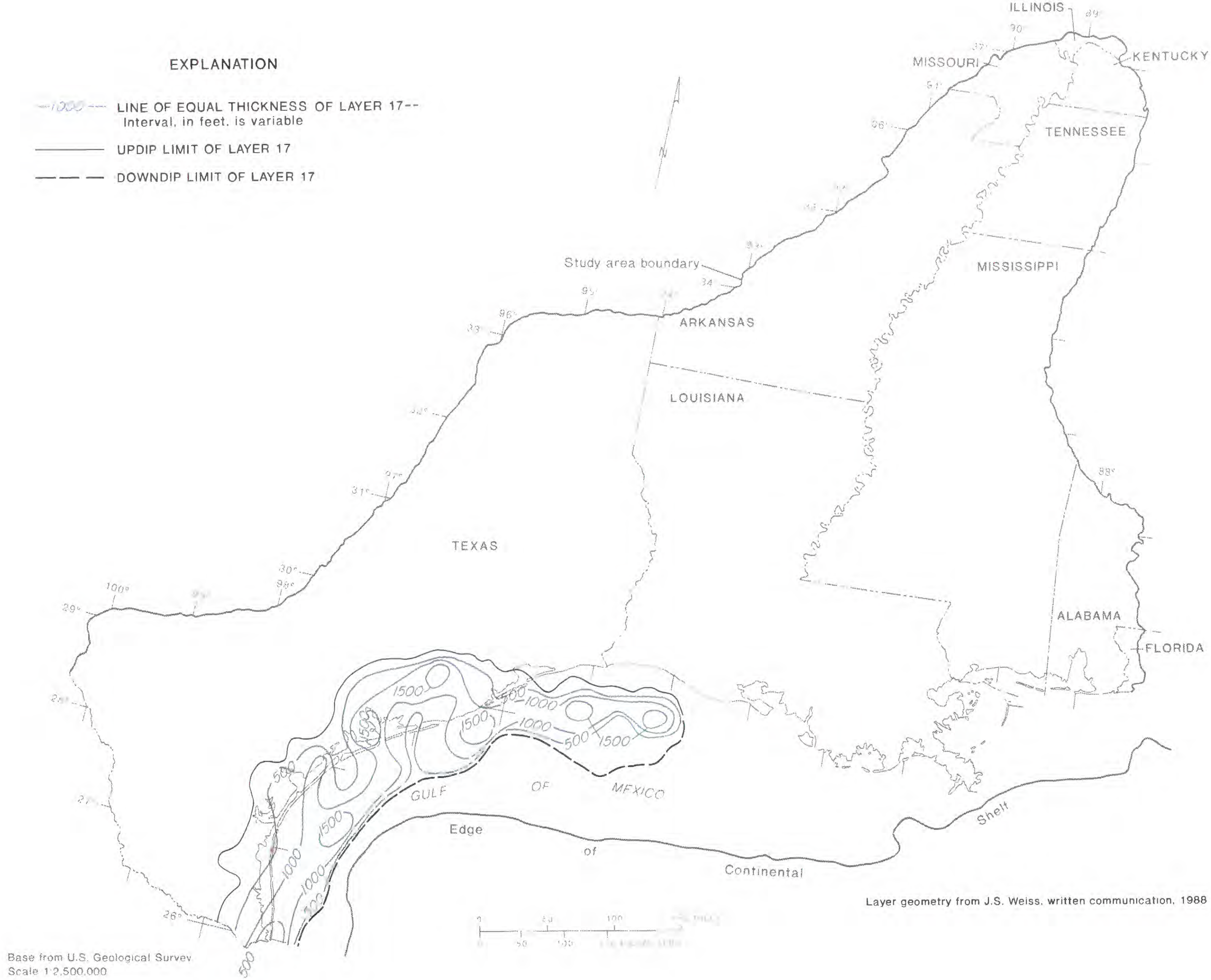


Figure 19l.--Thickness and extent of model layer 17 (zone D, confining unit).

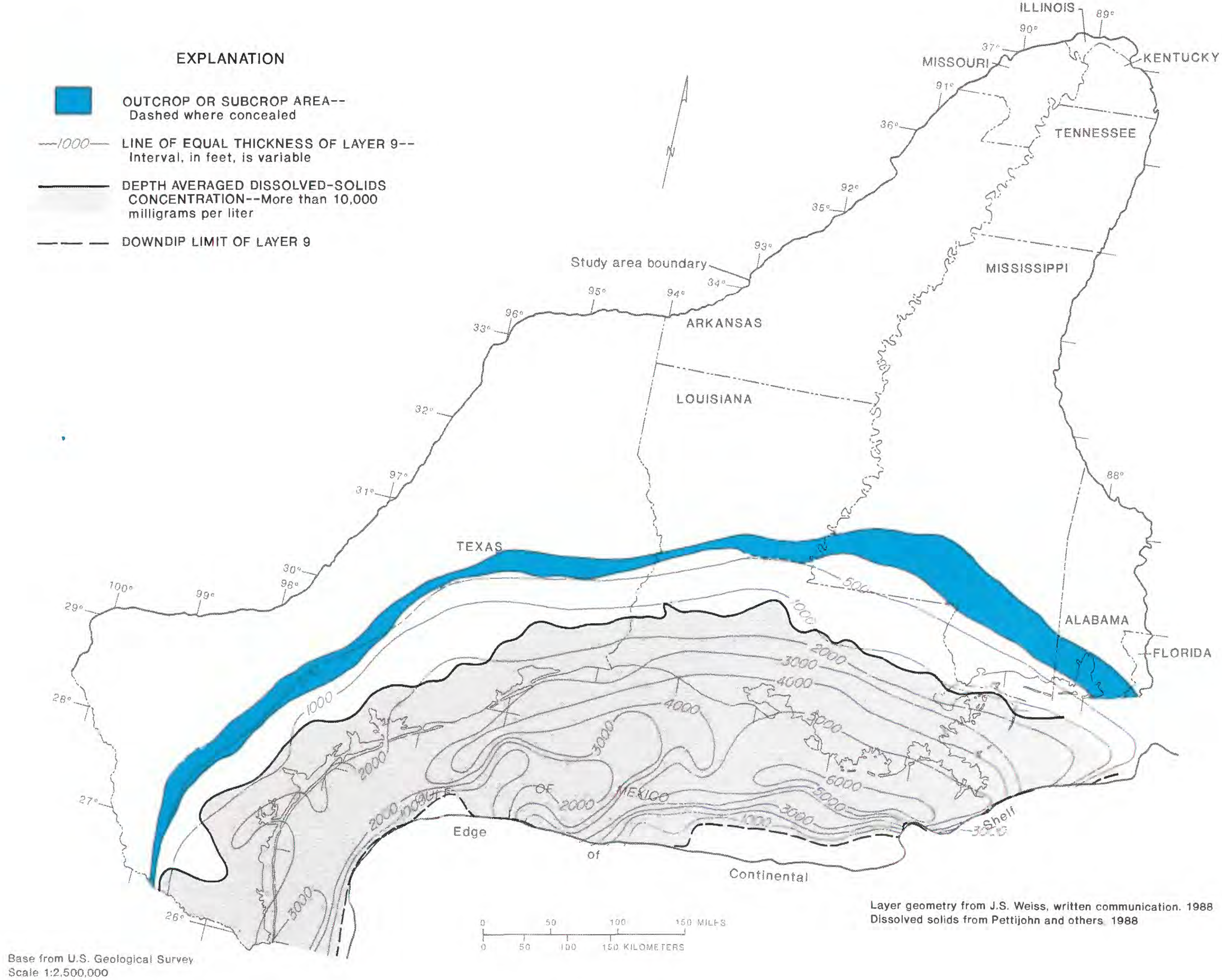


Figure 19m.--Thickness and extent of model layer 9 (permeable zone C, lower Pliocene-upper Miocene deposits).

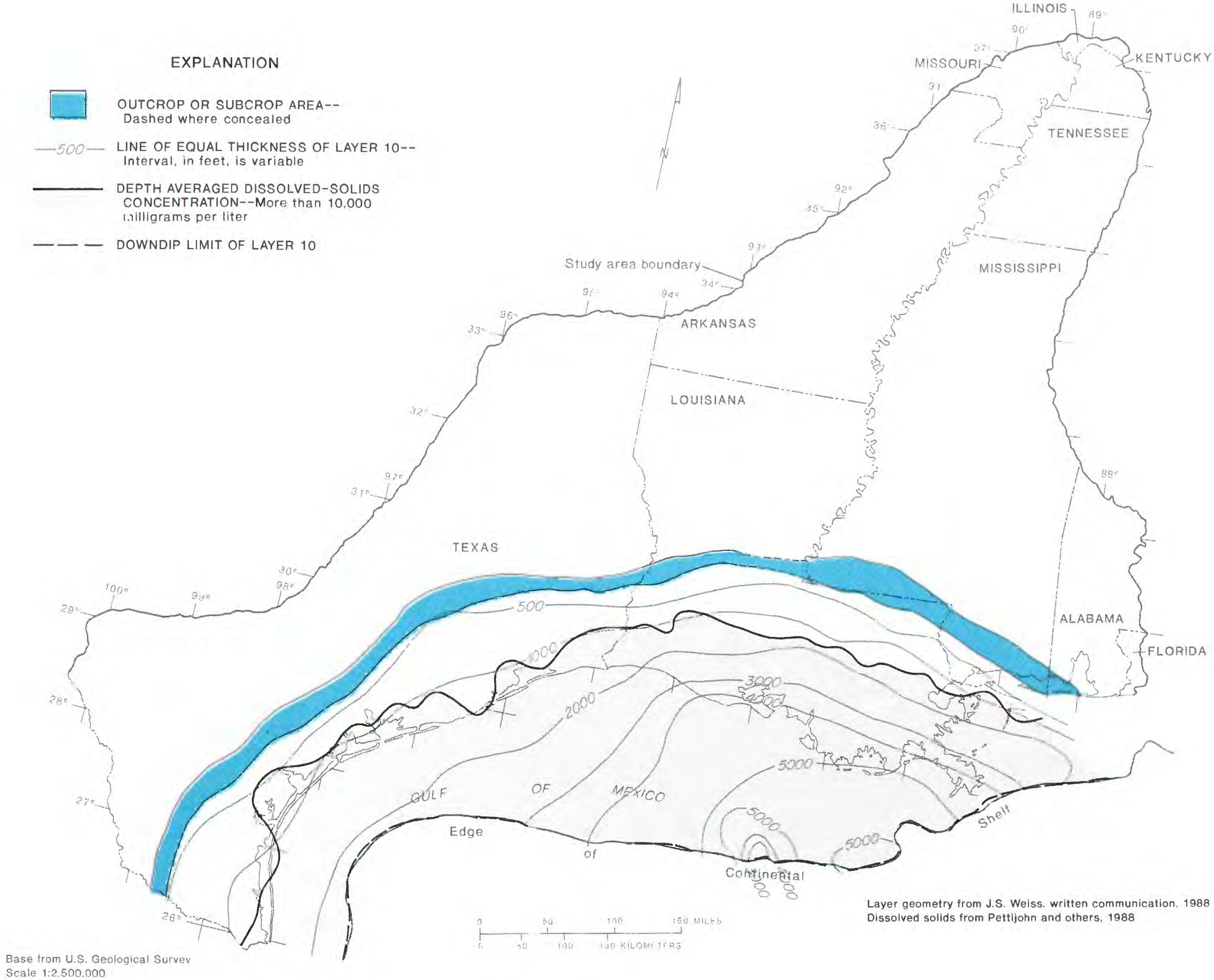


Figure 19n.--Thickness and extent of model layer 10 (permeable zone B, lower Pleistocene -upper Pliocene deposits).

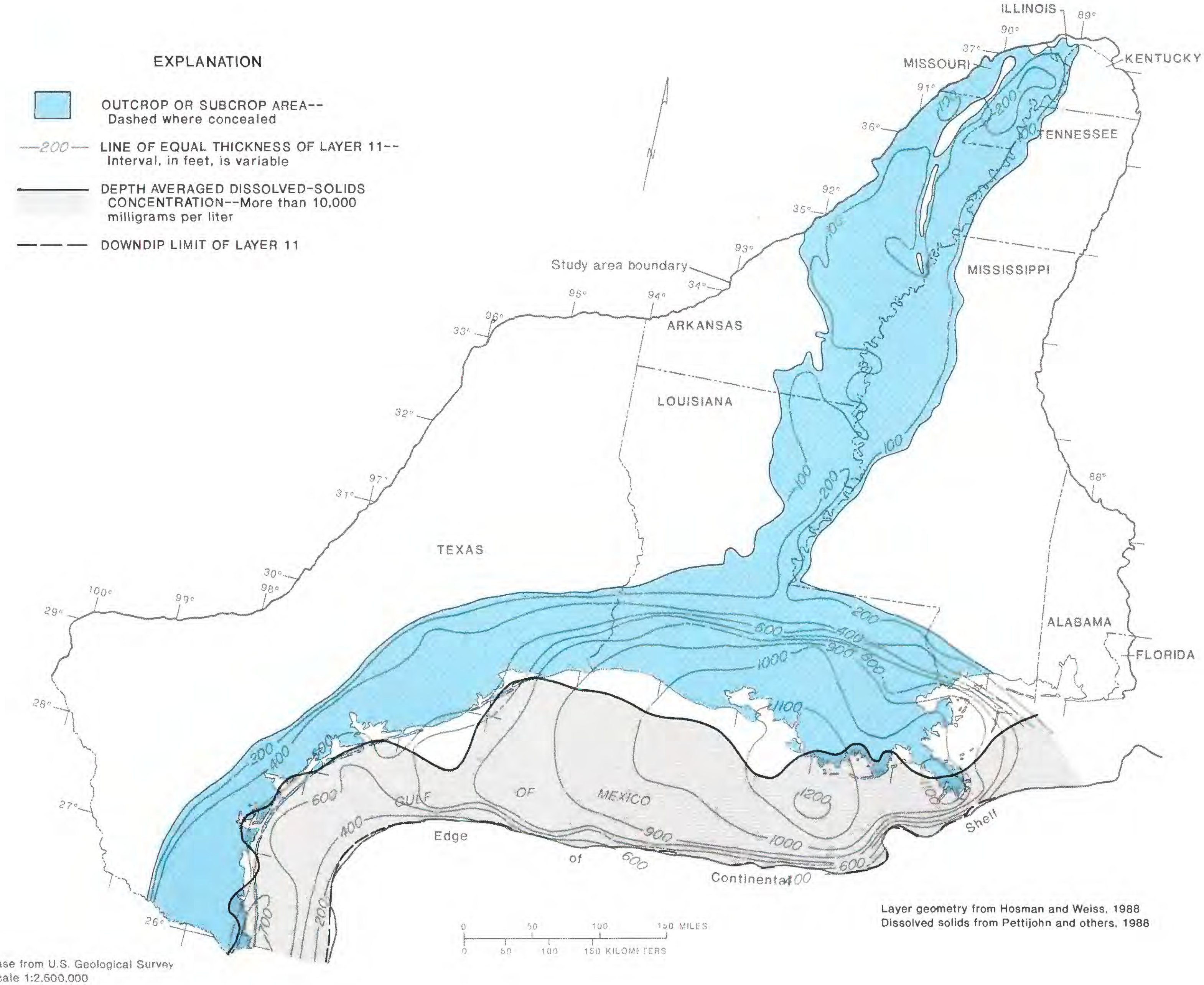


Figure 19o.--Thickness and extent of model layer 11 (Mississippi River Valley alluvial aquifer and permeable zone A, Holocene-upper Plesitocene deposits).

conductivity would be equal to the arithmetic mean of the conductivities of the individual beds. Horizontal ground-water flow on a regional scale represents a mixture of these two types of flow paths, so the effective conductivity should be between the values given by the two methods. The geometric mean (numerically equal to the anti-log of the arithmetic mean of log K) is a more representative effective average than an arithmetic mean of the conductivities, which is biased towards the larger values (Fogg, 1986b, p. 206).

Horizontal hydraulic conductivities were estimated from more than 1,500 aquifer tests and from specific capacity data from nearly 6,000 wells (D.E. Prudic, U.S. Geological Survey, written commun., 1988). The results of the estimates of hydraulic conductivity are summarized by layer in figure 20. Martin and Early (1987) previously estimated hydraulic conductivities from nearly 500 of these aquifer tests in Louisiana. Conductivities were estimated in this study by dividing transmissivity values by: (1) The screened interval for specific capacity data and for single well aquifer tests, (2) the thickness of the sand bed tapped by the well for aquifer tests which included observation wells, and (3) twice the screened interval for multi-well aquifer tests when the thickness of sand beds was unknown. The factor 2.0 is the average ratio of the sand thickness to screen interval for tests where both were known. Transmissivity was estimated from specific capacity data using the method described by Theis and others (1963, p. 331-341 and Prudic, written commun., 1988). Estimates of transmissivity from specific capacity data were first calculated from wells where both estimates of transmissivity from aquifer tests and specific capacity data were available. The estimates of transmissivity from specific capacity data were then compared to estimates of transmissivity from aquifer tests. The results suggest the estimates from specific capacity are generally less than estimates from aquifer tests by a factor of 1.43, although there is considerable scatter. The factor is slightly more than the range of 1 ± 0.3 suggested by Theis (1963, p. 335) for wells with a 1 ft diameter. All transmissivities estimated from specific capacity data were increased by the factor of 1.43.

The effect of different methods of calculating an average hydraulic conductivity for all of the aquifer tests within a layer is shown in figure 20. There is a much greater spread between the harmonic mean and the arithmetic mean for a unit with a high sand percentage like layer 11 than for a low sand percentage unit like layer 3 because a few small values force the harmonic mean to be much smaller and a few large values force the arithmetic mean to be much larger. There may be an additional bias introduced when comparing horizontal hydraulic conductivity (K_h) values from aquifer tests with K_h values used in the regional model due to the fact that wells are selectively completed in the sand bed which produces the most water. These factors generally lead to choosing K_h for the regional model generally slightly smaller than the lower 99 percent confidence limit for the geometric mean, but larger than the harmonic mean. Figure 20 shows the values selected for use during calibration and in the preliminary simulations.

It does not appear that any kind of trend can be mapped due to the large local variations. However, there are large and statistically significant differences between values averaged over geographic areas when compared to average values for adjacent areas. Surprisingly, when dividing the study area into nine areas, more variation was found between average values for an area ignoring layer subdivisions than was found between average values for layers ignoring area divisions. The estimated hydraulic conductivity increased about one order of magnitude from southwest to northeast following along strike approximately parallel to the coast. This areal variation is not included in these preliminary simulations, but will be considered in future simulations.

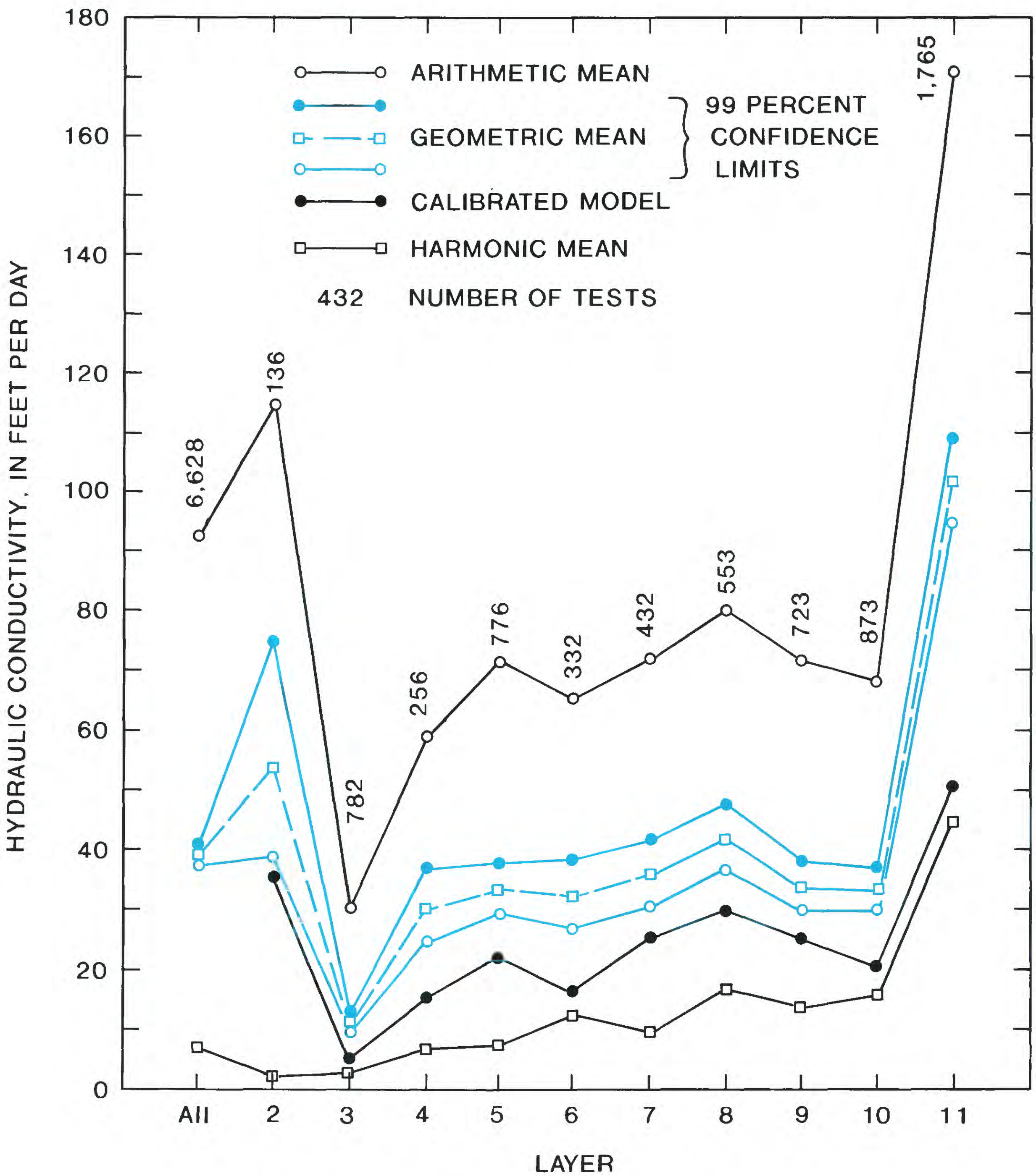


Figure 20.--Estimated horizontal hydraulic conductivity of permeable model layers, based on aquifer test data. Five different summary statistics are shown: arithmetic mean, geometric mean along with the upper and lower 99 percent confidence limits of that mean, harmonic mean (defined in text), as well as the value chosen for the regional model.

As discussed in the concepts section, it is assumed in this preliminary analysis that hydraulic conductivity does not vary as a function of depth. It is assumed that the increase in conductivity with depth due to temperature's effect on viscosity is approximately equal to the decrease in conductivity with depth due to compaction. Overall, conductivity probably does decrease with depth, however, it is not included in this analysis.

Changes in the altitude of the water table also affect the saturated thickness, and hence, the transmissivity of the uppermost aquifer layer. To simplify the model for this preliminary analysis, the effect of the change in saturated thickness of the uppermost aquifer layer due to water-table change is not modeled. The saturated thickness changes less than 20 percent except in a few areas, especially the cones of water level decline in the Mississippi River Valley alluvial aquifer (MRVAA) in central Arkansas. In these two places, one between the White River and Crowley's Ridge, and the other between the Arkansas and White Rivers, the change in saturated thickness is substantial (as much as 80 percent), although the area is small relative to the scale of this study (about 10 model blocks). Simulation of the alluvial aquifer at this scale is quite difficult because of the many large rivers crossing it. For a more detailed and accurate simulation of the MRVAA, see Ackerman (1989).

Vertical Leakance

Resistance to vertical flow is controlled by the ratio, called leakance, of the vertical hydraulic conductivity to the thickness of sediments through which vertical flow must occur. When a confining unit is simulated between two aquifers, the effective leakance is equal to the harmonic mean (because flow must occur in series) of the leakances of the lower half of the upper aquifer and the upper half of the lower aquifer and leakance of the confining unit. Kuiper's model (1985) calculates the total leakance between the centers of all adjacent model layers using the appropriate thicknesses and vertical hydraulic conductivities. The model allows entering either the ratio of the horizontal to vertical conductivity (anisotropy), or a specified vertical conductivity that may be unrelated to the horizontal conductivity. The latter method was used for more flexibility.

Vertical hydraulic conductivity is probably the most difficult characteristic of an aquifer system to measure or estimate, yet it is generally a very sensitive and significant parameter for regional models. The few good aquifer tests which can be used to estimate vertical conductivity are of little use because vertical conductivity usually varies significantly over short distances. Laboratory test results are of little help because fractures or leakier areas "short-circuit" the vertical flow path, causing an effective leakance for an area to be several orders of magnitude higher than that estimated for a typical sample in the lab. Regionally representative values are best determined from model calibration in an area where pumping is from an aquifer which is at least partially confined, and can be analyzed as if it were a huge aquifer test. The initial constant values of vertical conductivity for each layer were chosen based on previous investigations and judgment, then adjusted during calibration, and are shown in table 4. No attempt was made to vary these values areally because of the preliminary stage of these simulations.

Density

Density of the water in the aquifer systems was estimated based on estimates of dissolved solids, temperature, and the hydrostatic pressure at a depth corresponding to the middle of the model block. Dissolved-solids concentrations were estimated from

Table 4.--*Simulated horizontal and vertical hydraulic conductivities, (K) in feet per day, and their ratios for model layers. Horizontal K is the equivalent K for the coarse-grained part of the aquifers. Vertical K represents the confining units and fine-grained part of the aquifers*

Model layer	K_h Horizontal	K_v vertical		Anisotropy ratio K_h/K_v
		Confining unit	Fine-grained part of aquifer	
Source-sink			1.00×10^{-4}	---
11	50		0.30×10^{-2}	17,000
10	20		0.30×10^{-3}	67,000
9	25		0.47×10^{-3}	53,000
17		0.20×10^{-4}		
8	30		0.60×10^{-3}	50,000
16		0.20×10^{-3}		
7	25		1.20×10^{-4}	210,000
15		0.78×10^{-5}		
6	16		0.30×10^{-1}	530
14		0.20×10^{-4}		
5	22		0.30×10^{-3}	73,000
13		0.20×10^{-3}		
4	15		0.30×10^{-3}	50,000
3	5		0.10×10^{-3}	50,000
2	35		0.50×10^{-1}	700

water resistivities obtained from the spontaneous potential curve of 989 electric logs as described by Weiss (1987), using an algorithm of Bateman and Konen (1977) based on the original work by Gondouin and others (1957). The dissolved-solids concentration for a layer was assumed to be equal to the average of the values for each sand bed thicker than 20 ft. Temperatures were interpolated for the middle of a layer from bottom hole temperatures recorded on the electric logs on successive logging runs as the hole was deepened. Arrays of dissolved solids, temperature, and pressure were interpolated from the point log data as was previously described for thickness and sand percentage. These maps are presented in Pettijohn and others (1988). A computer program developed by Emanuel Weiss (1982) and modified by Kontis and Mandle (1988) was used to estimate ground-water density based on the equations and coefficients developed by Potter and Brown (1977). These equations require an iterative solution, because one of the independent variables in the equation, molality, is itself dependent on the density which is in turn dependent on the concentration expressed in mg/L.

A linear multiple regression using data for 15,200 5-mile model blocks, was used to estimate density within a standard error of estimate of 0.0020 g/cm^3 about the value estimated by the iterative technique (D.E. Prudic and T.A. Williams, U.S. Geological Survey, written commun., 1988), giving the equation:

$$DENGW = 0.000648\ TDS - 0.000368\ TEMPC + 0.0000491\ PBAR + 1.00472,$$

where,

DENGW = density of the ground water at the pressure and temperature in the aquifer, in g/cm^3

TDS = total dissolved solids, in mg/L,

TEMPC = temperature, in degrees C, and

PBAR = pressure, in bars.

Pumpage

Pumpage was estimated for each block for 1980 (Mesko and others, in press). Pumpage for 1960, 1965, 1970, 1975, and 1985 was estimated using the 1980 distribution of pumpage by block and the published total pumpage from each county for the respective year, assuming that the distribution of pumpage within each county has been constant through time. Two categories of ground-water pumpage data were distinguished:

1. Point pumpage, usually for public supply and industry, where little water can return to the aquifer system.

2. Areally dispersed pumpage for irrigation, fish farm, stock, and rural domestic uses, where there is a potential for a significant amount of water to seep back to the aquifer system.

Point-pumpage data for 1980 was tabulated for each of about 5,000 public supply and industrial wells, along with the wells's 5-mile block location and depth. These data are much more detailed for Harris and Galveston Counties in Texas, where law requires each of about 3,000 wells greater than 6 inches in diameter be metered and well discharge reported to the Harris-Galveston Coastal Subsidence District (obtained on computer tape from R.J. Neighbors, Harris-Galveston Coastal Subsidence District, written commun., 1985).

Pumpage from the wells in the Mississippi embayment aquifer system and Texas coastal uplands aquifer system was manually assigned to the appropriate model layer. The wells in the coastal lowlands aquifer system were assigned using a FORTRAN program which used all of the 5-mile grid data on layer thicknesses and elevations of the land surface. The altitude of the top of each layer at the well location was interpolated from the four adjacent 5-mile block centers, weighted by the distance from the well to each block center. The correct layer was found by checking the layer top altitudes, starting from the bottom, until one was found that was higher than the altitude of the bottom of the well. If the well bottom appeared to be in a confining unit, the pumpage was assigned to whichever adjacent aquifer layer was closer (vertically). This occurs in a few places because the regional confining units can contain small local aquifers which may contain wells and also because of errors interpolating the correct altitude of the tops of the layers at the well locations.

Estimation of areally dispersed pumpage is generally subject to greater errors because there are many more wells, most of which are not metered. Published data was used where it was available, generally as totals by county (references given by Mesko and others, in press). Estimates for agricultural use are generally based on irrigated areas, types of crops, and typical crop water use estimates. Domestic use estimates are generally based on the population not served by public systems and per capita use estimates. County total pumpage was apportioned to 5-mile blocks using estimates of the distribution to each block and layer in the county based on well records, land area proportions, and local knowledge of water use patterns.

Hydraulic Head

Estimating hydraulic-head values to be used in or compared with the regional flow model using head measurements in wells is complex: model analysis is conducted using discrete blocks of aquifer over discrete time intervals, requiring discretization of the water levels from wells over four dimensions, three in space and one in time. Horizontal discretization is complicated in places because the regional hydraulic gradient is relatively flat, but local gradients can be steep because of local topographic variation. In other places the estimation is complicated by steep depressions in head due to pumping which are areally small relative to the regional model block size. Vertical discretization is important but difficult because in some places layers are quite thick, and there is significant head difference vertically within the layer. Temporal discretization is likewise difficult because wells are not measured on the same days of the year or even every year as there are changes in the monitoring network design and operation.

Over one-half million water-level measurements taken from nearly 20,000 wells stored in the U.S. Geological Survey's WATSTORE file and the Texas Natural Resources Information System files (T.A. Williams, U.S. Geological Survey, written commun., 1987) were combined into one data base. A significant amount of error screening was done by several automatic methods in order to delete data which was obviously in error. In addition to internal checks for problems such as missing or impossible data values, locations were checked by computer program to test if they occurred in or near the county for which they were coded. About 1 percent of the wells which provided about 0.3 percent of the water-level measurements were ignored because of this location check. Land-surface altitudes were checked to test if they were within a reasonable range of the 5-mile land-surface altitude interpolated for the well location. About 1.6 percent of the wells which provided about 1 percent of the measurements were ignored because of this land-surface altitude check. The data were also plotted as hydrographs with many wells plotted on a page to check for extreme outliers.

Estimation of Predevelopment Water-Table Altitude

Average predevelopment water-table altitudes for 5-mile blocks were estimated by subtracting the estimated depth-to-water from the land-surface altitude calculated from detailed digital data (Williams and Williamson, 1989). Contouring measured water-table altitudes by hand was tried, but not used due to the large spacing of data relative to small-scale variations in land-surface altitude. The mean predevelopment depth-to-water in 6,825 wells less than 150 ft deep in the study area was 26 ft and the standard deviation was 20 ft. An even distribution would provide an average coverage of about one well per 5-mile block, however, the wells are not evenly distributed (Williams and Williamson, 1989, fig. 5). The land-surface altitude in the study area varies from sea level to over 800 ft. Therefore, most of the variation in predevelopment water-table altitude is a function of the variation in land-surface altitude, rather than in the depth-to-water. This fact and the availability of much more detailed data on land-surface altitude led to the abandonment of the manual mapping method.

Land-surface altitude was estimated using a digital data base which contained altitude data for every 30 seconds of latitude and longitude, originally digitized by the Defense Mapping Agency from 1:250,000 scale topographic maps (Godson, 1981). More than 90 land-surface altitude points per 5-mile grid block were averaged to develop a data file for use in this study. Comparison of the digitally averaged land-surface altitudes with water-table altitudes picked from a regional predevelopment water-table map prepared by the traditional method showed that in about one-third of the blocks the water-table was above land surface (Williams and Williamson, 1989, figs. 7 and 8). This is due to the inability of the sparse well spacing used in the manual method to account for small-scale variability in land-surface altitude.

A linear regression of depth-to-water as a function of land-surface altitude using the well data mentioned above yielded the equation: water-table altitude = 0.898 times land-surface altitude. This equation had a coefficient of determination (r-squared) of 0.992. The regression was improved using a measure of local topographic variation as a second independent variable and splitting the area up into five subareas. The multiple regression equations for the five subareas (Williams and Williamson, 1989, table 1) were used to estimate the predevelopment water-table altitude over the entire study area (fig. 21).

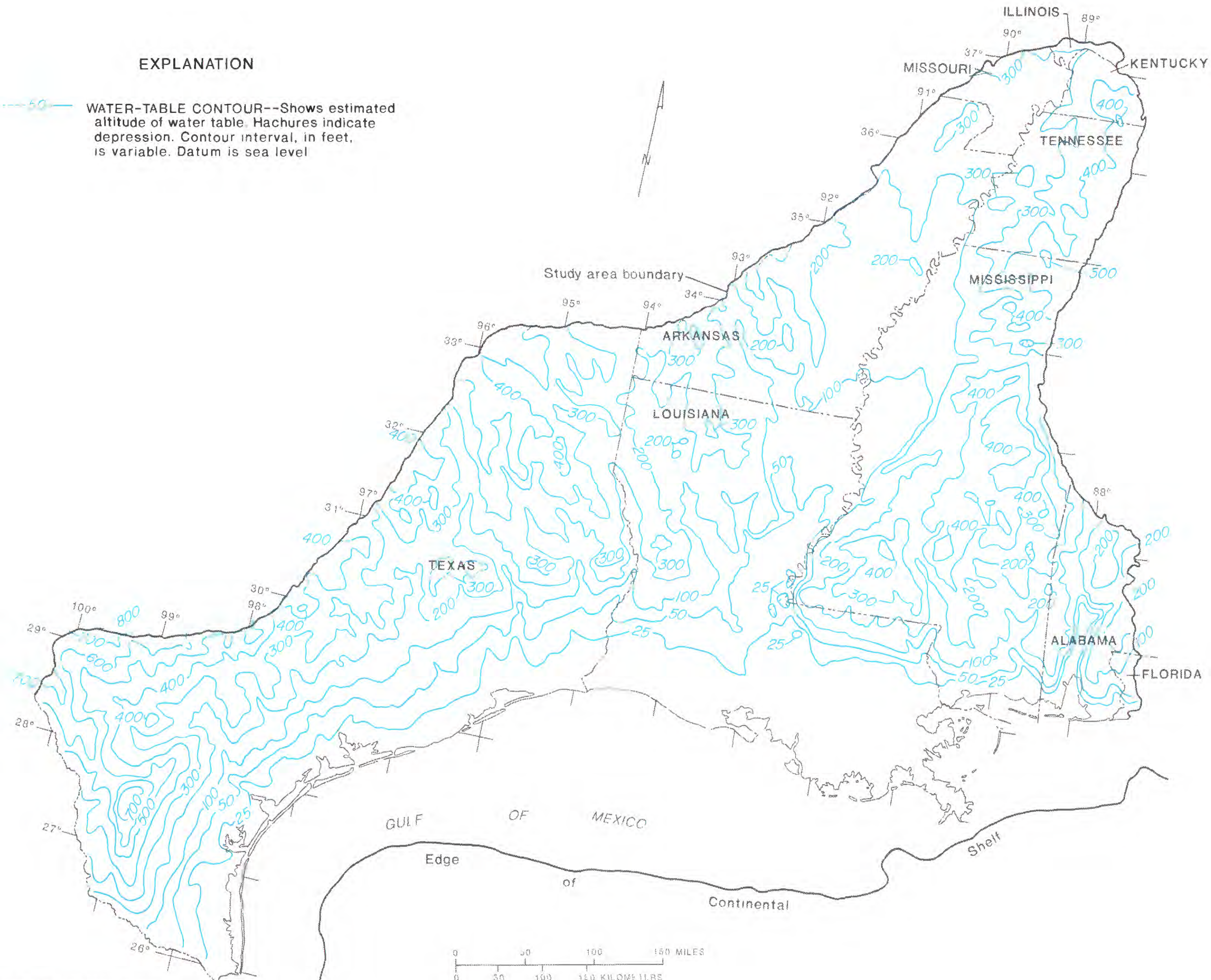
In offshore areas, the source-sink layer was assigned a constant head value of zero (sea level), a density of 1.025 (that of seawater), and a thickness equal to the ocean depth, so that the model would calculate equivalent freshwater hydraulic-head values to use in those constant head blocks.

1980 Hydraulic Head

Hydraulic-head data for 1980 were necessary to calibrate the model. There were more than 40,000 head measurements taken in about 7,000 wells from July 1979 through June 1981. Obviously, some wells were measured more than once in that time period, and if so, the measurements for that well were averaged. Then, 10-mile block-averaged head values were estimated using an inverse distance-squared weighted averaging method of the SURFACE II Graphics System (Sampson, 1978). Several different neighborhood criteria and fitting algorithms were attempted. This gridding algorithm was chosen because the block-averaged estimates should approximate an average over the block rather than an estimate of the head surface at the center of the block. The root mean squared error (RMSE) between the block-averaged observed heads (back interpolated to the well locations) and the well measurements was about 25 ft. This is so large

EXPLANATION

—50— WATER-TABLE CONTOUR--Shows estimated altitude of water table. Hachures indicate depression. Contour interval, in feet, is variable. Datum is sea level



Base from U.S. Geological Survey
Scale 1:2,500,000

Figure 21.--Estimated predevelopment water-table altitude.

because the horizontal and vertical hydraulic-head gradients are commonly very steep in the vicinity of pumping centers and there are abrupt changes in land-surface altitude, therefore, there is great difficulty in estimating an average head for a 100 mi^2 block within each layer. All model blocks which had at least one head measurement within 5 mi of the center of the block and at least three other measurements within a 12 mi radius were used in the comparison. The chosen set of block-averaged observed head data included 1,155 blocks among the 10 aquifer layers.

Storage

Assuming steady-state conditions (that is, no change over time), estimates of storage characteristics of an aquifer system are not necessary for simulations of ground-water flow. This assumption is probably valid for predevelopment conditions. It is also valid within the gulf coast if pumpage remains nearly constant for a long enough time because the flow system will adjust to pumpage by inducing additional recharge and (or) by decreasing natural discharge. Even in large pumping centers, ground-water flow can approach steady-state conditions in the middle of the cone of depression when recharge nearly equals discharge and only a very small proportion of the total pumpage is being supplied from aquifer storage. In addition, simulation assuming steady-state conditions demonstrates the maximum effect of a given rate of pumpage. Transient simulations were made assuming an elastic specific storage of the permeable zones of 3×10^{-6}, that when multiplied by the permeable zone thickness (fig. 19 and table 3), yields a range of storage coefficients from 0.0003 to 0.02, with an average of about 0.003.

Long term changes in depth-to-water in wells indicate trends in ground-water storage. A summary of about 200,000 measurements from about 2,500 wells shows a decline from the mid-1940's until the early 1970's when the water levels began to stabilize (fig. 22). This stabilization was partly due to the decrease in industrial pumpage in the areas which had previously had the largest declines. The areas with the largest increases in pumpage are mostly irrigated areas of the Mississippi River Valley alluvial aquifer which has large hydraulic conductivity, short distances to large rivers, and, in places, a large storage coefficient (specific yield) due to water table conditions. All of these factors contribute to a relatively small drawdown. This summary of changes in ground-water storage was made using the following procedure:

1. Data from a subset of wells which were measured for at least 13 years in the period 1958-83 were used. The average depth-to-water for each well for each year was calculated.

2. The average depth-to-water for each 10-mile grid block in each layer for each year was calculated from the well-year averages. Some wells have measurements in some years but not in others. Depth-to-water was used instead of hydraulic head because when estimating differences between annual averages, the depth-to-water average partially compensates for the differences caused by the annual variation in average land-surface altitude of the wells measured in a particular year.

3. The average depth-to-water of all the block averages for each year was calculated (fig. 22).

HYDRAULIC HEAD AND FLOW IN VARIABLE-DENSITY AQUIFERS

To understand the regional flow patterns in the gulf coast aquifer systems, one must understand how hydraulic head and flow are related in aquifers containing variable-density water. Unlike aquifers containing constant-density water, where flow

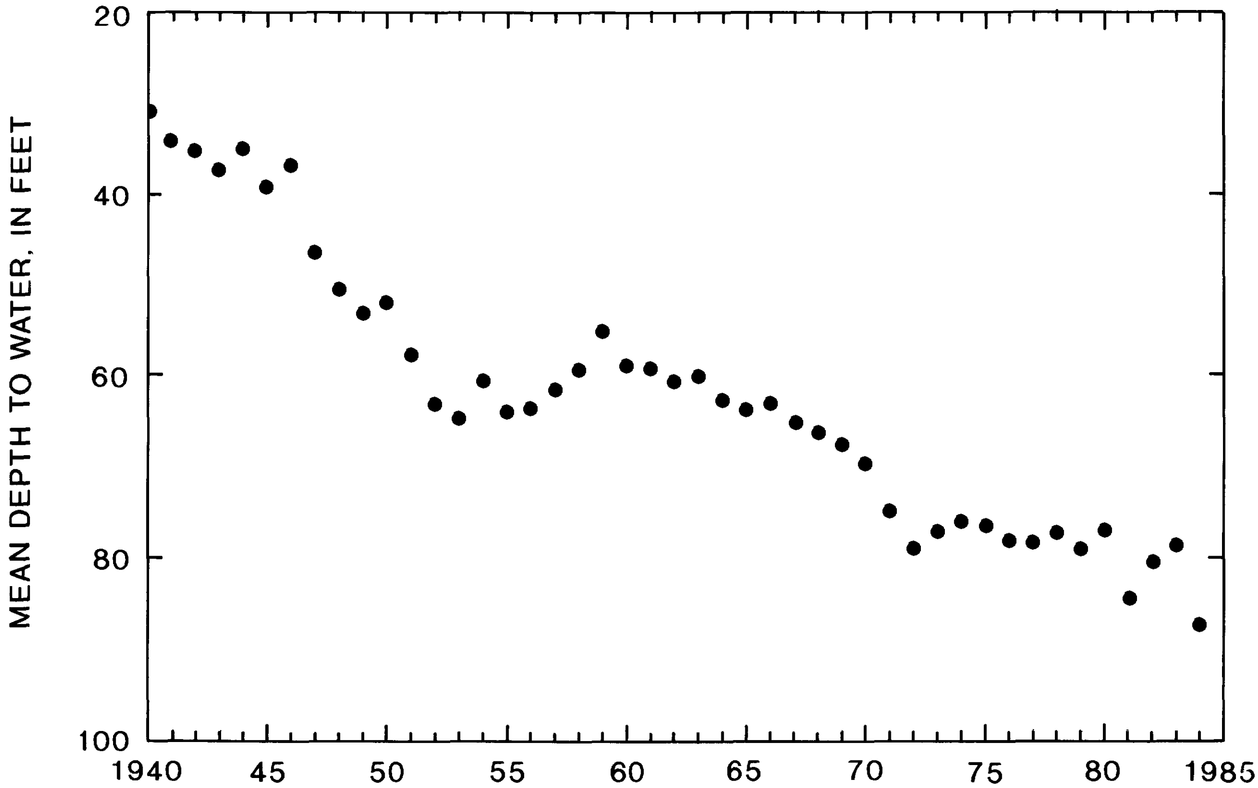


Figure 22.--Mean depth-to-water, 1940-84, averaged from about 2,500 wells.

is controlled only by the hydraulic head gradient and the hydraulic conductivity, variable-density flow is also affected by the density and change in altitude of the aquifer, which will be referred to as "gravitational effects". This second force affecting flow can be understood by taking the simple case where a bucket full of sand is saturated with freshwater except for one end which is saturated with a heavier fluid, such as a brine. Without any other force except for the difference in density, the brine will flow along the bottom of the bucket, displacing water vertically, until it fills the entire lower section of the bucket, with the freshwater "floating" on top of it. If the bucket is tilted, the brine will flow down until it fills the newly defined lowest section of the bucket. This is an exaggerated example of the force exerted in the aquifer system due to varying altitude and density. The actual flow in the aquifer will result from the vector sum of the two forces caused by: hydraulic head gradient; and gravitational effects due to varying density and altitude of the aquifer.

Therefore, in aquifers with variable-density water, hydraulic-head gradients, expressed either in terms of freshwater or formation water, do not necessarily indicate flow directions or magnitudes. Neither do bottom hole pressures, commonly measured by investigators in the oil industry, indicate flow directions or magnitudes (Hanor and Bailey, 1983). Bond (1972, 1973) combined the forces due to hydraulic-head gradients and gravitational effects, defining the "head available to cause flow", Psi, between points 1 and 2 in the aquifer, as:

$$\text{Psi} = \text{FWH2} - \text{FWH1} - \int_{2}^{1} (r - 1)\, dz$$

where,

FWH1,2 = Equivalent freshwater head at points 1 and 2.

$\int_{2}^{1}$ = integral along flow path from point 2 to 1,

r = relative density of the ground water under the temperature and pressure of its sampling, compared with the density of pure water at 75 F and 1 atm. pressure,

dz = the change in altitude in the aquifer between points 1 and 2.

The difference in the first two terms, FWH2 - FWH1, is the familiar difference in hydraulic head, with the minor adjustment of observed well water-level or pressure data to equivalent freshwater head. The integral term defines the force exerted due to the variable density and altitude in the aquifer. Bond (1973) gives exceptions to this equation for two conditions. The first exception is where the structure of the top and bottom of the aquifer involves troughs, saddles, anticlines, and synclines that can cause gravitational effects. In this case, the head differences that are caused by the structure usually reduce the head that is available to cause flow. The second exception is where the rock contains permeability barriers, which prevent horizontal flow within the aquifer or divert flow from horizontal paths. In this case, valid flow patterns cannot be deduced. These exceptions may be common because heterogeneity in aquifers is common (Bond, 1973).

The significance of the effect of density on hydraulic head and flow can be demonstrated using layer 8 (permeable zone D containing mostly middle Miocene deposits). Lusczynski, 1961, defined "point-water head" (referred to as "hydraulic head" by Kuiper, 1985) as the water level, referred to sea-level datum, in a well sufficiently filled with water from the formation at the well screen to balance the existing pressure. This will be called formation-water head in this report. The formation-water head for layer 8 under predevelopment conditions is shown in figure 23. Note that the formation-water head map (fig. 23) has depressions in the brine area that look like depressions in the head due to pumping, even though the map is from a steady-state, predevelopment simulation. The closed depressions are due to the variation in water density and the dip of the layer. Also, note that the freshwater head decreases from more than 300 ft in the higher parts of the outcrop area to nearly zero and then increases to more than 600 ft near the deepest downdip extent of the layer (fig. 25g). This seems impossible, but it is possible in a system with variable-density flow. The flow vectors (fig. 26g) are not necessarily perpendicular to either the freshwater head contours or the formation-water head contours.

PRELIMINARY RESULTS

Three simulations of different conditions that provided preliminary results are: predevelopment (prior to pumping); 1980 pumpage as steady-state; and 1935-1980 pumpage under transient conditions. Confidence in these results from a preliminary calibration is higher than might be expected because they are mainly based on relative comparisons of one area to another or one layer to another. A model prediction of hydraulic head at some time will contain errors in the conceptual model, errors in the system parameters, errors in the initial conditions, and errors in the pumpage data (Durbin and others, 1978, p. 126). Relative comparisons minimize errors due to initial conditions and pumpage data. As mentioned previously, Konikow (1978) reported what many investigators have found--that most significant changes are made during the first few calibration simulations, so the results reported here are expected to change little with more detailed work.

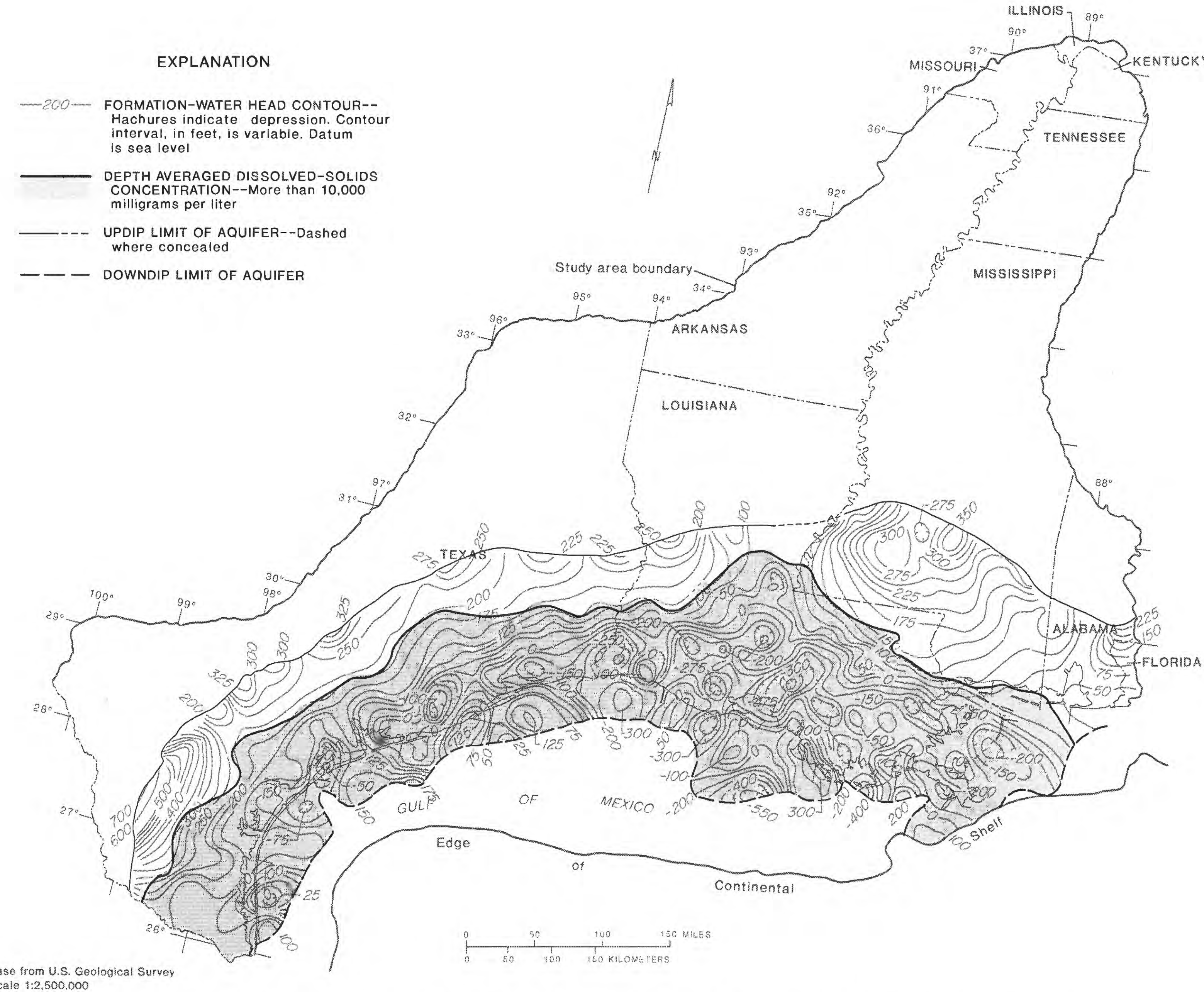


Figure 23.--Simulated predevelopment fromation-water head in model layer 8 (zone D of the coastal lowlands aquifer system containing mostly middle Miocene deposits).

Predevelopment flow was simulated assuming steady-state conditions (no change with time) with no pumping. Conditions for 1980 were also simulated as if steady-state conditions were reached using 90 percent of the 1980 pumping, assuming 10 percent of the 1980 pumpage was being supplied from storage, external to the simulation.

A transient simulation was made using the time periods and pumpage outlined below, predevelopment simulated hydraulic heads as initial conditions, and an elastic specific storage of 3×10^{-6}.

Transient-simulation time periods

Period ended	Length (years)	Pumpage used
1935	10	1960 * .149
1945	10	1960 * .310
1955	10	1960 * .569
1962	7	1960
1967	5	1965
1972	5	1970
1977	5	1975
1980	3	1980

The pumpage before 1960 was estimated using 1960, 1970, and 1980 pumpage totals and 1930-80 population data by simple linear regression.

Simulated heads using 1980 steady-state conditions were compared with observed data for preliminary model calibration because: 1) It is a simpler simulation with fewer unknowns, 2) more data were available for 1980 than were available for predevelopment conditions, and 3) more stress on the aquifer system from pumpage allowed a more rigorous test of the simulation. These advantages outweigh the disadvantage of making preliminary simulations of developed conditions ignoring the effects of changes in ground-water storage. Additionally, the effects of this shortcoming were mitigated somewhat by using only 90 percent of the 1980 pumpage in the steady-state 1980 simulation. Preliminary analysis of hydraulic-head data indicates on the average about 10 percent of the 1980 pumpage was derived from storage. This is close to the figure of 7 percent coming from storage in 1982 in a simulation of conditions from predevelopment through 1982 using an elastic specific storage of 1×10^{-6} and simulation of land subsidence by Kuiper (U.S. Geological Survey, written commun., 1988).

The fit of the 1980 steady-state simulated hydraulic head to the observed head data was adequate for this preliminary analysis (fig. 24). The overall root-mean-squared error between the observed and simulated heads in those 1,155 blocks was 56.9 ft. (table 5). The mean error was 7.4 ft, and the mean of the absolute values of the error was 39.2 ft. Considering the root-mean-squared error built into any set of block-averaged observed head data, and the simple level of calibration with constant vertical and horizontal hydraulic conductivities for each entire layer, this is quite good, and adequate for the results presented in the following sections. The shape of the simulated potentiometric surfaces agree with the surfaces derived from observed water levels.

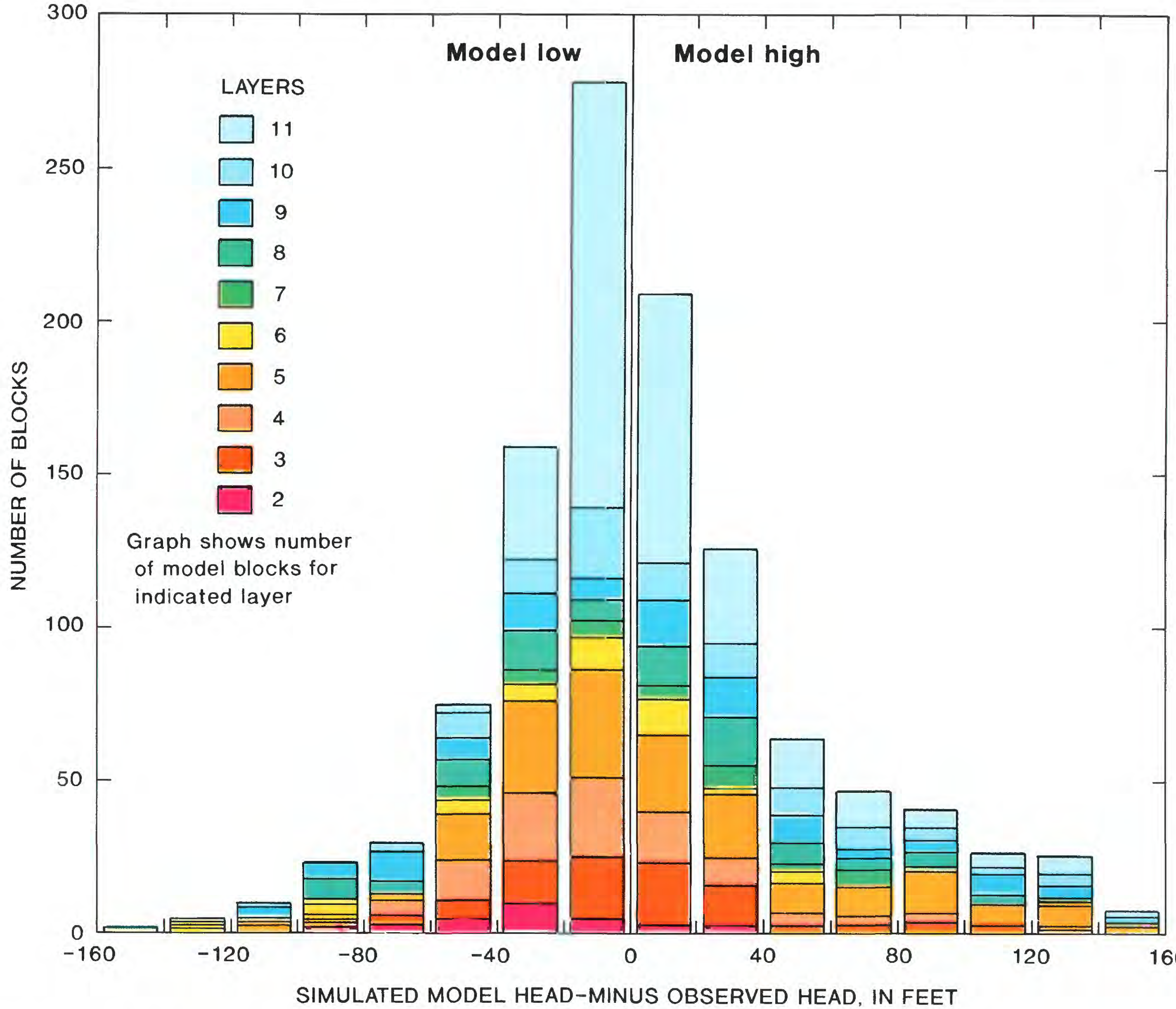


Figure 24.--Graph showing fit of simulated steady-state and observed 1980 hydraulic head.

Table 5.--*Fit of simulated hydraulic head to observed block-average head for 1980, assuming steady-state conditions and 90 percent of the 1980 pumpage, in feet*

[Std. dev., Standard deviation; MEAS, Mississippi embayment aquifer system; TCUAS, Texas coastal uplands aquifer system; CLAS, Coastal lowlands aquifer system; GC RASA, Gulf Coast Regional Aquifer-System Analysis]

Layer system	Number of blocks			Mean		Mean head difference			Head difference			Root mean square error
	Observed heads	Model high	Model low	Observed head	Model head	All blocks	Model high	Model low	Std. dev.	Min-imum	Max-imum	
2	35	9	26	196.3	172.4	-23.9	39.0	-45.6	47.1	-128.1	110.3	52.2
3	92	47	45	277.5	285.3	7.8	43.0	-29.1	52.8	-92.7	236.7	53.1
4	108	40	68	276.7	274.4	-2.2	44.4	-29.6	50.3	-103.4	193.7	50.1
5	190	99	91	123.9	135.3	11.4	58.1	-39.5	69.1	-247.4	202.9	69.9
6	46	18	28	128.5	109.8	-18.6	20.6	-43.8	48.9	-210.1	58.3	51.8
MEAS & TCUAS	471	213	258	194.8	196.8	2.0	48.2	-36.2	59.7	-247.4	236.7	59.7
7	36	21	15	162.6	176.5	13.8	50.9	-38.0	57.3	-129.1	158.3	58.1
8	93	50	43	155.3	157.3	2.0	46.6	-49.9	60.2	-148.8	165.9	59.9
9	104	58	46	54.3	64.3	10.0	60.3	-53.4	74.2	-121.9	282.6	74.5
10	102	55	47	0.4	21.6	21.1	64.2	-29.2	64.2	-177.7	223.3	67.3
11	349	170	179	34.0	44.8	10.8	36.7	-13.8	39.2	-50.4	198.4	40.6
CLAS	684	354	330	55.3	66 .5	11.2	47.1	-27.3	53.9	-177.7	282.6	55.0
GC RASA	1,155	567	588	112.2	119.6	7.4	47.5	-31.2	56.5	-247.4	282.6	56.9

Simulated Predevelopment Hydraulic-Head Distribution and Flow Direction

Simulated freshwater hydraulic-head for each aquifer is shown in figure 25 a-j and the regional ground-water flow directions and relative magnitudes are shown in figure 26 a-j for predevelopment conditions. The horizontal flow vectors are resolved at the intersection of four adjacent blocks to be comparable to vector maps of pumping conditions where pumping out of one block will be shown as the four surrounding vectors pointing toward the pumping block. The plotted vector equals the vector addition of the average of the two adjacent x-direction flows and the average of the two adjacent y-direction flows. The arrows show the direction of horizontal flow and its magnitude both qualitatively (by the arrowhead size) and quantitatively (by the arrow shaft length if it is longer than the arrowhead). The arrowhead size signifies the order of magnitude (integer component of the logarithm, base 10 (LOG 10)) of the flow and the overall length including the shaft is proportional to the mantissa of the LOG 10 flow. If the length is less than the arrowhead length, no shaft is shown. The type and color of the symbol on the flow maps shows the direction of vertical flow out of or into the bottom of each layer. The size of the symbol is proportional to the natural logarithm of the magnitude of the vertical flow. Flow vectors pointing toward the outside no flow boundary can be explained by the fact that the excess flow in the last model block goes up and (or) down into other blocks above or below.

Very little hydraulic-head data was available outside the outcrop of the individual layers for predevelopment conditions. Note that the model was preliminarily calibrated for 1980 conditions and then pumping removed to make these simulations. Therefore, large errors are possible especially on the small scale of individual flow vectors. More confidence can be placed in large, regional trends with many vectors following the same pattern and that appeared similar in most of the different calibration simulations. Only those significant features, which are likely to remain mostly unchanged through a range of simulation parameters, will be described in the text.

Layer 11 freshwater hydraulic-head in the coastal lowlands aquifer system never really gets to zero (minimum about 3 ft) before it starts increasing offshore. The flow map for layer 11 shows flow normal to the head contours and toward the coast (lower altitude) onshore as would be expected in a freshwater system. Near the coastline, the flow direction reverses and offshore, the flow direction is generally toward the coast. In the Mississippi River Valley alluvial aquifer, the flow directions are quite complex because the water moves in the direction of the shortest distance for flow to one of the major surface rivers or drains. Flow patterns in the deeper coastal lowlands aquifer system layers are more difficult to interpret. Layer 10 shows a pattern of hydraulic head and flow similar to layer 11 in the coastal lowlands aquifer system. The freshwater head rises substantially offshore due to the effects of density. Some circulation cells appear in that part of the layer containing highly saline water. Both of these features become more pronounced in layer 9 because it is deeper, more saline, and therefore more affected by the density driving force. These circulation cells can occur due to the density forces resulting from heat and solute gradients (L.K. Kuiper, U.S. Geological Survey, oral commun., 1985) as the denser, heavier waters overturn to get under the fresher, lighter waters. This can also occur in a continuing fashion somewhat analogous to convection currents in the atmosphere if a continuous force such as a heat source or salt dissolution exists. The pattern of hydraulic head in layers 7, 8, and 9 roughly follows what has already been described. The flow maps are more complex due to the high and highly variable density of the water. The part of the aquifers containing freshwater becomes fairly narrow compared to the brine section.

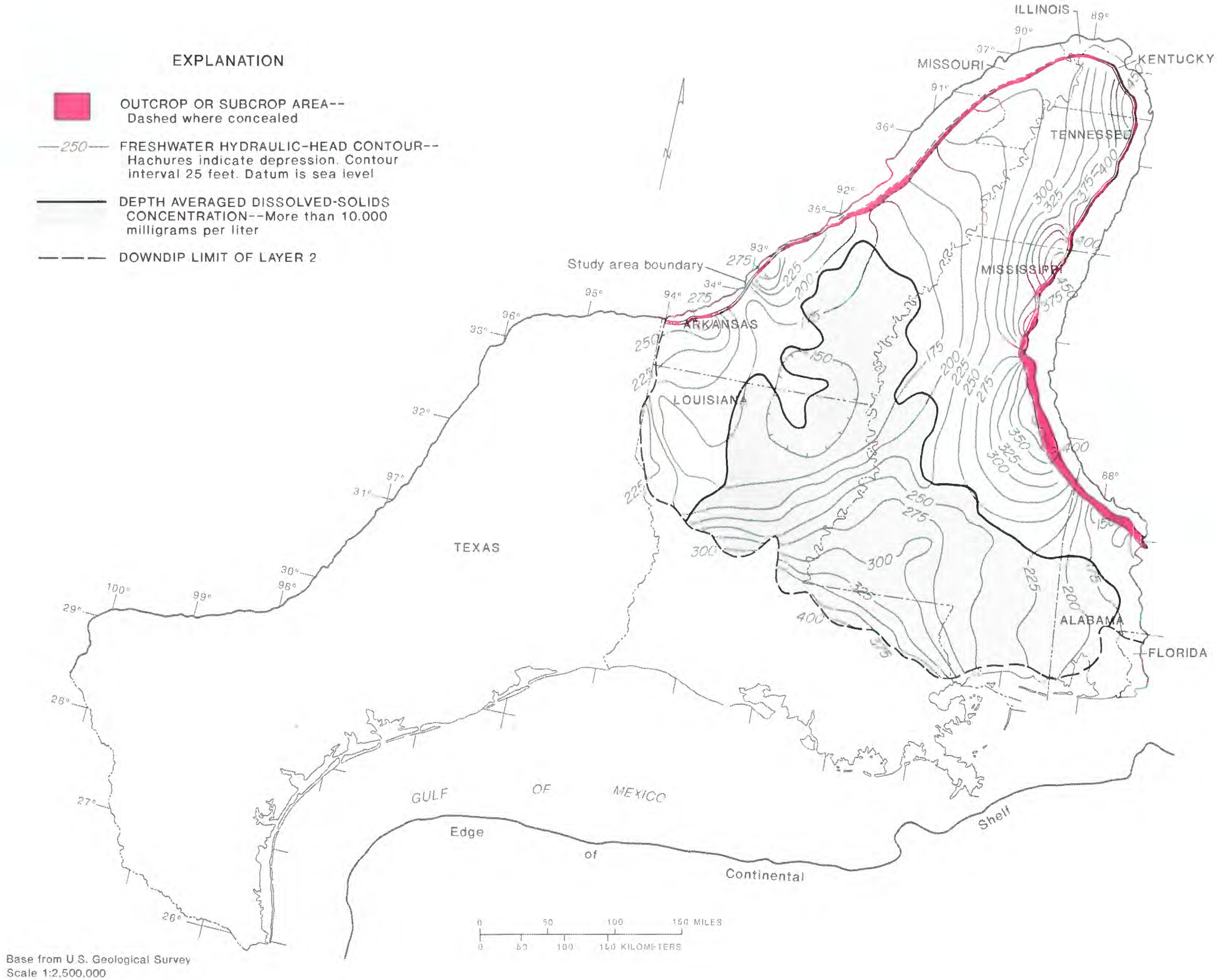


Figure 25a.--Simulated predevelopment freshwater head in model layer 2 (lower Wilcox aquifer).

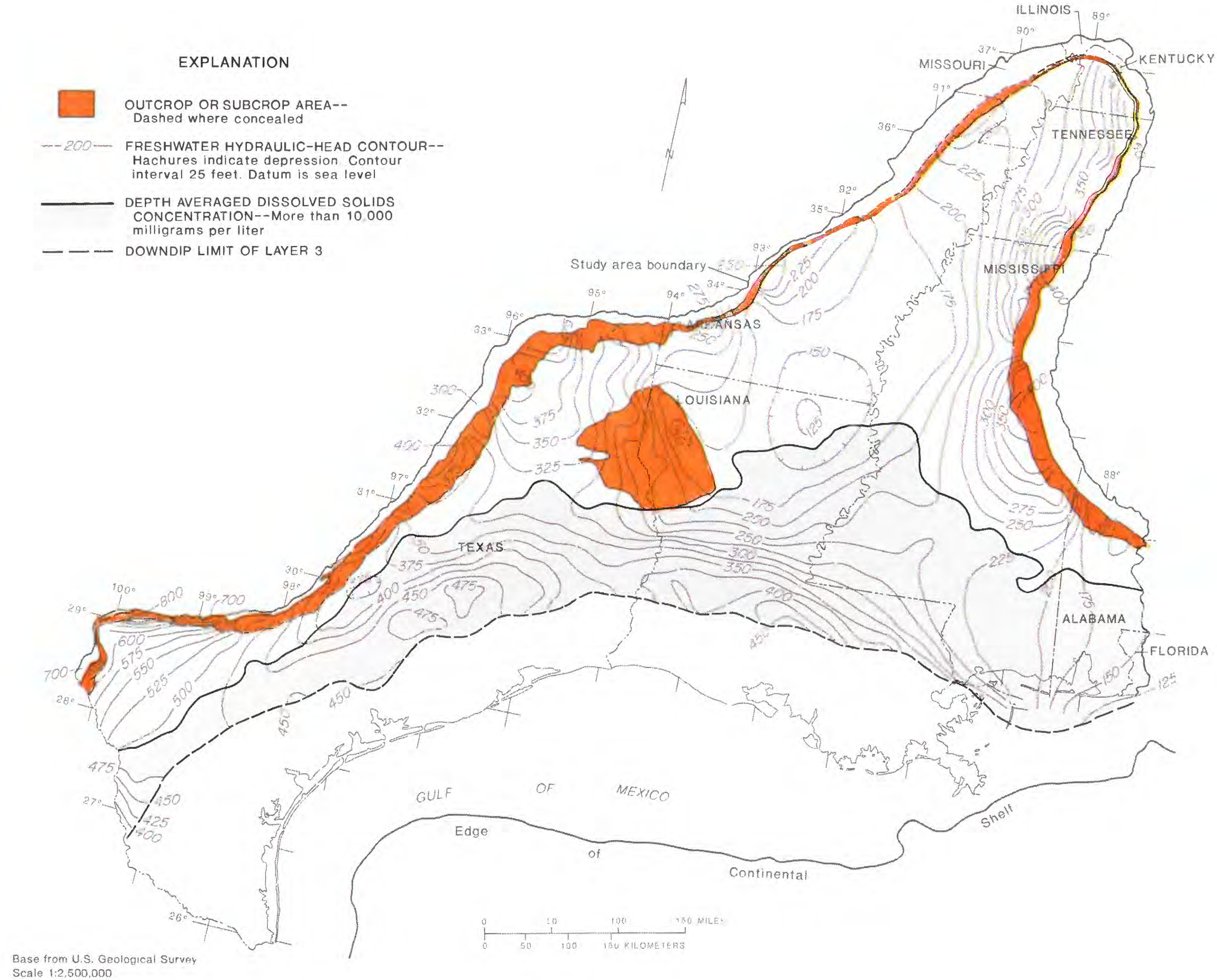


Figure 25b.--Simulated predevelopment freshwater head in model layer 3 (middle Wilcox aquifer).

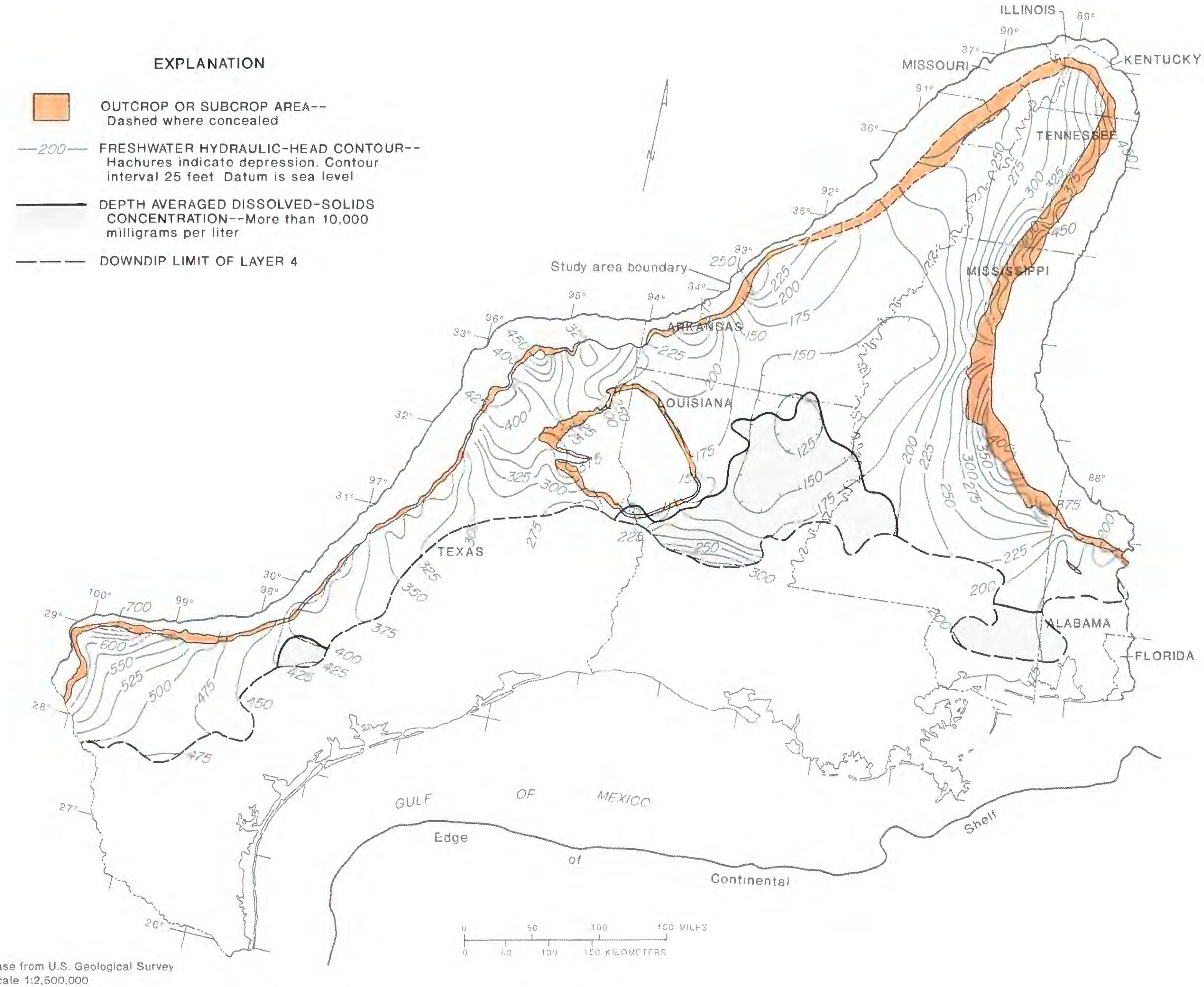


Figure 25c.--Simulated predevelopment freshwater head in model layer 4 (lower Claiborne-upper Wilcox aquifer).

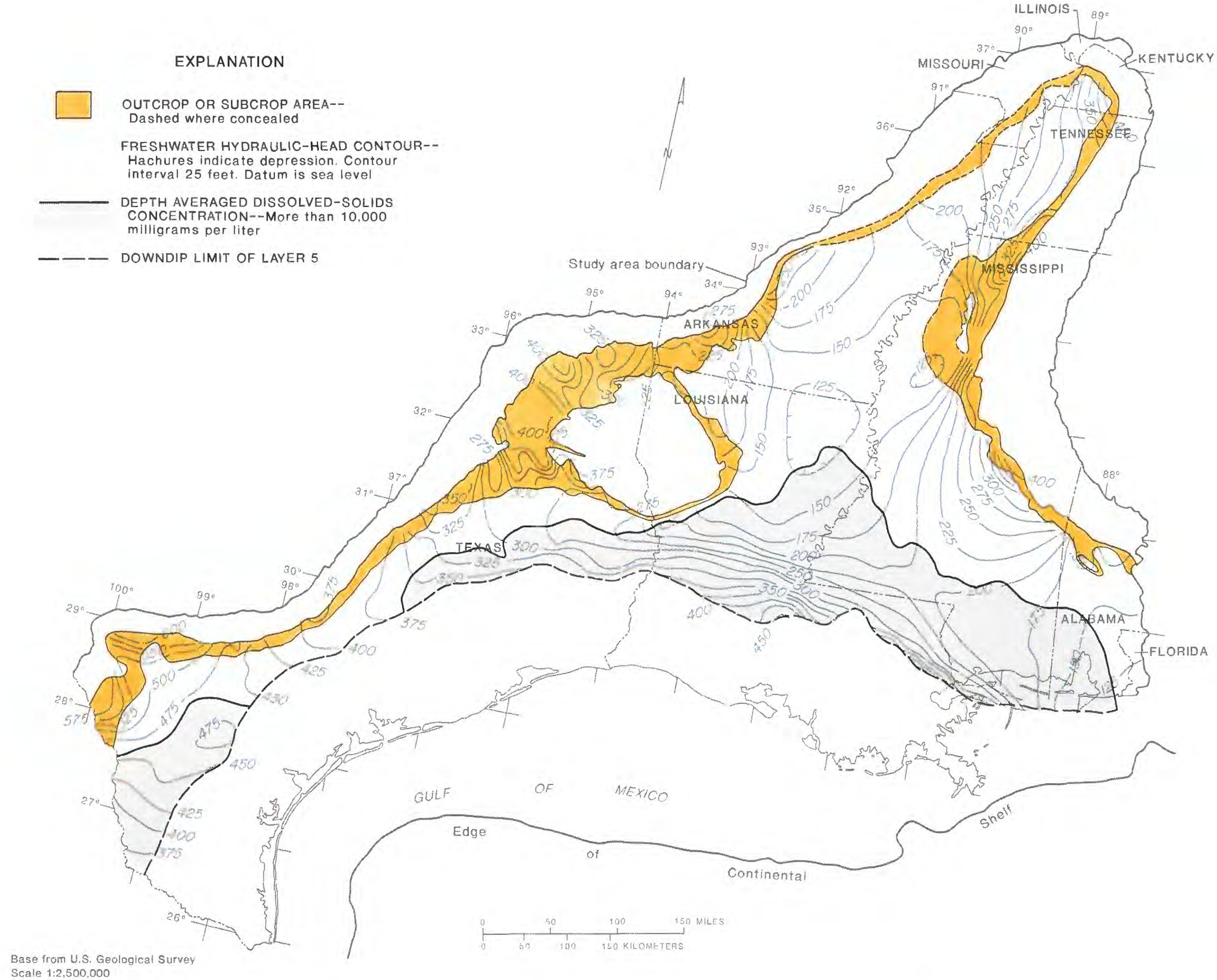


Figure 25d.--Simulated predevelopment freshwater head in model layer 5 (middle Claiborne aquifer).

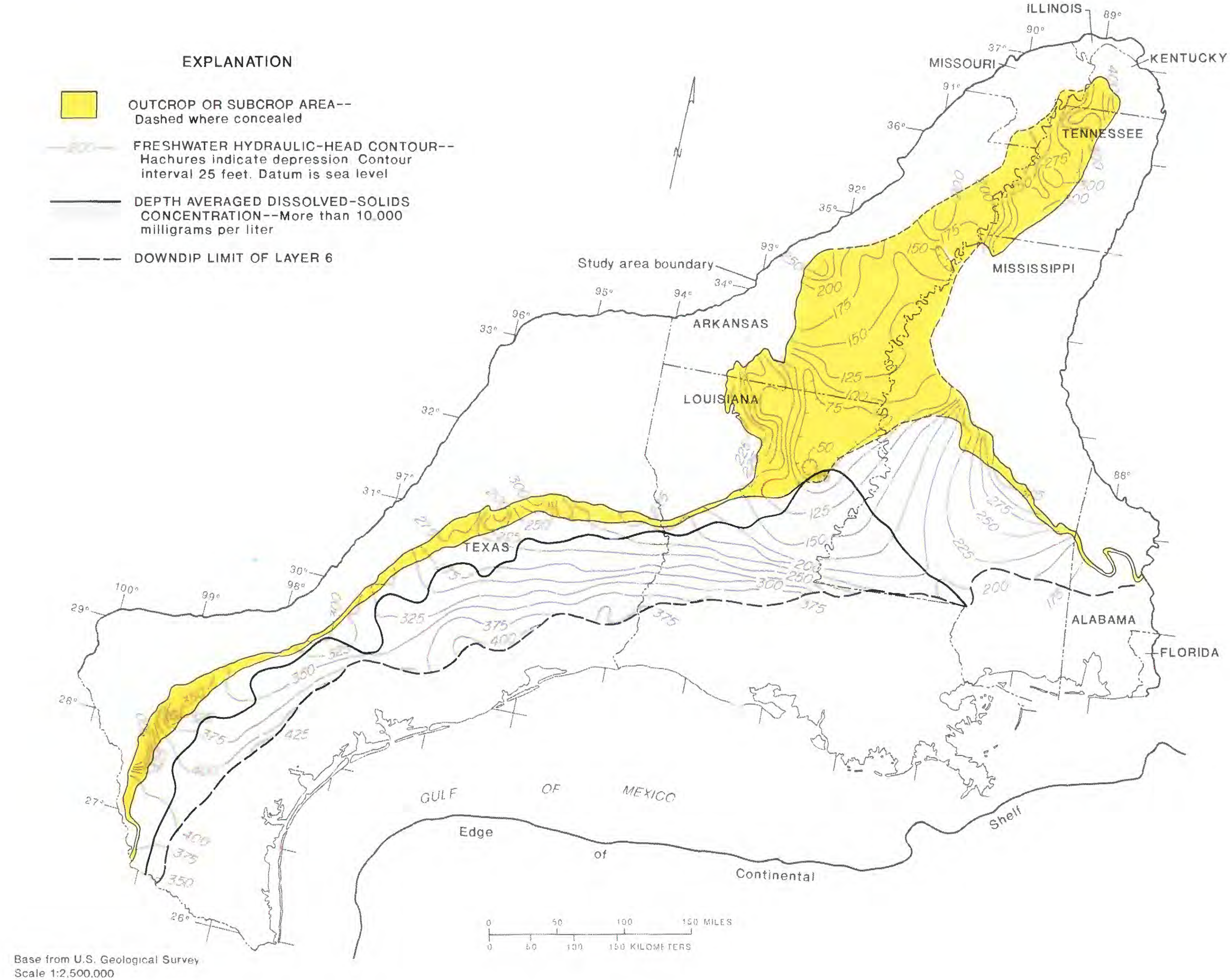


Figure 25e.--Simulated predevelopment freshwater head in model layer 6 (upper Claiborne aquifer).

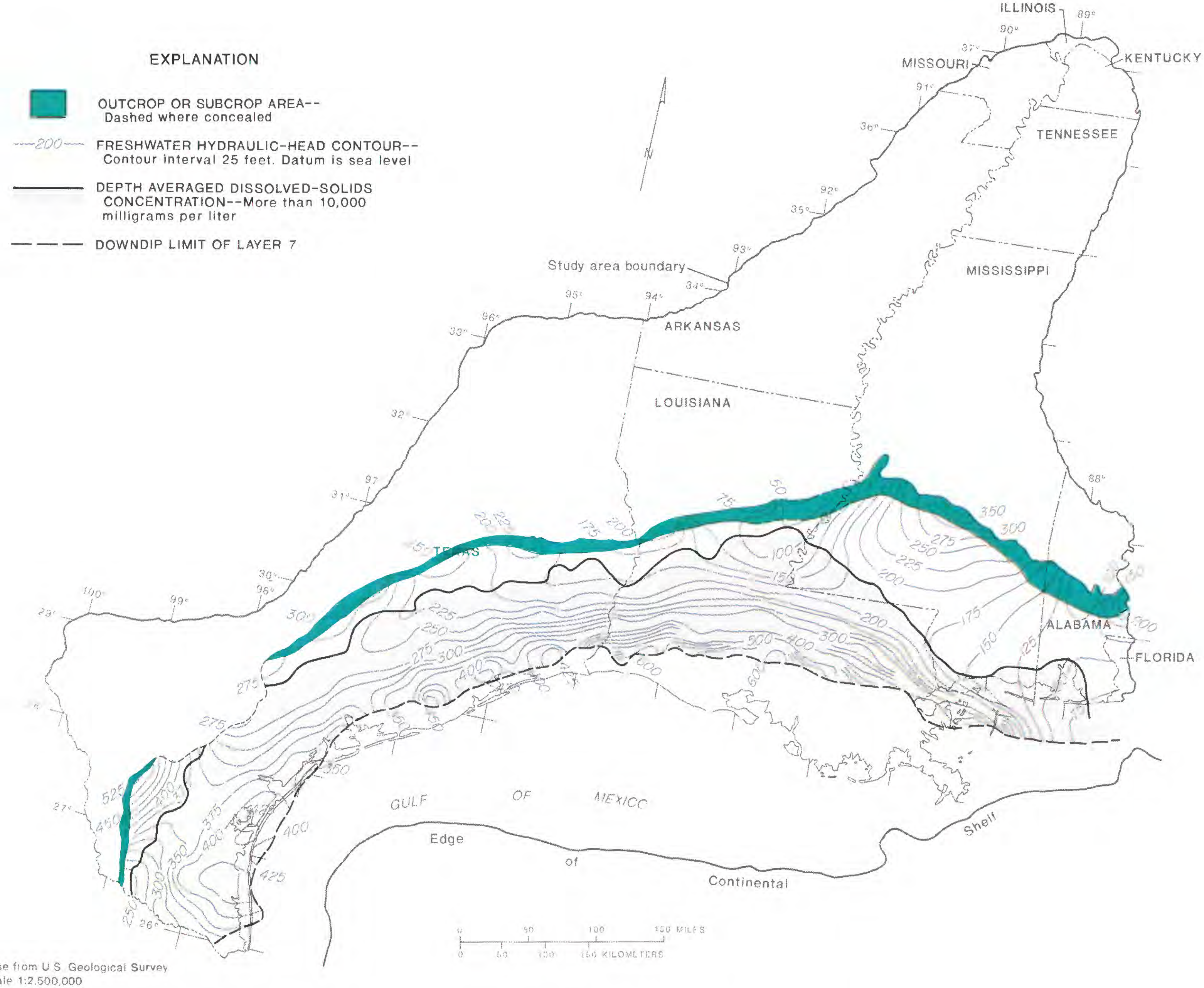


Figure 25f.--Simulated predevelopment freshwater head in model layer 7 (permeable zone E, lower Miocene-upper Oligocene).

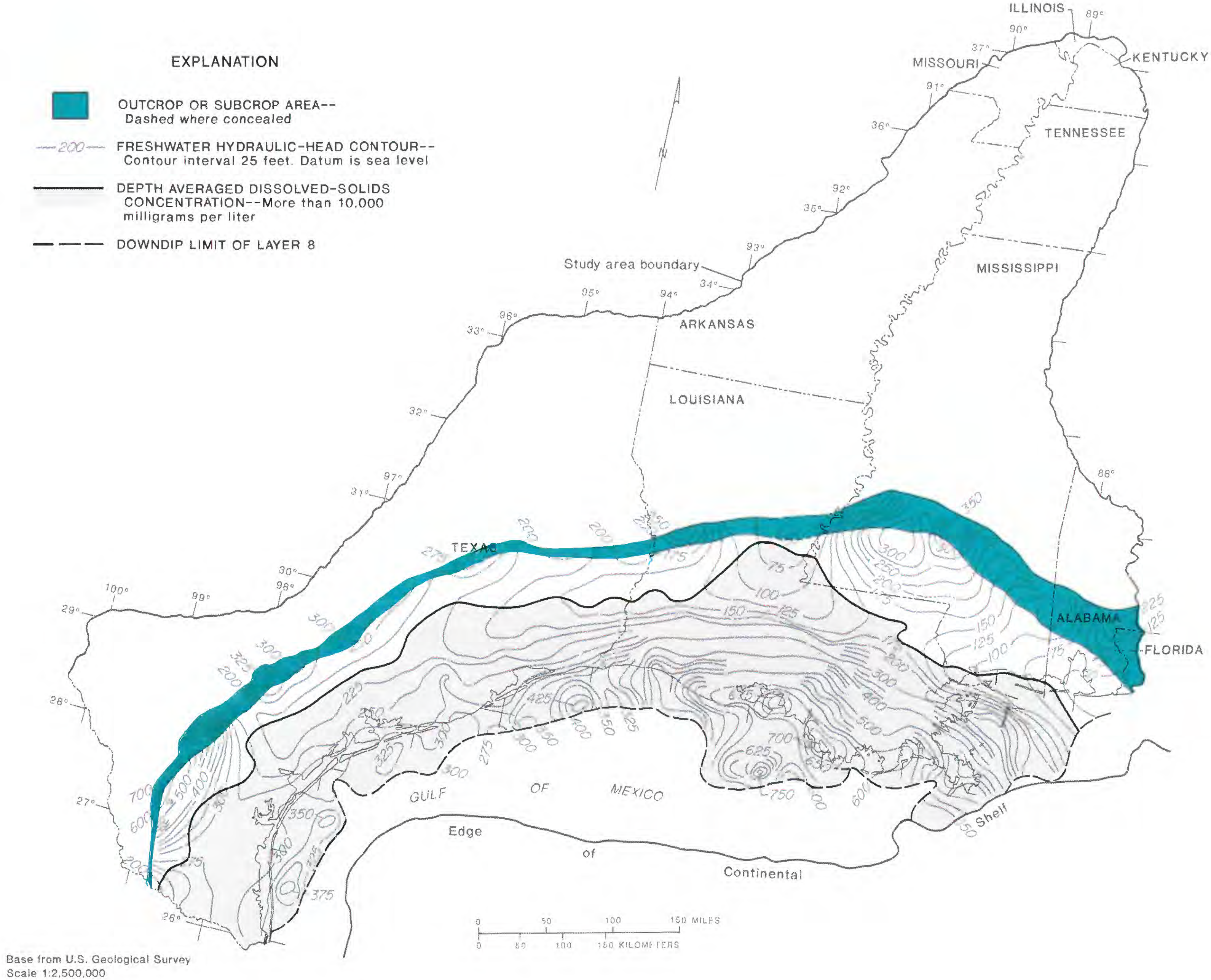


Figure 25g.--Simulated predevelopment freshwater head in model layer 8 (permeable zone D, middle Miocene deposits).

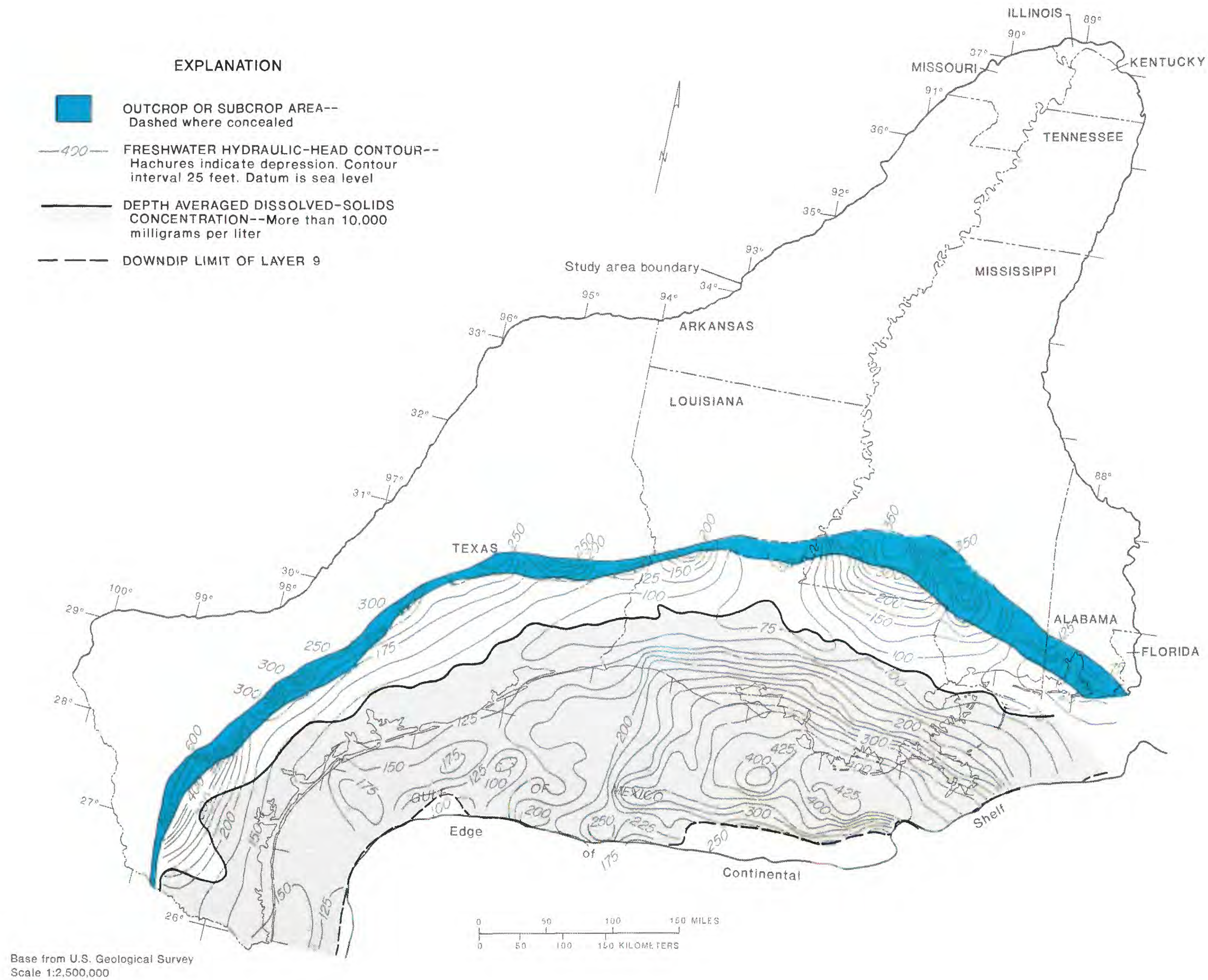


Figure 25h.--Simulated predevelopment freshwater head in model layer 9 (permeable zone C, lower Pliocene-upper Miocene deposits).

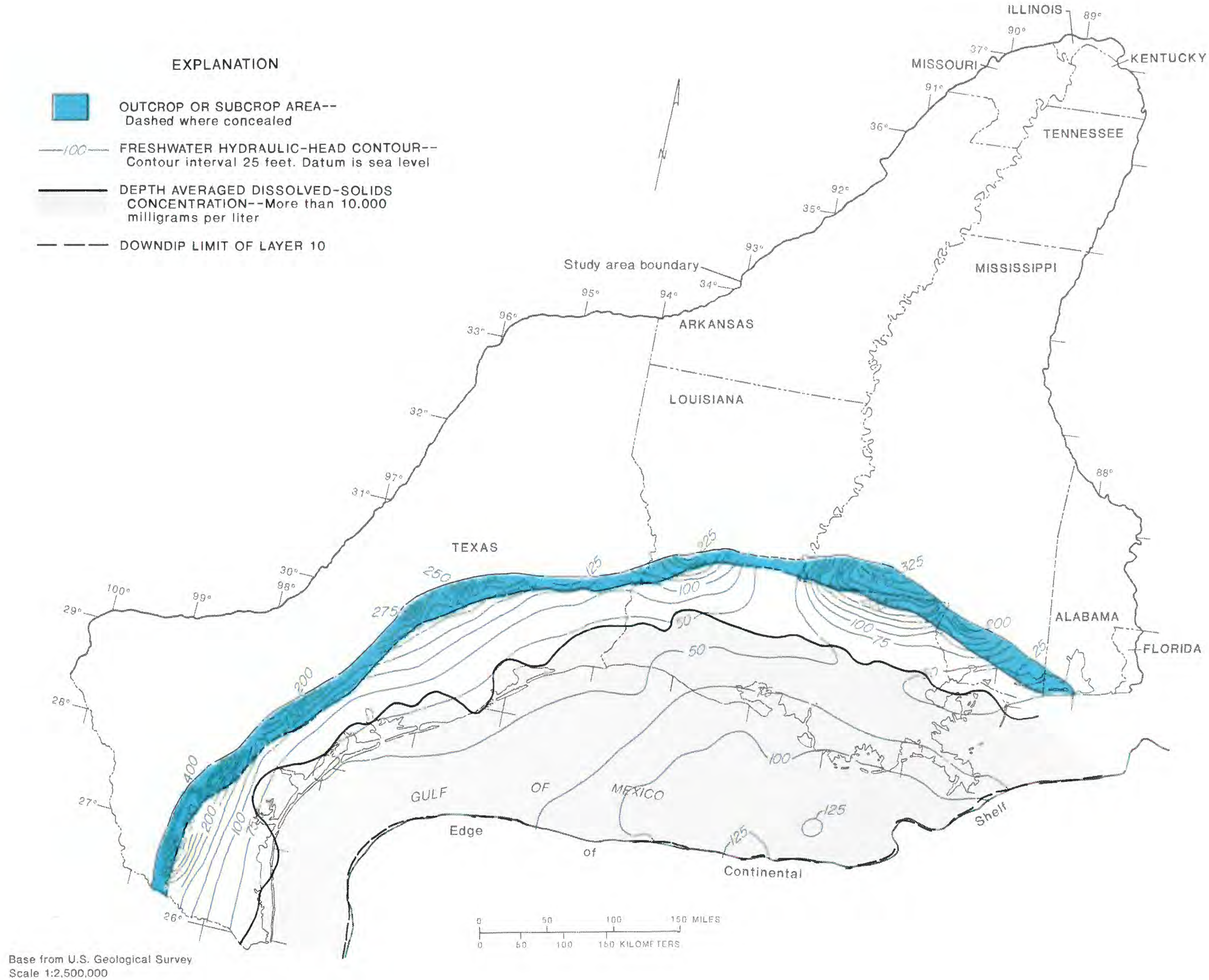


Figure 25i.--Simulated predevelopment freshwater head in model layer 10 (permeable zone B, lower Pleistocene-upper Pliocene deposits).

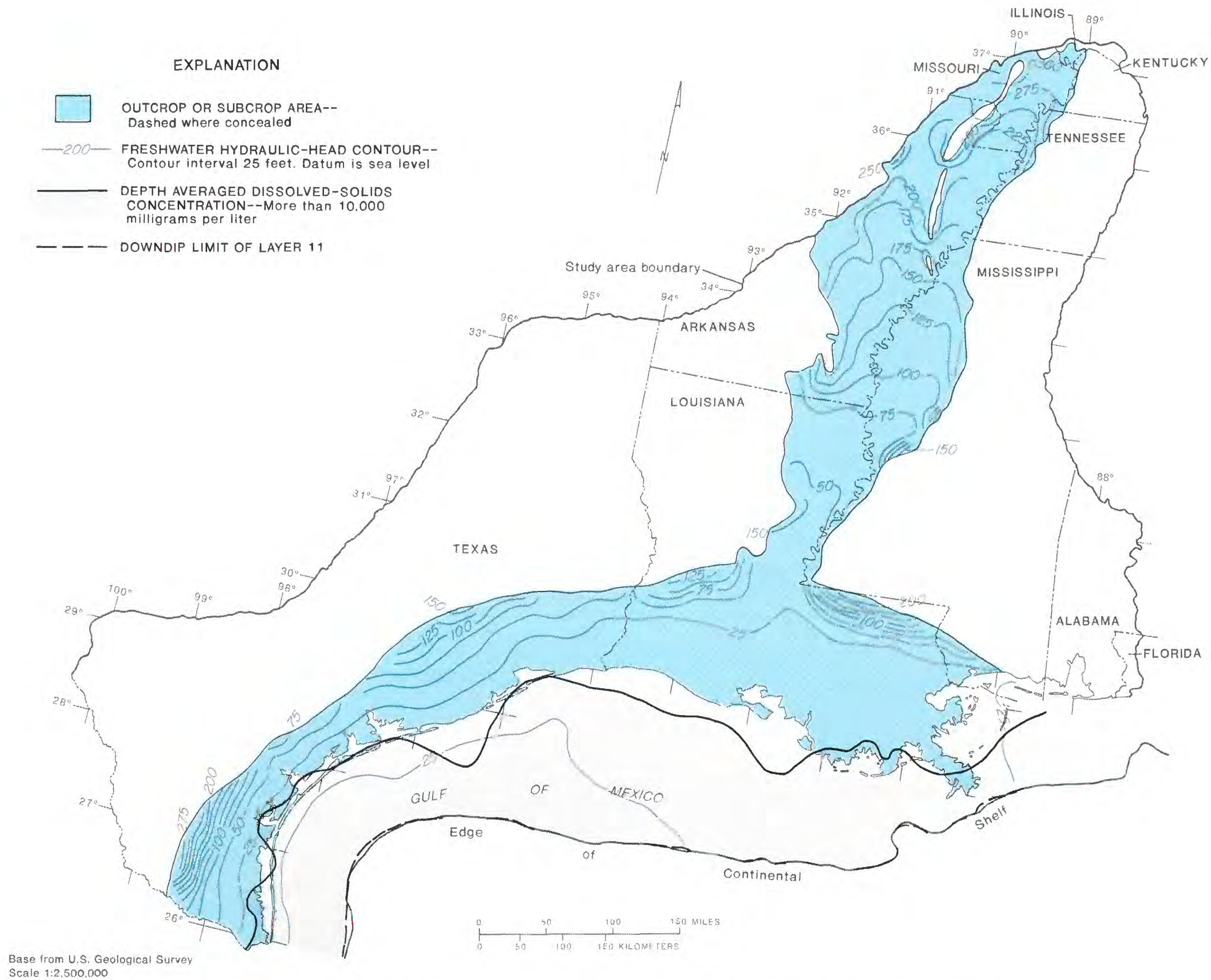


Figure 25j.--Simulated predevelopment freshwater head in model layer 11 (Mississippi River Valley alluvial aquifer and permeable zone A, Holocene-upper Pleistocene deposits).

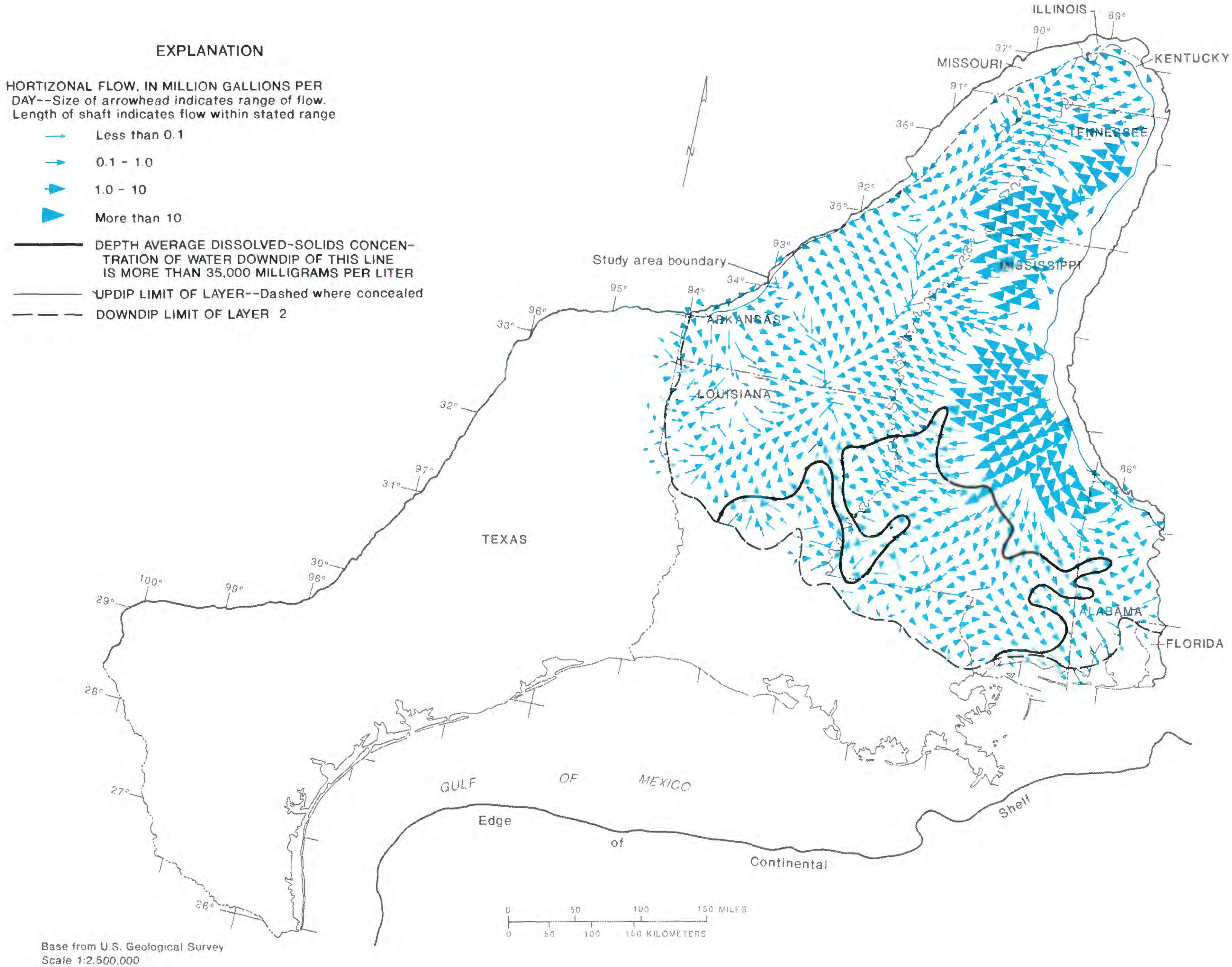


Figure 26a.--Simulated predevelopment regional ground-water flow directions and relative magnitudes in model layer 2 (lower Wilcox aquifer).

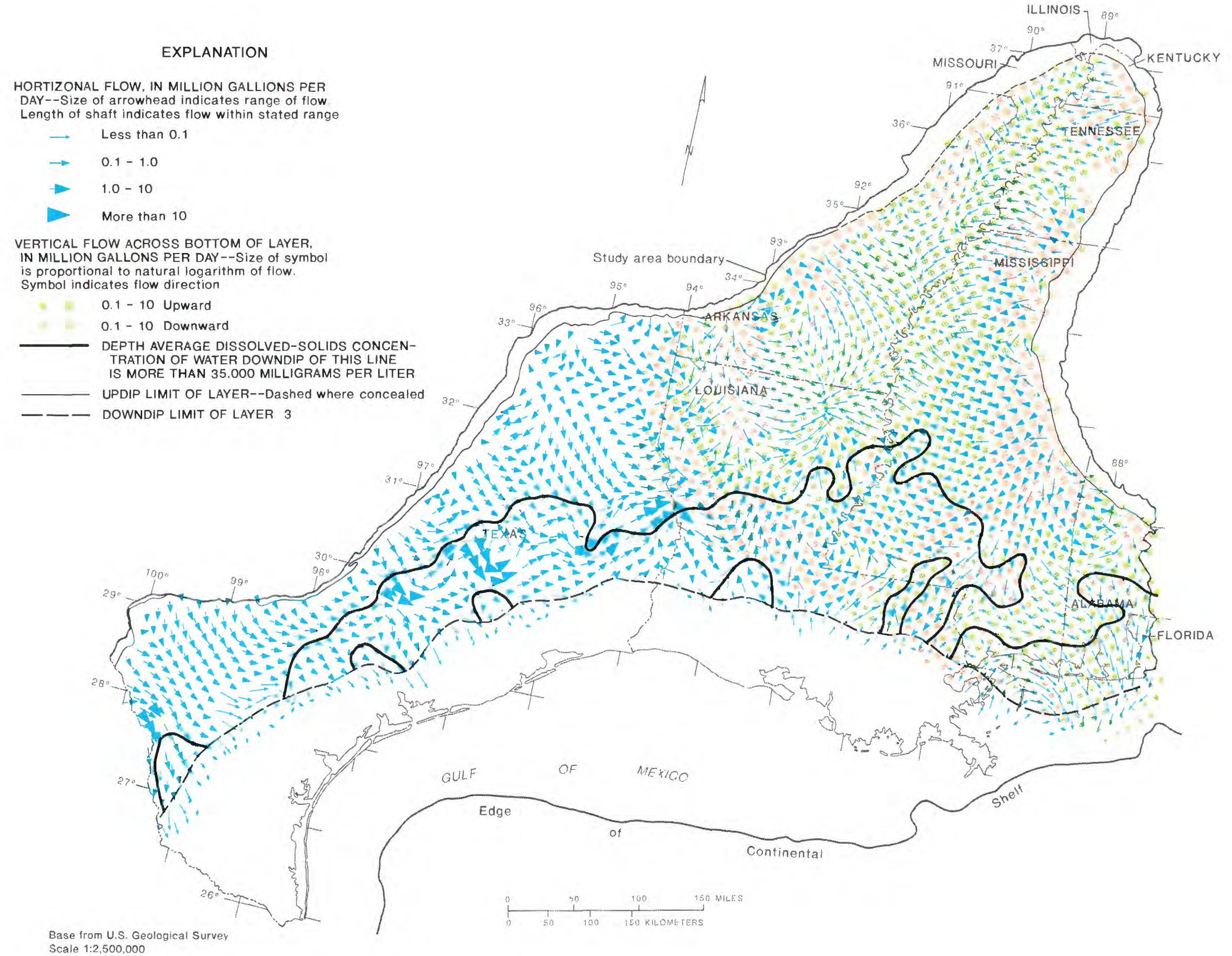


Figure 26b.--Simulated predevelopment regional ground-water flow directions and relative magnitudes in model layer 3 (middle Wilcox aquifer).

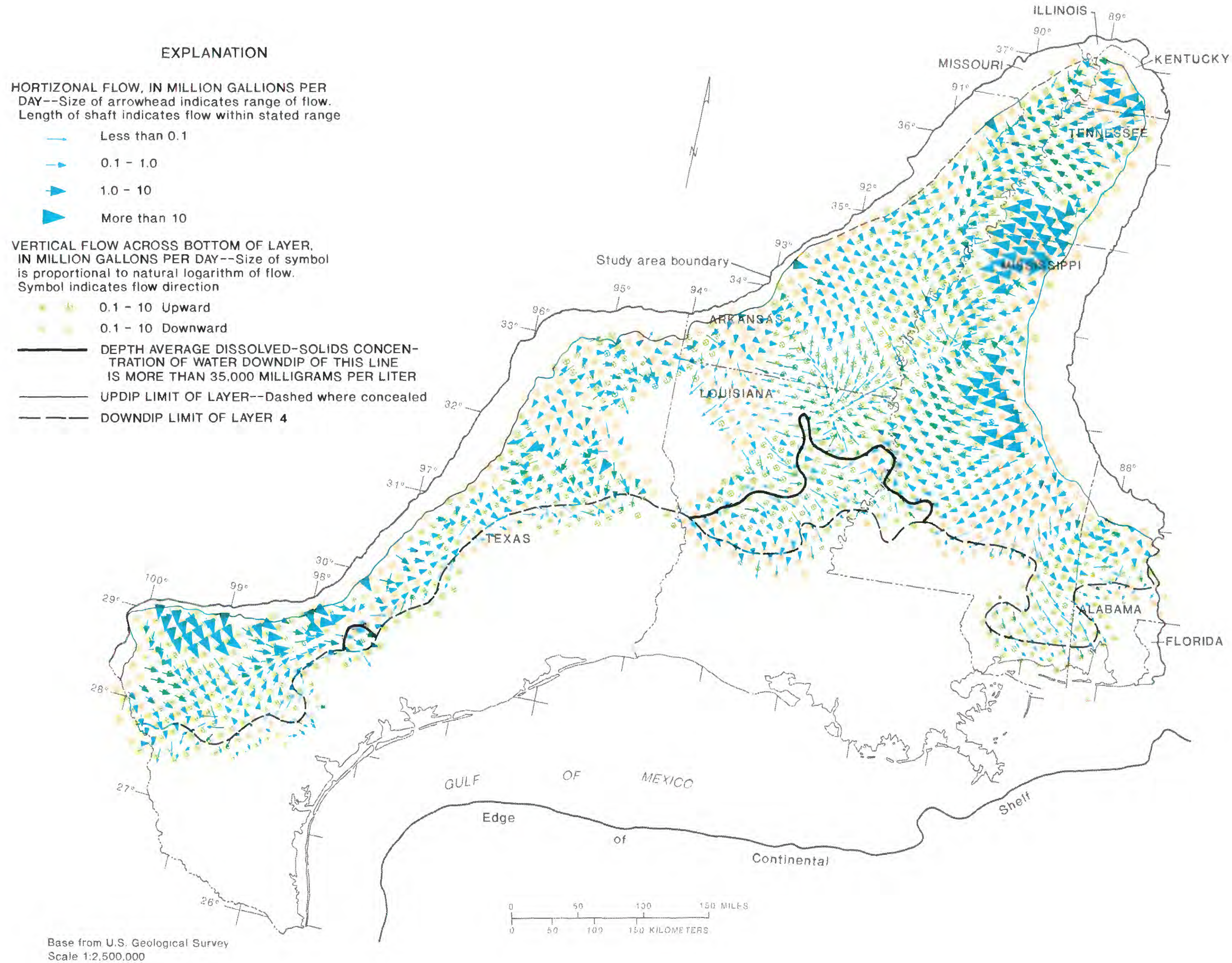


Figure 26c.--Simulated predevelopment regional ground-water flow directions and relative magnitudes in model layer 4 (lower Claiborne-upper Wilcox aquifer).

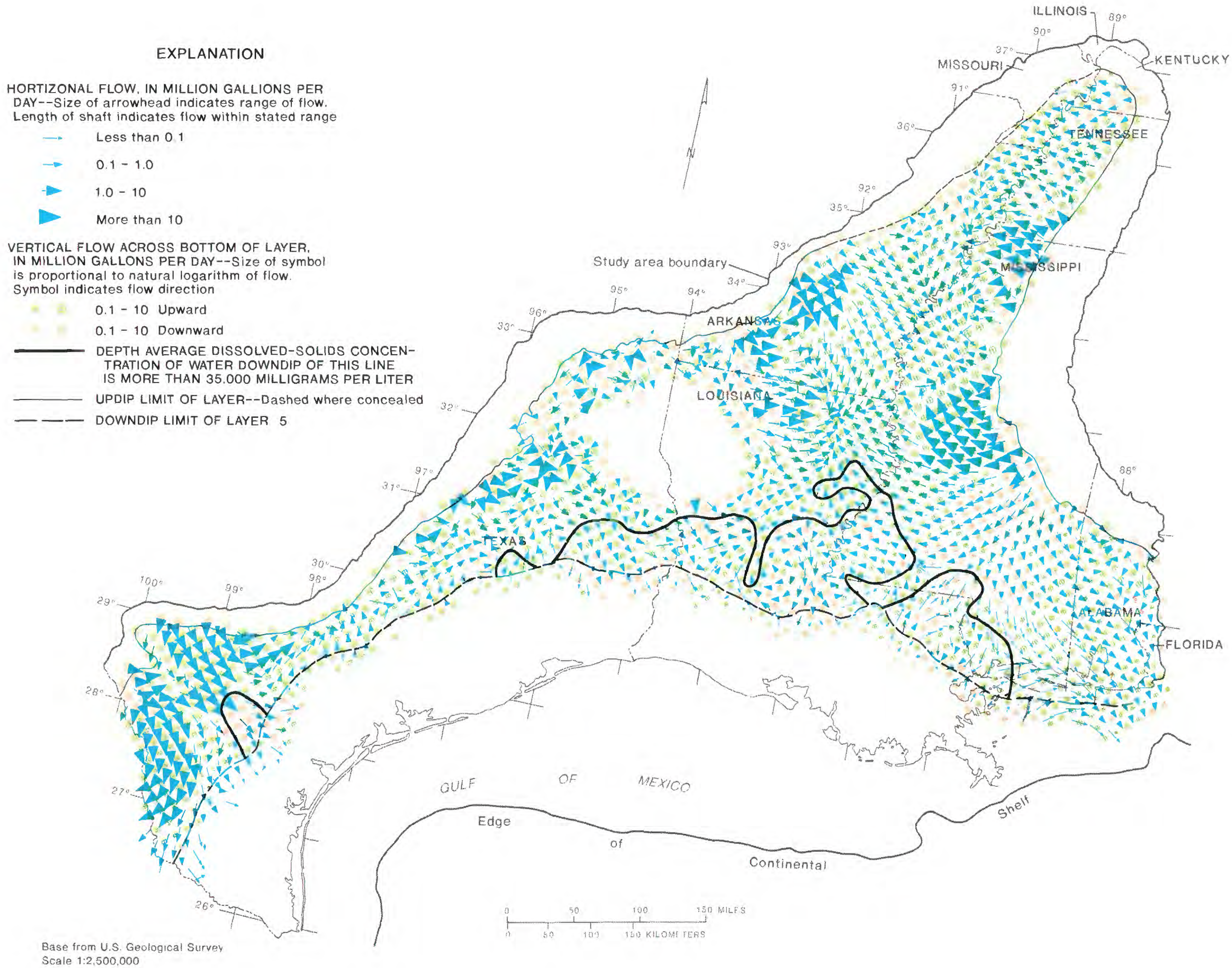


Figure 26d.--Simulated predevelopment regional ground-water flow directions and relative magnitudes in model layer 5 (middle Claiborne aquifer).

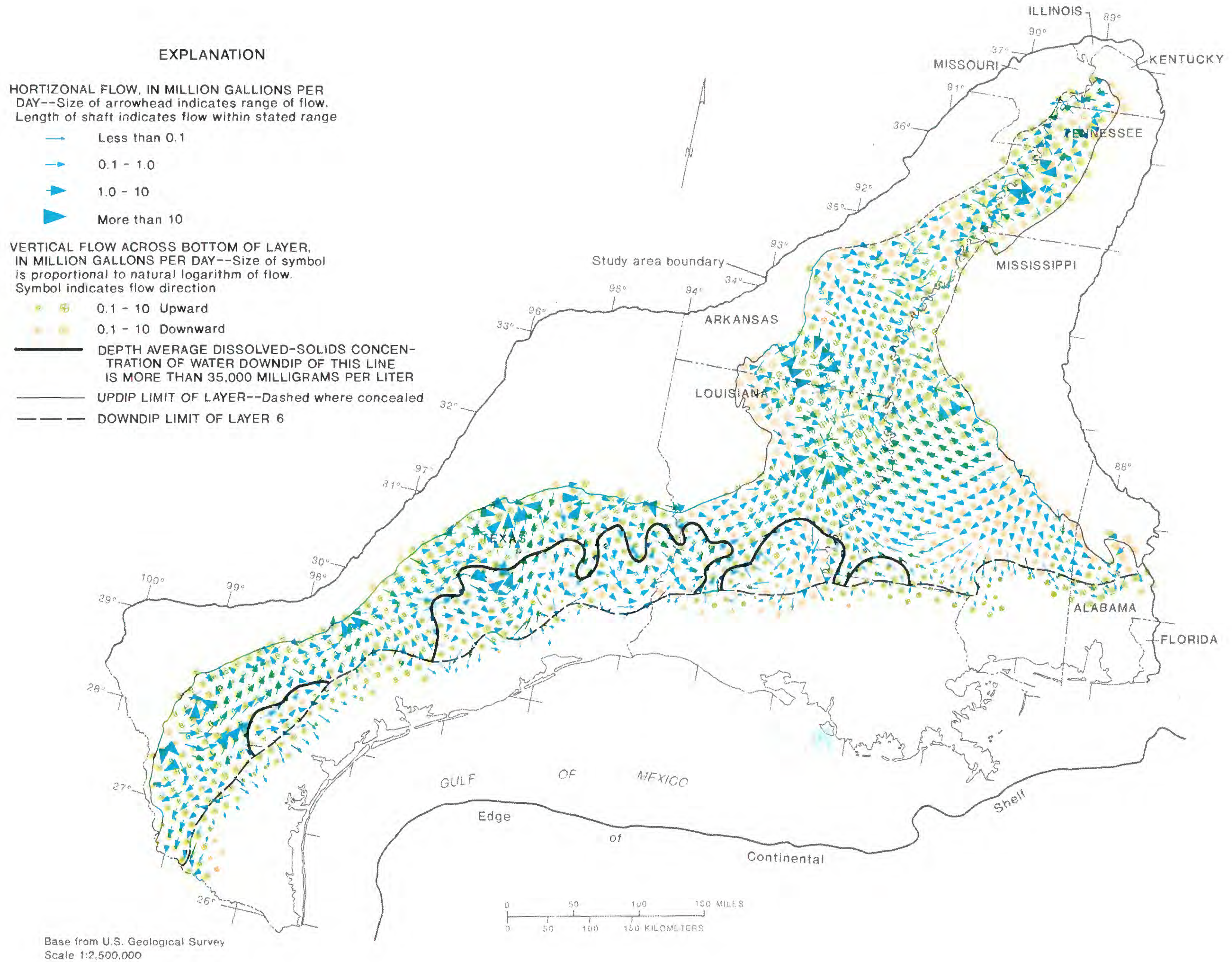


Figure 26e.--Simulated predevelopment regional ground-water flow directions and relative magnitudes in model layer 6 (upper Claiborne aquifer).

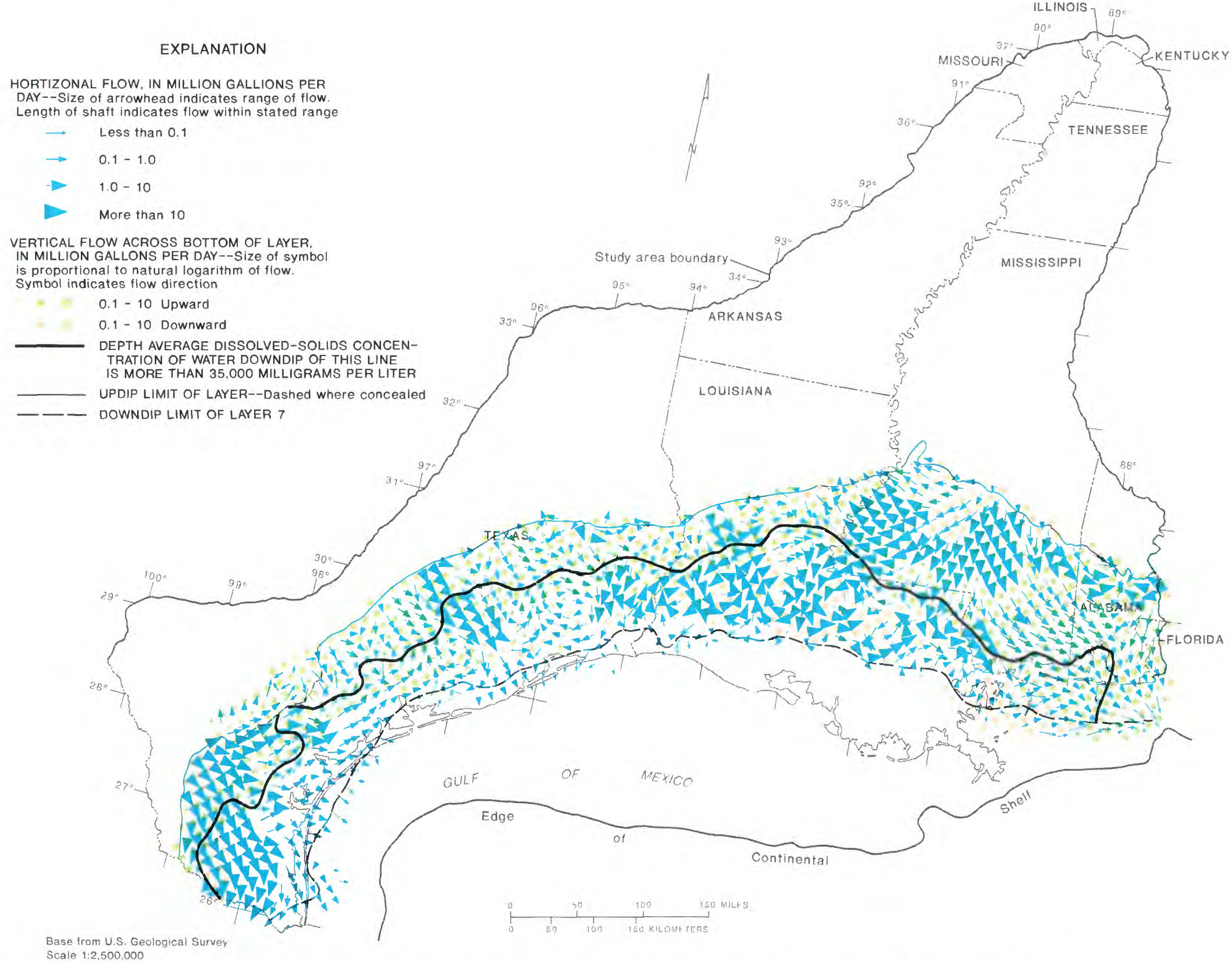


Figure 26f.--Simulated predevelopment regional ground-water flow directions and relative magnitudes in model layer 7 (permeable zone E, lower Miocene-upper Oligocene deposits).

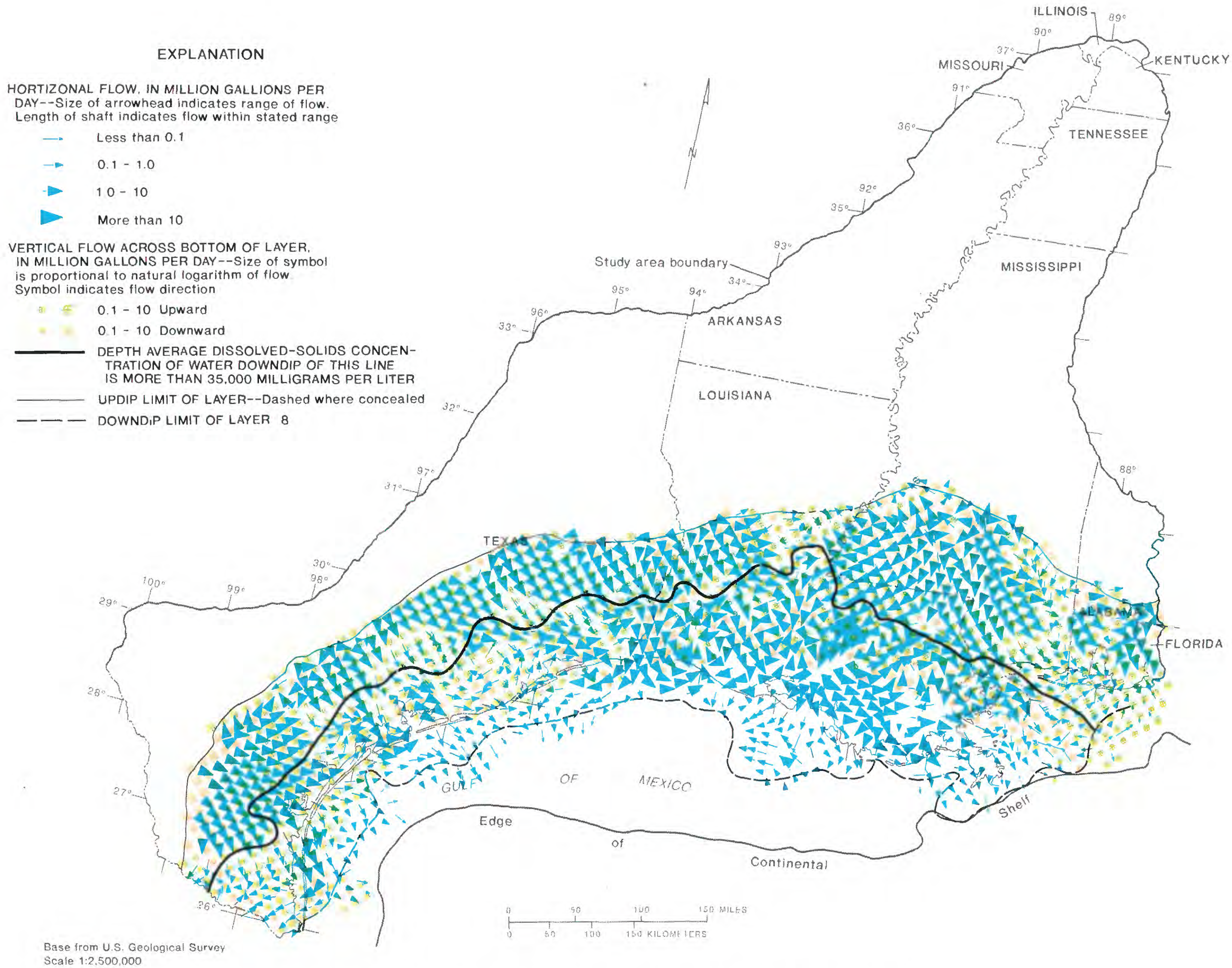


Figure 26g.--Simulated predevelopment regional ground-water flow directions and relative magnitudes in model layer 8 (permeable zone D, middle Miocene deposits).

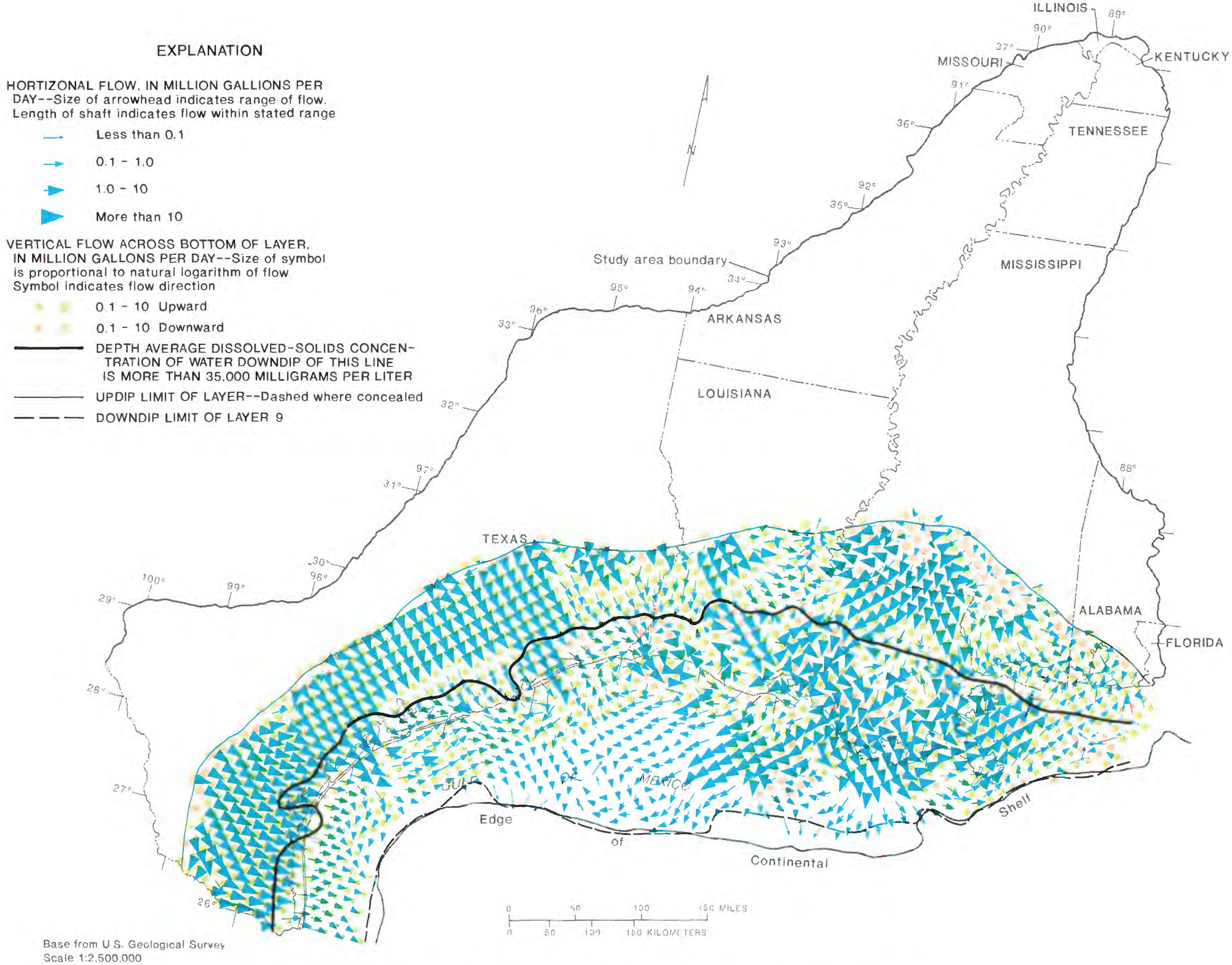


Figure 26h.--Simulated predevelopment regional ground-water flow directions and relative magnitudes in model layer 9 (permeable zone C, lower Pliocene-upper Miocene deposits).

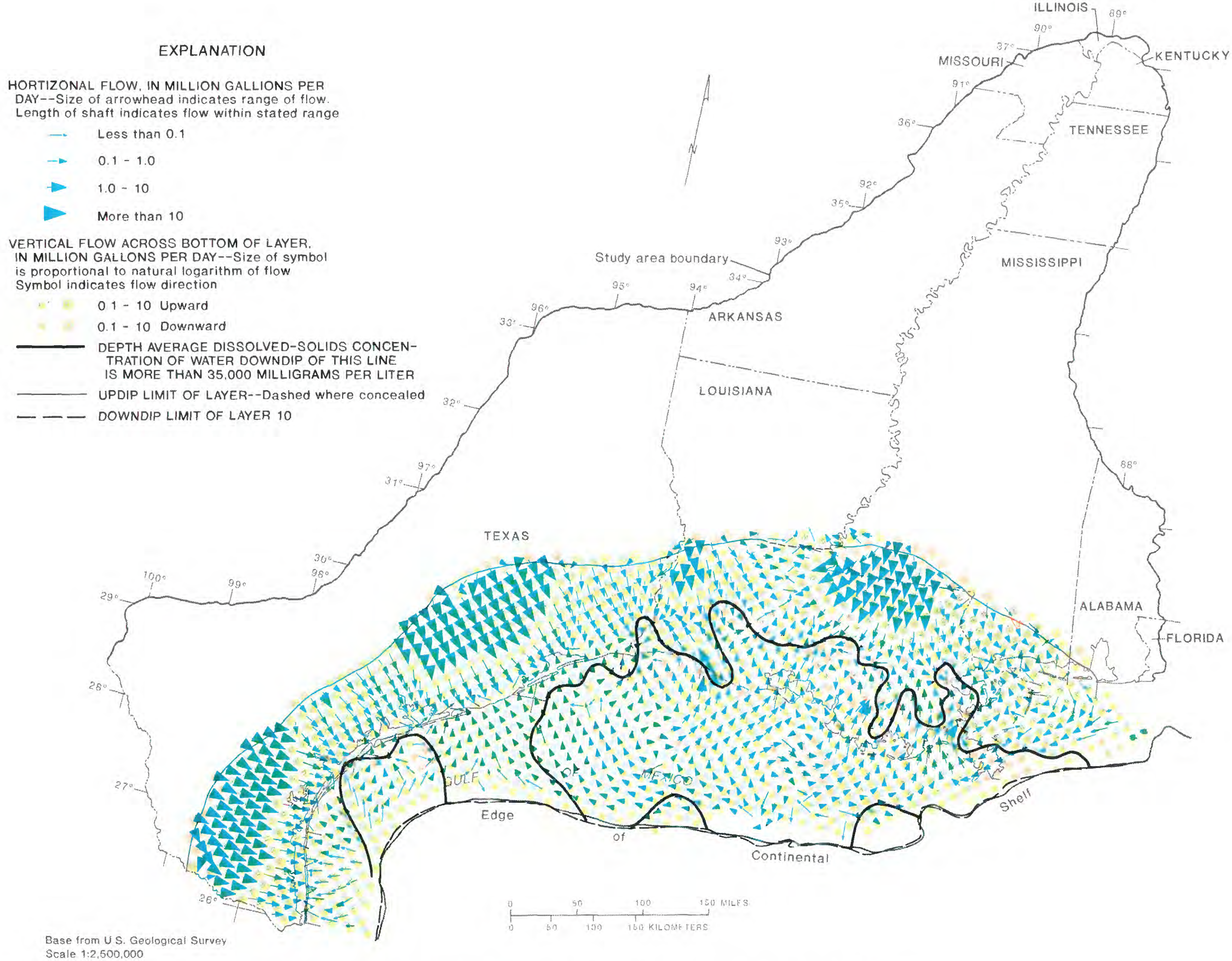


Figure 26i.--Simulated predevelopment regional ground-water flow directions and relative magnitudes in model layer 10 (permeable zone B, lower Pleistocene-upper Pliocene deposits).

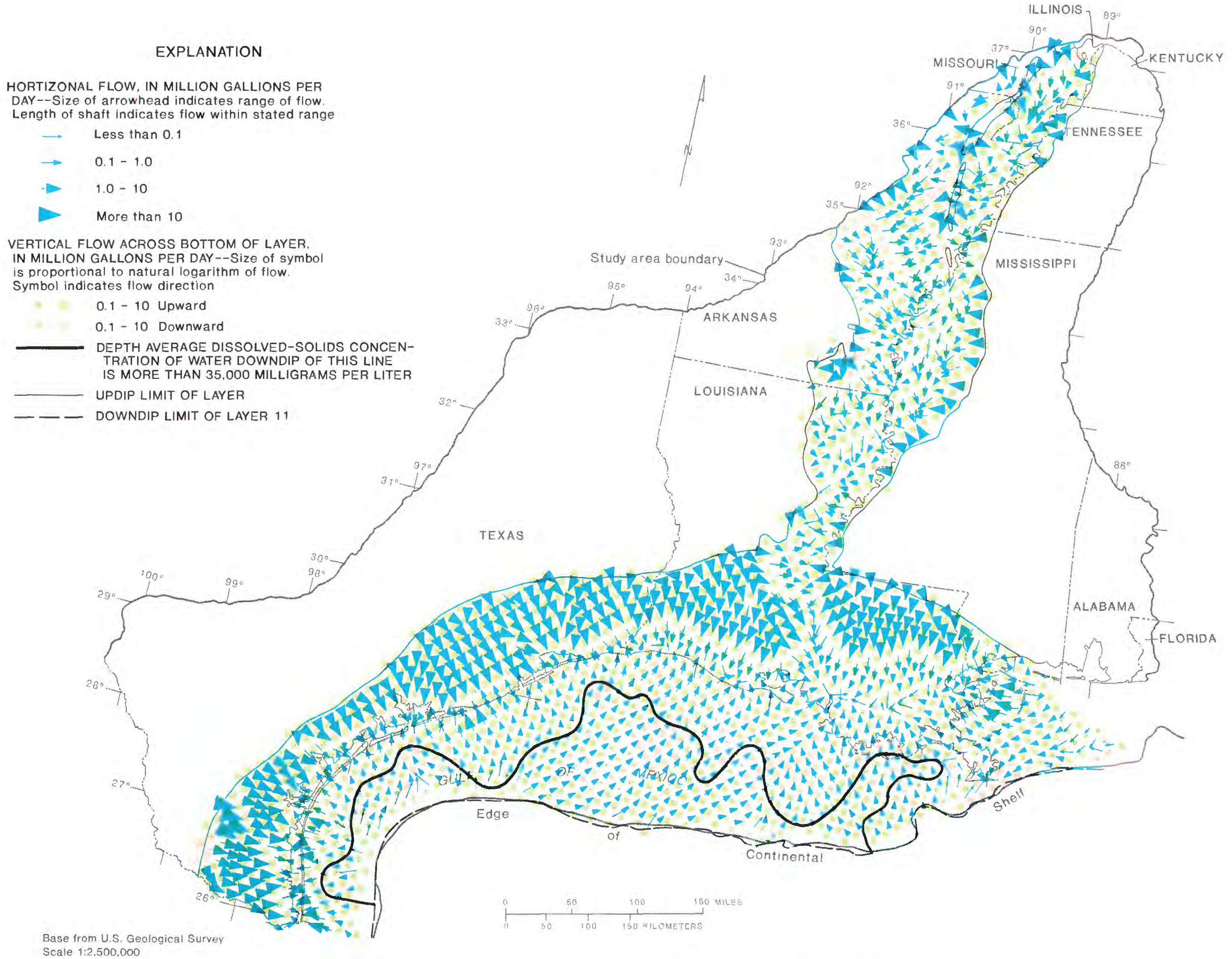


Figure 26j. --Simulated predevelopment regional ground-water flow directions and relative magnitudes in model layer 11 (Mississippi River Valley alluvial aquifer and permeable zone A, Holocene-upper Pleistocene-upper deposits).

One of the predominant flow paths throughout the aquifers of the Mississippi embayment aquifer system is from a major recharge area in south-central Mississippi southwestward and then curving northwestward to a discharge area in northeast Louisiana. In layer 6 this flow dips underneath the Vicksburg-Jackson confining unit before it is forced updip toward the less restricted pathway to the low-lying discharge area. Layer 6 subcrops beneath a wide area of the Mississippi River Valley alluvial aquifer and becomes a conduit from recharge areas to discharge areas. Throughout most of the areal extent of layer 6, the interaction with underlying aquifers is upward flow into the layer. In the northern part of the Mississippi embayment aquifer system, the dominant flow pattern in the layers that exist there is from east to west toward the flow axis of the embayment which is west of both the geological axis and the Mississippi River. In the Texas coastal uplands aquifer system, two major flow paths are toward the Rio Grande and toward the Nueces River.

Predevelopment Recharge and Discharge

The total amount of predevelopment regional recharge (equal to discharge) was about 1.9 Bgal/d. Recharge exceeded discharge in about one-half of the model area, while discharge exceeded recharge in the other half (fig. 27). The area with net regional recharge was about one-half of the total area and the unit net recharge averaged about 0.3 in/yr. This amounts to less than 0.5 percent of the total surface water flow in the study area (466.6 million acre-ft/yr or 523 Bgal/d from a drainage area of more than 1.5 million mi^2, table 1).

The two most important factors controlling predevelopment regional recharge and discharge (fig. 27) are topography and the outcrop pattern and geometry of permeable zones and confining units. Other factors which have been suggested to play a major role, such as geologic structure and precipitation, appear to play a minor role in the Gulf Coastal Plain. This conclusion agrees with the work of Freeze and Witherspoon, (1967). The major variations in subsurface permeability have been accounted for simply by delineation of aquifers and confining units (Grubb, 1986).

The topography (fig. 3) has a major effect on recharge, discharge, and ground-water flow. In a relatively humid environment where the aquifer system does not have very large transmissivity or steep gradient or both, aquifer systems are generally full-to-overflowing with ground water. The recharge map (fig. 27) shows two areas of the coastal lowlands aquifer system with the largest regional recharge; one in south Texas and one in the northern part of the "boot" in eastern Louisiana. These two areas are relatively large, topographically high areas adjacent to low areas where the flowpath is entirely in one layer. The flow paths are short, the gradients steep, and the resistance to flow is relatively small, partly because the vertical component of flow occurs all in one layer.

Although there are many large structural features like the Jackson dome, the Desha basin, and the San Marcos arch, which affect the regional geology, their effect on the predevelopment regional ground-water flow was not as significant as the configuration of the water-table and the subsurface permeability (Grubb, 1986). This can also be seen by comparing figure 27 with figures 8, 11, and 21.

The amount of recharge to the regional ground-water flow system was also probably not related to the average annual precipitation (fig. 4) over most of the study area because of the relatively humid climate. If regional ground-water recharge was controlled by the amount of precipitation, the depth to the water table should also be related to precipitation. This was tested by several types of regression analyses which indicated no relationship between the observed depth-to-water in measurements

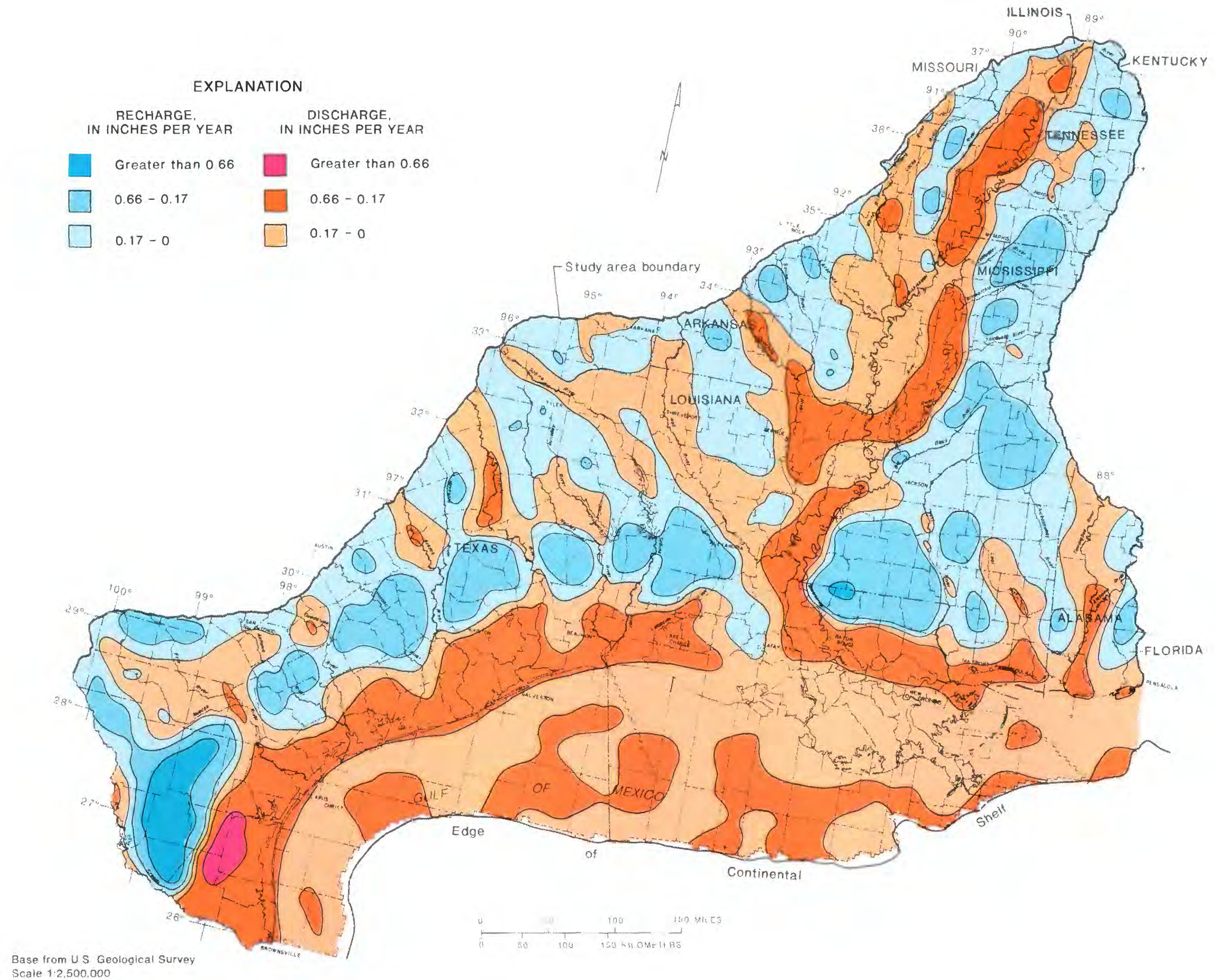


Figure 27.--Predevelopment regional ground-water recharge and discharge.

before 1960 from about 7,000 shallow wells and mean annual precipitation. It is more likely that the amount of precipitation which recharges the regional flow system was limited by the capacity of the flow system to carry ground water away from recharge areas to areas where it can discharge. Where the regional ground-water flow capacity was limited by flat regional hydraulic gradients or by the resistance to flow of the aquifer system, most of the water which infiltrated from precipitation as potential regional recharge was discharged to local surface-water bodies such as creeks and streams, or to evapotranspiration. In the extreme southwestern part of the area where the climate is much drier, recharge to the regional ground-water flow system probably was limited by the lesser precipitation, most of which falls in the hot summer months when much of it is quickly evaporated or transpired.

Distribution of Recharge and Discharge

The most significant regional recharge areas are the hills east of the Mississippi River Alluvial Plain, and some higher areas in a band about 75 mi inland from the coast where the distance between major river valleys is largest (fig. 27). One of the areas of largest simulated regional recharge was in south Texas in the coastal lowlands aquifer system where the precipitation is less. Although it is possible that the smaller recharge actually occurred due to less precipitation in this area, the vertical and horizontal hydraulic conductivity would have to be decreased in this area to simulate less regional recharge. The presence of caliche soils may also be evidence of a smaller vertical hydraulic conductivity in this area.

In the Mississippi embayment aquifer system, the predominant regional flow pattern was from recharge areas on the east side of the embayment to discharge areas in the Mississippi River Valley alluvial aquifer which is mostly in the central and western parts of the embayment. This flow pattern was caused by the topography (fig. 3) which is low and flat across the Mississippi River Valley alluvial aquifer with uplands covering most of the east side of the embayment. The topography of the Mississippi embayment is asymmetrical, in that the valley lies to the west side of the embayment and the topographically higher hills are mostly to the east side. The dominant feature of the topography is the flat, low-lying Mississippi River Alluvial Plain. The river itself, which is an important feature of the hydrologic system because of its size, generally traverses the east side of the plain. Some recharge was simulated on the western side of the embayment in the uplands south of the Arkansas River and on Crowley's Ridge, a 1- to 3-mile-wide ridge extending north to south about 100 mi in the north-central part of the Mississippi River Alluvial Plain.

In the Texas coastal uplands aquifer system, the predominant regional flow pattern was from recharge areas in bands perpendicular to the coastline between major river valleys to discharge areas along the valleys. The Vicksburg-Jackson confining unit covers a larger percentage of the Texas coastal uplands aquifer system than it does in the Mississippi embayment aquifer system. This had a significant effect on the flow pattern in addition to the effect of topography. The Vicksburg-Jackson confining unit impeded upward flow out of the downdip sections of the Texas coastal uplands aquifer system, effectively restricting the longer flow paths in the downdip direction. This is demonstrated by the fact that freshwater in layer 6 becomes saline close to the edge of the Vicksburg-Jackson confining unit except in one area of southwestern Mississippi (fig. 26e). There, the fresh ground water flowed downdip under the Vicksburg-Jackson confining unit and then turned to the northwest to flow updip from under the confining unit to discharge to the Mississippi River Valley alluvial aquifer in an area of low topography in northeastern Louisiana.

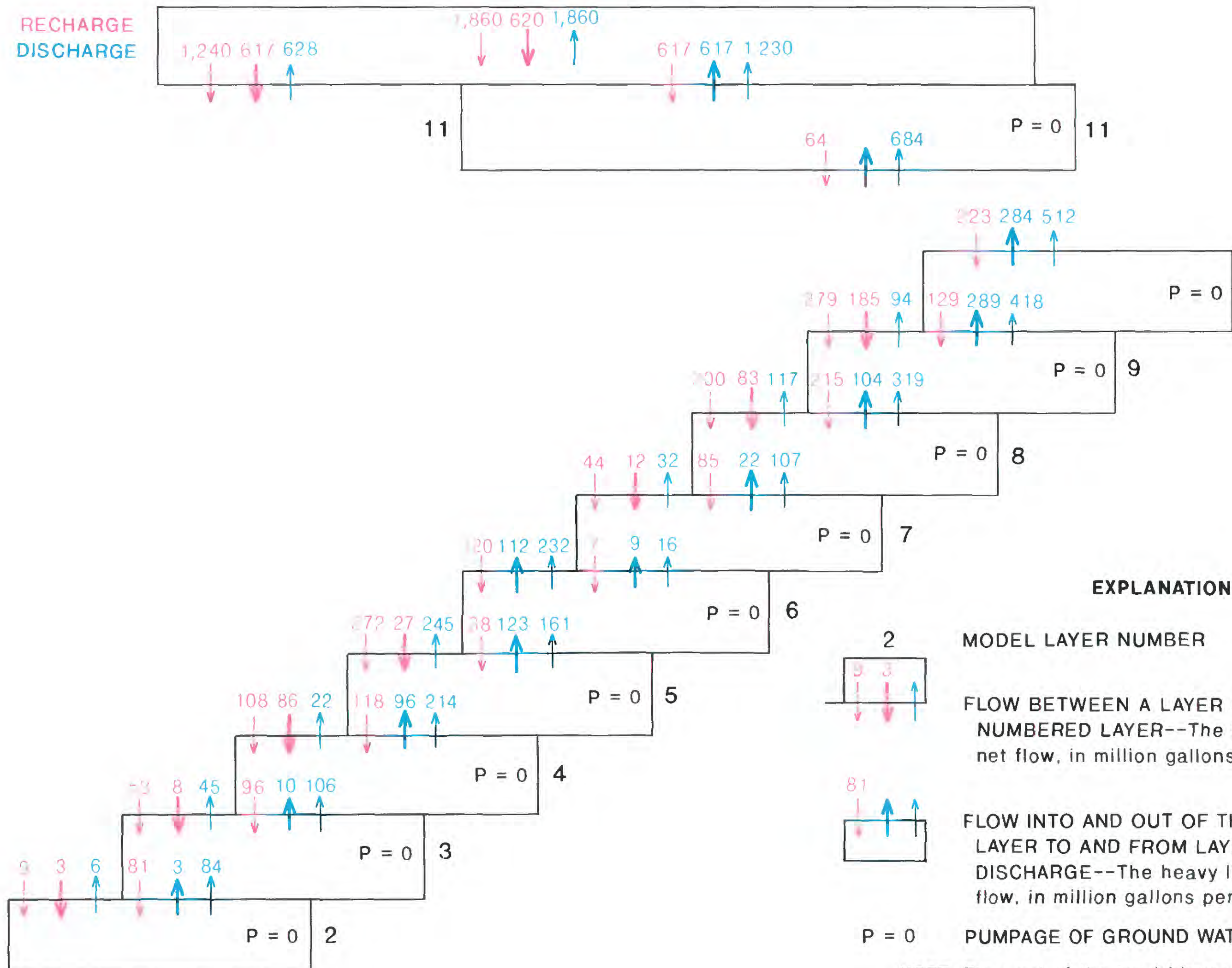
In the coastal lowlands aquifer system, the predominant flow pattern was from recharge areas in the updip and topographically higher areas along the inland edge of the system to discharge areas onshore towards the coast. The largest regional recharge area in the study occurred in southwestern Mississippi where the relatively high topography is a major driving force inducing regional recharge which flows downward to underlying layers and radiates horizontally in nearly all directions (figs. 26b,c,d,e). The location of maximum discharge occurred in a band parallel to the coastline, but inland. Almost all of the discharge occurred before the flow reaches the coast, except in southern Mississippi and Alabama, where higher ground extended closer to the coast, providing steeper gradients in the ground-water system causing some fresh ground water to discharge offshore. This is confirmed by the fact that fresh ground water occurs beneath some islands offshore from southern Mississippi, but does not occur offshore further to the west along the rest of the coastline in the project area.

Vertical Flow Between Model Layers

Layer 11, the Mississippi River Valley alluvial aquifer in the Mississippi embayment aquifer system and zone A (Holocene-upper Pleistocene deposits) in the coastal lowlands aquifer system, received as vertical flow from below nearly one-half of the total predevelopment ground-water discharge from all three aquifer systems (fig. 28a). Note that the individual flows across block faces in the model represent net flows across that particular face of the block; resulting from the sum of positive and negative flows across the area simulated by a block face which occur in the real aquifer system. The values in figure 28 were obtained by separately adding the negative and positive flows across all of the model block faces in a given layer. The net values equal the difference between the sum of the positive and the sum of the negative values.

In the coastal lowlands aquifer system, all of the net flows between layers were upward. This agrees with the overall pattern of recharge in the outcrop area, flow downdip, and discharge up through the overlying layers. This general flow pattern differs in places in other aquifer systems, especially in the Texas coastal uplands aquifer system, where the fresh ground water occurs only in narrow bands near the outcrop of each permeable zone. There are several examples of this in layer 5 in east Texas (fig. 26d). Where the outcropping layer occurs in highland areas, fresh ground water moves downdip and then flows parallel to the outcrop though slightly downdip from the outcrop, and finally moves updip to discharge in an area of the outcrop where the topography is low. An example of this in the Mississippi embayment aquifer system occurs in layer 6 in southwestern Mississippi and adjacent parts of Louisiana (fig. 26e).

Very little, less than 3 Mgal/d, leakage occurred upward through the Vicksburg-Jackson confining unit from the Mississippi embayment aquifer system and Texas coastal uplands aquifer system into the coastal lowlands aquifer system. This is due mostly to the large thickness and small vertical permeability of the confining unit. Some vertical flow does occur, though, between aquifer systems across the Vicksburg-Jackson confining unit; thus, it is not an impermeable barrier to flow as might be suggested by its thickness and lithology. The head gradient in layer 6 trends downdip from the outcrop, underneath the overlying confining unit, indicating flow away from the outcrop discharges up through the confining unit. This situation also exists below the Midway confining unit at the base of the aquifer systems. However, the effect of this upward leakage on the flow system above is much less than on the flow system below because at most locations the total flow in the upper system is much larger than flow in the lower system. This system therefore operates similarly to the regional confining units above the Cretaceous Dakota Sandstone discussed by Bredehoeft and others (1983).



EXPLANATION

2 MODEL LAYER NUMBER

FLOW BETWEEN A LAYER AND THE NEXT HIGHER NUMBERED LAYER--The heavy lined arrow represents net flow, in million gallons per day

81 FLOW INTO AND OUT OF THE TOP OF A PERMEABLE LAYER TO AND FROM LAYER 11 AND RECHARGE/ DISCHARGE--The heavy lined arrow represents net flow, in million gallons per day

P = 0 PUMPAGE OF GROUND WATER--In million gallons per day

NOTE: Flow out of the top of layer 10 and bottom of layer 11 are not separated as to flow into and out of each other versus other layers

Figure 28a.--Diagram showing vertical flow between model layers and recharge and discharge for predevelopment conditions.

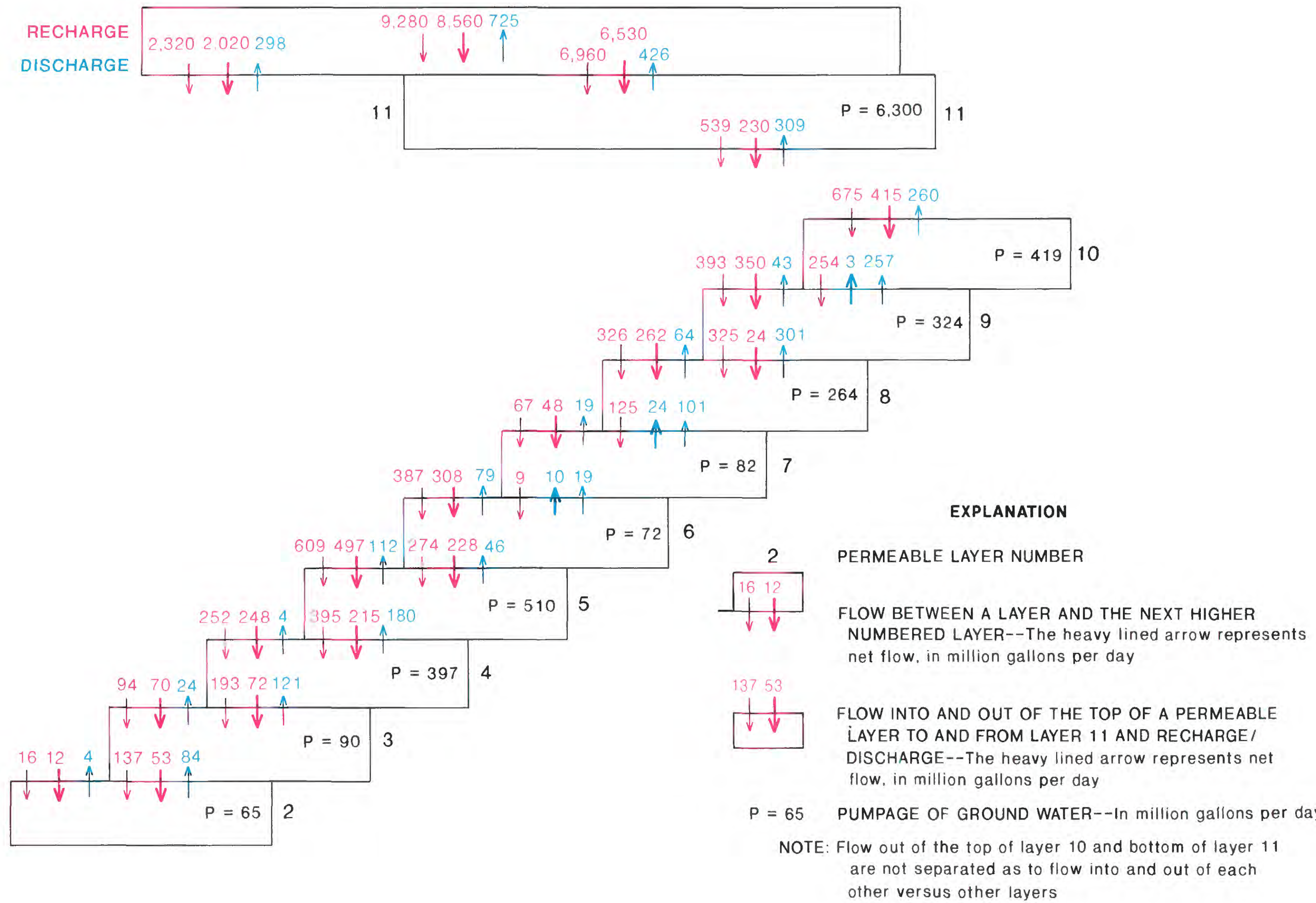


Figure 28b.--Diagram showing vertical flow between model layers and recharge and discharge for the steady-state simulation using 90 percent of the 1980 pumpage.

In the Texas coastal uplands aquifer system and Mississippi embayment aquifer system, only layers 4 and 5 had significant net flow upward to the layer above. The net flow upward out of layer 6 occurred mostly where it lies beneath and discharges to the Mississippi River Valley alluvial aquifer (layer 11). Layer 4 is the only layer with significant net recharge in its outcrop and subcrop areas, probably because it has a larger area of outcrop at a higher altitude than the other layers and has a relatively large horizontal hydraulic conductance. There was very little net flow between layers 2, 3, and 4, and layers 2 and 3 have very little net recharge.

Effect of the Approach to Layering on the Magnitude of Simulated Recharge and Discharge

Several approaches to subdividing the sediments of the coastal lowlands aquifer system into model layers were considered. One was to have a shallow surficial layer which covered the entire aquifer system, representing the top several hundred feet of the aquifers. Using the same horizontal and vertical conductivities, this approach allows a much easier flow path from recharge areas to discharge areas because it eliminates the resistance to vertical flow downdip in a layer. This had the effect of allowing much more recharge than was considered reasonable (in some cases approaching the value of annual precipitation). This was one of the reasons for abandoning this layering method in favor of one which followed the natural dip of the bedding planes in the aquifer system.

Effects of Density

The density of the saline waters in the deeper parts of the aquifer system probably has a significant effect on the ground-water flow in the system--especially in the saline part of the system, but also in the freshwater section. Two sets of simulations confirmed this effect. One compared two simulations which were identical, assuming steady-state 1980 pumpage conditions, except that one simulation had the density for all the model blocks set to 1.0 (the density of freshwater). Even at Memphis, which is about 200 mi updip of the limit of freshwater in layer 5, the difference in simulated head was nearly 1 ft. Assuming the entire aquifer system is composed of freshwater, there would be less drawdown caused by the pumping; an effect similar to moving a limited flow boundary further away.

The distribution of salinity has a high degree of local variability; consequently, the relatively sparse data available for this analysis has quite a bit of variability compared to the regional trend that is still discernable. To test what effect this local variation had on the flow simulation, the density datasets for each layer were smoothed using an areal moving average, and an otherwise identical simulation was made. Figure 29a and b demonstrate the degree of smoothing applied to the density data for layer 8. Although the small-scale variability of the direction of flow was reduced by the density smoothing, the regional flow patterns remained largely unchanged (figs. 26g and 30). Layer 8 was chosen to demonstrate this effect because it has some of the largest volumes of simulated flow in the brine area.

Predevelopment simulations indicate that significant quantities of flow were occurring in the saline part of the aquifer system (fig. 26g). Assumptions about how permeability varies with depth of burial, as a result of compaction, affects the model's sensitivity to density because most dense water occurs at depth. The amount of change of the permeability with depth is uncertain, yet it is certain that it decreases with depth due to loading and compaction, possibly as much as several orders of

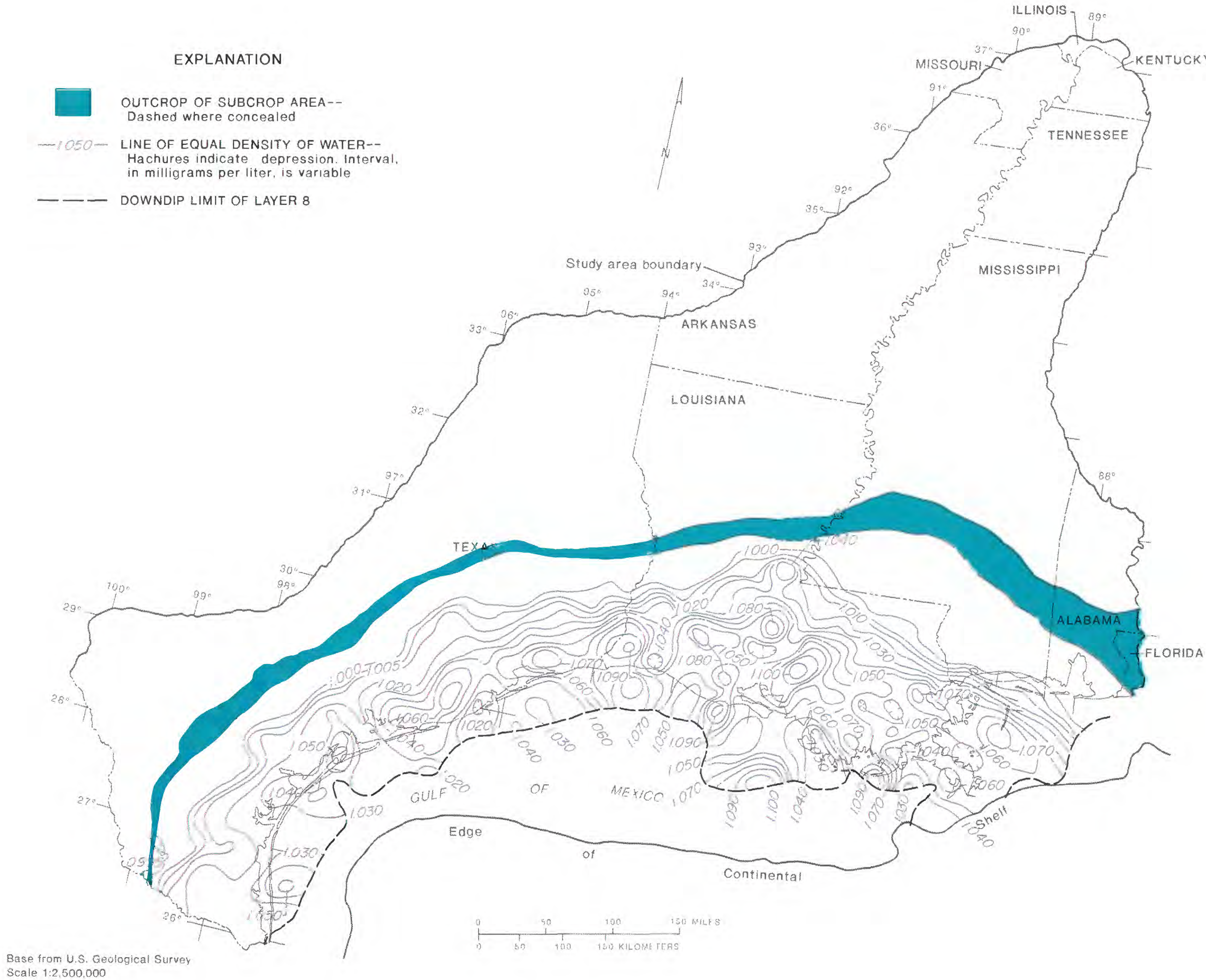


Figure 29a.--Density of water in model layer 8 (zone D of the coastal lowlands aquifer system) estimated from electric logs.

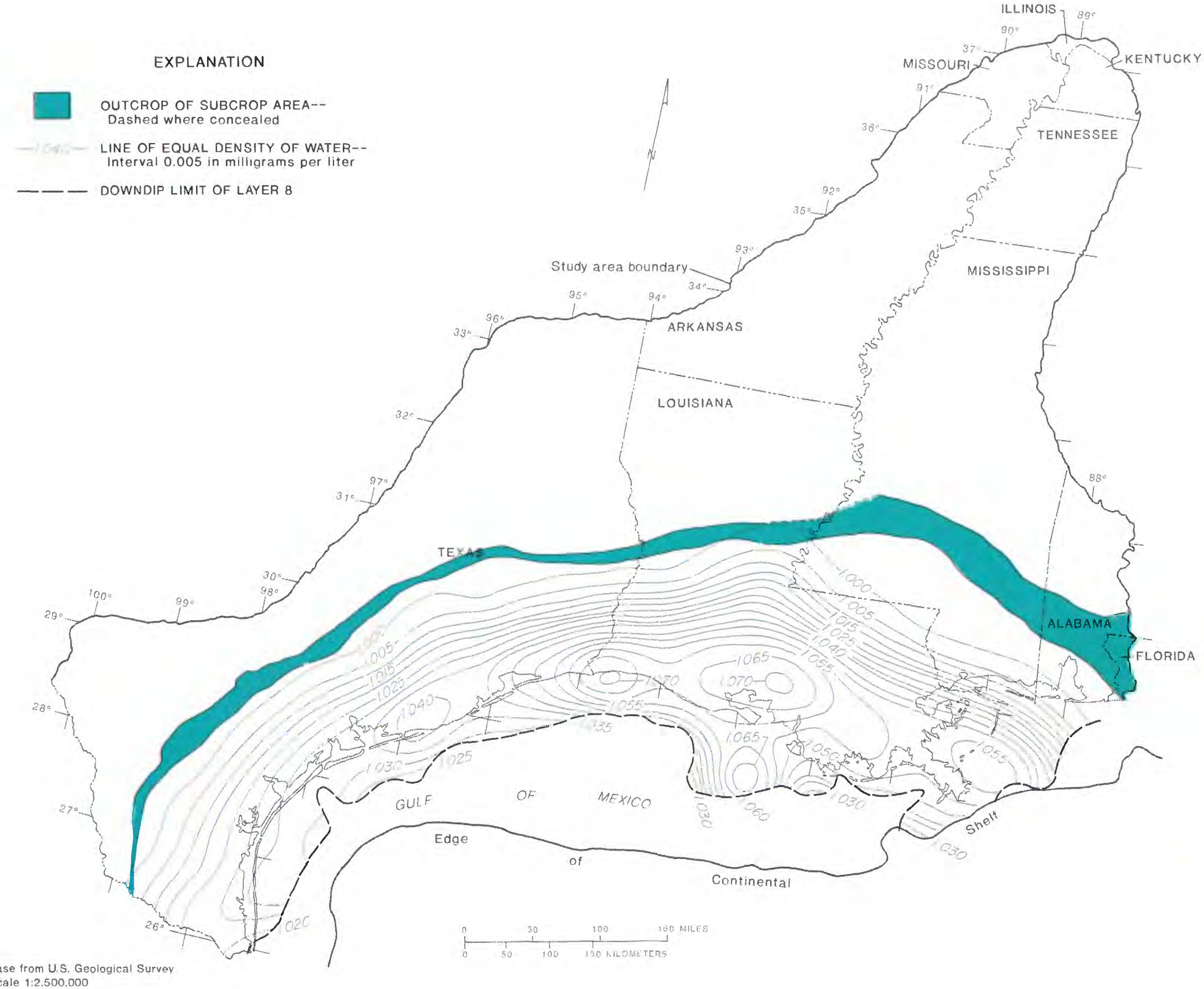


Figure 29b.--Density of water in model layer 8 (zone D of the coastal lowlands aquifer system) from smoothed estimates.

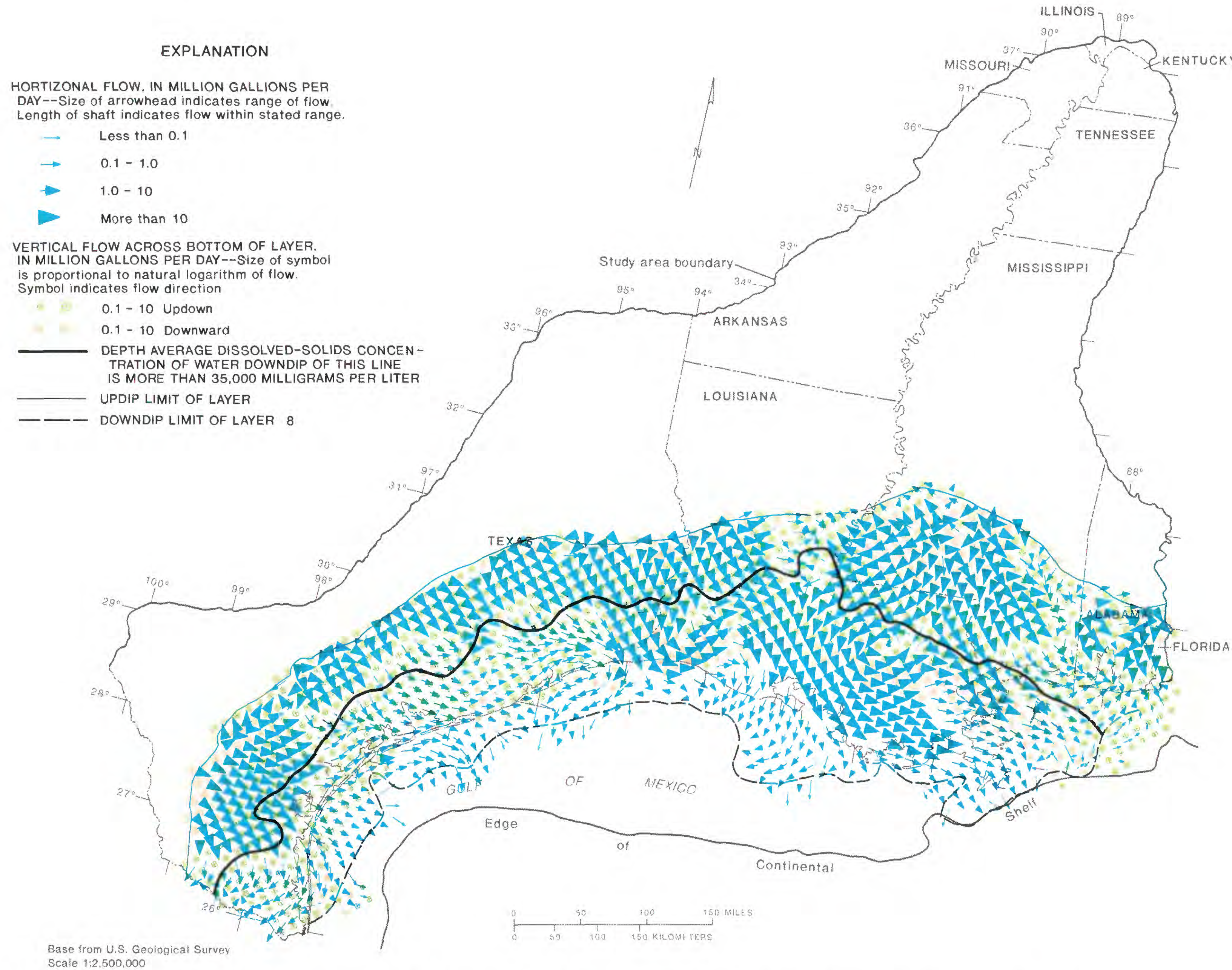


Figure 30.--Simulated predevelopment ground-water flow directions and relative magnitudes in model layer 8 (zone D of the coastal lowlands aquifer system) using smooth densities.

magnitude for fine-grained sediments at depths approaching the bottom of the system studied. For the simulations presented in this report, the decrease of viscosity with increasing depth (and hence increasing temperature) was assumed to balance the decreased permeability of the sediments to give essentially no change in hydraulic conductivity with depth. Hydraulic conductivity may well decrease with depth, decreasing the sensitivity of the model to density, and reducing the flow rates in the brine areas. However, simulations including the effect of compaction on permeability, still show brine areas with significant flow rates and velocities (L.K. Kuiper, written commun., 1988).

There has been much speculation about the source of salt causing wide distribution of the very saline waters found downdip in all waters of the gulf coast aquifer systems. Hanor (1987) has presented geochemical and physical evidence for the existence of density inversions in Gulf Coastal Plain sediments sufficient to drive large-scale convective fluid flow at rates of several feet per year. The inversions are caused in part by the dissolution of salt diapirs and the formation of dense brines at shallow depths. The circulation is thus considered thermohaline in nature. Hanor and others (1986) and Bennett and Hanor (1987) have suggested that these effects have regional significance in Gulf Coastal Plain sediments. Our simulations support this mechanism as a significant factor to distribute salts throughout the gulf coast aquifer systems, even though salt domes occur only in parts of the system. Additional sources of salt that have been given very little consideration are the massive salt formations which occur throughout the entire continental slope, downdip from this study area. These probably have been given little consideration previously because they occur so far offshore, away from the high dissolved solids concentrations much further updip and because of the common, though mistaken, assumption that ground water always flows downdip. The preliminary simulations have demonstrated the very large effects that heat and solutes have on the direction and magnitude of ground-water flow. The continental slope contains a very large source of dissolvable salt, also probably a source of abnormally high temperatures because of the salt, and it is a possible source for salts that have moved long distances coastward and updip.

Effects of Development

The rates (fig. 31) and distribution (fig. 27 and 32) of recharge and discharge were substantially changed by development (Williamson, 1987). These changes show some of the contrast between the predevelopment and 1980 steady-state simulations. The 1980 pumpage is about five times the predevelopment regional recharge. Development and the resulting declines in hydraulic head caused the regional recharge to increase by almost five times to over 9 Bgal/d and regional discharge to decrease to less than 1 Bgal/d, less than one-half of its predevelopment value. Figure 32 shows that in the 1980 steady-state simulation, almost all of the study area was characterized by net recharge as almost one-half the area converted from net regional discharge to net regional recharge due to development. Thus, about 80 percent of the 1980 pumpage was supplied from increased recharge, whereas less than 20 percent was captured natural discharge, and less than 10 percent was from ground-water storage. Most of the area which regionally discharged ground water before development had become a regional recharge area by 1980 to supply water for pumpage.

In the 1980 steady-state simulation, many of the net vertical flows between model layers 2 through 11 changed direction or at least approached zero (fig. 28b). The amount and even sign of the net vertical flows are somewhat sensitive to simulation parameters (such as hydraulic conductivity) chosen since they are a difference between

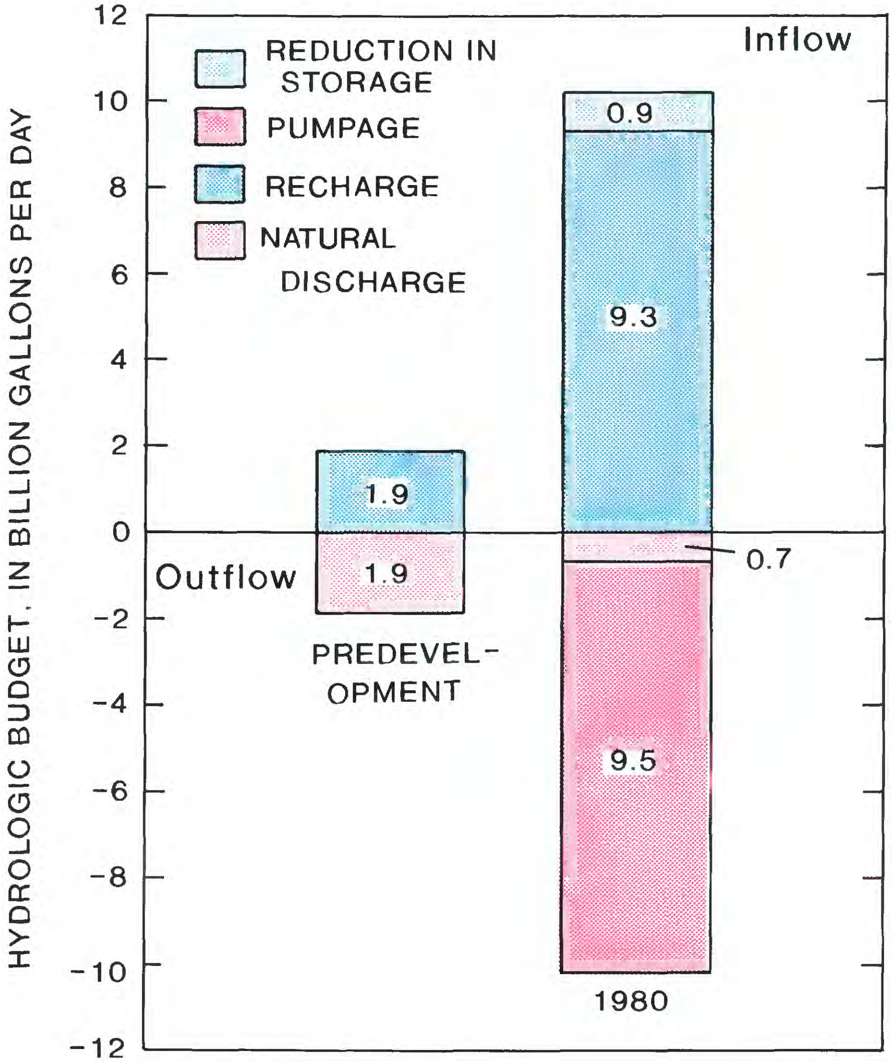
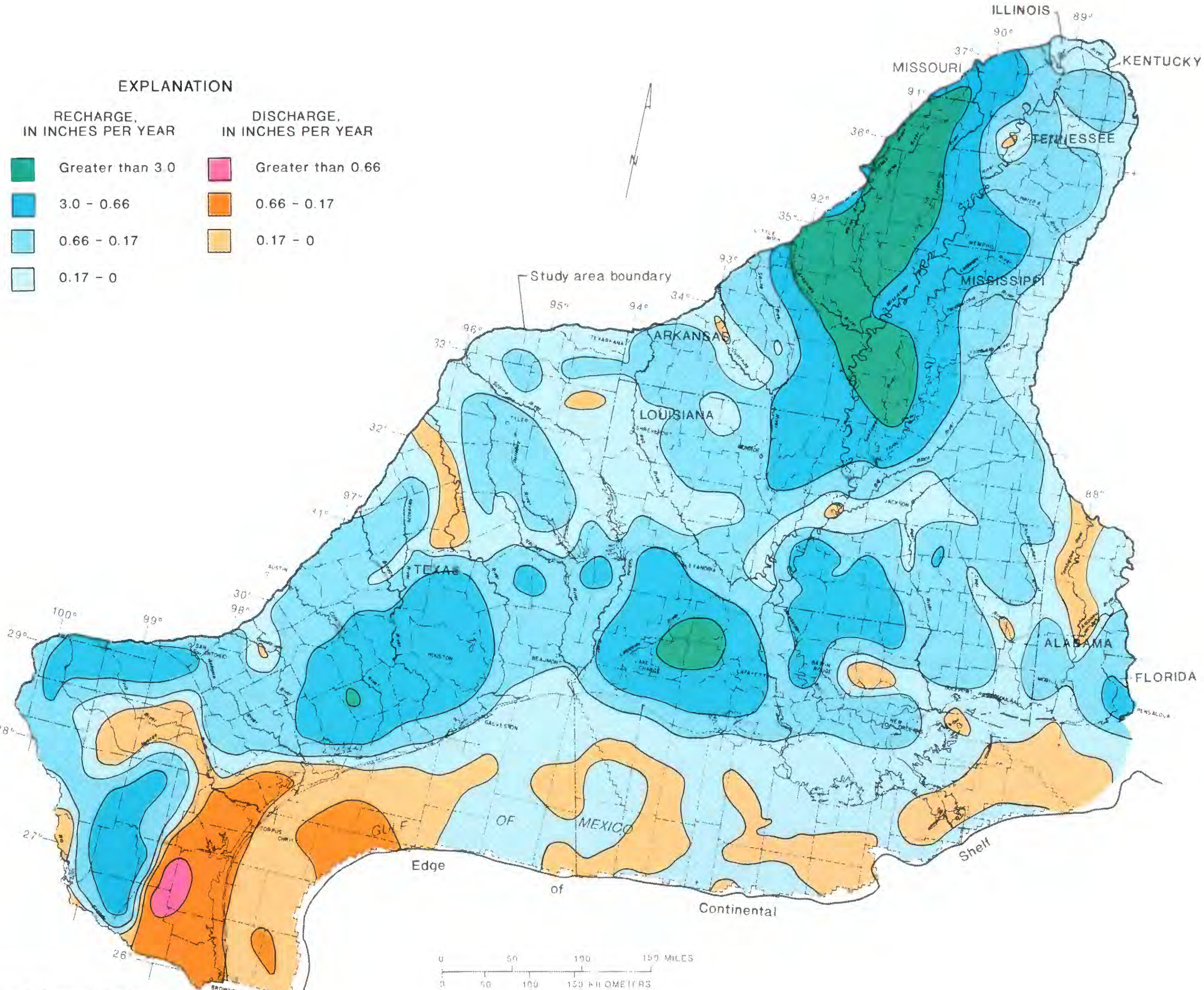


Figure 31.--Diagram showing change in regional ground-water inflow and outflow from predevelopment to 1980.



Base from U.S. Geological Survey
Scale 1:2,500,000

Figure 32.--Regional ground-water recharge and discharge for steady-state conditions using 90 percent of the 1980 pumpage.

two commonly larger numbers. The rates of upward and downward flow, considered individually, are less sensitive to aquifer parameters, especially if considered as relative numbers to compare to other layers or other pumping conditions. Most of the vertically upward flows between permeable model layers remained about the same as before development. Exceptions to this were upward flow into layer 6 which decreased by about two-thirds, and upward flow into and out of layer 10, which decreased nearly in half. Downward vertical flows between layers increased substantially in all of the Texas coastal uplands aquifer system and Mississippi embayment aquifer system, especially into model layers 5 and 4.

Model layers 5 and 8 through 11 have a ratio of the net recharge and discharge in the outcrop to pumpage of between 0.97 to 1.08 (fig. 28b). This indicates that on the average, these layers are self-supporting; in other words, most of their pumpage was supplied from increased recharge and decreased discharge within the layer itself. The ratio for model layer 2, which has a very narrow outcrop and subcrop band (fig. 11 and plate 4) was only 0.22, indicating that much of its pumpage had to be supplied by downward flow from layer 3. The ratio for model layer 6, which has both a very wide outcrop and subcrop band, was 4.3, indicating that it was capturing additional recharge and discharge. Model layer 6 then leaked the excess to supply pumpage in underlying layers, mainly layer 5. Model layer 5 has a large volume of discharge in its outcrop and subcrop area, probably due to the large pumpage in layer 11 which overlies layer 5 in a large area of the Mississippi embayment.

Even though, on the average, only a small proportion of pumping in 1980 is being supplied from aquifer storage, the extent of cones of depression are smaller than they would eventually become if 1980 pumping conditions were continued. This is because, as a cone of depression expands, its area increases greatly (in proportion to the square of the radius), so that a very large volume of aquifer storage is released from just a small change in hydraulic head. This effect can be demonstrated by comparing two simulations (fig. 32). The transient simulation of the aquifer system in 1980 involves some water being released from aquifer storage. The steady-state simulation involves no loss of aquifer storage so that the maximum steady-state drawdown is reached, using only 90 percent of the 1980 pumpage. Note how far offshore the cones of depression extend in figure 33b as compared to figure 33a. In figure 33b most of the offshore area has a drawdown exceeding 25 ft. Also note that the drawdown near the center of the cones is nearly equal.

Vertical Resistance to Flow in the Aquifer System

Resistance to vertical flow caused by many fine-grained beds within the permeable zones can be a significant part of the overall resistance to flow within the aquifer system. This is especially true in the coastal lowlands aquifer system where the regionally mappable confining units contain only a very small part of the total thickness of fine-grained sediments. There is considerable resistance to vertical flow caused by the many thin localized fine-grained beds which are dispersed throughout the permeable zones. The deep cones of depression at Houston and Baton Rouge demonstrate how much vertical resistance occurs. Hundreds of feet of vertical head difference has developed between the water table and the deep, intensively pumped sands (fig. 32).

This concept can be illustrated using the results of previous studies such as Wesselman (1967, p. 57 and fig. 22). He plotted hydraulic head versus depth throughout an interval containing the Chicot, Evangeline, and Jasper aquifers at Evadale, Texas. This location is near the type locality for the Burkeville "aquiclude" or confining

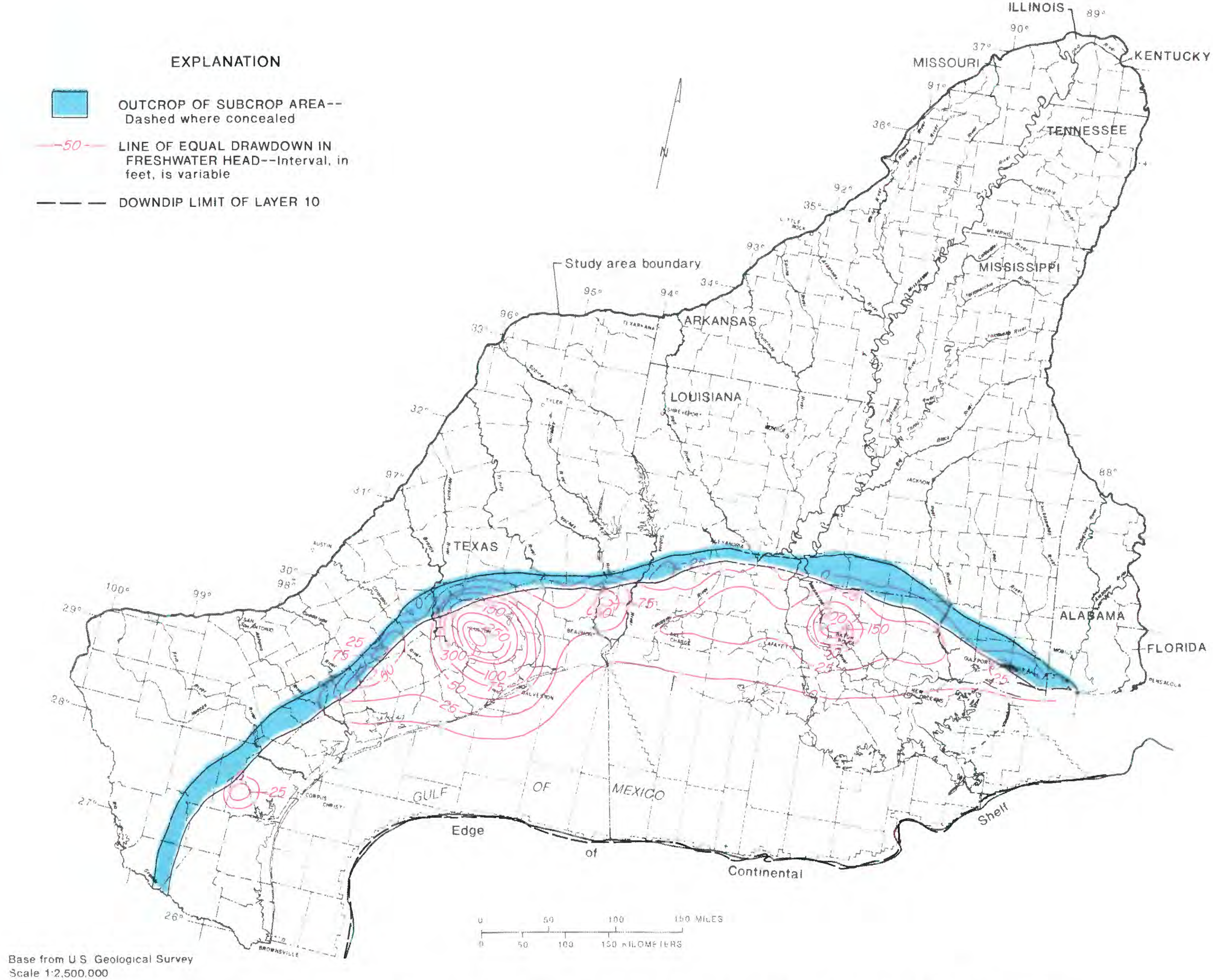


Figure 33a.--Drawdown due to pumpage in model layer 10 (zone B of the coastal lowlands aquifer system containing mostly lower Pleistocene and upper Pliocene deposits) assuming transient conditions in 1980.

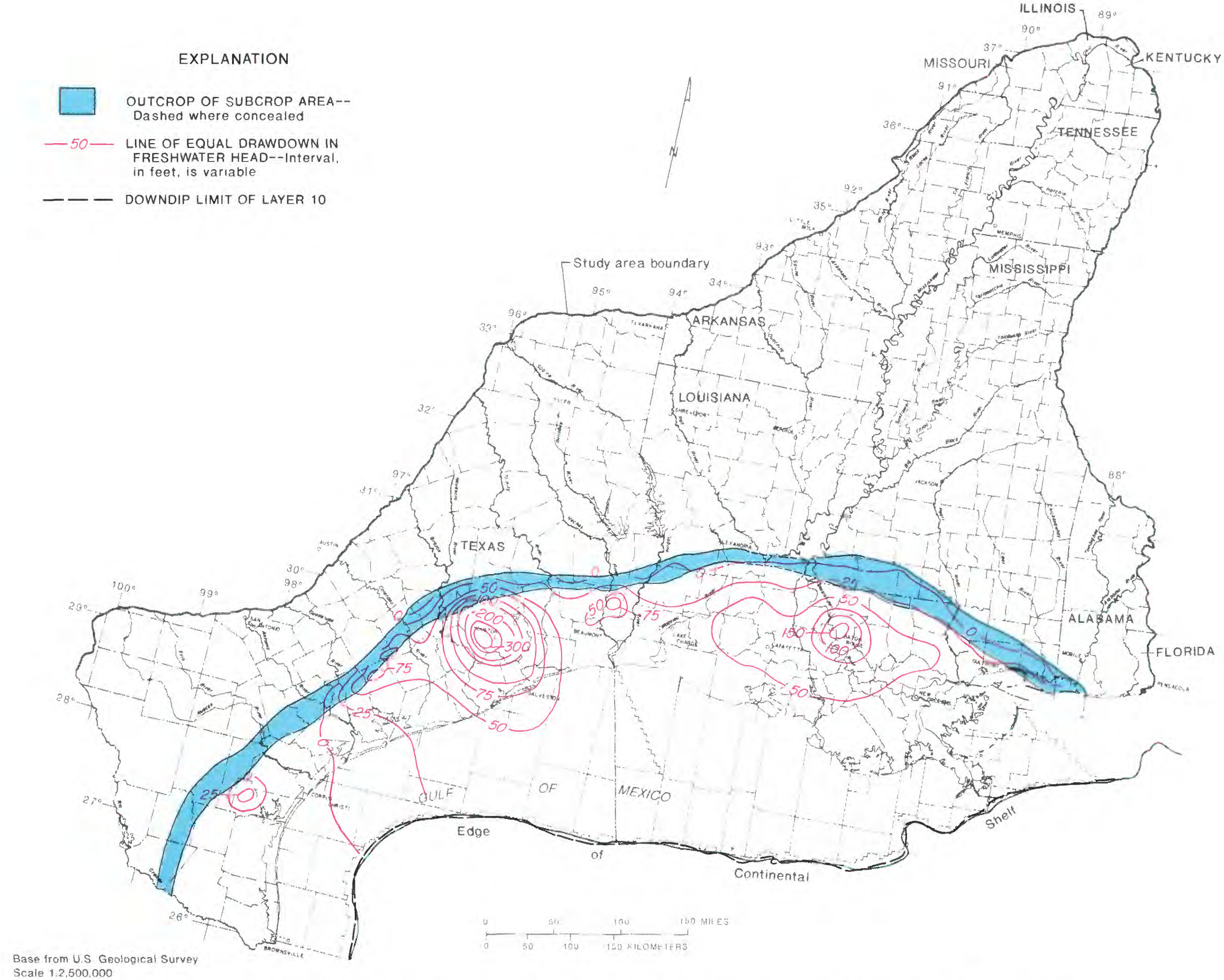


Figure 33b.--Drawdown due to pumpage in model layer 10 (zone B of the coastal lowlands aquifer system containing mostly lower Pleistocene and upper Pliocene deposits) assuming steady-state conditions are reached with 90 percent of the 1980 pumpage.

unit (located stratigraphically between the Jasper and the Evangeline aquifers). The plot shows a head difference of 67.2 ft over a depth interval of approximately 430 ft across the Burkeville, which is a vertical gradient of 0.16 ft/ft. However, on the same plot, a head change of 18.9 ft occurs across 130 ft of an unnamed clay bed in the middle part of the Evangeline aquifer, thus a gradient of 0.15 ft/ft. This similarity suggests that the numerous smaller clays within the aquifer system may provide nearly the same resistance to vertical flows as larger, regionally mappable confining units.

The calibrated vertical hydraulic conductivities for the fine-grained part of the permeable zones was not significantly higher than the calibrated value for the confining units (table 4). More than one-half of the thickness of fine-grained sediments in the aquifer systems is contained within the permeable zones (table 2), therefore, the average effective leakance of the fine-grained sediments (the non-sand part) within the aquifers is very significant when compared to the leakance of some of the regional confining units. This gives further support to the conclusion that many smaller fine-grained beds can have a significant effect on resistance to vertical flow.

NEEDS FOR FURTHER STUDY

A discussion of future direction is included because this is a preliminary report that describes results to date of an ongoing study. This will include ideas about methods which are in the process of being applied. Most of the ideas deal with different approaches to parts of the simulation whereas the last idea presented, ground-water management information using sensitivity to pumping stress and the principle of superposition, proposes a different use of ground-water flow simulations which should be more helpful to water management authorities.

Parameter Estimation

Calibration is an iterative process similar to the one used in parameter estimation models to select the combination of parameter values that give the best fit to available data. Automatic approaches by computer programs have the advantage of being able to test many more combinations and being more likely to find the "best fit" parameters. They have the disadvantage of not being able to recognize unrealistic parameters or combinations of parameters. This disadvantage can be minimized by careful application and by incorporating prior information about the parameters. Work is currently under way using a simple least-squares minimization method (Kuiper, 1986) applied to a transient simulation of the GC RASA (L.K. Kuiper, written commun., 1987). This was originally tried with a horizontal grid spacing of 30 mi to conserve computer time (L.K. Kuiper, written commun., 1986) and some preliminary results from estimation at both scales have been used in the simulations presented in this report.

Refine Hydraulic Conductivity Estimates

Several investigators (Desbarats, 1987, Dagan, 1986, and Fogg, 1986a, 1986b) have recently pointed out the importance and possibilities of estimating effective hydraulic conductivities using a statistical approach to the problem of sand-body interconnectedness. L.K. Kuiper (written commun., 1987) has incorporated the method of Desbarats (1987) into his parameter estimation model. Flow in Kuiper's simulation is not based on estimates of effective horizontal and vertical hydraulic conductivity. Instead, estimates of K_h for coarse-grained sediments and K_v conductivity for fine-grained sediments, the sand percentage, and a factor which is related to the scale of the model and the geometric proportions of typical sand bodies are used along with the equations developed statistically by Desbarats to calculate the effective horizontal and vertical conductivity. The parameter estimation program estimates values for the coarse- and fine-grained conductivities as well as the sand body factor. One major advantage to this approach is that the values used in the simulation for K_h of sand tend to be better correlated with values that are estimated from aquifer tests. This allows more direct comparison so that the range of confidence intervals at a certain probability (for example, 95 percent) could be used in the parameter estimation method as constraints on the selected K_h for estimating the mean for that particular area and layer. Alternatively, the mean aquifer test values for each area and layer could be used in the objective function as prior information. This technique is being refined.

In addition, the effects of depth of burial and consequent loading, compaction, and temperature on permeability and viscosity need to be analyzed. Only limited data on permeability versus depth is available in the literature. This analysis is complicated by the fact that effective vertical hydraulic conductivity, which is controlled mostly by the fine-grained sediments, will vary differently than effective horizontal conductivity, which is primarily controlled by the lateral conductivity of the coarse-grained sediments.

Incorporation and Analysis of Land Subsidence due to Ground-Water Withdrawal

Decreasing hydraulic head due to pumping causes increased loading on the matrix of aquifer material, and can cause compaction and land subsidence if the fine-grained deposits are deformable. This has a great effect on the ground-water hydrology because the effective storage coefficient during inelastic compaction is two or more orders-of-magnitude larger than the elastic storage coefficient of a confined aquifer, hence slowing the head decline by providing a source of water to the pumped wells. Prudic and Williamson (1986) used a technique to simulate a compacting regional aquifer in the Central Valley of California similar to the method of Meyer and Carr (1979) used in the Houston area. Recently, Leake and Prudic (1987, 1988) have enhanced the method to an implicit formulation which is numerically more stable. L.K. Kuiper (written commun., 1987) independently developed a similar technique and is applying it to the Gulf Coast RASA model.

Leakage from the Geopressured Zone and Effect of Deep Fluid Injection

Very high fluid pressures in the geopressured zone force some flow upward into the aquifer system. The rate of flow is thought to be small because of the very low vertical hydraulic conductivity of the predominantly shaly sediments; this is supported by the fact that the high pressures still exist. Hypotheses about the values of conductivities and flows will be tested by simulations, as well as the assumption used in this report that the flow would not be large enough to have a substantial effect on flow in the aquifer system.

Oil field brines and other hazardous wastes are injected into the saline section of the aquifer system. It has always been assumed that the volumes are not large enough to substantially affect upward migration of brine waters into the fresh part of the system. The preliminary evaluation indicates that this assumption could also be tested by simulation. Estimates of the volumes of fluid injected will be required. Also, estimates of withdrawals by oil wells could be used to test the significance of their effect on regional ground-water flow.

Ground-water Management Information Using Sensitivity to Pumping Stress and the Principle of Superposition

Historically, when ground-water development in a particular area expands enough to cause detrimental effects, evaluators of the situation may recommend that pumping be restricted and that alternative supply sources, generally surface water, be sought. Many ground-water studies describe only the effects of existing and proposed water development. Osborne and others (1986) have proven what many have suspected--that water-use forecasts, regardless of the method used, are likely always to be highly inaccurate, due to their dependence on unpredictable economic, social and political factors. Typically, ground-water flow models are used to simulate a few hypothetical future pumpage scenarios, which are difficult to estimate, may never occur, and still may not show where there is capacity for further development with minimum effects. Instead, studies need to contain the information water managers need to determine which areas or permeable zones have potential to supply more water for expanding development with minimum effects. Future studies need to include model simulations for an areal analysis of sensitivity to pumpage by layer. Results of model simulations need to include production of a map for each layer showing the water-level change in each block that would result from a unit increase in pumpage from current conditions. These maps could be prepared on the basis of simulations that ignore storage, which will maximize drawdown and will indicate what the maximum future effects might be. These maps will, therefore, allow a water manager to quickly identify the layers and areas from which pumpage increases need to be discouraged and where additional development can be tolerated or encouraged.

SUMMARY

A major objective of the Gulf Coast Regional Aquifer-Systems Analysis is to apply digital models to regional ground-water flow systems to improve understanding and management of the resource. Modeling is used because it is the best available tool to synthesize most of the known information about the aquifer systems and to test hypotheses about the relative importance of the components and processes of the systems. The study area includes 230,000 mi^2 onshore in the Gulf of Mexico Coastal Plain and 60,000 mi^2 offshore beneath the continental shelf. It includes deposits of Tertiary and younger age, which contain fresh and saline water. A 10-layer, finite-difference, variable-density model, with blocks 10 mi on a side, was used to simulate ground-water flow: before development (around 1900); in 1980, assuming steady-state conditions; and for transient pumping conditions from 1935-80. Model layer subdivisions were defined at regional confining units, where present, or major contrasts in hydraulic conductivity, or where many local fine-grained beds have created large vertical head differences in response to pumping.

Preliminary results indicate that the major factors controlling predevelopment regional flow are the topography and the outcrop pattern, and geometry of aquifers and confining units. Geologic structure and the distribution of precipitation were less significant factors. Prior to development, recharge to regional aquifers occurred in the hills east of the Mississippi River Alluvial Plain and some areas in a band parallel to and about 75 mi inland from the coast where the distance between major rivers is largest. Major regional discharge areas were the low-lying, flat Mississippi River Alluvial Plain, the coastal lowlands, and major river valleys. The largest area where relatively freshwater occurs under the Vicksburg-Jackson confining unit is in southwestern Mississippi where regional flow from the largest regional recharge area in the hills moves downdip under the confining unit. Very little water can escape upward so most of the predevelopment flow moved updip to the west and discharged into the Mississippi River Valley alluvial aquifer.

The variation in density of saline waters in the deeper parts of the aquifer system probably has a substantial effect on the ground-water flow of the system, even up in the freshwater part of the system. Water in the saline part of the system is not static, but is driven by large pressure and gravity forces. Even though the saline-freshwater interface has probably not moved very far during the past few decades, it is probably in motion in many places, at slow velocities. Highly variable water densities resulting from heat and salinity variations appear to be a significant driving force to transport dissolved salts from salt formations great distances in many directions, including updip.

The distribution and rates of regional recharge and discharge were substantially changed by development. The 1980 pumpage, which was about five times the value of predevelopment regional recharge, was being supplied mostly from increased recharge to the regional confined aquifers (about 80 percent), with lesser amounts from decreased natural discharge and aquifer storage. Under predevelopment conditions, the area of regional recharge was about equal to the area of regional discharge. By 1980, about 95 percent of the area onshore was functioning as a regional recharge area. Vertical flows between permeable zones changed substantially in both direction and magnitude during development. Simulation of 90 percent of the 1980 pumpage, assuming no change in storage in the aquifer system, resulted in a reasonable match of simulated and measured 1980 heads. Results also indicate that resistance to vertical flow caused by many smaller fine-grained beds within the permeable zones can be as important as resistance caused by regional confining units.

Needs for further study include:

1. Automatic calibration of the model using parameter estimation with constraints consisting of confidence limits on the mean, based on aquifer test data. The parameter estimation would also include a statistical approach to evaluating effective hydraulic conductivities based on sand body geometry.

2. Enhancing the simulation with other hydrologic processes, such as flow out of geo-pressured sediments, land subsidence, and possibly oil field withdrawals and injections.

3. Evaluating the aquifer system to produce maps for each layer showing suitability for future ground-water development.

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