U.S. Department of the Interior
U.S. Geological Survey

Ground-Water Flow in the Gulf Coast Aquifer Systems, South-Central United States

Regional Aquifer-System Analysis

Professional Paper 1416–F

USGS
science for a changing world

Availability of Publications of the U.S. Geological Survey

Order U.S. Geological Survey (USGS) publications by calling the toll-free telephone number 1-888-ASK-USGS or contacting the offices listed below. Detailed ordering instructions, along with prices of the last offerings, are given in the current-year issues of the catalog "New Publications of the U.S. Geological Survey."

Books, Maps, and Other Publications

By Mail

Books, maps, and other publications are available by mail from—

USGS Information Services
Box 25286, Federal Center
Denver, CO 80225

Publications include Professional Papers, Bulletins, Water-Supply Papers, Techniques of Water-Resources Investigations, Circulars, Fact Sheets, publications of general interest, single copies of permanent USGS catalogs, and topographic and thematic maps.

Over the Counter

Books, maps, and other publications of the U.S. Geological Survey are available over the counter at the following USGS Earth Science Information Centers (ESIC's), all of which are authorized agents of the Superintendent of Documents:

- Anchorage, Alaska–Rm. 101, 4230 University Dr.
- Denver, Colorado–Bldg. 810, Federal Center
- Menlo Park, California–Rm. 3128, Bldg. 3, 345 Middlefield Rd.
- Reston, Virginia–Rm. 1C402, USGS National Center, 12201 Sunrise Valley Dr.
- Salt Lake City, Utah–2222 West, 2300 South
- Spokane, Washington–Rm. 135, U.S. Post Office Building, 904 West Riverside Ave.
- Washington, D.C.–Rm. 2650, Main Interior Bldg., 18th and C Sts., NW.

Maps only may be purchased over the counter at the following USGS office:

- Rolla, Missouri–1400 Independence Rd.

Electronically

Some USGS publications, including the catalog "New Publications of the U.S. Geological Survey," are also available electronically on the USGS's World Wide Web home page at **http://www.usgs.gov**

Preliminary Determination of Epicenters

Subscriptions to the periodical "Preliminary Determination of Epicenters" can be obtained only from the Superintendent of Documents. Check or money order must be payable to the Superintendent of Documents. Order by mail from—

Superintendent of Documents
Government Printing Office
Washington, DC 20402

Information Periodicals

Many Information Periodicals products are available through the systems or formats listed below:

Printed Products

Printed copies of the Minerals Yearbook and the Mineral Commodity Summaries can be ordered from the Superintendent of Documents, Government Printing Office (address above). Printed copies of Metal Industry Indicators and Mineral Industry Surveys can be ordered from the Center for Disease Control and Prevention, National Institute for Occupational Safety and Health, Pittsburgh Research Center, P.O. Box 18070, Pittsburgh, PA 15236–0070.

Mines FaxBack: Return fax service

1. Use the touch-tone handset attached to your fax machine's telephone jack. (ISDN [digital] telephones cannot be used with fax machines.)
2. Dial (703) 648–4999.
3. Listen to the menu options and punch in the number of your selection, using the touch-tone telephone.
4. After completing your selection, press the start button on your fax machine.

CD-ROM

A disc containing chapters of the Minerals Yearbook (1993–95), the Mineral Commodity Summaries (1995–97), a statistical compendium (1970–90), and other publications is updated three times a year and sold by the Superintendent of Documents, Government Printing Office (address above).

World Wide Web

Minerals information is available electronically at **http://minerals.er.usgs.gov/minerals/**

Subscription to the catalog "New Publications of the U.S. Geological Survey"

Those wishing to be placed on a free subscription list for the catalog "New Publications of the U.S. Geological Survey" should write to—

U.S. Geological Survey
903 National Center
Reston, VA 20192

United States Department of the Interior

U. S. GEOLOGICAL SURVEY
Cartography & Publishing Program
505 Science Drive
Madison, WI 53711

November 30, 2001

To whom it may concern:

A short time ago, you received one or more copies of Professional Paper 1416-F, "Ground-Water Flow in the Gulf Coast Aquifer Systems, South-Central United States," by Alex K. Williamson and Hayes F. Grubb. Since that time, two errors were found on the title page that we believe should be corrected. "Coast" is misspelled and the report number is incorrect. Attached is a replacement page. Please replace the original title page with this new one. The information on the back of the title page is correct, so you can simply staple this new title page onto the original one. Thank you for your help with this matter.

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

By ALEX K. WILLIAMSON *and* HAYES F. GRUBB

REGIONAL AQUIFER-SYSTEM ANALYSIS—
GULF COASTAL PLAIN

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1416–F

GROUND-WATER FLOW IN THE GULF COST AQUIFER SYSTEMS, SOUTH-CENTRAL UNITED STATES

By ALEX K. WILLIAMSON *and* HAYES F. GRUBB

REGIONAL AQUIFER-SYSTEM ANALYSIS—
GULF COASTAL PLAIN

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1415–B

U.S. DEPARTMENT OF THE INTERIOR

GALE A. NORTON, ***Secretary***

U.S. GEOLOGICAL SURVEY

Charles G. Groat, ***Director***

Reston, Virginia 2001

Library of Congress Cataloging- in-Publication Data
Williamson, Alex K.
Ground-water flow in the Gulf Coast aquifer systems, south central United States / by Alex K. Williamson and Hayes F. Grubb.
p. cm. -- (Regional aquifer-system analysis --Gulf coastal plain) (U.S. Geological Survey professional paper: 1416-F)
Includes bibliographical references.
Supt. of Doc. no.: I 19.16:1416-F
1. Groundwater flow--Gulf Coast (U.S.) 2. Aquifers--Gulf Coast (U.S.) I. Grubb, Hayes F. II. Title. III. Series. IV. Series:
U.S. Geological Survey professional paper: 1416-F.
GB1197.7.W57 1998 97-34922
551.49'0976--dc21 CIP
ISBN 0-607-97120-7

For sale by the U.S. Geological Survey
Information Services, Box 25286,
Federal Center, Denver, CO 80225–0286

FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

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

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

Charles G. Groat

Director

CONTENTS

ILLUSTRATIONS

TABLES

CONVERSION FACTORS, VERTICAL DATUM, AND ABBREVIATED WATER-QUALITY UNITS

For readers who prefer to use metric units, conversion factors for terms used in this report are listed below:

Multiply	*By*	*To obtain*
inch	25.40	millimeter
inch per year (in/yr)	25.40	millimeter per year
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer
cubic mile (mi^3)	4.168	cubic kilometer
million acre-feet	1,233	cubic hectometer
million gallons per day (Mgal/d)	0.04381	cubic meter per second
billion gallons per day (Ggal/d)	43.81	cubic meter per second
pounds per square inch per foot [(lb/in^2)/ft]	6.895	kilopascal per meter

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

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

$$°F = 1.8\,(°C) + 32$$

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

By ALEX K. WILLIAMSON and HAYES F. GRUBB

ABSTRACT

The gulf coast regional aquifer systems comprise one of the largest, most complicated, and interdependent aquifer systems in the United States. The Gulf Coast Regional Aquifer-System Analysis (Gulf Coast RASA) study area encompasses approximately 230,000 square miles onshore in parts of Alabama, Arkansas, Florida, Illinois, Kentucky, Mississippi, Missouri, Tennessee, Texas, and all of Louisiana. The aquifer systems (and the study area) extend offshore beneath the Gulf of Mexico to include an additional 60,000 square miles and are truncated at the edge of the Continental Shelf. The Gulf Coast RASA study is limited to coastal plain sediments mostly of Cenozoic age except for the northernmost part of the area, where it includes Late Cretaceous rocks. The thickness of the aquifer system increases toward the Gulf of Mexico and exceeds 17,000 feet near the coastline of southeastern Louisiana. The shallower parts of the aquifers contain freshwater, but the deeper and offshore parts contain mostly mineralized water or brine.

Nearly 10 billion gallons per day of ground water was withdrawn from the aquifers in 1985; most of the water was for irrigation, but substantial quantities were used for municipal and industrial purposes. 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. Although all three aquifer systems are mixtures of fine- and coarse-grained clastic sediments deposited in continental and marine environments, the clay and silt beds in the coastal lowlands aquifer system are generally thinner, dispersed vertically throughout the aquifers, and not areally as extensive as in the two other aquifer systems.

The base of the gulf coast regional aquifer systems in the northern part of the area is at the top of the Midway confining unit and in the southern part is at the transition zone into geopressured sediments. Faults are common, but fault throws generally are not great enough to entirely offset the regional hydrogeologic units described in this report, although individual beds could be offset. Many salt domes occur in several basins of the gulf coast.

The gulf coast regional aquifer systems were divided into 10 aquifers and 5 regional confining units. Each of the aquifers was simulated as a separate layer in a three-dimensional, variable density, finite-difference ground-water flow model with 10-mile-grid spacing. The model accounted for inelastic compaction of fine-grained beds, resulting in more water being released from storage than would be released from coarse-grained sediments alone, and also resulting in land subsidence. Many of the aquifer characteristics, such as thickness, sand percentage, water density (based on concentration of dissolved solids), temperature, and pressure, were derived from a computerized file of 989 geophysical logs.

The factors that controlled regional ground-water flow in the aquifer systems under predevelopment conditions (before 1925) are, in order of importance: (1) topography; (2) outcrop and subcrop pattern and geometry of aquifers, permeable zones, and confining units; (3) variation of hydraulic properties of aquifers, permeable zones, and confining units; (4) distribution of density and brines; and (5) downdip limits of geohydrologic units and geologic structure. Topography is the most significant factor controlling ground-water flow because the relatively humid climate maintains the aquifer systems generally full to overflowing with ground water. Therefore, the amount and distribution of regional recharge and discharge is generally proportional to the topographic gradient and aquifer conductivity rather than being controlled by precipitation or other factors. The major variations in subsurface permeability have been accounted for by delineation of aquifers and confining units. Hydraulic conductivity of sand beds increases from the western side of the study area toward the eastern side. Effective hydraulic conductivity varies as a power function of sand percentage because as the sand percentage decreases, the degree of hydraulic connection among sand beds also decreases. Hydraulic conductivity also decreases with depth due to compaction.

About one-third of the volume of the aquifer systems studied contains freshwater; in the remaining two-thirds, saline water increases rapidly in concentration with distance, generally downdip, to more than 100,000 milligrams per liter. This increase causes large differences in density that have substantial effects on ground-water flow both in the saline and the freshwater part of the system. Many forces apparently have prevented the more saline or brine part of the system from reaching equilibrium, such as heat and salt sources (salt domes), sediment compaction, geopressured sediments, changes of sea level through geologic time, and an extensive continental shelf requiring long flow paths along which ground water must move to adjust to these forces. Thus, the primary source of brines most likely is dissolution of salt domes. Geologic structure seems to have a small effect on ground-water flow except for the indirect effect on the geometry of the geohydrologic units. The effect of the small amount of flow leaking up through the basal Midway confining unit, which is composed of marine clays several hundred feet thick, was found to be insignificant compared to the volume of flow in the aquifer above it. This also seems to be true for the small amounts of flow leaking up from the geopressured zone farther downdip toward the coastline.

Predevelopment net regional recharge was occurring in 41 percent of the aquifer system at the average rate of about 0.48 inch per year. The highest predevelopment net recharge rate was about 6 inches per year in southwestern Mississippi. Predevelopment net regional discharge was occurring in 59 percent of the area at the rate of about 0.35 inch per year. The largest net regional discharge was in the Mississippi River Valley alluvial aquifer at about 4 inches per year. The distribution of freshwater (dissolved-solids concentration less than 10,000 milligrams per liter) thickness correlated well with the simulated net regional recharge and discharge, indicating that the regional flow has flushed saline water (dissolved-solids concentration greater than

10,000 milligrams per liter) out of the sediments, creating a thicker section of freshwater. Some ground water flows offshore and is discharged to the ocean, but most is discharged before it reaches the coastline.

Major changes have occurred in the gulf coast regional aquifer systems in response to large-scale development of ground water throughout much of the study area. Development has led to lowering of hydraulic heads, changes in recharge and discharge, changes in flow and velocity both in direction and magnitude, land subsidence, and changes in water quality. Total pumpage increased by a factor of 2.5 during 1960–80 but decreased by 7 percent from 1980 to 1985.

Every aquifer or permeable zone has more than one major cone of drawdown where drawdown exceeds 80 feet. The 1985 pumpage is more than three times the rate of predevelopment regional recharge. Ground-water pumpage and the resulting declines in hydraulic head tripled the regional recharge to more than 9 billion gallons per day. The increase in regional recharge was derived from both an increase in natural recharge and capture of water that was previously locally discharged. The regional discharge decreased to about 1 billion gallons per day, nearly one-third of its predevelopment rate. Less than 10 percent of the pumpage in 1985 was supplied from ground-water storage.

By 1985, the development caused the area of net regional recharge to increase from 41 percent to 66 percent of the study area. The area that changed from having net regional discharge to net regional recharge includes most of the area underlain by the Mississippi River Valley alluvial aquifer and most of the area underlain by the coastal lowlands within about 40 miles of the coastline except in southern Texas. Offshore, near Houston, Texas, there has been a change in the direction of flow from onshore areas to offshore areas during predevelopment, whereas in 1985, the simulated flow was from the offshore areas inland.

Downward vertical flows between layers increased substantially in all of the aquifers and permeable zones as a result of ground-water pumpage and resultant reversal of vertical gradients. The upper Claiborne aquifer continued to have the most flow interactions with the Mississippi River Valley alluvial aquifer, probably due to the large area of contact. The upper Claiborne aquifer has very little pumpage but has large inflows and outflows. The middle Wilcox, middle Claiborne, permeable zone D, and the Mississippi River Valley alluvial aquifer are mostly self supporting; in other words, most of their pumpage was supplied from increased recharge and decreased discharge within the layer itself. In the lower Wilcox aquifer and permeable zone E, which have very narrow outcrop bands, much of the pumpage had to be supplied by vertical flow from adjacent layers. The upper Claiborne aquifer, which has a very wide outcrop and subcrop band, captures additional recharge and discharges and transmits to adjacent aquifers.

Large-scale ground-water pumpage in the freshwater part of the aquifer systems has markedly changed flow patterns in the brine (dissolved-solids concentration greater than 35,000 milligrams per liter) part of the system and induced flow updip toward pumping areas. However, it will take decades or centuries for the brine water to reach these pumping areas because of the relatively slow ground-water velocities when compared to the long flow paths.

Simulation indicates that there is great potential for continued further development of the ground-water resource. By carefully designing the pattern of pumping additions, doubling the pumpage to 20 billion gallons per day could be accomplished with minimal impacts (average of less than 60 feet of drawdown over the study area).

INTRODUCTION

Smaller scale studies of parts of aquifer systems commonly are hampered by not knowing how the hydrology of the study area relates to the hydrology and development of adjacent areas and by not having enough data to do useful statistical analyses and comparisons. Hydrologic boundary conditions for smaller study areas, though important, generally are unknown. Commonly, data or time, or both, are insufficient to thoroughly study certain types of data or to consider different methods of analysis. The Regional Aquifer-System Analysis (RASA) Program, of which this study is a part, was designed to minimize these problems.

The RASA Program, which was started in 1978 by the U.S. Geological Survey, resulted from a congressional mandate to develop quantitative appraisals of the major ground-water systems of the United States (see Foreword). More than 20 major aquifer systems have been studied (Sun, 1986; Sun and Weeks, 1991). 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). To reach these objectives, the use of computer simulation of ground-water flow in the aquifer system under natural, undisturbed conditions and conditions affected by human activities has proven to be an important tool.

The 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 systems (and the study area) extend offshore beneath the Gulf of Mexico to include an additional 60,000 mi^2 and are truncated at the edge of the Continental Shelf. The study area and aquifer systems have been terminated at natural hydrologic boundaries rather than political boundaries.

The Gulf Coast RASA study is limited to coastal plain sediments of mostly Cenozoic age except for the northernmost part of the area, where it includes rocks of Late Cretaceous age. This report does not include the Upper Cretaceous rocks because they were the subject of detailed studies reported elsewhere (Brahana and Mesko, 1988). The aquifer systems thicken toward the Gulf of Mexico in a general wedge shape and have 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. An overview of the aquifer systems and the plan of study for the Gulf Coast RASA were described by Grubb (1984, 1985, and 1987).

Ground water is an important resource in the study area though rainfall and surface water are relatively abundant. Nearly 10 Ggal/d of ground water was withdrawn from the aquifers during 1980 and 1985, mostly for irrigation. Public supply and industrial uses of

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

ground water also are very important in the region. The effect of ground-water withdrawals has become regional in nature (Grubb, 1984, p. 1). 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 saltwater intrusion are significant side effects of ground-water-level declines 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 on the geologic structure and sedimentation pattern, the separation of aquifer systems by significant regional confining units, and the presence of more than two significant regional aquifers within each aquifer system. The topography and outcrop pattern that control the regional ground-water flow in the Mississippi embayment have been substantially affected by the synclinal structure of the embayment. Although all three aquifer systems are mixtures of fine- and coarse-grained clastic sediments, the fine-grained beds in the coastal lowlands aquifer system are generally thinner, dispersed vertically throughout the aquifers, and not areally as extensive as in the other two aquifer systems.

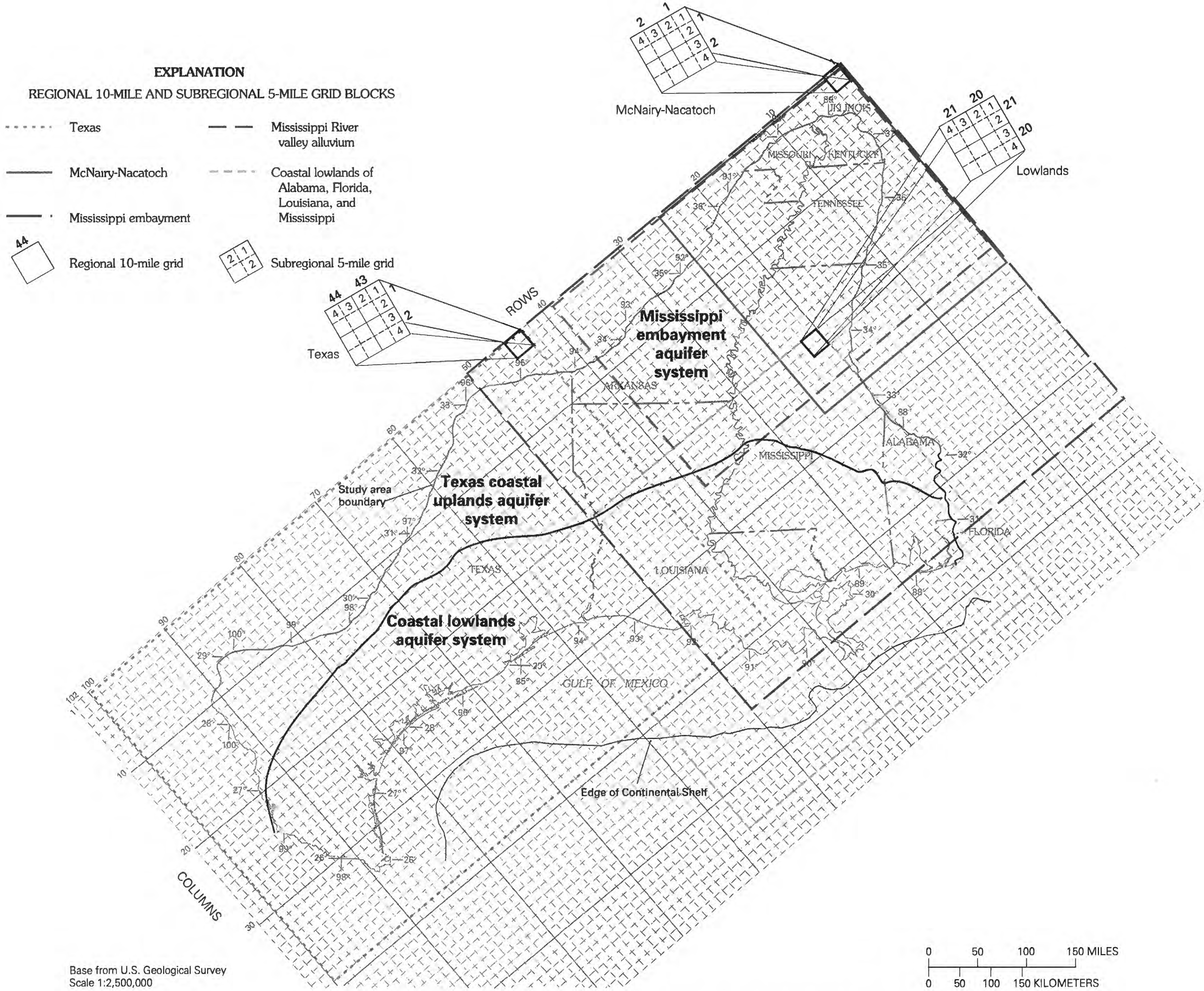


FIGURE 2.—Gulf coast regional aquifer systems and regional and subregional model areas.

PURPOSE AND SCOPE

The purpose of this report is to present an analysis of the regional flow in the gulf coast regional aquifer systems from predevelopment through 1987, to develop an understanding of the controlling factors, and to evaluate the potential for further development. To accomplish this, regional flow paths were determined, and characteristics of the system were compared from one area or aquifer to another. The significant human effects on flow are explained. The effects of future development on ground-water flow and the related problems are presented.

Computer simulation of ground-water flow is the best available tool for this study to (1) integrate most of the known information about the aquifer system, (2) test hypotheses about the regional flow system, and (3) provide a common basis for comparing system characteristics from one area or aquifer to another. Therefore, most of the results presented in this report are based on digital simulations. The approach and model construction are summarized in this report. Owing to the complexity of the aquifer system and the amount of information, the details of the methods used in the modeling that have not been presented elsewhere (Williamson and others, 1990; Williamson, 1987) will be presented in the supplement "Simulation Details" at the end of this report.

ACKNOWLEDGMENTS

Text editing was done by Gertrude MacDonnell. Many of the figures were prepared or modified by April Wild and Matt Eli. We thank them for their willing assistance and accept the responsibility for the errors and omissions in the report.

RELATED INVESTIGATIONS

Numerous related investigations have been done owing to the size of the study area, the complexity of the aquifer systems, and the intensity of development of the ground-water resources in the area. There are four adjacent RASA studies as well as five subregional studies done as part of the Gulf Coast RASA study. Numerous investigations were made, but only those most closely related to this report are referenced here.

Several other reports from the Gulf Coast RASA study are closely related to this report. Results from a ground-water flow model developed by Kuiper (1994), which used nonlinear regression to estimate parameter values, were used extensively in the simulations presented in this report. The geohydrologic units of the gulf coast regional aquifer systems are described by Weiss (1992) and Hosman and Weiss (1991), and the geology is discussed by Hosman (1993). The geochemistry of these aquifer systems is described by Pettijohn (in press), and the maps showing the concentration of dissolved solids and temperature are presented by Pettijohn and others (1988).

ADJACENT REGIONAL AQUIFER-SYSTEM ANALYSIS STUDIES

The Gulf Coast RASA study area is adjacent to four other regional aquifer systems with completed or ongoing studies under the RASA Program (fig. 1). They are the Edwards-Trinity, Central Midwest, Southeastern Coastal Plain, and Floridan aquifer systems.

EDWARDS-TRINITY

The Edwards-Trinity Regional Aquifer-System Analysis includes Cretaceous carbonate aquifers which dip underneath the gulf coast aquifers in Texas and adjacent States (Bush, 1986; Barker and others, *in press*). The vertical flow from or to the Edwards-Trinity aquifer system is thought to have a small effect on flow in the lower part of the Texas coastal uplands aquifer system owing to the great thickness of fine-grained marine sediments of the Midway Group, which separates the two aquifer systems.

CENTRAL MIDWEST

The Central Midwest Regional Aquifer-System Analysis includes the Paleozoic rocks west of the northern tip of the Gulf Coast RASA study area (Jorgensen and others, 1993; Imes and Emmett, 1994). 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 more productive 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 Regional Aquifer-System Analysis includes Cretaceous and Tertiary sediments east of the Gulf Coast RASA study area (Renken, 1984; Barker and Pernik, 1994). The study areas overlap in western Alabama and eastern Mississippi; however, the Southeastern Coastal Plain RASA study is restricted to Cretaceous aquifers that dip beneath the clays of the Midway Group at the base of the gulf coast regional aquifer systems. The Midway Group becomes more permeable due to facies changes and yields water in Georgia and east-central Alabama; but westward towards the gulf coast regional aquifer systems it is predominantly composed of thick marine clays yielding small amounts of water.

FLORIDAN

The Floridan Regional Aquifer-System Analysis includes the major Tertiary carbonate aquifers of the coastal Southeastern United States (Bush and Johnston, 1988; Johnston and Bush, 1988). The boundary between the Floridan and the Gulf Coast RASA study areas occurs where the more permeable carbonate facies changes to the less permeable marine clays of the Jackson and Vicksburg Groups, which form a regional confining unit in the Gulf Coast RASA study area.

GULF COAST RASA SUBREGIONAL STUDIES

More detailed investigations of the flow systems in subareas or groups of aquifers were conducted in the gulf coast by using subregional models (at one-half of the grid spacing; fig. 2). The subregional models, except for the gulf coast aquifer in Texas, simulate only the part of the aquifer system containing freshwater, whereas the regional model includes the entire thickness of Tertiary and younger sediments except where they are geopressured. The regional model includes all of the aquifers included in any of the subregional studies, except for the McNairy-Nacatoch aquifer (contained in rocks of Late Cretaceous age) in the northernmost part of the area. The subregional models are described separately:

McNairy-Nacatoch aquifer (Brahana and Mesko, 1988; Brahana and Mesko, written commun.)

Mississippi River Valley alluvial aquifer (Ackerman, 1989; Ackerman, 1996)

Mississippi embayment aquifer system (Arthur and Taylor; 1990, Arthur and Taylor, *1998*)

Coastal lowlands aquifer system of Alabama, Florida, Louisiana, and Mississippi (Martin and Whiteman, 1989; Martin and Whiteman, *1999*)

Gulf coast aquifers in Texas (Ryder, 1988; Ryder and Ardis, 1991)

PREVIOUS INVESTIGATIONS

A detailed discussion of many of the studies since 1901 of ground-water flow in the gulf coast regional aquifer systems was given by Williamson and others (1990). Only the more recent and comprehensive ones will be summarized here. By the mid-1970's reports documenting intensive studies of the ground-water resources had been published for many of the 343 counties in the study area. Also many studies of special ground-water problems were documented; for example, saltwater encroachment in the Houston-Galveston area (Winslow and others, 1957), effects of faults on the ground-water flow system (Whiteman, 1979; Gabrysch, 1984), and subsidence due to pumping ground water (Winslow and Doyel, 1954; Winslow and Wood, 1959; Gabrysch, 1969, 1977, 1982; Gabrysch and Bonnet, 1975; 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 described in a series of reports (Cushing and others, 1970; Boswell and others, 1965, 1968; Hosman and others, 1968). A multi-State 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 was published during the 1970's (West and Broadhurst, 1975; Baker and Wall, 1976; Bedinger and Sniegocki, 1976; Zurawski, 1978; 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 electric analog simulation techniques in the study area. The following analog model studies were made: the Houston District, Texas, by Wood and Gabrysch (1965); Jorgensen (1973, 1975), and Jorgensen and Gabrysch (1974); the Sparta Sand in the Mississippi embayment by Reed (1972); and the Bayou Bartholomew alluvium in Arkansas by Broom and Reed (1973).

By the mid-1970's, studies began to appear where digital models were used to simulate ground-water flow, such as: an alluvial aquifer and adjacent Carrizo-Wilcox aquifers near a reservoir on the Trinity River in eastern Texas by Garza (1974); the Carrizo aquifer in southern Texas by Klemt and others (1976); along the Colorado River of Texas by Thorkildsen and others (1989); the Bayou Bartholomew alluvium in Arkansas by Reed and Broom (1979); and the heavily pumped Houston area of Texas by Meyer and Carr (1979) and Jorgensen (1981).

The number of areas and the types of problems to which simulation techniques were applied continued to increase in the decade of the 1980's, and the digital model replaced the electric analog model. The use of simulation techniques within the study area is discussed herein under five categories: (1) shallow alluvial aquifers, (2) aquifers simulated with multilayer models, (3) effects of pumpage on water levels and salinity in a single layer, (4) effects of development on other resources of the ground-water system, and (5) the use of optimization models for managing ground-water resources.

Parts of the shallow Mississippi River Valley alluvial aquifer were the subject of two separate ground-water simulation studies—the Cache and St. Francis River basin in northeastern Arkansas by Broom and Lyford (1981) and the "Delta" area in western Mississippi by Sumner and Wasson (1984a, 1984b).

Aquifers simulated by multilayer models include the Texas Gulf Coastal Plain underlain by the Chicot and Evangeline aquifers (Carr and others, 1985); three zones in the Baton Rouge area, Louisiana (Torak and Whiteman, 1982; Huntzinger and others, 1985); the Memphis Sand of the Memphis area, Tennessee (Brahana, 1982); and the Jasper aquifer in eastern Texas (Baker, 1986).

Studies involving water-level declines and resultant movement of saline water into freshwater due to ground-water pumping also have been conducted. Groschen (1985) studied the Chicot and Evangeline aquifers southwest of Corpus Christi, Texas, and Trudeau and Buono (1985) studied the Sparta aquifer near West Monroe, Louisiana.

The effects of development of other resources on the ground-water flow system have been evaluated using simulation techniques. A multilayer model study of the Wilcox Group near the Oakwood salt dome in eastern Texas was made by Fogg and others (1983); a study of ground-water flow along a fault zone where uranium mining was being conducted by the solution method in Miocene-age sediments in southern Texas was reported by Henry and others (1982); and the effects of open-pit lignite mining and in-situ gasification of lignite on ground-water flow in Eocene-age sediments in east-central Texas was made by Charbeneau and Wright (1983).

Optimization models to manage pumpage such that a minimum saturated thickness is maintained have been proposed by Peralta and Peralta (1984), Peralta and Killian (1985), Peralta and others (1986), and Yazdanian and Peralta (1986) for the Grand Prairie area of Arkansas where the Mississippi River Valley alluvial aquifer is heavily pumped.

The primary purpose of most of the studies cited above has been to address a specific water-management problem, and only secondarily to develop an understanding of the flow in the aquifer under study and how it interacts 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.

PHYSICAL SETTING

The study area is a gently sloping coastal plain in a mostly humid temperate climate and is underlain by thick, clastic sediments deposited in a gulfward offlapping sequence. Much agricultural and industrial development has occurred in the area, supporting a fast-growing population of more than 18 million as of 1980 (Mesko and others, 1990) and requiring substantial ground- and surface-water resources.

TOPOGRAPHY

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 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 Mississippi River is an important feature of the hydrologic system because of its size; it generally traverses the east side of the alluvial plain. Toward the coast, the alluvial plain slopes toward the Gulf of Mexico and is cut through by large stream valleys that are approximately perpendicular to the coastline.

CLIMATE

The warm, relatively humid climate of the area is very favorable for agriculture. Mean annual precipitation in the area averages 48 inches and ranges from about 24 inches near the Mexico border to about 64 inches along the gulf coast of Louisiana, Mississippi, and Alabama (fig. 4). The precipitation is evenly distributed throughout the year, though the spring is somewhat wetter and the summer and fall are somewhat drier, except in the western part of the area where most of the rain falls in the spring and summer (fig. 4). The mean annual temperature ranges from about 16°C in the north to about 21°C in the south, and the potential evapotranspiration is less variable than precipitation across the area.

In part of the area or during part of the time, or both, the actual evapotranspiration is limited to the amount of rainfall, which is less than the potential evapotranspiration. Pan evaporation, which is generally higher but closely related to potential evapotranspiration, ranges from about 45 in/yr to about 100 in/yr and averages about 64 in/yr (fig. 5). In the western part of the area in Texas, potential evapotranspiration generally exceeds the rainfall (figs. 4, 5), especially in the summer and fall, so most of the rainfall returns to the atmosphere in a short time. During wetter times or in wetter areas, or

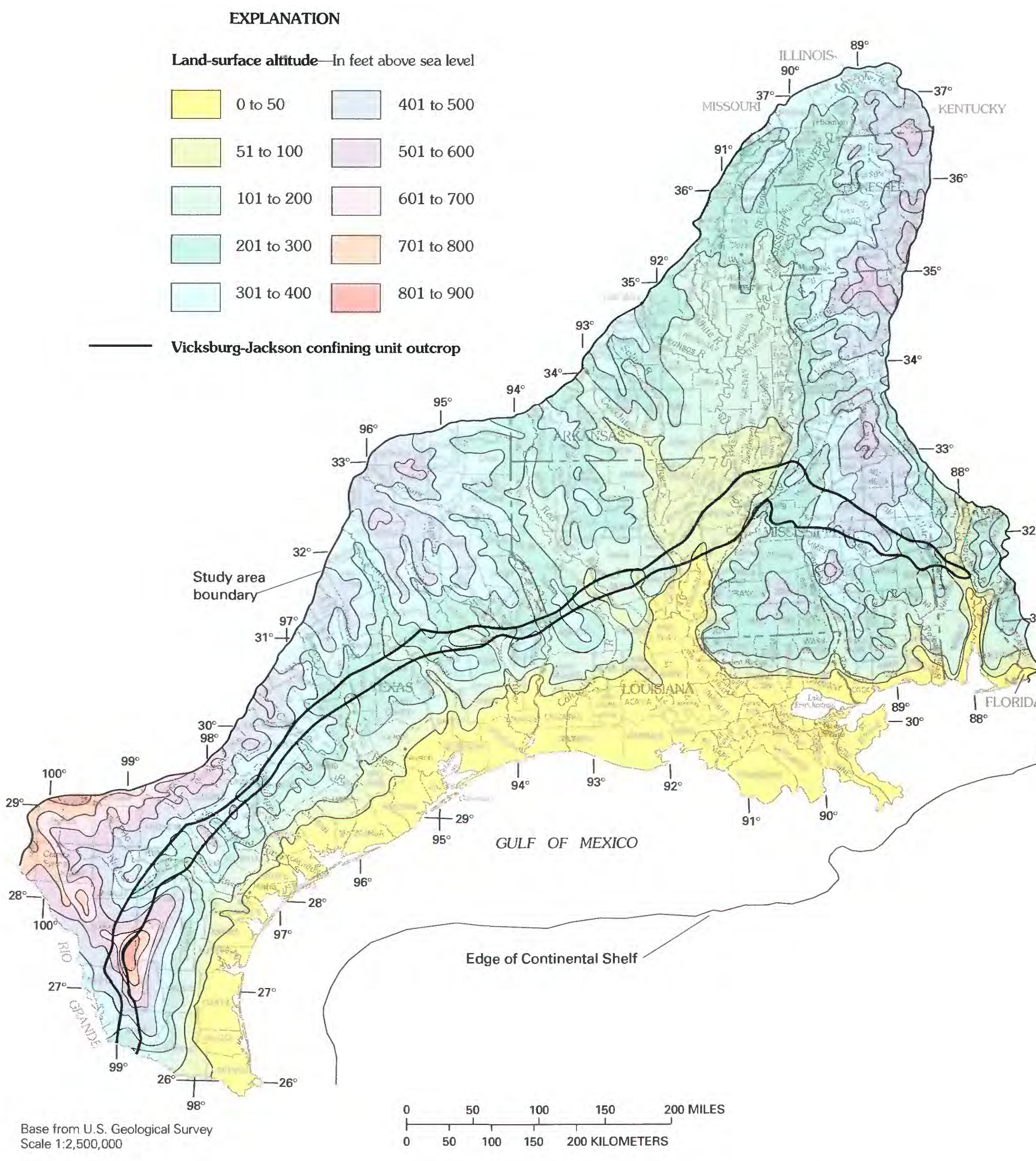


FIGURE 3.—Generalized average land-surface altitude and outcrop of Vicksburg-Jackson confining unit.

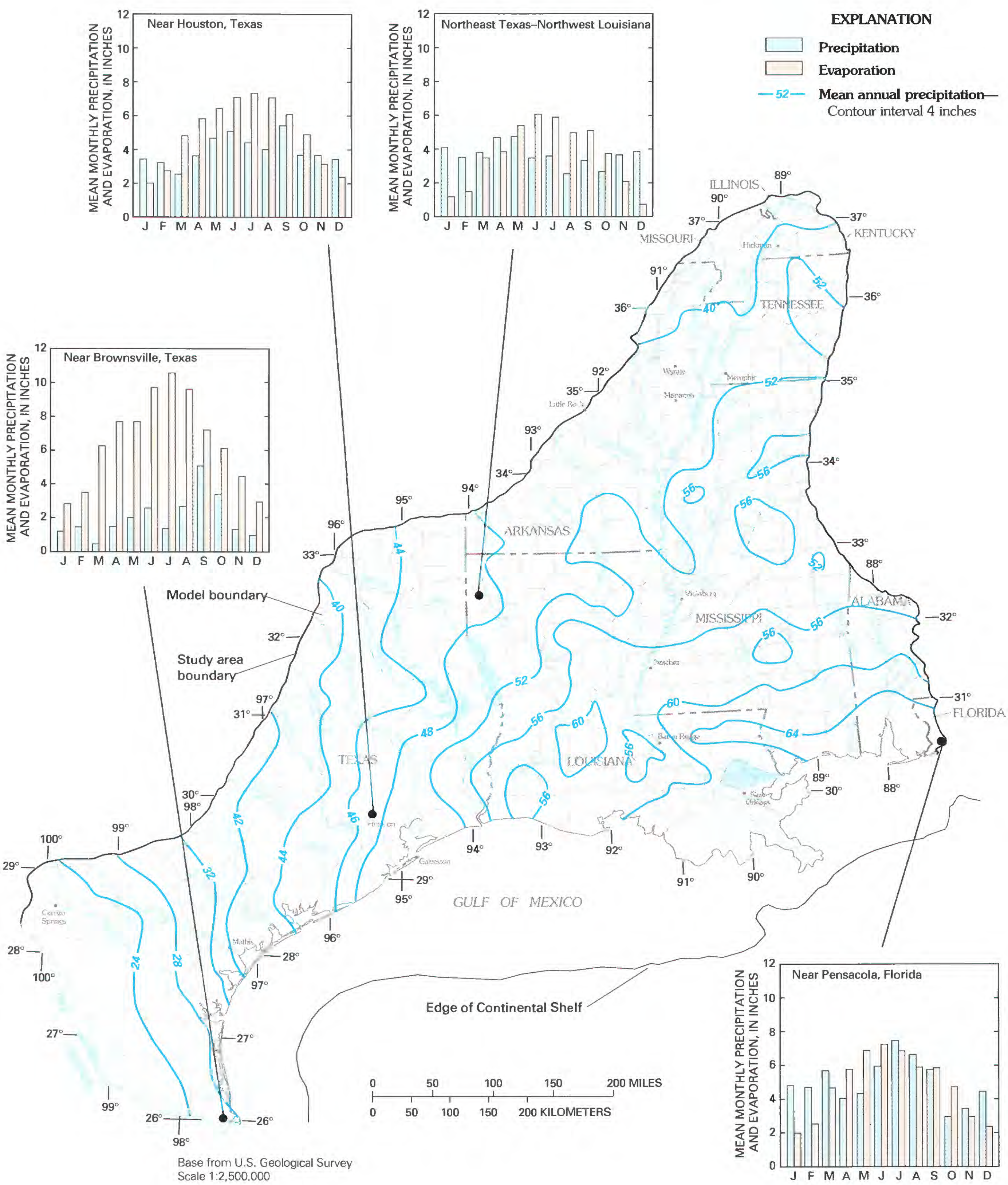


FIGURE 4.—Mean annual precipitation, 1951–80, and mean monthly precipitation and pan evaporation at four locations.

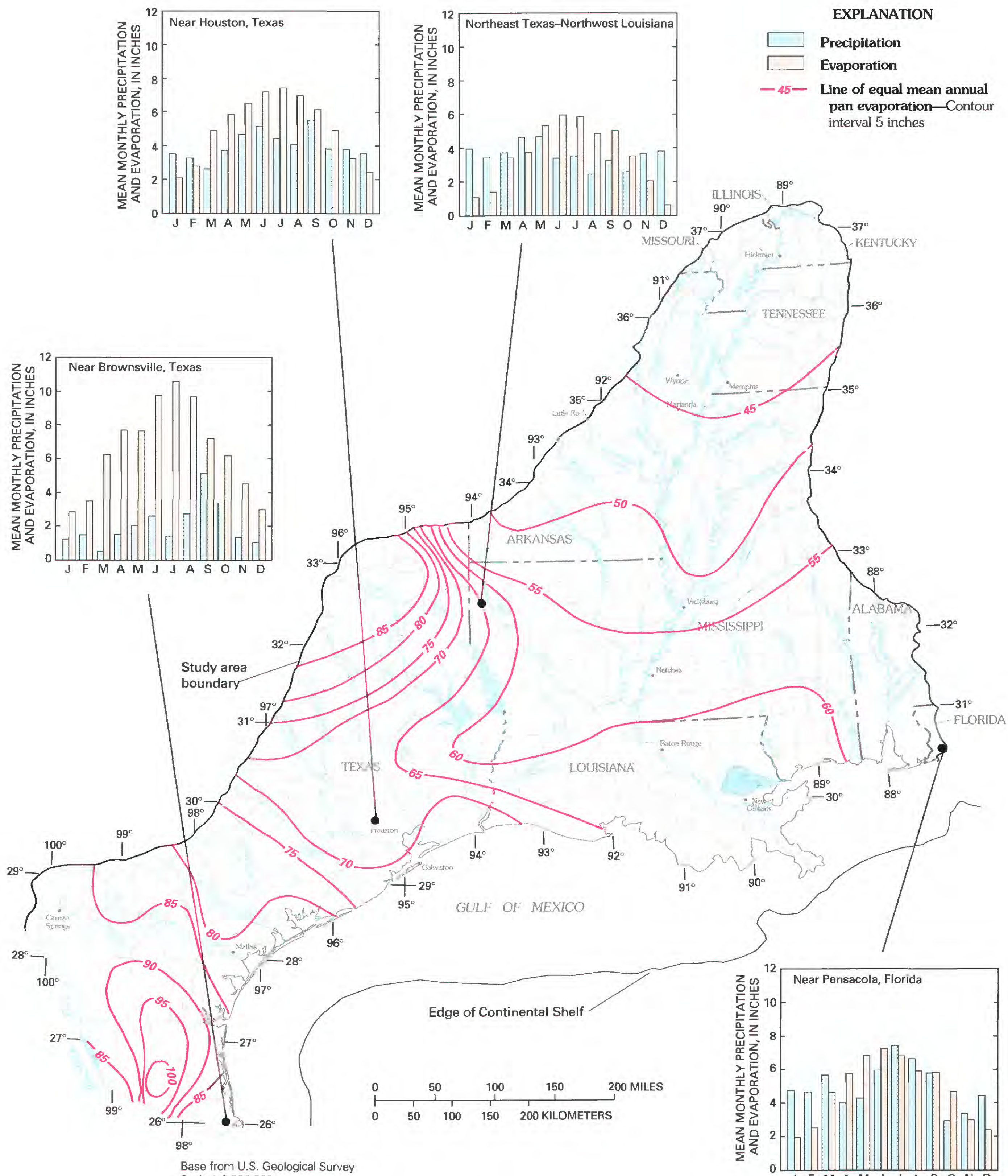


FIGURE 5.—Mean annual pan evaporation, 1951–80.

both, when rainfall exceeds potential evapotranspiration, the remainder either infiltrates directly into the ground or runs off into streams. In the eastern part of the area, rainfall exceeds potential evapotranspiration and provides abundant surface-water runoff. The part of rainfall that recharges the aquifers is usually only a small part of the total, so it is difficult to estimate from a water-budget approach.

SURFACE WATER

Many large rivers flow through the study area. The unit runoff (mean annual flow divided by drainage area and converted to a depth over drainage area) (fig. 6) varies from about 0.2 in/yr in the southwestern part of the area to about 20 in/yr in the northern and eastern parts and averages about 14 in/yr. The mean annual flow of the larger rivers is over 467 million acre-ft/yr or

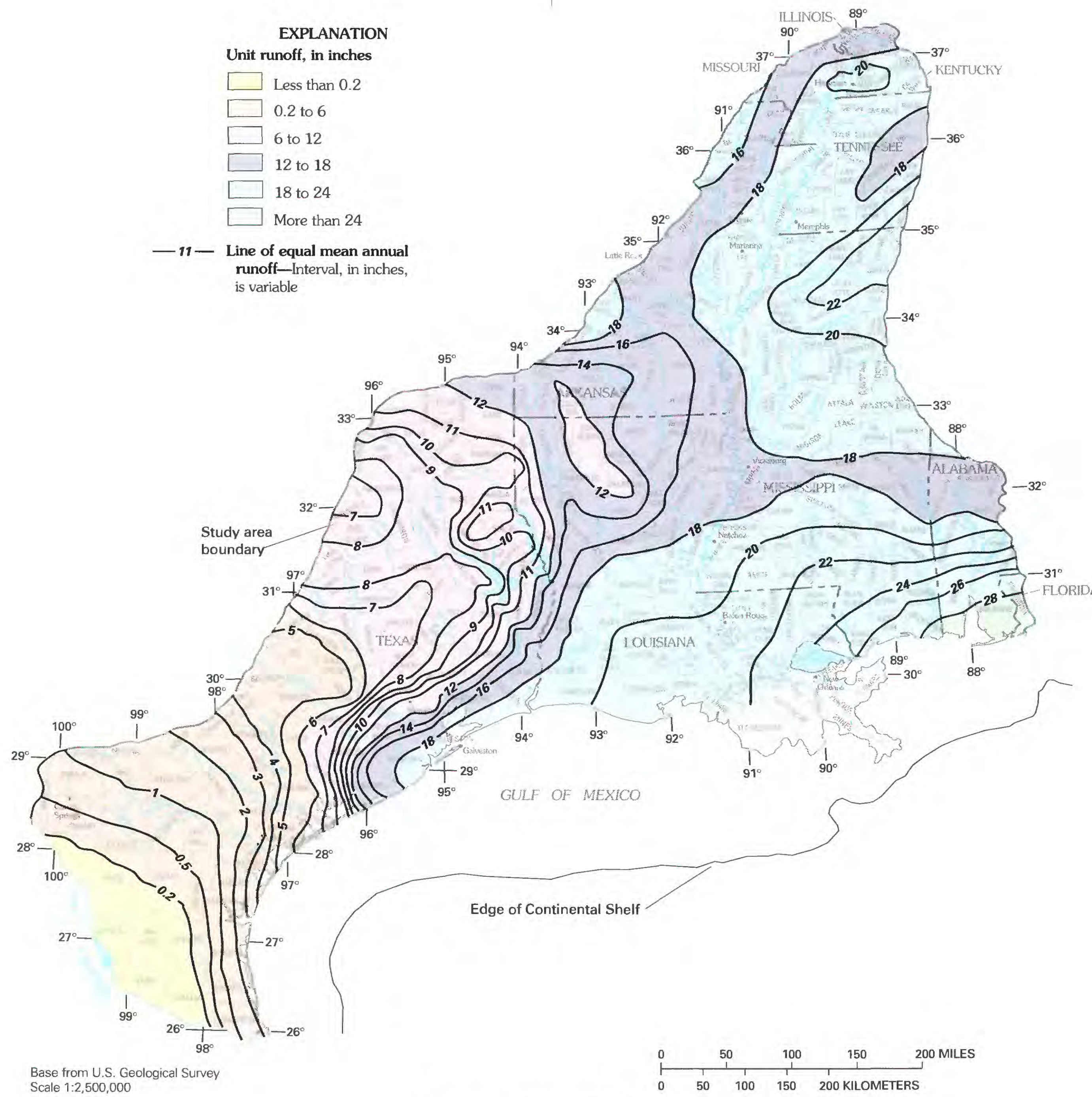


FIGURE 6.—Unit runoff, 1951–80.

416 Ggal/d from a combined drainage area of more than 1.5 million mi^2 (Williamson and others, 1990, table 1). Streamflow rates are so large that the portion moving into or out of the aquifer is not detectable, within the streamflow measurement error, because the ground-water surface-water interflow is generally less than 1 percent of the streamflow. The combined storage (conservation pool or nonflood storage) in major reservoirs in or near the study area is more than 30 million acre-ft (Williamson and others, 1990, table 1). However, the surface-water storage is very small relative to the volume, more than 2,000 million acre-ft, of freshwater in storage in the top 200 ft of the aquifer systems (assuming an aquifer outcrop area of 180,000 mi^2, a depth to the water table 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) (Mesko and others, 1990).

GEOHYDROLOGIC FRAMEWORK

The sediments of the gulf coast regional aquifer systems were deposited mostly during Cenozoic time. Deposition was 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 sea level resulted in transgressions and regressions of the sea and thus cyclical sedimentation. Depositional environments alternated from predominantly continental to predominantly marine. In general, the more sandy continental deposits have higher permeabilities characteristic of aquifers, and the more clayey marine deposits have lower permeabilities characteristic of confining units. Sedimentary units crop out in roughly parallel bands that are progressively younger gulfward in a typical offlap sequence.

The Gulf Coast geosyncline and the Mississippi embayment are the major structural features of the study area (fig. 7) and largely control the pattern and thickness of sedimentation (fig. 8). The base of the aquifer system (fig. 8) in the northern part of the area is at the top of the Midway confining unit and in the southern is specified as the transition zone into geopressured sediments. Except where affected by local uplift, the general pattern of sedimentation is one of increasing thickness in a gulfward, downdip direction. Hosman

TABLE 1.—*Summary of thickness, areal extent, and sand percentage of model layers*

[From Williamson and others, 1990, table 3. Some numbers may not agree due to rounding]

Layer number (see fig. 9)	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 (silt and clay)
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	7,850	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	3440	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

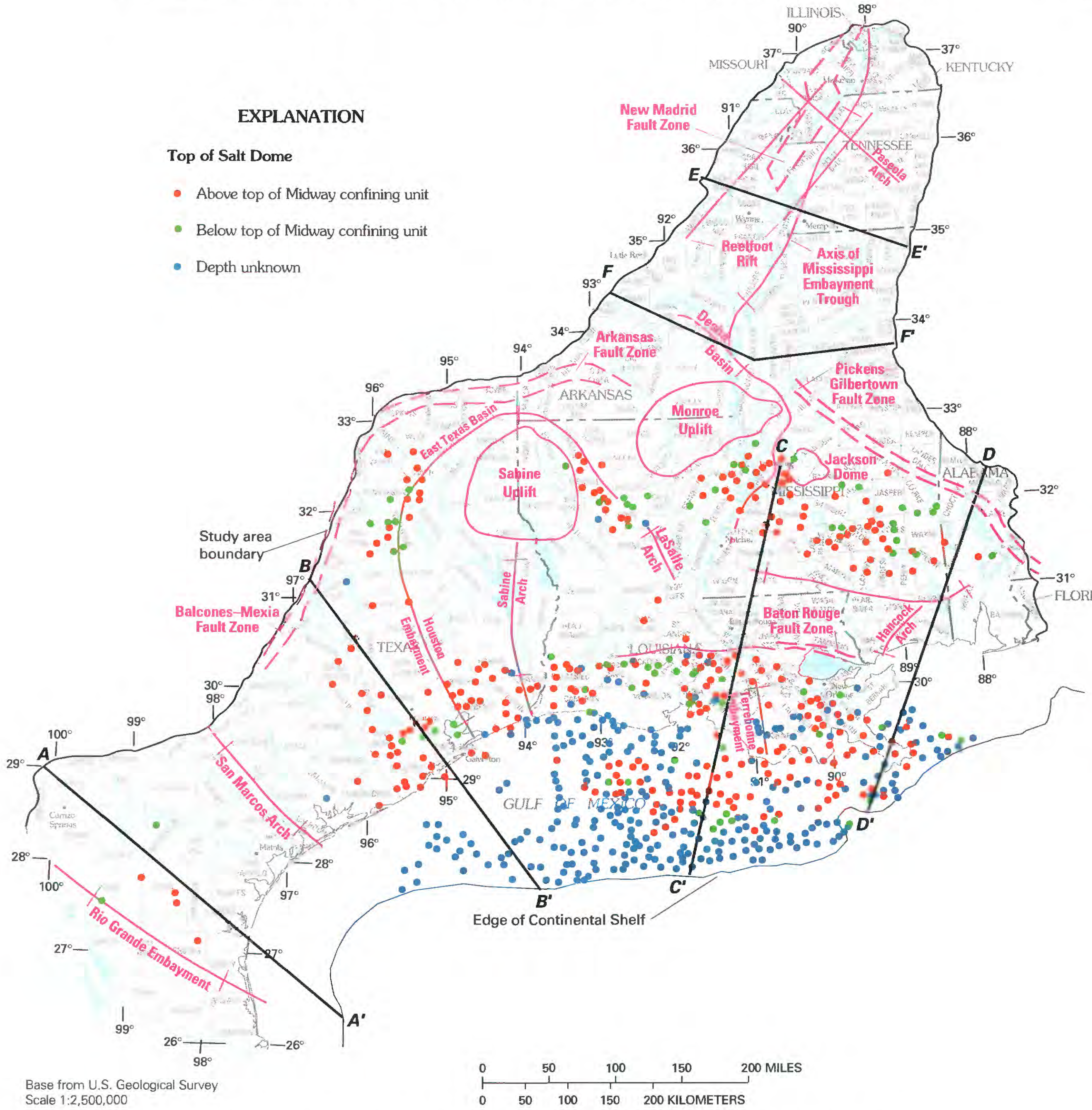


FIGURE 7.—Location of structural features, salt domes, and geohydrologic sections (shown in fig. 24 and plate 1).

(1988) presents a fence diagram of the study area and describes the lithology of the geohydrologic units.

Faults are common throughout the area, although their effect on regional ground-water movement is not well understood. In general, fault throws are not great enough to entirely offset the regional geohydrologic units described in this report, although individual beds could be offset. Numerous growth faults, which occur contemporaneously with 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 could act similarly. Due to the scale of this analysis, the effects of faults were not considered on an individual basis.

Geopressured zones (fig. 8) have fluid pressures that are substantially greater than the normal hydrostatic pressure (Jones, 1969). Geopressured zones encroach younger units as the geopressured zones occur higher in the section in a gulfward direction. The most probable cause for the development of these abnormally high

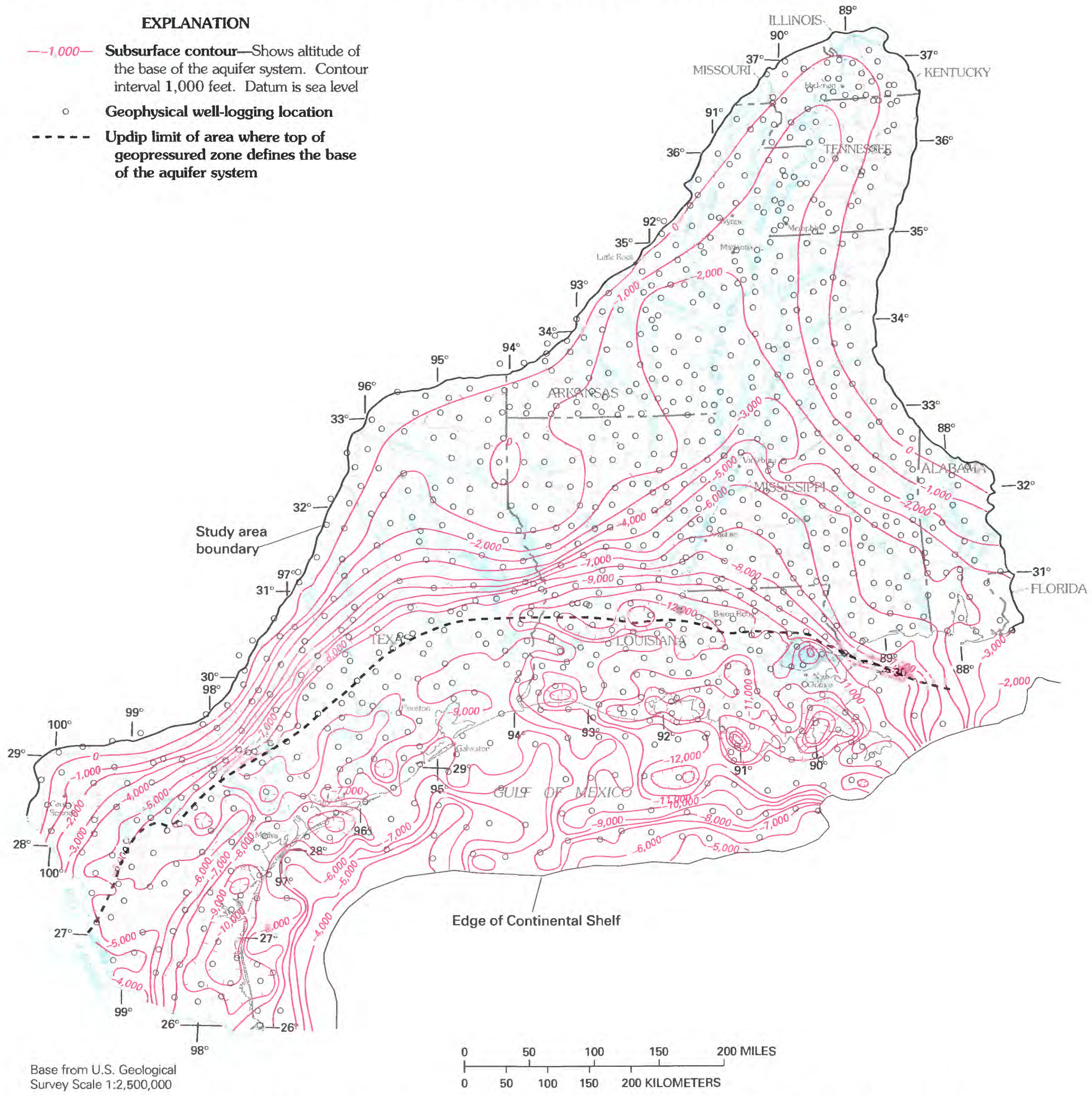


FIGURE 8.—Altitude of the base of the aquifer systems, area of occurrence of geopressure, and geophysical well-logging locations.

fluid pressures is the restriction of the escape of fluids 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, and given a very long time, free from other influences, the abnormally high pressures will eventually reduce to near normal hydrostatic pressures.

Salt domes occur throughout the gulf coast regional aquifer systems (fig. 7), particularly in bands near and roughly parallel to the coastline. The source of salt for the domes is the deeply buried Louann Salt of Jurassic age. The salt is extruded in diapirs that penetrate varying thicknesses of Cenozoic strata. The extent to which each dome penetrates the gulf coast regional aquifer systems is described by Beckman and Williamson (1990). The structural effects of the domes are relatively localized. However, the domes can have a substantial effect on water quality in the deeper sediments due to dissolution of salt by ground water. Simulation demonstrates that the highly mineralized ground water affects regional flow by creating areas with high fluid densities.

The gulf coast regional aquifer systems were subdivided into geohydrologic units for analysis in the Gulf Coast RASA study. Weiss and Williamson (1985) described the methods used for subdividing the thick sequence of sediments. The relations of the geohydrologic units to aquifer systems and a selection of previously named units and the model layer numbering system used in this study are shown in figure 9. The geohydrologic units are referred to in the text by their layer number for convenience, whereas both number and names generally are used in figures and tables. The term "layer" is used in place of "aquifer or permeable zone" for convenience as well. The relation of the various units are shown in the hydrologic sections on plate 1. The thickness of each unit is a block average for blocks spaced 5 mi apart. Layer 11 includes 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 areas of occurrence (fig. 12) of layer 11 are connected horizontally across a narrow band in central Louisiana. 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 geohydrologic units.

Three approaches were used in three situations to subdivide the aquifer systems into geohydrologic units (Weiss and Williamson, 1985):

1. Borehole geophysical logs were used to map regionally significant confining units (model layers 12–17), with aquifers in between (model layers 5 and 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 (model layers 2 and 4 that represent massive sands and layer 3 that represents thin, complexly interbedded coarse- and fine-grained deposits).
3. In most of the coastal lowlands aquifer system (model layers 7–11) where approaches 1 and 2 could not be used, hydraulic-head and well-opening data at various depths were used to subdivide the system so that the minimum vertical head change occurs within a unit. 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.

APPROACH: DIGITAL MODELING OF GROUND-WATER FLOW

A computer program (Kuiper, 1983, 1985) was used to simulate variable-density ground-water flow in three dimensions using the finite-difference method. Effective vertical hydraulic conductivity is specified for active aquifer layers as well as for inactive confining units. Inactive confining units have no storage nor horizontal flow. Kuiper's model calculates the total leakance (effective vertical conductivity divided by thickness) between the centers of all adjacent model layers by using the appropriate thicknesses and vertical hydraulic conductivities and their harmonic means. The water density, although variable in space, is assumed to be constant in time. This assumption simplifies the simulation substantially and is justified because model simulations of several decades are relatively short spans of time compared to the amount of time necessary for the average density simulated in very large model blocks to be considerably altered by the transport of dissolved solids. The model also accounted for inelastic compaction of fine-grained beds, resulting in more water being released from storage and in land subsidence (Kuiper, 1994).

The areal extent of the gulf coast regional aquifer systems required that the regional model horizontal grid spacing be large (fig. 2)—10 mi on a side—and vertical discretization be limited to 10 active aquifer 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-third of these are active due to the geometry of the systems. The horizontal grid is oriented approximately 45 degrees from north-south (fig. 2). This orientation was chosen to minimize the size of the matrices required to fit the entire areal extent of the aquifer systems.

The geohydrologic unit (model layer) tops, thicknesses, and sand percentages used in these simulations are described in detail by Hosman and Weiss (1991) and Weiss (1992). The one exception is in the Mississippi embayment aquifer system north of the 34th parallel, where Hosman and Weiss describe the Memphis aquifer that is equivalent to both active layers 4 and 5 as modeled in this study. Layer 5 was defined as the top one-third of the Memphis aquifer with layer 4 composing the lower two-thirds of the aquifer for the purpose of preserving the horizontal hydraulic continuity of the model layers. South of the 34th parallel, layers 4 and 5

A

Mississippi embayment and Texas coastal uplands aquifer systems

GEOLOGIC UNIT: SYSTEM	GEOLOGIC UNIT: SERIES	GEOLOGIC UNIT: GROUP	GEOHYDROLOGIC UNITS DEFINED BY PREVIOUS STUDIES	GULF COAST REGIONAL AQUIFER-SYSTEM ANALYSIS: MODEL LAYER NUMBER	GULF COAST REGIONAL AQUIFER-SYSTEM ANALYSIS: 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]
	EOCENE	CLAIBORNE	Cockfield aquifer system (Payne, 1970) Cockfield Formation (Hosman and others, 1968)	6	Upper Claiborne aquifer
			Cook Mountain Formation	14	Middle Claiborne confining unit
			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
			Cane River Formation	13	Lower Claiborne confining unit
			Carrizo and Meridian Sand aquifer (Payne, 1975) Carrizo Sand and Meridian-upper Wilcox aquifer (Hosman and others, 1968)	4	Lower Claiborne-upper Wilcox aquifer
		WILCOX	Wilcox Group (Hosman and others, 1968)	3	Middle Wilcox aquifer
	PALEOCENE		Lower Wilcox aquifer (Hosman and others, 1968)	2	Lower Wilcox aquifer*
		MIDWAY	Midway Group	12	Midway 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 (1984, p. 11).

* Not present in the Texas coastal uplands aquifer system.

FIGURE 9*A*.—Correlation of aquifer systems and geohydrologic units (model layers) with previously named units.

B

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		Upper Chicot aquifer (Jorgensen, 1975) Chicot aquifer (Meyer and Carr, 1979)	11	Permeable zone A (Holocene-upper Pleistocene deposits)
QUATERNARY / TERTIARY	PLEISTOCENE AND HOLOCENE / PLIOCENE			10	Permeable zone B (Lower Pleistocene-upper Pliocene deposits)
TERTIARY	PLIOCENE		Evangeline aquifer (Whitfield, 1975) (Meyer and Carr, 1979)	9	Permeable zone C (Lower Pleistocene-upper Miocene deposits)
TERTIARY	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
TERTIARY	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 (1984, p. 11).

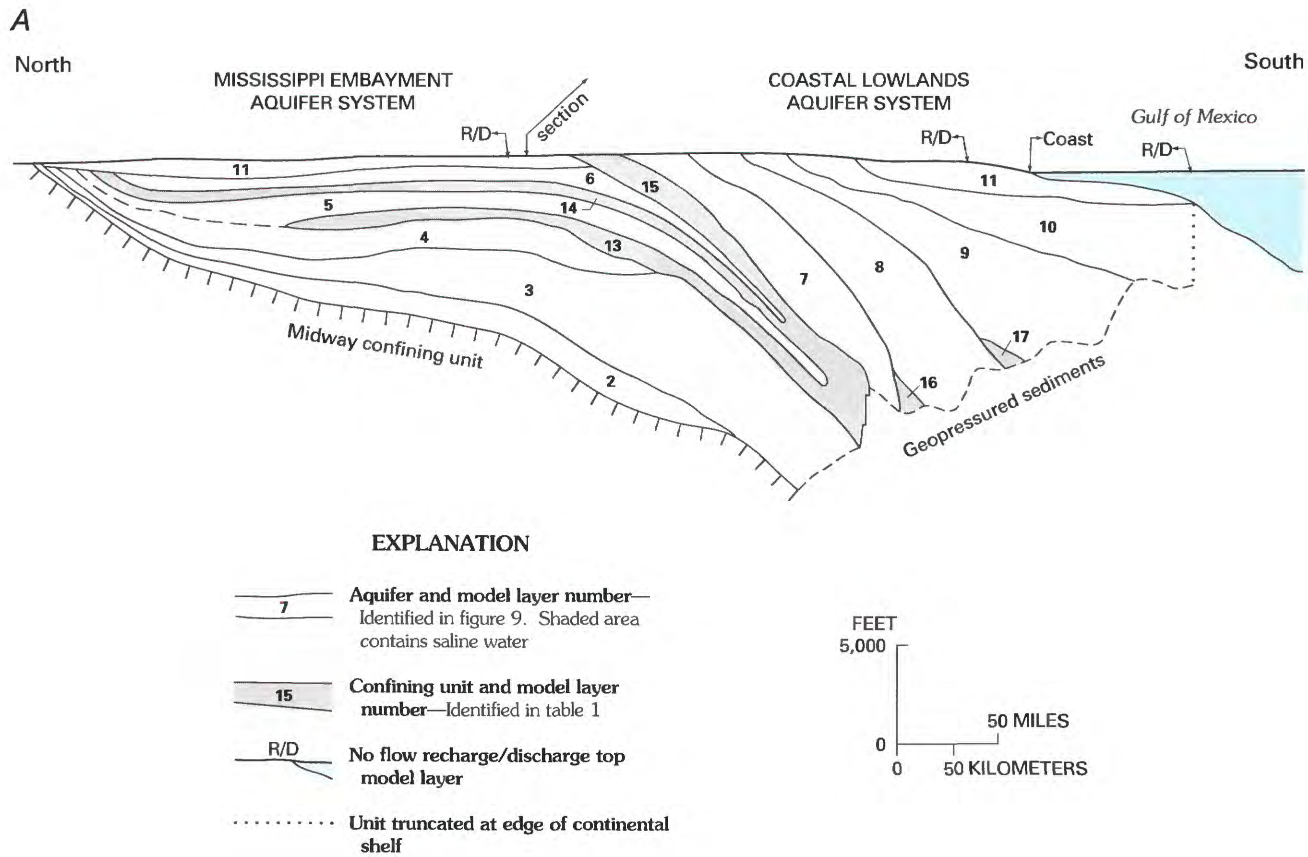
FIGURE 9*B*.—Correlation of aquifer systems and geohydrologic units (model layers) with previously named units.

have an intervening confining unit, the lower Claiborne confining unit (layer 14).

Many of the aquifer characteristics such as thickness, sand percentage, and water density (based on concentration of total dissolved solids and temperature) were derived from a computerized file of 989 geophysical logs (fig. 8) (Wilson and Hosman, 1988). The logs were chosen to show regional trends of both onshore (895 logs) and offshore (94 logs) parts of the aquifer system. Williamson and others (1990) gave the procedures for preparing the data sets for the model by using computer contouring algorithms.

The comparison of the vertical relations of geohydrologic units within the aquifer systems (*A*) with the representation of the geohydrologic units in the regional flow model (*B*) is shown in figure 10. The section shown in figure 10 is diagrammatic, but it shows the general position of aquifers and confining units along the axis of the Mississippi embayment. Where interior layers pinch out, the adjacent layers that still exist are directly connected with no more additional resistance to flow than is inherent in the conductivities and thicknesses of the respective units which do exist. The condition of resistance to flow also applies to the Mississippi River Valley alluvial aquifer (layer 11) that overlies the Mississippi embayment aquifer system (layers 2–6).

The base of the flow system for the regional flow model is assumed to be zero flow and is located at the top of the Midway confining unit, or the top of the zone of geopressure, whichever is shallowest (fig. 8). All of the lateral boundaries of the model were assumed to be no-flow. The eastern boundary was assumed to be no-flow because the aquifers undergo a facies change to a much less permeable unit. The approach used at the southwestern boundary in Mexico was to use estimates



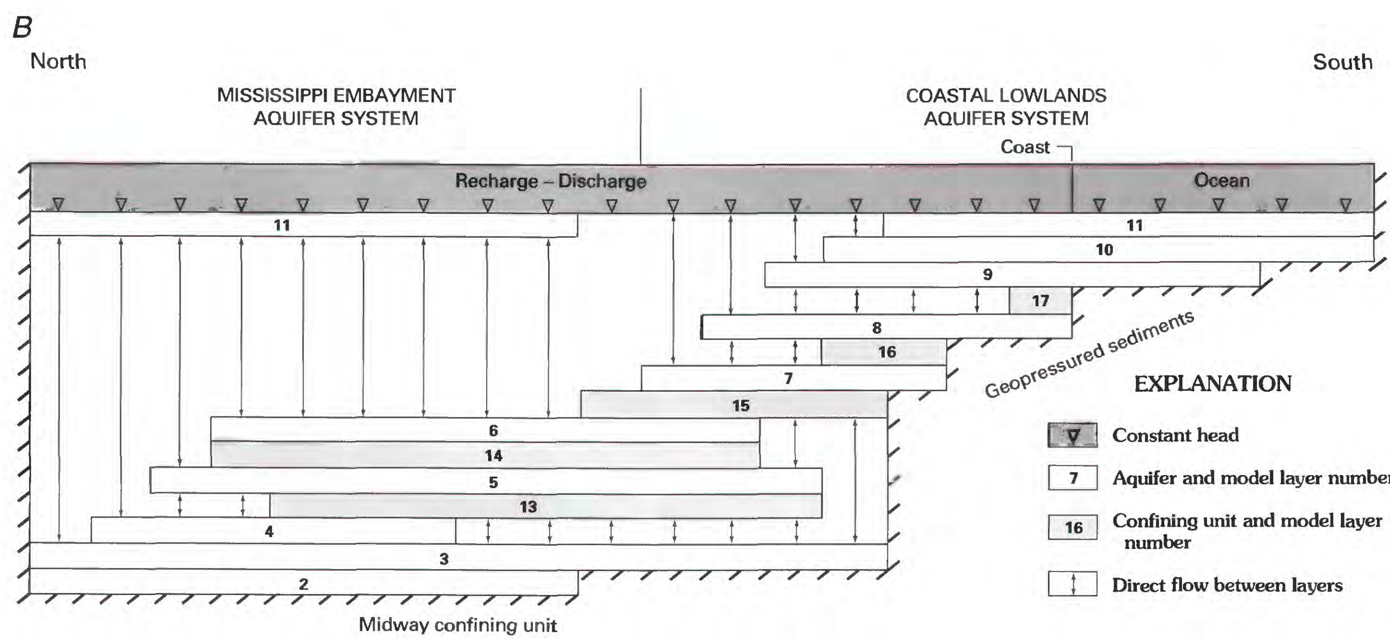


FIGURE 10.—*A*. Vertical relations of the confining units, aquifers, and permeable zones; and *B*. idealized representation of the geohydrologic units in the regional flow model.

in Mexico to extend the model to the aquifer system outcrop rather than try to force an arbitrary boundary at the United States border.

A constant head boundary was simulated in a layer above the top aquifer layer to simulate recharge and discharge. The head value specified was that of the water table (Williams and Williamson, 1989). The flow into or out of the top layer is simulated by using the difference in head between the first active aquifer block in the appropriate outcropping layer and the water-table head value above it times the conductance for vertical flow between the two blocks.

Most of the model calibration depended on automatic parameter estimation results described by Kuiper (1994). The regression methodology allowed testing of a wide range of parameters for simulation of the aquifer system. It also provided estimates of the accuracy of results and a mechanism to determine sources of model error. Kuiper used more than 40 different multiple-regression models with 2 to 31 regression parameters. More than 3,000 values for grid-element volume-averaged head and volume-averaged hydraulic conductivity were used as observations for the regression models. He estimated error bands for each model. Kuiper found that the root mean squared-weighted residual decreases little with an increase in the number of parameters, so that the models with a small number of parameters appear to be the most reliable for prediction of head.

Further discussions of the simulations are presented in the sections to which they relate, and the differences between simulations described in this report from the preliminary simulations of Williamson and others (1990) are discussed in the supplementary section "Simulation Details" at the back of this report. All of the data sets used in the model, along with original data sets used to map the geohydrologic units and ground-water chemistry, are archived in ASCII files on a WORM (Write Once, Read Many) disk that was prepared by Kirkpatrick (1993).

FACTORS CONTROLLING REGIONAL GROUND-WATER FLOW

The factors which controlled regional ground-water flow in the aquifer systems under predevelopment conditions are, in order of significance of effect, (1) topography; (2) outcrop and subcrop pattern and geometry of aquifers, permeable zones, and confining units; (3) variation of hydraulic properties of aquifers, permeable zones, and confining units; (4) distribution of brine water and water density; and (5) downdip limits of aquifers, permeable zones, and geologic structure. They will be discussed in the following sections in the order of the significance of their effect on predevelopment regional ground-water flow.

TOPOGRAPHY

Topography has a major effect on the water-table altitude, which in turn has an effect on natural regional ground-water flow (Williams and Williamson, 1989; Williamson and others, 1990, p. 97). In terrains of humid climate and gentle hydraulic gradients, aquifer systems are generally full to overflowing with ground water. Therefore, the amount of recharge to the regional ground-water flow system is probably not related to the mean annual rainfall over most of the study area. Williamson (1987) reported that there was no correlation between the depth to the water table and the mean annual rainfall in this study area. It is more likely that the amount of recharge to the regional flow system is limited by the capacity of the regional flow system to transmit ground water away from recharge areas to discharge areas. Where the flow capacity is limited by the low regional hydraulic gradients or by the low conductance of the aquifer system, rainfall, which is potential regional recharge, is discharged to local surface-water bodies such as creeks and streams or to evapotranspiration. Under predevelopment conditions, areas with high land-surface altitude (fig. 4) were regional recharge areas, and areas of low land-surface altitude were regional discharge areas. The water-table altitude follows the land-surface altitude. The distribution of the thickness of relatively fresh water also demonstrates that the regional recharge areas are determined by their high relief, causing the deep brines to be more flushed out than in discharge areas. In the western part of the study area, where the climate is much drier (fig. 4), recharge to the regional ground-water flow system probably is limited by the smaller rainfall amount, most of which falls in the hot summer months when it is quickly evaporated or transpired.

ESTIMATION OF PREDEVELOPMENT WATER-TABLE ALTITUDE

Average predevelopment water-table altitudes for 25-mi^2 blocks were estimated by subtracting an estimated depth to water from the land-surface altitude calculated from very detailed digital data (Williams and Williamson, 1989; summarized in Williamson and others, 1990, p. 67). The traditional method of manually contouring measured water-table altitudes at wells was tried but not used owing to the large spacing of wells relative to small-scale variations in land-surface altitude. Most of the variation in predevelopment water-table altitude is a function of the variation in land-surface altitude rather than the variation of depth to water. Linear regression equations of depth to water were developed for five subareas on the basis of data from 6,825 wells and average land-surface altitude (Williams and Williamson, 1989.) These equations also included a measure of local topographic variation. The resulting estimate of predevelopment water-table altitude over the entire study area is a subdued replica of the topography (fig. 3) and was presented by Williamson and others (1990, fig. 21).

In offshore areas, the source-sink layer was assigned a constant head value of 0 (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 the constant head blocks.

OUTCROP AND SUBCROP PATTERN AND GEOMETRY OF AQUIFERS, PERMEABLE ZONES, AND CONFINING UNITS

The second most significant factor controlling regional ground-water flow is the outcrop and subcrop pattern (fig. 11) and geometry of the aquifers, permeable zones, and confining beds. The major variations in subsurface permeability have been accounted for simply by delineation of aquifers and confining units (Grubb, 1986). Grubb qualitatively showed how the outcrop pattern, together with land-surface altitude, controlled the regional recharge and discharge areas. In many localities the water that was recharged to the regional flow system moved laterally to an area within the same outcrop area and was discharged at a lower altitude. Martin and Whiteman (in press) state that 59 percent of the predevelopment recharge in the outcrop areas was also discharged in the outcrop areas in the coastal lowlands aquifer system in Louisiana, Mississippi, Alabama, and Florida. Some examples of the effect of the outcrop pattern will be discussed later in the section "Predevelopment Ground-Water Flow."

The areal variation of the thickness of the regional aquifers and permeable zones and confining units (fig. 12) is also a very important factor determining regional flow. For example, the middle Wilcox aquifer (layer 3) 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, is composed predominantly of interbedded coarse- and fine-grained beds. The coarse-grained beds have varying degrees of hydraulic connection and therefore have a relatively low effective horizontal permeability. In many locations, the middle Wilcox is considered to be a confining unit. However, because of its vast thickness (fig. 12, table 1), the horizontal component of regional flow was substantial (pl. 6). Although recognized in this study as an aquifer, the fine-grained beds within the middle Wilcox aquifer are the major restriction to vertical flow between overlying and underlying aquifers.

HYDRAULIC PROPERTIES OF AQUIFERS, PERMEABLE ZONES, AND CONFINING UNITS

The third most significant factor controlling flow in the aquifer systems is hydraulic conductivity and its areal distribution within the aquifers, permeable zones, and confining units. Hydraulic conductivity (K) of entire aquifers, permeable zones, and confining units is difficult to measure and may vary widely across relatively short distances. However, maximum use was made of field measurements of horizontal hydraulic conductivity (Kh) to estimate effective horizontal hydraulic conductivity (Kh_{eff}) in order to avoid including model simulation errors of all kinds in model-derived values of Kh_{eff}. The Kh values used are directly related to aquifer properties, and they have a greater potential transfer value to other aquifer studies than model-derived Kh_{eff} values. The available data used included sand-bed thicknesses at locations, aquifer-test and specific-capacity data for several thousand wells, and the geometry of the units. The approach used to obtain estimates of Kh is discussed below and can be summarized as follows: (1) use aquifer-test and specific-capacity data to develop areal averages for sand-bed Kh for each model layer, (2) map sand percentages from the geophysical-log data, (3) develop a relation between sand percentage and Kh_{eff}, and (4) adjust the Kh_{eff} estimates for the variation in K with depth. Model-derived values for vertical K of confining units were used because the field data are sparse, highly variable, and difficult to extrapolate from specific locations to block-averaged values.

HORIZONTAL HYDRAULIC CONDUCTIVITY OF AQUIFERS AND PERMEABLE ZONES

The horizontal hydraulic conductivity (Kh) of sand beds is many times larger than the conductivity of fine-grained beds; therefore, the thickness of the fine-grained beds can be ignored in the calculation of transmissivity of the aquifers. The transmissivity of the aquifer layers can be calculated as the product of the total thickness of the sand beds in a layer and the

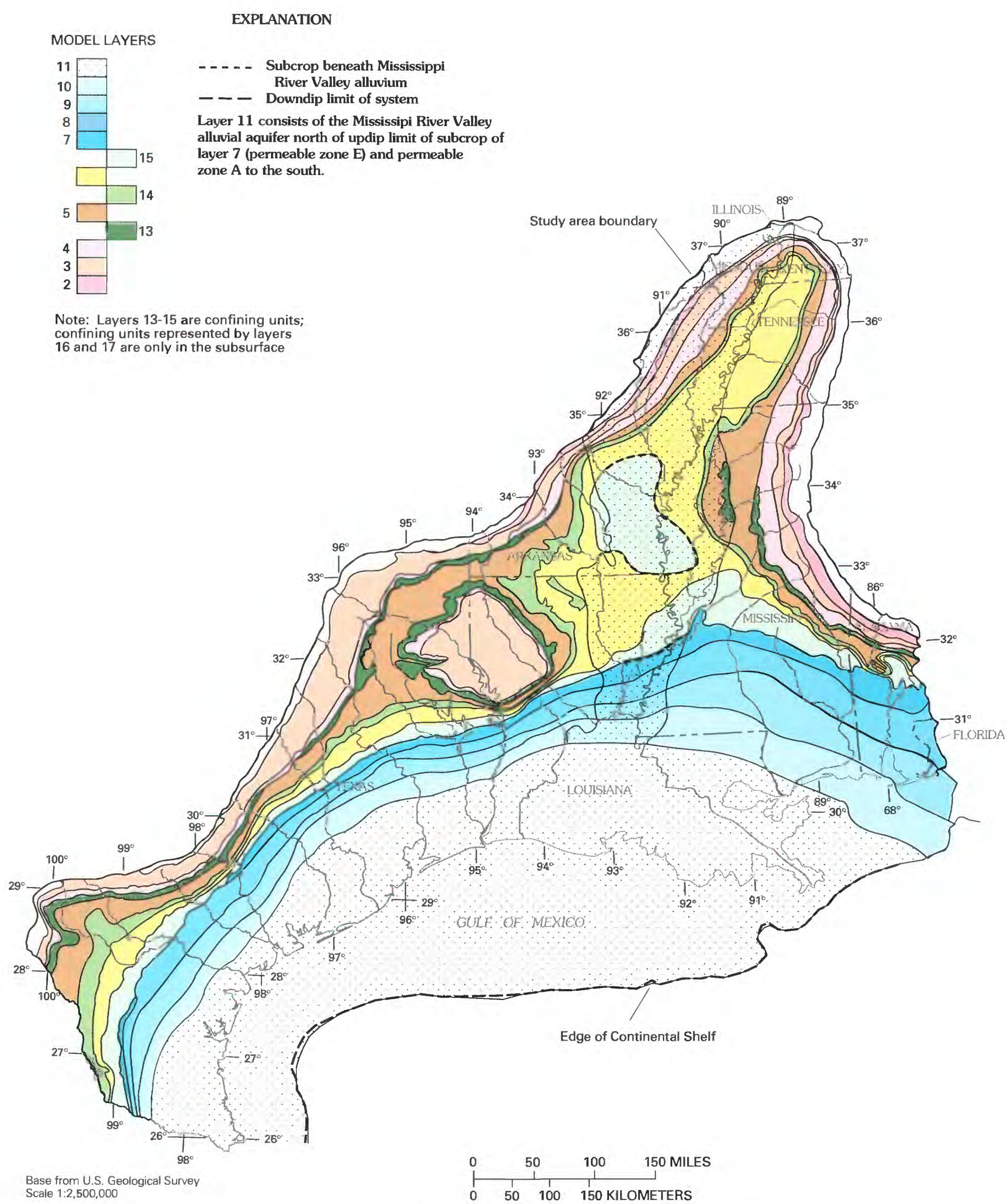


FIGURE 11.—Outcrop and subcrop of aquifers, permeable zones, and confining units (model layers).

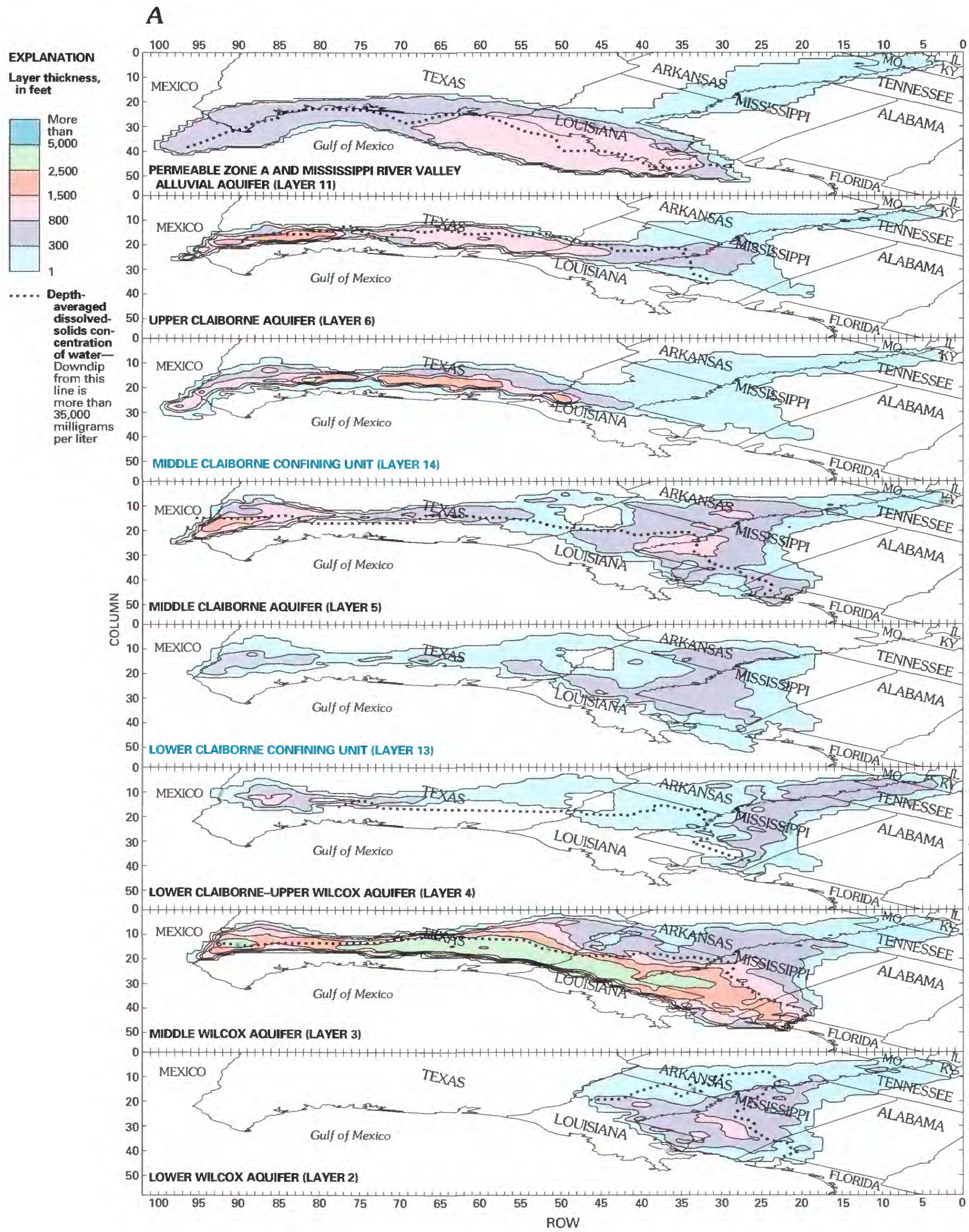


FIGURE 12.—Perspective diagrams showing extent and thickness of aquifers, permeable zones, confining units, and saline/freshwater interface of: *A*. the Mississippi embayment and Texas coastal uplands aquifer systems, and *B*. the coastal lowlands aquifer system.

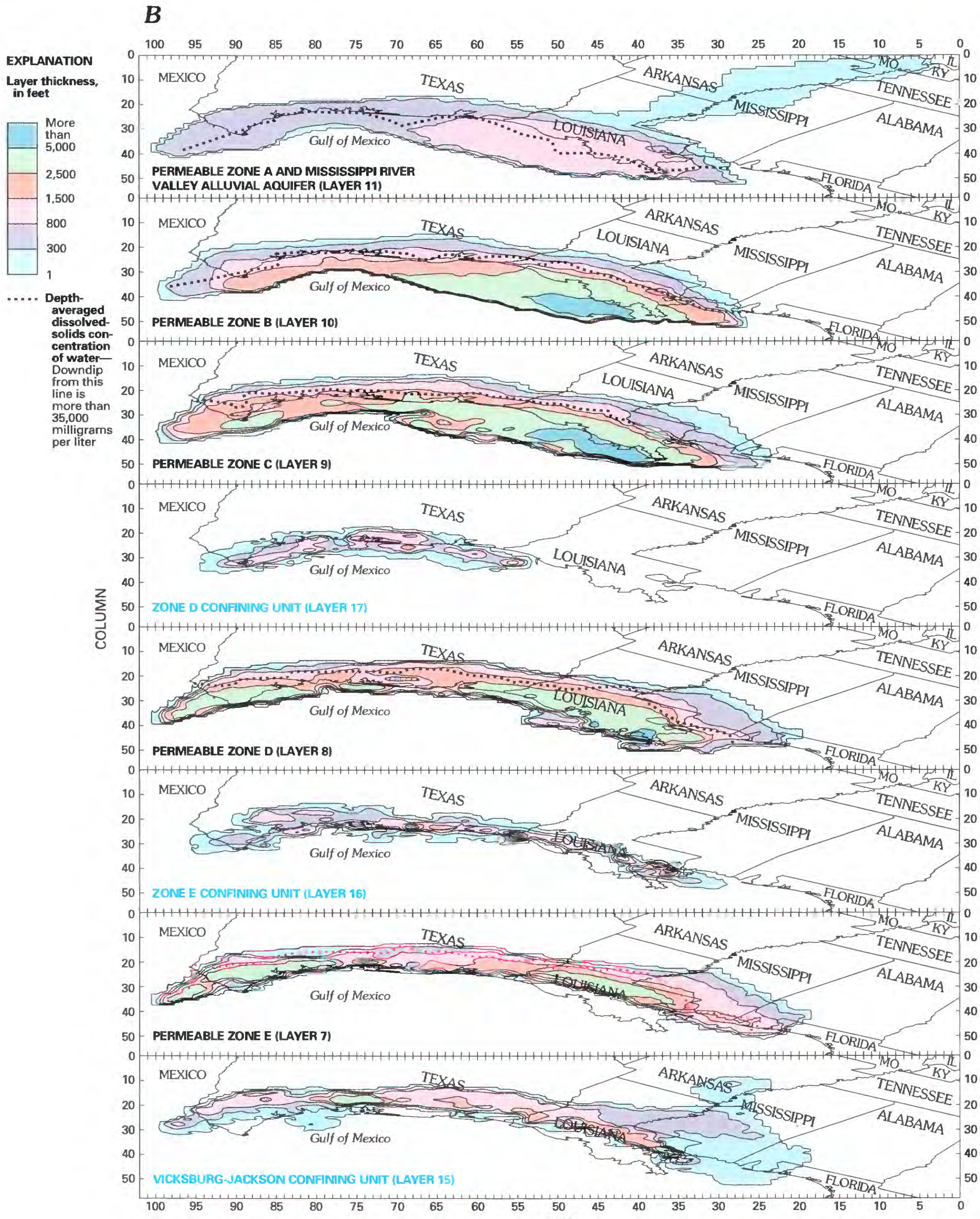


FIGURE 12.—Perspective diagrams showing extent and thickness of aquifers, permeable zones, confining units, and saline/freshwater interface of: *A*. the Mississippi embayment and Texas coastal uplands aquifer systems, and *B*. the coastal lowlands aquifer system—Continued.

effective hydraulic conductivity of the sand beds. The effective hydraulic conductivity (K_{eff}) of a geohydrologic unit is the K value that would produce the same hydraulic gradients in a model (with one value per model block) as those in the aquifer, though the aquifer is composed of many kinds of sediments with different Kh values. The Kh_{eff} varies widely in the aquifer systems and is difficult to estimate. The factors that control Kh_{eff} are: Kh of sand beds, the sand-bed connectedness, and the depth. These factors will be discussed in the following sections.

HORIZONTAL HYDRAULIC CONDUCTIVITY OF SAND BEDS

Prudic (1991) summarized estimates of horizontal hydraulic conductivity for sand beds from about 1,500 aquifer-test analyses and more than 5,000 specific-capacity tests in the study area. He found that there was a significant regional pattern of variation, although the extremely heterogeneous nature of variations at very close distances prohibited mapping the conductivities. The variation of geometric mean hydraulic conductivity of sand (K_{sand}) between areas, ignoring layer differences, was as great as the variation between layers, ignoring areal differences (Prudic, 1991, p. 23).

The study area was divided into nine geographic areas for subsequent discussions of areal comparisons (fig. 13). The regional variation showed up very clearly when all of the data within one geographic area (fig. 13) and layer were averaged and compared with averages for 40 or so other area-layer combinations (fig. 14). Prudic (1991) chose the geometric mean for the averaging statistic to compare because of the log-normal distribution of the hydraulic conductivities.

The regression modeling of Kuiper (1994) indicated that the models with fewer numbers of parameters produced a narrower confidence interval in predicted head; yet the great range between the geometric mean hydraulic conductivities of some area and layer combinations, when compared to others (fig. 14), indicates a compromise solution (discussed in the supplement "Simulation Details"). Grouping the area and layer combinations according to their mean conductivity into six groups (table 2) minimized the number of parameters for the regression model while preserving the differences in mean hydraulic conductivity indicated by the aquifer-test data. The regression model then estimated conductivities for each group which were near the estimated values, yet allowed the model to optimize the fit of hydraulic-head data. The model calibration was done using an initial estimate of the geometric mean K_{sand} (table 2) for each group from Prudic (D.E. Prudic, U.S. Geological Survey, written commun., 1990). Prudic (1991) revised his estimates using a slightly different method of calculating K from specific-capacity data, resulting in a general overall increase of about 30 percent in the value of K. The model calibration was not redone because the change would not substantially affect the results except to increase, by about 30 percent, the K and hence flow and velocity. This amount of change is within the estimated error bands of the regression model.

EFFECTIVE HORIZONTAL HYDRAULIC CONDUCTIVITY OF REGIONAL AQUIFERS

The sand and fine-grained beds of the aquifer systems in many areas were deposited in geometrically complex shapes and highly variable sizes in three dimensions. The ratio of effective hydraulic conductivity (K_{eff}) to K_{sand} depends on the sand-bed connectedness, which depends on the geometry of the beds and the scale used in the simulation of ground-water flow (Freeze and Cherry, 1979, p. 32–34; Bear, 1972, p. 153–155; Bear and Verruijt, 1987, p. 35–37). What is referred to here as K_{eff} is called equivalent conductivity by Freeze and Cherry (1979).

In such a complex depositional environment, even a simple question such as, "Can the lithology be described as sand beds in a fine-grained unit or as fine-grained beds in a sand unit?" is actually difficult to answer. This question does not have a yes or no answer, but rather the answer lies somewhere in between the two extremes. In nearly every locality, both extremes exist. In areas where deposition is predominantly sand, the sediments could be characterized generally as fine-grained beds in a sand unit, where the reverse would be true in other areas (fig. 15). Figure 15 is an attempt to represent the observed pattern of sand and fine-grained beds that compose the gulf coast regional aquifer systems. It was produced by moving through an array representing a cross section of an aquifer and randomly locating sand, or fine-grained beds, or both, using the sand percentage as the probability of existence of a bed in any one location. The geometry of the beds (length to width ratio) and spatial correlation are implicit due to the vertical exaggeration of the hypothetical sections.

The appropriate method of calculating K_{eff} for a model depends on the geometry of coarse- and fine-grained beds in an aquifer. In the case where a heterogeneous aquifer is composed of alternating sand and fine-grained beds whose extent is larger than that of a model block (fig. 16*A*), the K_{eff} is calculated as follows. If a given flow path went through a specific pattern of very low-conductivity clay beds and varying conductivity sand beds (analogous to series flow in electrical circuits), the effective conductivity would be equal to the harmonic mean (inverse of the mean of the inverses

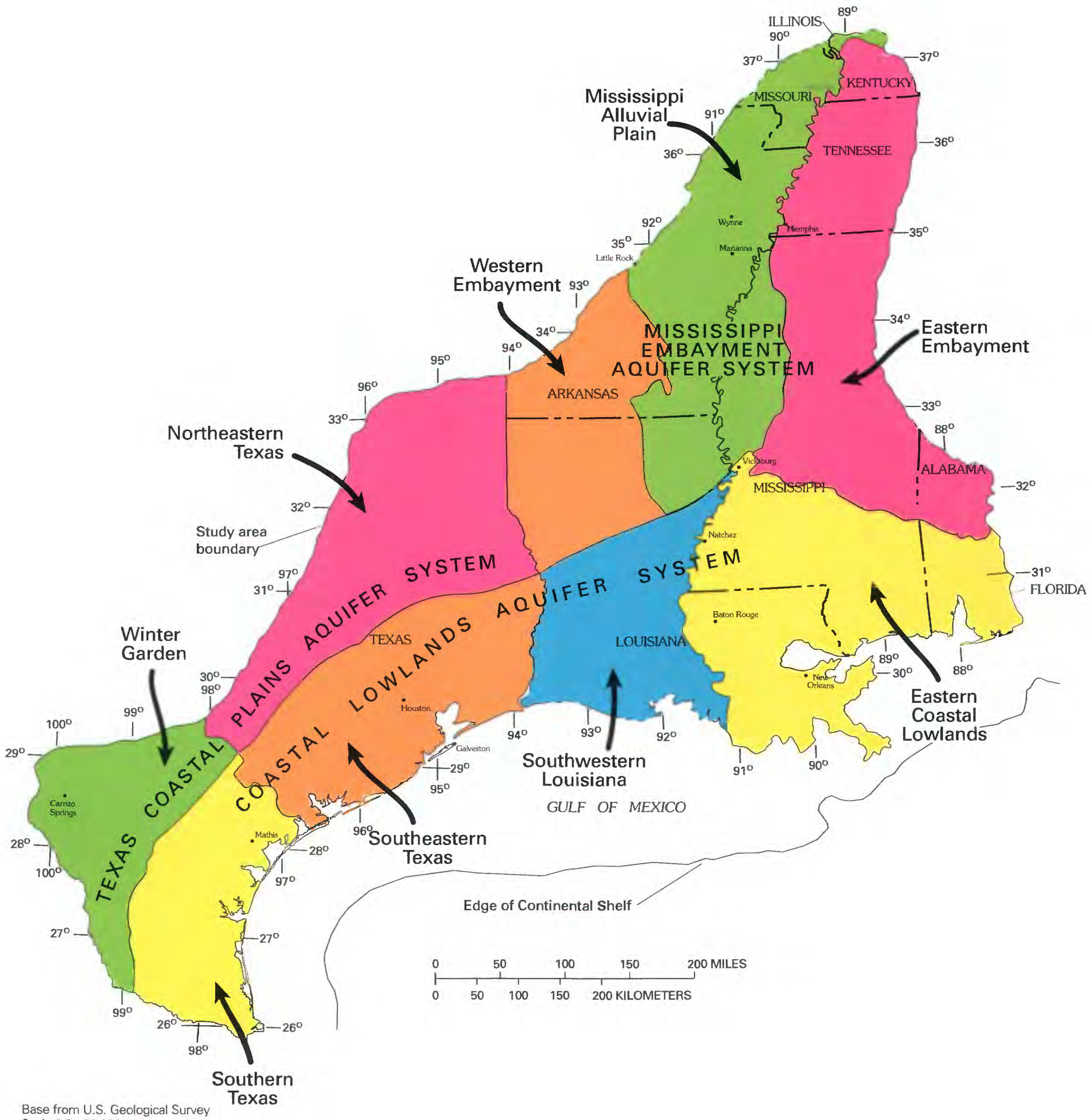


FIGURE 13.—Geographic areas used for comparisons of sand-bed hydraulic conductivities and flow summaries.

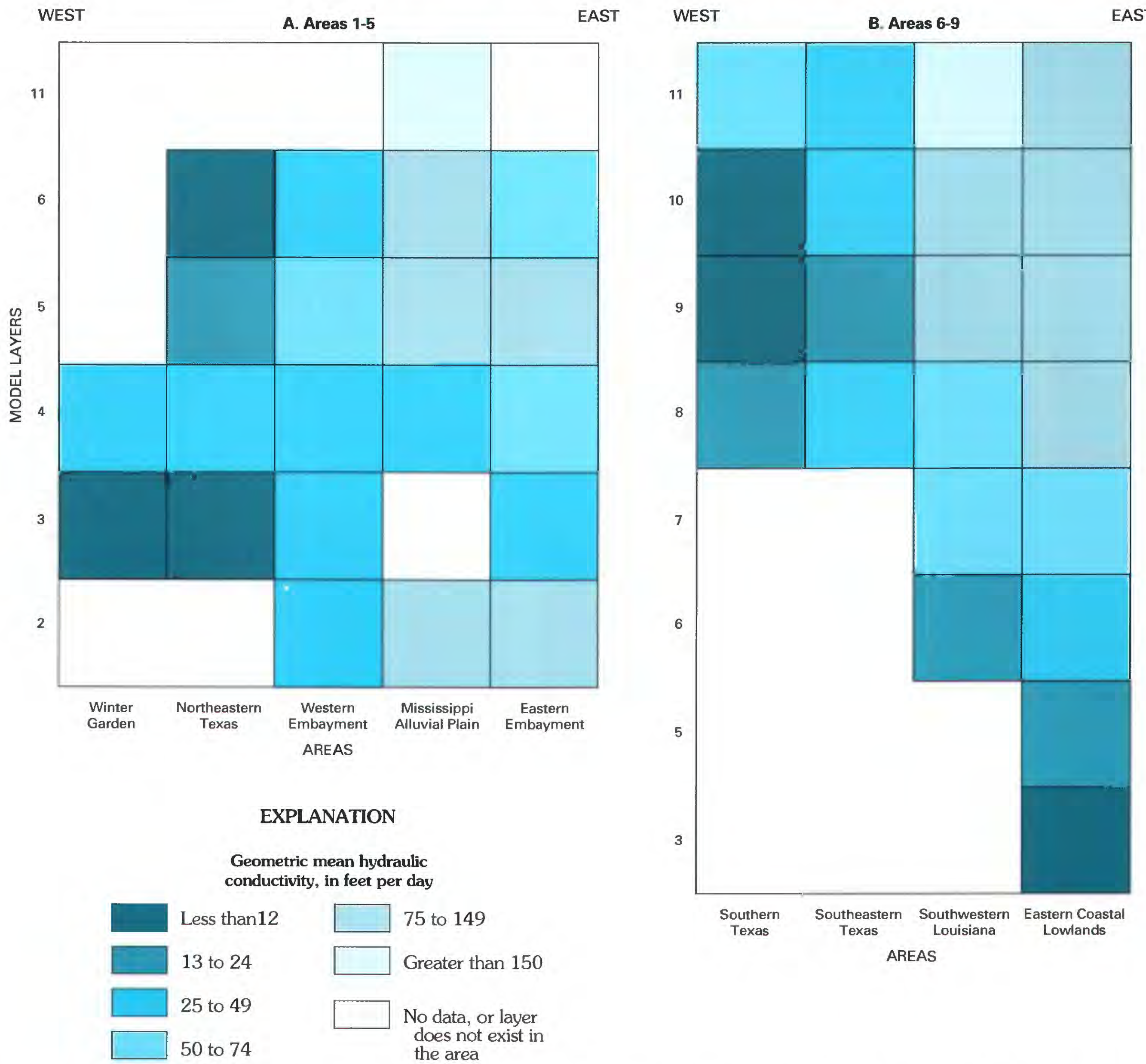


FIGURE 14.—Geometric mean horizontal hydraulic conductivity of sand beds by layer and area, estimated from aquifer-test analyses and specific-capacity data of: *A*. the Mississippi embayment and Texas coastal uplands aquifer systems, and *B*. the coastal lowlands aquifer system. Areas represent east-west variations.

of the values) of the individual bed conductivities. If the flow could go through any of a group of beds (analogous to parallel flows in electrical circuits), the effective conductivity would be equal to the arithmetic mean of the individual bed conductivities. For the case where the extent of the beds was much less than the size of the model blocks (fig. 16C) and the beds are randomly distributed and randomly oriented, the K_{eff} in any direction is equal to the geometric mean of all of the K values. However, for the case where the extent of individual beds varies from smaller to larger than the extent of the model blocks (fig. 16*B*), the horizontal K_{eff} is between the geometric and arithmetic means, and for the vertical, K_{eff} is between the geometric and harmonic means (Fogg, 1989, p. 46). This is the case in this study area, as well as many others, because there is so much variability in the extent of the beds, as well as the large variation in K values. As early as 1945, Cardwell and Parsons (1945) recognized that K_{eff} could not be calculated either by the harmonic or the arithmetic means but was somewhere in between.

TABLE 2.—*Comparison of hydraulic conductivity of sand beds (K_{sand}) by model calibration with mean estimated K_{sand} from aquifer-test analyses and specific-capacity data by group of model layers and areas*

[Estimates from Prudic, 1991, table 6. All values are in feet per day.[1] Summaries are not included for layers in an area with less than five estimates of hydraulic conductivity]

Group	Area[2]	Layer	Number of values	Final estimates[3] Arithmetic mean	Standard deviation	Harmonic mean	Geometric mean	Initial estimated geometric mean	Model calibrated
1	4	11	374	376	203	227	316	240	
Group composite:			374	376	203	227	316	240	247
2	4	2	18	187	193	95	128	111	
2	5	5	71	170	142	42	106	89	
2	8	11	601	270	207	103	189	144	
2	9	11	515	194	175	84	432	104	
Group composite:			1,205	231	194	87	156	122	123
3	4	5	147	119	123	41	75	52	
3	4	6	174	116	113	51	78	58	
3	5	2	107	145	162	5.4	79	65	
3	5	4	78	129	146	25	71	57	
3	6	11	16	201	235	15	69	62	
3	8	9	152	141	168	41	82	56	
3	8	10	293	147	154	48	88	60	
3	9	8	326	123	140	45	78	60	
3	9	9	334	129	121	45	85	67	
3	9	10	219	126	122	39	82	63	
Group composite:			1,846	131	139	30	81	61	64
4	1	4	44	63	56	25	42	36	
4	2	4	38	40	31	14	29	24	
4	3	5	492	94	112	30	57	39	
4	3	6	73	64	75	24	41	27	
4	4	4	31	76	73	24	48	35	
4	5	3	70	92	137	9.8	40	29	
4	5	6	32	90	73	19	60	47	
4	7	8	25	50	43	13	30	30	
4	7	11	231	79	96	30	49	42	
4	8	7	242	103	121	36	62	42	
4	8	8	196	97	116	26	59	41	
4	9	6	16	49	38	14	33	25	
4	9	7	186	95	110	15	56	42	
Group composite:			1,724	87	106	23	52	38	37
5	6	8	6	19	9.1	15	17	17	
5	7	9	213	24	18	15	19	16	
5	7	10	330	35	39	20	26	21	
5	8	6	11	28	13	19	24	16	
5	8	6	11	28	13	19	24	16	
5	9	5	29	65	70	1.9	20	13	
5	3	2	10	42	54	17	25	15	
5	3	3	458	56	85	8.3	26	16	
5	3	4	18	47	29	5.3	31	19	

TABLE 2.—*Comparison of hydraulic conductivity of sand beds (K_{sand}) by model calibration with mean estimated K_{sand} from aquifer-test analyses and specific-capacity data by group of model layers and areas*—Continued

[Estimates from Prudic, 1991, table 6. All values are in feet per day.[1] Summaries are not included for layers in an area with less than five estimates of hydraulic conductivity]

Group	Area[2]	Layer	Number of values	Final estimates[3]				Initial estimated geometric mean	Model calibrated
				Arithmetic mean	Standard deviation	Harmonic mean	Geometric mean		
Group composite:			1,088	42	62	10	24	17	17
6	6	9	23	8.6	7.2	3.9	5.8	5.2	
6	6	10	27	17	18	8.1	12	9.5	
6	9	3	10	13	14	.9	4.9	2.3	
6	1	3	10	17	23	3.2	8.3	4.7	
6	2	3	17	18	24	5.5	10	7.9	
6	2	5	224	19	17	10	14	12	
6	2	6	32	16	15	5.3	11	8.5	
Group composite:			22	17	21	5	10	7	7

[1]Hydraulic conductivities exceeding 1,000 feet per day were not included in the analyses.

[2]Areas shown in figure 13.

[3]Initial and final estimates refer to a change in the method of calculation of K_{sand} from aquifer-test analyses and specific-capacity data. Model was calibrated to match initial estimates before final estimates were available. See text, p.24.

The effect of different methods of calculating an effective hydraulic conductivity for model blocks where the K_{sand}, K_{fine}, and sand percentage are either known or estimated is shown in figure 17, where K_{fine} denotes the hydraulic conductivity of fine-grained beds. Sand percentage is the only readily usable and consistent measurement of the degree of connectedness between sand beds in the study area. Desbarats (1987) has presented a statistically based procedure for calculating grid-element effective hydraulic conductivity values, which in effect is similar to the method presented here, except that the method presented here is based on data (sand percentage) that are easily calculated. Effective horizontal and vertical hydraulic conductivity values, Kh and Kv, are calculated as functions (eqs. 1, 2) of sand percentage (S%), expressed as a ratio, and shown in figure 18, for the case where (K_{fine}/K_{sand})= 0.0001. Kh_{eff} and Kv_{eff} can be calculated from:

$$Kh_{eff}/ K_{sand} = [K_{fine}/K_{sand}]^{(S\%)^4} \quad (1)$$

$$Kv_{eff}/ K_{sand} = [K_{fine}/K_{sand}]^{(S\%)^{0.25}} \quad (2)$$

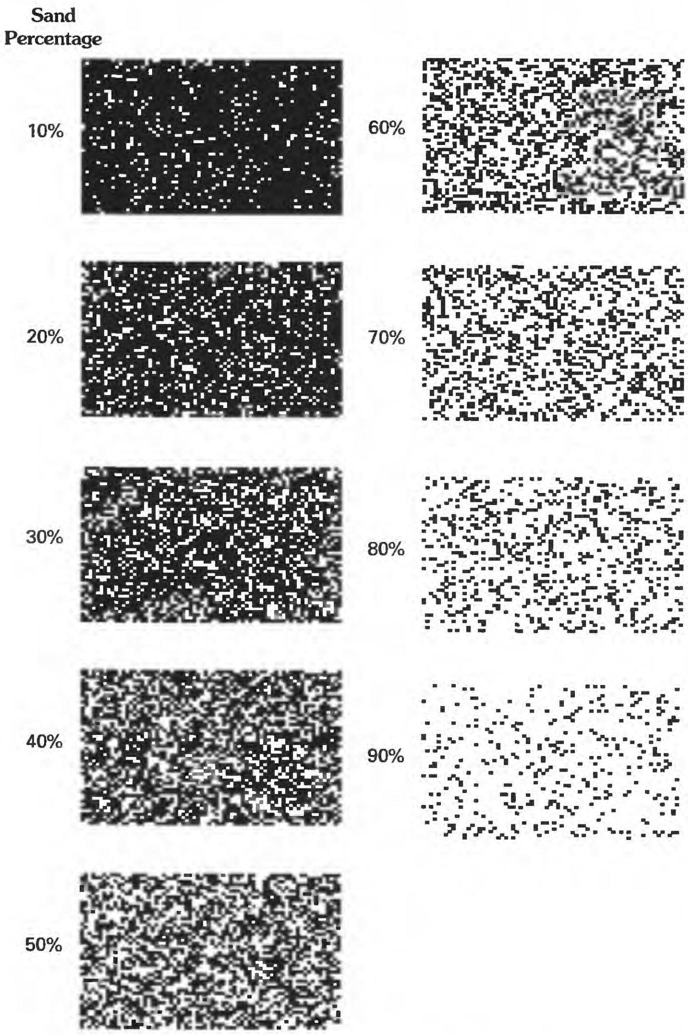
The exponents 4 and 0.25 were derived by Kuiper's (1994) regression model.

Sand percentage (S%) also varies widely among aquifers and across the study area (fig. 18). The S% implies the statistical frequency with which coarse-grained beds are interconnected, which largely determines the effective K and hence why it is a power function in equations 1 and 2.

VERTICAL HYDRAULIC CONDUCTIVITY OF FINE-GRAINED SEDIMENTS

Resistance to vertical flow is of two types: across regionally identifiable confining units, and within aquifers and permeable zones due to many beds of fine-grained materials that may have considerable areal extent yet are not traceable over multicounty areas. The deep cones of drawdown resulting from large ground-water pumpage at Houston and Baton Rouge demonstrate that large vertical resistance can occur even when only the latter type of resistance occurs. Several hundred feet of vertical-head difference has developed between the water table and the deep, intensively pumped sands, although there are no regionally identifiable confining units.

Vertical hydraulic conductivity (Kv) is nearly impossible to estimate because (1) laboratory values are seldom representative of undisturbed field values, (2) values range over many orders of magnitude (Freeze and Cherry, 1979, p. 29; Wolff, 1982, p. 36–90), and (3) the large variations generally can occur over short distances, making estimation of the effective value even more difficult. Kuiper's (1994) regression model estimated 0.11×10^{-3} ft/d as the best estimate of vertical hydraulic conductivity for fine-grained materials within confining units and 0.22×10^{-3} ft/d for fine-grained materials contained within aquifers. Note that the two values are quite close to each other considering Kv can vary several orders of magnitude. Less than 22 percent (Texas coastal uplands aquifer system and Mississippi embayment aquifer system) or 6 percent



Each of nine hypothetical sections through parts of the aquifer system with differing sand percentages represents hundreds of feet vertically and tens of miles horizontally. White denotes sand. Black denotes fine-grained beds

FIGURE 15.—Continuity of sand beds and fine-grained (clay and silt) beds in hypothetical sections composed of 10- to 90-percent sand.

(coastal lowlands aquifer system) of the fine-grained sediments is contained within the regional confining units. Most of the fine-grained beds occur within the regional aquifers and permeable zones. Vertical resistance to flow can be proportional to the net thickness of fine-grained beds whether they have regional lateral extent or not. Regionally mappable, fine-grained units may cause most of the overall resistance to vertical flow where they exist, especially if they comprise the bulk of the thickness of fine-grained material at those locations.

The concept of large vertical resistances to flow due to fine-grained beds within the aquifers can be illustrated using the results of previous studies such as that

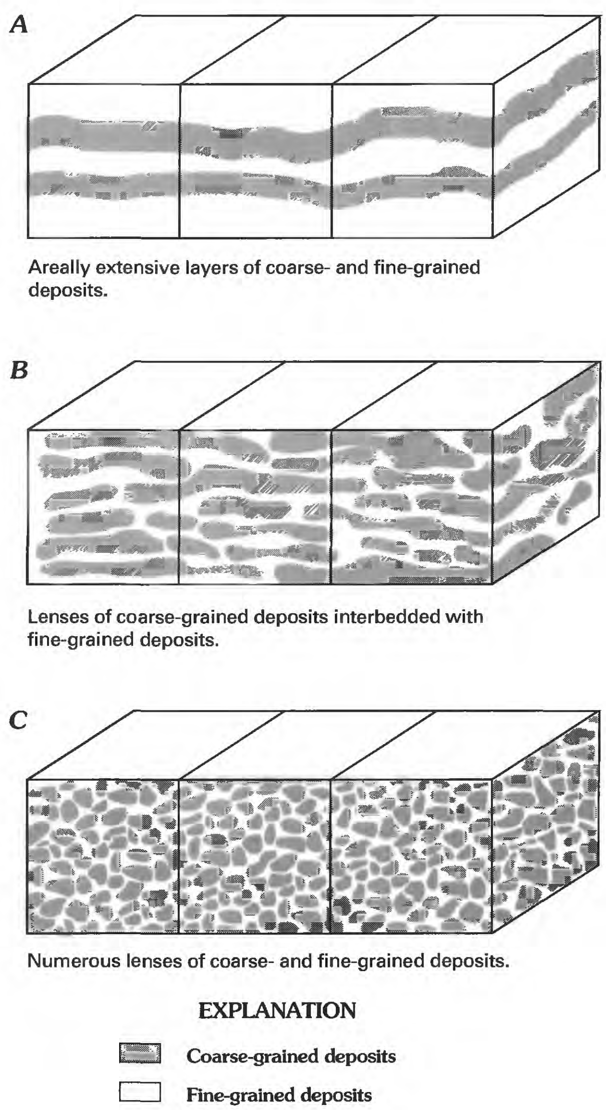
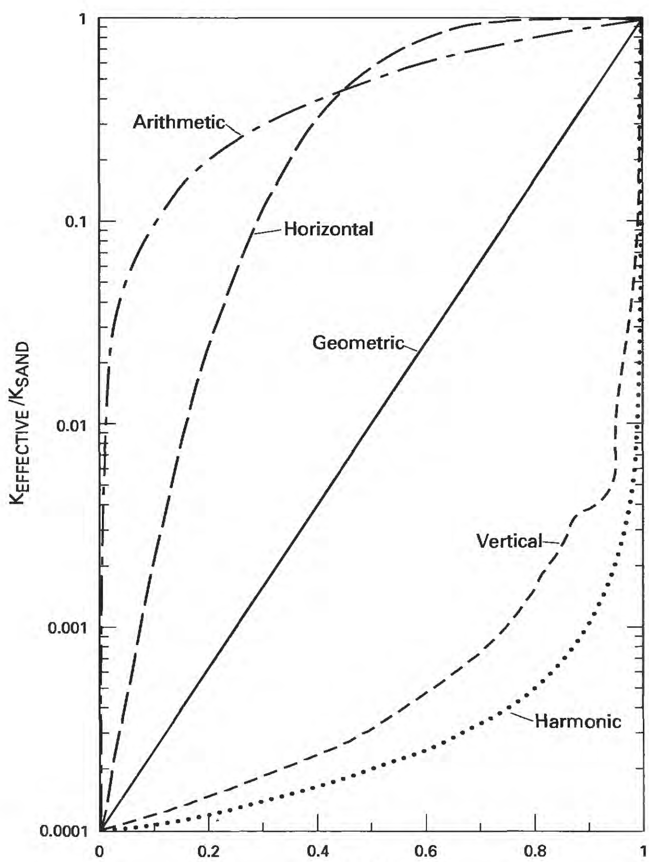


FIGURE 16.—Model blocks with varying extents of coarse- and fine-grained beds: *A*. beds extend beyond model blocks, *B*. most beds do not extend beyond model blocks, and *C*. extent of beds generally much smaller than model blocks.



Graph construction assumption is that K_{sand}/K_{fine} equals 10,000, although vertical axis could be adjusted to other ratios.

FIGURE 17.—Effective hydraulic conductivity (K_{eff}) as a function of the ratio of sand to total volume.

by Wesselman (1967). He plotted hydraulic head relative to depth (Wesselman, 1967, p. 57) for an interval containing the Chicot, Evangeline, and Jasper aquifers and the Burkeville confining unit at Evadale, Texas, near the type locality of the Burkeville confining unit. The plot shows a hydraulic head change of 67.2 ft over a depth interval of approximately 430 ft across the Burkeville, which is a vertical hydraulic gradient of 0.16 ft/ft. However, on the same plot, a hydraulic-head change of 18.9 ft occurs across 130 ft of an unnamed fine-grained bed in the middle part of the Evangeline aquifer, indicating a hydraulic gradient of 0.15 ft/ft. This similarity in hydraulic gradients suggests that the numerous smaller fine-grained beds within the aquifer system may cumulatively provide nearly the same resistance to vertical flows as larger, more regionally mappable confining units. This is only one example, but numerous other examples exist.

DEPTH DEPENDENCE OF HYDRAULIC CONDUCTIVITY

The hydraulic conductivity of the sediments tends to decrease with depth due to increasing overburden pressure causing compaction and a corresponding decrease in porosity. However, this decrease is moderated because hydraulic conductivity also tends to increase with depth due to higher natural geothermal temperature and the resulting decrease in fluid viscosity. This temperature effect can change the viscosity by a factor of nearly five (Rouse, 1950, p. 1008; Weiss, 1982) from near land surface where the water temperature is about 21°C to a depth of over 10,000 ft where the temperature can be as high as 149°C (Wilson and Hosman, 1988). The decrease in viscosity with depth is partly compensated by an increase in density of water downdip, due to increased salinity. The increase in salinity causes a slight increase in viscosity and a corresponding decrease in hydraulic conductivity. The net effect of these factors, however, is toward decreasing hydraulic conductivity with depth.

Kuiper (1994) estimated that K_{sand} varies by depth in proportion to $10^{(-0.000243D)}$, where D is depth in feet. Kuiper used data from Loucks and others (1986) and Lake and Carroll (1986). Data on the sand hydraulic conductivity of deposits similar to those of this study are shown in figure 19, combining data from oil-well tests (Loucks and others, 1986) and from water-well aquifer tests (Prudic, 1991). Kuiper's curve, $(30)10^{(-0.0002493D)}$, which is an approximation assuming the constant for K_{sand} is 30 ft/d, is also shown in figure 19. The exponent of this curve was assigned a regression parameter by Kuiper (1994), but because of the shallow depth (D) of most head observations, the parameter had such a large confidence interval that it was decided that the model would be more accurate if the curve was assumed fixed. Prudic (1991) found the variation of K_{sand} with depth to be significant at the 95-percent confidence interval but only in an analysis of variance where each area-layer combination was treated separately. However, about 9 of the more than 40 area and layer combinations had signs for the coefficients in the equations, indicating the wrong direction of the relation. These tended to be groups of data where there was lower confidence in the relation as well. A simple log linear regression was performed on both data sets shown in figure 19, yielding the equation, $K = (54.5)10^{(-0.000309D)}$, with an adjusted r^2 of 0.736. This equation is similar to Kuiper's (also shown in fig. 19), so Kuiper's equation was used.

The vertical hydraulic conductivity of the fine-grained beds, like that of the sands, also tends to decrease with depth. The primary cause for this is compaction. Neglia (1979) presented data for the

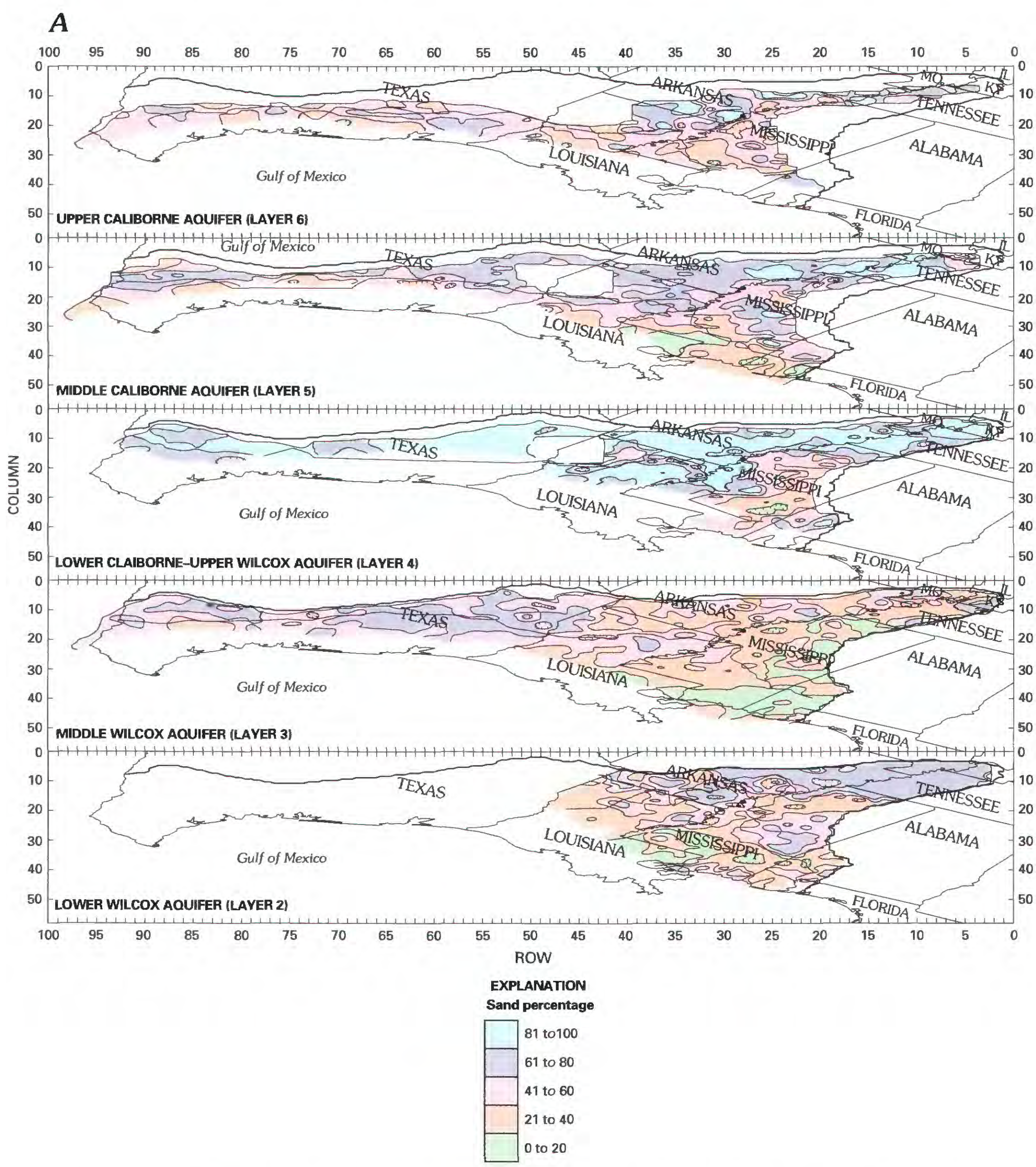


FIGURE 18.—Perspective diagrams showing sand percentage in each aquifer or permeable zone (layer) of the gulf coast aquifer systems in: *A*. the Mississippi embayment and Texas coastal uplands aquifer systems, and *B*. the coastal lowlands aquifer system.

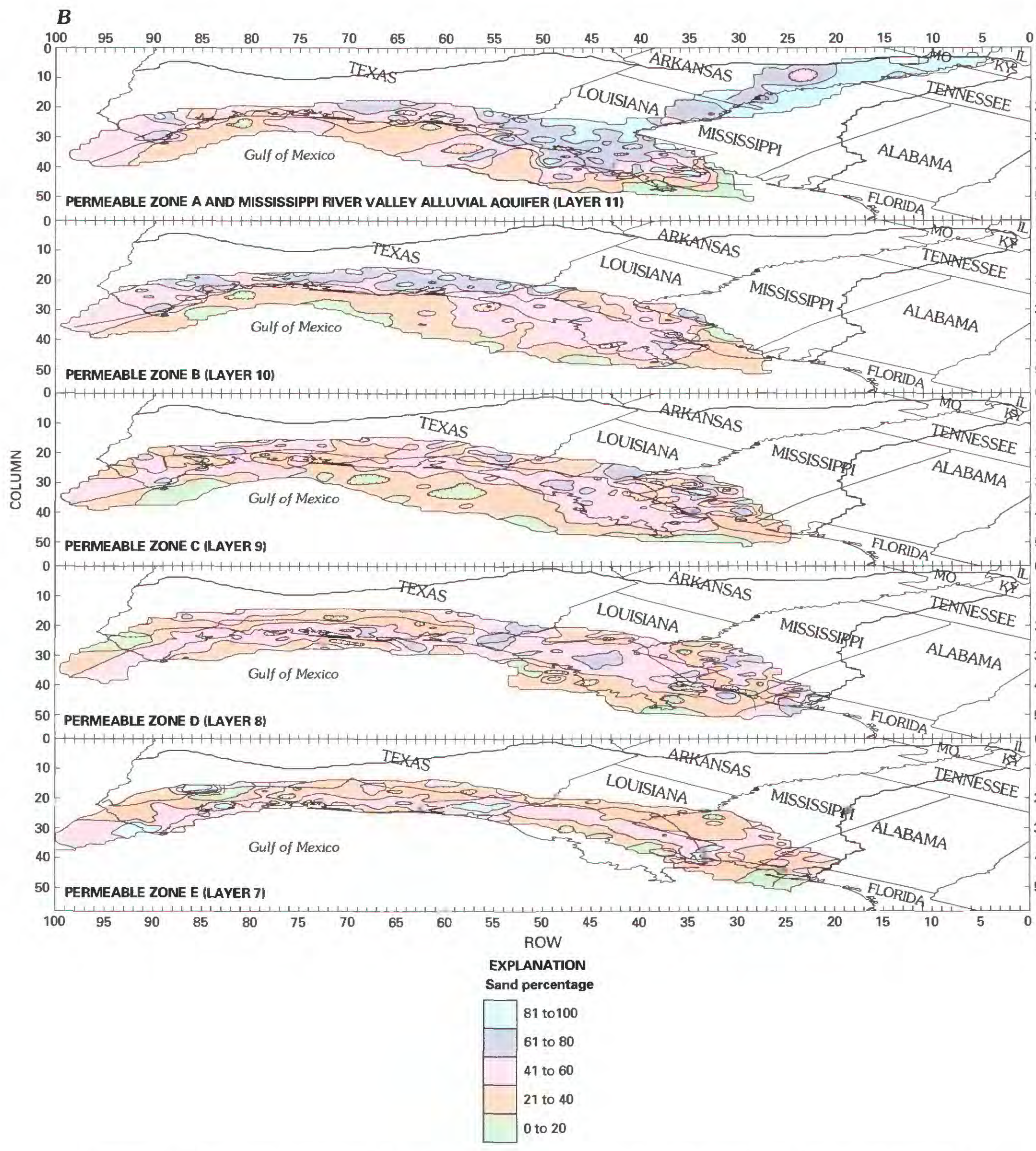


FIGURE 18.—Perspective diagrams showing sand percentage in each aquifer or permeable zone (layer) of the gulf coast aquifer systems in: *A*. the Mississippi embayment and Texas coastal uplands aquifer systems, and *B*. the coastal lowlands aquifer system—Continued.

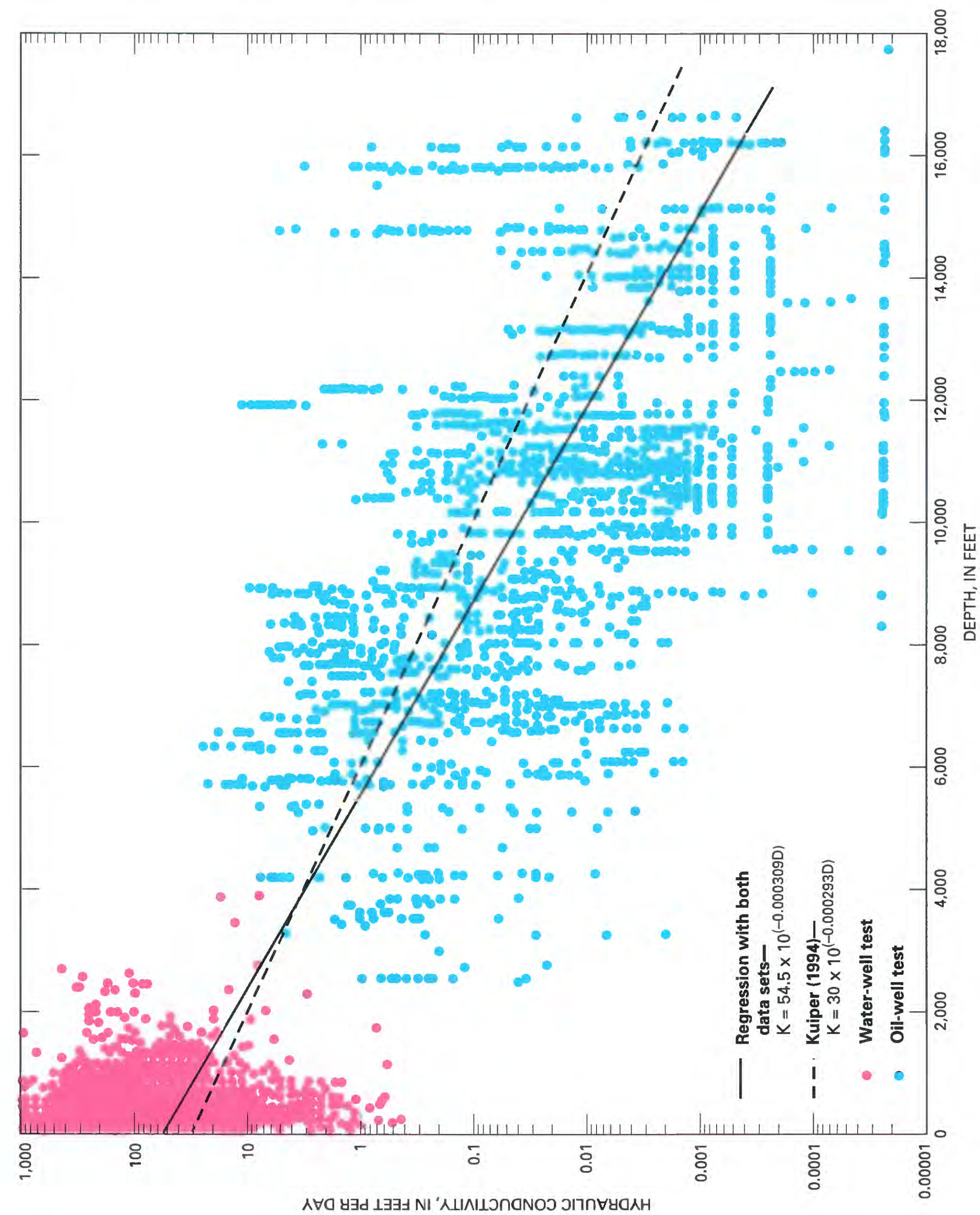


FIGURE 19.—Decrease of sand-bed hydraulic conductivity with depth.

decrease with depth of the hydraulic conductivity of various clay samples. Kuiper (1994, p. 48 and fig. 30) used Neglia's data (1979) and formulated a regression equation to approximate the measured values, (constant)$10^{(-0.000356\ D\ +\ 0.0000254\ D^{-2})}$, where D is in feet, which was used in this study. As with the decrease of sand hydraulic conductivity with depth, a regression parameter that Kuiper (1994) placed into the exponent had a very large confidence interval and was removed so that the depth relation was assumed fixed.

After all of these factors are accounted for in estimating regional effective hydraulic conductivity in the horizontal and vertical directions for each model block in every layer, the resulting K_{eff} values span a wide range (fig. 20). Both Kh_{eff} and Kv_{eff} span 6 to 7 orders of magnitude, although most of the blocks are in a more limited spread. The approach used maximizes the use of measurable data to estimate K_{eff} across a broad range of values (fig. 20).

The areal distribution of effective horizontal hydraulic conductivity shows some interesting patterns, more obvious than any shown by component distribution (fig. 21). The west-to-east trend of increasing sand-bed hydraulic conductivity and also sand percentage combine to show an even more dominant east-west trend of increasing effective conductivity. The pattern in layer 11 also shows an increased conductivity in the Rio Grande embayment (fig. 7), probably due to deposits from higher energy stream deposits, similar to what appears in the Mississippi embayment, but to a lesser extent. A lower conductivity north-south band occurs in layers 2 and 4 in north Louisiana and south-central Arkansas.

BRINES AND DENSITY

Variations in concentration of dissolved solids, and hence density, have a substantial effect on flow in the aquifer system because about two-thirds of the volume of the aquifer systems studied contain saline water—that is, water having a concentration of dissolved solids greater than 10,000 mg/L (Kuiper, 1994, p. 12 and fig. 8). About one-half of the volume of the aquifer systems contains water saltier than seawater or brine (fig. 22). Generally, the transition from a relatively fresh water (dissolved-solids concentration less than

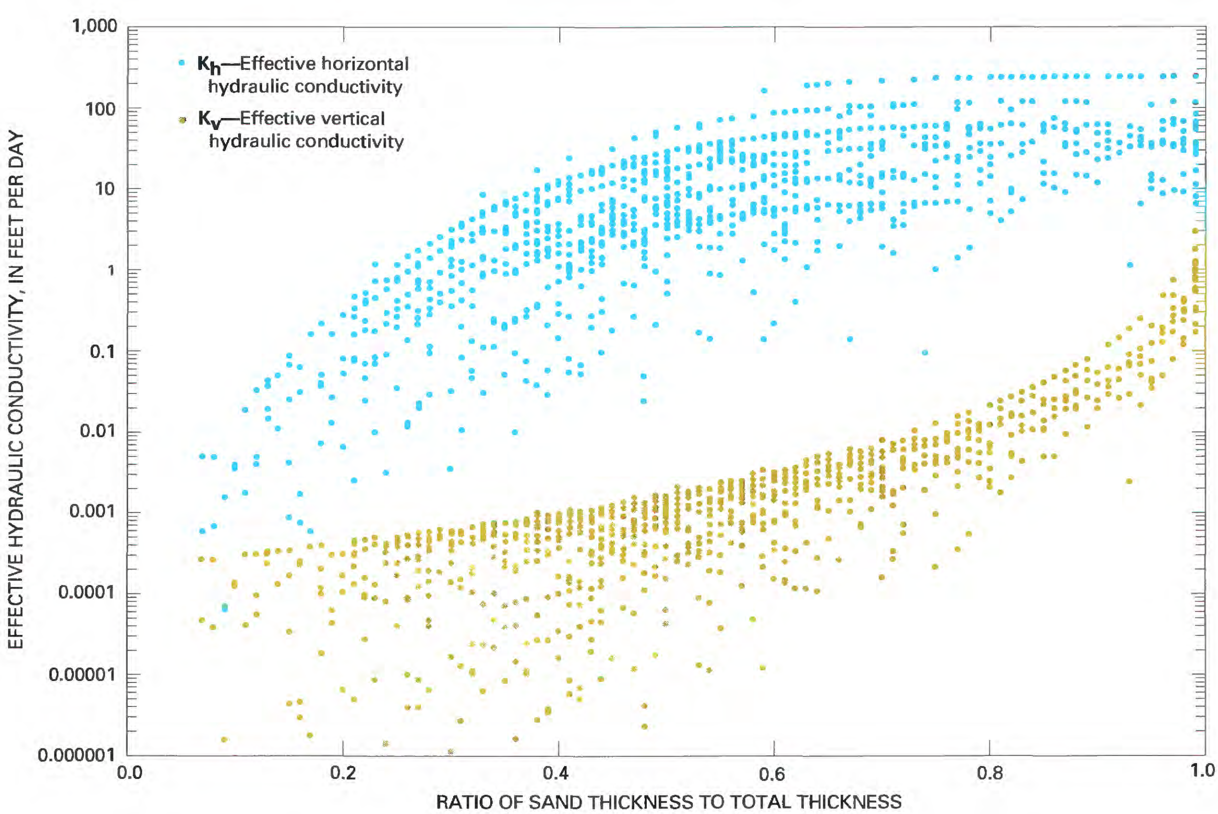


FIGURE 20.—Relation of horizontal and vertical effective hydraulic conductivity to sand proportion in the gulf coast aquifers systems.

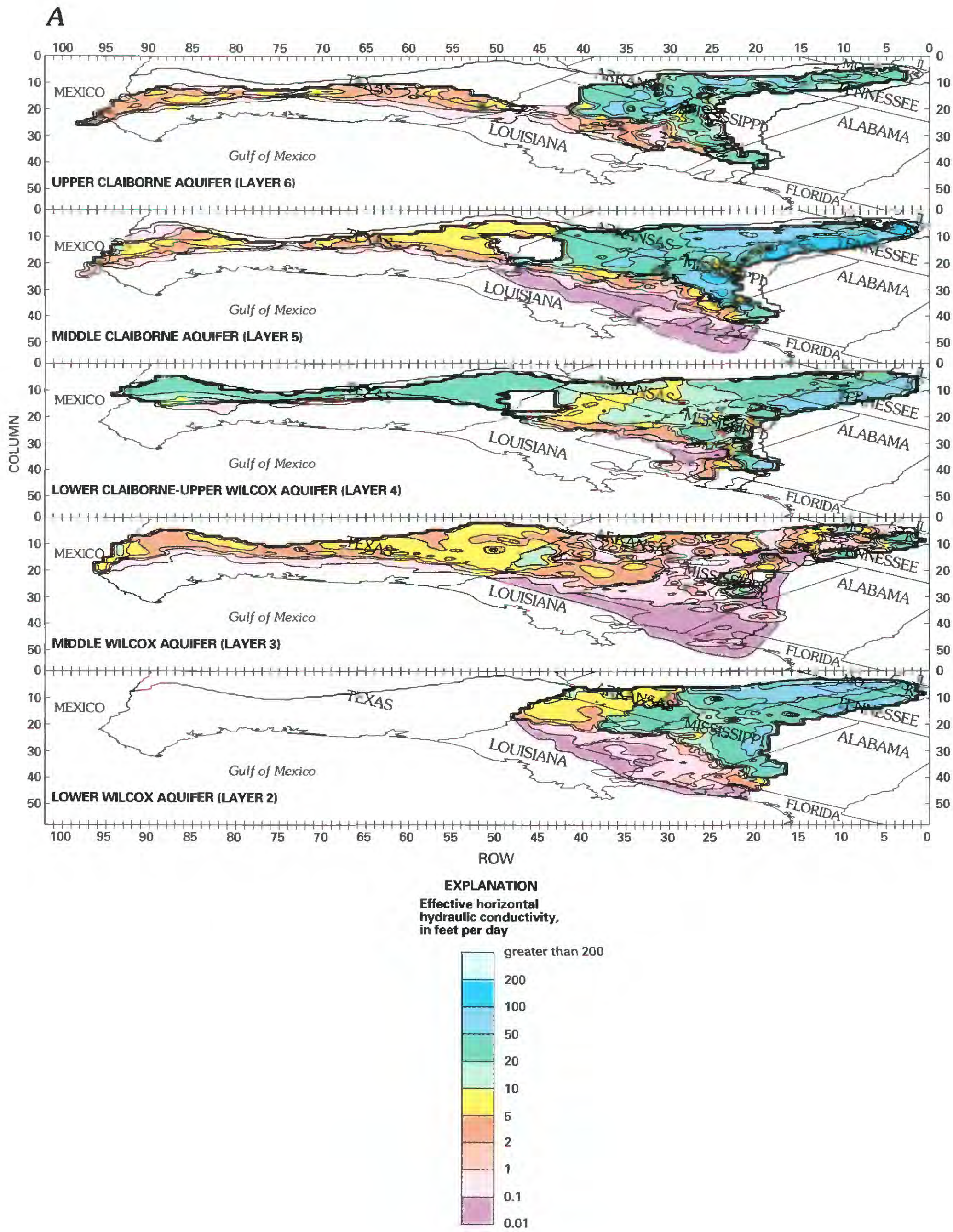


FIGURE 21.—Perspective diagrams showing regional effective horizontal hydraulic conductivity in each aquifer and permeable zone (layer) of the gulf coast aquifer systems in: *A*. the Mississippi embayment and Texas coastal uplands aquifer systems, and *B*. the coastal lowlands aquifer system.

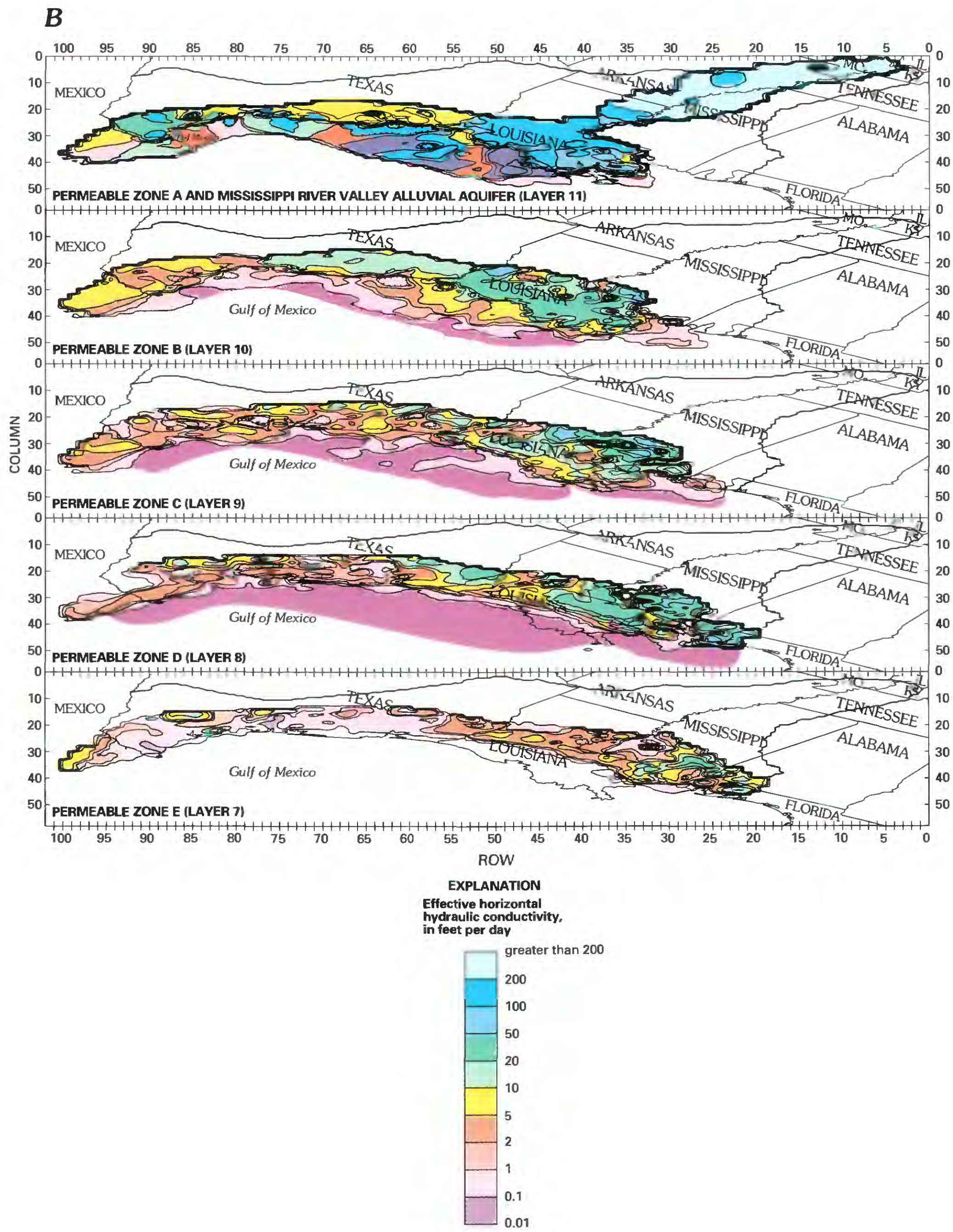


FIGURE 21.—Perspective diagrams showing regional effective horizontal hydraulic conductivity in each aquifer and permeable zone (layer) of the gulf coast aquifer systems in: *A*. the Mississippi embayment and Texas coastal uplands aquifer systems, and *B*. the coastal lowlands aquifer system—Continued.

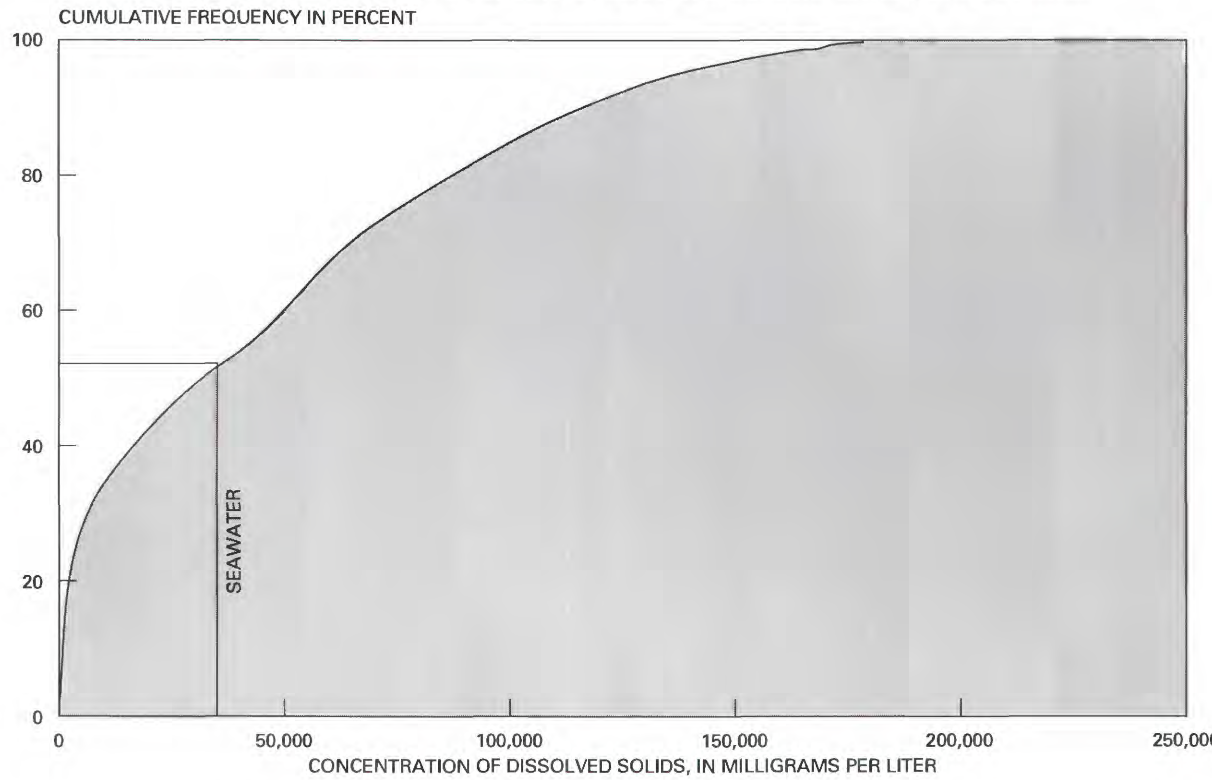


FIGURE 22.—Frequency distribution of concentrations of dissolved solids in water from about 20,000 sand beds in the gulf coast aquifers.

10,000 mg/L) to a brine (dissolved-solids concentration greater than 35,000 mg/L) occurs within a distance of about one to three model-block widths (10–30 mi).

Dissolved solids and temperature were mapped by Pettijohn and others (1988). The complicated distribution of dissolved solids in the aquifers and permeable zones is shown in figure 23. The approximate location of the 35,000-mg/L dissolved-solids concentration line (saline/freshwater interface) is also shown in figure 12 for each layer. The distribution of brine concentrations also is variable when looking at a vertical slice of the aquifer system (fig. 24). For example, the well log at mile 40 of section G–G' shows substantial variability in dissolved-solids concentrations between adjacent sand beds, especially in layer 3. The vertically weighted average concentration of dissolved solids for each permeable zone at well locations is plotted on the sections shown on plate 1.

ESTIMATION OF DENSITY

Water density is affected by the concentration of dissolved solids, temperature, and hydrostatic pressure. Dissolved-solids concentrations were estimated by Williamson and others (1990, p. 63–65) from water resistivities obtained from the spontaneous potential curve of the 989 electric logs as described by Weiss (1987). A computer program requiring an iterative solution, developed by Weiss (1982) and modified by Kontis and Mandle (1988), was used to estimate water density from concentration, temperature, and formation pressure based on equations and coefficients of Potter and Brown (1977). To simplify this calculation, Williamson and others (1990) presented a linear multiple regression using these data and estimates of density for 15,200 model blocks that have horizontal dimensions of 5 mi on each side. The regression equation estimated density within a standard error of estimate of 0.0020 g/cm^3, about the value estimated by the iterative program. The regression equation is:

$$DENGW = 0.000648\ TDS - 0.000368\ TEMPC + 0.0000015\ D + 1.00472 \tag{3}$$

where

DENGW = density of the ground water at the pressure and temperature in the aquifer, in grams per cubic centimeter,

TDS = dissolved solids, in grams per liter,

TEMPC = temperature, in degrees Celsius, and

D = depth, in feet.

The equation in Williamson and others (1990, p. 65) is given in terms of pressure in bars rather than depth in feet. Equation (3) was converted to the more convenient value of depth by assuming a hydrostatic gradient of 0.444 pound per square inch per foot [(lb/in^2)/ft] that

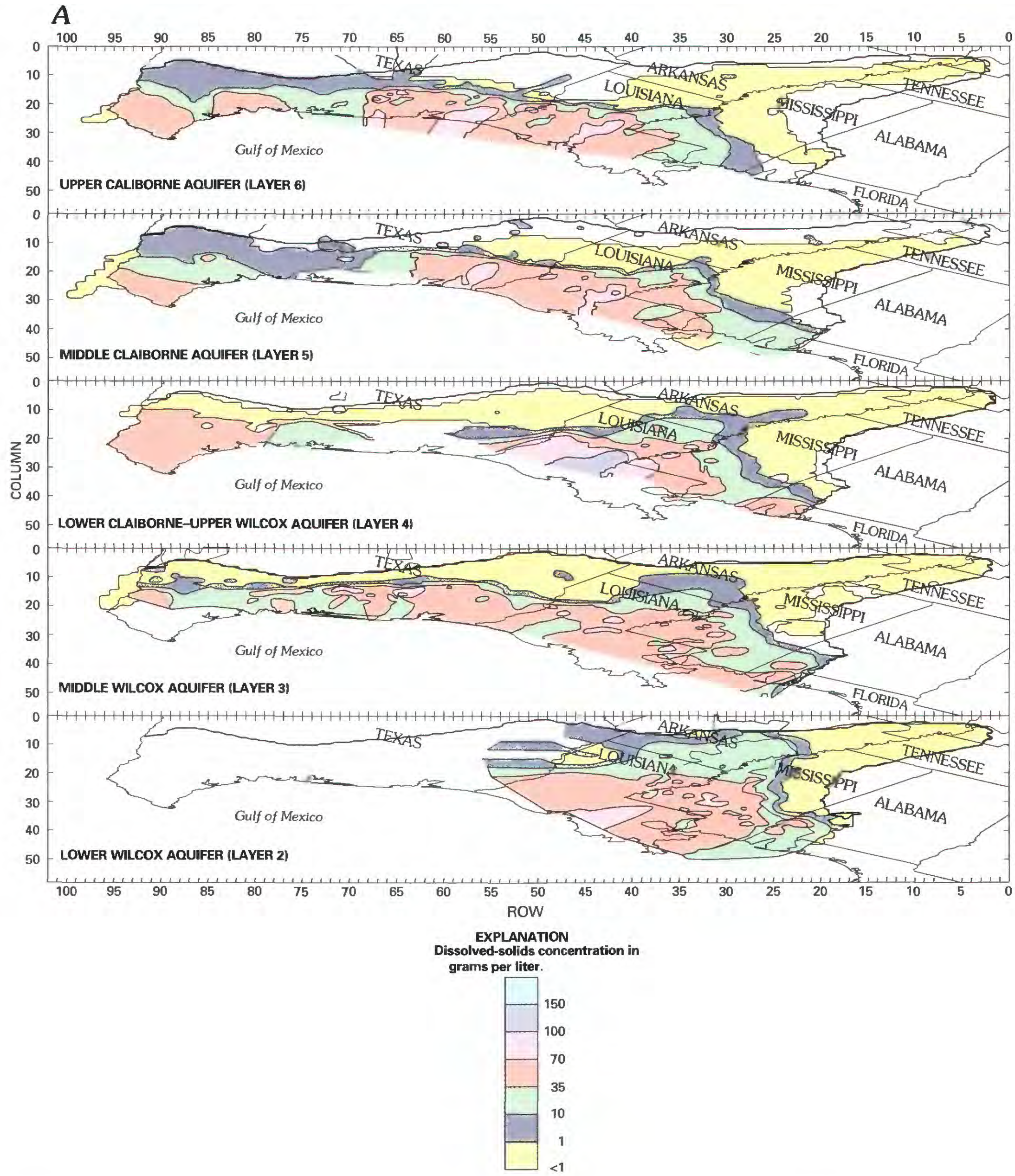


FIGURE 23.—Perspective diagrams showing distribution of dissolved-solids concentrations in: *A*. the Mississippi embayment and Texas coastal uplands aquifer systems, and *B*. the coastal lowlands aquifer system.

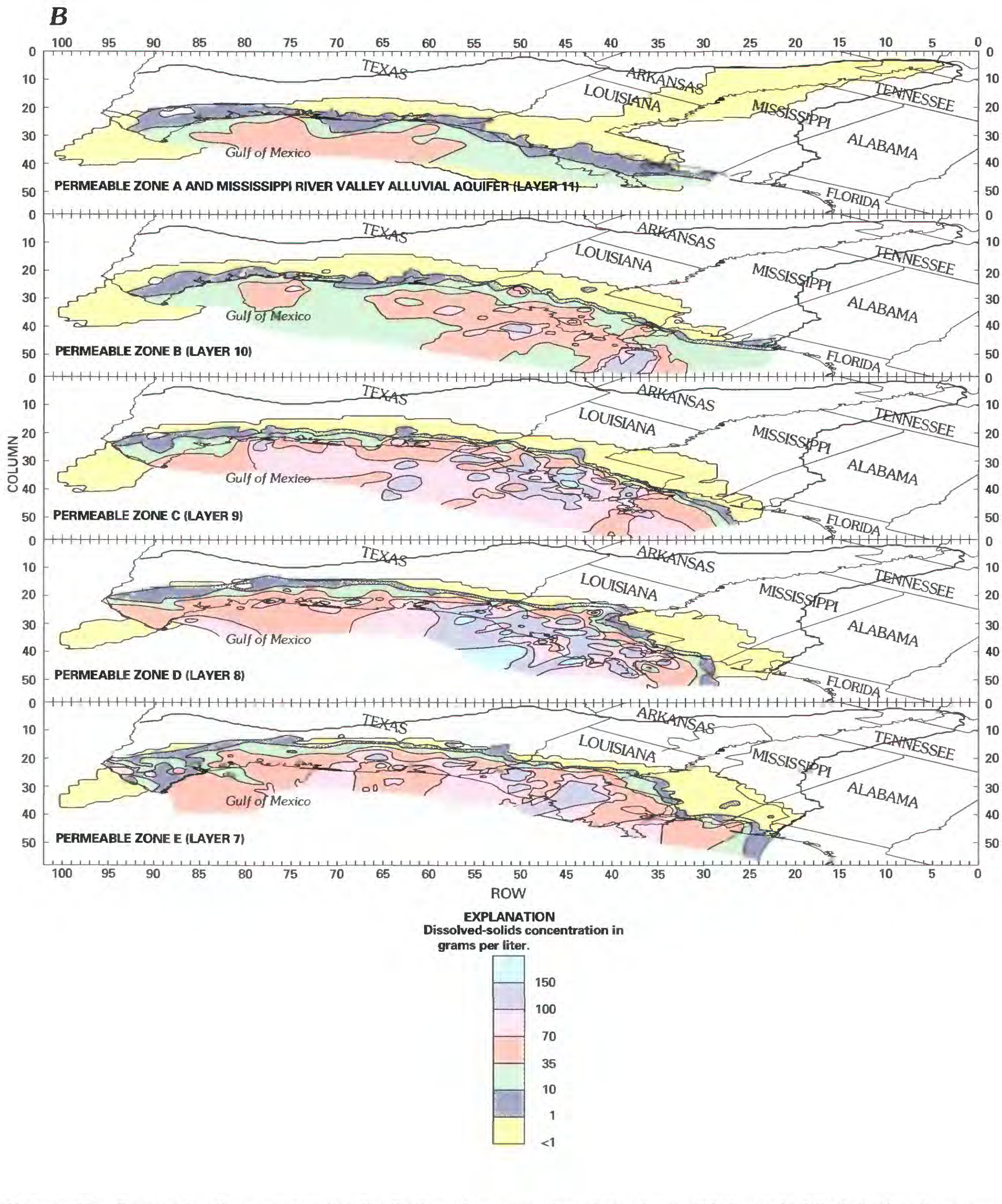


FIGURE 23.—Perspective diagrams showing distribution of concentration of dissolved solids in: *A*. the Mississippi embayment and Texas coastal uplands aquifer systems, and *B*. the coastal lowlands aquifer system—Continued.

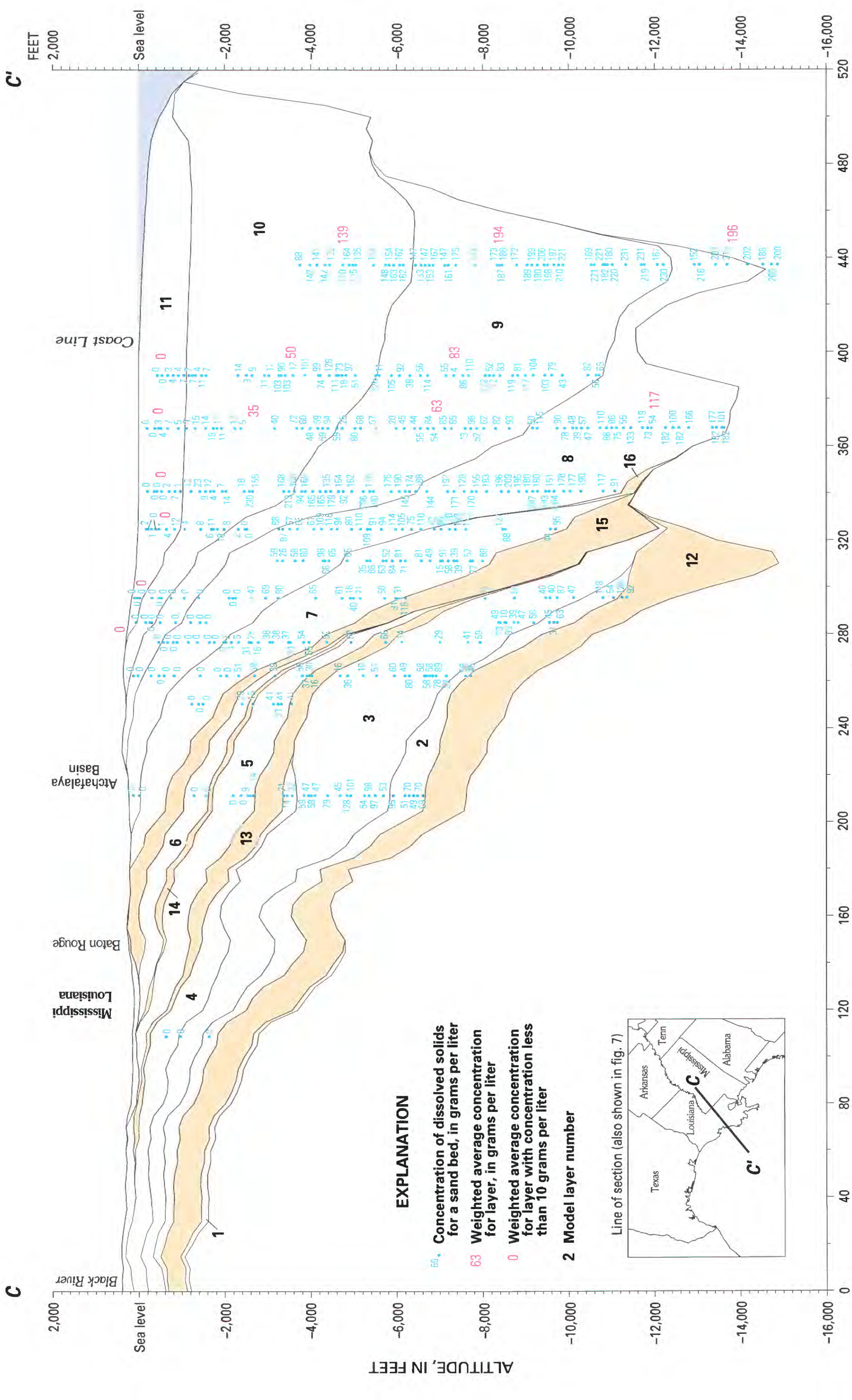


FIGURE 24.—Vertical distribution of dissolved-solids concentrations in sand beds thicker than 20 feet along section C–C'.

corresponds to a density of 1.025 g/cm^3. This approximation is adequate in many cases because pressure is the least sensitive factor. The variation of densities of water from the gulf coast regional aquifer systems is shown in figure 25. A substantial volume of the water has density less than 1.0 g/cm^3 because temperature is elevated at depth and dissolved-solids concentrations are minimal at some locations. This is especially true in some of the deeper parts of the Mississippi embayment aquifer system, which contains mostly freshwater.

HYDRAULIC HEAD AND FLOW IN VARIABLE-DENSITY AQUIFERS

To understand the regional flow patterns in the gulf coast regional 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 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 top and bottom, which will be referred to as "gravitational effects." This second force affecting flow can be understood by taking the simple case where a bucketful 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—hydraulic gradient and gravitational effects due to differences in water density and altitude of the aquifer.

Therefore, in aquifers with variable-density water, hydraulic gradients, expressed either in terms of equivalent freshwater head or formation-water head, do not necessarily indicate flow directions or magnitudes. Neither do bottom hole pressures, commonly measured during drill-stem tests 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," Φ, between points *a* and *b* in the aquifer, as:

$$\Phi = (Ha - Hb) - \int_a^b (r-1)dz \qquad (4)$$

where

Ha, Hb	= equivalent freshwater head = (Z, altitude, plus P/rg, the pressure head), at points *a* and *b*,
$\int_a^b$	= integral along flow path from point *b* to *a*,
r	= relative density of the ground water under the temperature and pressure of its sampling, compared with the density of pure water at 24°C and 1 atmosphere pressure, and
dz	= the change in altitude in the aquifer between points *a* and *b*.

The difference in the first two terms, $Hb - Ha$, is the familiar difference in hydraulic head, with the minor adjustment of observed water level in a well 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 that restrict 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). 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 filled sufficiently 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 variation of formation-water head for layer 8 based on simulation under predevelopment conditions is shown in figure 26. Note that the formation-water head map (fig. 26) has depressions in the brine area that appear

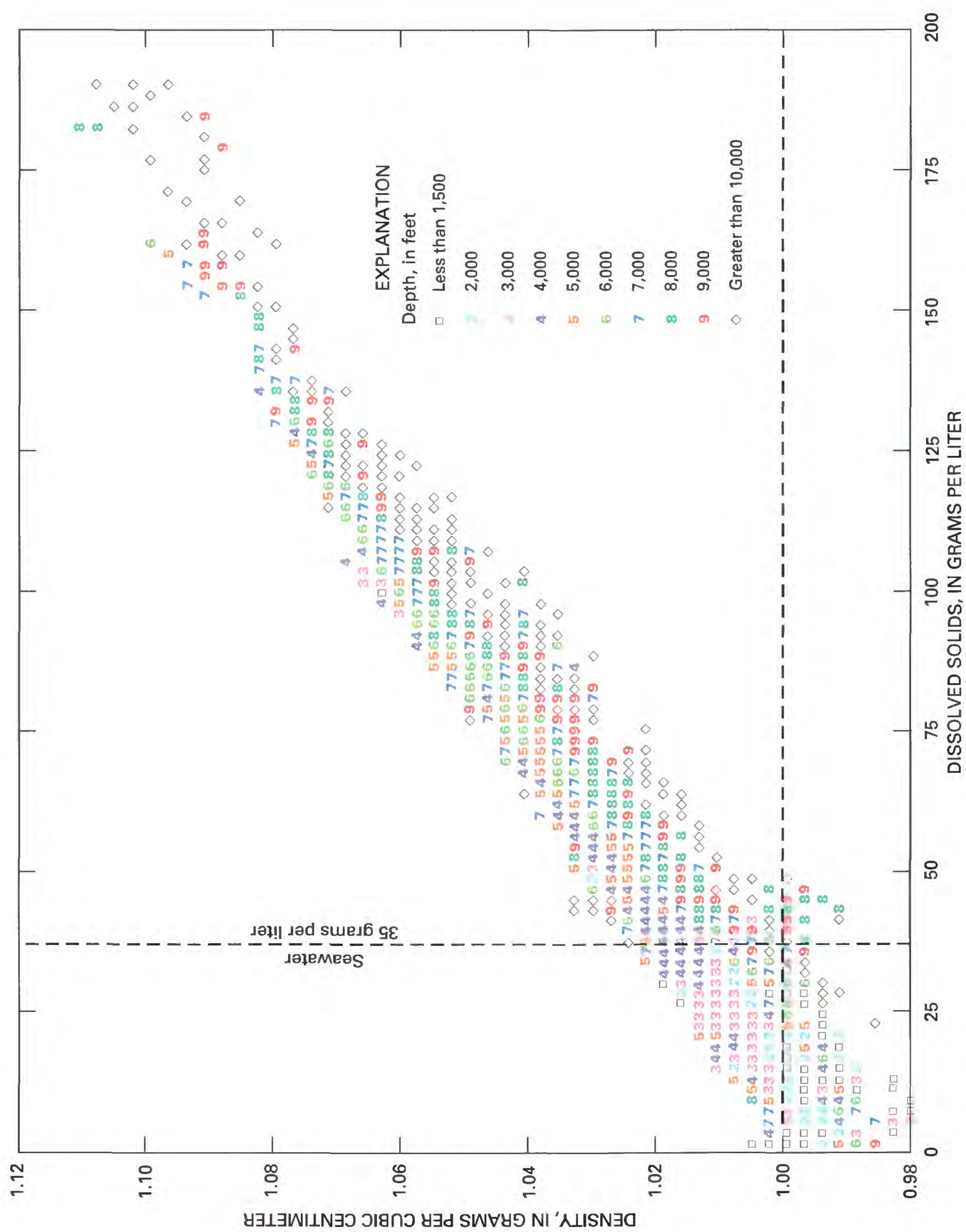


FIGURE 25.—Density of water in the gulf coast aquifer systems as a function of concentration of dissolved solids and depth.

like cones of drawdown, even though the map is from a steady-state, predevelopment simulation. The closed depressions are due to the variation in water density and sloping of the layers.

The freshwater head declines from more than 300 ft in the higher parts of the outcrop area to nearly zero and increases to more than 700 ft near the deepest downdip extent of the layer (pl. 2*D*). This would be impossible in a freshwater aquifer, but it is possible in a system with variable density flow. The flow vectors (pl. 3*D*) are not necessarily perpendicular either to the freshwater head contours or the formation-water head contours in the brine part of the aquifer system.

EFFECTS OF DENSITY

The density of the water in the deeper parts of the aquifer systems has a substantial effect on the ground-water flow in the systems—especially in the saline section, but also in the freshwater section. The simulations presented in this report do not assume that the brine is stagnant. Although often assumed, it is unknown whether the brine part of the system is in equilibrium with other large and long-term changes such as sea-level change and natural sediment compaction. Equilibrium conditions probably have not yet been achieved due to the relatively low flow velocities and the long flow paths to a discharge point.

The effect of spatially variable density is usually ignored in ground-water flow models by imposing a no-flow boundary at the saline/freshwater interface. In certain cases this sharp interface assumption is justified. The sharp interface assumption requires that the brine part of the system be in equilibrium with the freshwater part of the system so that freshwater flow is the only dynamic driving force. This assumption is adequate in cases like small-scale seawater intrusion where there are no other heat or salt sources so that the saline ground water has had enough time to reach equilibrium. In the study area, there are many forces which apparently have prevented the brine part of the system from reaching equilibrium, such as heat and salt sources, sediment compaction, geopressured sediments, changing sea level, and a very wide continental shelf that makes for longer flow paths to discharge water.

Although density is assumed to vary spatially, it is assumed to be constant in time so that the effect of solute transport on ground-water flow can be ignored. This assumption is valid over the relatively short (a few decades) time scales used in the transient simulations because the volume of water that can move into an adjacent block at maximum flow rates will not substantially affect the average density of the water in these very large blocks in the regional model.

Due to the complexity of the aquifer systems and the effect of density on ground-water flow, the flow paths in and near the area containing brine are very complex in all three dimensions (fig. 27). The complicated three-dimensional flow pattern is difficult to display and fully visualize; however, some patterns can be seen by looking at Darcy velocity in three dimensions as shown in figure 27. Recharge in outcrop areas of each layer near the Mississippi-Louisiana State line moves downdip and south and then to the west, discharging in east-central Louisiana.

Several types of simulations confirmed the large effect of density on flow. One test compared two simulations that were identical except that one simulation specified the density for all the model blocks to 1 g/cm^3 (the density of freshwater) instead of the actual estimated density. Even at Memphis, Tennessee, which is about 200 mi updip from the limit of freshwater in layer 5 (fig. 12), the difference in simulated head was nearly 1 ft (Williamson and others, 1990, p. 103). 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 farther away.

In addition, all simulations indicate that substantial volumes of flow occur in the saline part of the aquifer system (pls. 3, 6). The distribution of salinity has a high degree of local variability (pl. 1, figs. 23, 24); consequently, the relatively sparse spatial-data density (fig. 8) available for this analysis contains variability in addition to the regional trend that is still discernible. Williamson and others (1990, p. 103–106) tested the effect that local variation in density had on the flow simulation by smoothing the density data sets for each layer using an areal moving average. The smoothed and unsmoothed density data sets were used to make otherwise identical simulations. Some small-scale variations were eliminated by the smoothed density data sets; however, the regional pattern of flow was very similar to the flow using the unsmoothed density data sets (Williamson and others, 1990, figs. 26*G*, 30). All simulations presented in this report were made using the smoothed density data sets to show the effects of the regional trends in density.

Flow in the brine part of the aquifer systems appears not to have reached equilibrium. The time required for the flow system to adjust to processes such as sea-level change, additional sedimentation and compaction, and salt dissolution is long, perhaps thousands of years. Such time spans are owing to the very long flow paths necessary for water to move to a discharge point and

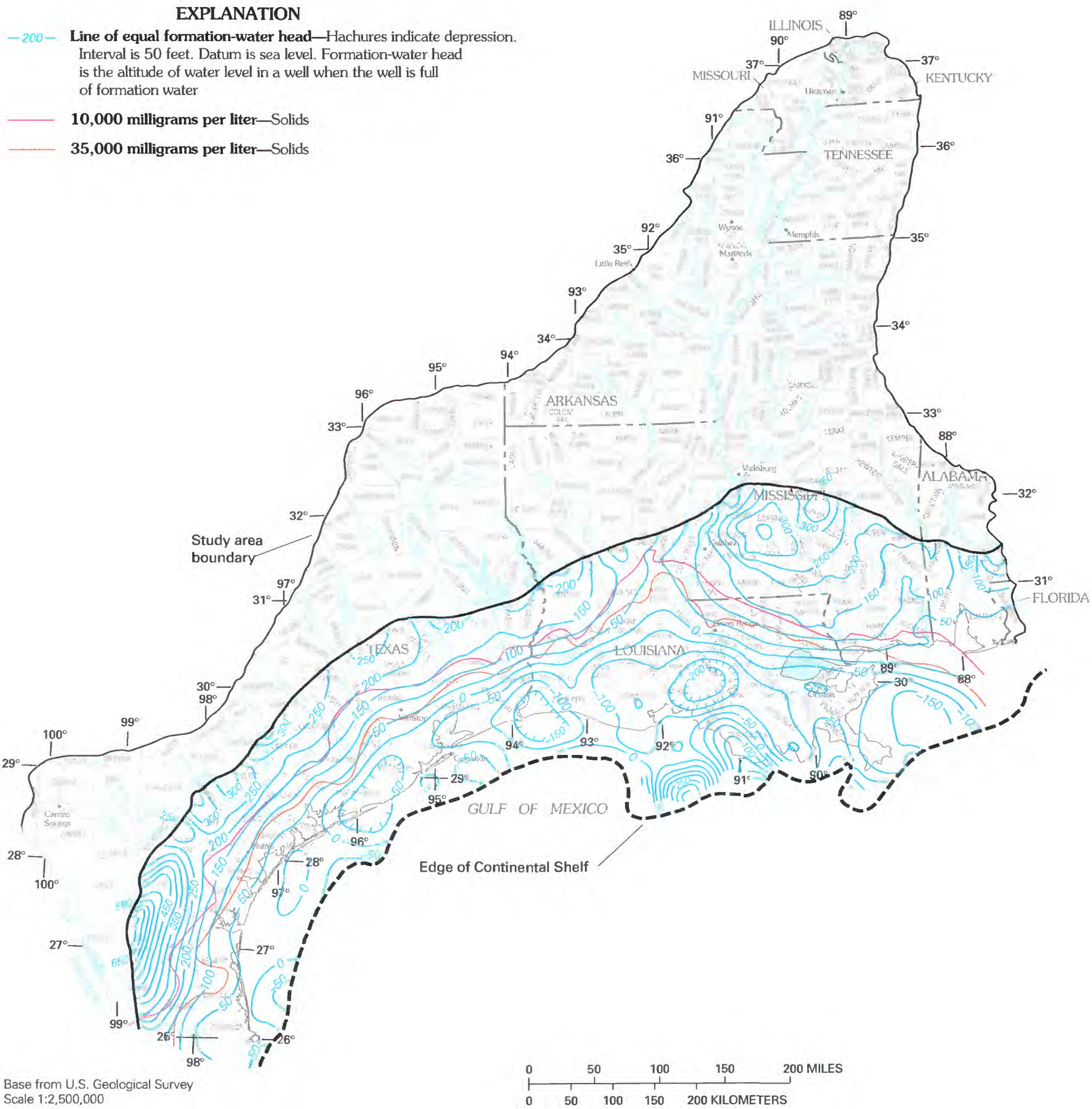


FIGURE 26.—Simulated predevelopment hydraulic head of formation water in layer 8 (permeable zone D of the coastal lowlands aquifer system).

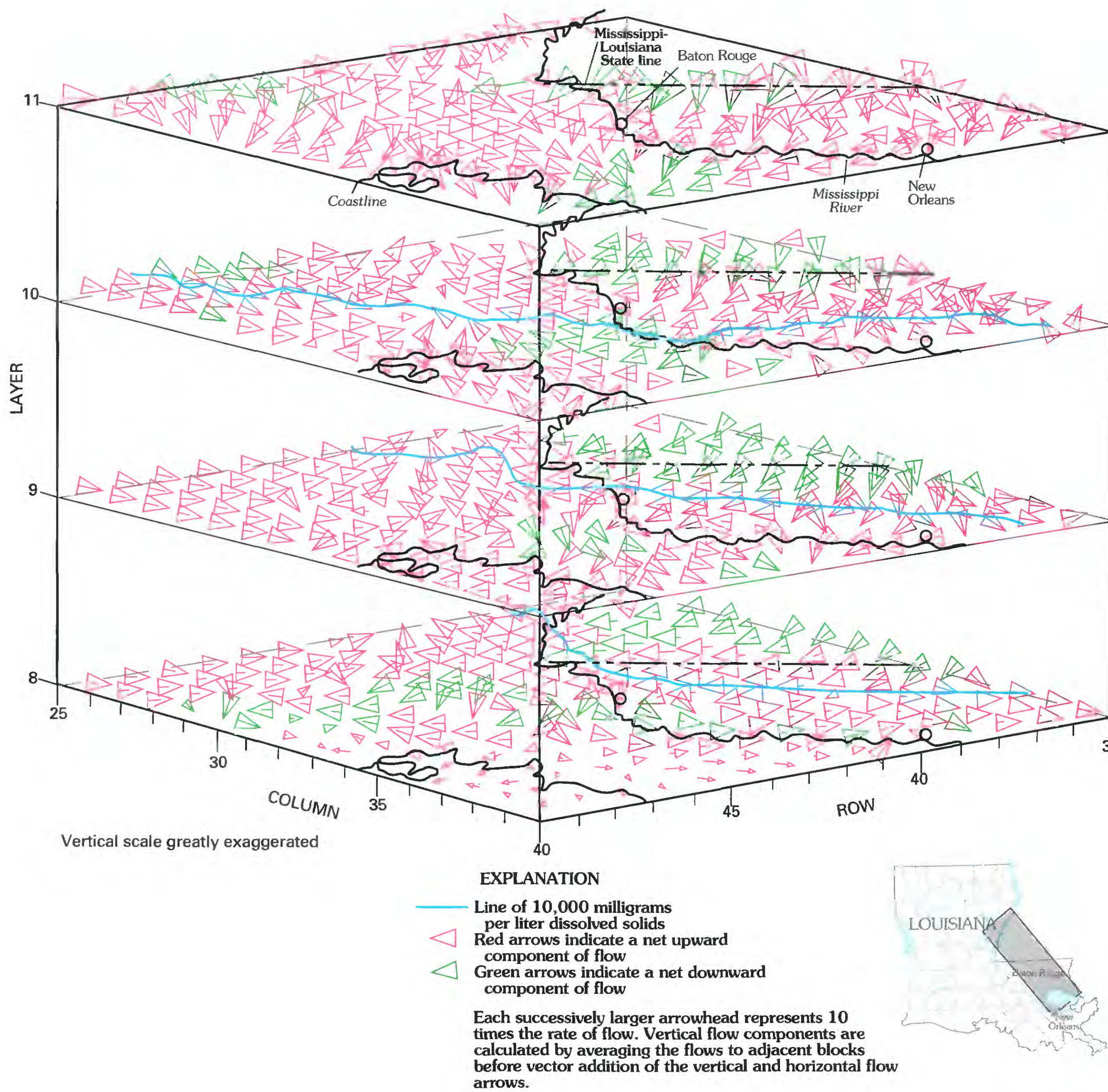


FIGURE 27.—Perspective view of Darcy flow rates and directions in layers 8 through 11 in a 22,500-square-mile area around Baton Rouge, Louisiana.

the relatively slow ground-water velocity. Predevelopment simulations indicate that there is a complex, three-dimensional flow pattern of brine (pl. 3) and that average velocities of about a meter per year occur in some blocks containing brine. The flow pattern in the brine part of the system is largely controlled by the distribution of ground-water density, as evidenced by looking at the results of a large number of simulations with varying parameters. Although density estimates vary substantially from one sand bed to the next and from one well log to the next, there is confidence in the regional trend (fig. 23) in density due to the large number of logs used (fig. 8).

Assumptions about how hydraulic conductivity varies with depth of burial, as a result of compaction, affect the model's sensitivity to density because most dense water occurs at depth. Therefore if it is assumed that conductivity decreases rapidly with depth, all of the flows at depth will appear to be much smaller and hence have less effect on flow in the freshwater part of the aquifer systems. The amount of change of the hydraulic conductivity with depth is uncertain. Williamson and others (1990, p. 107), for simplicity, assumed that hydraulic conductivity did not vary with depth, so the results presented in the preliminary report would indicate a larger value of average regional flow in the brine parts of the aquifer system. The simulations presented in this report assume that the K decreases substantially with depth (see section "Depth Dependence of Hydraulic Conductivity"). Therefore, much smaller flows are shown in the deeper parts of the aquifer system on the vector maps (pls. 3, 6).

Large differences in water chemistry on opposite sides of confining units or fine-grained beds can introduce osmotic gradients (Jones, 1969, p. 63, 81–86; Glasstone, 1946, p. 651; Young and Low, 1965). 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 may induce substantial flow. 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, osmotic effects have been ignored in this study.

FORMATION OF BRINE WATERS

There has been much speculation about the source of salt and mechanisms causing widespread occurrence of the brine waters found downdip in all waters of the gulf coast regional aquifer systems. Jones (1969, p. 59) accepted the theory of widespread diagenesis (De Sitter, 1947, p. 2040) of formation waters whose original source was connate seawater (Clayton and others, 1966, p. 3873). Bredehoeft and others (1963) discussed the possibility of clay beds acting as membrane filters to concentrate brines on the side that flow is coming from. Hanor (1987) presented geochemical and physical evidence for the existence of density inversions in Gulf Coastal Plain sediments sufficient to drive large-scale convective fluid flow. The inversions are caused in part by the dissolution of salt domes 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) suggested that these effects have regional significance in Gulf Coastal Plain sediments. Simulations done for the Gulf Coast RASA indicate that this is a possible mechanism to distribute brines throughout the gulf coast regional aquifer systems, even though salt domes occur only in parts of the systems.

The possibility of dissolution of salt domes as the primary source of brine in the gulf coast regional aquifer systems was explored using location (fig. 7) and other general information on all recognized salt domes in the Gulf Coastal Plain and adjacent Continental Shelf. The caprock is composed chiefly of anhydrite (Halbouty, 1979, p. 45), whereas less than 3 percent of the salt dome content is anhydrite (Kupfer, 1963). Assuming that the existing caprock is whatever remains from previous dissolution of salt (Goldman, 1952), which is the most common view (Halbouty, 1979, p. 45), an estimate of the volume of salt that has dissolved to date can be made. Beckman and Williamson (1990) compiled the information from eight published sources, recognizing 624 salt domes. Data on the depth to the top of salt were available for 349 of these domes (fig. 28). Data on the depth to the top of the caprock were available for 185 of these domes. Estimated average diameter was available for most of the domes whose depth was known. For others, the average diameter of 1.25 mi was used. The total volume of caprock in the 185 domes with sufficient information is 15.5 mi^3. Dividing the total volume of anhydrite by 3 percent yields about 500 mi^3 of dissolved salt. If that volume were prorated to estimate the total, including domes whose caprock depth is unknown (624/185), the volume would be about 1,700 mi^3.

Assuming most of the sediments containing brines today were originally deposited in marine environments and thus contained seawater, it is possible to calculate the volume of salt needed to get the distribution of salt concentration that exists today. Using the aquifer geometry and estimated concentration for each model block to calculate that volume of salt, when summed for all of the blocks, yields about 600 mi^3. Admittedly, both

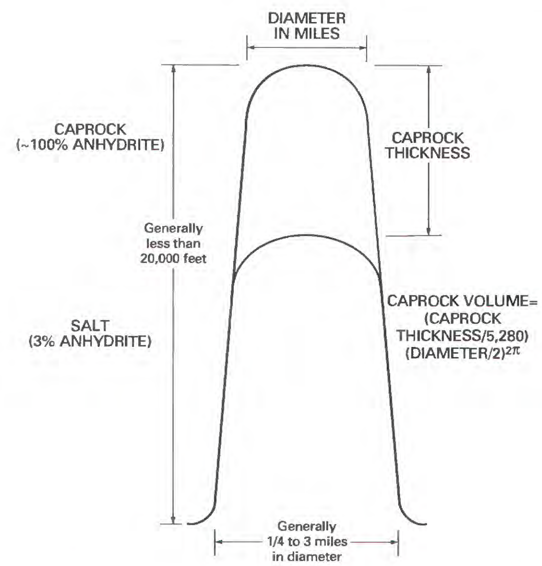


FIGURE 28.—Idealized geometry of salt dome and caprock.

of these calculations are only rough approximations; however, it is very interesting to note that the two volumes are in general agreement and demonstrate the possibility that the major source of salt for the brine ground waters could have been dissolution from salt domes.

More evidence that the dissolution of salt domes is the primary source of salt in the brines is the chemical composition and areal distribution. Pettijohn (in press) has shown, based on discriminant and factor analysis of tens of thousands of water analyses, that the major geochemical process controlling chemistry of the ground water in the brine area is dissolution of halite (salt). With a few exceptions, the distribution of salt domes is closely correlated with the areal distribution of brines with concentrations in excess of 35,000 mg/L dissolved solids (fig. 29, calculated as described in the previous paragraph). The exceptions are in the east Texas and north Louisiana salt basins (Beckman and Williamson, 1990) (fig. 29). In both of these basins, there is very little ground water with dissolved-solids concentrations greater than 35,000 mg/L; however, the sediments are predominantly much older and shallower than in the other salt basins. This anomaly could be explained, at least in part, because much of the brine in these areas could have been diluted and flowed out to the surface or could have moved downdip over the very long time since deposition and subsequent salt-dome penetration and dissolution. Another anomalous condition might be explained by this hypothesis. Many salt domes whose tops are near land surface appear to be entirely surrounded by freshwater. It seems possible that the caprock that develops on top also develops on the sides as a thin veneer. Anhydrite is 30 times less soluble than salt, therefore forming a casing and preventing more salt from dissolving. The diverse nature of salt-dome uplift timing and resultant structure provides for much variation in the thickness of the caprock as well as the side casing of anhydrite.

An additional source of salts that has been given very little consideration is the massive salt formations that occur throughout the entire Continental Slope (Martin, 1980), 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 farther updip and the common, though possibly mistaken, assumption that ground water always flows downdip. The Continental Slope contains a very large source of dissolvable salt and probably is a source of abnormally high temperatures because of the salt. Therefore the Continental Slope is a possible source of salts that have moved long distances coastward and updip, driven by density differences.

LEAKAGE AT BASE OF FLOW SYSTEM

FLOW ACROSS MIDWAY CONFINING UNIT

The base of the flow system for the regional flow model is assumed to be the top of the Midway confining unit or the top of the zone of geopressure, whichever is shallower (fig. 8). 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 was done as part of the analysis of McNairy-Nacatoch aquifer, which directly underlies the Midway confining unit in the northern Mississippi embayment (Brahana, 1987; Brahana and Mesko, 1988). The McNairy-Nacatoch aquifer in this area is composed predominantly of sand and is distinctly different from the equivalent Upper Cretaceous limestone in the southern Mississippi embayment (approximately south of the 35th parallel) and below the Texas coastal uplands aquifer system.

Under predevelopment conditions, there was less than 12 Mgal/d (18.3 ft^3/s) (Brahana and Mesko, written commun., fig. 43) flowing up through the Midway confining unit, which is small relative to horizontal flow in the lower Wilcox aquifer (122 Mgal/d, fig. 33, pl. 3) (Arthur and Taylor, 1990; Arthur and Taylor, 1998). Under 1985 conditions, there was still less than 17 Mgal/d (26 ft^3/s) (Brahana and Mesko, 1988, fig. 64; Brahana and Mesko, 1998, fig. 52) flowing up through

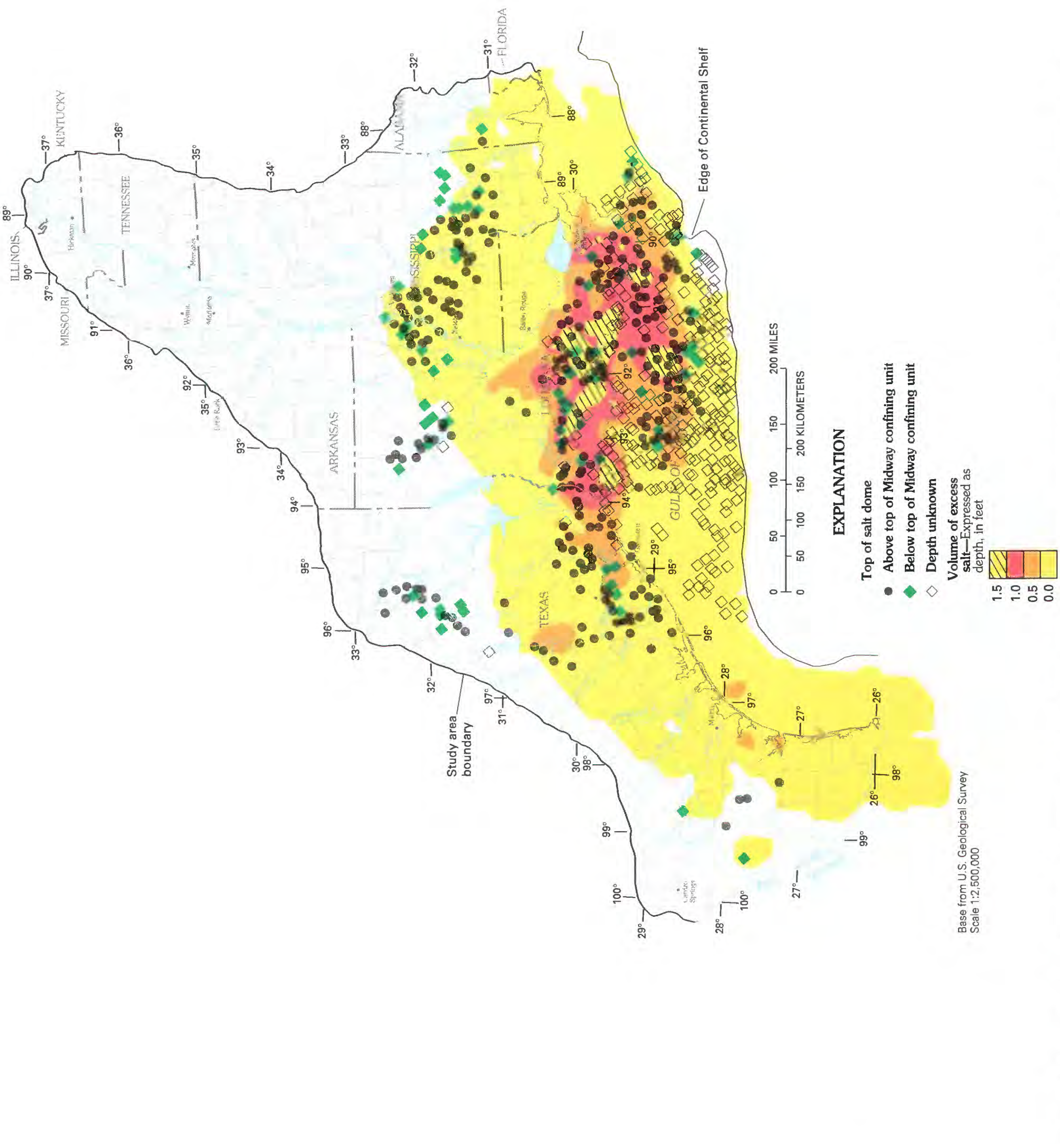


FIGURE 29.—Location of salt domes and distribution of volume of dissolved solids from ground water with concentrations greater than 35,000 milligrams per liter.

the Midway confining unit and any fracture zones penetrating it, which is small relative to the horizontal flow in the lower Wilcox aquifer (173 Mgal/d, fig. 33, pl. 6) (Arthur and Taylor, 1990; Arthur and Taylor, in press). Therefore, the Midway confining unit has been treated as a no-flow boundary in the simulations presented in this report. If larger volumes of pumpage were simulated in the lower Wilcox aquifer, this no-flow assumption would need modification.

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-Nacatoch aquifer and Paleozoic rocks of the Central Midwest aquifer system. According to Brahana and Mesko, pinching out of the Midway and occurrence of the McNairy-Nacatoch aquifer directly above the Paleozoic rocks allow substantial flow between the Paleozoic rocks and the Mississippi embayment aquifer system (Brahana and Mesko, written commun.). The flow is predominantly from aquifers in Paleozoic rocks discharging to the Mississippi River Valley alluvial aquifer, which in turn discharges as rather large flows to streams. Nearly all of these flows occur within the same regional model block and therefore are not considered regional flow.

LEAKAGE FROM THE GEOPRESSURED ZONE AND EFFECT OF DEEP FLUID INJECTION

Downdip and gulfward in the gulf coast regional aquifer systems, the top of the zone of geopressure is above the top of the Midway confining unit (fig. 8). Geopressured zones are defined as areas where the pressure substantially exceeds the hydrostatic pressures caused by the regional ground-water flow system. The high pressures were created by other forces. Geopressured zones exist in the transition from a predominantly coarse-grained facies to a predominantly fine-grained facies. Substantial 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 minimal vertical hydraulic conductivity of the predominantly shaly sediments, which is confirmed by the fact that the high pressures still exist. Hypotheses about the values of conductivities and flows were tested by Kuiper (1994) using various simulations. He concluded that the sparsity of reliable deep-head data precluded the possibility of determining the volume of flow from the geopressured zone on the basis of simulations. Most of his model simulations assumed zero flow from the geopressured zone as does this report.

Oil-field brines and other hazardous wastes are injected into the brine section of the aquifer system, and oil and brine are pumped out of the same section. This study generally assumed that the volumes are not large enough to substantially affect regional ground-water flow or quality of water in the freshwater part of the system. Using data on oil production by the Louisiana Department of Natural Resources (1986), a simple comparison of pumpage totals for Louisiana indicates that this pumpage is less than 10 percent of the volume of the freshwater pumpage for the same period of time. Crude oil production for the State of Louisiana in 1985 was about 185 million barrels (28 Mgal/d) (Louisiana Department of Natural Resources, 1986, p. 146), including oil and condensate but not including gas volumes. Ground-water pumpage for the same year for all of Louisiana was about 1,800 Mgal/d (Mesko and others, 1990). Assuming a 3:1 ratio of brine to oil withdrawn, which varies widely but is thought to be an average, the total withdrawal of fluids would have been about 110 Mgal/d, or about 6 percent of the volume of the freshwater pumpage for Louisiana for 1985. This would still be a small percentage of the total withdrawals, even after increasing this volume somewhat to account for gas production.

PREDEVELOPMENT GROUND-WATER FLOW

The dating of "predevelopment" conditions is complicated by the fact that ground-water development in a few areas, such as at Memphis, Tennessee, dates back to decades before 1900. In most of the study area, however, ground-water development did not occur until decades after 1900, with large-scale developments not active until the 1930's or 1940's. Predevelopment conditions in this report generally refer to conditions in 1925 or before.

HYDRAULIC HEAD AND HORIZONTAL FLOW

Simulated predevelopment freshwater hydraulic head for each aquifer is shown on plate 2, and the regional ground-water flow directions and relative magnitudes are shown on plate 3. 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 arrowshaft length as well). The arrowhead size signifies the order of magnitude [(integer component of the logarithm, base 10 ($\log_{10}$)], of the flow, and the shaft length is proportional to the mantissa (decimal component) of the $\log_{10}$ flow. The type and color of the round symbols on the flow maps show 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.

Hydraulic-head data were sparse outside the outcrop of the individual aquifers for predevelopment conditions. The model was primarily calibrated using 1960–85 hydraulic-head data. Predevelopment conditions were simulated by removing ground-water withdrawal. Flow patterns presented in this report are more complex than those presented in the preliminary report (Williamson and others, 1990) because of the addition to the simulations of variations of hydraulic conductivity (K) by area sand percentage and depth. In the earlier report, for simplicity, K was assumed to be constant for each aquifer. The simulation results should be used and interpreted with caution, especially on the small scale of individual flow values. More confidence can be placed in the large, regional flow trends composed of several to many arrows, which are noticeable on plate 3. The regional flow trends were similar in most of the different calibration attempts. Only such features that are more likely to remain constant through a broad range of simulation parameters will be described in the text.

In the Mississippi embayment aquifer system, the predominant regional flow pattern, especially in layers 2, 4, and 5, is from major recharge areas on the eastern side of the embayment to discharge areas in layer 11, which is mostly in the central and western parts of the embayment (pl. 3). This flow pattern primarily is caused by the topography (fig. 3), which is asymmetrical, in that the valley lies on the western side of the embayment and the topographically higher hills are mostly on the eastern side. The dominant feature of the topography is the flat, low-lying Mississippi Alluvial Plain. The river, because of its size, is an important feature of the hydrologic system. It generally traverses the eastern side of the plain. Some recharge is simulated on the western side of the embayment in the uplands south of the Arkansas River and on Crowleys Ridge, a 1- to 3-mi-wide ridge extending north to south about 100 mi in the north-central part of the Mississippi Alluvial Plain. The horizontal flow rates in layer 3 are substantially larger in the Texas coastal uplands aquifer system than in the Mississippi embayment aquifer system because of the larger hydraulic conductivity (fig. 21).

One of the more noticeable flow paths in the aquifers of the Mississippi embayment aquifer system is from a major recharge area in south-central Mississippi that radiates outward to the south and to the west (pl. 3). A large part of this flow moves southwestward and then curves northwestward to the major discharge area in northeastern Louisiana. This curvature in flow paths is partly because of the restriction of vertical flow downdip. More importantly, what might have been a discharge area for the Mississippi embayment aquifers in southern and western Mississippi is really a regional recharge area for the coastal lowlands aquifers due to high land-surface (and water-table) altitude, reversing the flow direction to downward. This gradient forces the flow in the lower units to turn westward to discharge in the lowland area of northeastern Louisiana. In layer 6, the fresh ground water flows downdip under the Vicksburg-Jackson confining unit and turns 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. Throughout most of the areal extent of layer 6 there is upward flow from underlying aquifers.

Another major predevelopment flow pattern involves recharge in the outcrop of each aquifer in south-central Arkansas, downward flow into layers 6, 4, and especially 5, and then southeastward flow across Arkansas, towards the regional discharge area in northeastern Louisiana (pl. 3). In the northern part of the Mississippi embayment aquifer system, the dominant flow pattern is from east to west toward the flow axis of the embayment, which is west of both the geological axis and the Mississippi River (fig. 7). In the Mississippi River Valley alluvial aquifer, the flow directions are quite complex because the water moves in the direction of the shortest circuit for flow to one of the major surface rivers or drains. The lowest freshwater head in nearly every Mississippi embayment aquifer occurs in northeastern Louisiana (pl. 2).

One of the interesting smaller features of predevelopment flow is in southern Arkansas and northern Louisiana in layers 3 and 2 (pls. 3*I*, 3*J*). Notice that the flow lines converge approximately toward the axis of the Ouachita River and that the line of equal concentration of dissolved solids takes a downdip bend at the same location, even though layers 3 or 2 do not outcrop in this area. Apparently, the water-table altitude decreases due to the Ouachita River Valley increasing leakage from layer 2 to the surface and drawing fresher water farther downdip in the section at this location. A similar flow pattern is seen along the Red River (pls. 3*I*, 3*J*).

In the Texas coastal uplands aquifer system, the predominant regional flow pattern is from recharge areas between major river valleys laterally to discharge areas along the valleys (pl. 3). In southern and south-central Texas, there is a pronounced regional flow pattern. Flow is lateral (parallel to the strike of the outcrops of the units) from the southwest toward the northeast, as indicated on both the head (pl. 2) and flow (pl. 3) maps, especially in layers 3, 4, and 5. The Vicksburg-Jackson confining unit covers a larger percentage of the Texas coastal uplands aquifer system than it does of the Mississippi embayment aquifer system. The Vicksburg-Jackson confining unit has a substantial effect on the flow pattern in addition to the effect of topography. It impedes upward flow out of the downdip parts of the Texas coastal uplands aquifer system, effectively restricting longer flow paths in the downdip direction. This is demonstrated by noting that freshwater in layer 6 becomes saline close to the outcrop except for the area in southwestern Mississippi, as described previously (pl. 3*F*). In the Texas coastal uplands aquifer system, several major flow paths are toward the Nueces, Brazos, Neches, and Sabine Rivers and the Rio Grande.

In the coastal lowlands aquifer system, the predominant flow pattern is from recharge areas in the updip, topographically higher areas along the inland edge of the system to discharge areas onshore near the coast. The horizontal flows radiate outward from the higher topography recharge areas in southwestern Mississippi, eastern Texas, west-central Louisiana, and southern Texas (pl. 3). In layers 7 through 10, the widest band of freshwater occurs where the radial flow pattern in southwestern Mississippi shows the effects of the largest area of regional recharge in the hills. Generally, the downdip flow pattern continues across the interface of saltier water, though the flow is diminished because at this point the aquifer is quite deep and the hydraulic conductivity has diminished.

The minimum freshwater hydraulic head in layer 11 in the coastal lowlands aquifer system never reaches sea level (minimum about 3 ft) before it starts increasing offshore due to density differences (pl. 2). The flow map for layer 11 (pl. 3*A*) shows flow perpendicular to the head contours (pl. 2*A*) and toward the coast (lower altitude) onshore as would be expected in a freshwater system. Near the coastline in several places, the flow direction reverses. Offshore, the flow direction is quite varied, with several areas of converging flows where the water moves downward into the lower units.

Flow patterns in the deeper coastal lowlands aquifer system layers are more difficult to interpret because of the large variability due to the complex three-dimensional distribution of ground-water density. Layer 10 shows a pattern of hydraulic head and flow similar to layer 11. There is a striking uniformity of directions of flow in two freshwater parts of layer 10, toward the southeast in the area west and south of Houston, Texas, and toward the east in southern Texas. Unlike layer 11, but more like the lower layers, the freshwater head in layer 10 rises substantially offshore due to the effects of density. Some circulation cells appear in that part of the layer containing brine water. It is interesting to note that off the coast of south-central Louisiana, where layer 11 shows convergent horizontal flow and downward flow out the bottom, layer 10 shows divergent flow paths and downward flow out the bottom.

Regional ground-water flow in layer 9 generally moves downdip from the outcrop except in the discharge area adjacent to the Mississippi River. The variability of flow far downdip from the outcrop area becomes more pronounced in layer 9 because it is deeper and more saline than overlying layers and therefore, more affected by the density driving force. Circulation cells that appear on plate 3C can occur solely due to the density forces resulting from heat and solute gradients (L.K. Kuiper, oral commun., 1985) as the denser, heavier water overturns to flow under the fresher, lighter water. These circulation cells can continue indefinitely, somewhat analogous to convection currents in the atmosphere if a continuous driving force is present in the ground-water flow system such as a heat source or salt dissolution. The pattern of hydraulic head in layers 7, 8, and 9 approximately follows what has already been described for layers 10 and 11. The flow maps are more complex due to the high density of the water and the substantial variability in the density of the water spatially. The part of the aquifers containing freshwater is fairly narrow compared to the part containing brine.

RECHARGE AND DISCHARGE BUDGET

Ground-water flow can be considered at various scales, from small valleys with recharge on a hill and discharge to a nearby creek in the valley, to regional flow patterns across counties or States (fig. 30). This investigation is concerned with regional flow that generally involves flow paths of tens of miles or more. The small-scale (or local) flow paths, which generally have higher rates because the flow paths are shorter (a few miles or less) and the gradients are steeper, will not be shown or discussed in this report. They were not simulated because the short flow paths do not cross a model block face (fig. 30). Only the net flow across a model block face is considered in these simulations. In addition, there is some canceling of effects of recharge and discharge within a model block. Local recharge would cancel out an equal amount of local discharge (fig. 30) and therefore would not be simulated in these regional models. These factors should be considered when comparing water budgets and rates of regional recharge and discharge from this report to detailed studies of smaller

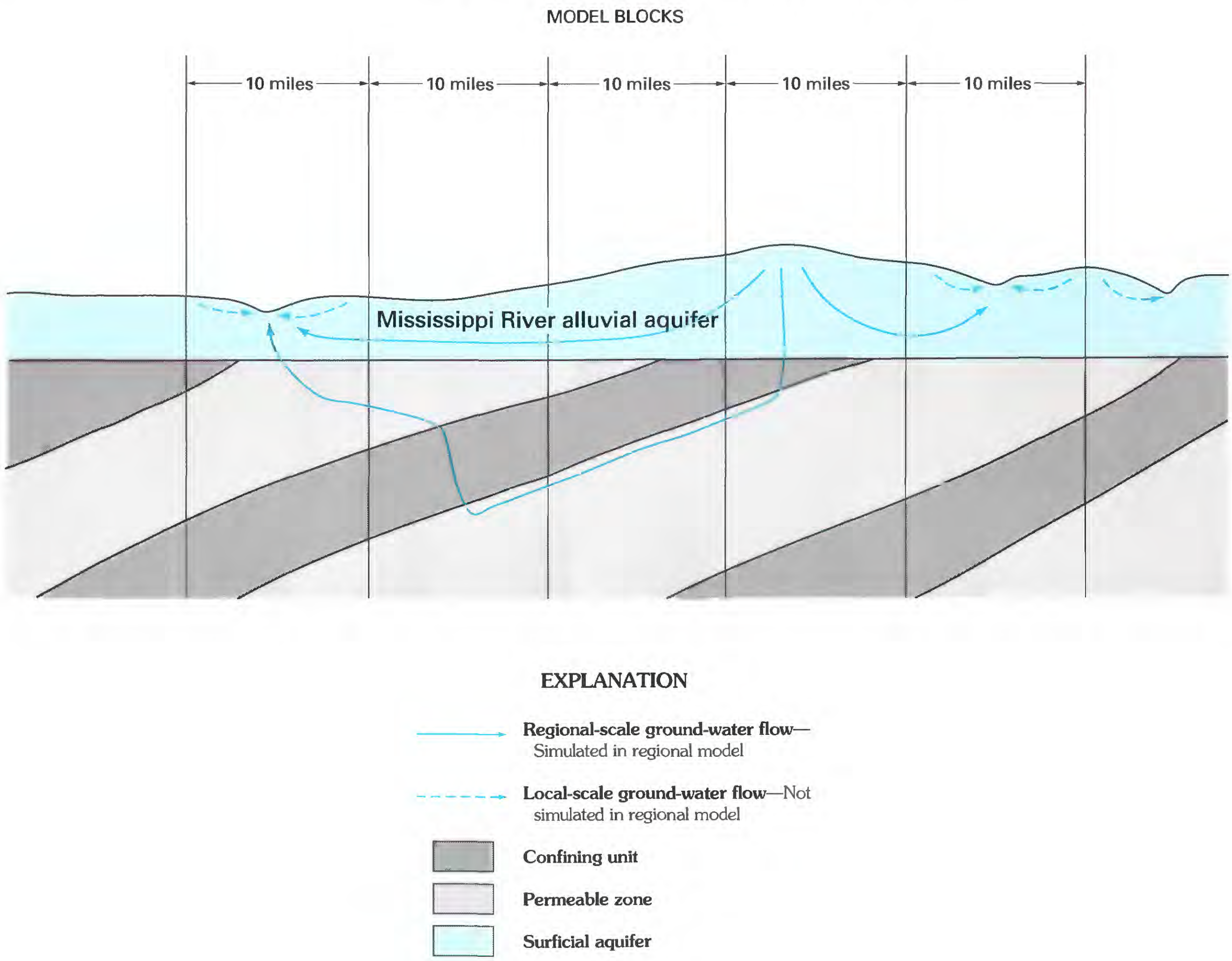


Figure 30.—Generalized local- and regional-scale ground-water flows.

areas that probably show higher rates of recharge and discharge.

Simulation indicates that the total predevelopment regional recharge is about 2.9 Ggal/d. This equals regional net discharge assuming steady-state conditions. Although some effects of long-term changes like sea-level change may exist, the effect of these transient conditions on the regional budget is probably minimal because of the size of the study area. Regional recharge equals only 0.6 percent of the total precipitation (about 500 Ggal/d, fig. 4) in the study area. Most of the precipitation returns to the atmosphere in a short time as evapotranspiration (ET) because much of the precipitation occurs during the hot months when ET rates are highest (fig. 5). Most of the remaining precipitation becomes runoff (about 140 Ggal/d) (fig. 6) either directly or by recharging small, local ground-water flow systems that discharge to creeks and streams nearby. The net regional recharge amounts to only about 0.5 percent of the total surface-water flow (523 Ggal/d) from the many large streams that cross the study area (Williamson and others, 1990, table 1, p. 97). Regional recharge cannot reliably be estimated by stream budgets because of the contrast in sizes of the flows.

Predevelopment regional ground-water budgets (fig. 31) by geographic area (fig. 13) show that the Mississippi Alluvial Plain has the largest total flow. In the Mississippi Alluvial Plain, southwestern Louisiana, and to some extent the Winter Garden area, inflow from adjacent geographic areas is a substantial source of water. In the other areas, net regional recharge is the dominant source of water. In several geographic areas including the eastern embayment, western embayment, eastern coastal lowlands, and the Winter Garden area (listed in order of proportions), outflow is dominantly to adjacent areas rather than to net regional discharge.

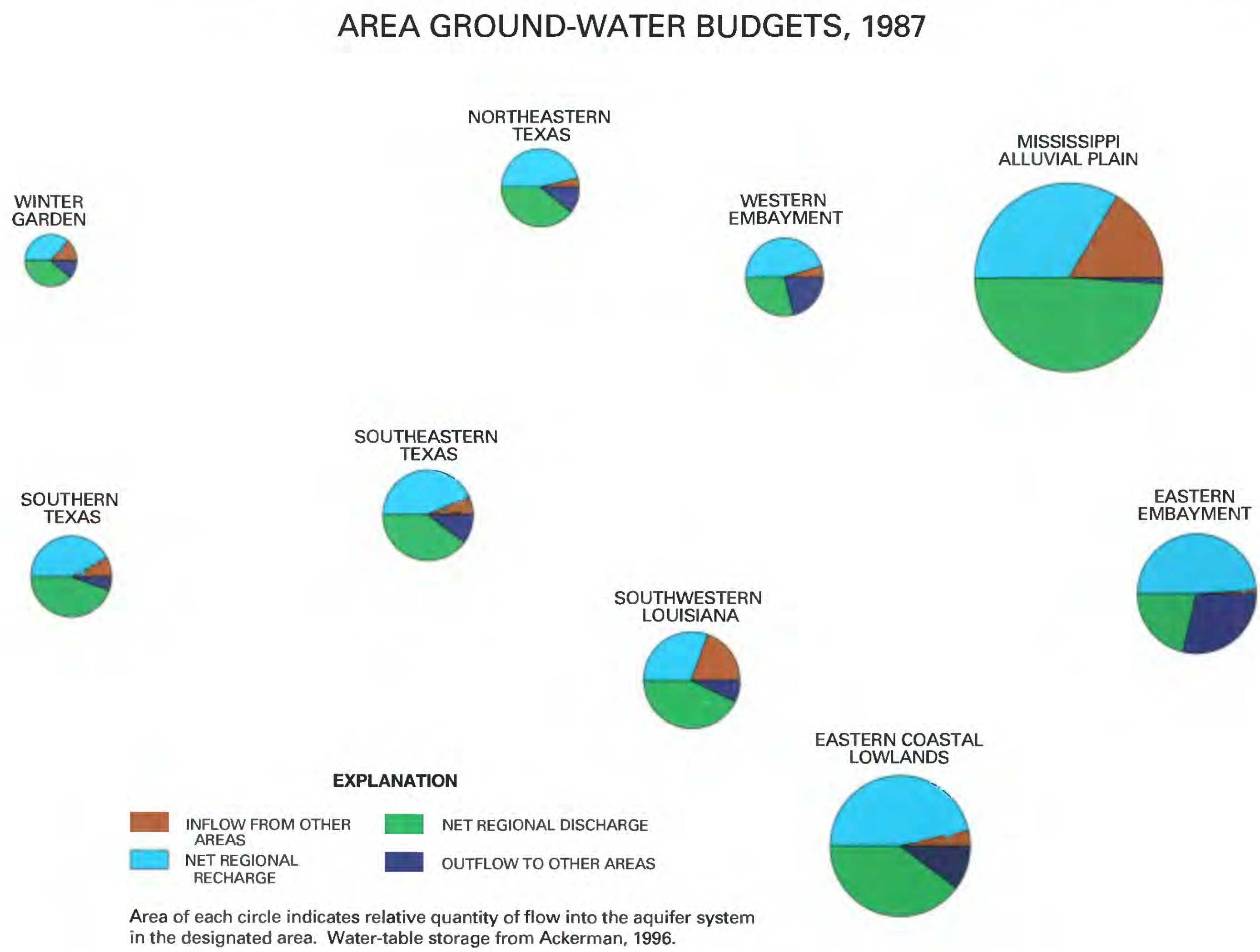


FIGURE 31.—Simulated predevelopment ground-water budgets, by area.

DISTRIBUTION OF RECHARGE AND DISCHARGE

Predevelopment net regional recharge occurred in 41 percent of the modeled area at the average rate of about 0.48 in/yr (fig. 32). The largest rate was in southwestern Mississippi at about 6 in/yr. Predevelopment net regional discharge was occurring on 59 percent of the area at the average rate of about 0.35 in/yr (fig. 32). The largest net regional discharge rate was in the Mississippi Alluvial Plain at about 4 in/yr.

The distribution of regional net recharge and discharge (fig. 32) shows a high degree of correlation with the altitude of the predevelopment water table and the land-surface altitude (fig. 3). Areas with the largest regional recharge are in the northern part of the "boot" in eastern Louisiana, an area in south-central Louisiana, and one in southern Texas. The areas with substantial regional recharge rates are relatively large, topographically high areas adjacent to low areas where the regional flow paths are generally 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. The largest regional recharge area in the study occurs in southwestern Mississippi and adjacent Louisiana where the relatively high topography is a major driving force, inducing regional recharge that flows downward to

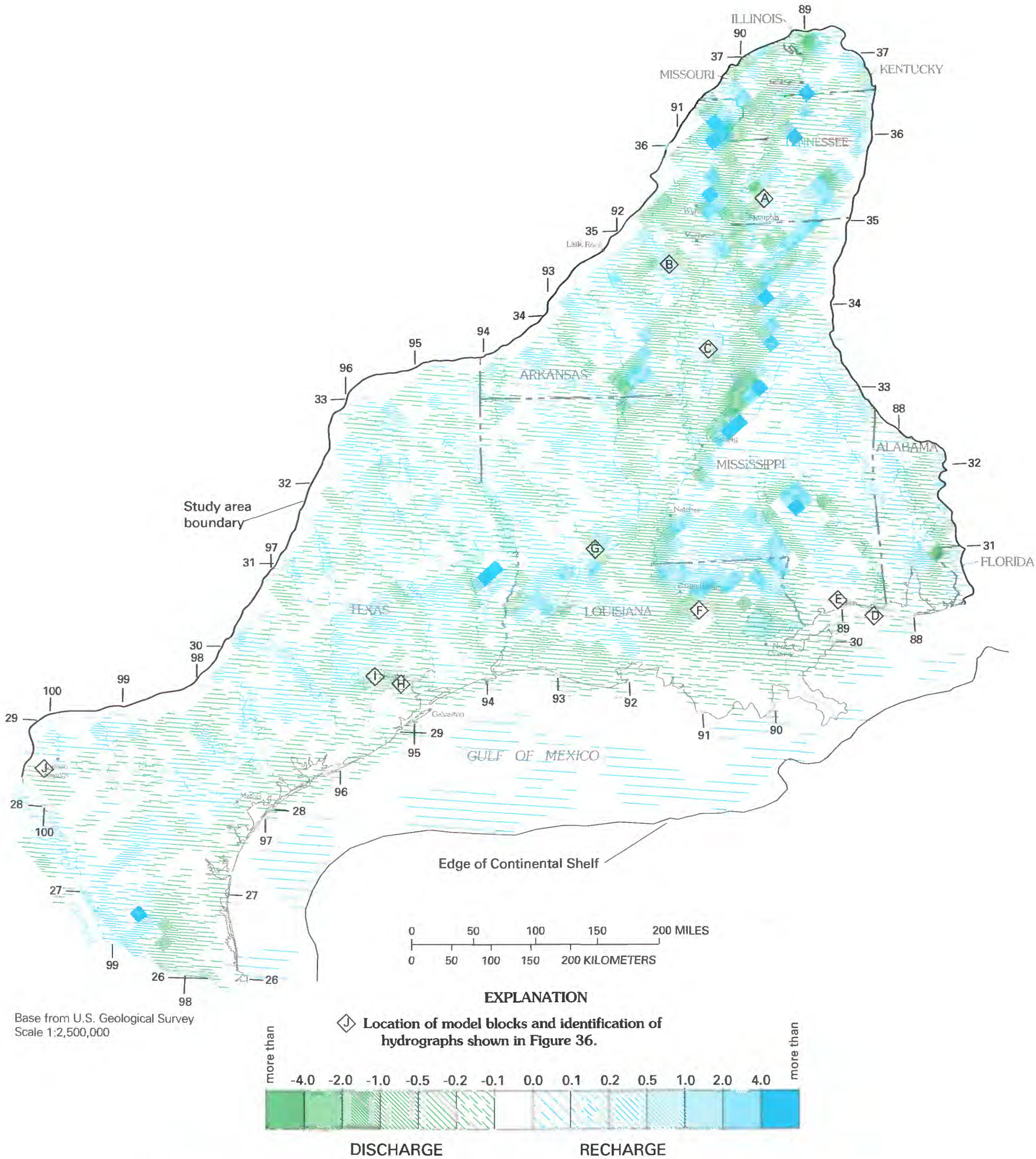


FIGURE 32.—Simulated predevelopment net regional ground-water recharge and discharge, and location of hydrographs.

underlying layers and radiates horizontally in nearly all directions (pl. 3). The large area of regional recharge in southwestern Mississippi and adjacent Louisiana generally is coincident with the greatest thickness of freshwater mapped in Mississippi by Gandl (1982) and in Louisiana by Winslow and others (1968).

The location of the Vicksburg-Jackson confining unit (layer 15 in fig. 3) has a substantial effect on the distribution of recharge and discharge. Areas updip from the outcrop of the unit are regional discharge zones, probably due to two factors. As described previously, upward vertical discharge from the underlying aquifers is impeded by the Vicksburg-Jackson confining unit, forcing more discharge immediately updip from its outcrop. Also, in some places, particularly eastern and southern Texas, the Vicksburg-Jackson confining unit outcrop coincides with a band of higher topography; therefore, these areas are regional recharge areas.

There are many large structural features in the study area (fig. 7), such as the Jackson dome, the Desha Basin, and the San Marcos Arch, which affect the regional geology. The effect of these structural features on the predevelopment regional ground-water flow is not as significant as the configuration of the water table and the permeability of the geohydrologic units (Grubb, 1986, p. 18). This is also indicated by comparing figure 32 with figure 7.

The amount of recharge to the regional ground-water flow system is also probably not related to the mean annual precipitation (fig. 4) over most of the study area because of the relatively humid climate. If regional ground-water recharge is controlled by the amount of precipitation, the depth to the water table should be related to precipitation. Williamson and others (1990, p. 97–98) tested this possibility by several types of regression analyses that indicated no relationship between the measured depth to water and mean annual precipitation based on measurements of water levels made before 1960 in about 7,000 shallow wells.

It is more likely that the amount of precipitation that recharges the regional flow system is limited by the capacity of the flow system to transmit water from recharge areas to discharge areas. Where the regional ground-water flow capacity is limited by flat regional hydraulic gradients or by the resistance to flow in the aquifer system, most of the water that infiltrates from precipitation as potential regional recharge is 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 may be limited by the lesser precipitation, much of which falls in the hot summer months and is quickly evaporated or transpired.

Some ground water flows offshore before being discharged to the ocean, but most is discharged before it reaches the coastline (fig. 32). Maximum discharge occurs in an inland band parallel to the coastline. In southern Mississippi, higher topography exists closer to the coast than in Texas. The higher topography produces steeper horizontal gradients in the ground-water system, steeper upward vertical gradients offshore (pl. 1*D*), and more fresh ground-water discharge from offshore Mississippi than from offshore Texas (pl. 1*A*). Fresh ground water extends beneath some islands offshore from southern Mississippi (Sumner and others, 1987, p. 7, figs. 27–28); but to the west, along the Louisiana and Texas coastline, the only freshwater offshore is a relatively thin lens that is recharged and discharged locally (see layer 10 in fig. 12*B*). Deep, fresh ground water does not extend offshore to the east, but probably for a different reason. The aquifers that are onshore decrease in hydraulic conductivity to the east and offshore. On Dauphin Island, offshore from Mobile, Alabama, the only fresh ground water found was a relatively thin lens of freshwater in a water-table aquifer recharged on the island (Kidd, 1987).

VERTICAL FLOW BETWEEN LAYERS

Although vertical hydraulic conductivity is commonly several orders of magnitude smaller than horizontal hydraulic conductivity, the geometry of the aquifer system where the width greatly exceeds the thickness of each unit causes a large area for vertical flow and a large head gradient because vertical flow paths are relatively short. These factors cause the vertical flow to be about the same magnitude as the horizontal flow despite the large difference in conductivity. The upward flows between layers were generally larger than the downward flows (fig. 33) under predevelopment conditions. The individual flows across block faces in the model represent net flows across that particular face of the block. The net flow is the difference between positive and negative flow across the individual block face. The flows in figure 33 were obtained by adding separately the negative and positive net flows across all of the model block faces between adjacent layers. The net flows between layers (not shown in fig. 33 due to lack of room, but easily calculated) equal the difference between the upward and the downward flows.

In the coastal lowlands aquifer system, all of the net flows between layers are upward, the largest between layers 10 and 11. The flows become progressively smaller down through the layers, 9–10, 8–9, 7–8 (fig. 33). Recharge in the topographically higher outcrop areas flows downdip and discharges upward through the

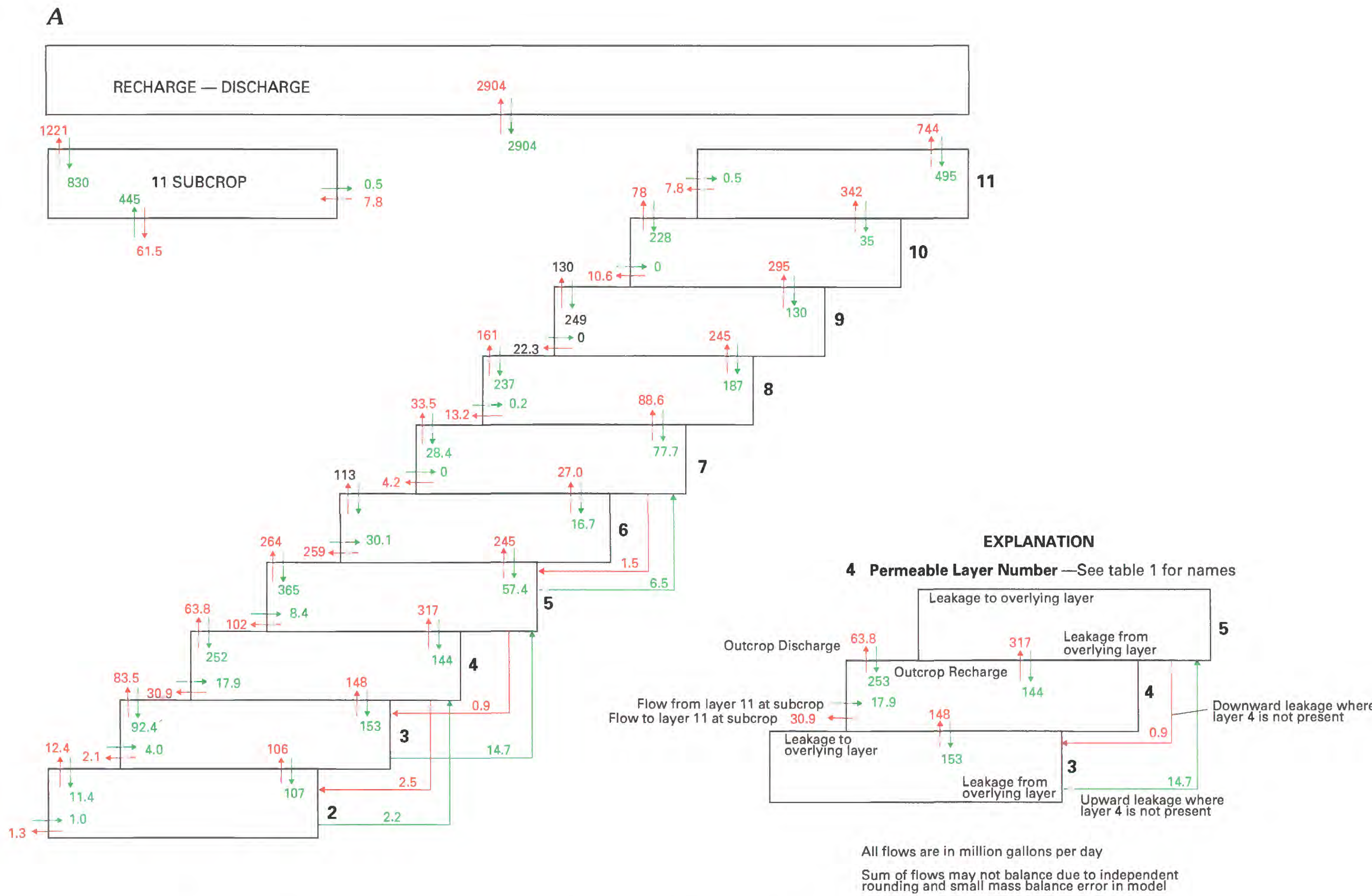


FIGURE 33.—Simulated vertical flow between layers and recharge and discharge for: *A*. predevelopment conditions; and *B*. 1987 conditions.

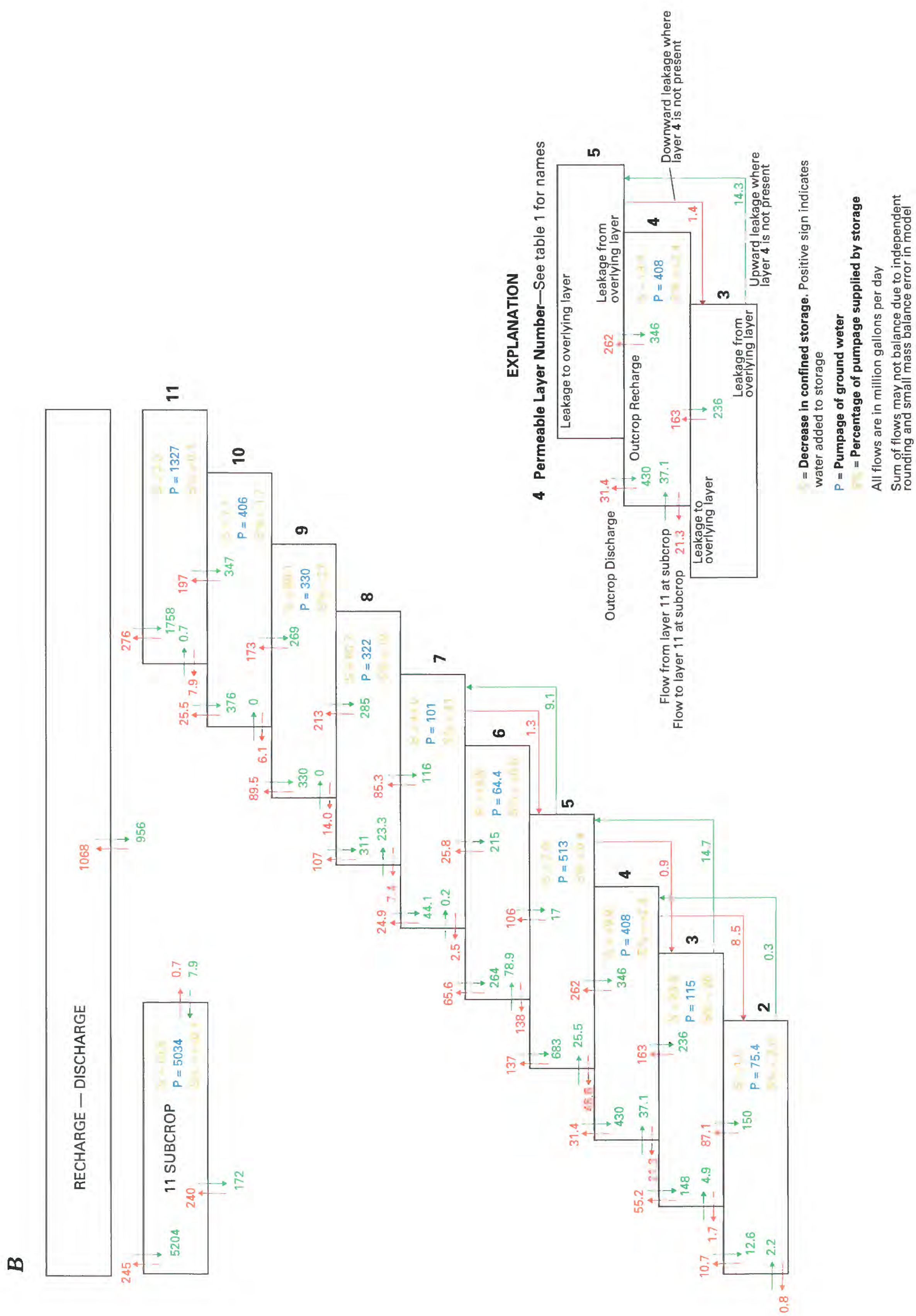


FIGURE 33.—Simulated vertical flow between layers and recharge and discharge for: *A*, predevelopment conditions; and *B*, 1987 conditions—Continued.

overlying layers at lower altitudes. Layers 8, 9, and 10 all have similar amounts of recharge in the outcrop areas. Layer 7 has about one-tenth as much recharge as the other three layers, probably because of its much smaller outcrop area and the increased resistance to vertical flow caused by the overlying layers. Potential regional recharge is discharged locally owing to the limited capacity of layer 7 for regional flow. Layer 11 has about twice as much recharge because of its wide outcrop and easy discharge path.

In the Texas coastal uplands aquifer system, the general flow pattern from updip highlands to downdip lowlands is affected by many large stream valleys. Fresh ground water occurs only in narrow bands near the outcrop of each permeable zone. Several examples of this occur in layer 2, the middle Claiborne aquifer, in eastern Texas (pl. 3*G*). 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 occurs in the Mississippi embayment aquifer system in layer 6 in southwestern Mississippi and adjacent parts of Louisiana (pl. 3*F*).

Less than 30 Mgal/d of leakage occurs upward through the Vicksburg-Jackson confining unit (layer 15) from the Mississippi embayment aquifer system and Texas coastal uplands aquifer system into the coastal lowlands system (between layers 5 and 7) under predevelopment conditions (fig. 33*A*). The small rate of flow is due mostly to the large thickness and small vertical permeability of the confining unit. Some vertical flow does occur between aquifer systems across the Vicksburg-Jackson confining unit; thus, the unit is not an impermeable barrier to flow as might be suggested by considering only its thickness and lithology. The head gradient in layer 6 trends downdip from the outcrop, underneath the overlying confining unit (pl. 3), indicating flow away from the outcrop that discharges up through the confining unit. A similar situation also exists below the Midway confining unit at the base of the aquifer system. However, the effect of this upward leakage on the flow system above the Midway confining unit is much less than on the flow system below. If not for this upward leakage, there would be very little or no circulation of ground water below the confining unit. Without upward leakage, the flow patterns in the upper system would change little because of other sources of flow that usually are larger. The Vicksburg-Jackson confining unit operates similarly to the regional confining units above the Dakota Sandstone discussed by Bredehoeft and others (1983) in that they do conduct significant quantities of water vertically even though their hydraulic conductivities are low.

The Mississippi River Valley alluvial aquifer (part of layer 11) receives a substantial amount of flow (more than 400 Mgal/d) from the older units of the Mississippi embayment aquifer system (fig. 33*A*). In the Texas coastal uplands and Mississippi embayment aquifer systems, only layers 4 and 5 have substantial net flow upward to the layer above. The net flow upward out of layer 6, the upper Claiborne aquifer, occurs mostly where it subcrops beneath and discharges to the Mississippi River Valley alluvial aquifer (layer 11). The flow discharging upward from layer 6 to the subcrop is approximately equal to the flow upward into layer 6 from layer 5 below. Layer 5 has the largest amount of recharge and the largest amount of discharge in its outcrop area; much of the discharge is in outcrop areas of eastern and southern Texas. Layer 4 has the largest net recharge in its outcrop area, probably because it has a large area of outcrop at a higher altitude than the other layers and has a relatively large horizontal hydraulic conductivity. Upward flow is nearly balanced by the downward flow between layers 2, 3, and 4. Layers 2 and 3 have very little net recharge in their respective outcrops.

The vertical flow between layers can be analyzed by looking at hydraulic-head profiles of all aquifers shown on one graph (pl. 1). Where the head decreases with depth (decreasing layer number) and the density is nearly constant, downward vertical flow and regional recharge are indicated, as on the topographic high shown on plate 1*A* in southern Texas. Where the head increases with depth (increasing layer number) and the density is nearly constant, upward vertical flow and regional discharge are indicated, as between the Neches River and Kingsville in southern Texas (pl. 1*A*). At this location, the heads in layers 3 and 4 are more than 200 ft above the land surface. The head in layer 3 nearly equals the head in layer 4 because of a fairly high sand percentage, which indicates a high vertical leakance.

In the coastal lowlands aquifer system, where there is a much greater range of water density, the vertical relations are more difficult to interpret. At the point where the dissolved-solids concentration and hence density increase in a layer, the freshwater head rises substantially (pls. 1*A*, *B*, *C*, *D*).

CHANGES DUE TO GROUND-WATER DEVELOPMENT

Massive changes occurred in the gulf coast regional aquifer systems in response to large-scale development of the ground-water resources throughout much of the study area. Development of ground water has led to lowering of hydraulic heads and changes in aquifer storage, changes in rates and locations of recharge and discharge, changes in flow direction and magnitude, changes in flow velocity, changes in water quality, and land subsidence. Digital simulations were made using estimates of ground-water pumpage available at 5-yr intervals. It was assumed that 1985 pumpage at a steady rate would simulate changes in ground-water levels that occurred between 1982 and 1987; therefore, flow quantities are attributed to 5-yr intervals (for example, 1980, 1985), whereas water-level comparisons are lagged two years (for example, 1982, 1987).

GROUND-WATER PUMPAGE

About 9 Ggal/d of ground water was pumped from the three aquifer systems in 1985 for use by about 19 million people inhabiting the area (Mesko and others, 1990). Total pumpage increased by 250 percent during 1960–80 but decreased by 7 percent by 1985 (fig. 34). This was consistent with a nationwide decline of about 12 percent from 1980 to 1985 (Solley and others, 1988). Irrigated agriculture uses the largest volume (about three-fourths) of ground water in the study area (Grubb, 1984, p. 17) and also represents the fastest growth in ground-water pumpage (fig. 34) during the past three decades. Much of the increase in agricultural pumpage was between 1970 and 1980, especially in the Mississippi River Valley alluvial aquifer. Other areas with large use of ground water for irrigation are southwestern Louisiana and the Winter Garden area of southern Texas. Figure 35 shows the distribution of groundwater pumpage in 1985 and the predominantly pumped layer at each location.

Over 60 percent of the municipal and industrial pumpage is withdrawn from the coastal lowlands aquifer system (Mesko and others, 1990). Ground-water usage for public supply has continued to increase (fig. 34) as the population has grown. Industrial pumpage at many locations has declined since 1970 (fig. 34), mainly in response to large drawdowns, and in some areas because of land subsidence and other adverse effects, as well as socioeconomic factors. About three-fourths of the total pumpage is withdrawn from layer 11 (Mississippi River Valley alluvial aquifer and permeable zone A). The remainder of the pumpage is generally withdrawn from whatever shallower zones exist in an area. The upward trend of pumpage has been unequal across the study area. These trends in pumpage in different areas are discussed in a later section "Potential and Limitations for Future Development" and shown in the area budget-trend bar graphs (figs. 42–50).

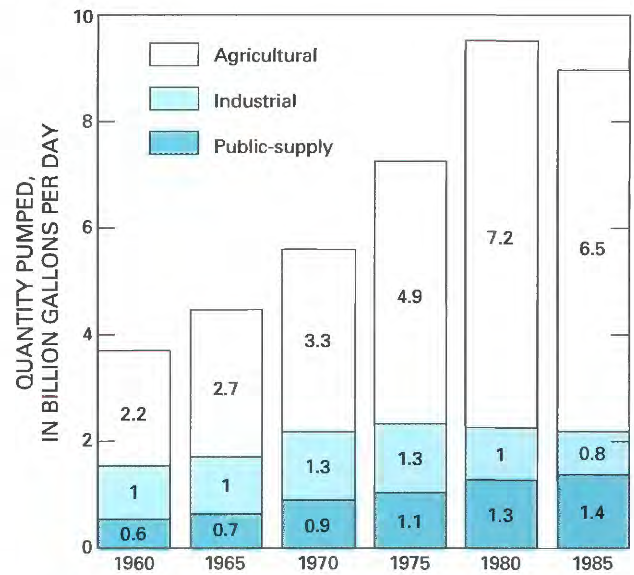


FIGURE 34.—Ground-water pumpage by use, 1960–85.

Although most areas showed a general increase in pumpage throughout the period 1960–85, there were some decreases in pumpage after 1965 in southern Texas (fig. 47).

The pumpage data used in this study were compiled by Mesko and others (1990) for 1960–85 at 5-yr intervals by using a detailed 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, with a few exceptions. Two categories of ground-water pumpage data were chosen:

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

More information about the method of estimating pumpage is discussed in the supplement "Simulation Details" and more detailed pumpage information is available in Williamson and others (1990, p. 65–66) and Mesko and others (1990, p. 25–29).

WATER-LEVEL CHANGES

Ground-water pumpage has caused regional-scale water-level declines in most of the aquifers (fig. 36). Every aquifer or permeable zone has more than one major cone of drawdown and has maximum drawdowns (block-averaged) exceeding 80 ft. Note that

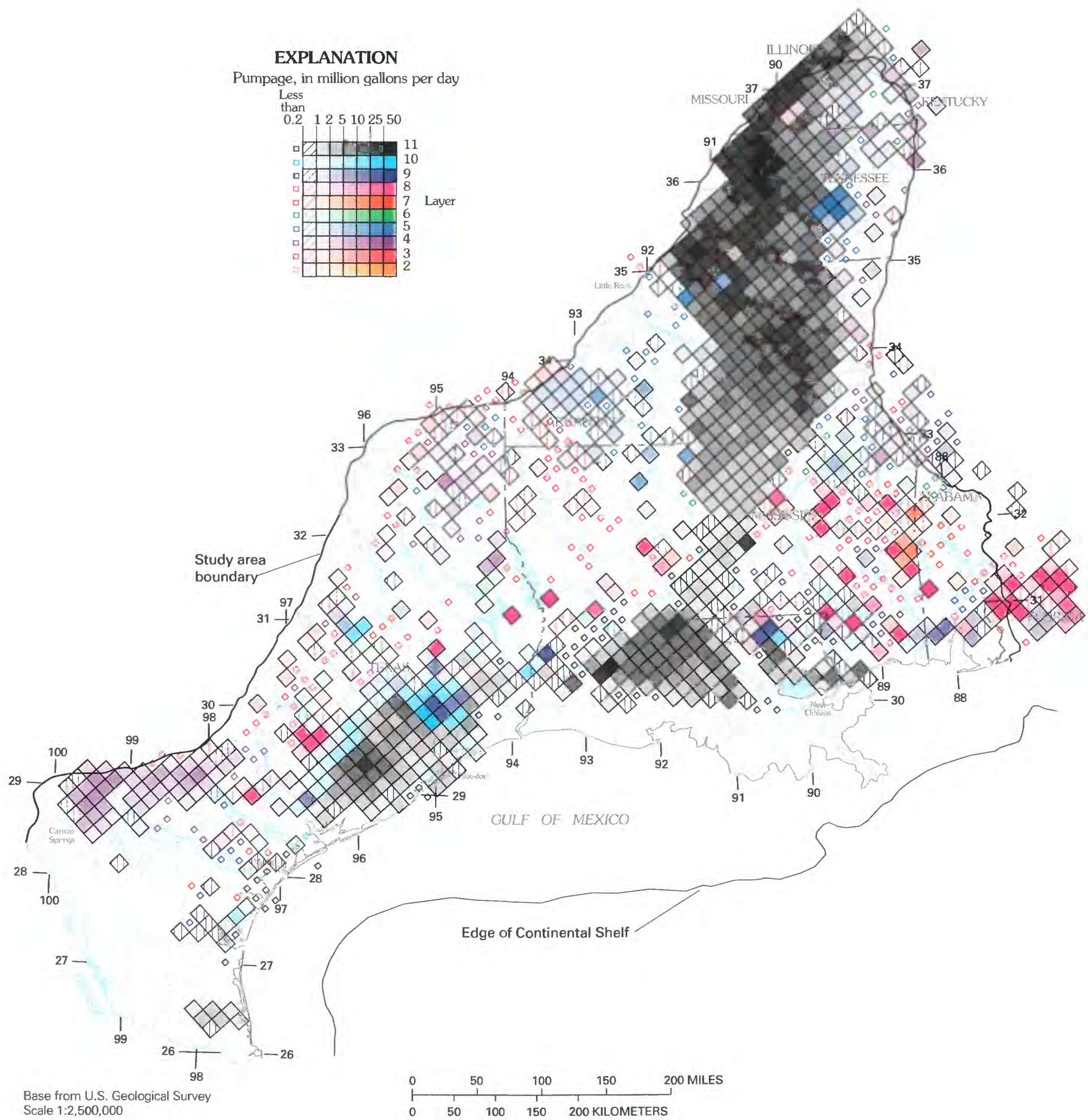


FIGURE 35.—Ground-water pumpage, 1985.

plate 4 shows simulated drawdown from predevelopment to 1987. Total drawdown has not been observed in most places because little hydraulic-head data exist for predevelopment conditions. The standard error of estimate [or root mean-squared error (RMSE)] of the simulated 1987 hydraulic head is about 50 ft, based on the calibration of the model against 1972 and 1980 head estimates (discussed in the supplement "Simulation Details").

Differences in the patterns of drawdown exist between the coastal lowlands aquifer system and the Mississippi embayment aquifer system and the Texas coastal uplands aquifer system (fig. 36). In most of layers 2 to 6, the 1987 drawdown near the updip limit of the aquifer exceeds 5 ft, whereas in layers 7 through 11, the drawdown around the updip limit of the aquifer is less than 5 ft. In southern Texas and adjacent offshore Texas, there are large areas where drawdown is less than 5 ft, and the drawdown more gradually increases away from the updip limit of layers 7 through 11 than in layers 2 through 6. This can be demonstrated more quantitatively by using a cumulative frequency distribution of drawdown relative to area of the layer at that level of drawdown (fig. 37). Layers 3 and 4 have the largest areas of drawdown throughout the whole range of plotted drawdowns. Layer 3 has the largest cumulative volume of drawdown, and it is exceeded in pumpage by all of the layers except 2, 6, and 7 (table 3). This implies that throughout the large areal extent of layer 3 (the largest area—170,000 mi^2, table 1), it is affected by pumping in adjacent aquifers. Layer 5 has the next largest cumulative volume of drawdown and has the second largest volume of pumpage (table 3). Layer 9 is fourth largest in both pumping volume (table 3) and cumulative volume of drawdown (fig. 37). Layers 6 and 7 have the least drawdown, but also have small pumpage. Layer 11 has one of the least cumulative volumes of drawdown (fig. 37) although it has the largest volume of pumpage (table 3). The relatively small cumulative volume of drawdown is due to the shallow depth of the aquifer and hence close proximity to sources of recharge, as well as its large horizontal hydraulic conductivity and the change in water-table storage.

Changes in simulated and observed hydraulic head over time for selected 10-mi blocks having substantial drawdown are demonstrated by the hydrographs in figure 36. The observed block-averaged heads were estimated by averaging for each well, the measurements made during each 2-yr period centered at the 5-yr intervals of plate 5 and then averaging at least two wells in a block in the same layer (discussed in the supplement "Simulation Details"). At most locations, the rate of drawdown increased around 1960 and again during the 1970's because of large increases in pumpage. At most locations, there is drawdown in nearly every permeable zone, due to pumpage, that may be several hundred or thousand feet vertically and several layers removed from the layer pumped (fig. 36).

In 1985, ground-water pumpage and resulting drawdown decreased, leveled off, or the rate of increase declined at most locations with considerable pumpage (fig. 35–36). Water levels recovered substantially to the east of Houston, Texas (fig. 36*H*) after 1977. Several feet of land subsidence had occurred in this area during the 1960's and 1970's due to ground-water pumping. The Harris-Galveston Coastal Subsidence District was formed in the mid-1970's to reduce pumping and subsidence. Water levels also recovered in the Grand Prairie region of central Arkansas (fig. 36*B*), at Greenville, Mississippi (fig. 36*C*), and at Pascagoula, Mississippi (fig. 36*D*). Pumping and resulting drawdown at Baton Rouge, Louisiana, decreased during the 1970's (fig. 36*F*).

Another interesting feature of the hydrographs is the time at which the vertical-head gradient reverses. In every block shown in figure 36 there is at least one layer whose head crosses the head for another layer or layers at some point in time, indicating a reversal of the vertical-head gradient. Commonly the direction of these reversals is from upward flow to downward flow after pumping begins. In many places the reversal changes differently above and below the major pumping zone so that all flows move toward that zone, such as at Memphis, Tennessee (fig. 36*A*), where the vertical gradient is downward above layer 5 and upward below it.

Hydraulic-head changes in one layer caused by pumpage in an adjacent layer are also shown by the head profiles on plate 1. Most of the pumpage in the Winter Garden area of southern Texas occurs in layer 4, but the head in layer 3 follows the head in layer 4 closely (pl. 1*A*) as the leakance is substantial due to the high sand percentage. The heads in all the layers of the Texas coastal uplands aquifer system decline in this area, largely due to pumpage in layer 4. At Kingsville, also in southern Texas but nearer to the coast, the pumping is in layer 10, but drawdown occurs in layers 9 and 11 as well (pl. 1*A*). Here, the model did not simulate enough drawdown because the area of intense pumping is much smaller than a model block. In addition, the pumping is allocated to more than one block because it does not coincide with the center of a block, further minimizing the simulated drawdown.

The multicounty cone of drawdown at Houston, most extreme in layer 9, causes drawdown in all of the permeable zones of the coastal lowlands aquifer system, and the effects extend as far down in the section as layer

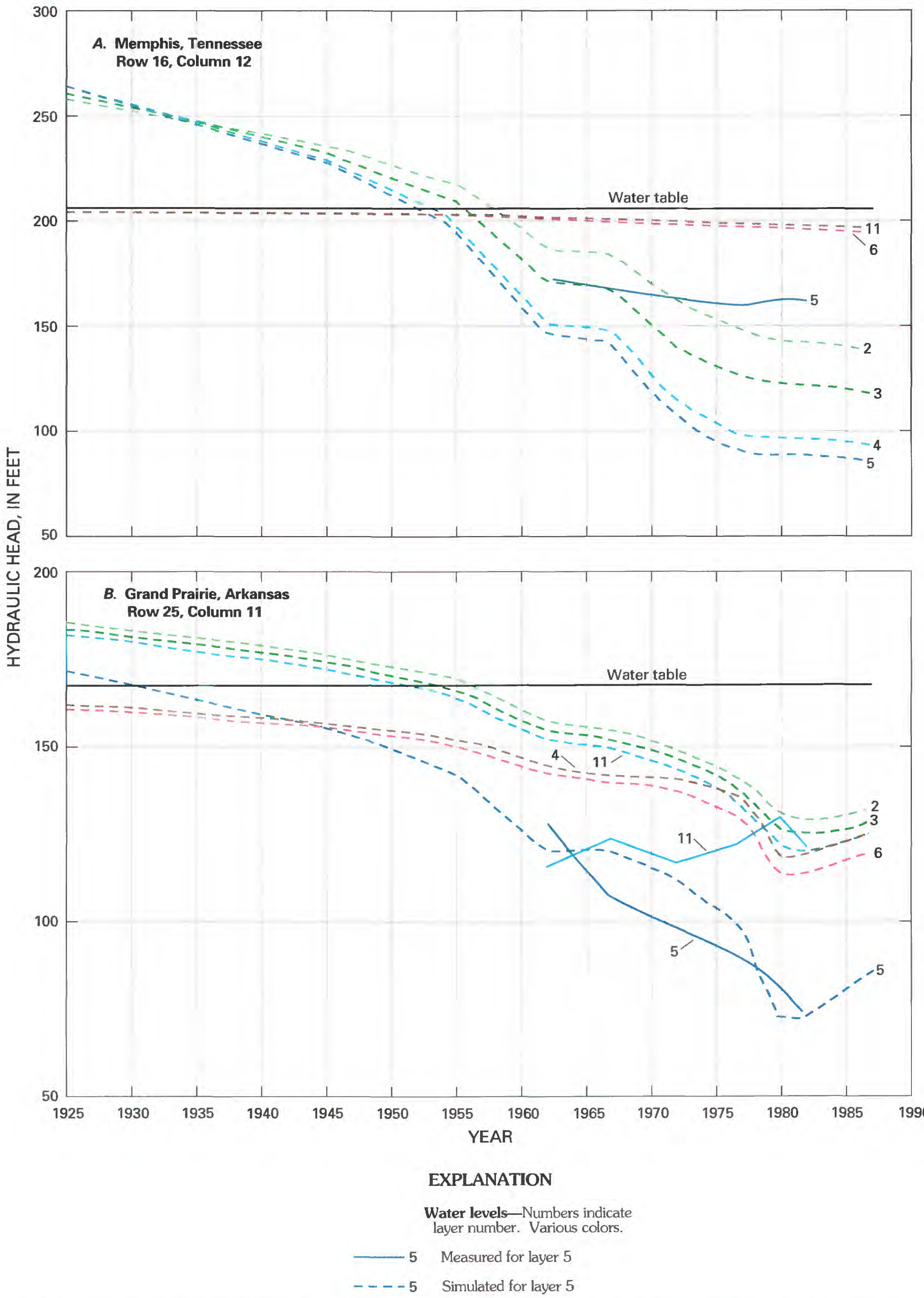


FIGURE 36.—Hydrographs showing simulated and observed block-averaged water levels in selected areas with substantial drawdown, 1925–87: *A*. Memphis, Tennessee, *B*. Grand Prairie, Arkansas; *C*. Greenville, Mississippi; *D*. Pascagoula, Mississippi; *E*. Gulfport, Mississippi; *F*. Baton Rouge, Louisiana; *G*. Alexandria, Louisiana; *H*. Southeast Houston, Texas; *I*. West Houston, Texas; *J*. Winter Garden area, Texas.

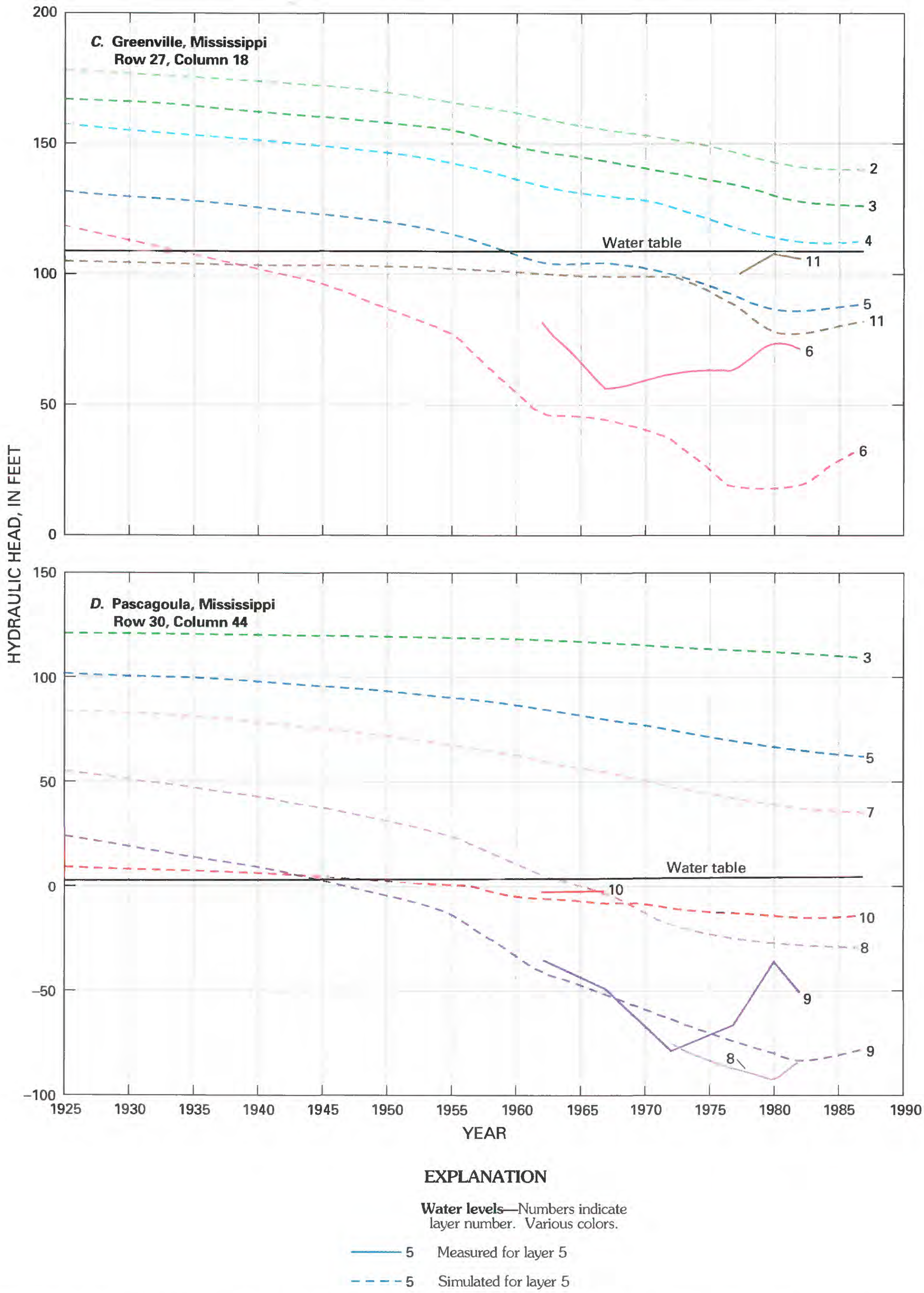


FIGURE 36.—Hydrographs showing simulated and observed block-averaged water levels in selected areas with substantial drawdown, 1925–87: *A*. Memphis, Tennessee, *B*. Grand Prairie, Arkansas; *C*. Greenville, Mississippi; *D*. Pascagoula, Mississippi; *E*. Gulfport, Mississippi; *F*. Baton Rouge, Louisiana; *G*. Alexandria, Louisiana; *H*. Southeast Houston, Texas; *I*. West Houston, Texas; *J*. Winter Garden area, Texas—Continued.

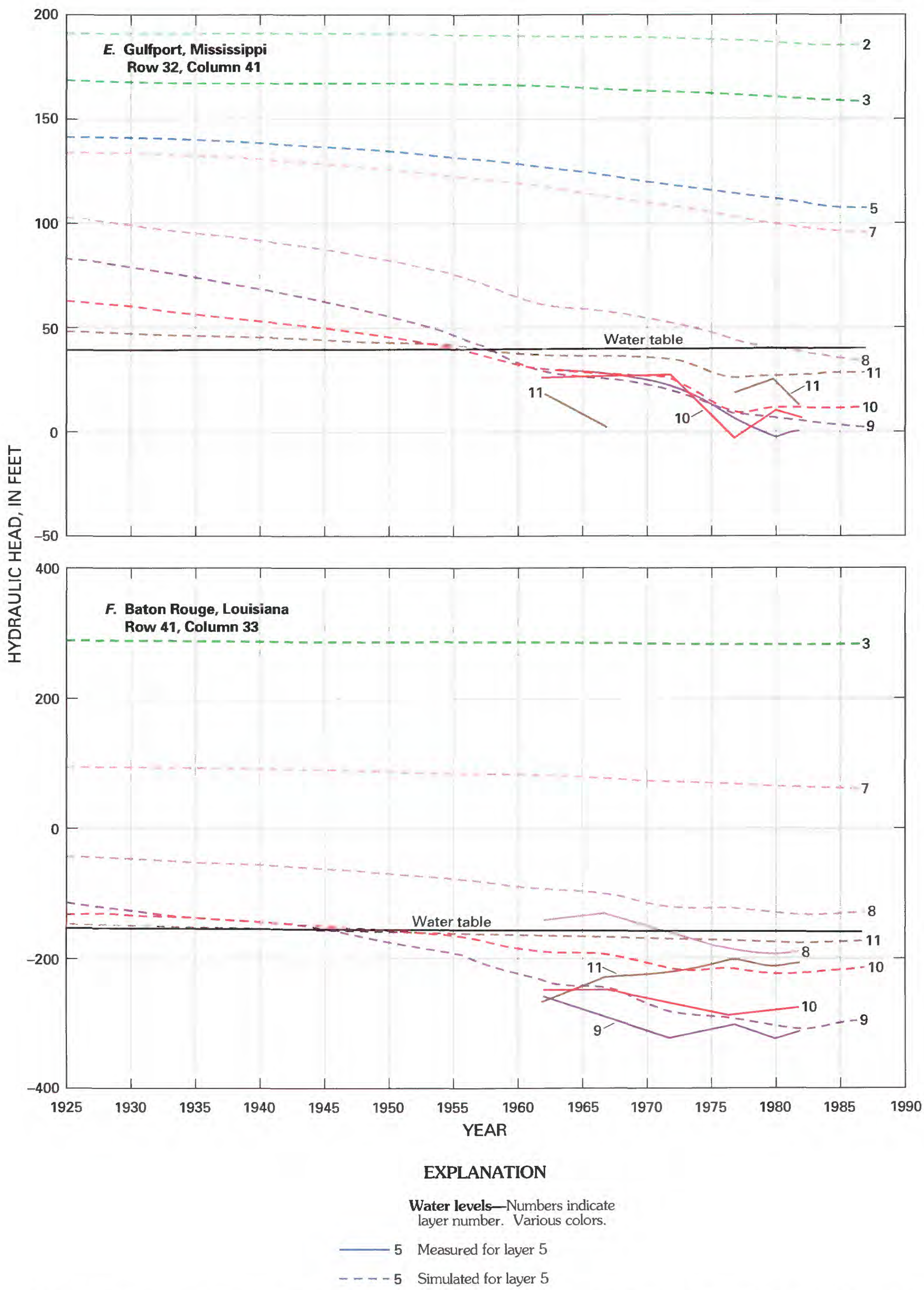
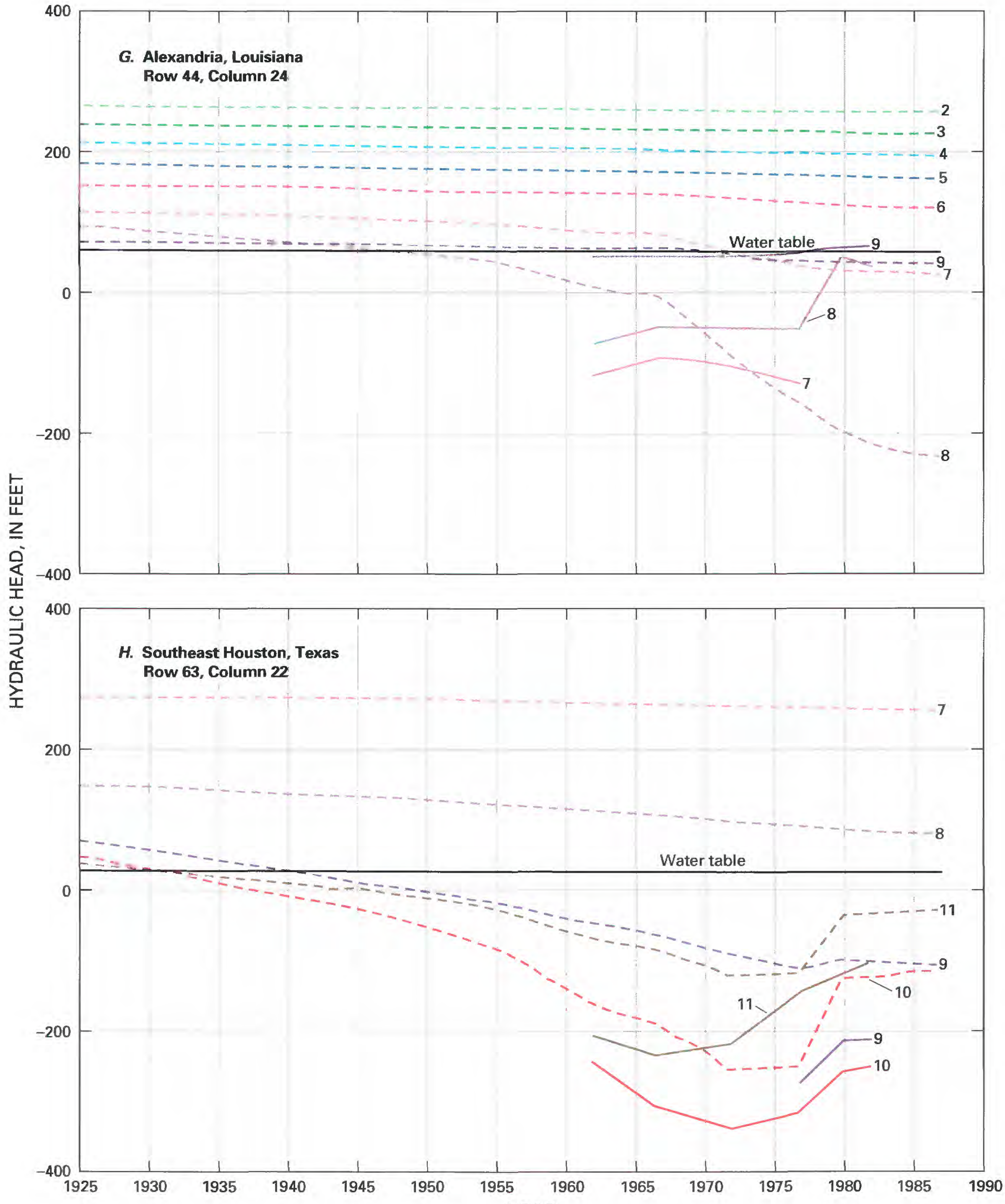


FIGURE 36.—Hydrographs showing simulated and observed block-averaged water levels in selected areas with substantial drawdown, 1925–87: *A*. Memphis, Tennessee, *B*. Grand Prairie, Arkansas; *C*. Greenville, Mississippi; *D*. Pascagoula, Mississippi; *E*. Gulfport, Mississippi; *F*. Baton Rouge, Louisiana; *G*. Alexandria, Louisiana; *H*. Southeast Houston, Texas; *I*. West Houston, Texas; *J*. Winter Garden area, Texas—Continued.



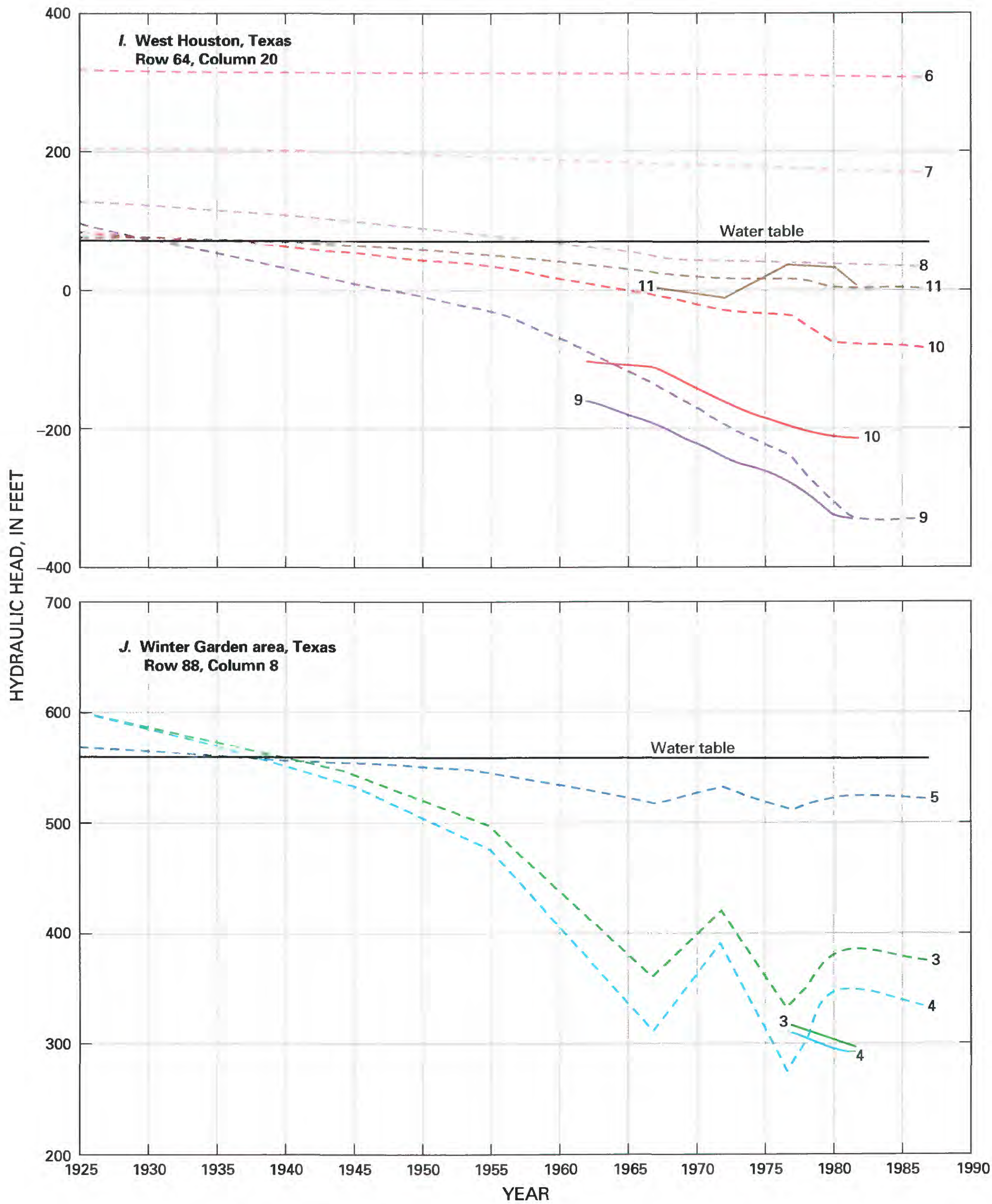
EXPLANATION

Water levels—Numbers indicate layer number. Various colors.

——— 5 Measured for layer 5

– – – – 5 Simulated for layer 5

FIGURE 36.—Hydrographs showing simulated and observed block-averaged water levels in selected areas with substantial drawdown, 1925–87: *A*. Memphis, Tennessee, *B*. Grand Prairie, Arkansas; *C*. Greenville, Mississippi; *D*. Pascagoula, Mississippi; *E*. Gulfport, Mississippi; *F*. Baton Rouge, Louisiana; *G*. Alexandria, Louisiana; *H*. Southeast Houston, Texas; *I*. West Houston, Texas; *J*. Winter Garden area, Texas—Continued.



EXPLANATION

Water levels—Numbers indicate layer number. Various colors.

5 Measured for layer 5

5 Simulated for layer 5

FIGURE 36.—Hydrographs showing simulated and observed block-averaged water levels in selected areas with substantial drawdown, 1925–87: *A*. Memphis, Tennessee, *B*. Grand Prairie, Arkansas; *C*. Greenville, Mississippi; *D*. Pascagoula, Mississippi; *E*. Gulfport, Mississippi; *F*. Baton Rouge, Louisiana; *G*. Alexandria, Louisiana; *H*. Southeast Houston, Texas; *I*. West Houston, Texas; *J*. Winter Garden area, Texas—Continued.

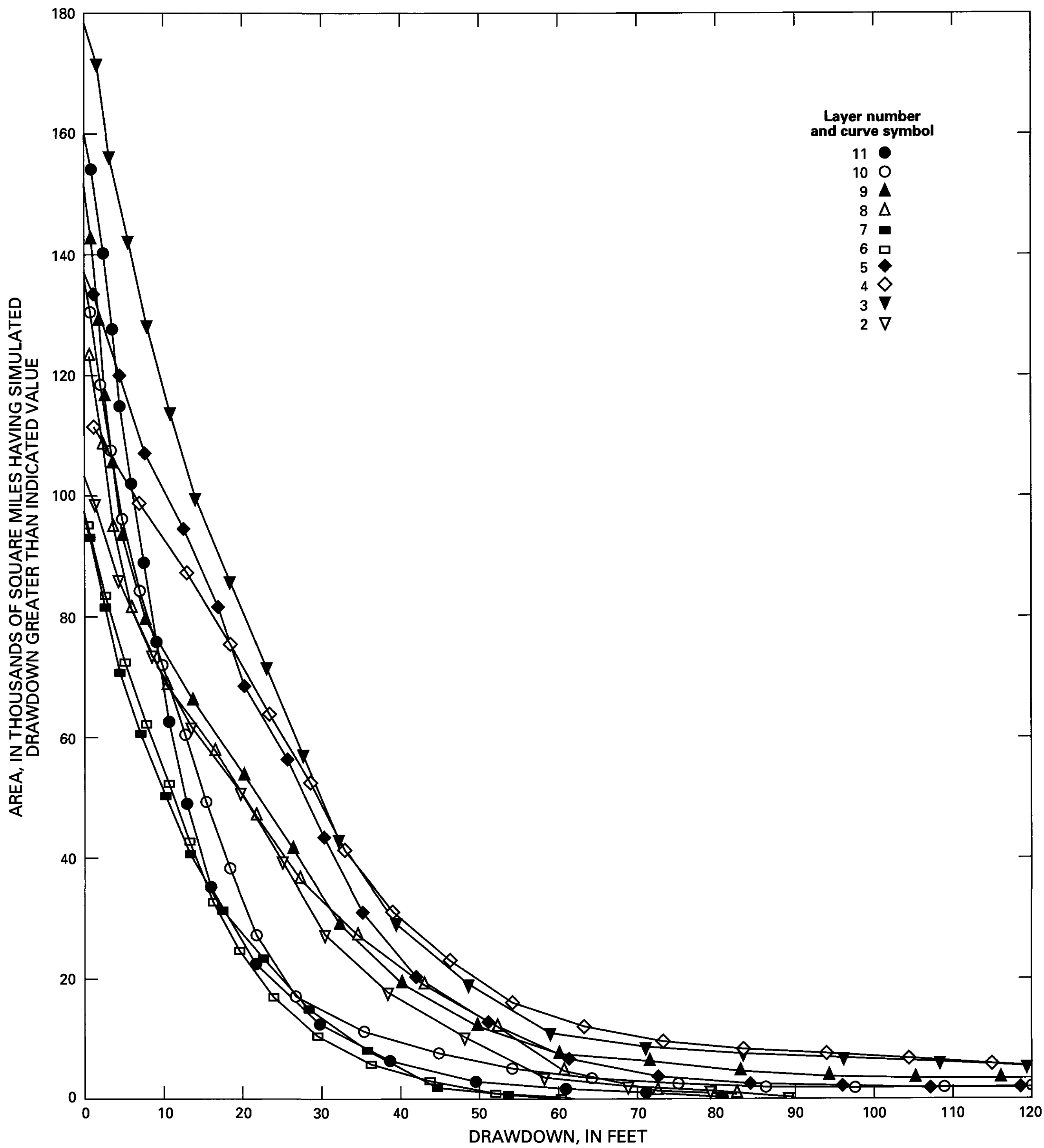


FIGURE 37.—Cumulative distributions of areas of each aquifer (layer) for predevelopment to 1987 simulated drawdown levels.

TABLE 3.—*Summary of 1985 ground-water pumpage by area, type of withdrawal, and model layer, in million gallons per day*

[P, point withdrawals; pumpage is mostly for public supply and industrial use. A, areal withdrawals; pumpage is mostly for irrigation use; quantities from data in Mesko and others (1990). The symbol x denotes absence of layer in the indicated area. Leaders (---) indicate no pumpage in the area. Zero indicates pumpage less than 0.5 million gallons per day. Totals may not agree due to independent rounding]

Area[1]	Type	Model layer 2	3	4	5	6	7	8	9	10	11	All layers
1. Winter Garden	P	x	1	26	1	1	x	x	x	x	x	29
	A	x	4	186	1	0	x	x	x	x	x	191
2. Northeastern Texas	P	x	61	68	9	7	x	x	x	x	x	145
	A	x	24	12	20	8	x	x	x	x	x	64
3. Western embayment	P	0	6	0	65	1	x	x	x	x	x	74
	A	1	13	7	34	2	x	x	x	x	x	56
4. Mississippi alluvial	P	32	1	20	173	23	x	x	x	x	68	317
plain	A	11	2	12	57	3	x	x	x	x	[2]4,910	5,000
5. Eastern embayment	P	29	4	73	143	6	x	x	x	x	x	254
	A	3	1	6	8	3	x	x	x	x	x	21
6. Southern Texas	P	x	---	0	---	---	3	3	5	13	5	29
	A	x	---	0	0	0	0	3	4	5	20	32
7. Southeastern Texas	P	x	---	---	0	2	4	14	193	267	89	568
	A	x	---	0	0	2	3	6	7	54	386	458
8. Southwestern	P	---	---	---	---	0	8	61	6	8	144	226
Louisiana	A	---	---	---	0	0	2	2	0	5	596	605
9. Eastern coastal	P	---	---	---	5	6	76	213	105	45	123	572
lowlands	A	---	0	0	1	1	5	20	20	10	20	77
Total point		61	73	187	396	46	91	291	309	333	429	2,210
Total areal		15	44	223	121	19	10	31	31	74	5,930	6,500
Grand Total		76	117	410	517	65	101	322	340	407	6,360	8,710

[1]See figure 13 for area boundaries.

[2] Small difference in number here and in figure 33*B* is due to southern boundary of area placed at updip limit of subcrop of Vicksburg-Jackson confining unit for modeling purposes rather than at downdip limit as used for this compilation.

6 in the Texas coastal uplands aquifer system, where the water is saline (pl. 1*B*). Pumping near Pascagoula, on the coast of Mississippi, in layer 9 causes drawdown in adjacent layers 8 and 10 (pl. 1*D*).

Layers 4 and 5 at Memphis, Tennessee, comprise the Memphis aquifer, and simulated heads in both layers are very similar throughout time (pl. 1*E*). The pumping in layers 4 and 5 in Fulton County, Kentucky, appears to cause drawdown in layers 3 and 2 as shown by the narrowing drawdown contours near where there is no pumpage in layer 2 or 3 (pl. 4). Farther south in the embayment, the head profiles in all layers are lowered all the way across the embayment (pl. 1*F*).

Although the thickness and sand percentage of the aquifers may be less significant than other factors affecting regional flow, they may have a large effect on the shape and size of a cone of drawdown, as shown by comparing the cones in layer 8 in Louisiana at Alexandria and DeRidder (pl. 4*D*). Pumpage at these two locations is about equal (30 Mgal/d), yet the cone of drawdown at Alexandria is about four times deeper. An explanation is that the sand percentage is about three times larger and the aquifer thickness is about twice as large at DeRidder, resulting in a much larger transmissivity at DeRidder as compared to Alexandria.

HYDRAULIC HEAD AND HORIZONTAL FLOW

Ground-water pumping has largely affected the potentiometric surface (pl. 5) and the pattern of ground-water flow (pl. 6) of the 10 layers and permeable zones. The horizontal flow vectors (pls. 3, 5, 6) are resolved at the intersection of four adjacent blocks to facilitate display of pumping conditions where pumping out of one block will be shown as the four surrounding vectors point toward the pumping block. In many locations the hydraulic gradient and horizontal flow directions have reversed from predevelopment conditions. In other cases the location of lowest head and consequently the regional discharge point have moved substantially.

Plates 5 and 6 also show lines of equal dissolved-solids concentration of 10,000 and 35,000 mg/L. These lines represent the average concentration within 100-mi^2 model blocks. In many areas, where it appears that the brine interface intrudes into areas of current pumping, this may not be true. Actually, most of the pumping is from the top part of the permeable zone where the water is fresher than what is shown as the average for the block, or the pumping is concentrated near the corner of the block.

Although random and unknown bias errors are present in the simulations, relative comparisons between one simulation and another or between one location and another have greater certainty because they are not absolute; therefore, similar errors in both parts of a comparison would offset each other. Simulated hydraulic-head maps also contain uncertainty because of errors in the model design and assumptions as well as errors in estimating relevant hydraulic-head values for 100-mi^2 blocks using observed well data for appropriate comparison with simulated heads from the regional flow model. Estimating observed head values for a regional ground-water model is a complex problem owing to a variety of reasons. Models assume discrete blocks of aquifer over discrete time units, requiring discretization of the point well-measurement data over four dimensions—three in space and one in time. Horizontal discretization is complicated due to the relatively flat regional hydraulic gradient, which is affected by large local topographic variations. It also is complicated by the steep horizontal-head gradients occurring around the steep cones of drawdown. Vertical discretization is important but difficult because in some places layers are quite thick, and considerable head difference occurs vertically within the layer. In addition, wells are not measured on the same days of the year or even every year because of changes in the monitoring network design and operation. The details of estimating block-averaged observed hydraulic head are given in the supplement "Simulation Details."

MISSISSIPPI RIVER VALLEY ALLUVIAL AQUIFER AND PERMEABLE ZONE A (HOLOCENE-UPPER PLEISTOCENE DEPOSITS)

In layer 11, the Mississippi River Valley alluvial aquifer in the northern part of the study area and permeable zone A in the southern part, the general pattern of hydraulic head (pl. 5*A*) and flow (pl. 6*A*) has been altered in several areas with large pumpage. Areas with some of the largest ground-water withdrawals from layer 11 are Houston, Texas, and a farming area about 80 mi southwest of Houston, southwestern Louisiana, and the Grand Prairie area of central Arkansas. There would have been much more drawdown from this huge volume of pumping except that the layer is close to surface sources of recharge, has a high sand percentage, and has substantial hydraulic conductivity. Water-table storage may be a factor, but it is thought to be a minor contributor because the water levels have begun to reach steady state. Throughout most of coastal eastern Texas, coastal Mississippi, and some parts of southern Louisiana, upward flow into layer 11 before development became downward flow out of the layer (pl. 6*A*) as ground-water withdrawals increased. Vertical flow direction reversed throughout most of the area underlain by the Mississippi River Valley alluvial aquifer as well. The offshore horizontal flows in layer 11 (permeable zone A) are about the same as existed prior to development.

PERMEABLE ZONE B (LOWER PLEISTOCENE-UPPER PLIOCENE DEPOSITS)

The 1985 offshore flow in layer 10 (permeable zone B) is substantially altered from the predevelopment condition (pls. 3*B*, 6*B*). Many of the cones of drawdown have lowered the hydraulic head to well below sea level (pl. 5*B*). In an area that extends up to 70 mi offshore from the Houston area, the horizontal flow toward the coast is larger than predevelopment flow, and in that part of the area extending up to 50 mi offshore, former upward flow from layer 9 becomes downward flow out of layer 10 and into layer 9. Horizontal flow in an inland direction increased along the coast of Louisiana. Horizontal flow toward the coastline also increased offshore from southern Mississippi.

Layer 10 contains more than 10 separate cones of drawdown, of which 3 have simulated drawdown exceeding 100 ft since predevelopment (pl. 4*B*) near Houston, Texas, at Baton Rouge, Louisiana, and Gulfport, Mississippi. Hydraulic head had dropped below sea level by 1987 at these locations and also north of Beaumont, Texas. The model does not simulate enough drawdown at Kingsville, Texas: the observed drawdown is more than 100 ft and the 1987 head is also below sea level. Despite all of the drawdown in layer 10, throughout most of coastal eastern Texas, coastal Mississippi, and several parts of southern Louisiana, former upward flow into the bottom of layer 10 (pl. 3*B*) during predevelopment had reversed to downward flow out of the layer (pl. 6*B*) by 1987. The flow out of layer 10 is due to drawdown in the underlying layer 9.

PERMEABLE ZONE C (LOWER PLIOCENE-UPPER MIOCENE DEPOSITS)

Layer 9, or permeable zone C, has the maximum simulated drawdown of any layer, with more than 400 ft near Houston, Texas, and with five cones having more than 100 ft of simulated drawdown (pl. 4*C*). The four other areas with large drawdowns are north of Beaumont in Texas, Baton Rouge in Louisiana, and Gulfport and Pascagoula in Mississippi. At all five locations the simulated 1987 hydraulic head is below sea level (pl. 5*C*).

In a few places along the updip extent of layer 9, especially in central Louisiana, there is more downward vertical flow out of layer 9 in 1987 than before development. There were very few other vertical flow reversals at the base of the layer, except that many of the former upward vertical flows under predevelopment conditions (pl. 3*C*) became much larger (pl. 6*C*) after ground-water withdrawals began. Pumping onshore in and around Houston has changed flow directions more than 45 degrees or flow rates more than 50 percent in the brine part of layer 10. The area where flow directions have changed is about 6,000 to 8,000 mi^2 and extends as far as about 60 mi offshore. Changes in flow directions and volumes of similar magnitude were simulated throughout a large area in the brine part of layer 9 in south-central Louisiana.

PERMEABLE ZONE D (MIDDLE MIOCENE DEPOSITS)

Although there is little pumpage from layer 8 (permeable zone D) in the Houston area in Texas, very large drawdown is simulated in this area in response to pumping in shallower layers (pl. 4*D*). A similar situation also exists to some extent in layer 8 near the Baton Rouge area in Louisiana. Pumping in layer 8 has created cones of drawdown with drawdowns of over 200 ft at Pascagoula, Mississippi, and at Alexandria, Louisiana, and the simulated 1987 head is below sea level (pl. 5*D*). Several other areas have lesser but still substantial drawdown (pl. 4*D*). Drawdown in layer 8 (pls. 4*D*, 5*D*) in Claiborne County, Mississippi, is due to withdrawal of ground water from the Mississippi River Valley alluvial aquifer for a powerplant, beginning in the early 1980's.

The distribution of vertical flow relative to the base of layer 8 did not change much from predevelopment conditions (pl. 3*D*) to 1987 conditions (pl. 6*D*), except for an area in south-central Louisiana where small downward flows reversed to upward flows. In this area, the horizontal flow magnitude also increased and flow direction shifted to the east, toward the cone of drawdown created by pumping in shallower layers at Baton Rouge. Another change in horizontal flow was in the cone of drawdown surrounding Bogalusa, in Washington Parish, Louisiana, on the Louisiana-Mississippi border.

PERMEABLE ZONE E (LOWER MIOCENE-UPPER OLIGOCENE DEPOSITS)

Layer 7 (permeable zone E) has numerous cones of simulated drawdown (pl. 4*E*), none of which exceed 100 ft. The simulated cone at Laurel, Mississippi, underestimates the observed drawdown, probably because the pumpage occurs from a small area that is not accurately represented in the coarse-mesh model. The largest and most extreme simulated drawdown occurs in the Alexandria area of Louisiana. A rather large cone exists to the west of Houston, Texas, but it is not as deep as the one at Alexandria.

The distribution of vertical flow across the Vicksburg-Jackson confining unit (upward into layer 7) is not greatly changed from predevelopment conditions. In a few parts of southern Louisiana, Mississippi, and Alabama, and in the western tip of Florida, small downward flows (pl. 3*E*) have reversed direction to upward flows (pl. 6*E*). Horizontal flows in layer 7 have not changed greatly either, although there are many occurrences of flow being diverted towards one of the many small pumping cones.

UPPER CLAIBORNE AQUIFER

By 1987, layer 6 (upper Claiborne aquifer) had at least seven areas with regional drawdown even though most of the cones of drawdown are related to pumping in other layers. Very little ground water is withdrawn from layer 6 (fig. 35, pl. 4*F*). For example, at Alexandria, Louisiana, there is drawdown in layer 6 because of pumping in the layers above. The drawdown at Alexandria is interesting because the ground-water withdrawals are from layer 8 above the thick Vicksburg-Jackson confining unit. At Bryan, Texas, and in central Arkansas, drawdown in layer 6 is mostly in response to pumping in layer 5 below. At Jackson and Greenville in Mississippi, the drawdown is due to pumping from layer 6. The shape of the regional hydraulic-head surface of layer 6 has not greatly changed due to ground-water pumping (pls. 2*F*, 5*F*).

Layer 6 subcrops beneath the Mississippi River Valley alluvial aquifer throughout a broad area and becomes a conduit from recharge areas to discharge areas associated with other aquifers. The vertical flow across the base of layer 6 was different in 1985 (pl. 6*F*) than it was under predevelopment conditions (pl. 3*F*), especially in the Mississippi embayment. In and around Memphis, Tennessee, and across western Arkansas, the upward flow of ground water to the regional discharge area was reversed to downward flow into layer 5, which is heavily pumped. In the Texas coastal uplands aquifer system, the pattern of flow is not greatly changed. The horizontal flows are not changed as much as the vertical flows in the Mississippi embayment aquifer system. The largest regional discharge area still occurs in northern Louisiana.

MIDDLE CLAIBORNE AQUIFER

Layer 5, or the middle Claiborne aquifer, had more than 15 different cones of drawdown in 1987 (pl. 4*G*). This includes several large cones where large quantities of water are pumped out of layer 5, such as at Memphis, Tennessee, central Arkansas, and northern Louisiana. The simulated drawdown in layer 5 in central Arkansas and northern Louisiana substantially underestimates the observed drawdown. This error will be discussed in the supplement "Simulation Details." Layer 5 also has some large cones of drawdown where much of the pumping is from adjacent aquifers, such as layer 4 in the Winter Garden area of southern Texas and also to some extent in layer 6 in central Mississippi near Jackson. Despite all of the drawdown in layer 5, the 1987 potentiometric surface (pl. 5*G*) still generally resembles the predevelopment potentiometric surface (pl. 2*G*) with the large regional flow area in southern Arkansas and northern Louisiana.

Vertical flow between layer 5 and the underlying layer changed considerably in certain areas from predevelopment (pl. 3*G*) to 1985 (pl. 6*G*). Most notable was the large-scale reversal of flow in the Winter Garden area where generally upward vertical flow changed to large downward flow throughout an extensive area. Other areas where vertical flow direction changed from upward to downward are in northwestern Arkansas, western Tennessee, eastern Texas, and central Mississippi. In the Memphis area of Tennessee, small upward vertical flows became very large, maintaining the upward direction. Large patterns of concentric horizontal flows that did not exist prior to development (pl. 3*G*) can be seen on the 1985 simulated flow map (pl. 6*G*).

LOWER CLAIBORNE-UPPER WILCOX AQUIFER

Layer 4, the lower Claiborne-upper Wilcox aquifer, contains the largest area of any aquifers with more than 40 ft of simulated drawdown from predevelopment to 1987 (fig. 37). The extensive area having 40 ft or more drawdown is due to the large area of public-supply pumping around Memphis, Tennessee, in the Mississippi embayment aquifer system, and in the Winter Garden area of southern Texas, where intensive pumping for irrigation causes widespread drawdown (pl. 4*H*). Also, more than 80 percent of the area underlain by layer 4 had more than 10 ft of simulated drawdown from predevelopment to 1987, a much higher percentage than any other layer (fig. 37). This is due in part to the model design in which the massive sand unit at the top of the Wilcox Group is defined as layer 4 and has a more limited extent, especially downdip.

Throughout a large part of the Mississippi embayment aquifer system, the drawdown in layer 4 is caused by pumping in layer 5 because the two layers either are not separated by a regional confining unit north of 35° latitude (Hosman and Weiss, 1991) or are separated by a leaky confining unit. Note, however, that where the two layers are not separated, the heads are not exactly equal (pls. 5*G*, 5*H*), but the difference may be too small to be observed onsite when combined with other factors causing differences (horizontal location of wells and timing of measurements). Even though there is no intervening regional confining unit and the units are generally sandy, the interbedded fine-grained sediments do tend to restrict vertical flow.

The simulated potentiometric surface for layer 4 in 1987 (pl. 5*H*) is largely changed from the predevelopment surface (pl. 2*H*) in the Winter Garden area, the Memphis area, and in a couple of cones of drawdown in eastern Texas. In other areas, the configuration of the surface is largely similar to what it was in predevelopment except that most of the heads are lowered.

Over the study area, downward flow between layer 4 and the layer below it increased from 153 Mgal/d in predevelopment to 236 Mgal/d in 1985 while upward flow stayed about the same (fig. 33). The direction of vertical flow also changed in many locations. In the Winter Garden area, changes in vertical flow volume and direction developed (pls. 3*H*, 6*H*). The horizontal flow in the Winter Garden area also changed considerably from a southward and westward direction in predevelopment to generally horizontal radial flow in 1985 toward a point about 30 mi downdip from the landward extent of the aquifer. Horizontal flow direction changed all the way to the downdip limit of the aquifer. In the general vicinity of the Brazos River, the vertical flow also reversed to become downward. Around Tyler in northeastern Texas, the vertical flow reversed from downward to upward due to pumping in layer 4 where there was also horizontal radial flow toward the center of pumping.

In the Mississippi embayment aquifer system, in layer 4 south of about 33.5° latitude, there are no noticeable differences in flow in the downdip part of the layer except for an increase in downward flow in south-central Mississippi. Farther north in the embayment, pumping causes several sizable changes in ground-water flow. West of the Mississippi River across from Tennessee, the vertical flow relative to the base of layer 4 reversed from upward to downward in 1985. The flow reversal affected four adjacent aquifers, with water flowing downward through layers 4 and 3 to satisfy a small volume of pumping from layer 2. Apparently the distances to layer 2 outcrop areas with direct recharge are so great that most of the flow to the pumping centers has to be drawn vertically from above where the flow paths are short and the hydraulic gradients relatively steep even though the vertical permeability is low. The reversal of vertical flow directions is also shown in the head profiles shown on section N–N' (pl. 1*N*). However, just east of the Mississippi River at Memphis, the vertical flow changed from downward in predevelopment to upward in 1985 to supply pumping in layer 5. This flow reversal also affected the vertical flows between layers 2 through 5.

MIDDLE WILCOX AQUIFER

Layer 3, or the middle Wilcox aquifer, has a pattern of drawdown that is very similar to the pattern of drawdown at the same locations in layer 4 (pl. 4*I*). In general, the drawdown in layer 3 is slightly less than the drawdown at the same location in layer 4. The total pumpage in layer 3 is about one-fourth of the amount of pumpage in layer 4, and much of the pumpage occurs in the top part of layer 3 adjacent to pumping in layer 4. One exception is in Brazos County in east-central Texas at Bryan and College Station. Here, the most productive part of layer 3 is a massive sand locally known as the Simsboro Formation (G.E. Fogg, The University of Texas at Austin Bureau of Economic Geology, oral commun., 1988). The drawdown in layer 3 is greater than in layer 4 because almost all of the ground water withdrawn at this location is from layer 3. Except as noted, the changes in the hydraulic gradient in layer 3 are similar to the changes in layer 4.

In 1985, just as in predevelopment times (pl. 3*I*), the horizontal flow in the middle Wilcox aquifer, layer 3, is much larger in the Texas coastal uplands aquifer system than it is in the Mississippi embayment aquifer system (pl. 6*I*). The layer flows are mainly due to larger hydraulic conductivity (fig. 14) and thickness of layer 3 in Texas (fig. 12*A*) than to the east. In the Mississippi embayment aquifer system, some hydrologists consider the middle Wilcox aquifer (layer 3) to be a confining unit. It does act like a confining unit in many locations when considered on a local scale. The horizontal flow in layer 3 in Texas did not change much during development except for flow to some pumping centers, especially the Winter Garden area, where flows from the outcrop on the northern side of the Winter Garden area increased considerably (pls. 3*I*, 6*I*). The lateral flows in the Mississippi embayment aquifer system were generally small during predevelopment except for a few locations (pl. 3*I*), and the changes that occurred were also small (pl. 6*I*). Vertical flow relative to the base of layer 3 was not simulated in the Texas coastal uplands aquifer system because layer 2 does not exist in Texas. There were no substantial changes in vertical flow in layer 3 in the Mississippi embayment aquifer system other than the ones noted the discussion of layer 4.

LOWER WILCOX AQUIFER

There are only a few cones of drawdown in the lower Wilcox aquifer (layer 2) (pl. 4*J*), near Memphis, Tennessee, central Arkansas, northern Louisiana, and east-central Mississippi. The simulated drawdown at Memphis exceeded 100 ft, whereas the other areas had drawdowns of only 30–50 ft. At the two locations west of the Mississippi River, the main pumping is much shallower, yet the drawdown extends far downdip where the concentration of dissolved solids in ground water exceeds 10,000 mg/L (pl. 4*J*). The shape of the horizontal hydraulic-head gradient in 1987 (pl. 5*J*) is very similar to what it was in predevelopment times (pl. 2*J*) except for the cone of drawdown at Memphis.

Some differences in horizontal flow due to ground-water pumpage in layer 2 (pl. 6*J*) were simulated. Although the general pattern in 1985 is not greatly different from predevelopment, the flow that originally moved across the northern embayment to discharge to the subcrop of layer 2 is captured by the pumpage at west Memphis and other well fields in layer 2 located in Arkansas. The change is indicated by the flow vectors pointing in the opposite direction from the predevelopment flow direction in the area west of Memphis (pl. 6*J*).

REGIONAL RECHARGE AND DISCHARGE

The rates (fig. 38) and distribution (fig. 39) of regional recharge and discharge were substantially changed by development of the ground-water resource. The 1985 pumpage was about three times greater than the rate of predevelopment regional recharge. Pumpage and the resulting declines in hydraulic head increased the regional recharge by more than three times to more than 9 Ggal/d. Part of the increase in recharge is the capture of water that was previously locally discharged. Regional discharge decreased to about 1 Ggal/d, nearly one-third of its predevelopment rate. Thus, more than 65 percent of the simulated 1985 pumpage was supplied from increased regional recharge. About 20 percent of the pumpage was captured regional discharge, and less than 15 percent was from decrease of ground-water storage (fig. 38). A large part of the increased regional recharge is captured local discharge. Much of the increased recharge in the irrigated areas comes from seepage of irrigation water past the root zone. Jorgensen (1975, p. 55) estimated that as much as 30 percent of the ground water pumped for irrigation in the Katy area, west of Houston, Texas, returned to the shallow aquifer. The volumes of water from aquifer storage shown in figures 38 and 40 include confined storage summed from the simulation reported here as well as loss of unconfined storage in the Mississippi River Valley alluvial aquifer simulated by Ackerman (in press). The water-budget values shown in figures 38 and 40 were calculated by taking the regional model-simula-

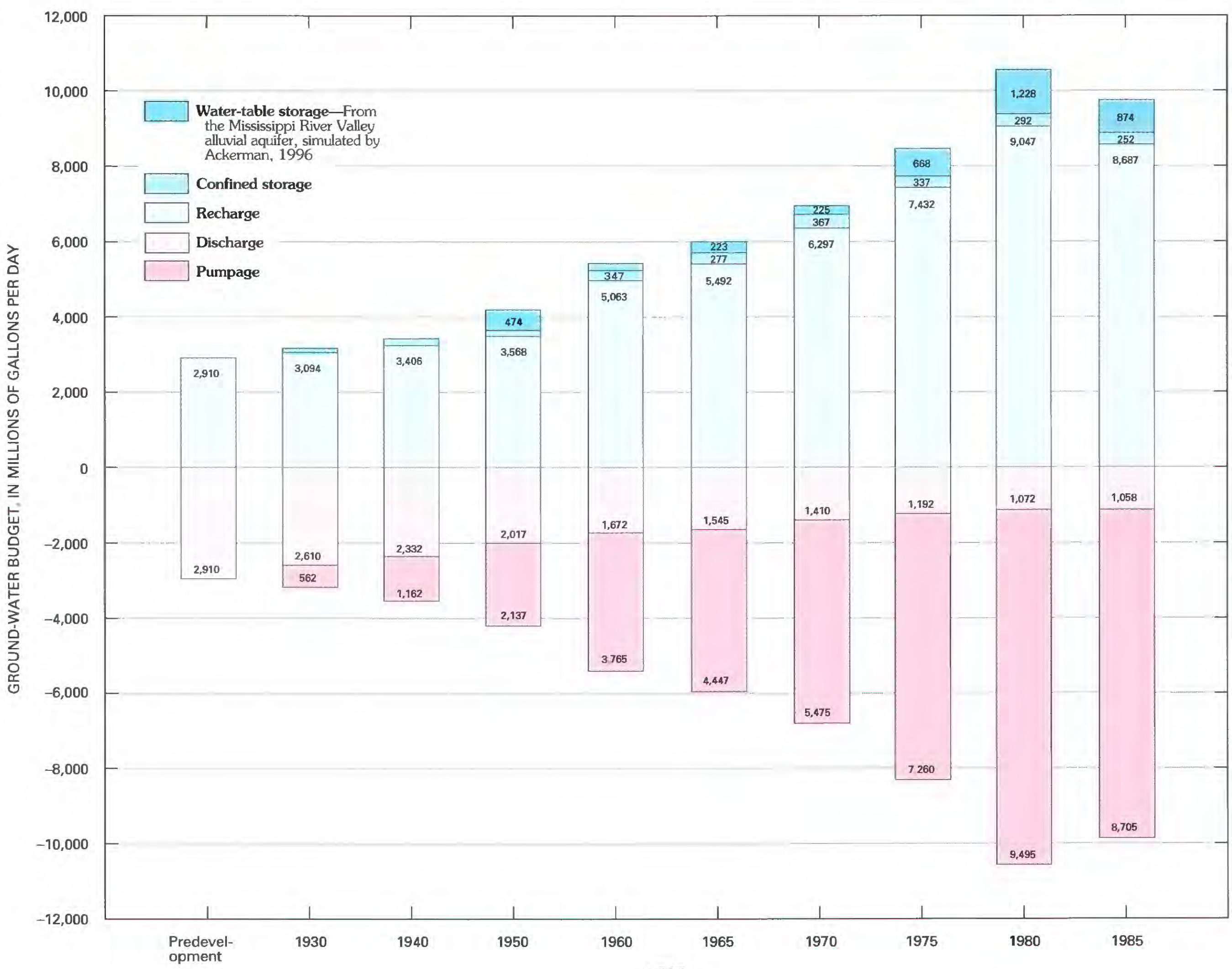


FIGURE 38.—Change in simulated regional ground-water budget from predevelopment to 1985.

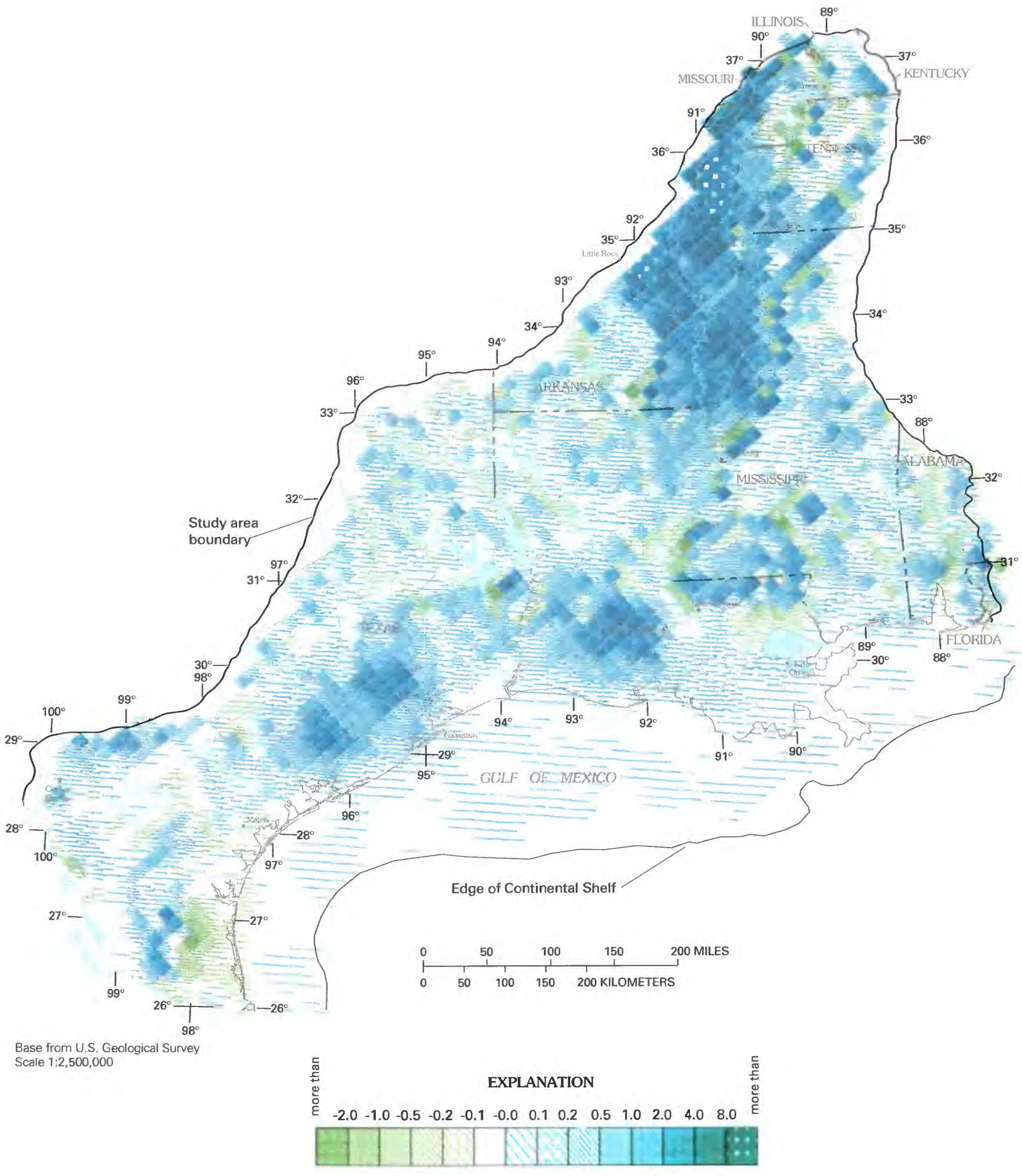


FIGURE 39.—Simulated 1985 net regional ground-water recharge and discharge.

tion results and adding Ackerman's water-table storage values to simulated storage and subtracting the same values for simulated recharge.

Although there were large changes in the ground-water budget caused by ground-water withdrawals, the changes are proportionately smaller when put in perspective of the overall hydrologic budget of the study area. The volumes of precipitation (500 Ggal/d summed from fig. 4) and runoff (140 Ggal/d summed from fig. 6) are much greater than the 6.7 Ggal/d change in simulated ground-water recharge. The major stream inflow to the study area (415 Ggal/d) as estimated by Williamson and others (1990, table 1) is almost as large as the volume of precipitation. Natural evapotranspiration is difficult to estimate but probably is 300 to 400 Ggal/d.

As of 1985, the Mississippi Alluvial Plain still had the largest volume of ground-water flow (fig. 40), as it did before development (fig. 31). However, the southeastern Texas area had the second largest total flow in 1985. Areas in southwestern Louisiana, Winter Garden, and northeastern Texas also had large increases in total flow. The eastern coastal lowlands area had the largest proportion of flow coming from confined storage. Most of the change in storage in the Mississippi Alluvial Plain is from water-table changes that are unevaluated for the other areas. Many of the interarea flows stayed relatively constant despite substantial changes in other components of the ground-water budget.

DISTRIBUTION OF REGIONAL RECHARGE AND DISCHARGE

In 1985, the proportion of the study area that contributed net regional recharge increased from 41 percent of the study area to 66 percent of the study area (figs. 32, 39). The area that changed from having net regional discharge to net regional recharge includes most of the area underlain by the Mississippi River Valley alluvial aquifer and most of the area underlain by the coastal lowlands aquifer system within about 40 mi of the coastline, except in southern Texas. In southern Texas along the coast, there is still net regional discharge because of little ground-water pumpage. Even offshore, there has been a change from discharge to recharge. Off the coast near Houston during predevelopment, flow was discharging from the aquifers to the Gulf of Mexico, whereas in 1985 the simulated flow was from the Gulf of Mexico into the aquifer. On the north and west edges of the Winter Garden area there have been changes from net discharge to net regional recharge. In the areas of intensive irrigation such as the Mississippi Alluvial Plain, southwestern Louisiana, and west and south of Houston, Texas, a substantial portion of the simulated net regional recharge actually may be irrigation return flow.

VERTICAL FLOW BETWEEN LAYERS

By 1985, most of the simulated net vertical flows between layers changed direction or changed considerably in magnitude (fig. 33). The amount and direction of the net vertical flows are sensitive to the aquifer parameters chosen for the simulation because net flows are a difference between two commonly larger numbers. The absolute rates of upward and downward flow considered separately are less sensitive to simulation parameters and hence more certain than the net vertical flows, especially if considered as relative numbers to compare to other layers or prepumping conditions, or both. The magnitudes of vertically upward flow between layers remained about the same as before development in most layers. Exceptions to this were upward flow into layer 6, which decreased to less than one-half, and upward flow into and out of layer 10, which decreased to slightly more than one-half. This is probably because in many places with little or moderate pumpage, the vertical flow remained similar. However, in layers 6 and 10, much water that had moved upward into the layers during predevelopment conditions was moving to the large pumping areas in layers 5 and 9 by 1985.

Downward vertical flows between layers increased substantially in all of the aquifers and permeable zones (fig. 33) as a result of ground-water pumpage and resultant reversal of vertical gradients in many places (pl. 1). Downward vertical flow increased so much that all the net flows between l ayers, which were generally upward before development, became downward except between layers 6 and 7. Flow across the Vicksburg-Jackson confining unit between layers 6 and 7 increased slightly in both directions due to development. The smallest changes in vertical flow were across this unit because there is minimal pumpage in the two adjacent permeable layers.

In every layer, the regional recharge increased and some of the regional discharge in the outcrop areas was captured owing to ground-water pumpage (fig. 33). The Mississippi River Valley alluvial aquifer and middle Claiborne aquifer had the largest amounts of pumpage and the largest increases in regional recharge. The size ranking of all of the layers shown in figure 33 would be very similar when ranking amount of pumpage, or 1985 recharge, or change in recharge from predevelopment. This indicates that increased pumpage is correlated with increased regional recharge.

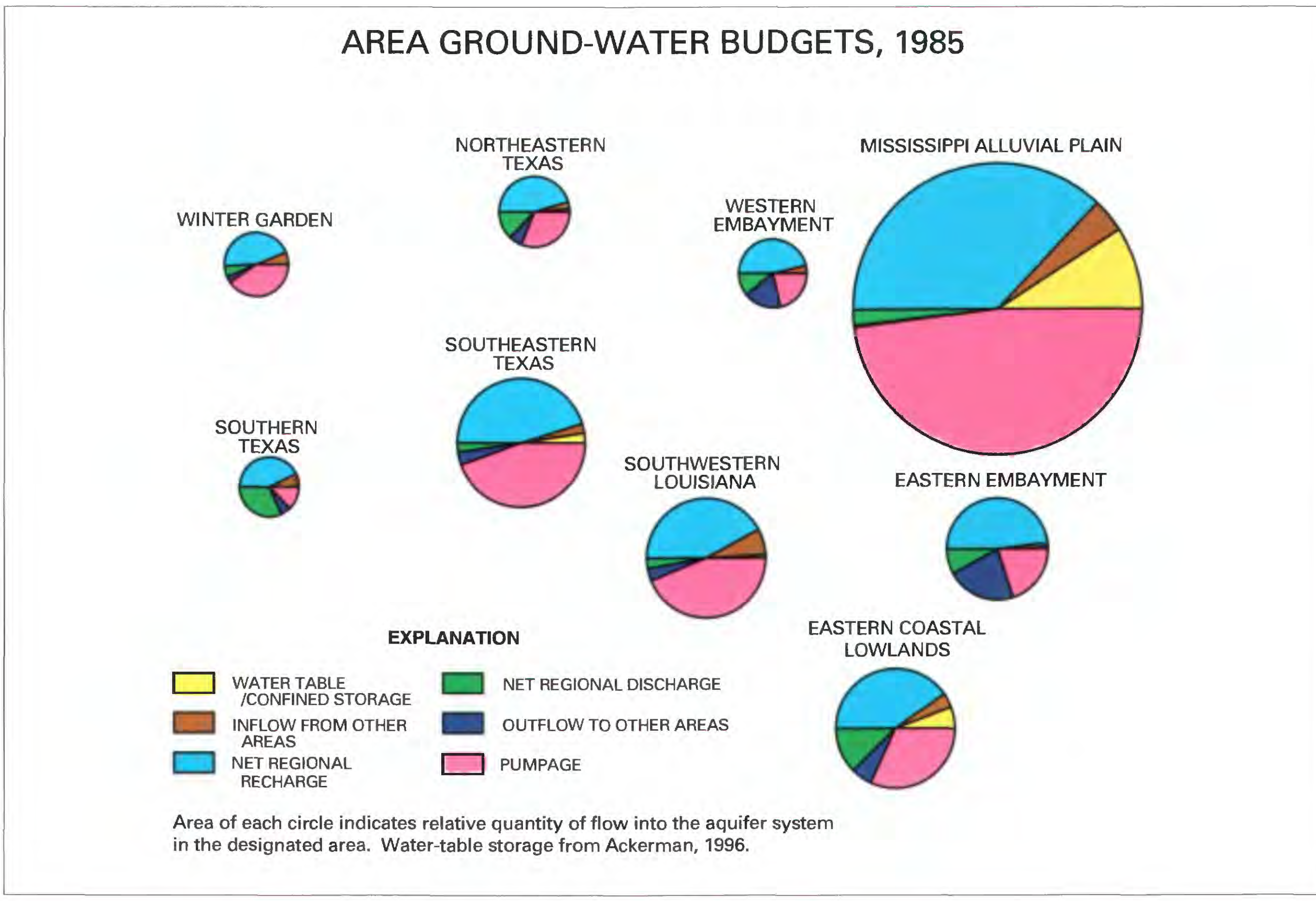


FIGURE 40.—Simulated 1985 ground-water budgets, by area.

Upward vertical discharge from the layers subcropping the Mississippi River Valley alluvial aquifer decreased by nearly one-half, and downward flow into the subcropping layers increased to more than double its predevelopment value (fig. 33). Layer 6 continued to have the most flow interactions with the alluvium, probably due to the large area of contact. Layer 6 has minimal pumpage yet has large inflows and outflows. It is the only layer in which the change in recharge was much larger than the 1985 pumpage. Layer 6 acts as a large regional conduit, carrying water from recharge areas along regional flow paths (pl. 6) to cones of drawdown caused by pumping in deeper layers.

Layers 3 through 5 and 8 through 11 have ratios of net recharge (total recharge minus total discharge) in the outcrop (and subcrop of layer 11) to pumpage of between 0.68 and 1.01. These ratios indicate that on the average, most of the water withdrawn from the layer was supplied from increased recharge and decreased discharge within the layer itself rather than by inflow from adjacent layers. The ratios for layers 2 and 7, which have very narrow outcrop and subcrop bands (fig. 11), were only 0.04 and 0.17, indicating that most of the pumpage was supplied by vertical flow from adjacent layers. The ratio for layer 6, which has a very wide outcrop and subcrop band, was 2.16, indicating that it was capturing additional recharge and discharge. Layer 6 supplied the excess flow as leakage to underlying layers to satisfy pumpage from other layers, primarily layer 5.

The direction of vertical flow and an indication of its magnitude during both predevelopment and 1985 conditions are shown on the sections that also show both hydraulic-head profiles (pl. 1). The head profiles do not always indicate flow directions in the brine part of the aquifer system. The location of brine water can be approximated by the concentrations of dissolved solids that exceed 35,000 mg/L as posted on the sections.

AQUIFER STORAGE

The simulated proportion of pumpage being supplied from confined aquifer storage in 1985 was less than 3 percent (fig. 38). This is a small percentage, as was the figure of 7 percent supplied by storage in a simulation of 1985 conditions by Kuiper (1994). The total simulated change in confined storage for the period 1925–87 was about 3.5×10^{12} gal, or about 4 percent of the total pumpage of 88×10^{12} gal. The change in water-table storage was not simulated because a constant-head layer was used to simulate regional recharge and discharge and because the observed change in water-table storage volume was not estimated due to the lack of water-table measurements over time. This limitation will be discussed more fully in the supplement "Simulation Details." In addition to the simulated change in confined storage, figures 38 and 40 also show changes in water-table storage simulated by a subregional model for the Mississippi River Valley alluvial aquifer (Ackerman, 1996).

Although only a small proportion of 1985 pumping was supplied from aquifer storage, on the average the lateral extents of cones of drawdown are far from where they would eventually stabilize if 1985 pumping conditions were continued until equilibrium was reached. As a cone of drawdown expands, its area increases greatly (in proportion to the square of the radius), so a very large volume of water from aquifer storage is released with a small change in hydraulic head. Therefore, when the cone becomes large, its rate of lateral expansion slows substantially. If no boundaries or sources are intersected in a large flow system, the cone will keep extending very slowly over a very long period of time. This effect was demonstrated by Williamson and others (1990, p. 110–112) by comparing two simulations. One simulated the aquifer system under 1980 conditions with some water being released from aquifer storage, and the other simulated no aquifer storage available so that the maximum steady-state drawdown was reached. The 50-ft contour of the cone of drawdown in layer 10 moved from 5 mi offshore in 1980 to 30 mi offshore when steady-state conditions were reached. The 25-ft contour of drawdown in layer 10 moved from 30 mi offshore in 1980 to near the edge of the Continental Shelf about 100 mi offshore when steady-state conditions were reached. The drawdown near the center of the cones is nearly equal in both simulations (Williamson and others, 1990).

The area with the largest proportion of flow coming from confined storage in 1987 is the eastern coastal lowlands area (fig. 40). This is partly due to the fact that withdrawals in some other areas that are more extensively developed actually declined in the 1980's. Most of the change in storage in the Mississippi Alluvial Plain area is from water-table changes (Ackerman, 1996) that are unevaluated for the other areas, as was mentioned at the beginning of this section. Many of the interarea flows stayed relatively constant (figs. 31, 40) despite large changes in other components of the ground-water budget.

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 shallow wells shows a decline from the mid-1940's until the early 1970's, when the water levels began to stabilize (Williamson and others, 1990, fig. 22). This stabilization probably was due to the decrease in industrial pumpage in the areas that had previously had the largest declines. The areas with the largest increases in pumpage are mostly irrigated areas underlain by the Mississippi River Valley alluvial aquifer, which has large hydraulic conductivity, short distances to large surface-water bodies, and, in places, a large storage coefficient, 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 a procedure outlined by Williamson and others (1990, p. 69).

LAND SUBSIDENCE

Decreases in hydraulic head due to pumping cause increased loading on the skeleton of the aquifers and can cause compaction and land subsidence if the fine-grained deposits are deformable. This has a great effect on the aquifers because the effective storage coefficient during compaction could be two or more orders of magnitude larger than the elastic storage coefficient of a confined aquifer, thus slowing the head decline by providing a source of water to the pumping wells. Prudic and Williamson (1986) used a technique to simulate a compacting regional aquifer in the Central Valley of California similar to the method Meyer and Carr (1979) used in the Houston area of Texas. More recently, Leake and Prudic (1987, 1988) and Leake (1990) have enhanced the method to an implicit formulation that is more accurate. Kuiper (1994) independently developed a similar technique that is used in this study.

The subsidence algorithm in the model operates so that if the hydraulic head in the aquifer drops more than a set amount, called the critical head (80 ft was used), the elastic storage coefficient is switched to an inelastic value, which is commonly two or more orders of magnitude larger than the confined storage value. The critical head is then reset to the new lowest head value in that model block. Thus the process inside fine-grained beds of the aquifer system is simulated where there is compaction. If the head rises above the previous lowest value, called the critical head, then the storage switches back to the confined value. The simulated land subsidence is estimated by summing the volumes of change in confined storage in a grid cell and dividing by the area to get a depth of water removed from storage. The amount of water removed from confined storage approximately equals the volume of land subsidence over a large area. The part of this land-surface change that is caused by change in elastic confined storage would rebound if the water levels rose again, as is the case in the San Joaquin Valley in California and elsewhere.

Simulated land subsidence caused by withdrawal of ground water generally approximates the major features of estimated land subsidence in the study area (fig. 41). Subsidence has been observed for the period 1906–83 in the Houston area (Gabrysch, 1969, 1977, 1982, 1984), and although it was simulated for the period 1925–87, both periods should show similar subsidence. A maximum of 1.2 ft of land subsidence has also been observed in the Baton Rouge area of Louisiana from 1934 to 1976 (Whiteman, 1980, fig. 1). Although subsidence has only been documented in some detail in those two areas, it may have occurred elsewhere. Subsidence over large areas commonly goes unnoticed until it reaches several feet in magnitude, especially in an inland area where it will not contribute to flooding. At inland locations, subsidence may occur without any other effects if it proceeds slowly enough and over a large enough area, as is commonly the case when it is caused by widespread ground-water withdrawals.

Two national studies (Chi and Reilinger, 1984; Holdahl and Morrison, 1974) of geodetic data for evidence of land subsidence at benchmarks have isolated vertical movements that might be related to land subsidence in the study area (fig. 41, table 4). At Robstown, Texas, the estimated subsidence probably resulted from oil production (R.K. Gabrysch, U.S. Geological Survey, oral commun., 1991). Near New Orleans, Louisiana, the estimated land subsidence is probably mostly due to compaction of peat soils (Snowden, Simmons, and others, 1977; Snowden, Ward and Studlick, 1980). No inelastic compaction was simulated in the Eocene-age sediments because it was presumed that they had already undergone previous consolidation stresses. That assumption may not hold for all locations, which could account for some of the discrepancies shown in figure 41.

Little of the water pumped comes from confined aquifer storage even when the water from inelastic storage is included in the totals. In the Houston area, only 8 percent of the pumpage was from confined and compaction storage by 1987 (fig. 53). Nearly all of the water derived from storage in the Houston area was from layer 9. About 31 percent of the water pumped from layer 9 in this area was from confined and inelastic storage combined.

GROUND-WATER VELOCITY AND POTENTIAL FOR SALINE-WATER INTRUSION

Large-scale ground-water pumpage in the freshwater part of the aquifer systems has markedly changed flow patterns in the brine part of the systems and induced flow updip toward pumping areas (compare pls. 3, 6). However, it will take decades or centuries for

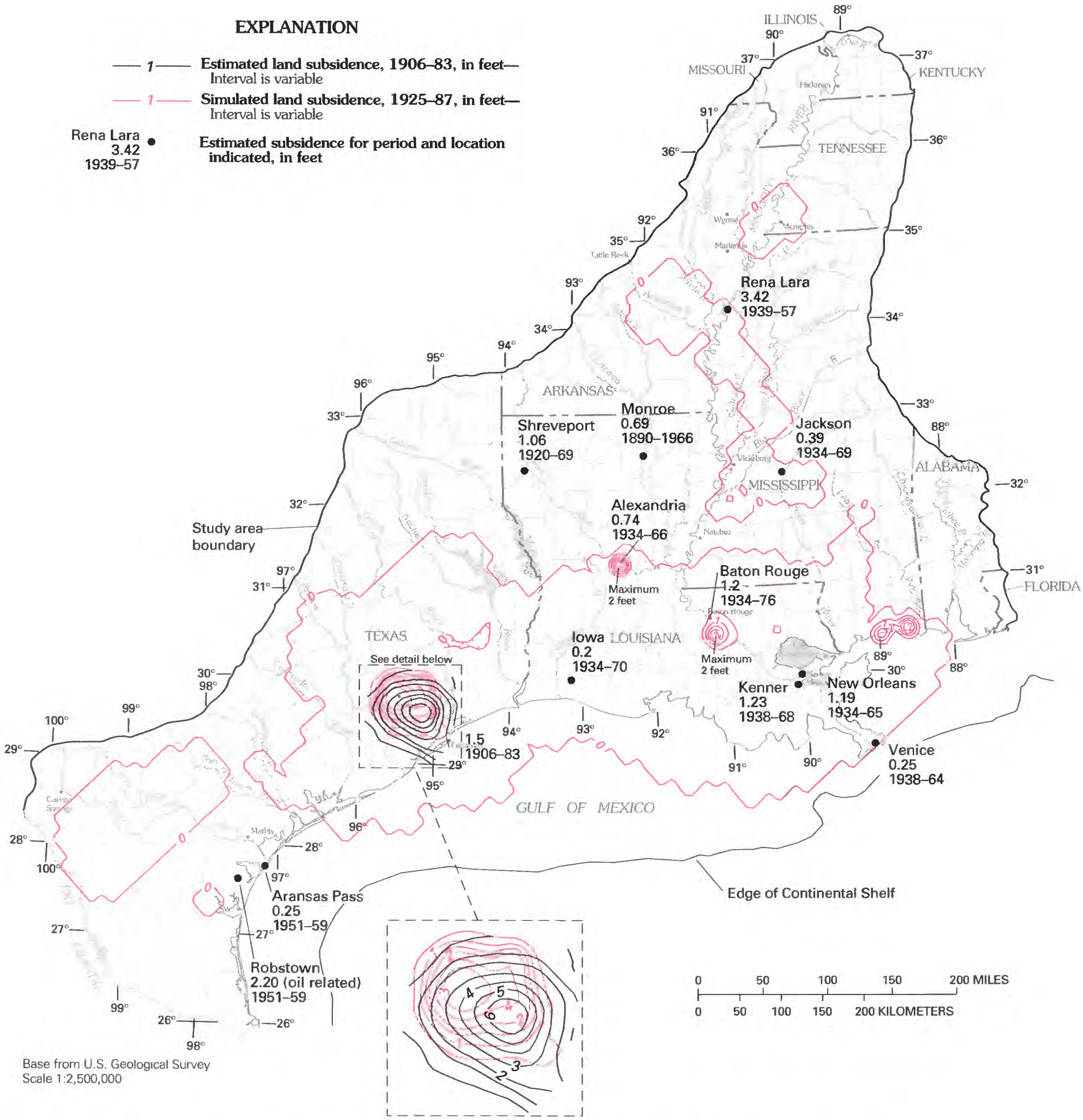


FIGURE 41.—Estimated land subsidence and simulated land subsidence due to ground-water withdrawals.

TABLE 4.—*Summary of indicators of estimated land subsidence in the study area*

[Negative numbers indicate land-surface rise; ---, no data]

Location	Source 1[a]			Source 2[b]	
	Period	Subsidence (feet)	Extent (square mile) [1]	Subsidence rate (foot per decade)	Standard deviation (foot per decade)
Alexandria, La.	1934–66	0.74	13	0.11	0.04
Aransas Pass, Tex.	1951–59	.25	12	---	---
Baton Rouge, La.[2]	1934–76	1.20	20	.16	.04
Beaumont, Tex.	---	---	---	.08	.03
Biloxi, Miss.	---	---	---	.02	.04
Galveston, Tex.[3]	1906–78	1.40	10	.16	.02
Houston, Tex.[3]	1906–78	9.00	80	.53	.02
Iowa, La.	1934–70	.20	12	---	---
Jackson, Miss.	1934–69	.39	20	---	---
Kenner, La.	1938–68	1.23	20	---	---
Meridian, Miss.	---	---	---	-.03	.04
Mobile, Ala.	---	---	---	-.06	.04
Monroe, La.	1890–1966	---	---	---	---
New Orleans, La.	1934–69	1.19	31	.25	.04
Pensacola, Fla.	---	---	---	.04	.02
Port Isabel, Tex.	---	---	---	.00	.04
Robstown, Tex.[4]	1951–59	2.20	12	---	---
Shreveport, La.	1920–69	1.06	17	.03	.05
Venice, La.	1938–64	.25	2	.32	.06

[a]From Chi and Reilinger, 1984.
[b]From Holdahl and Morrison, 1974.
[1]Approximate diameter of the subsiding area.
[2]From Whiteman, 1980, figure 1.
[3]From Gabrysch, 1982, figure 13.
[4]Related to oil withdrawals.

the brine water to reach these pumping areas because of the relatively slow ground-water velocities and the long flow paths. For example, freshwater pumping onshore in and around Houston had changed flow directions more than 45 degrees or flow rates more than 50 percent in the brine part of layer 10 or permeable zone B (lower Pleistocene-upper Pliocene deposits) throughout an area of 6,000–8,000 mi^2 that extends as far as about 60 mi offshore. Block-averaged Darcian velocities (specific discharge) in that part of layer 10 containing brine water approach 10 ft/yr at a maximum. Assuming an effective porosity of 20 percent, this would mean average particle velocities of about 50 ft/yr or about 100 yr to move 1 mi. The actual ground-water velocities could differ from these estimates by an order of magnitude or more because these estimates are averages across a 10-mi model block face. The large variability resulting from heterogeneity of the deposits allows for large differences in velocities between different pathways that water may actually take. Therefore, the velocities will generally give traveltimes slower than what actually occurs in the aquifer. However, they do indicate the possibility of long-term problems that may not be noticed for years or even decades, yet which need consideration for prudent management of the ground-water resource.

POTENTIAL AND LIMITATIONS FOR FUTURE DEVELOPMENT

Historically, when ground-water development in a particular area expands enough to cause detrimental effects, water managers usually recommend that pumping be restricted and that alternative supply sources, generally surface water, be sought (Harris-Galveston Coastal Subsidence District, written commun., 1987; Mississippi Water Resources Research Institute, 1987, p. 3). Many ground-water studies describe only the effects of existing and proposed water development. Osborne and others (1986) have shown that water-use forecasts, regardless of the method used, are likely to be highly inaccurate owing to their dependence on unpredictable economics and politics. 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 do not show where there is capacity for further development with minimal effects. Instead, studies need to contain the information water managers could use to determine what areas and permeable zones have potential to supply more water for expanding ground-water development with minimal adverse effects.

A method was developed to estimate relative potential for ground-water development using the ground-water flow model that will be described in the following section. The potential for development will then be described for each of the nine geographic areas delineated in figure 13. As each geographic area is discussed, the historical trends of the ground-water budget of the respective area will be discussed to put the potential for ground-water development in the perspective of the effects of the current development. The changes from predevelopment to 1985 in water budgets prepared for parts of the aquifer systems (fig. 13) are shown in figures 42–50. Those changes will be discussed as the potential of each geographic area is discussed in the following sections. The pumpage trends by geographic area are also described by Mesko and others (1990, p. 15–19, fig. 12, table 6). There are some general similarities among all of the trends shown in the water budgets for the different areas. The budget trend in all geographic areas shows a common pattern of generally increasing pumpage causing decreased natural regional discharge and increased net regional recharge. Confined storage, the storage shown in all budgets except the one for the Mississippi Alluvial Plain, is a small fraction of the total aquifer-system water budget. Changes in water-table storage are considered with regional recharge or discharge due to the way the flow system was conceptualized (see supplement "Simulation Details"). The only alternative is to measure hundreds or thousands of wells and study the observed changes in water levels in very shallow wells over a long period of time. The exact proportions of the simulated net recharge that actually represents change in water-table storage is probably small because in most areas, the water-table altitude has probably changed very little since predevelopment time. There are some notable exceptions to this, such as in the Mississippi River Valley alluvial aquifer, in a large area in southwestern Louisiana where rice is irrigated, and in the Winter Garden area of southern Texas.

GROUND-WATER MANAGEMENT INFORMATION USING SENSITIVITY TO PUMPING STRESS

Simulated drawdown from a uniform increase in pumpage is an integrated result of most of the hydrologic properties known to affect the hydraulic response, such as distance to recharge sources, aquifer thickness,

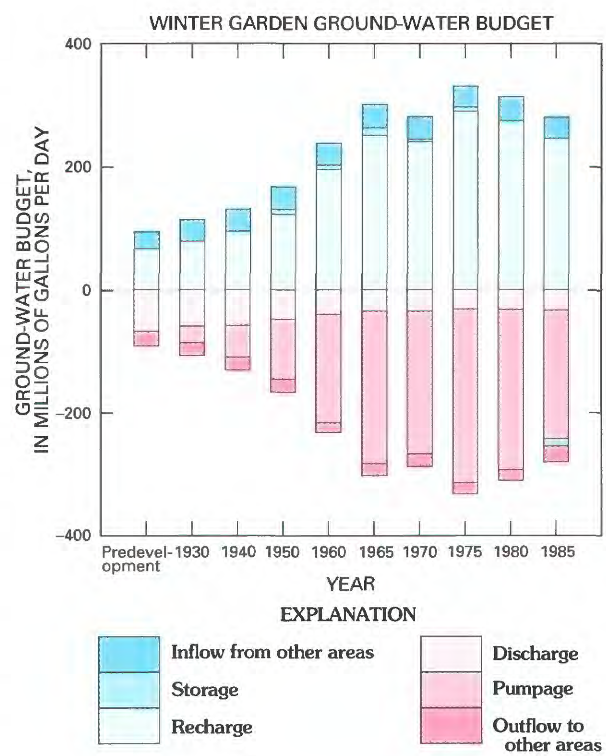


FIGURE 42.—Changes in the Winter Garden area regional ground-water budget from predevelopment to 1985.

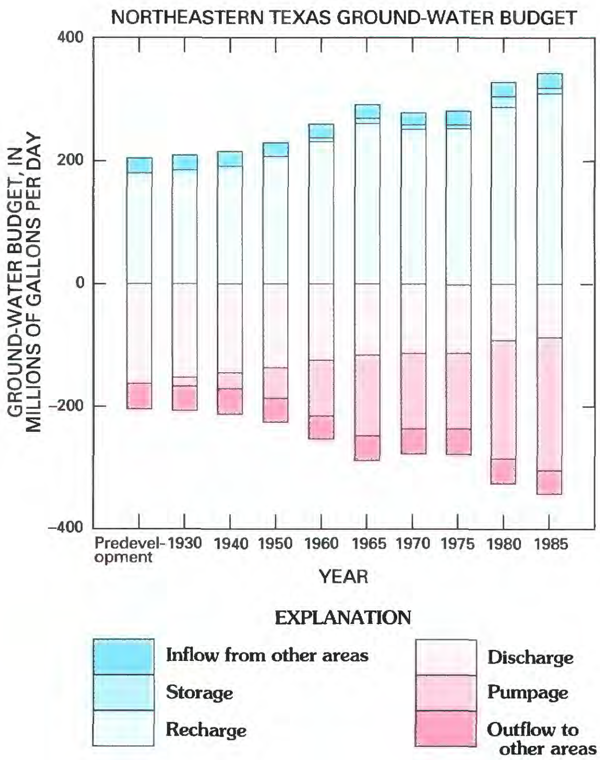


FIGURE 43.—Changes in the northeastern Texas area regional ground-water budget from predevelopment to 1985.

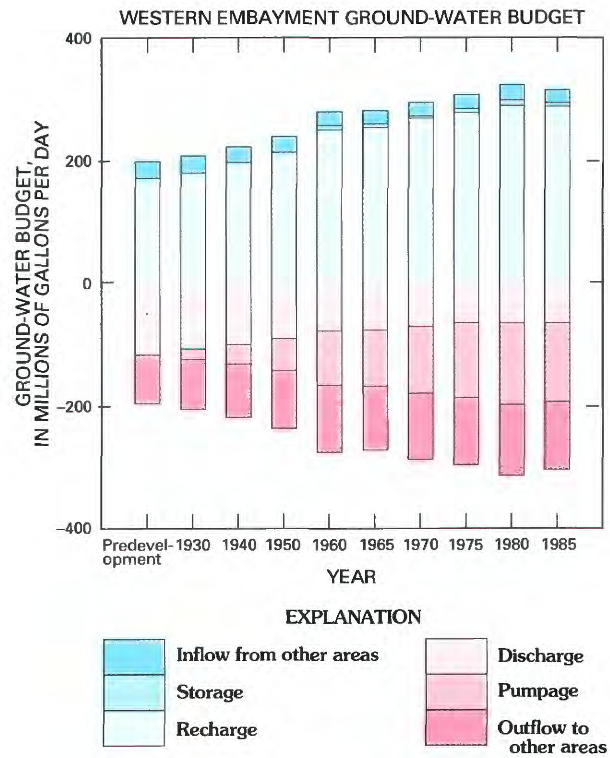


FIGURE 44.—Changes in the western embayment area regional ground-water budget from predevelopment to 1985.

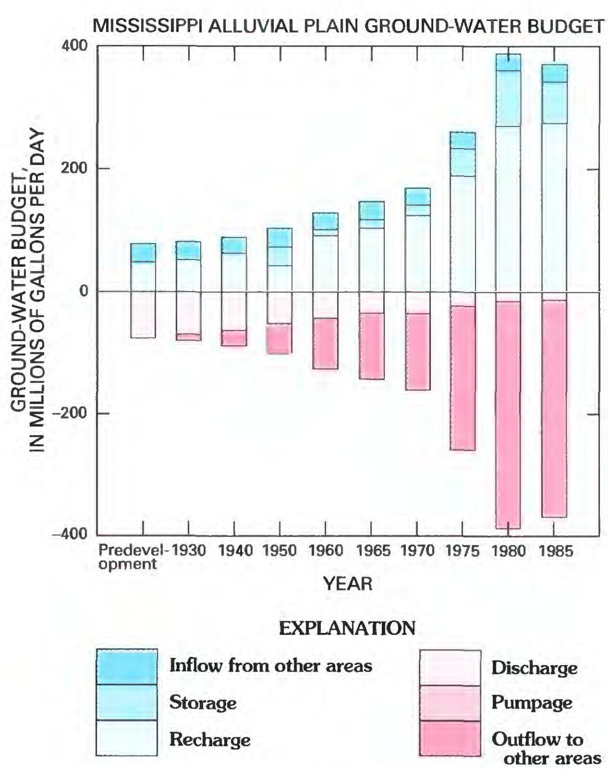


FIGURE 45.—Changes in the Mississippi Alluvial Plain area regional ground-water budget from predevelopment to 1985.

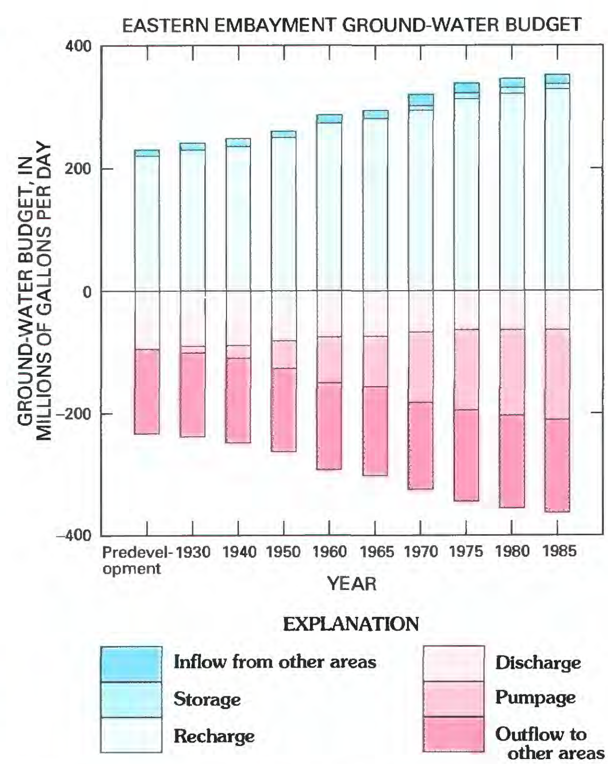


FIGURE 46.—Changes in the eastern embayment area regional ground-water budget from predevelopment to 1985.

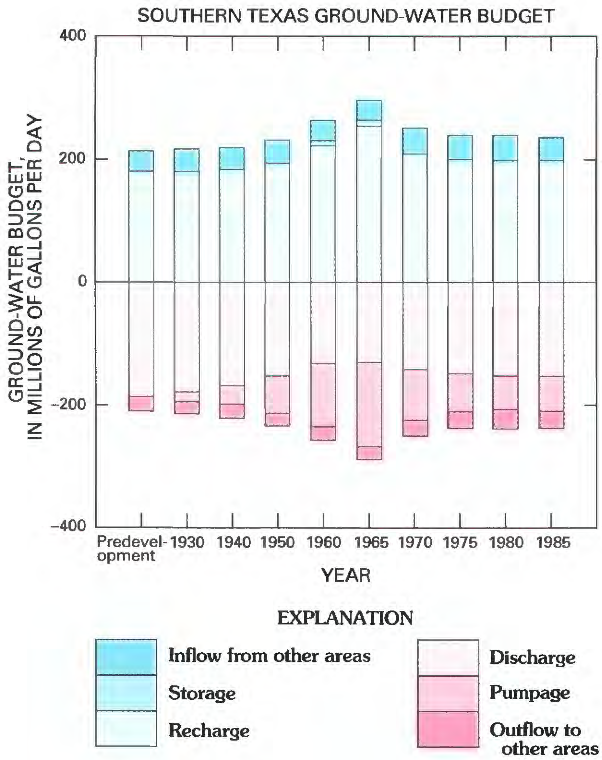


FIGURE 47.—Changes in the southern Texas area regional ground-water budget from predevelopment to 1985.

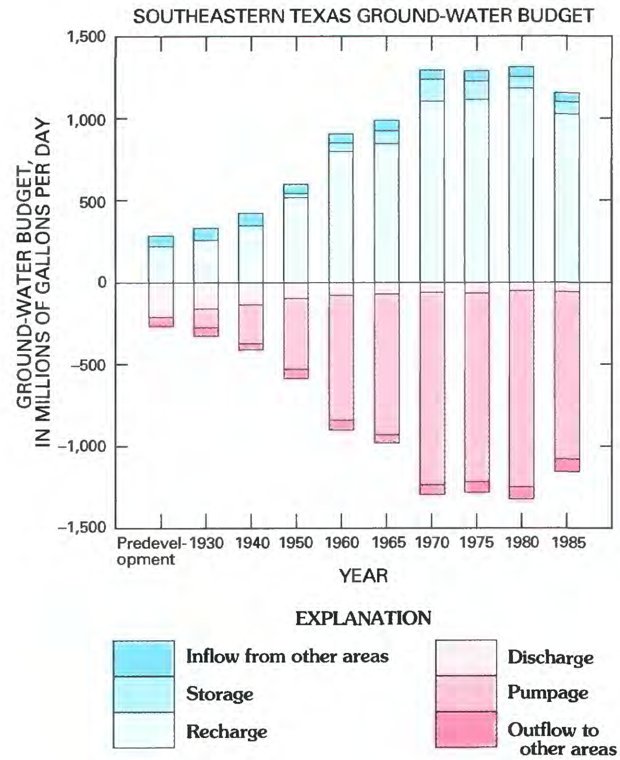


FIGURE 48.—Changes in the southeastern Texas area regional ground-water budget from predevelopment to 1985.

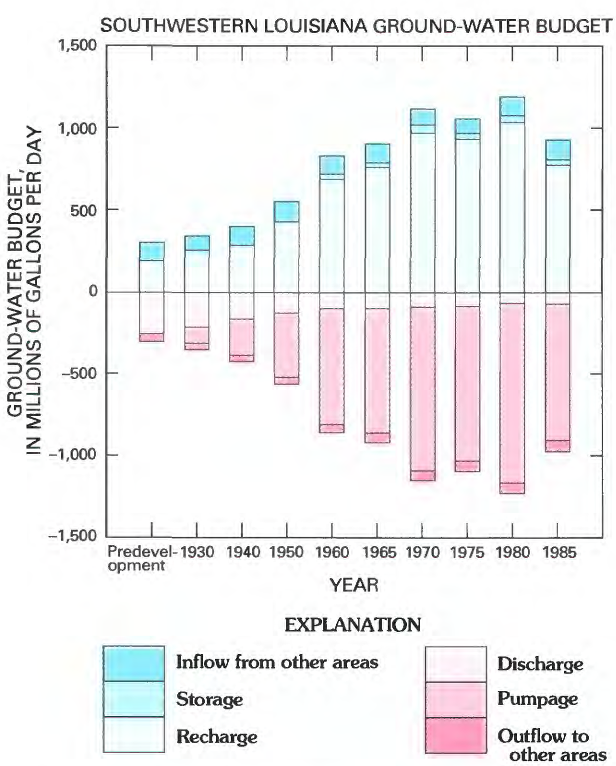


FIGURE 49.—Changes in the southwestern Louisiana area regional ground-water budget from predevelopment to 1985.

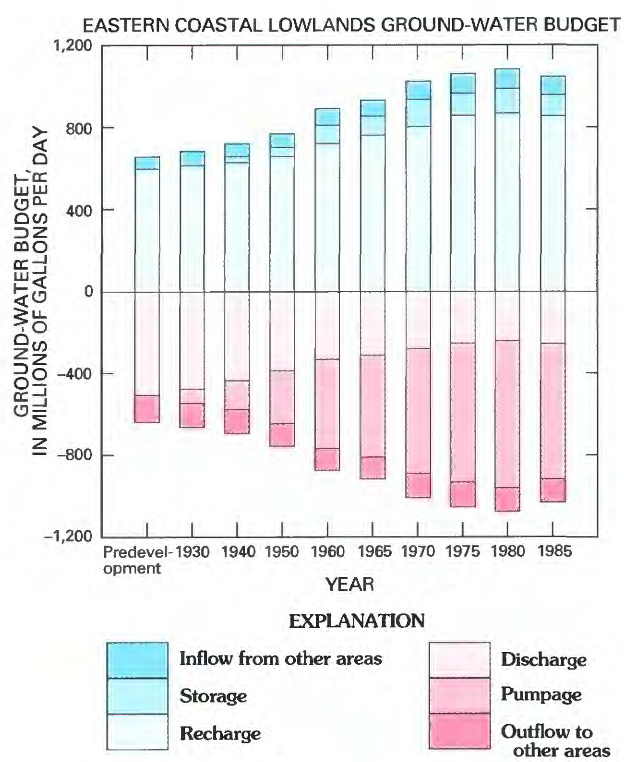


FIGURE 50.—Changes in the eastern coastal lowlands area regional ground-water budget from predevelopment to 1985.

sand percentage, hydraulic conductivity, vertical leakage, and aquifer storage. Therefore, the model was used to predict the relative potential for increased ground-water pumpage by doing an areal analysis of sensitivity to pumpage by layer. The aquifer system's response to a unit increase in pumpage was simulated for each model block containing relatively fresh water. The simulations were done with two slightly different approaches, both described in the supplement "Simulation Details." The product is a map for each layer (fig. 51) showing the potential for future ground-water development. The five levels of potential, as shown in the following table represent the volume of drawdown in each block that would result from an increase in pumpage of 1 Mgal/d per 100-mi^2 block.

Level	Potential for development	Drawdown (feet)
5	Excellent	0 to 2
4	Good	2 to 5
3	Fair	50 to 10
2	Poor	10 to 25
1	Virtually none	More than 25

These simulations, which ignore aquifer storage and compute maximum steady-state drawdown, will indicate what the maximum future effects might be. These maps will therefore allow a water manager to identify the layers and(or) geographic areas in which additional pumpage may have relatively greater effect and geographic areas or layers where additional ground-water development may have relatively less effect.

Every 10-mi model block with estimated concentration of dissolved solids less than 10,000 mg/L and a thickness greater than 25 ft was stressed by simulation of an additional 1 Mgal/d. There are 7,324 model blocks that contain ground water with a concentration of dissolved solids less than 10,000 mg/L. This dissolved-solids concentration was used because it was convenient and generally was identified within a few miles of dissolved-solids concentrations of 3,000 mg/L, a concentration value commonly used to delineate "usable water" in the western part of the study area. At each model grid location, equal additional pumping in each block of each layer with relatively fresh water is simulated independently (discussed in the supplement "Simulation Details").

A composite map (fig. 52) was made showing the layer at each model grid location that had the least sensitivity to pumpage (least drawdown) from the data shown in figure 51. Theoretically, this would be the layer for water managers to consider first when additional ground-water supplies are needed because drawdown would be minimized for a given amount of pumpage if the pumpage came from this layer. Several practical aspects limit the application of this information. In some cases it may be more expensive to develop wells in the layer shown rather than in a shallower layer. However, developing a deeper well field might still be much less costly than developing an alternative source of water. The reason well installation costs could be greater is related to the scale of this analysis. This analysis considers only the regional perspective of large-scale development of aquifer systems. A permeable zone that appears the best from this analysis, such as layer 6 in parts of the Mississippi embayment aquifer system, may have only thin sand beds in which to place well screens, which might limit the well's yield. However, because of some other characteristic of that permeable zone at that location, such as proximity to a recharge source, the layer might still have better potential for ground-water development if the water could be withdrawn using a larger number of smaller capacity wells than might be needed in some other layer with thicker sand beds.

Water quality and land subsidence are factors to consider along with potential for ground-water development. Attempting to minimize drawdown for a given level of pumpage by locating well fields in a new area or layer may increase the total contributing area (and hence spread the effects of drawdown). This has a disadvantage when trying to protect ground water in the areas contributing to well fields from becoming contaminated. There is a corresponding advantage; with a larger contributing area, there is a greater chance that any water-quality degradation will be more diluted by freshwater coming from the remaining contributing area. In areas where land subsidence has occurred, it may recur in the future if drawdown is greater than it was previously. In other parts of the coastal lowlands aquifer system, land subsidence may occur if drawdown from predevelopment exceeds a critical value. In calibrating the model, the average critical value was found to be about 80 ft, but that value is approximate and could vary spatially.

WINTER GARDEN AREA

In the Winter Garden area, the peak ground-water pumpage occurred in 1975 (which was only slightly more than in 1965) and then decreased slightly thereafter (fig. 42). Since pumpage began, there has been slightly more inflow from other areas than outflow to other areas. Natural regional discharge decreased to less than one-half of its predevelopment value. Regional recharge increased by about a factor of four. In

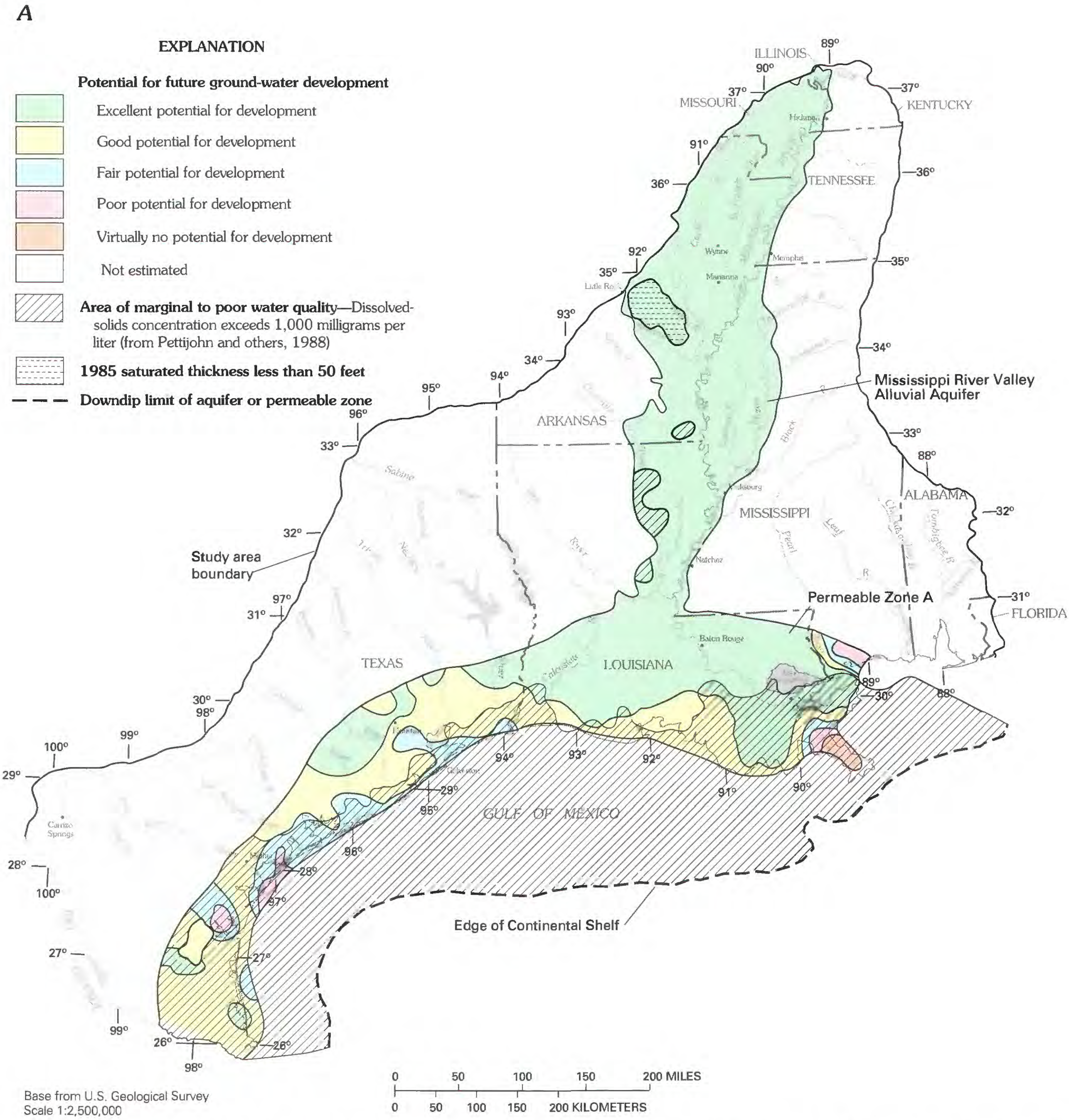


FIGURE 51.—Potential for additional development of the ground-water resource: *A*. permeable zone A and Mississippi River Valley alluvial aquifer (layer 11); *B*. permeable zone B (layer 10); *C*. permeable zone C (layer 9); *D*. permeable zone D (layer 8); *E*. permeable zone E (layer 7); *F*. upper Claiborne aquifer; (layer 6); *G*. middle Claiborne aquifer (layer 5); *H*. lower Claiborne-upper Wilcox aquifer (layer 4); *I*. middle Wilcox aquifer (layer 3); *J*. lower Wilcox aquifer (layer 2).

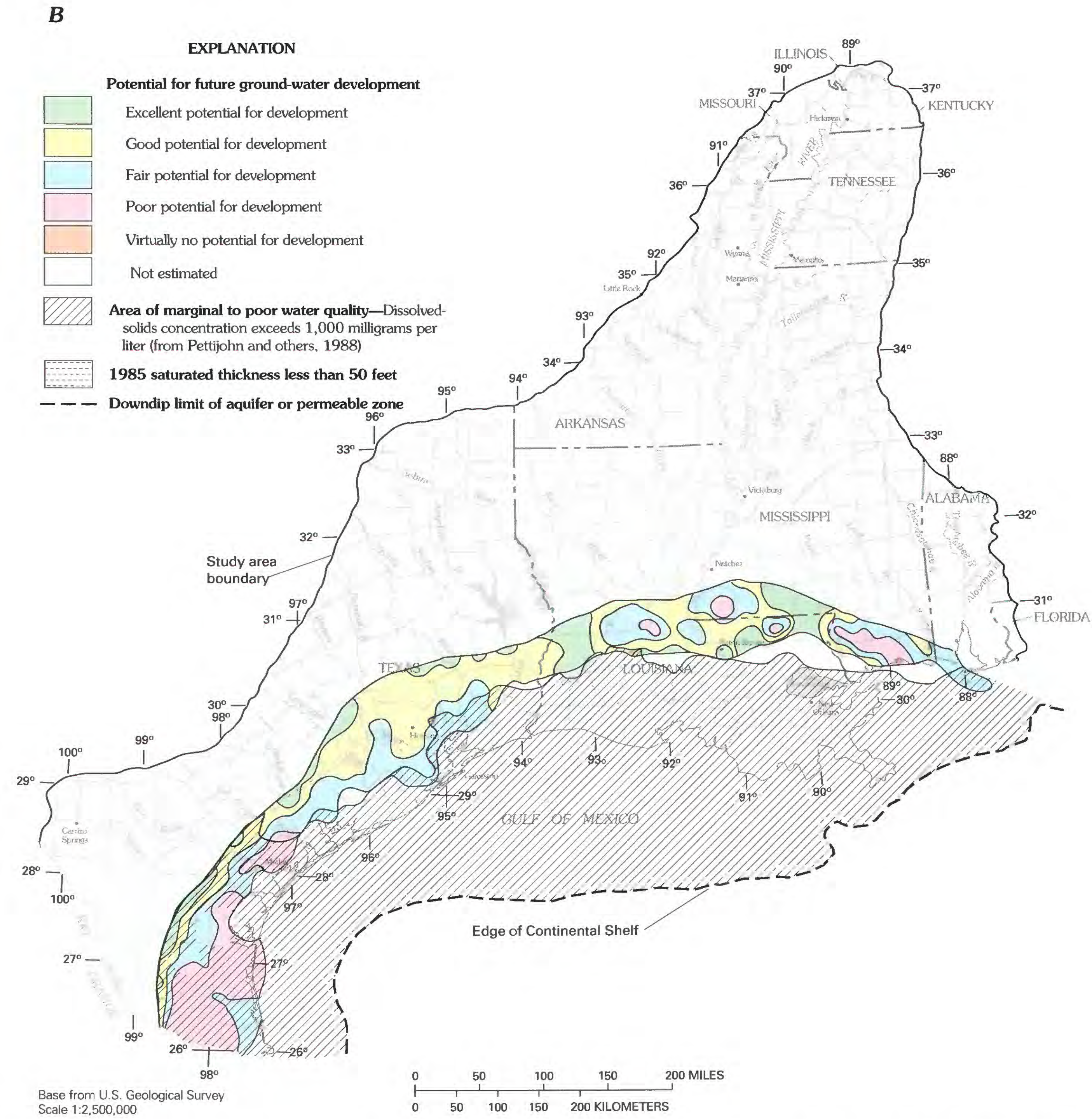


FIGURE 51.—Potential for additional development of the ground-water resource: *A*. permeable zone A and Mississippi River Valley alluvial aquifer (layer 11); *B*. permeable zone B (layer 10); *C*. permeable zone C (layer 9); *D*. permeable zone D (layer 8); *E*. permeable zone E (layer 7); *F*. upper Claiborne aquifer; (layer 6); *G*. middle Claiborne aquifer (layer 5); *H*. lower Claiborne-upper Wilcox aquifer (layer 4); *I*. middle Wilcox aquifer (layer 3); *J*. lower Wilcox aquifer (layer 2)—Continued.

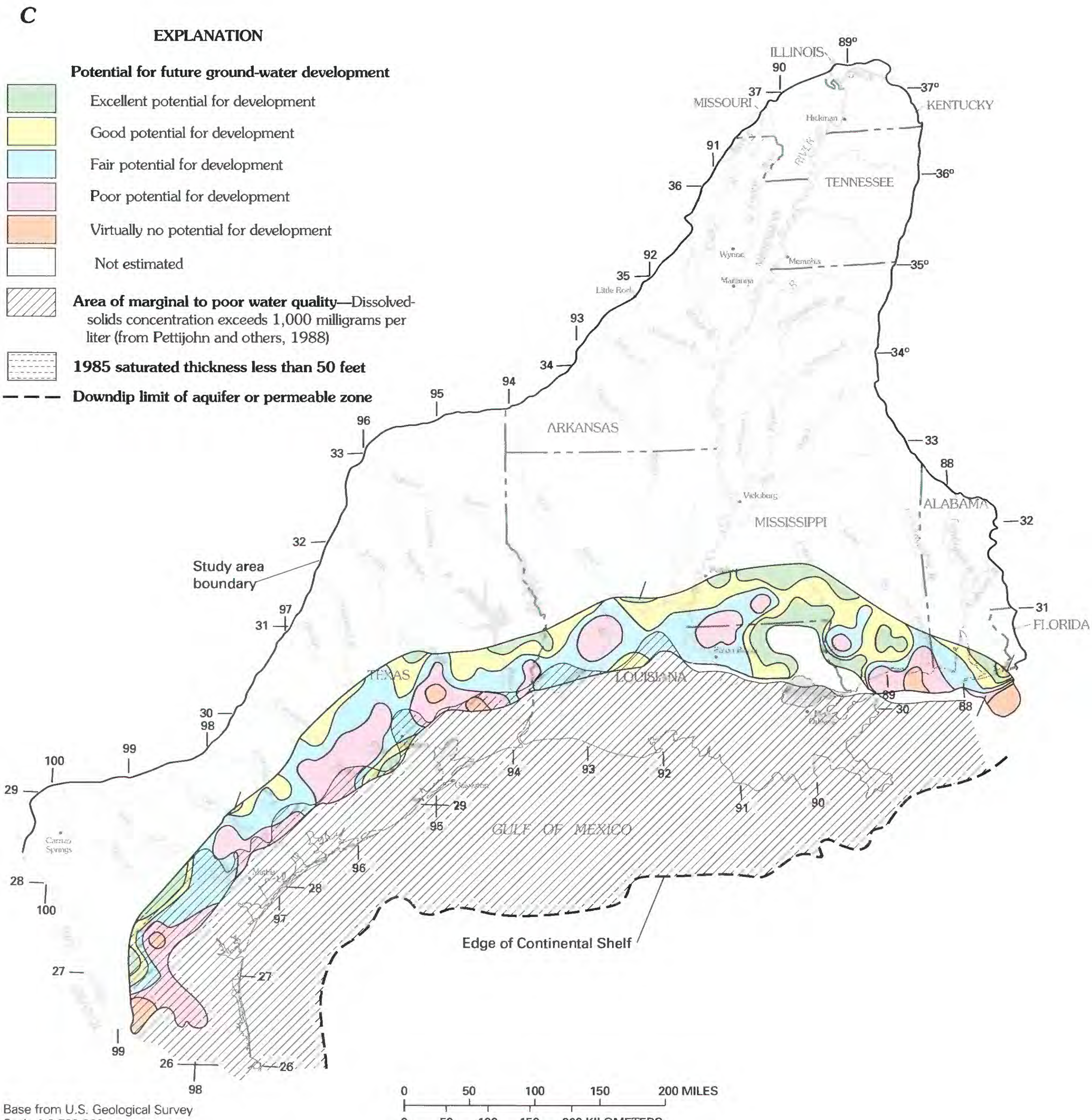


FIGURE 51.—Potential for additional development of the ground-water resource: *A*. permeable zone A and Mississippi River Valley alluvial aquifer (layer 11); *B*. permeable zone B (layer 10); *C*. permeable zone C (layer 9); *D*. permeable zone D (layer 8); *E*. permeable zone E (layer 7); *F*. upper Claiborne aquifer; (layer 6); *G*. middle Claiborne aquifer (layer 5); *H*. lower Claiborne-upper Wilcox aquifer (layer 4); *I*. middle Wilcox aquifer (layer 3); *J*. lower Wilcox aquifer (layer 2)—Continued.

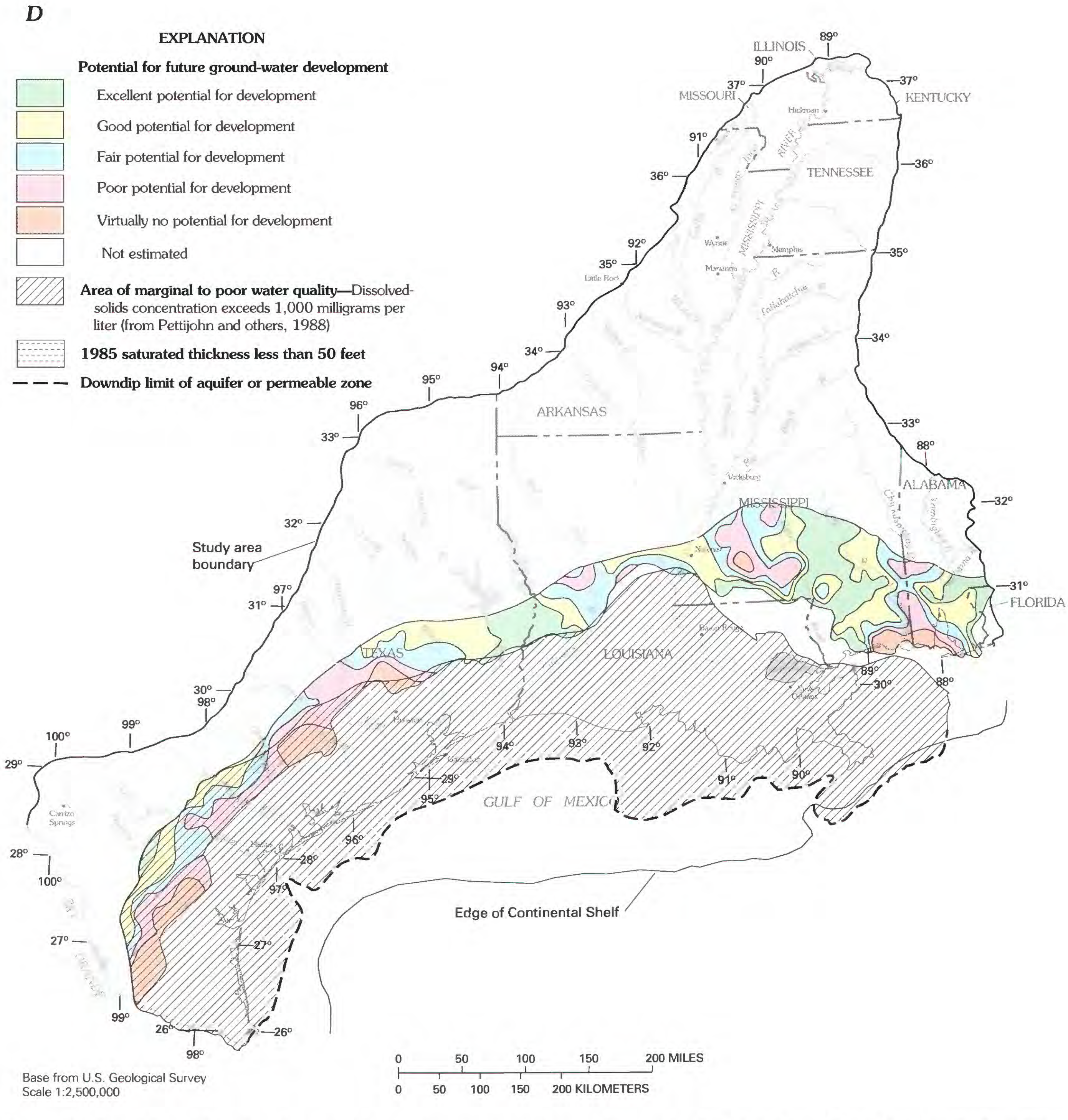


FIGURE 51.—Potential for additional development of the ground-water resource: *A*. permeable zone A and Mississippi River Valley alluvial aquifer (layer 11); *B*. permeable zone B (layer 10); *C*. permeable zone C (layer 9); *D*. permeable zone D (layer 8); *E*. permeable zone E (layer 7); *F*. upper Claiborne aquifer; (layer 6); *G*. middle Claiborne aquifer (layer 5); *H*. lower Claiborne-upper Wilcox aquifer (layer 4); *I*. middle Wilcox aquifer (layer 3); *J*. lower Wilcox aquifer (layer 2)—Continued.

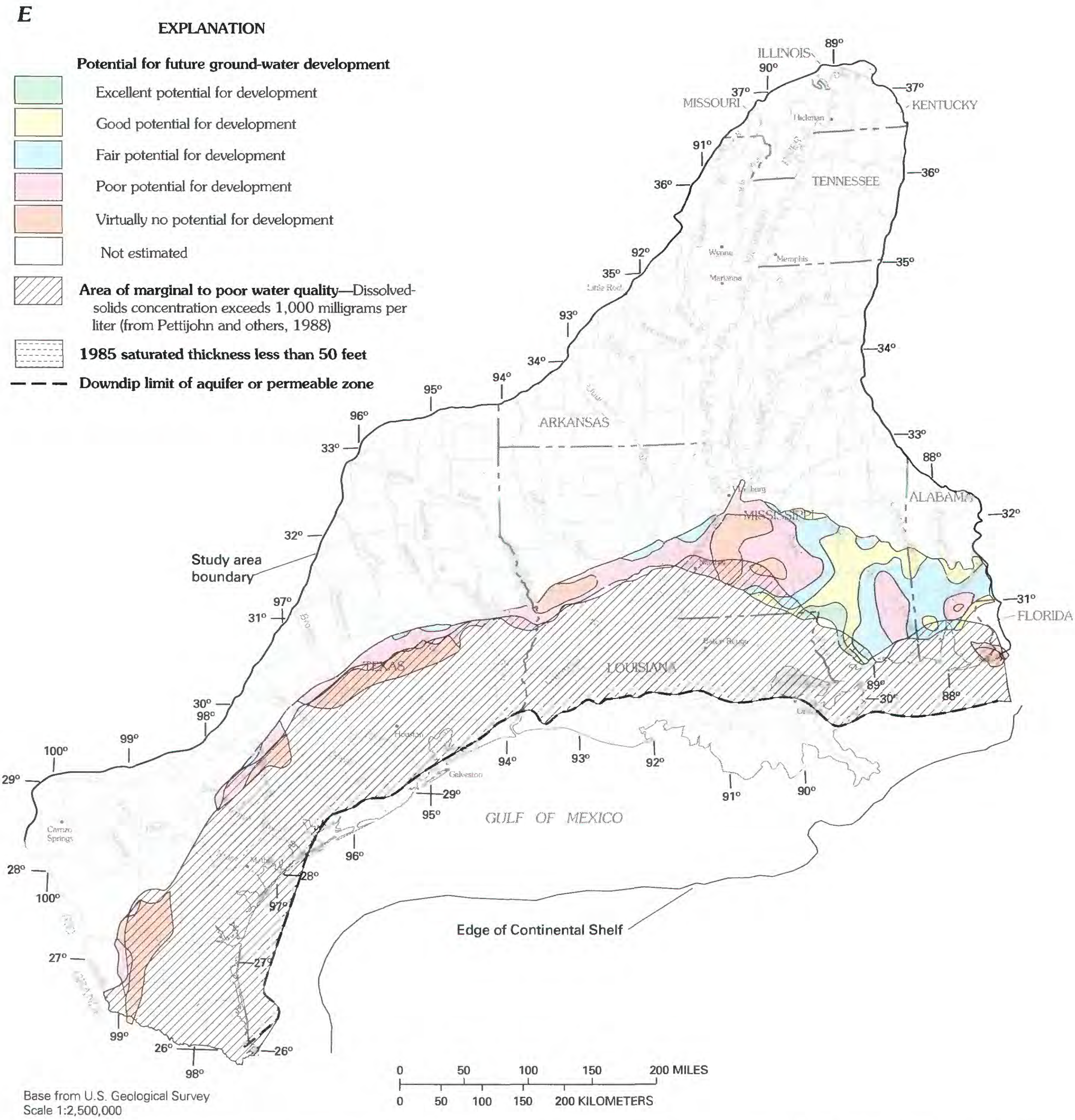


FIGURE 51.—Potential for additional development of the ground-water resource: *A*. permeable zone A and Mississippi River Valley alluvial aquifer (layer 11); *B*. permeable zone B (layer 10); *C*. permeable zone C (layer 9); *D*. permeable zone D (layer 8); *E*. permeable zone E (layer 7); *F*. upper Claiborne aquifer; (layer 6); *G*. middle Claiborne aquifer (layer 5); *H*. lower Claiborne-upper Wilcox aquifer (layer 4); *I*. middle Wilcox aquifer (layer 3); *J*. lower Wilcox aquifer (layer 2)—Continued.

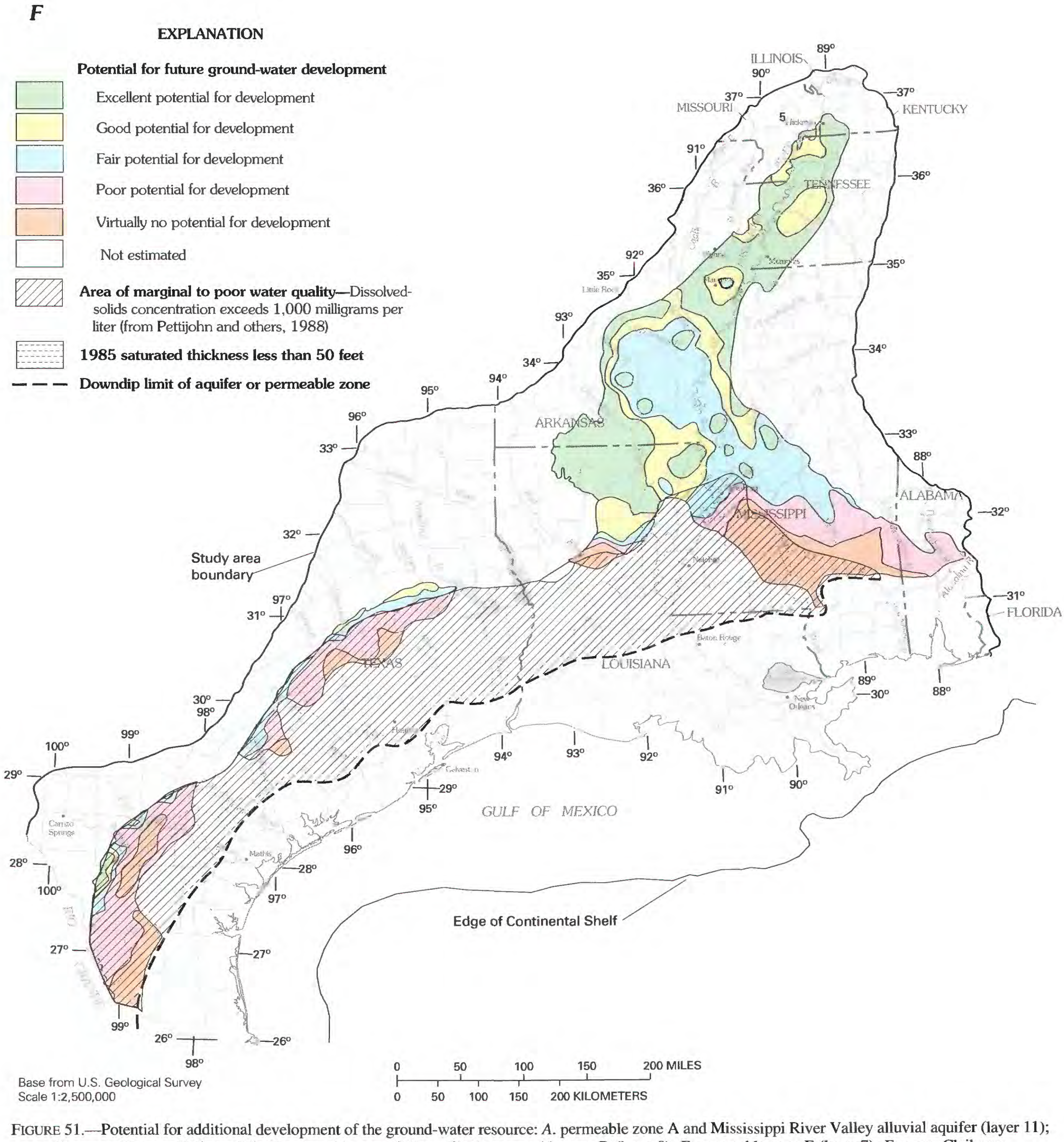


FIGURE 51.—Potential for additional development of the ground-water resource: *A*. permeable zone A and Mississippi River Valley alluvial aquifer (layer 11); *B*. permeable zone B (layer 10); *C*. permeable zone C (layer 9); *D*. permeable zone D (layer 8); *E*. permeable zone E (layer 7); *F*. upper Claiborne aquifer; (layer 6); *G*. middle Claiborne aquifer (layer 5); *H*. lower Claiborne-upper Wilcox aquifer (layer 4); *I*. middle Wilcox aquifer (layer 3); *J*. lower Wilcox aquifer (layer 2)—Continued.

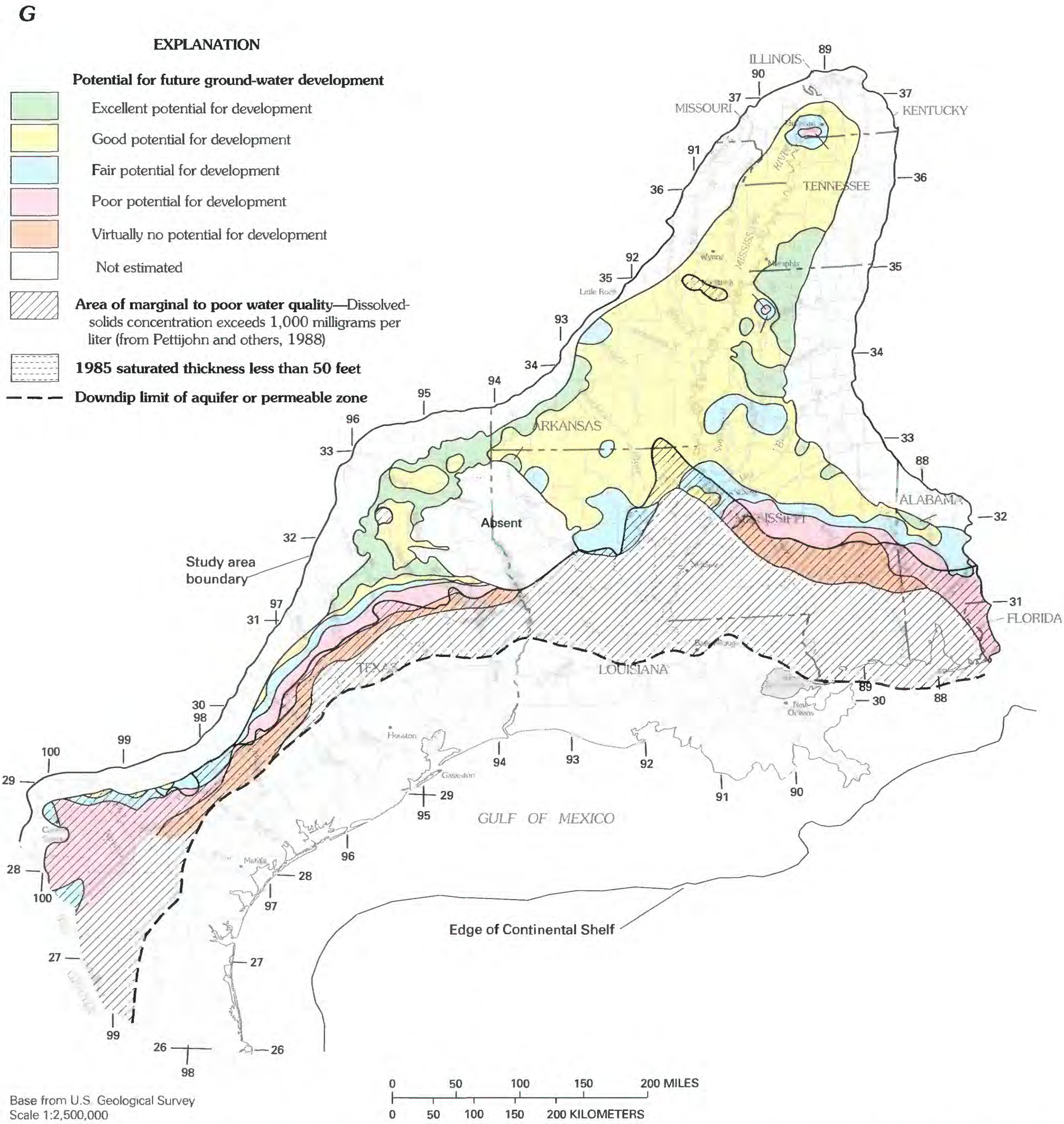


FIGURE 51.—Potential for additional development of the ground-water resource: *A*. permeable zone A and Mississippi River Valley alluvial aquifer (layer 11); *B*. permeable zone B (layer 10); *C*. permeable zone C (layer 9); *D*. permeable zone D (layer 8); *E*. permeable zone E (layer 7); *F*. upper Claiborne aquifer; (layer 6); *G*. middle Claiborne aquifer (layer 5); *H*. lower Claiborne-upper Wilcox aquifer (layer 4); *I*. middle Wilcox aquifer (layer 3); *J*. lower Wilcox aquifer (layer 2)—Continued.

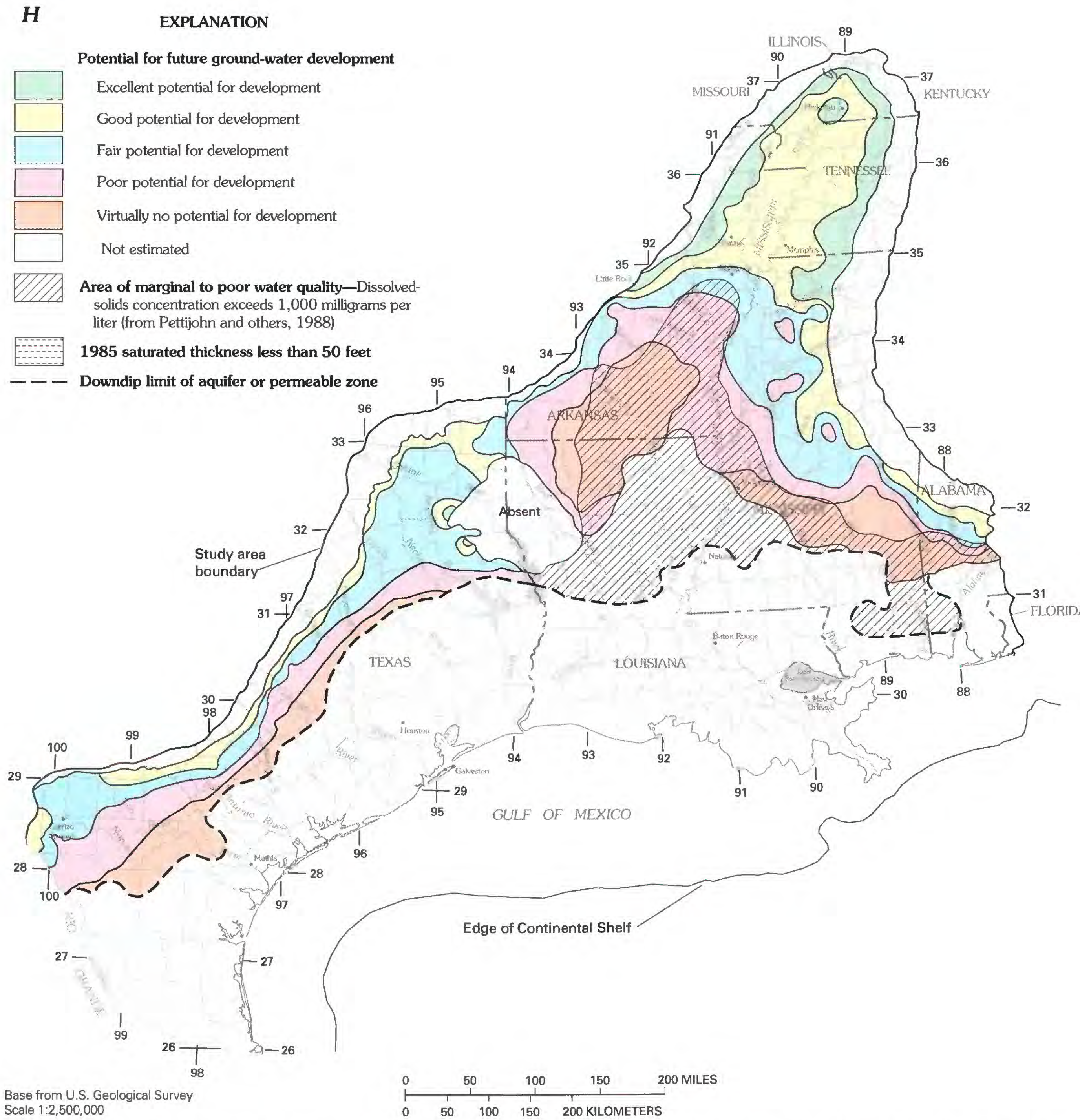


FIGURE 51.—Potential for additional development of the ground-water resource: *A*. permeable zone A and Mississippi River Valley alluvial aquifer (layer 11); *B*. permeable zone B (layer 10); *C*. permeable zone C (layer 9); *D*. permeable zone D (layer 8); *E*. permeable zone E (layer 7); *F*. upper Claiborne aquifer; (layer 6); *G*. middle Claiborne aquifer (layer 5); *H*. lower Claiborne-upper Wilcox aquifer (layer 4); *I*. middle Wilcox aquifer (layer 3); *J*. lower Wilcox aquifer (layer 2)—Continued.

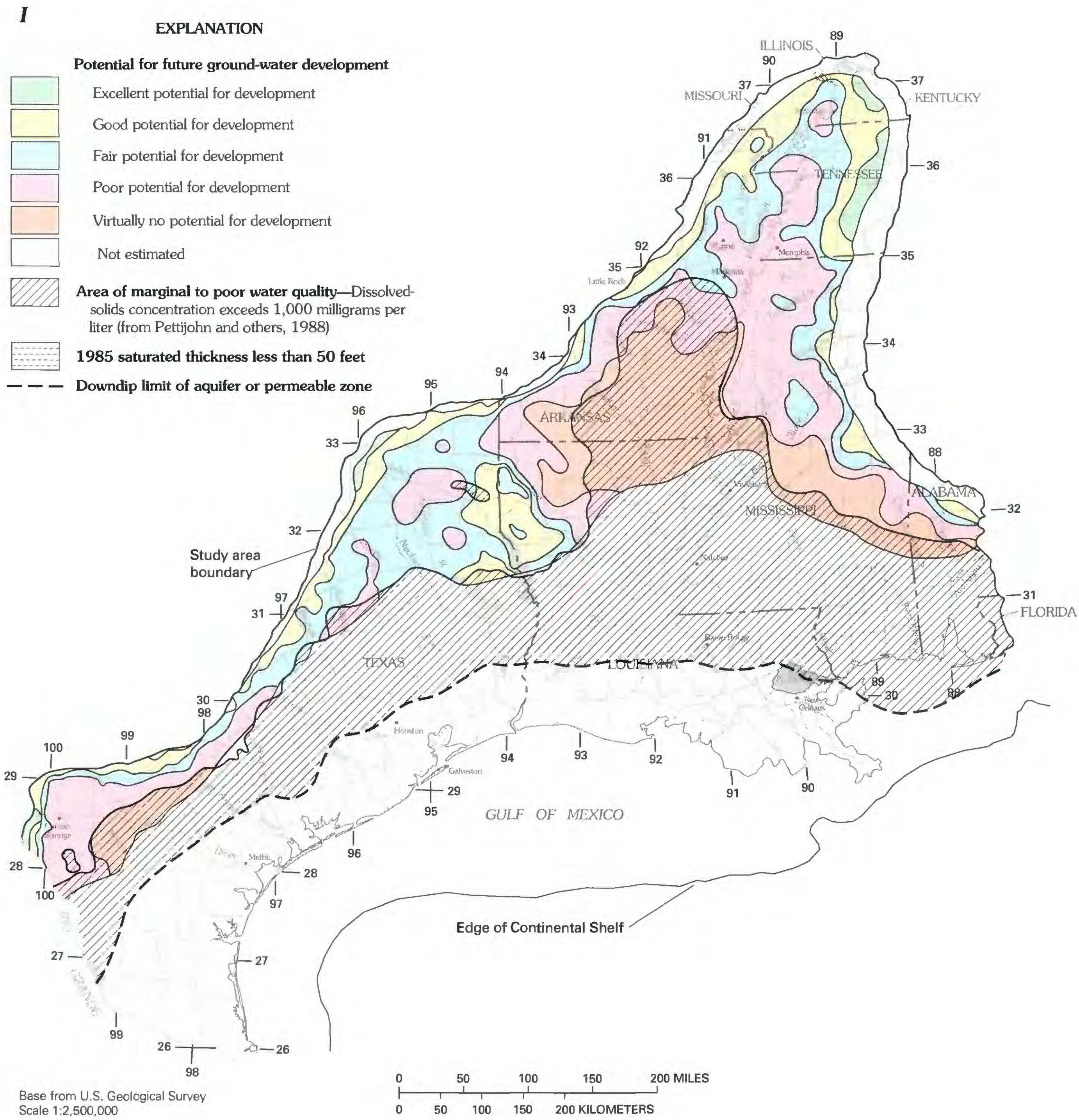


FIGURE 51.—Potential for additional development of the ground-water resource: *A*. permeable zone A and Mississippi River Valley alluvial aquifer (layer 11); *B*. permeable zone B (layer 10); *C*. permeable zone C (layer 9); *D*. permeable zone D (layer 8); *E*. permeable zone E (layer 7); *F*. upper Claiborne aquifer; (layer 6); *G*. middle Claiborne aquifer (layer 5); *H*. lower Claiborne-upper Wilcox aquifer (layer 4); *I*. middle Wilcox aquifer (layer 3); *J*. lower Wilcox aquifer (layer 2)—Continued.

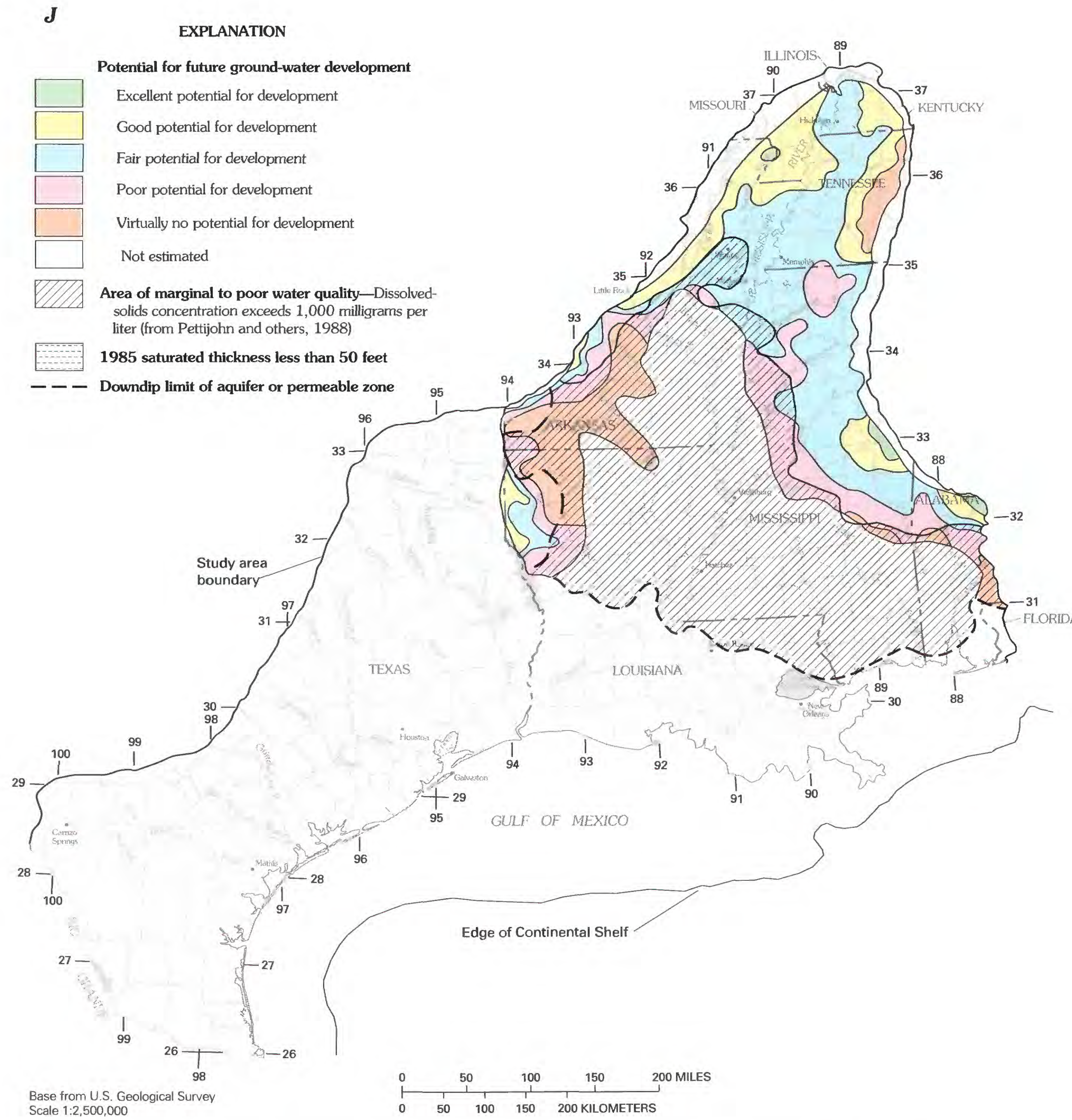


FIGURE 51.—Potential for additional development of the ground-water resource: *A*. permeable zone A and Mississippi River Valley alluvial aquifer (layer 11); *B*. permeable zone B (layer 10); *C*. permeable zone C (layer 9); *D*. permeable zone D (layer 8); *E*. permeable zone E (layer 7); *F*. upper Claiborne aquifer; (layer 6); *G*. middle Claiborne aquifer (layer 5); *H*. lower Claiborne-upper Wilcox aquifer (layer 4); *I*. middle Wilcox aquifer (layer 3); *J*. lower Wilcox aquifer (layer 2)—Continued.

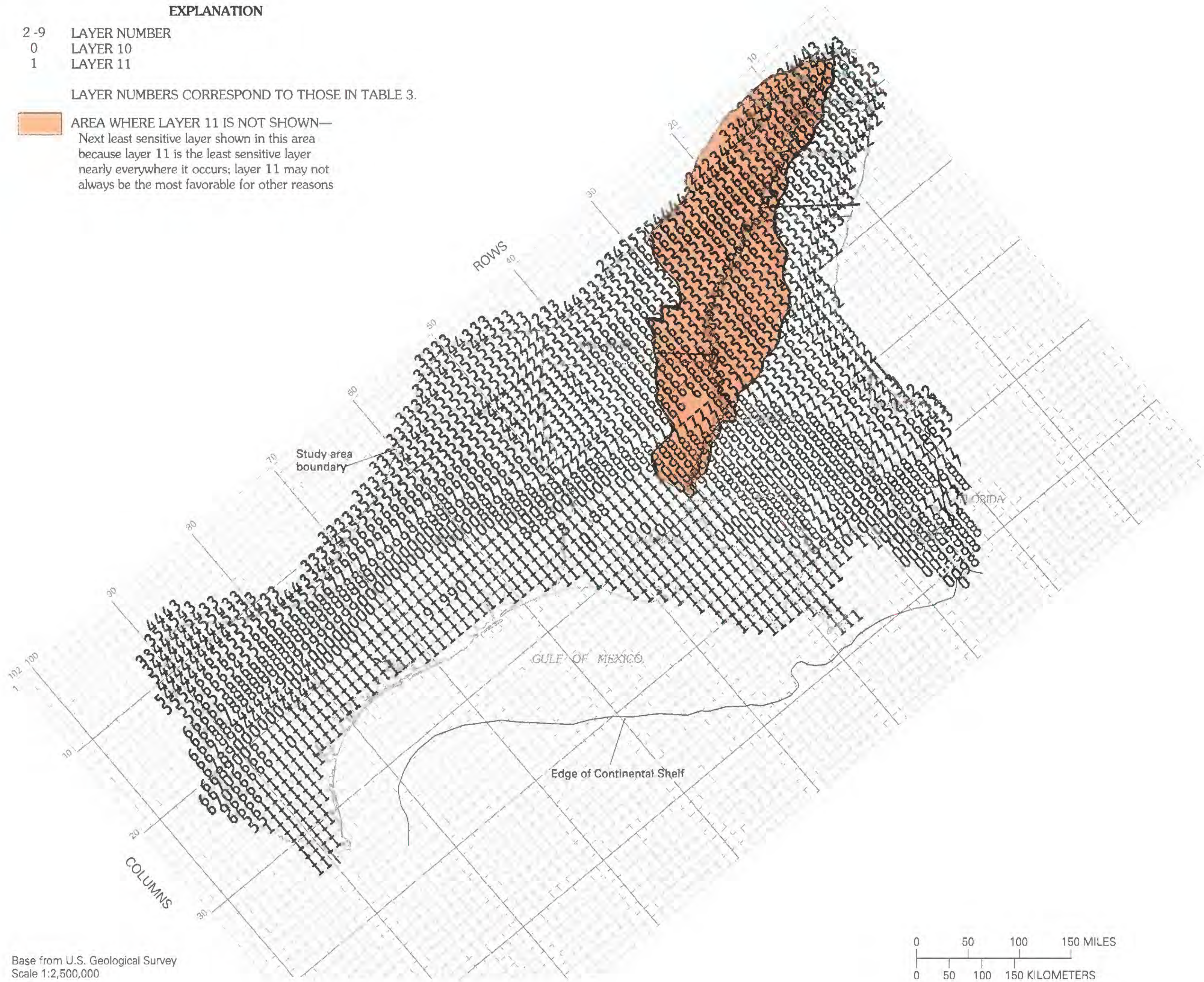


FIGURE 52.—Areal locations of layers that are least sensitive to drawdown from increased pumpage.

fact, the maximum pumpage exceeded predevelopment regional recharge by more than a factor of four. Contribution from confined storage reached a peak in 1965 and was 6 percent of the pumpage. This geographic area has had large water-table declines (Klemt and others, 1976) and, therefore, changes in water-table storage, which were not simulated separately in the regional model. Ryder and Ardis (in press) reported results from their subregional simulation with a 5-mi grid spacing, indicating that 49 percent of the 1980 pumpage in the lower Claiborne-upper Wilcox aquifer (layer 4) was supplied from water-table and confined storage. That amount, 121 Mgal/d, equals about 46 percent of the total pumpage in the Winter Garden area during 1980. In 1985, confined storage was being replenished as water levels recovered.

In the Winter Garden area of southern Texas, the lower Claiborne-upper Wilcox aquifer has the most potential for ground-water development (fig. 52). Not coincidentally, this also is the aquifer that has been most heavily developed to date, with large water-level drawdowns. The middle Claiborne (layer 5) and upper Claiborne (layer 6) aquifers, in areas closer to the coast and toward the Mexico border, have little potential for development because of their proximity to saline water (fig. 52, figs. 51*F*, 51*G*). The middle Wilcox aquifer (layer 3), at locations farther inland, has good and fair potential for development (fig. 51*C*).

NORTHEASTERN TEXAS AREA

In the northeastern Texas area, ground-water pumpage has not had as large an effect on the ground-water budget as in the Winter Garden area. Regional ground-water recharge has increased by two-thirds from predevelopment, while natural regional discharge has decreased by almost one-half (fig. 43). Pumpage peaked in 1985, whereas the maximum change in confined storage was in 1980 and was about 9 percent of the pumpage. Both inflow and outflow to other geographic areas were steady with outflow exceeding inflow by about a factor of two. The 1985 ground-water pumpage was larger than the value of predevelopment regional recharge in this area.

Throughout a large part of the northeastern Texas area, the middle Claiborne aquifer (layer 5) has the best relative potential for ground-water development (fig. 52). The relative potential for development in this aquifer is good to excellent (fig. 51*G*). Large areas underlain by the lower Claiborne-upper Wilcox aquifers and the middle Wilcox aquifer (layers 4, 3) have a relative potential for ground-water development that ranges from fair to good (fig. 51*H*, 51*I*). The good potential in the middle Wilcox aquifer (layer 3) occurs where the aquifer is thickest (fig. 12*A*).

WESTERN EMBAYMENT AREA

Ground-water pumpage in the western embayment area has caused the regional recharge to increase by more than one-half (fig. 44). Pumpage has captured about one-half of the natural regional ground-water discharge. Ground-water pumpage in 1985 was more than three-fourths of the volume of predevelopment regional recharge in the area. Although peak ground-water pumpage in the western embayment occurred in 1985, the peak change in confined storage came in 1980 and was about 8 percent of the pumpage. Outflow from the area (mainly to the Mississippi Alluvial Plain area) exceeds inflow by a factor of four or five, and both outflow and inflow have changed slowly over time.

Simulations indicate that the upper Claiborne aquifer (layer 6, fig. 52) has the most potential for ground-water development in the western embayment area. Towards the eastern side of this area, it has excellent to good relative potential for development (fig. 51*F*). In south-central Arkansas, the potential for development in this aquifer is diminished substantially by the presence of an isolated part of the Vicksburg-Jackson confining unit (fig. 11), which decreases the vertical connection to recharge sources. The middle Claiborne aquifer (layer 5) has good relative potential for ground-water development in much of the western embayment area (fig. 51*G*). The middle Wilcox aquifer (layer 3) has good relative potential for development in the southwestern part of this area (fig. 51*I*).

MISSISSIPPI ALLUVIAL PLAIN AREA

Much ground-water pumpage in the Mississippi Alluvial Plain area has caused the regional recharge to increase by more than a factor of five (fig. 45). Pumpage also has captured about 80 percent of the natural regional ground-water discharge. The maximum ground-water pumpage is more than seven times larger than the predevelopment regional recharge rate. The peak ground-water pumpage and change in ground-water storage was in 1980, when the decrease in confined and water-table storage was about 25 percent of the pumpage (Ackerman, 1996, fig. 24). Inflow from adjacent areas (mainly from the eastern embayment area) is much larger than outflow to adjacent areas and has changed slowly over time. The water-budget values shown in figure 45 were calculated by taking the regional model-simulation results and adding water-table storage values from the subregional model (Ackerman, 1996) to simulated storage while subtracting the same values from simulated recharge.

Simulations indicate that the aquifer with the most potential for ground-water development in the Mississippi Alluvial Plain area is the Mississippi River Valley alluvial aquifer (layer 11), followed by the middle and upper Claiborne aquifers (layers 5, 6) (fig. 52). The Mississippi River Valley alluvial aquifer has excellent potential for development throughout much of the area, probably because of its proximity to surface-water sources of recharge and also because of its position as the natural regional discharge of most of the underlying aquifers. The major limitations to further development of this aquifer are its current level of drawdown, which is large in several places (pl. 7*A*), and marginally poor quality of water at some locations (fig. 51*A*) (Pettijohn and others, 1988). The upper Claiborne aquifer (layer 6) has several areas with excellent and good potential for development (fig. 51*F*). The middle Claiborne aquifer (layer 5) has good potential for ground-water development throughout most of the Mississippi Alluvial Plain area (fig. 51*G*). This includes the northern part of the embayment where the Memphis aquifer includes layer 4 and layer 5. The middle Wilcox aquifer (layer 3) has some areas of poor and fair potential with smaller areas of good and even excellent potential for ground-water development (fig.51*I*). The lower Wilcox aquifer (layer 2) has large areas with fair and some areas with good potential for ground-water development (fig. 51*J*).

EASTERN EMBAYMENT AREA

Ground-water pumpage in the eastern embayment area has caused the regional recharge to increase by about one-half the predevelopment recharge rate (fig. 46) while capturing nearly one-half of the natural regional ground-water discharge. The 1985 ground-water pumpage is about two-thirds of the volume of predevelopment regional recharge. The peak ground-water pumpage and change in confined storage (to date) came in 1985 and was about 7 percent of the pumpage. Outflow from the area (mainly to the Mississippi Alluvial Plain area) exceeds inflow by more than 10 times, and both outflow and inflow have changed slowly with time.

Simulations indicate that each of the aquifers in the eastern embayment area have good to excellent potential for development at some locations (fig. 52). The middle Claiborne aquifer (layer 5) has good to excellent potential for ground-water development throughout much of the eastern embayment area (fig. 51*G*). This includes the northern part of the embayment where the Memphis aquifer includes layers 4 and 5. The middle Wilcox aquifer (layer 3) has some areas of poor and fair potential with smaller areas of good and excellent potential for ground-water development (fig. 51*I*). The lower Wilcox aquifer (layer 2) has large areas with fair and some areas with good and excellent potential for ground-water development (fig. 51*J*).

SOUTHERN TEXAS AREA

Some pumping developed in the early 1960's along the Rio Grande but began decreasing by the late 1960's (Mesko and others, 1990, fig. 6). Of all the areas in the study area, ground-water pumpage has least affected the hydrologic budget of the southern Texas area. The maximum ground-water pumpage is only about three-fourths of the rate of predevelopment regional recharge. Pumpage has caused regional recharge to increase (fig. 47) while natural regional ground-water discharge has decreased. The peak of ground-water pumpage and change in confined storage came in 1965 and was about 5 percent of the pumpage from permeable zone C (layer 9). Outflow and inflow to the area are about equal, and both have changed slowly over time.

Compared to other areas in the Gulf Coastal Plain, the southern Texas area has less potential for ground-water development (fig. 51) because this area has lower sand percentage and lower hydraulic conductivity of sand than other areas as well as less rainfall (figs. 4, 14, 18, 21). In the few areas where the aquifer has some potential for ground-water development, the potential is limited by dissolved-solids concentrations that generally exceed 3,000 mg/L (fig. 51) (Pettijohn and others, 1988).

SOUTHEASTERN TEXAS AREA

In the southeastern Texas area, peak ground-water pumpage occurred in 1980 (which was only slightly more than in 1970) and then decreased in 1985 (fig. 48). Natural regional discharge decreased to less than one-fourth of its predevelopment value. Regional recharge increased by more than a factor of four. The maximum pumpage is more than five times the predevelopment regional recharge rate. Contribution from confined storage, both elastic and inelastic, reached a peak in 1970 and was about 11 percent of the pumpage. Water withdrawn from inelastic storage has caused land subsidence, as described previously. Since pumpage began, there has been slightly more outflow to other areas than inflow from other areas.

Focusing on the ground-water budget of Houston and the surrounding area (fig. 53) shows some interest-

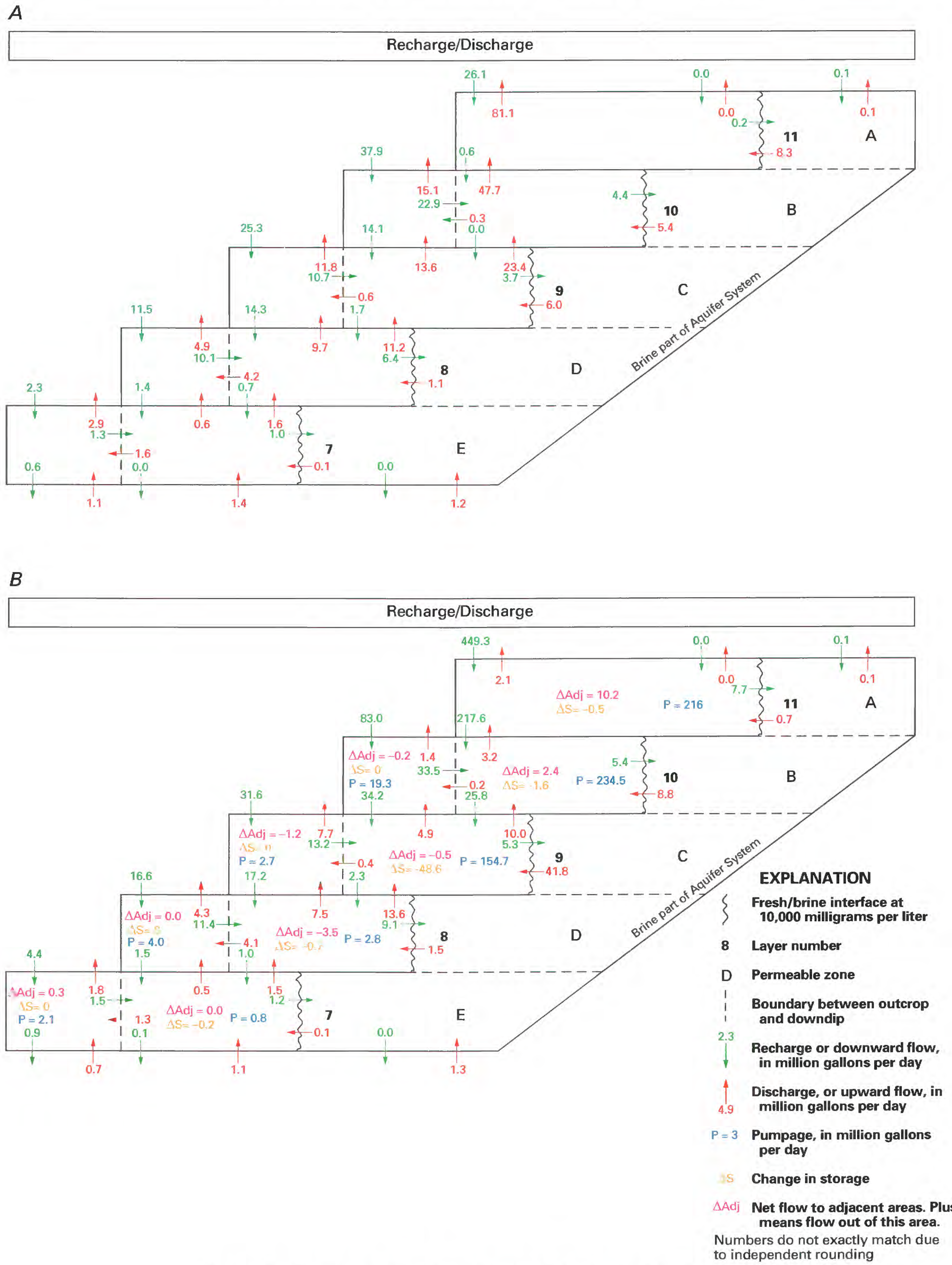


FIGURE 53.—Ground-water budget by model layer of the southeastern Texas area around Houston: *A*. predevelopment, and *B*. 1985.

ing effects of development that are more striking than the budget for the entire southeastern Texas area (fig. 48). The Houston area is defined for this purpose as the part of the coastal lowlands aquifer system between model rows 56 and 70 (fig. 2). The outcropping part of each layer is defined separately from the downdip part of the respective model layers for the summaries of flows presented in figure 53. Most of the change in confined storage in 1982–87 in the southeastern Texas area (81 Mgal/d, fig. 48) was in the downdip portion of permeable zone C (layer 9) in the Houston area (49 Mgal/d, fig. 53). Change in confined storage accounts for about 31 percent of the water pumped out of the nonoutcropping part of permeable zone C (layer 9) in the Houston area.

Another striking change in permeable zone C (layer 9) was the flow from the brine part of the layer toward the part that has relatively fresh water. The flow increased by a factor of 7 to nearly 42 Mgal/d in 1985 (fig. 53). This brine flow accounted for about 27 percent of the volume of flow toward the pumping wells in the nonoutcropping part of permeable zone C (layer 9). Owing to the large distances (tens of miles) from the locations with brine water to pumping centers and the relatively slow movement of ground water, wells will not necessarily be withdrawing any part of the brine water now or possibly even in the next few decades. Another large component of flow toward the cone of drawdown in permeable zone C (layer 9) was net downward flow from permeable zone B (layer 10), which amounted to about 29 percent of the pumpage in permeable zone C (fig. 53). Only about 8 percent of the pumpage from permeable zone C (layer 9) was supplied by recharge in the outcrop and flow downdip in the zone (fig. 53).

There were also changes in the flows in permeable zone B (layer 10) in the Houston area (fig. 53). As above, the net vertical flow was from the underlying permeable zone C (layer 9) in predevelopment conditions. By 1987 the flow direction had largely reversed, with flow moving down from permeable zone B (layer 10) to permeable zone C (layer 9). The net recharge to outcrop areas of model permeable zone B (layer 10) increased by a factor of four, to about 82 Mgal/d. Net vertical flow upward to permeable zone A (layer 11), which had been about 47 Mgal/d, reversed to about 210 Mgal/d downward. Interestingly, the flow into the brine area of permeable zone A (layer 11) increased due to development, probably to replenish water that flowed out of the brine area of permeable zone A (layer 11) and down into the fresher part of permeable zone B (layer 10). There were only small changes in the flows within permeable zones D and E (layers 7, 8) in the Houston area because there was little pumpage from those zones.

Simulations indicate that no single permeable zone has the most potential for ground-water development in the southeastern Texas area (fig. 52). With a few exceptions the permeable zone with the most potential is generally the shallowest one that exists at that location. One exception is west of Houston, where permeable zone A (layer 11) is at the surface, but permeable zone C (layer 9) has the most potential for ground-water development in three blocks and permeable zone B (layer 10) in another block (fig. 52). Permeable zone A (layer 11) has the largest outcrop area and hence the most potential for ground-water development, generally good or excellent except west and south of Houston and in some areas along the coast (fig. 51*A*). Permeable zone B (layer 10) has a considerable area with good potential for ground-water development (fig. 51*B*). Permeable zone C (layer 9) has a narrow strip with good potential for ground-water development (fig. 51*C*). Permeable zone D (layer 8) has some area with good and some area with excellent potential for ground-water development, especially toward the east (fig. 51*D*). Permeable zone E (layer 7) has very little potential for ground-water development in the southeastern Texas area (fig. 51*E*).

Land subsidence is a possible effect of future development of ground water, as it has been in the past in the southeastern Texas area. Subsidence is less likely at locations with good potential for ground-water development shown on the maps (fig. 51*A–E*) because those areas are in close proximity to sources of recharge.

SOUTHWESTERN LOUISIANA AREA

In the southwestern Louisiana area, ground-water pumpage peaked in 1980 (which was only slightly more than in 1970) and then decreased slightly during 1985 (fig. 49). Natural regional discharge decreased to about one-fifth of its predevelopment value. Regional recharge increased by more than a factor of four. The maximum pumpage is nearly six times the value of predevelopment regional recharge. Contributions from confined storage reached a peak in 1970 and were about 3 percent of the pumpage. Inflow from other areas was more than twice the outflow to other areas and was increasing somewhat throughout the whole period.

The greater relative potential for ground-water development is toward the western Louisiana border for most permeable zones. Permeable zone A (layer 11) has the largest outcrop area and hence the greatest potential for ground-water development, generally excellent or good (fig. 51*A*). Permeable zone B (layer 10) has some area with good and some area (toward the west) with excellent potential for ground-water development (fig. 51*B*). One area in the central part of layer 10 in southwestern Louisiana has only fair potential. Permeable zone C (layer 9) has two strips with good potential for ground-water development, one toward the west and one toward the east (fig. 51*C*). Permeable zone D (layer 8) exhibits a pattern similar to layer 9 (fig. 51*D*). Permeable zone E (layer 7) has very little potential for ground-water development in the southwestern Louisiana area (fig. 51*E*).

EASTERN COASTAL LOWLANDS AREA

In the eastern coastal lowlands area, ground-water pumpage peaked in 1980 and decreased slightly by 1985 (fig. 50). Natural regional discharge decreased to about one-half of its predevelopment value. Regional recharge increased by almost 50 percent. The maximum pumpage is slightly larger than the value of predevelopment regional recharge. Contribution from confined storage reached a peak in 1970 and was about 22 percent of the pumpage. This is the largest proportion of pumpage coming from confined storage of any area, possibly because ground-water development has increased more recently in this area. Inflow from other areas was somewhat less than the outflow to other areas. Inflow nearly doubled, and outflow decreased slightly during the period.

Nearly all of the permeable zones exhibit a north-south trend of greater potential for ground-water development along or slightly to the west of the Pearl River. Simulations show that no single permeable zone has the most potential for ground-water development in the eastern coastal lowlands area (fig. 52). With a few exceptions, the permeable zone with the most potential is generally the shallowest one that exists at that location. Permeable zone A (layer 11) has the largest outcrop area and hence the most potential for ground-water development, generally excellent or good except in the birdfoot delta area of the Mississippi River and in coastal Hancock County in Mississippi (fig. 51*A*). Permeable zone B (layer 10) has an area with excellent potential for ground-water development north of Lake Pontchartrain and several narrow bands with good potential (fig. 51*B*). Permeable zone C (layer 9) has several bands with good potential for ground-water development and some narrower areas with excellent potential in southern Mississippi and Alabama (fig. 51*C*). Permeable zone D (layer 8) has a considerable area with excellent potential for ground-water development in a wide strip along the Pearl River and in southern Alabama (fig. 51*D*). Adjoining those areas are strips with good potential for ground-water development in layer 8. Permeable zone E (layer 7) has a strip along the Pearl River with good potential for ground-water development (fig. 51*E*).

There is a large potential for further development of the entire ground-water resource. Doubling the pumpage to 20 Ggal/d might be accomplished without unreasonable consequences if the pattern of additional pumping was carefully designed. Doubling the pumpage was tested by using the pumping additions proportional to the potential for development (fig. 51). Therefore, model blocks were assigned pumpage according to table 5. The resulting pumpage distribution was simulated in the model, resulting in an average drawdown of 60 ft. Although this distribution of pumpage is impractical, it demonstrates the large potential for development.

TABLE 5.—*Distribution of pumpage in maximum development simulation according to the estimated potential for ground-water development*[1]

Potential for future ground-water development	Pumpage range (million gallons per day per 100-square-mile block)
Excellent	15 to 38
Good	1.8 to 15
Fair	0.35 to 1.8
Poor	0.04 to 0.35
Virtually none	0 to 0.04

[1]Potential ranges taken from figure 51. Pumpage range calculated from log-log curve relating pumpage to the drawdown shown in figure 51 from constant incremental pumping. Pumpage limited to 38 million gallons per day. These pumpages were for 100-square-mile block values for each block where relatively fresh water (less than 10,000 milligrams per liter) is present.

SUMMARY AND CONCLUSIONS

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 systems (and the study area) extend offshore beneath the Gulf of Mexico to include an additional 60,000 mi^2 and are truncated at the edge of the Continental Shelf. The Gulf Coast RASA study is limited to coastal plain sediments of mostly Cenozoic age except for the northernmost part of the area where it includes sediments of Late Cretaceous age. The geohydrologic units generally thicken toward the Gulf of Mexico and reach a combined 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. Nearly 10 Ggal/d of ground water was withdrawn from the aquifers in 1985, mostly for irrigation, but public supply and industrial uses are also important.

The study area is a gently sloping coastal plain in a mostly humid, temperate climate, underlain by thick, mostly clastic sediments deposited in a gulfward offlapping sequence. Historical sea-level oscillations caused cyclical sedimentation. Depositional environments alternating from predominantly continental to predominantly marine resulted in alternating fine-grained and more coarse-grained deposits. The base of the aquifer system in the north is at the top of the Midway confining unit and in the southern part of the area is at the transition zone into geopressured sediments. Though faults are common, fault throws are generally not great enough to entirely offset the full thickness of the regional hydrologic units described in this report, although individual beds could be offset. Many salt domes are present in several basins of the gulf coast regional aquifer systems. The highly mineralized ground water, which probably has resulted from salt-dome dissolution, largely affects regional flow by creating areas with high fluid densities.

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. 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. The aquifer systems were divided into 10 aquifers and 5 regional confining units, each of which was defined as a separate layer in a three-dimensional, variable density, finite-difference ground-water flow model. The model grid consists of 102 by 58 cells that are each 10 mi on a side. The water density, although variable in space, is assumed to be approximately constant in time, a valid assumption given the large model blocks and simulation time that is short relative to transport times. The model accounted for irreversible compaction of fine-grained beds, resulting in water being released from storage in fine-grained beds and in land subsidence. Many of the aquifer characteristics such as thickness, sand percentage, and water density (based on concentration of dissolved solids, temperature, and pressure) were derived from a computerized file of 989 geophysical logs.

The factors that controlled regional ground-water flow in the aquifer systems under predevelopment conditions are, in order of importance of effect (1) topography; (2) outcrop and subcrop pattern and geometry of aquifers, permeable zones, and confining units; (3) variation of hydraulic properties of aquifers, permeable zones, and confining units; (4) distribution of density and brines; and (5) downdip limits of geohydrologic units and geologic structure.

Topography is the most significant factor controlling predevelopment ground-water flow because the relatively humid climate maintains the aquifer systems generally full to overflowing with ground water. Therefore, the amount and distribution of regional recharge and discharge were proportional to the topographic gradient and the conductance of the aquifer. The major variations in subsurface permeability have been accounted for simply by delineation of aquifers and confining units. Sand-bed hydraulic conductivity increases from the western side of the study area toward the eastern side. Regional effective hydraulic conductivity varies as a power function of sand percentage because as the sand percentage decreases, the degree of hydraulic connection of the sand beds also decreases. Conductivity also decreases with depth due to compaction.

About one-third of the aquifer system studied contains freshwater, whereas the dissolved-solids concentrations of the saline water increases over relatively short distances to more than 100,000 mg/L. The resulting large differences in water density has substantial effects on ground-water flow both in the saline and the freshwater part of the system. Many forces apparently have prevented the more concentrated saline part of the system from reaching equilibrium. These forces are heat and salt sources (salt domes), sediment compaction, geopressured sediments, changing sea level, and a very large continental shelf requiring long flow paths to move brine to make adjustments. It is possible that most of the salt contained in high-concentration brines came from dissolution of salt domes. Geologic structure seems to have a small effect on ground-water flow except indirectly through the geometry of geohydrologic units. 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, was found to be insignificant compared to the volume of flow in the aquifer above it. This may also be true for the small amounts of flow leaking up out of the geopressured zone farther downdip.

One of the most prominent predevelopment flow paths in the Mississippi embayment aquifer system is from a major recharge area in south-central Mississippi, radiating outward with a major part curving to the south and then west. A large part of this flow moves southwestward and then curves northwestward to the regional discharge area in northeastern Louisiana. This curvature in flow paths is due partly to the resistance to vertical flow downdip. In the Texas coastal uplands aquifer system, the predominant regional flow pattern is from recharge areas between major river valleys laterally to discharge areas along the valleys. In the coastal lowlands aquifer system, the predominant flow pattern is from recharge areas in the updip and topographically higher areas along the inland edge of the system to discharge areas onshore towards the coast. Flow patterns in the deeper coastal lowlands aquifer system are more variable and difficult to interpret because of the large variability due to the complex three-dimensional distribution of ground-water density.

Predevelopment net regional recharge occurred in 41 percent of the modeled area at the average rate of about 0.48 in/yr. The model block with the highest rate was in Mississippi at about 6 in/yr. Predevelopment net regional discharge occurred on 59 percent of the area at the rate of about 0.35 in/yr. The model block with the largest regional discharge was in the Mississippi River Valley alluvial aquifer at about 4 in/yr. The distribution of freshwater thickness correlates well with the simulated net regional flow and discharge, indicating that the regional recharge has flushed saline water and created a thick section of freshwater. Some ground water flows offshore before it is discharged to the ocean, but most of it discharges before it reaches the coastline.

Massive changes have occurred in the gulf coast regional aquifer systems in response to large-scale development of ground water throughout much of the study area. Development has led to lowering of hydraulic heads, changes in recharge and discharge, changes in flow velocity, changes in flow direction and magnitude, changes in water quality, and land subsidence. Total pumpage increased by a factor of 2.5 during 1960–80 but decreased by 7 percent to 1985. Irrigated agriculture uses the largest volume (about three-fourths) of ground water in the study area, and also represents the fastest growth in ground-water pumpage for the period 1960 to 1985. Much of the increase in agricultural pumpage was between 1970 and 1980, especially in the Mississippi Alluvial Plain. Other areas with large use of ground water for irrigation are southwestern Louisiana and southern Texas. More than 60 percent of the municipal and industrial pumpage is withdrawn from the coastal lowlands aquifer system. Ground-water usage for public supply has continued to increase with the population, whereas industrial pumpage at many locations has declined since 1970, largely spurred by the adverse effects of pumping and economic factors.

Every aquifer or permeable zone has more than one major cone of drawdown and has maximum block-averaged drawdowns exceeding 80 ft. The 1985 pumpage is more than three times the rate of predevelopment regional recharge. Pumpage and the resulting declines in hydraulic head caused the regional recharge to increase by more than a factor of three to more than 9 Ggal/d by increasing natural recharge and capturing water that was previously locally discharged. Meanwhile, the regional discharge decreased to about 1 Ggal/d, nearly one-third of its predevelopment value. Thus, most of the simulated 1985 pumpage was supplied from increased regional recharge. Less than 20 percent of the pumpage was captured natural regional discharge, and less than 10 percent was from ground-water storage.

By 1987, the proportion of the study area that contributed net regional recharge had increased by more than 50 percent, from 41 percent of the study area to 66 percent of the study area. The area that changed from having net regional discharge to net regional recharge includes most of the area underlain by the Mississippi River Valley alluvial aquifer and most of the lowlands area within about 40 mi of the coastline except in southern Texas. Even offshore, there has been a reversal in the direction of simulated flow from onshore to the Gulf of Mexico off the coast near Houston during predevelopment to flow from the Gulf of Mexico toward the coastline in 1985.

By 1985, most of the net vertical flows between aquifers changed direction, or at least the magnitude changed considerably. Downward vertical flows between aquifers increased substantially in all of the aquifers and permeable zones as a result of ground-water pumpage and resultant reversal of vertical-head gradients. The upper Claiborne aquifer continued to have the most flow interchange with the Mississippi River Valley alluvial aquifer, probably due to the large area of contact. The upper Claiborne has very little pumpage yet has large inflows and outflows, capturing additional recharge and discharge and transmitting them to adjacent layers. The other aquifers are mostly self-supporting, meaning that most of their pumpage was supplied from increased recharge and decreased discharge within the aquifer. In the lower Wilcox aquifer and permeable zone E, which have very narrow outcrop and subcrop bands, much of the pumpage had to be supplied by vertical flow from adjacent layers.

In the Houston, Texas area, in the nonoutcropping part of permeable zone C, change in storage accounted for 31 percent of the flow toward the pumping wells. The flow from the brine area accounted for 27 percent of the volume of pumpage. However, wells will not necessarily be withdrawing any part of the brine water immediately or possibly even in the near future owing to the distances from the brine water and the relatively slow movement of ground water. Another large component of flow toward the cone of drawdown in permeable zone C was downward flow from layer 10, which amounted to 29 percent of the pumpage. Only 8 percent of the pumpage was supplied by recharge in the outcrop flowing downdip to the pumping center in permeable zone C.

Large-scale ground-water pumpage in the freshwater part of the aquifer systems has markedly changed flow patterns in the brine part and induced flow updip toward pumping centers. However, it will take decades or centuries for the brine to reach these pumping areas because of the relatively slow ground-water velocities and long flow paths. For example, freshwater pumping onshore in and around Houston had changed flow directions more than 45 degrees or flow rates more than 50 percent in the brine part of the lower Pleistocene and upper Pliocene deposits (permeable zone B; layer 10) throughout an area of 6,000 to 8,000 mi^2 that extends as far as about 60 mi offshore. Block-averaged Darcian velocities in the brine approach 10 ft/yr. Assuming an effective porosity of 20 percent, this would mean average particle velocities of about 50 ft/yr, or about 100 yr to move 1 mi. These estimates could be in error by an order of magnitude or more. However, they indicate a possibility of long-term problems that may not be noticed for decades, yet which need consideration for prudent management of the resource.

The model was used to predict the relative potential for increased ground-water pumpage by making an areal analysis by aquifer of sensitivity to pumpage by simulating a unit increase in pumpage in each model block containing relatively fresh water. In addition to making maps by aquifer of relative potential for ground-water development, a composite map was made showing the layer in each grid block which had the least sensitivity to pumpage (least drawdown) and therefore the most potential for development.

Simulation indicates there is great potential for continued additional development of the ground-water resource in the study area. Doubling the pumpage to 20 Ggal/d produced minimal consequences (average 60 ft of drawdown over the study area) if the pattern of added pumping was carefully designed. While this distribution of pumping is not very practical, it does demonstrate the potential for development.

SELECTED REFERENCES

Ackerman, D.J., 1989, Hydrology of the Mississippi River Valley alluvial aquifer, south-central United States—A preliminary assessment of the regional flow system: U.S. Geological Survey Water-Resources Investigations Report 88–4028, 74 p.

———1996, Hydrology of the Mississippi River Valley alluvial aquifer, south-central United States: U.S. Geological Survey Professional Paper 1416–D, p. D1–D56, 8 pls. in separate case.

Arthur, J.K., and Taylor, R.E., 1990, Definition of the geohydrologic framework and preliminary simulation of ground-water flow in the Mississippi embayment aquifer system, United States: U.S. Geological Survey Water-Resources Investigations Report 86–4364, 97 p.

Arthur, J.K., and Taylor, R.E., in press, Ground-water flow analysis of the Mississippi embayment aquifer system, south-central United States: U.S. Geological Survey Professional Paper 1416–I.

Baker, E.T., Jr., 1986, Hydrology of the Jasper aquifer in the southeast Texas Coastal Plain: Texas Water Development Board Report 295, 64 p.

Baker, E.T., Jr., and Wall, J.R., 1976, Summary appraisals of the Nation's ground-water resources, Texas-Gulf region: U.S. Geological Survey Professional Paper 813–F, 29 p.

Barker, R.A., Bush, P.W., and Baker, E.T, Jr., 1994, Geologic history and hydrogeologic setting of the Edwards-Trinity aquifer system, west-central Texas: U.S. Geological Survey Water-Resources Investigations Report 94–4039, 50 p.

Barker, R.A., and Pernik, Maribeth, 1994, Regional geohydrology and computer-model simulation of deep ground-water flow within the Southeastern Coastal Plain aquifer system of the United States: U.S. Geological Survey Professional Paper 1410–C, C1–C–87, 10 pls. in separate case.

Bateman, R.M., and Konen, C.E., 1977, The log analyst and the programmable calculator: The Log Analyst, v. 18, no. 5, p. 3–11.

Bear, Jacob, 1972, Dynamics of fluids in porous media: New York, American Elsevier, p. 153–155.

Bear, Jacob, and Verruijt, Arnold, 1987, Modeling groundwater flow and pollution: Dordrecht, Holland, D. Reidel Publishing Company, p. 35–37.

Beckman, J.D., and Williamson, A.K., 1990, Salt dome locations in the Gulf Coastal Plain, south-central United States: U.S. Geological Survey Water-Resources Investigations Report 90–4060, 44 p.

Bedinger, M.S., and Sniegocki, R.T., 1976, Summary appraisals of the Nation's ground-water resources—Arkansas-White-Red region: U.S. Geological Survey Professional Paper 813–H, 31 p.

Bennett, G.D., 1979, Regional ground-water systems analysis: U.S. Army Corps of Engineers, Water Support Center, Fort Belvoir, Va., Water Spectrum, v. 11, no. 4, p. 36–42.

Bennett, S.S., and Hanor, J.S., 1987, Dynamics of subsurface salt dissolution at the Welsh dome, Louisiana Gulf Coast, *in* Lerche, I., and O'Brien, J.S., eds., Dynamical geology of salt and related structures: Orlando, Fla., Academy Press, p. 653–677.

Bond, D.C., 1972, Hydrodynamics in deep aquifers of the Illinois basin: Illinois Geological Survey Circular 470, 69 p.

———1973, Deduction of flow patterns in variable-density aquifers from pressure and water-level observations, *in* Underground waste water and artificial recharge, preprints and papers: American Association of Petroleum Geologists, v. 1, p. 357–378.

Boswell, E.H., Cushing, E.M., and Hosman, R.L., 1968, Quaternary aquifers in the Mississippi embayment, *with a discussion of* Quality of the water, by H.G. Jeffery: U.S. Geological Survey Professional Paper 448–E, 15 p.

Boswell, E.H., Moore, G.K, and MacCary, L.M., 1965, Cretaceous aquifers in the Mississippi embayment, *with a discussion of* Quality of the water, by H.G. Jeffery: U.S. Geological Survey Professional Paper 448–C, 37 p.

Brahana, J.V., 1982, Two-dimensional digital ground-water model of the Memphis Sand and equivalent units, Tennessee-Arkansas-Mississippi: U.S. Geological Survey Open-File Report 82–0099, 55 p.

———1987, The role of multilayer model in refining understanding of deep regional ground-water flow in a tectonically active area, *in* Conference on Solving Ground Water Problems with Models, Denver, Colo., February 10–12, 1987, Proceedings: Association of Ground Water Scientists and Engineers, v. 2, p. 1051–1070.

Brahana, J.V., and Mesko, T.O., 1988, Hydrogeology and preliminary assessment of regional flow in the Upper Cretaceous and adjacent aquifers, northern Mississippi embayment: U.S. Geological Survey Water-Resources Investigations Report 87–4000, 65 p.

Bredehoeft, J.D., Back, William, and Hanshaw, B.B., 1982, Regional ground-water flow concepts in the United States; historical perspective, *in* Narasimhan, T.N., ed., Recent trends in hydrogeology: Geological Society of America, Special Paper 189, p. 297–316.

Bredehoeft, J.D., Blyth, C.R., White, W.A., and Maxey, G.B., 1963, Possible mechanism for concentration of brines in subsurface formations: American Association of Petroleum Geologists Bulletin, v. 47, p. 257–269.

Bredehoeft, J.D., Neuzil, C.E., and Milly, P.C.D., 1983, Regional flow in the Dakota Aquifer—A study in the role of confining layers: U.S. Geological Survey Water-Supply Paper 2237, 45 p.

Broom, M.E., and Lyford, F.P., 1981, Alluvial aquifer of the Cache and St. Francis River Basins, northeastern Arkansas: U.S. Geological Survey Open-File Report 81–0476, 48 p.

Broom, M.E., and Reed, J.E., 1973, Hydrology of the Bayou Bartholomew alluvial aquifer-stream system, Arkansas: U.S. Geological Survey Open-File Report 73–0034, 91 p.

Bush, P.W., 1986, Planning report for the Edwards-Trinity Regional Aquifer-System Analysis in central Texas, southeast Oklahoma, and southwest Arkansas: U.S. Geological Survey Water-Resources Investigations Report 86–4343, 15 p.

Bush, P.W., and Johnston, R.H., 1988, Ground-water hydraulics, regional flow, and ground-water development of the Floridan aquifer system in Florida and in parts of Georgia, South Carolina, and Alabama: U.S. Geological Survey Professional Paper 1403–C, 80 p.

Cardwell, W.T., Jr., and Parsons, R.L., 1945, Average permeabilities of heterogenous oil sands: American Institute of Mining and Metallurgical Engineers Transactions, fall meeting, Los Angeles, Calif., October 19–20, 1944, v. 160, p. 283–291.

Carr, J.E., Meyer, W.R., Sandeen, W.M., and McLane, I.R., 1985, Digital models for simulation of ground-water hydrology of the Chicot and Evangeline aquifers along the Gulf Coast of Texas: Texas Department of Water Resources Report 289, 101 p.

Cederstrom, D.J., Boswell, E.H., and Tarver, G.R., 1979, Summary appraisals of the Nation's ground-water resources—South Atlantic-Gulf region: U.S. Geological Survey Professional Paper 813–O, 35 p.

Charbeneau, R.J., and Wright, S.G., 1983, Hydrologic site selection for mining of deep basin lignite in Texas: The University of Texas at Austin Center for Research in Water Resources Technical Report 198, 113 p.

Chi, S.C., and Reilinger, R.E., 1984, Geodetic evidence for subsidence due to groundwater withdrawal in many parts of the United States of America: Journal of Hydrology, v. 67, p. 155–182.

Clayton, R.N., Friedman, Irving, Graf, D.L., Mayeda, T.K., Meents, W.F., and Shimp, N.F., 1966, The origin of saline formation waters: Journal of Geophysical Research, v. 71, p. 3869–3882.

Cooley, R.L., 1979, A method of estimating parameters and assessing reliability for models of steady state ground-water flow; 2. Application of statistical analysis: Water Resources Research, v. 15, no. 3, p. 603–617.

———1982, Incorporation of prior information on parameters into nonlinear regression ground-water flow models; 1, Theory: Water Resources Research, v. 18, no. 4, p. 965–976.

Cooley, R.L., Konikow, L.F., and Naff, R.L., 1986, Nonlinear-regression groundwater flow modeling of a deep regional aquifer system: Water Resources Research, v. 22, no. 13, p. 1759–1778.

Crider, A.F., and Johnson, L.C., 1906, Summary of the underground-water resources of Mississippi: U.S. Geological Survey Water-Supply Paper 159, 86 p.

Cushing, E.M., Boswell, E.H., Speer, P.R., and Hosman, R.L., 1970, Availability of water in the Mississippi embayment: U.S. Geological Survey Professional Paper 448–A, 13 p.

Dagan, Gedeon, 1986, Statistical theory of groundwater flow and transport—Pore to laboratory, laboratory to formation, and formation to regional scale: Water Resources Research, v. 22, no. 9, p. 120S–134S.

De Sitter, L.U., 1947, Diagenesis of oil-filled brines: Bulletin of the American Association of Petroleum Geologists, v. 31, p. 2030–2040.

Desbarats, A.J., 1987, Numerical estimation of effective permeability in sand-shale formations: Water Resources Research, v. 23, no. 2, p. 273–286.

Deussen, Alexander, 1914, Geology and underground water of the southeastern part of the Texas Coastal Plain: U.S. Geological Survey Water-Supply Paper 335, 365 p.

———1924, Geology of the Coastal Plain of Texas west of Brazos River: U.S. Geological Survey Professional Paper 126, 144 p.

Deussen, Alexander, and Dole, R.B., 1916, Ground water in LaSalle and McMullen Counties, Texas, in Contributions to the hydrology of the United States, 1915: U.S. Geological Survey Water-Supply Paper 375, p. 141–177.

Dirección General de Geográfica del Territorio Nacional, 1984, Carta Geologica, Monterrey and Mexico sheets: Mexico City, Mexico, 2 sheets, scale 1:1,000,000.

Durbin, T.J., Kapple, G.W., and Freckleton, J.R., 1978, Development and use of two-dimensional digital models for the Salinas Valley ground-water basin, California; Part 1, Development of ground-water flow models: U.S. Geological Survey Water-Resources Investigations 78–0113, 134 p.

Fenneman, N.M., 1938, Physiography of eastern United States: New York, McGraw-Hill Book Company, 714 p.

Fertl, W.H., 1976, Abnormal formation pressures; implications to exploration, drilling, and production of oil and gas resources: New York, Elsevier Scientific Publishing Co., 382 p.

Fogg, G.E., Jr., 1986a, Stochastic analysis of aquifer interconnectedness, with a test case in the Wilcox group, east Texas: Austin, The University of Texas at Austin, unpublished Ph.D. dissertation, 216 p.

———1986b Groundwater flow and sand body interconnectedness in thick, multiple-aquifer system: Water Resources Research, v. 22, no. 5, p. 679–694.

———1989, Stochastic analysis of aquifer interconnectedness—Wilcox Group, Trawich area, East Texas: Austin, The University of Texas at Austin, Bureau of Economic Geology Report of Investigations 189, 68 p.

Fogg, G.E., Jr., Seni, S.J., and Kreitler, C.W., 1983, Three-dimensional ground-water modeling in depositional systems—Wilcox Group, Oakwood Salt Dome area, east Texas: Austin, The University of Texas at Austin, Bureau of Economic Geology Report of Investigations 133, 55 p.

Franke, O.L., Reilly, T.E., and Bennett, G.D., 1987, Definition of boundary and initial conditions in the analysis of saturated ground-water flow systems—An introduction: U.S. Geological Survey Techniques of Water Resources Investigations, book 3, chap. B5, 15 p.

Freeze, R.A., and Cherry, J.A., 1979, Groundwater: Englewood Cliffs, N.J., Prentice-Hall, Inc., 604 p.

Freeze, R.A., and Witherspoon, P.A., 1967, Theoretical analysis of groundwater flow; 2. Effect of water-table configuration and subsurface permeability variation: Water Resources Research, v. 3, no. 2, p. 623–634.

Fuller, M.L., 1904, Contributions to the hydrology of the eastern United States, 1903: U.S. Geological Survey Water-Supply Paper 102, 522 p.

———1905, Underground waters of the eastern United States: U.S. Geological Survey Water-Supply Paper 114, 285 p.

Gabrysch, R.K., 1969, Land-surface subsidence in the Houston-Galveston region, Texas, *in* International Symposium on Land Subsidence, 1st, Tokyo, Japan, September 1969, Proceedings: International Association of Hydrological Sciences, Publication 88, p. 43–54.

———1977, Land-surface subsidence in the Houston-Galveston region, Texas, *in* International Symposium on Land Subsidence, 2d, Anaheim, Calif., December 13–17, 1976, Proceedings: International Association of Hydrological Sciences, Publication 121, v. 1, p. 16–24.

———1982, Ground-water withdrawals and land-surface subsidence in the Houston-Galveston region, Texas, 1906–80: U.S. Geological Survey Open-File Report 82–0571, 68 p.

———1984, Ground-water withdrawals and changes in water levels in the Houston District, Texas, 1975–79: Texas Department of Water Resources Report 286, 42 p.

Gabrysch, R.K., and Bonnet, C.W., 1975, Land-surface subsidence in the Houston-Galveston region, Texas: Texas Water Development Board Report 188, 19 p.

Gandl, L.A., 1982, Characterization of aquifers designated as potential drinking water sources in Mississippi: U.S. Geological Survey Open-File Report 81–550, 90 p.

Garza, Sergio, 1974, Projected effects of the proposed Tennessee Colony Reservoir on ground-water conditions in the Carrizo-Wilcox aquifer and Trinity River alluvium, Trinidad-Oakwood area, Texas: U.S. Geological Survey Open-File Report 74–0297, 33 p.

Gebert, W.A., Graczyk, D.J., and Krug, W.R., 1987, Average annual runoff in the United States, 1951–80: U.S. Geological Survey Hydrologic Investigations Atlas HA–710, 1 sheet, scale 1:7,500,00.

Glasstone, Samuel, 1946, Textbook of physical chemistry (2d ed.): New York, Van Nostrand, 1,320 p.

Glenn, L.C., 1906, Underground water of Tennessee and Kentucky west of the Tennessee River and of an adjacent area in Illinois: U.S. Geological Survey Water-Supply Paper 164, 173 p.

Godson, R.H., 1981, Digital terrain map of the United States: U.S. Geological Survey Miscellaneous Investigations Map I–1318, 1 sheet, scale 1:7,500,000.

Goldman, M.I., 1952, Deformation, metamorphism, and mineralization in gypsum-anhydrite cap rock, Sulphur salt dome, Louisiana: Geological Society of America, Memoir 50, 169 p.

Gondouin, M., Tixier, M.P., and Simard, G.L., 1957, An experimental study on the influence of the chemical composition of electrolytes on the SP curve: American Institute of Mining and Metallurgical Engineers, Petroleum Transactions, v. 210, p. 58–72.

Groschen, G.E., 1985, Simulated effects of projected pumping on the availability of freshwater in the Evangeline aquifer in an area southwest of Corpus Christi, Texas: U.S. Geological Survey Water-Resources Investigations Report 85–4182, 103 p.

Grubb, H.F., 1984, Planning report for the Gulf Coast Regional Aquifer-System Analysis, Gulf of Mexico Coastal Plain, United States: U.S. Geological Survey Water-Resources Investigations Report 84–4291, 30 p.

———1985, The Gulf Coast Regional Aquifer-System Analysis, an overview, *in* Smerdon, E.T., and Jordon, W.R., eds., Issues in groundwater management: The University of Texas at Austin, Center for Research in Water Resources, Water Resources Symposium Twelve, San Antonio, Texas, October 29–31, 1984, p. 71–91.

———1986, Gulf Coast Regional Aquifer-System Analysis—A Mississippi perspective: U.S. Geological Survey Water-Resources Investigations Report 86–4162, 22 p.

———1987, Overview of the Gulf Coast Regional Aquifer-System Analysis, south-central United States, *in* Vecchioli, John, and Johnson, A.I., eds., Regional aquifer systems of the United States, Aquifers of the Atlantic and Gulf Coastal Plain: American Water Resources Association Monograph 9, p. 101–118.

Halbouty, M.T., 1979, Salt domes, Gulf Region, United States and Mexico: Houston, Texas, Gulf Publishing Co., 561 p.

Hanor, J.S., 1987, Kilometre-scale thermohaline overturn of pore waters in the Louisiana Gulf Coast: Nature, v. 327, p. 501–503.

Hanor, J.S., and Bailey, J.E., 1983, Use of hydraulic head and hydraulic gradient to characterize geopressured sediments and the direction of fluid migration in the Louisiana Gulf Coast: Gulf Coast Association of Geological Societies Transactions, v. 33, p. 115–122.

Hanor, J.S., Bailey, J.E., Rogers, M.C., and Milner, L.R., 1986, Regional variations in physical and chemical properties of south Louisiana oil field brines: Gulf Coast Association of Geological Societies Transactions, v. 36, p. 143–149.

Harris, G.D., 1904, Underground waters of southern Louisiana, with discussions of their uses for water supplies and for rice irrigation by M.L. Fuller: U.S. Geological Survey Water-Supply Paper 101, 98 p.

Henry, D.C., Galloway, W.E., and Smith, G.E., 1982, Considerations in the extraction of uranium from a fresh-water aquifer—Miocene Oakville Sandstone, south Texas: Austin, The University of Texas at Austin Bureau of Economic Geology Report of Investigations 126, 36 p.

Hill, R.T., 1901, Geography and geology of the Black and Grand Prairies, Texas: U.S. Geological Survey Twenty-first Annual Report, pt. 7, 666 p.

Holdahl, S.R., and Morrison, N.L., 1974, Regional investigations of vertical crustal movements in the U.S., using precise relevelings and mareograph data: Tectonophysics, v. 23, p. 373–390.

Hosman, R.L., 1988, Geohydrologic framework of the Gulf Coastal Plain: U.S. Geological Survey Hydrologic Investigations Atlas HA–695, 2 sheets, scale 1:2,500,000.

———1993, Regional stratigraphy and subsurface geology of Cenozoic deposits, Gulf Coastal Plain, south-central United States: U.S. Geological Survey Professional Paper 1416–G, 34 p.

Hosman, R.L., Long, A.T., Lambert, T.W., and others, 1968, Tertiary aquifers in the Mississippi embayment, *with discussions of* Quality of the water, by H.G. Jeffery: U.S. Geological Survey Professional Paper 448–D, 19 p.

Hosman, R.L., and Weiss, J.S., 1991, Geohydrologic units of the Mississippi embayment and Texas coastal uplands aquifer systems, south-central United States: U.S. Geological Survey Professional Paper 1416–B, 19 p.

Huntzinger, T.L., Whiteman, C.D., and Knochenmus, D.D., 1985, Simulation of ground-water movement in the "1,500- and 1,700-foot" aquifer of the Baton Rouge area, Louisiana: Louisiana Department of Transportation and Development, Office of Public Works Technical Report 34, 52 p.

Imes, J.L., 1989, Geohydrology and hydrochemistry of the Ozark Plateaus aquifer system, *in* Swain, L.A., and Johnson, Ivan A., eds., Regional aquifer systems of the United States—Aquifers of the midwestern area: American Water Resources Association Monograph 13, p. 165–178.

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

Johnston, R.H., and Bush, P.W., 1988, Summary of the hydrology of the Floridan aquifer system in Florida and in parts of Georgia, South Carolina, and Alabama: U.S. Geological Survey Professional Paper 1403–A, 24 p.

Jones, P.H., 1969, Hydrology of Neogene deposits in the northern Gulf of Mexico basin: Louisiana State University, Louisiana Water Resources Research Institute Bulletin GT–2, 105 p.

Jorgensen, D.G., 1973, Analog-model studies of the effects of recharge wells along the Houston ship channel on potentiometric surfaces of the Chicot and Evangeline aquifers, Houston, Texas: U.S. Geological Survey Open-File Report 73–0138, 2 sheets, scale 1:130,000.

———1975, Analog-model studies of ground-water hydrology in the Houston District, Texas: Texas Water Development Board Report 190, 84 p.

———1981, Geohydrologic models of the Houston District, Texas: Ground Water, v. 19, no. 4, p. 418–428.

Jorgensen, D.G., and Gabrysch, R.K., 1974, Simulated water-level changes resulting from proposed changes in ground-water pumping in the Houston, Texas, area: U.S. Geological Survey Open-File Report 75–0020, 3 sheets, scale 1:820,000.

Jorgensen, D.G., and Signor, D.C., 1981, Plan of study for the Central Midwest Regional Aquifer System analysis in parts of Arkansas, Colorado, Kansas, Missouri, Nebraska, New Mexico, Oklahoma, South Dakota, and Texas: U.S. Geological Survey Water-Resources Investigations Report 81–0206, 28 p.

Jorgensen, D.G., Helgesen, J.O., Signor, D.C., Leonard, R.B., Imes, J.L., and Christenson, S.C., 1993, Analysis of regional aquifers in the central Midwest of the United States in Kansas, Nebraska, and parts of Arkansas, Colorado, Missouri, New Mexico, Oklahoma, South Dakota, Texas, and Wyoming—Summary: U.S. Geological Survey Professional Paper 1414–A, 67 p.

Kidd, R.E., 1987, Hydrogeology and water-supply potential of the water-table aquifer on Dauphin Island, Alabama: U.S. Geological Survey Water-Resources Investigations Report 87–4283, 49 p.

Kirkpatrick, K.A., 1993, Archiving data from Gulf Coast Regional Aquifer System Analysis study: U.S. Geological Survey, Open-File Report 92–661, 18 p., 1 WORM disk.

Kitanidis, P.K., and Vomvoris, E.G., 1983, A geostatistical approach to the inverse problem in groundwater modeling (steady state) and one-dimensional simulations: Water Resources Research, v. 19, no. 3, p. 677–690.

Klemt, W.B., Duffin, G.L., and Elder, G.R., 1976, Ground-water resources of the Carrizo aquifer in the Winter Garden area of Texas, v. 1: Texas Water Development Board Report 210, 66 p.

Konikow, L.F., 1978, Calibration of ground-water models, *in* Conference on Verification of Mathematical and Physical Models in Hydraulic Engineering, College Park, Md., August 9–11, 1978, Proceedings: American Society of Civil Engineers, p. 87–93.

Kontis, A.L., and Mandle, R.J., 1988, Modifications of a three-dimensional ground-water flow model to account for variable water density and effects of multiaquifer wells: U.S. Geological Survey Water-Resources Investigations Report 87–4265, 78 p.

Kuiper, L.K., 1981, A comparison of the incomplete Cholesky-Conjugate gradient method with the strongly implicit method as applied to the solution of two-dimensional groundwater flow equations: Water Resources Research, v. 17, no. 4, p. 1082–1086.

———1983, A numerical procedure for the solution of the steady-state variable density groundwater flow equation: Water Resources Research, v. 19, no. 1, p. 234–240.

———1985, Documentation of a numerical code for the simulation of variable density ground-water flow in three dimensions: U.S. Geological Survey Water-Resources Investigations Report 84–4302, 90 p.

———1986, A comparison of several methods for the solution of the inverse problem in two-dimensional steady-state groundwater flow modeling: Water Resources Research, v. 22, no. 5, p. 705–714.

———1987, A comparison of iterative methods as applied to the solution of the three-dimensional ground-water flow equation: Society of Industrial and Applied Mathematics Journal on Scientific and Statistical Computing, v. 8, no. 4, p. 521–528.

———1994, Nonlinear regression flow model, gulf coast aquifer systems, south-central United States: U.S. Geological Survey Water-Resources Investigations Report 93–4020, 171 p.

Kupfer, D.H., 1963, Structure of salt in Gulf Coast domes: The American Association of Petroleum Geologists Bulletin, v. 46, no. 8, p. 1460–1467.

Lake, L.W., and Carroll, H.B., Jr., 1986, Reservoir characterization: New York, Academic Press, 660 p.

Leake, S.A., 1990, Interbed storage changes and compaction in models of regional groundwater flow: Water Resources Research, v. 26, no. 9, p. 1939–1950.

Leake, S.A., and Prudic, D.E., 1987, Non-linear subsidence module for three-dimensional finite-difference ground-water flow model [abs.]: American Geophysical Union Transactions, fall meeting, San Francisco, Calif., December 7–11, 1987, v. 68, no. 44, p. 1301.

———1988, Documentation of a computer program to simulate aquifer-system compaction using the modular finite-difference ground-water flow model: U.S. Geological Survey Open-File Report 88–482, 80 p.

Livingston, Penn, and Bridges, T.W., 1936, Ground-water resources of Kleberg County, Texas: U.S. Geological Survey Water-Supply Paper 773–D, p. 197–232.

Lobmeyer, D.H., 1985, Freshwater heads and ground-water temperatures in aquifers of the Northern Great Plains in parts of Montana, North Dakota, South Dakota, and Wyoming: U.S. Geological Survey Professional Paper 1402–D, 11 p.

Lonsdale, J.T., 1935, Geology and ground-water resources of Atascosa and Frio Counties, Texas: U.S. Geological Survey Water-Supply Paper 676, 90 p.

Lonsdale, J.T., and Day, J.R., 1937, Geology and ground-water resources of Webb County, Texas: U.S. Geological Survey Water-Supply Paper 778, 104 p.

Loucks, R.G., Dodge, M.M., and Galloway, W.E., 1986, Controls on porosity and permeability of hydrocarbon reservoirs in lower Tertiary sandstones along the Texas Gulf Coast: Austin, The University of Texas at Austin, Bureau of Economic Geology Report of Investigations 149, 78 p.

Louisiana Department of Natural Resources, 1986, Louisiana annual oil and gas report—1985: Baton Rouge, La., Department of Natural Resources, Office of Conservation, p. 146.

Lusczynski, N.J., 1961, Head and flow of ground water of variable density: Journal of Geophysical Research, v. 66, no. 12, p. 4247–4256.

Martin, Angel, Jr., and Early, D.A., 1987, Statistical summary of aquifer-test data for Louisiana: U.S. Geological Survey Water-Resources Investigations Report 87–4001, 26 p.

Martin, Angel, Jr., and Whiteman, C.D., Jr., 1989, Geohydrology and regional ground-water flow of the coastal lowlands aquifer system in parts of Louisiana, Mississippi, Alabama, and Florida—A preliminary analysis: U.S. Geological Survey Water-Resources Investigations Report 88–4100, 88 p.

———1999, Hydrology of the coastal lowlands aquifer system in parts of Alabama, Florida, Louisiana, and Mississippi: U.S. Geological Survey Professional Paper 1416–H.

Martin, R.G., 1980, Distribution of salt structures in the Gulf of Mexico—Map and descriptive text: U.S. Geological Survey Miscellaneous Field Studies Map 1213, 8 p., 2 sheets, scale 1:2,500,00.

Matson, G.C., and Sanford, Samuel, 1913, Geology and ground waters of Florida: U.S. Geological Survey Water-Supply Paper 319, 445 p.

Matthews, C.S., and Russell, D.G., 1967, Pressure buildup and flow tests in wells, Appendix G: New York, Society of Petroleum Engineers of the American Institute of Mining, Metallurgical, and Petroleum Engineers, Monograph 1, p. 155–163.

McDonald, M.G., and Harbaugh, A.W., 1988, A modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Techniques of Water-Resources Investigations , book 6, chap. A1, p. 5–1 to 5–24.

McKelvey, J.G., and Milne, I.H., 1962, Flow of salt solutions through compacted clay, *in* Clays and clay minerals—National Conference on Clays and Clay Minerals, 9th, 1960, Proceedings: New York, Pergamon Press, Earth Science Series Monograph 11, p. 248–259.

Mesko, T.O., Williams, T.A., Ackerman, D.J., and Williamson, A.K., 1990, Ground-water pumpage from the gulf coast aquifer systems, 1960–85, south-central United States: U.S. Geological Survey Water-Resources Investigations Report 89–4180, 177 p.

Meyer, W.R., and Carr, J.E., 1979, A digital model for simulation of ground-water hydrology in the Houston area, Texas: Texas Department of Water Resources LP–103, 133 p.

Mississippi Water Resources Research Institute, 1987, 1987 in review: Mississippi State, Mississippi State University Water Resources Research Institute, 13 p.

Neglia, Stefino, 1979, Migration of fluids in sedimentary basins: The American Association of Petroleum Geologists Bulletin, v. 63, no. 4, p. 582.

Neuman, S.P., 1980, A statistical approach to the inverse problem of aquifer hydrology; 3, Improved solution method and added perspective: Water Resources Research, v. 16, no. 2, p. 331–346.

Osborn, C.T., Schefter, J.E., and Shabman, Leonard, 1986, The accuracy of water use forecasts—Evaluation and implications: American Water Resources Association Water Resources Bulletin, v. 22, no. 1, p. 101–109.

Payne, J.N., 1968, Hydrologic significance of the lithofacies of the Sparta Sand in Arkansas, Louisiana, Mississippi, and Texas: U.S. Geological Survey Professional Paper 569–A, 17 p.

———1970, Geohydrologic significance of lithofacies of the Cockfield Formation of Louisiana and Mississippi and of the Yegua Formation of Texas: U.S. Geological Survey Professional Paper 569–B, 14 p.

———1972, Hydrologic significance of lithofacies of the Cane River Formation or equivalents of Arkansas, Louisiana, Mississippi, and Texas: U.S. Geological Survey Professional Paper 569–C, 17 p.

———1975, Geohydrologic significance of lithofacies of the Carrizo Sand of Arkansas, Louisiana, and Texas and the Meridian Sand of Mississippi: U.S. Geological Survey Professional Paper 569–D, 11 p.

Peralta, R.C., Dutram, P.W., Peralta, A.W., and Yazdanian, Amin, 1986, Saturated thickness for drought and litigation protection: Ground Water, v. 24, no. 3, p. 357–364.

Peralta, R.C., and Killian, P.J., 1985, Optimal regional potentiometric surface design—Least-cost water-supply/sustained ground-water yield: American Society of Agriculture Engineers Transactions, v. 28, no. 4, p. 1098–1107.

Peralta, R.C., and Peralta, A.W., 1984, Arkansas ground-water management via target levels: American Society of Agriculture Engineers Transactions, v. 27, no. 6, p. 1696–1703.

Petroleum Information, 1990, Well history control system manual: Petroleum Information, Littleton, Colo., 1176 p.

Pettijohn, R.A., 1996, Geochemistry of ground water in the gulf coast aquifer systems, south-central United States: U.S. Geological Survey Water-Resources Investigation Report 96–4107, 158 p.

Pettijohn, R.A., Williamson, A.K., and Weiss, J.S., 1988, Dissolved-solids concentrations and water temperature for aquifers in the gulf coast aquifer systems, south-central United States: U.S. Geological Survey Water-Resources Investigations Report 88–4082, 5 sheets, scale 1:3,500,000.

Potter, R.W., III, and Brown, D.L., 1977, The volumetric properties of aqueous sodium chloride solutions from 0°C to 500°C at pressures up to 2000 bars based on a regression of available data in the literature: U.S. Geological Survey Bulletin 1421–C, 35 p.

Price, L.C., 1985, A critical overview of and proposed working model for hydrocarbon microseepage: U.S. Geological Survey Open-File Report 85–271, 83 p.

Prudic, D.E., 1991, Estimates of hydraulic conductivity from aquifer-test analyses and specific-capacity data, gulf coast regional aquifer systems, south-central United States: U.S. Geological Survey Water-Resources Investigations Report 90–4121, 38 p.

Prudic, D.E., and Williamson, A.K., 1986, Evaluation of a technique for simulating a compacting aquifer system in the Central Valley of California, U.S.A, *in* Johnson, A.I., Carbognin, Laura, and Ubertini, L., eds., Land Subsidence: International Symposium on Land Subsidence, 3d, Venice, Italy, March 1984, Proceedings, International Association of Hydrological Sciences Publication 151, p. 53–64.

Reed, J.E., 1972, Analog simulation of water-level declines in the Sparta Sand, Mississippi embayment: U.S. Geological Survey Hydrologic Investigations Atlas HA–434, 1 sheet, scale 1:2,500,000.

Reed, J.E., and Broom, M.E., 1979, Digital model of the Bayou Bartholomew alluvial aquifer-stream system, Arkansas: U.S. Geological Survey Open-File Report 79–0685, 37 p.

Reilly, T.E., Franke, O.L., and Bennett, G.D., 1987, The principle of superposition and its application in ground-water hydraulics: U.S. Geological Survey Techniques of Water-Resources Investigations, book 3, chap. B6, 28 p.

Renken, R.A., 1984, The hydrogeologic framework for the Southeastern Coastal Plain aquifer system of the United States: U.S. Geological Survey Water-Resources Investigations Report 84–4243, 26 p.

Robertson Research, 1980, Framework for oil and gas occurrence in the Gulf Coast Tertiary: Robertson Research Proprietary Report, p. 133–139, pls. 10–11.

Rouse, Hunter, ed., 1950, Engineering hydraulics: New York, John Wiley and Sons, 1,008 p.

Ryder, P.D., 1988, Hydrogeology and predevelopment flow in the Texas Gulf Coast aquifer systems: U.S. Geological Survey Water-Resources Investigations Report 87–4248, 109 p.

Ryder, P.D., and Ardis, A.F., 1991, Hydrology of the Texas Gulf Coast aquifer systems: U.S. Geological Survey Open-File Report 91–64, 147 p.

Sampson, R.J., 1978, The Surface II graphics system: Lawrence, Kansas Geological Survey, 240 p.

———1988, The Surface III graphics system: Lawrence, Kansas Geological Survey, 277 p.

Sayre, A.N., 1937, Geology and ground-water resources of Duval County, Texas: U.S. Geological Survey Water-Supply Paper 776, 116 p.

Shepard, E.M., 1907, Underground waters of Missouri, their geology and utilization: U.S. Geological Survey Water-Supply Paper 195, 224 p.

Signor, D.C., and Imes, J.L., 1989, Geohydrology of regional aquifer systems in Cretaceous and older rocks underlying the central United States, *in* Swain, L.A., and Johnson, I.A., eds., Regional aquifer systems of the United States—Aquifers of the midwestern area: American Water Resources Association Monograph 13, p. 149–163.

Smith, E.A., 1907, The underground water resources of Alabama: Alabama Geological Survey Monograph 6, 388 p.

Snowden, J.O., Simmons, W.B., Traughber, E.B., and Stephens, R.W., 1977, Differential subsidence of marshland peat as a geologic hazard in the greater New Orleans area, Louisiana: Gulf Coast Association of Geological Societies Transactions, v. 27, p. 169–179.

Snowden, J.O., Ward, W.C., and Studlick, J.R.J., 1980, Geology of greater New Orleans—Its relationship to land subsidence and flooding: The New Orleans Geological Society, 25 p.

Solley, W.B., Chase, E.B., and Mann, W.B., IV, 1988, Estimated use of water in the United States in 1985: U.S. Geological Survey Circular 1004, 82 p.

Stephenson, L.W., and Crider, A.F., 1916, Geology and ground waters of northeastern Arkansas, *with a discussion of* The chemical character of the water by R.B. Dole: U.S. Geological Survey Water-Supply Paper 399, 315 p.

Sumner, D.M., and Wasson, B.E., 1984a, Summary of results of an investigation to define the geohydrology and simulate the effects of large ground-water withdrawals on the Mississippi River alluvial aquifer in northwestern Mississippi: U.S. Geological Survey Water-Resources Investigations Report 84–4343, 17 p.

———1984b, Geohydrology and simulated effects of large ground-water withdrawals on the Mississippi River alluvial aquifer in northwestern Mississippi: U.S. Geological Survey Open-File Report 84–822, 83 p.

Sumner, D.M., Wasson, B.E., and Kalkhoff, S.J., 1987, Geohydrology and simulated effects of withdrawals on the Miocene aquifer system in the Mississippi Gulf Coast area: U.S. Geological Survey Water-Resources Investigations Report 87–4172, 203 p.

Sun, R.J., 1986, Regional Aquifer-System Analysis program of the U.S. Geological Survey—Summary of projects, 1978–84: U.S. Geological Survey Circular 1002, 263 p.

Sun, R.J., and Weeks, J.B., 1991, Bibliography of Regional Aquifer-System Analysis program of the U.S. Geological Survey, 1978–91: U.S. Geological Survey Water-Resources Investigations Report 91–4122, 92 p.

Taylor, T.U., 1907, Underground waters of the Coastal Plain of Texas: U.S. Geological Survey Water-Supply Paper 190, 73 p.

Terry, J.E., Hosman, R.L., and Bryant, C.T., 1979, Summary appraisal of the Nation's ground-water resources—Lower Mississippi region: U.S. Geological Survey Professional Paper 813–N, 41 p.

Theis, C.V., Brown, R.H., and Meyer, R.R., 1963, Estimating the transmissibility of aquifers from the specific capacity of wells, *in* Bentall, Ray, compiler, Methods of determining permeability, transmissibility, and drawdown: U.S. Geological Survey Water-Supply Paper 1536–I, p. 331–341.

Thorkildsen, David, Quincy, Roger, and Preston, Richard, 1989, A digital model of the Carrizo-Wilcox aquifer within the Colorado River Basin of Texas: Texas Water Development Board Report LP–208, 67 p.

Torak, L.J., and Whiteman, C.D., Jr., 1982, Applications of digital modeling for evaluating the ground-water resources of the "2,000-foot" sand of the Baton Rouge area, Louisiana: Louisiana Department of Transportation and Development Water Resources Technical Report 27, 87 p.

Trudeau, D.A., and Buono, Anthony, 1985, Projected effects of proposed increased pumpage on water levels and salinity in the Sparta aquifer near West Monroe, Louisiana: Louisiana Department of Transportation and Development Water Resources Technical Report 39, 70 p.

Veatch, A.C., 1906, Geology and underground water resources of northern Louisiana and southern Arkansas: U.S. Geological Survey Professional Paper 46, 422 p.

Vecchia, A.V., and Cooley, R.L., 1987, Simultaneous confidence and prediction intervals for nonlinear regression models with application to a groundwater flow model: Water Resources Research, v. 23, no. 7, p. 1237–1250.

Wallace, J.R., and Grant, J.L., 1977, A least squares method for computing statistical tolerance limits: Water Resources Research, v. 13, no. 5, p. 819–823.

Weeks, J.B., and Sun, R.J., 1987, Regional Aquifer-System Analysis program of the U.S. Geological Survey—bibliography, 1978–86: U.S. Geological Survey Water-Resources Investigations Report 87–4138, 81 p.

Weiss, E.J., 1982, A model for the simulation of variable-density ground water in three dimensions under steady-state conditions: U.S. Geological Survey Open-File Report 82–0352, 59 p.

Weiss, J.S., 1987, Determining dissolved-solids concentrations in highly mineralized ground water of the gulf coast aquifer systems using electric logs, *in* Vecchioli, John, and Johnson, A.I., eds., Regional aquifer systems of the United States, Aquifers of the Atlantic and Gulf Coastal Plain: American Water Resources Association Monograph 9, p. 139–150.

———1992, Geohydrologic units of the coastal lowlands aquifer system, south-central United States: U.S. Geological Survey Professional Paper 1416–C, 32 p.

Weiss, J.S., and Williamson, A.K., 1985, Subdivision of thick sedimentary units into model layers for simulation of ground-water flow: Ground Water, v. 23, no. 6, p. 767–774.

Wells, F.G., 1932, A preliminary report on the artesian water supply of Memphis, Tennessee: U.S. Geological Survey Water-Supply Paper 638A, 34 p.

Wesselman, J.B., 1967, Ground-water resources of Jasper and Newton Counties, Texas: Texas Water Development Board Report 59, 167 p.

West, S.W., and Broadhurst, W.L., 1975, Summary appraisal of the Nation's ground-water resources—Rio Grande region: U.S. Geological Survey Professional Paper 813–D, 39 p.

White, W.N., Rose, N.A., and Guyton, W.F., 1944, Ground-water resources of the Houston District, Texas: U.S. Geological Survey Water-Supply Paper 889–C, p. 141–288.

Whiteman, C.D., Jr., 1979, Saltwater encroachment in the "600-foot" and "1,500-foot" sands of the Baton Rouge area, Louisiana, 1966–78 including a discussion of saltwater in other sands: Louisiana Department of Transportation and Development, Office of Public Works Water-Resources Technical Report 19, 49 p.

———1980, Measuring local subsidence with extensometers in the Baton Rouge area, Louisiana, 1975–79: Louisiana Department of Transportation and Development, Office of Public Works Water-Resources Technical Report 20, 18 p.

Whitfield, M.S., Jr., 1975, Geohydrology of the Evangeline and Jasper aquifers of southwestern Louisiana: Louisiana Department of Conservation, Louisiana Geological Survey and Louisiana Department of Public Works Water-Resources Bulletin 20, 72 p.

Williams, T.A., and Williamson, A.K., 1989, Estimating water-table altitudes for regional ground-water flow modeling, U.S. Gulf Coast: Ground Water, v. 27, no. 3, p. 333–340.

Williamson, A.K., 1987, Preliminary simulation of ground-water flow in the gulf coast aquifer systems, south-central U.S., *in* Vecchioli, John, and Johnson, A.I., eds., Regional aquifer systems of the United States, Aquifers of the Atlantic and Gulf Coastal Plain: American Water Resources Association Monograph 9, p. 119–137.

Williamson, A.K., 1989, Possible cause and substantial effects of brine distribution in the gulf coast aquifer systems, south-central United States [abs.]: International Geological Congress, 28th, Washington, D.C., July 9–19, 1989, v. 3, p. 364–365.

Williamson, A.K., Grubb, H.F., and Weiss, J.S., 1990, 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, 125 p.

Wilson, T.A., and Hosman, R.L., 1988, Geophysical well-log database for the gulf coast aquifer systems, south-central United States: U.S. Geological Survey Open-File Report 87–0677, 213 p.

Winslow, A.G., and Doyel, W.W., 1954, Land-surface subsidence and its relation to the withdrawal of ground water in the Houston-Galveston region, Texas: Economic Geology, v. 49, no. 4, p. 413–422.

Winslow, A.G., Doyel, W.W., and Wood, L.A., 1957, Salt water and its relation to fresh ground water in Harris County, Texas: U.S. Geological Survey Water-Supply Paper 1360–F, p. 375–407.

Winslow, A.G., Hillier, D.E., and Turcan, A.N., Jr., 1968, Saline ground water in Louisiana: U.S. Geological Survey Hydrologic Investigations Atlas HA–310, 4 sheets, scale 1:750,000.

Winslow, A.G., and Wood, L.A., 1959, Relation of land subsidence to ground-water withdrawals in the upper Gulf Coast region Texas: Mining Engineer, v. 11, no. 10, p. 1030–1034.

Wolff, R.G., 1982, Physical properties of rocks; porosity, permeability, distribution coefficients, and dispersivity: U.S. Geological Survey Open-File Report 82–16, 123 p.

Wood, L.A., and Gabrysch, R.K., 1965, Analog model study of ground water in the Houston District, Texas, *with a section on* Design, construction, and use of electric analog models, by E.P. Patten, Jr.: Texas Water Commission Bulletin S6508, 103 p.

Yazdanian, Amin, and Peralta, R.C., 1986, Sustained-yield ground-water planning by goal programming: Ground Water, v. 24, no. 2, p. 157–165.

Yeh, W.W-G., and Yoon, Y.S., 1981, Aquifer parameter identification with optimum dimension in parameterization: Water Resources Research, v. 17, no. 3, p. 664–672.

Young, Allen, and Low, P.F., 1965, Osmosis in argillaceous rocks: Bulletin of the American Association of Petroleum Geologists, v. 49, p. 1005–1007.

Zurawski, Ann, 1978, Summary appraisals of the Nation's ground-water resources—Tennessee Region: U.S. Geological Survey Professional Paper 813–L, 35 p.

SUPPLEMENT: SIMULATION DETAILS

SIMULATION DETAILS

Details of the methods and data used to prepare the simulations are included in this section. The information provided here is helpful for making an evaluation of the simulation techniques used in the study and provides a basis for continued research.

DIFFERENCES IN SIMULATION FROM PRELIMINARY REPORT

Most of the details of data sources and compilation and other simulation items are contained in a preliminary report by Williamson and others (1990). All of the model input and output files as well as original data files were archived in ASCII files on a WORM (Write Once, Ready Many) disk by Kirkpatrick (1993). This section will only document differences between simulations presented in this report and those from the preliminary report. The differences include transient simulations with both elastic and inelastic (compaction) storage, areally variable hydraulic conductivity related to sand percentage and depth, and the use of automatic parameter-estimation methods.

TRANSIENT FLOW UNDER ELASTIC AND INELASTIC STORAGE CONDITIONS

The preliminary report (Williamson and others, 1990) described several simulations, most of which assumed steady-state conditions. Most of the simulations described in this report, except those for predevelopment conditions and estimating potential for development, assume transient conditions where the pumpage and possibly other factors are changing between each pumping period. This necessitates estimates of the storage properties of the aquifers. Three types of aquifer storage are possible: water-table storage (elastic) where the change in storage represents the upper part of the aquifer being dewatered and sometimes wetted again; elastic confined storage, which occurs at any point where the aquifer is confined; and inelastic storage with irreversible compaction of fine-grained beds, which provides a source of water to wells. Only the latter two types were directly simulated in this study. In most areas, little is known about the changes in water-table altitudes over time because few studies have attempted to measure them. It was beyond the scope of this study to collect data on water-table altitudes through time because of the large area and complexity of the aquifer system. Instead, change in water-table storage was ignored in the model, implying that some part (probably small) of the water volume simulated as regional recharge actually comes from water-table storage.

Kuiper (1994, table 4) found in his parameter-estimation study that the model calibration was most representative of observed data when the specific storage value of 0.51×10^{-6} ft^{-1} was used. Specific storage multiplied by the thickness of the aquifer or permeable zone equals the storage coefficient. Kuiper found that this was a relatively sensitive parameter in his simulations and that over a broad range of parameterization approaches (Kuiper, 1994, table 4), the regression always indicated specific storage values at the smallest limit without violating physical principles. The only parameters more sensitive were hydraulic conductivity of sand and hydraulic conductivity of fine-grained beds. Kuiper's parameter-estimation method will be discussed in the "Summary of Parameter-Estimation Results." Elastic storage in the confining units is assumed to be zero in the simulations, an assumption that should have little effect on the simulation results because the total volume of regional confining units is only about one-fifth of the total volume of sediments in the aquifer systems studied. However, inelastic storage in confining beds, which contributes to land subsidence, is simulated.

PUMPAGE ESTIMATION

Pumpage was estimated for 1960–85 at 5-yr intervals using a distribution to model blocks based on detailed pumpage estimates for 1980 and the total county pumpage for the respective year (Mesko and others, 1990, p. 25–29). Some uncertainty is introduced by the assumption that the percentage distribution of pumpage within a county has been equal to the 1980 distribution over the 1960–85 period. Exceptions to this assumption were made in the central Arkansas and Houston, Texas, areas because the distribution was known to change over time (Mesko and others, 1990, p. 28–29). Two categories of ground-water pumpage data were distinguished in the proration of county data to model blocks:

1. Point pumpage, primarily water pumped for public supply and industrial uses, where little water returns to the aquifer system; and
2. areally dispersed pumpage, primarily for irrigation, fish farms, stock, and rural domestic uses, where there is a potential for a substantial amount of return flow to the aquifer system.

Point-pumpage data for 1980 were tabulated for each of about 5,000 municipal and industrial wells, along with the well's 5-mi block location, depth, and model layer, if known. These data included annual (1976–85) withdrawals for Harris and Galveston Counties in Texas, where the law requires each of about 3,000 wells to be metered and reported (obtained on computer disk from Ronald J. Neighbors, Harris-Galveston Coastal Subsidence District, written commun., 1985).

The wells in the Mississippi embayment aquifer system and Texas coastal uplands aquifer system were manually assigned to appropriate model layers. The wells in the coastal lowlands aquifer system were assigned using a FORTRAN program that used all of the 5-mi 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-mi 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 was calculated to be in a confining unit, it was assigned to the closest adjacent aquifer layer.

Estimation of areally dispersed pumpage is generally subject to greater errors than the point pumpage because there are many more wells, most of which are not metered, and well locations are usually not as accurately known as industrial or municipal well locations. Published data were used where available, generally as totals by county (references given by Mesko and others, 1990). Estimates for agricultural use were generally based on irrigated areas, types of crops, and typical crop water-use estimates. Rural domestic-use estimates generally were based on the population not served by public systems and per capita use estimates. County total pumpage was apportioned to 5-mi blocks by using estimates of the distribution to each block and layer in the county based on well records and local knowledge of water-use patterns.

The transient simulations were made using the time periods and pumpage outlined below, with predevelopment simulated hydraulic heads as initial conditions. The pumpage factors before 1960 were extrapolated by simple linear regression using 1960, 1970, and 1980 pumpage totals and 1930–80 population data.

Transient simulation time periods

Period ended	Length (in years)	Pumpage used
1925	---	None
1935	10	1960 pumpage multiplied by factor 0.149
1945	10	1960 pumpage multiplied by 0.310
1955	10	1960 pumpage multiplied by 0.569
1962	7	1960
1967	5	1965
1972	5	1970
1977	5	1975
1980	3	1980
1982	2	1980
1987	5	1985

ESTIMATION OF OBSERVED 1962–87 HYDRAULIC HEAD

Estimates of hydraulic head for model blocks at 5-yr intervals were needed for comparison with simulated heads. Over one-half million water-level measurements taken from data on nearly 20,000 wells stored in the U.S. Geological Survey's WATSTORE data base and including over 100,000 measurements from the Texas Natural Resources Information System files (Thomas A. Williams, U.S. Geological Survey, written commun., 1987) were combined into one data base. Each well in the data base was assigned a model layer by a computer program using the location, altitude, and depth of the well. A considerable amount of error screening was done by several automatic methods in order to delete data that were obviously in error. In addition to internal checks such as missing or impossible data values, well locations were checked by computer program to test if they were in or near the county specified in the data base. About 1 percent of the wells, which contained about 0.3 percent of the measurements, were not used because of failure to pass this location check. Land-surface altitudes were checked to test if they were within a reasonable range of the 5-mi land-surface altitude interpolated to the well location. About 1.6 percent of the wells, which contained about 1 percent of the measurements, were ignored because of this land-surface altitude check. The data were then plotted as hydrographs, with many wells plotted on a page, to check for extreme outliers.

The desired intervals for head data were offset from the pumpage data by 2.5 yr. For example, pumpage data for 1980 were assumed to approximate the average for the period 1977–82, and simulation with 1980 pumpage should produce water-level changes from 1977 to 1982. All of the measurements for each well during the 2-yr time period on either side of the end of the pumping period (1981–83) were used to calculate an average head for the well at the end of the pumping period. The number of wells measured during each period varied from about 4,000 in 1957 to a maximum of about 9,500 in 1972, decreasing to about 7,000 in 1982. The 2-yr well averages were used to interpolate heads for each block that had sufficient head data in and near it.

Using 1982 as an example, the combined data base contained over 37,000 head measurements taken in over 7,000 wells between July 1981 and June 1983. Obviously, some wells were measured several times in that time period, and if so, the measurements for those wells were averaged.

As is generally the case, the numerical flow model used in this study produces an approximation to block-averaged head values that are not equal to values for head at a single point in the aquifer system (Kuiper, 1994, p. 55). Observed hydraulic-head data were averaged for each block to assure comparison of values of head expected to be representative of what the model computes. Ten-mile block-averaged head values were estimated using an inverse distance-squared weighted-averaging scheme of the SURFACE III computer program (Sampson, 1978, 1988). Several different neighbor selection criteria (to select which data points will be used to estimate a grid cell value) and gridding algorithms were attempted. Two different gridding algorithms were compared, one using a simple distance-weighted average of nearby sample values (Sampson, 1988, fig. 13) and the other using distance-weighted averages of the projection of dips of the surface at the nearby values (Sampson, 1988, figs. 14, 15; Sampson, 1978, fig. 10). The projection of dip averages would be the best estimate of the observed head surface at the center of the block, whereas the simple averages would approximate the average head for the block. The root mean squared error (RMSE) between these two sets of block-averaged observed heads was 65 ft. The RMSE between simulated 1982 heads and dip projection-averaged observed heads was 85.7 ft, and for the simple averages, 57.4 ft. This indicates that the simple-average head, approximating the block-averaged head, estimates values that compare better with the model-computed heads than those the method designed to estimate the head value at the center of the model block. The RMSE between different sets of neighbor selection criteria for simple block-averaged observed heads was about 25 ft. These RMSE values for different estimates of observed heads are large because the horizontal and vertical hydraulic-head gradients are commonly very steep in the vicinity of pumping centers and there are many abrupt changes in land-surface altitude, causing great difficulty in estimating an average head for a 100-mi^2 block for a layer. All model blocks that 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 for 1982 included 1,110 blocks among the 10 permeable layers. Kuiper (1994, p. 55–58) used the same raw data with different culling criteria and averaging method to come up with water-level estimates for 1,432 blocks for 1972 and 1,675 blocks for 1982.

DEEP PRESSURE DATA

Nearly all of the hydraulic-head measurements mentioned above are taken from the upper parts of the aquifer systems, which contain relatively fresh water. Calibration of aquifer characteristics in the deep, brine parts of the aquifer systems (which comprise about two-thirds of the aquifer-system volume) is dependent on the availability of pressure or head estimates. More than 40,000 final shut-in pressure measurements from drill-stem tests were available in a file subset from the Petroleum Information file (1990). Several problems complicate the use of these data. Many of the pressure measurements are affected by oil and gas withdrawals. Some measurements have low pressures because not enough time was allowed after the well was shut in for the pressure to build up to static condition or because of equipment problems, leaking packers, and so forth. Deeper measurements, near the bottom of the systems studied, are from near the geopressured zone where the fluid pressures are in transition from hydrostatic to geopressure and are much greater than immediately above. Pressures in the geopressured zones are commonly high enough to make equivalent freshwater heads several thousand feet above sea level. A few pressure measurements from the transition zone can greatly affect any areal averages or interpolations.

Culling the pressure data by retaining those having equivalent freshwater heads between 300 below to 1,500 ft above sea level reduced the number of measurements to about 19,500. The lower bound was chosen to retain data affected by head declines resulting from freshwater pumpage and to eliminate most of the data strongly affected by oil pumping or test deficiencies. The upper bound was set to a few hundred feet above the maximum head calculated from the hydrostatic gradient in the modeled area to eliminate pressures affected by the transition to the geopressured zone. All but 8,933 of these pressure data were from below the aquifer systems modeled. Using a similar procedure as mentioned above for freshwater well data, equivalent freshwater heads were interpolated for 288 blocks that had at least seven pressure measurements each. There was generally very large variability among the seven or more pressure measurements for a single block.

Kuiper (1994, p. 55) used the same raw data with different culling criteria and averaging method to estimate deep heads for 586 blocks. Including the deep head estimates did not materially affect the parameter estimation results (Kuiper, 1994, p. 94). This is probably because even with the culling and averaging, the data are still quite noisy and cannot be simulated well by the model regardless of the parameters used.

AREALLY VARIABLE HYDRAULIC CONDUCTIVITY RELATED TO SAND PERCENTAGE AND DEPTH

Most of the details about the methods and results of estimating effective hydraulic conductivity, K_{eff} are in the body of this report. However, Kuiper (written commun., 1987) did test other areal groupings for potential improvement in the model calibration. Kuiper found that the area boundaries (fig. 13) used as presented in this report produced a model with better fit and narrower confidence limits on parameters than with any other of several combinations.

Kuiper also tested the exponent used in the power function relating sand percentage to effective hydraulic conductivity, K_{eff}. He found that the model was quite sensitive to this parameter. Under at least 10 different schemes of parameter choices, the model consistently estimated a value within the range of 3.52 to 4.36 for the exponent of horizontal K and the reciprocal of those values for the vertical K (Kuiper, 1994, table 4).

The function relating K_{eff} to depth was not a very sensitive parameter in the model according to Kuiper (1994, p. 73). This is probably because most of the reliable head data was quite shallow in the aquifer system compared to the total depth of the system modeled.

APPROACH TO CALIBRATION AND PARAMETER ESTIMATION

The purpose of the modeling in this study was to understand regional-scale flow. Therefore, no attempt was made to alter the parameters for an individual block or even groups of blocks unless data errors were found. Small improvements in the model fit to historical observed data would not substantially improve the understanding of the system. Instead, more emphasis was given to the design of the approach to estimating model parameters such as hydraulic conductivity to maximize the information content of the available data. An additional reason for not attempting alteration of individual model block parameters during calibration was the size and complexity of the system modeled.

Manual model calibration is an iterative process similar to the one used by parameter-estimation models to find the combination of parameter values that gives the best fit to available data. Automatic approaches by computer programs have the advantage of being able to test many more values 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 constraining the parameter selection to within reasonable ranges.

METHOD TO ESTIMATE AREAL SENSITIVITY TO PUMPING FOR DEVELOPMENT POTENTIAL

A starting head equal to zero was used for these simulations for easy comparison. Therefore, drawdown is equal to the opposite sign of hydraulic head. Running the simulation to steady state gives quicker runs and the full and final effects of the pumpage. If subsidence was simulated in the sensitivity to pumpage runs (described previously in the text), areas of potential subsidence would look more favorable for development than would be true for a steady-state condition, due to less drawdown.

To analyze effects of pumping individual blocks, interference of nearby pumped blocks must be minimized. In the first attempt at the sensitivity to pumpage runs, every fourth block in every other layer was pumped. The pumped blocks in every other layer were offset two blocks, so the nearest interference from a neighboring pumping block was at least four blocks away, horizontally and vertically. Enough interference was noted that a new design was developed that pumped only one layer at a time and only every ninth block in each direction. This took 810 simulations, but because the simulations were all steady state, computer run time was small and the whole group of simulations executed in one night. Every model block with an average concentration of dissolved solids less than 10,000 mg/L was included in the analysis, a total of 7,324 blocks.

LIMITATIONS OF THE SIMULATION

Uncertainty is present in every scientific finding. This section will discuss some of the known sources of uncertainties in the findings of this study.

MODEL FIT

Hydraulic head is the only model-output quantity that is observable because of the previously mentioned accuracy problems in estimating stream gains and losses or other flows. Comparison of simulated with observed hydraulic head is a very complicated process when so much observed data are available and the system simulated is so complex. Data could be compared at six time periods for each of 10 layers in tens or hundreds of blocks in each layer, totaling many thousands of comparisons. A summary of comparisons for 1980 is shown in table 6 because the most observed-head data were available for that year. The model fit is adequate considering the method and approach to calibration. No attempt was made to alter the parameters for individual blocks or even groups of blocks unless data errors were found, as described in the previous section on calibration.

The overall model-fit root mean squared error (RMSE) is only slightly better than it was in the preliminary model (comparing table 6 with Williamson and

TABLE 6.—*Comparison of simulated hydraulic head to estimated block-averaged head for 1980, in feet*

[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			Mean absolute value error	Root mean-squared error
	Estimated heads	Model high	Model low	Estimated head	Model head	All blocks	Model high	Model low	Standard deviation	Minimum	Maximum		
2	24	13	11	209.9	225.5	15.6	39.7	12.9	35.3	–25.9	106.4	27.4	37.9
3	105	79	26	283.6	306.2	22.6	39.2	27.9	47.0	–88.4	202.3	36.4	51.9
4	81	56	25	272.1	287.8	15.7	32.8	22.7	43.8	–209.7	160.5	29.7	46.3
5	170	125	45	139.6	179.2	39.6	61.2	20.2	58.0	–75.5	245.4	50.3	70.1
6	55	28	27	138.0	149.4	11.4	43.4	21.9	39.3	–55.1	91.0	32.8	40.6
MEAS and TCUAS	435	301	134	202.7	228.9	26.2	47.6	21.9	50.8	–209.7	245.4	39.7	57.1
7	37	30	7	165.0	191.6	26.6	38.3	23.5	35.3	–52.5	103.7	35.5	43.8
8	69	45	24	148.8	159.4	10.6	37.3	39.6	52.0	–248.8	115.6	38.1	52.7
9	96	62	34	51.2	69.6	18.4	47.2	34.1	52.6	130.8	187.3	42.6	55.5
10	96	72	24	9.2	44.2	34.9	54.8	24.7	52.6	–58.3	164.4	47.3	62.9
11	345	164	181	43.7	48.7	5.1	27.5	15.3	31.3	–61.9	176.1	21.1	31.7
CLAS	643	373	270	57.9	71.3	13.3	38.1	20.9	42.7	–248.8	187.3	30.9	44.7
GC RASA	1,078	674	404	116.3	134.9	18.5	42.3	21.2	46.5	–248.8	245.4	34.4	50.1

others, 1990, table 5), with a few exceptions. Confirming what Konikow (1978) reported—that is, most improvement in model fit is obtained in the early stages of model calibration. There is improvement in the overall standard deviation of the errors from 56.5 ft to 46.5 ft. However, there is an increase in the mean error, from 7.4 to 18.5 ft, which causes the RMSE to be only slightly less, 56.9 to 50.1. The distribution of errors shows that there is some bias to the errors (fig. 54). The leakance (hydraulic conductivity multiplied by the flow area divided by the flow-path length) between the source sink layer and the top model layer was adjusted to reduce the mean error, but as mean error decreased, the RMSE increased. Kuiper adjusted for this by decreasing the estimated water-table altitude used in the source and(or) sink layer by a factor ranging from 0.834 to 0.932 in the various model runs reported (Kuiper, 1994, table 4). This adjustment was not made in the simulations presented in this report because no physical explanation for this kind of adjustment is known and because it would only decrease the measurement of error without producing other significant improvements. The model fit was improved over the preliminary model for layers 2, 9, and 11, was worse for layers 5, 4, and 3, and was similar for layers 6, 7, 8, and 10.

The details of the model fit are discussed, layer by layer, by comparing contours of simulated hydraulic head with posted values of block-averaged estimated head, as shown on plate 7*A*–*J*. The block-averaged estimated heads are not contoured because that is not possible without including further averaging and extrapolation. The block-averaged heads were estimated using the Surface III gridding algorithm described previously. Generally, the discussion is focused at locations where the model fit is less than ideal. At some locations, it appears that the simulated drawdown is not coincident with the estimated drawdown. This is usually due to the independent discretization of pumpage and estimated head data. Both data sets have some additional error introduced that is inherent to the discretization process. The values in one data set may be shifted in one direction to the nearest block, whereas the values in the other data set may be shifted in the other direction, magnifying the apparent error. An example is shown on plate 7*G* where pumping at Nacogdoches, in eastern Texas, is shifted to the nearest 10-mi block to the south (estimated head = −178). A large simulated drawdown is shown in a slightly different location than the maximum block-averaged drawdown, which was shifted to the nearest block center north of the locations of estimated well data.

The simulated 1980 hydraulic head for layer 11 matches the estimated head in permeable zone A somewhat better than the RMSE for the whole model (table 6) because the layer is relatively shallow, and the estimated drawdown is not very large. The model underestimates the depth of the drawdown in several areas (pl. 7*A*). A steep drawdown identified at New Orleans is not simulated very well because the cone is small and because the existing wells pump only from the top part of the layer where the transmissivity is effectively smaller (C.D. Whiteman, U.S. Geological Survey, oral commun., 1989). The model also underestimates the drawdown in southwestern Louisiana and the vicinity of Houston, Texas.

In Louisiana and Mississippi, the simulated 1980 hydraulic head in layer 10 generally matches the estimated head in permeable zone B (pl. 7*B*). The simulated drawdown around Houston is large and quite deep, but not as deep as the observed cone. The simulated drawdown near Kingsville (the first area north of Mexico with abundant estimated heads) is also not as deep as estimated data indicate (pl. 7*B*). The situation is similar to that at New Orleans, where the cone is also quite small and the pumping is from part of the layer, making simulation difficult.

The simulated 1980 hydraulic head for layer 9 generally follows the estimated head estimates in permeable zone C (pl. 7*C*). Simulation of the cones of depression along the coast at Gulfport and Pascagoula, Mississippi, and Baton Rouge, Louisiana, generally matches the observed cones except for some differences in shape. The very large cone at Houston has a few more differences in shape. There are some minor differences in the simulated heads farther south in Texas.

Less pumpage and therefore fewer and smaller cones of depression are present in the 1982 heads for layers 6, 7, and 8 (pls. 7*D*, *E*, *F*). In and near the outcrop areas of the layers, the simulated water levels closely approximate observed water levels except where local variations in the land-surface altitude are strongly reflected in observed water levels. The maximum drawdown at Alexandria in central Louisiana is simulated by the model in layer 8, whereas the maximum observed block-averaged drawdown is in layer 7 (fig. 36*G*). The failure to match drawdowns is because layer 8 is very thin at this location, resulting in a discretization error in allocating excessive pumpage to layer 8.

The model overestimates the 1980 head in layer 8 below the very large pumping at Baton Rouge. The cone along the Mississippi Gulf Coast is offset somewhat from the estimated drawdown but is similar in magnitude. The model underestimates drawdown at some areas in layer 7 in southern Mississippi. Layer 6 heads generally are simulated well by the model except for underestimating the drawdown in central Mississippi.

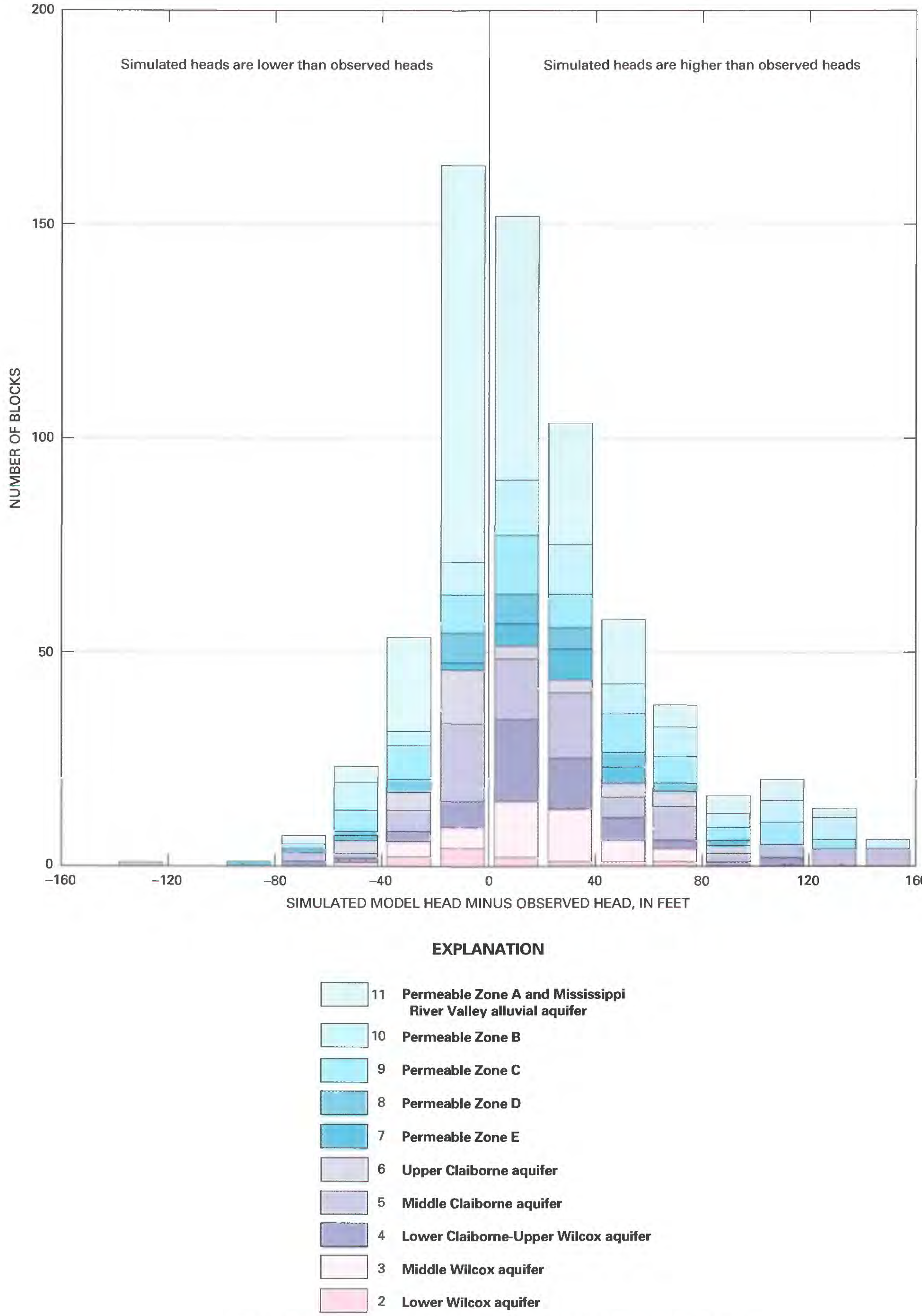


FIGURE 54.—Comparison of simulated and estimated block-averaged head, 1980.

The model fit is poorest in layer 5 (table 6), especially in southern Arkansas and northern Louisiana where the estimated hydraulic heads vary greatly from one block to the next and the simulated heads are generally too high (pl. 7*G*). In this area, layer 5 contains a substantial confining unit, effectively dividing it into two permeable zones with nearly all of the pumpage from the upper zone. The upper zone is much thinner than the whole unit, causing more actual drawdown (due to the reduced transmissivity) than is simulated in the model, which uses the transmissivity of the entire thickness of layer 5. The simulated heads near Memphis, Tennessee, match the estimated heads well.

In layer 4, the simulated hydraulic heads generally match the estimated heads in the Mississippi embayment aquifer system but underestimate the drawdown in the Winter Garden area of southern Texas (pl. 7*H*). The fit (RMSE) is poorest in the western part of La Salle County and adjacent Dimmit County, an area with some of the lowest estimated heads. The trough of lowest simulated head is much farther to the east, possibly due to errors in the distribution of observed pumpage.

In layer 3, most of the estimated hydraulic-head data are in eastern Texas, where the estimated head varies considerably from one block to the next (pl. 7*I*). The simulation for eastern Texas is neither consistently high nor low. However, due to the varied nature of the estimated heads, the model fit is poor at some of the locations.

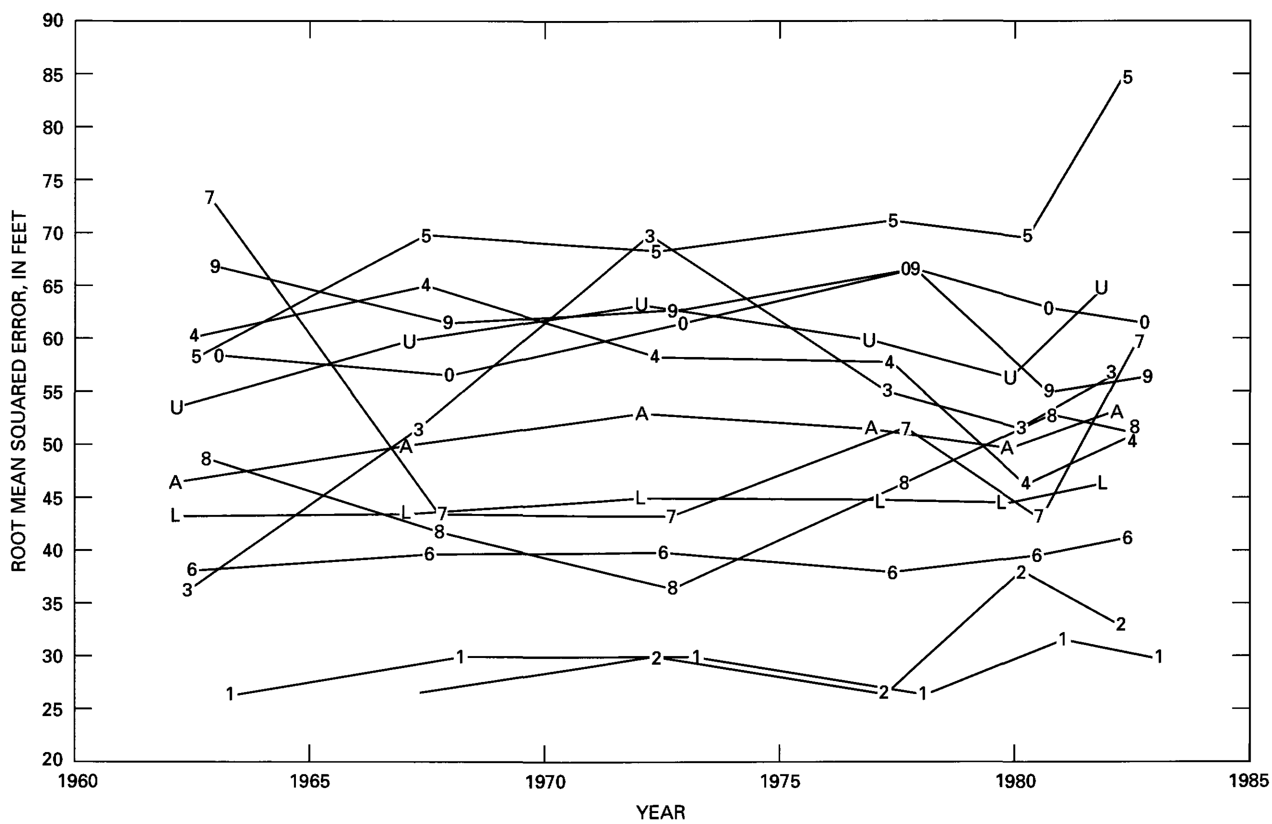
The estimated hydraulic heads are matched best (lowest RMSE) by the simulated heads in layer 2 (table 6). Most of the estimated block-averaged heads are within 15 ft of the model-simulated head for the same location for 1980 (pl. 7*J*).

Another check of the accuracy of the simulation is to examine model fit over time for any trends that would indicate the model did not simulate some time-sensitive feature of the aquifer system. Usually, this is done by comparing a few available well hydrographs with simulated hydrographs for the nearest block. The large model blocks of this regional model preclude comparing individual well hydrographs because the model simulates many hydrologic complexities within each model block. An alternative approach was chosen, comparing block-averaged estimated heads, wherever and whenever they existed, with the equivalent model block head and then plotting different statistics of the fit over time. Simulated trends exhibit little bias over time (fig. 55). The RMSE is plotted for each layer and groups of layers in figure 55. All of the statistical measures included in table 6 were plotted similarly to figure 55. The mean differences were steady over time. The statistics that measure variability in errors, like mean absolute error and standard deviation, showed trends in time very similar to the ones shown in figure 55. In some layers the model fits the observed data consistently better, and in other layers the fits are consistently poorer. The estimated hydraulic-head data used to construct figure 55 include more than 5,000 individual block-year comparisons for the six time periods shown, with about 500 blocks compared in 1962, increasing to a maximum of 1,078 blocks in 1980, and decreasing slightly in 1982. Additionally, the block-averaged estimated and simulated heads for all layers in 10 selected blocks having large drawdowns are shown in figure 36.

SUMMARY OF PARAMETER-ESTIMATION RESULTS

Kuiper's (1994) parameter-estimation results are summarized here. Parameter values from Kuiper's 11-parameter model (model number 9) (Kuiper, 1994, table 4) were chosen for simulations presented in this report. Kuiper (1994, p. 75) states that the best model to predict future hydraulic heads and estimates of effective conductivity was his model using four parameters. Those four parameters were horizontal hydrologic conductivity of sand, vertical hydraulic conductivity of fine-grained deposits, specific storage, and a multiplier for the water-table head used for the constant-head source and(or) sink layer. Additional parameters chosen for this report were used to divide up the Kh for sand into six groups divided by combinations of area and layer, ordered by estimated Kh from aquifer tests (Prudic, 1991, table 5, fig. 15). The layer-area combinations having the largest estimated Kh were combined in group 1, and so on, through the six groups with successively smaller values of Kh. Kuiper's model 9, chosen for use in this study, also included two groups of Kv, one for fine-grained deposits in the regional confining units and one for the fine-grained deposits contained within the regional aquifers and permeable zones. The only considerable change from Kuiper's model 9 is that the simulations presented in this report do not use a multiplier for the water-table head used in the constant- head source and(or) sink layer. Use of this parameter would have improved the fit of the simulated to estimated hydraulic head by lowering the mean difference from 18.5 ft (table 6) to 3.7 ft (Kuiper, 1994, table 4). However, little else would have changed and there is no physical explanation for use of this multiplier. For comparison purposes, note that Kuiper's weighted RMSE (36.4 ft for model 9) is equivalent to an RMSE using nonweighted head data alone of 51.6 ft.

Part of the difference in model fit in this study compared to Kuiper's is due to Kuiper's culling of estimated-head data that appear to be outliers because of steep cones of drawdown. His culling criteria only



EXPLANATION

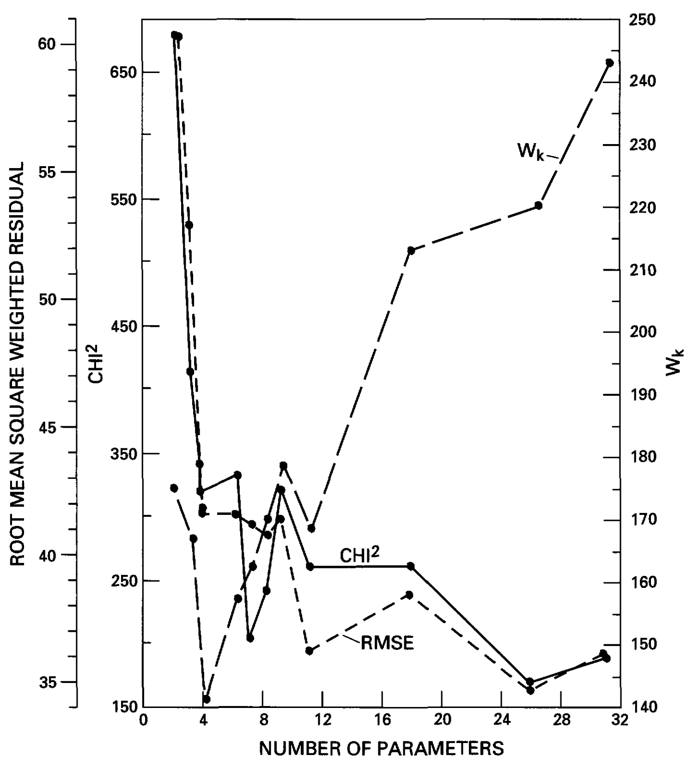
Layer	Geohydrologic unit
6	Upper Claiborne aquifer
5	Middle Claiborne aquifer
4	Lower Claiborne-Upper Wilcox aquifer
3	Middle Wilcox aquifer
2	Lower Wilcox aquifer
U	All layers in Mississippi Embayment and Texas Coastal Uplands aquifer system
11	Permeable Zone A and Mississippi River Valley alluvial aquifer
10	Permeable Zone B
9	Permeable Zone C
8	Permeable Zone D
7	Permeable Zone E
L	All layers in Coastal Lowlands aquifer system
A	All layers

FIGURE 55.—Model fit as a function of simulation time, 1962–82, by layer and groups of layers.

allowed data showing head gradients of 35 ft/mi or less (Kuiper, 1994).

The confidence limits (W_k) of predicted head of model 9 are 18 percent worse than the confidence limits for model 4 (fig. 56), which is not good. However the Chi^2 value is lower and closer to a value that would prove that the errors were normally distributed and therefore random. In both cases, the Chi^2 value is too high to accept the hypothesis of normally distributed errors, introducing some uncertainty about which model is the best to choose (Kuiper, 1994, p. 75–76). Kuiper's 11-parameter model was chosen for use in this report because the RMSE and Chi^2 were less than for the four-parameter model, and the W_k value was not much greater (fig. 56).

Approximately 95 percent of the prediction interval half-widths for volume-averaged freshwater heads exceed 108 ft. This means that about 95 percent of the predicted freshwater-head values would be within 108 ft of the actual head value. For volume-averaged base 10 logarithm hydraulic conductivity, the prediction interval half-widths exceed 0.89 (Kuiper, 1994, table 4). All of the models are unreliable for the prediction of head and ground-water flow in the deeper parts of the aquifer system, including the amount of flow coming from the geopressured zone beneath the aquifer system. A simulation was done to test the effect of assuming freshwater flow with a no-flow boundary at the brine interface. This was done by truncating the domain of solution to exclude that part of the system having a ground-water density greater than 1.005 g/cm^3 and setting the density to that of freshwater in the remaining shallow part. This change does not appreciably change the results for head and ground-water flow from the model, except for locations close to the truncation surface (Kuiper, 1994).



EXPLANATION

W_k = confidence limits
CHI^2 =
RMSE = root mean squared error

FIGURE 56.—Results of parameter estimation by regression, comparing number of parameters with statistics of fit.

SCALE EFFECTS

The scale chosen to simulate ground-water flow systems has a significant effect on several aspects of the study, including recharge and discharge estimates, effective hydraulic conductivity as described in the body of this report, and water density. The following discussions are provided to familiarize the reader with the basis for these differences in order to properly interpret the conclusions presented in this report.

RECHARGE AND DISCHARGE ESTIMATES

As was stated in the body of the report, out of necessity, the only flows simulated were those that crossed block faces. Flows within a single model block were not considered (fig. 30). This has a sizable effect on estimates of recharge and discharge because the small-scale local flow systems commonly have large recharge and discharge. The head gradients in these small-scale systems are steep due to short flow paths. These factors are important considerations when comparing flows described in this report with those of smaller scale studies.

DENSITY

As mentioned previously, the distribution of concentration of dissolved solids has a very large degree of local as well as regional variability, both in the vertical and horizontal directions. To limit the effect of local variations in the point data used to estimate the regional trends of density, the density data were smoothed using an areal moving average as described by Williamson and others (1990, p. 103). Kuiper (1994) found that smoothing the density data had almost no effect on the parameter-estimation fit or prediction. The smoothed density-data sets were used in all simulations presented in this report in order to make the best approximation at a regional scale.

BOUNDARIES

Boundary conditions are commonly a source of uncertainty in ground-water simulations (Franke and others, 1987). Simulating a very large portion of the entire regional aquifer system minimizes the uncertainty due to boundaries. The following description is provided to alert the reader to limitations in applying the results of this study near the boundaries where true boundary conditions in the aquifer system were simulated as nearly as was practical.

LATERAL

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 fine-grained clastic deposits of the Vicksburg-Jackson confining unit gradually change into the 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 the study area. This boundary was treated as a no-flow boundary for the preliminary simulations (Williamson, 1990, p. 36).

Transient simulations highlighted the need to change the eastern boundary from a no-flow boundary to some other type; otherwise, the model would incorrectly simulate drawdowns of hundreds of feet near Mobile, Alabama. The eastern boundary was changed to a constant head boundary in layers 3, 5, 7, and 8, a poor approximation of conditions near this boundary but an improvement compared to a no-flow boundary.

The only data available on the gulf coast regional aquifer systems in Mexico was topography and a surficial geology map (Direcciôn 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 condition of some kind 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.

TOP RECHARGE AND DISCHARGE BOUNDARY AND WATER-TABLE STORAGE

The constant-head layer simulating regional recharge and discharge was used in this study because estimates of the value of the net recharge and(or) discharge and its components are not 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. The flow into or out of the source and(or) sink layer was simulated by the model using the difference in head between the first active aquifer block in the appropriate outcrop layer and the constant-head value above it multiplied by the conductance of 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 one-half of the constant-head block. For simplicity, the vertical conductance for the constant-head layer is uniform all over the study area ($Kv = 3 \times 10^{-5}$ ft/d and thickness = 1 ft) and simulates resistance to water entering or leaving the aquifer such as streambeds, the soil zone, lake bottoms, and so forth. The constant-head layer accounts for ground-water and surface-water interaction from all surface sources and sinks such as rainfall that has infiltrated into the soil, seepage into or out of lakes and streams, and evapotranspiration from ground water.

The method works relatively well for estimating net recharge and discharge for steady-state conditions, especially because there are no alternative methods that will work at the scale of the model. Other methods frequently estimate net recharge and discharge values with the wrong signs—that is, recharge where there should be discharge, or the reverse. This is because, in a relatively humid climate with large streams crossing the aquifer system, the magnitude of all surface volumes of water is so large compared to the regional ground-water component that the regional ground-water component is usually of the same order of magnitude or much smaller than the error component of surface-water measurements. Under transient conditions, the source-sink layer method used can overestimate the volume of regional recharge unless it is limited to a maximum reasonable amount. Under the range of observed stressed conditions, the simulation required amounts of regional recharge that were possible, given the rainfall and location of perennial streams.

An area of research that has not been pursued in the Gulf Coastal Plain is the subject of water-table change over time. The method used for simulating regional recharge and discharge in this report has a limitation because the constant-head source-sink layer tends to restrict head changes in the top aquifer layer and because some water is simulated as increased recharge or decreasing discharge, which really is withdrawal from water-table storage. Therefore, those two water-budget components cannot be calculated separately. Few investigations in the study area have addressed the changes in water levels in shallow wells, which would indicate how much change there has been in water-table storage. A much better understanding of the way the ground-water system responds to stress would be possible if locations with a declining or rising water table were identified. In the Houston area, there appear to be very large areal differences in the depth to the water table as of 1991 (Dana Barbie, U.S. Geological Survey, written commun., 1994), probably resulting from increased downward flow in some areas of lower vertical resistance.

BOTTOM NO-FLOW BOUNDARY

The effects and likelihood of the bottom boundary acting as a no-flow boundary or as an upward leakage boundary from the abnormally high geopressured zone below the aquifer systems in the downdip areas were tested by Kuiper (1994). He found that, based on known data and the simulation results, the upward flow could be as high as 108 ft^3/d, or about 7 percent of the total regional recharge in 1982 (Kuiper, 1994, table 5) without materially affecting the fit for the rest of the aquifer system, mostly as observed in the shallow part. A flow of this magnitude is possible with an upward vertical-head gradient of several thousand feet of head and could probably be maintained for about 100 million years, assuming a large value of specific storage in the geopressured zone (Kuiper, 1994). Flow from the geopressured zone could be ignored without materially affecting the model fit of the shallow head data, and the overall model fit was not very sensitive to two different types of model truncation (Kuiper, 1994). The truncations were to test the effects of the brine part of the system on the freshwater part of the system without modeling the density effects. One truncation was at the shallowest depth where water density was 1.005 g/cm^3. The other truncation was at a fixed depth of 3,000 ft. Although the overall model fit was not very different when treating the saline/freshwater interface as a no-flow boundary, the flows near the boundary were substantially different. Obviously, treating the saline/freshwater interface as a no-flow boundary would grossly underestimate the potential for saltwater intrusion into aquifers that previously held freshwater.

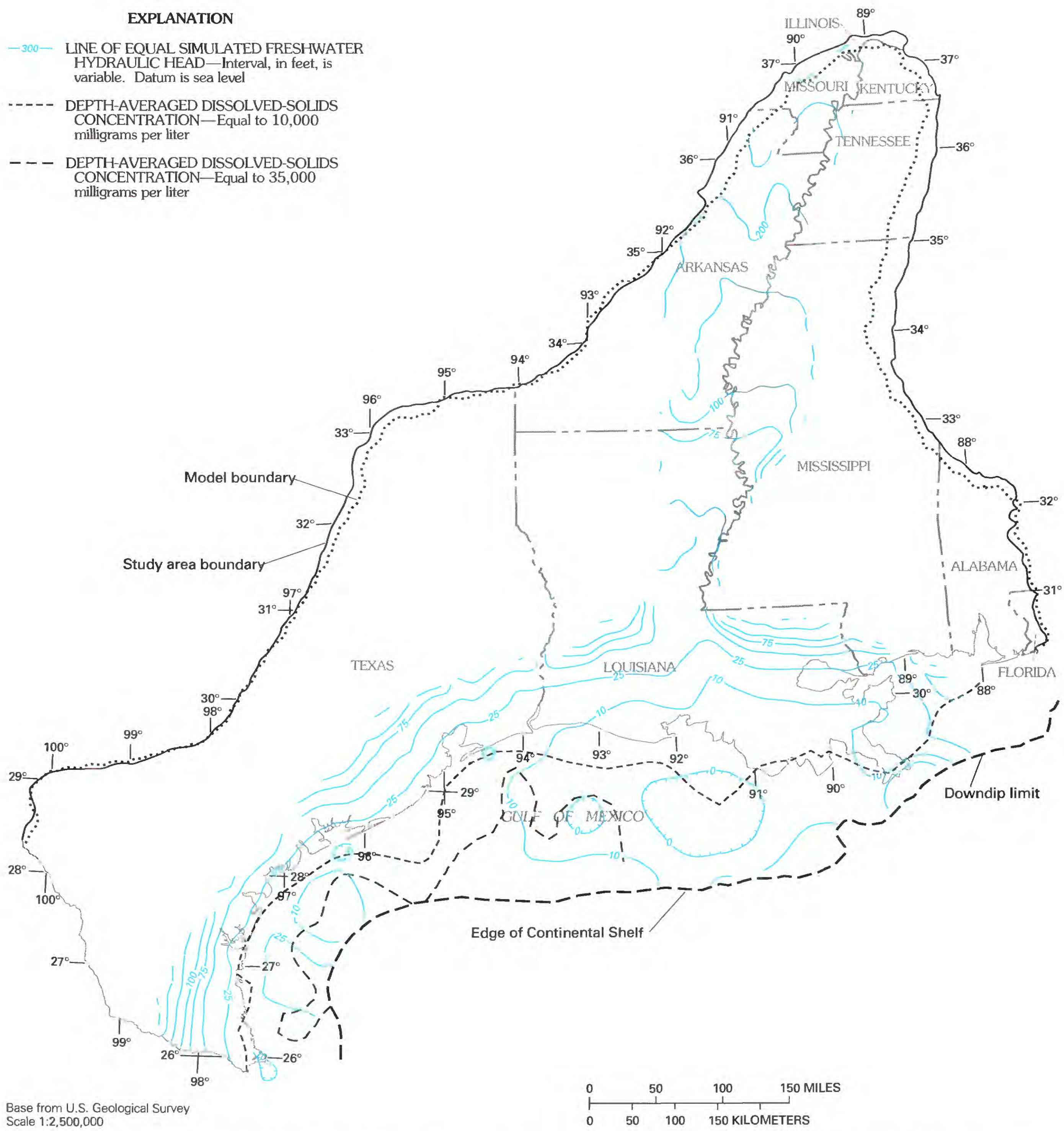


Plate 2A. Simulated predevelopment freshwater hydraulic head and depth-averaged dissolved-solids concentrations: Layer 11, permeable zone A, and Mississippi River Valley alluvial aquifer.

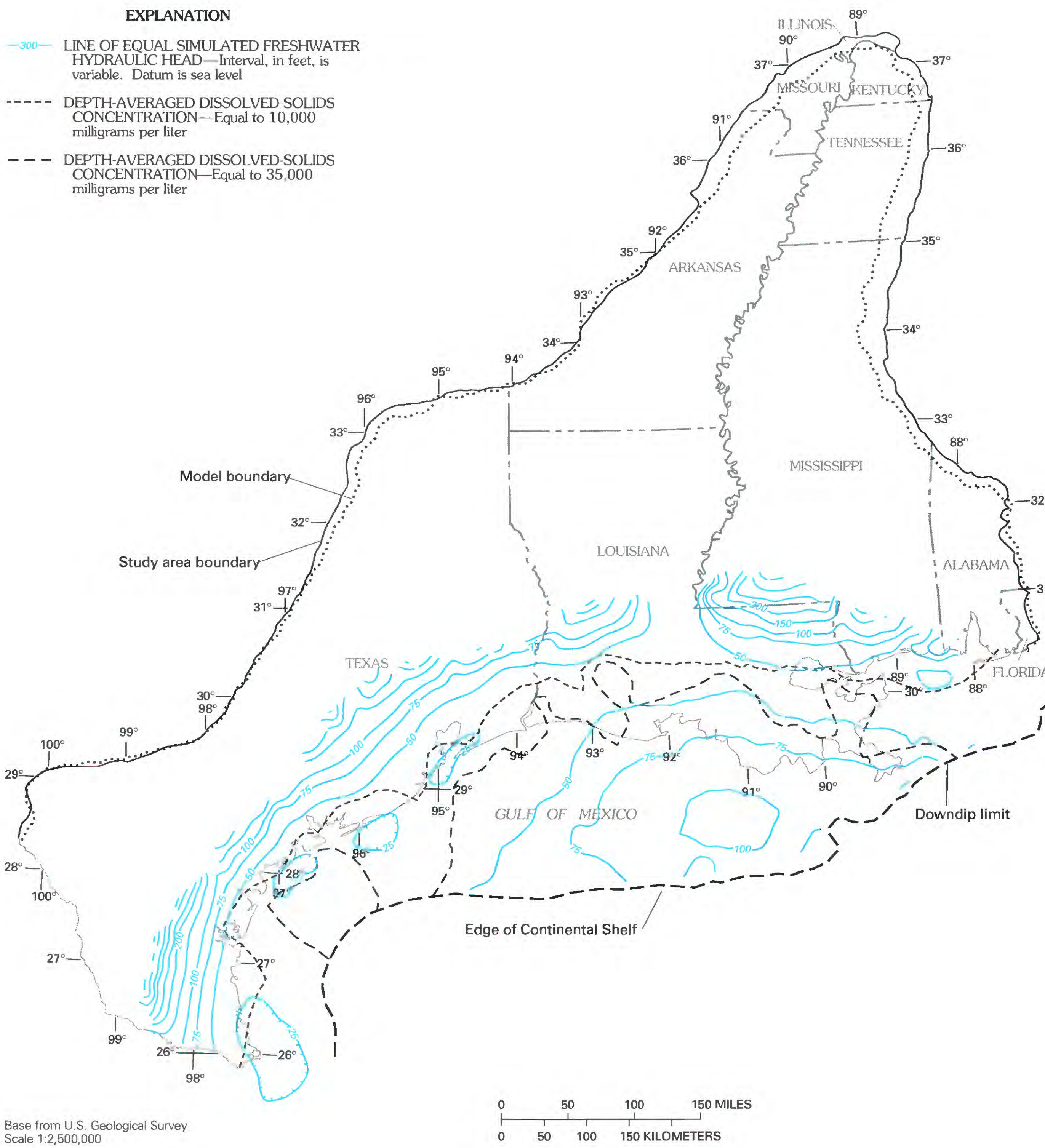


Plate 2B. Simulated predevelopment freshwater hydraulic head and depth-averaged dissolved-solids concentrations: Layer 10, permeable zone B.

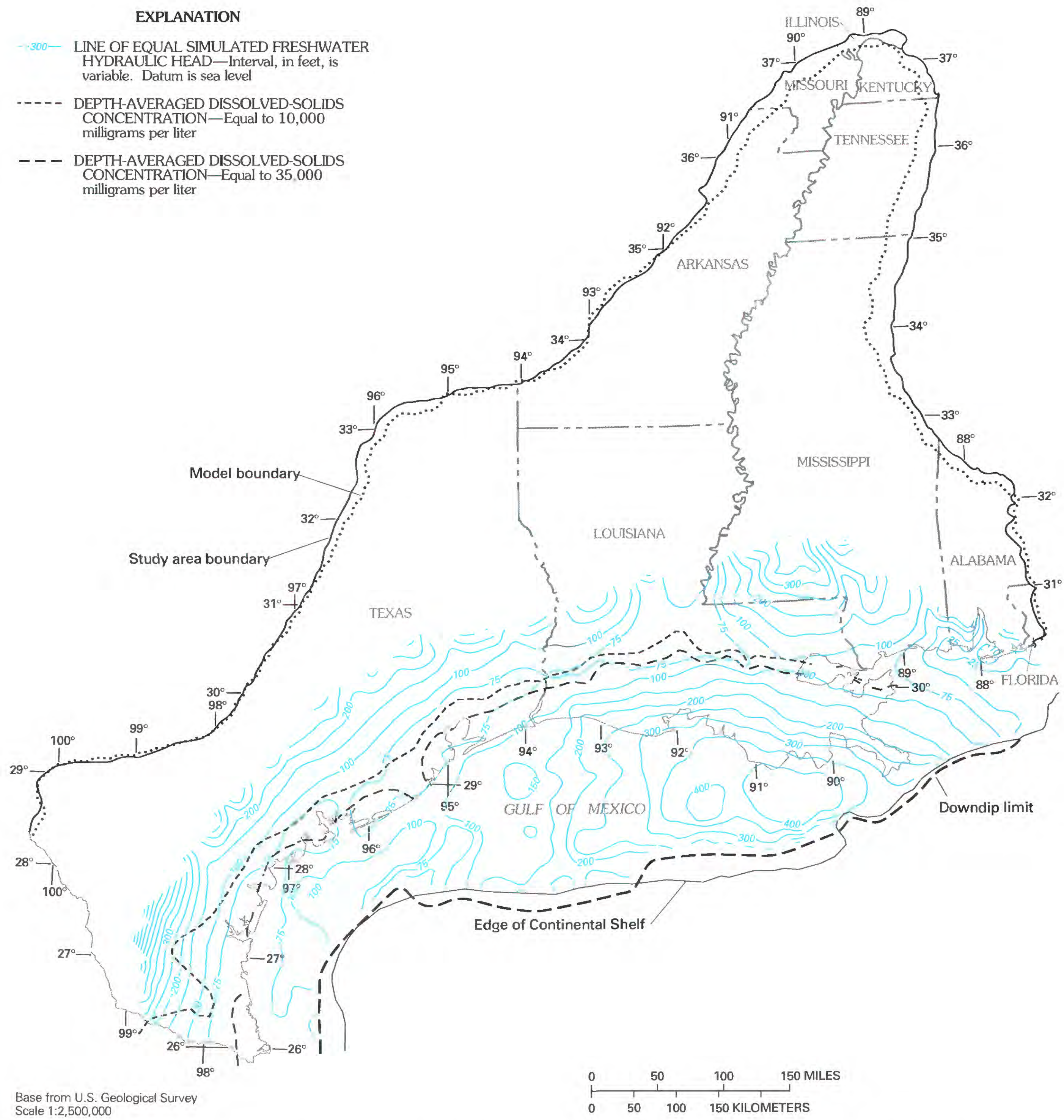


Plate 2C. Simulated predevelopment freshwater hydraulic head and depth-averaged dissolved-solids concentrations: Layer 9, permeable zone C.

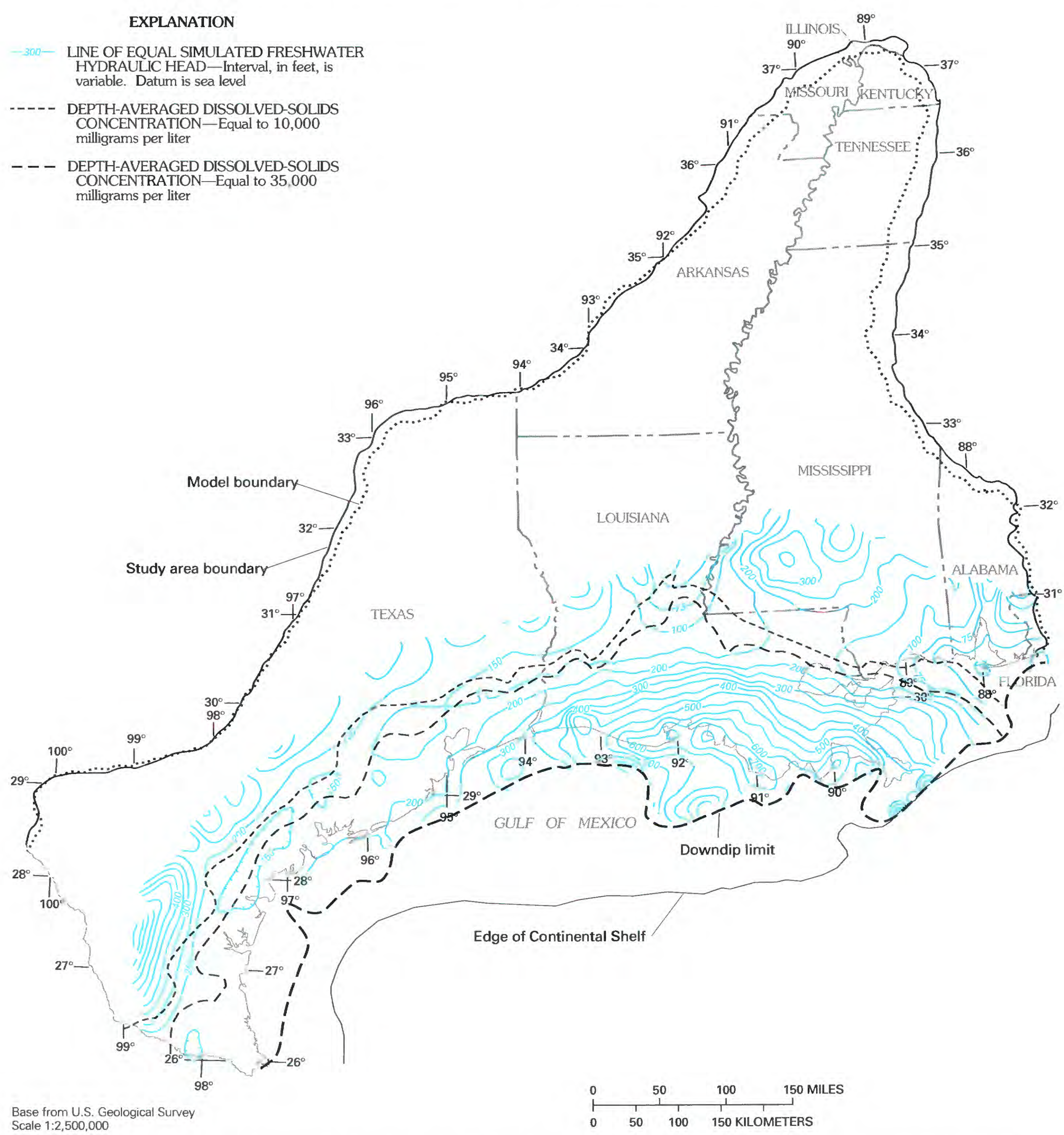


Plate 2D. Simulated predevelopment freshwater hydraulic head and depth-averaged dissolved-solids concentrations: Layer 8, permeable zone D.

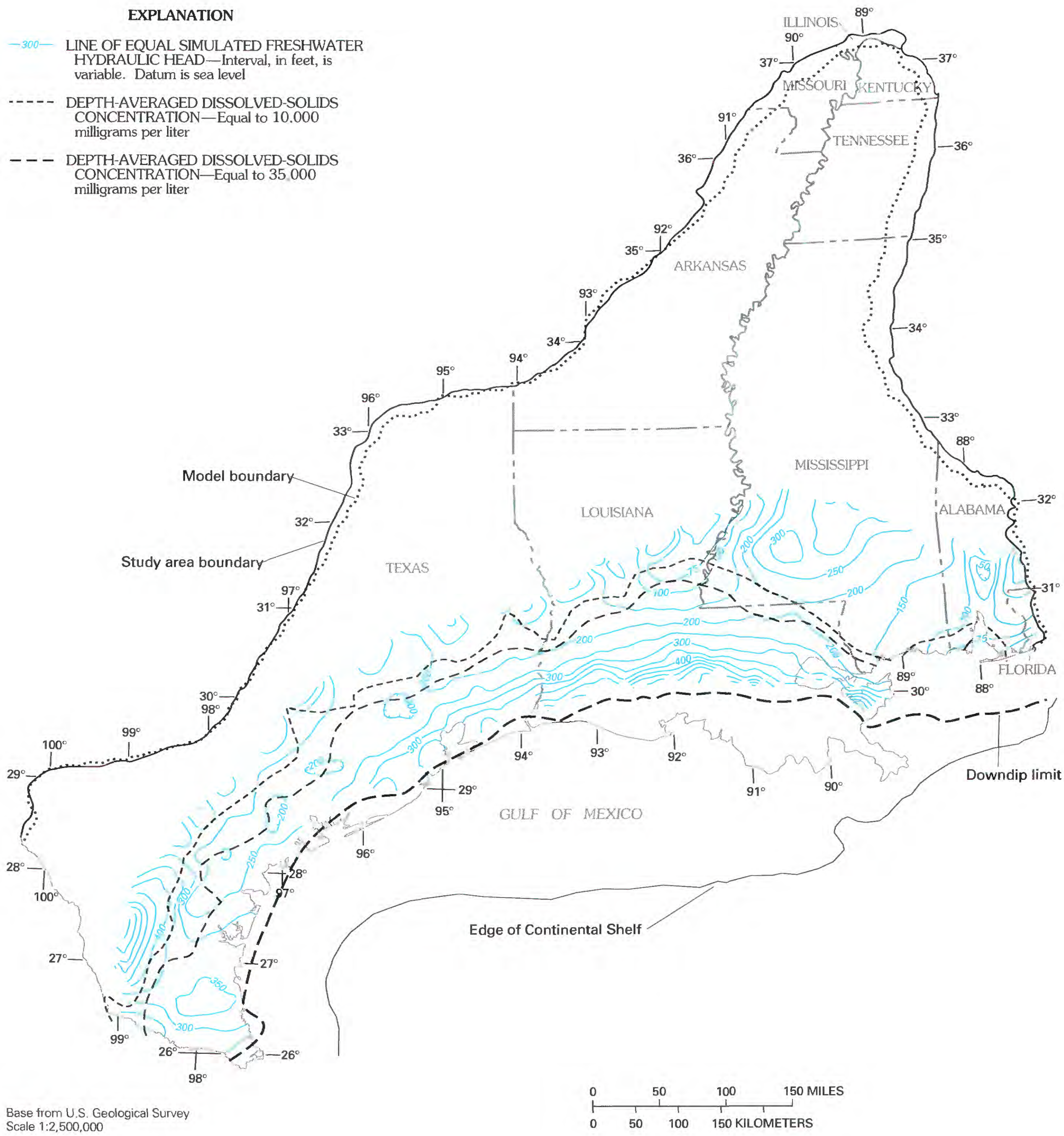


Plate 2E. Simulated predevelopment freshwater hydraulic head and depth-averaged dissolved-solids concentrations: Layer 7, permeable zone E.

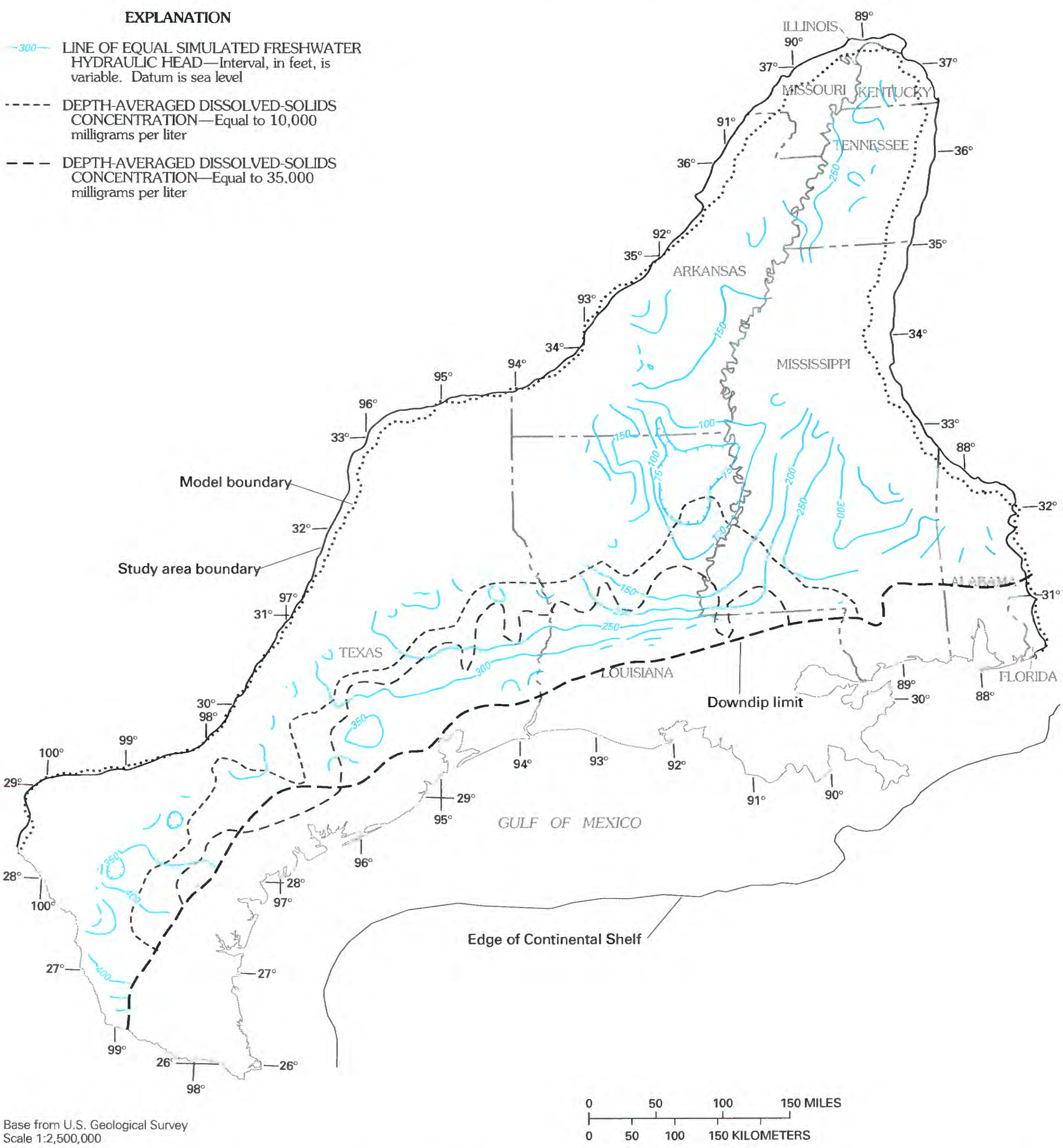


Plate 2F. Simulated predevelopment freshwater hydraulic head and depth-averaged dissolved-solids concentrations: Layer 6, upper Claiborne aquifer.

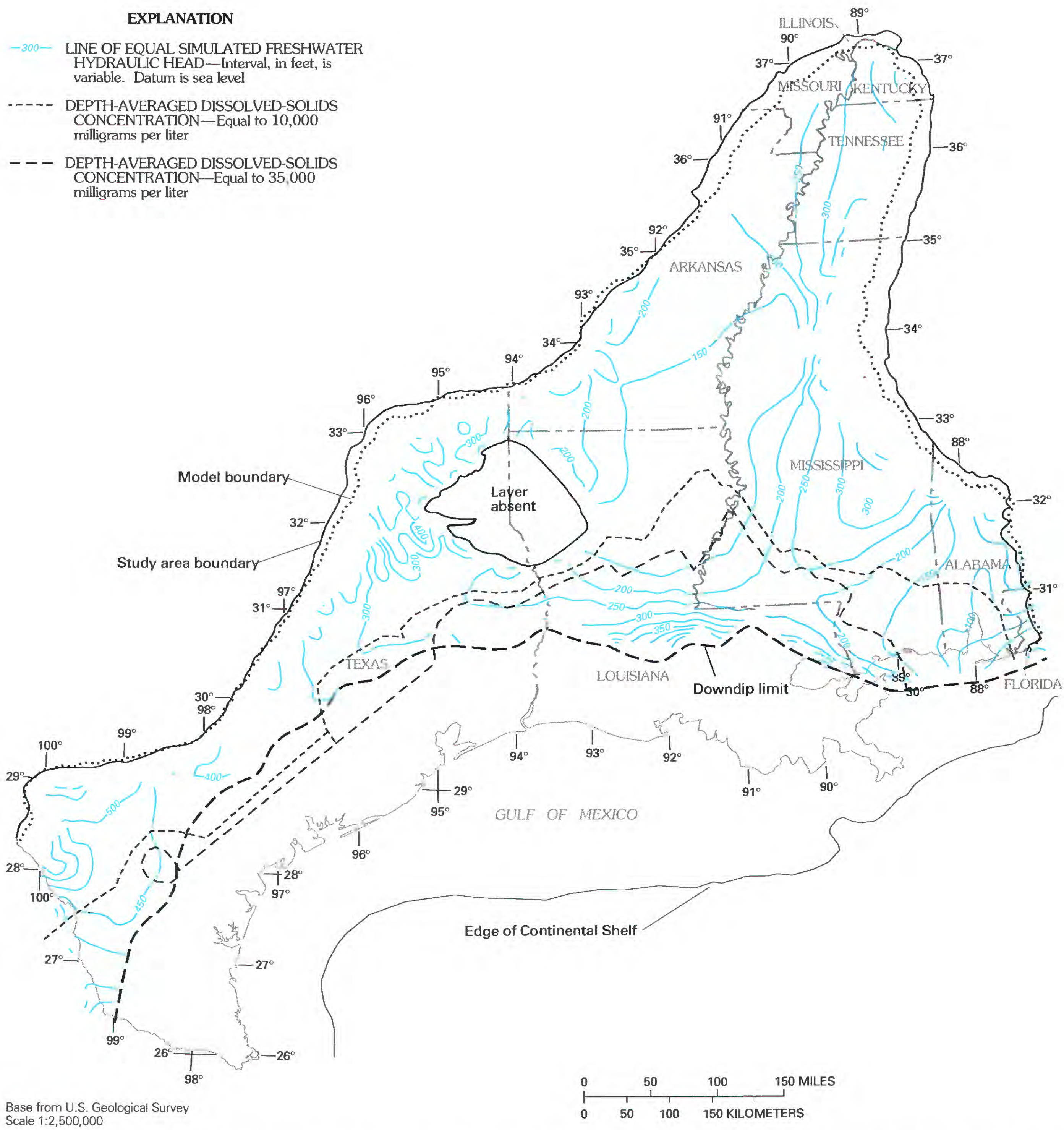


Plate 2G. Simulated predevelopment freshwater hydraulic head and depth-averaged dissolved-solids concentrations: Layer 5, middle Claiborne aquifer.

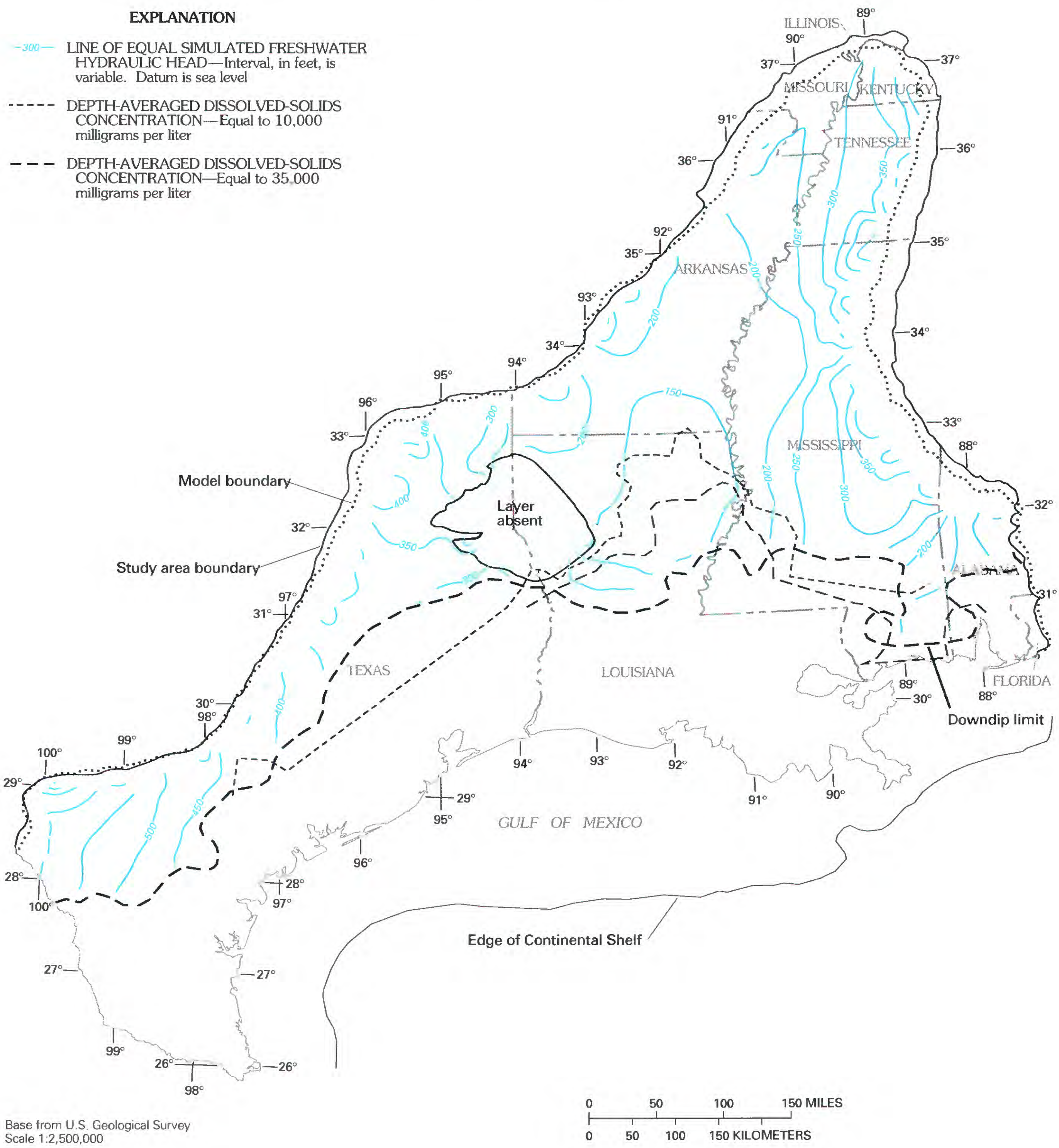


Plate 2H. Simulated predevelopment freshwater hydraulic head and depth-averaged dissolved-solids concentrations: Layer 4, lower Claiborne-upper Wilcox aquifer.

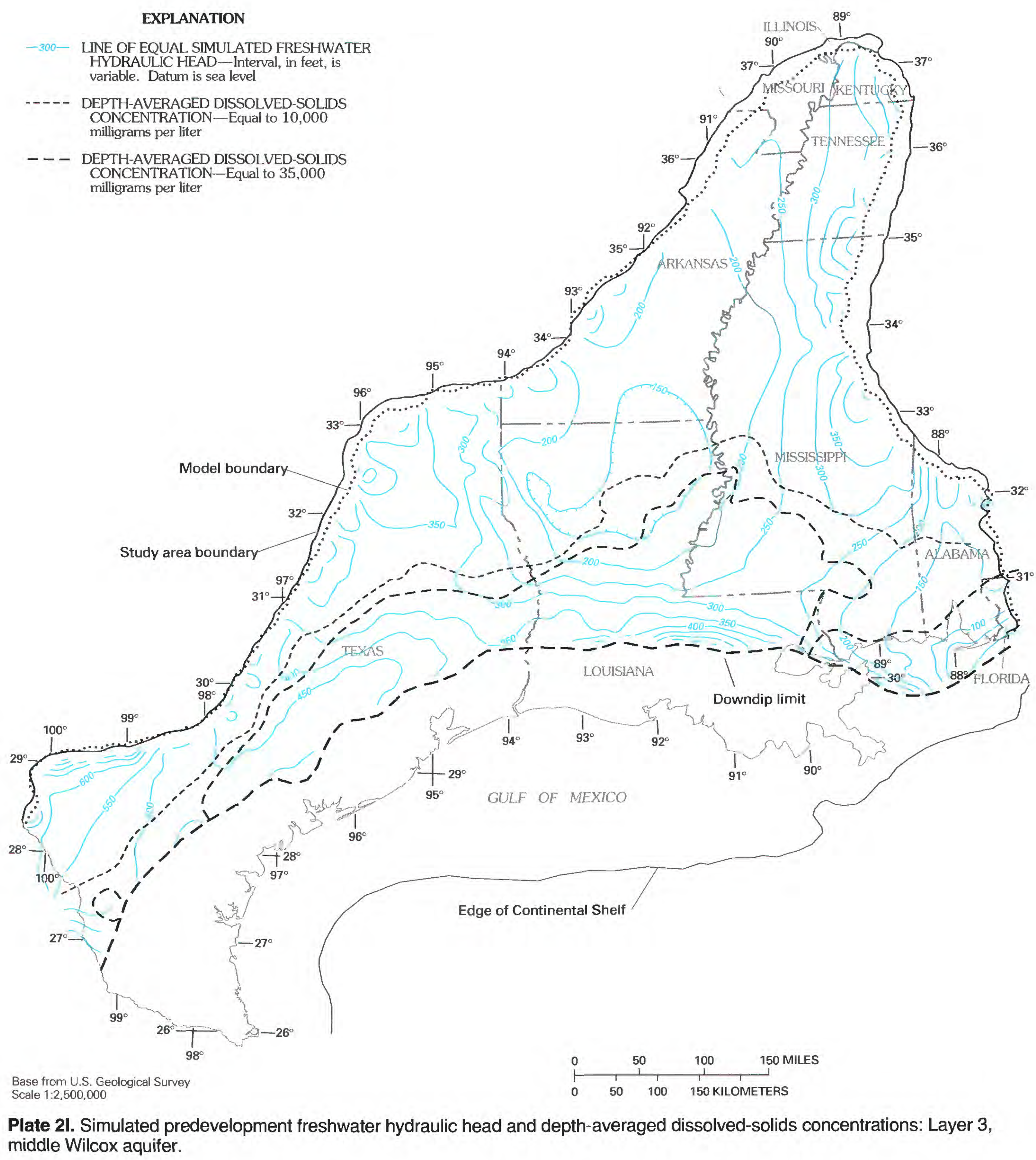


Plate 2I. Simulated predevelopment freshwater hydraulic head and depth-averaged dissolved-solids concentrations: Layer 3, middle Wilcox aquifer.

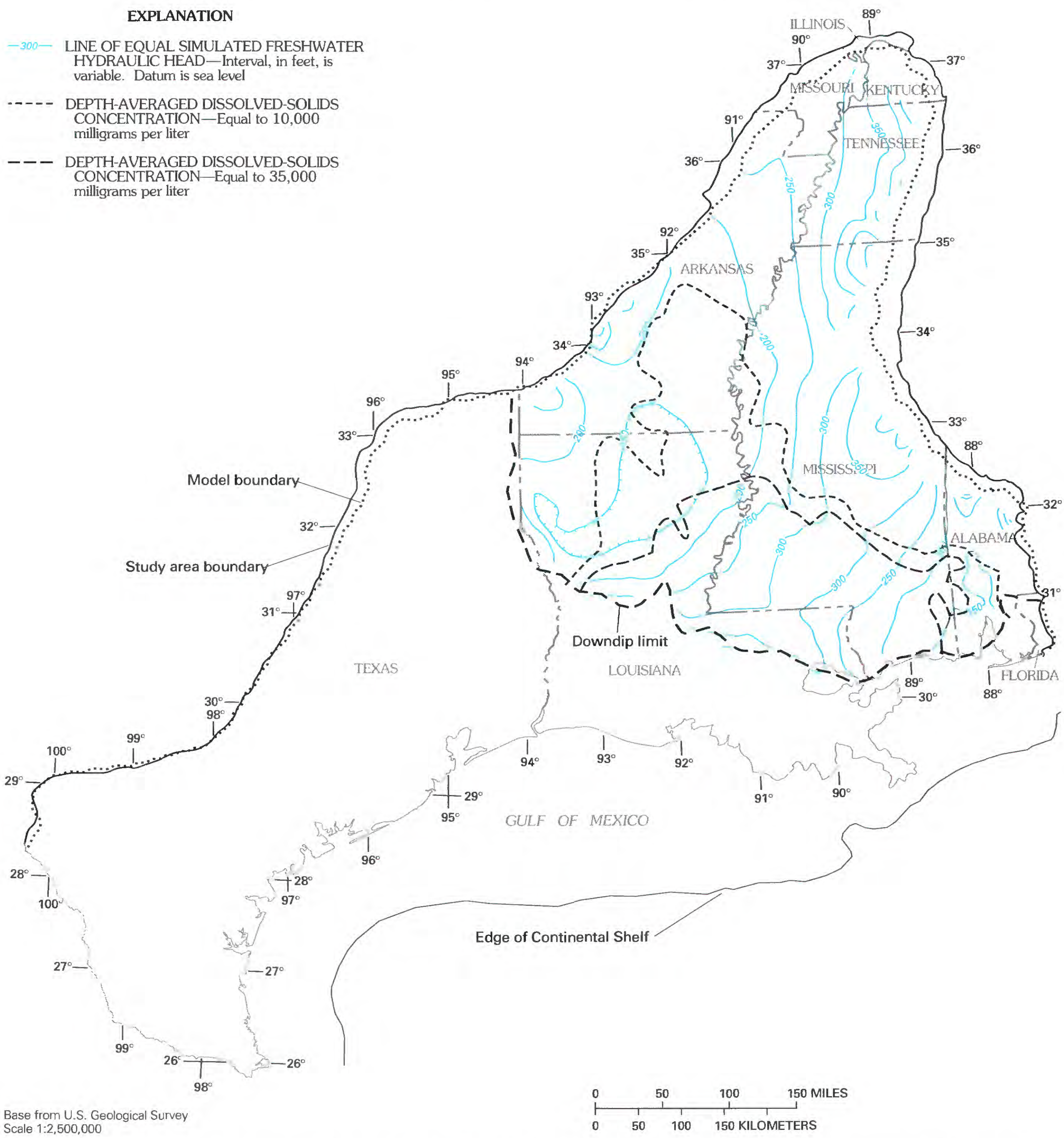


Plate 2J. Simulated predevelopment freshwater hydraulic head and depth-averaged dissolved-solids concentrations: Layer 2, lower Wilcox aquifer.

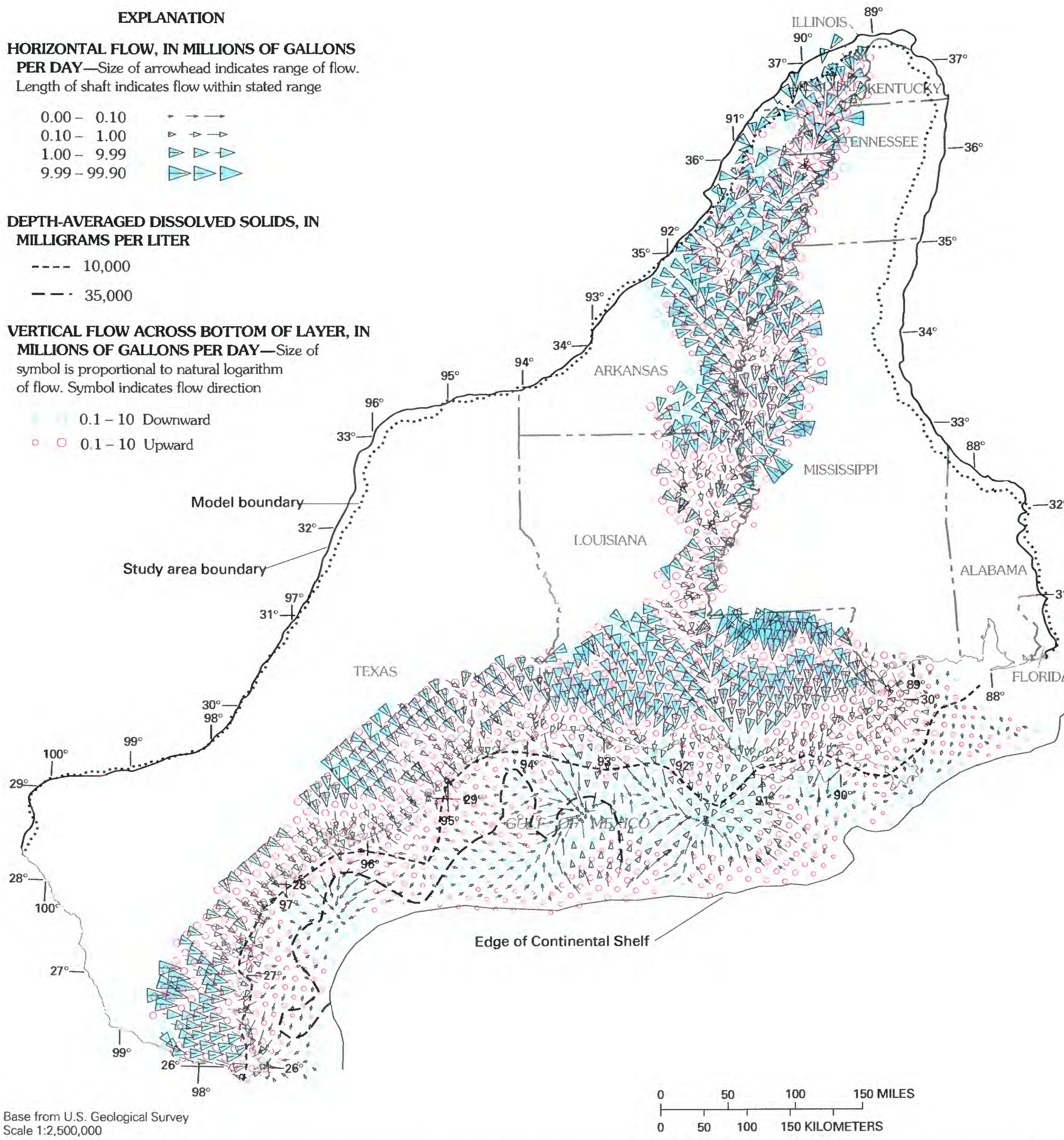


Plate 3A. Simulated predevelopment horizontal and vertical ground-water flow direction and magnitude, and depth-averaged dissolved-solids concentration: Layer 11, permeable zone A, and Mississippi River Valley alluvial aquifer.

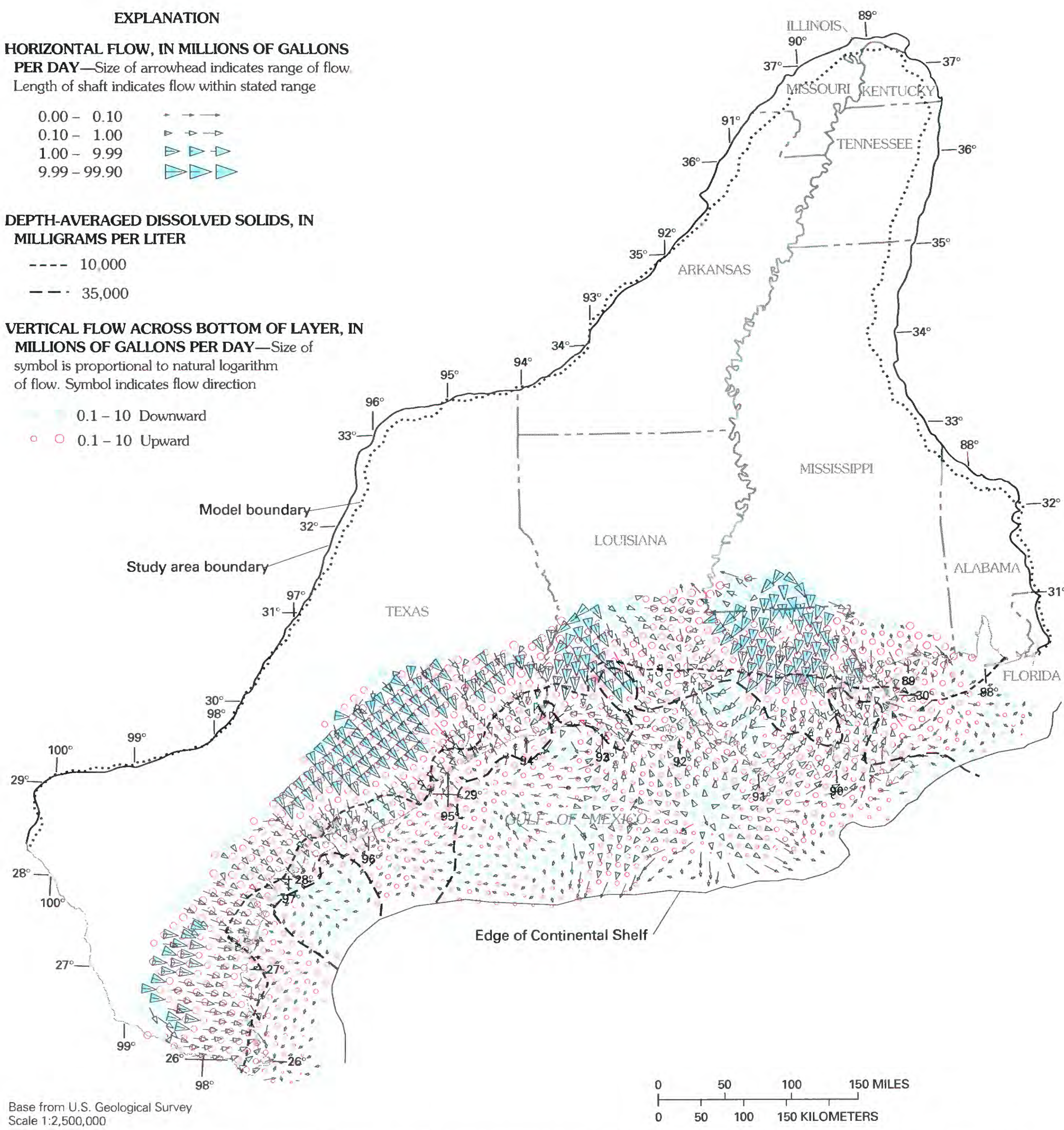


Plate 3B. Simulated predevelopment horizontal and vertical ground-water flow direction and magnitude, and depth-averaged dissolved-solids concentration: Layer 10, permeable zone B.

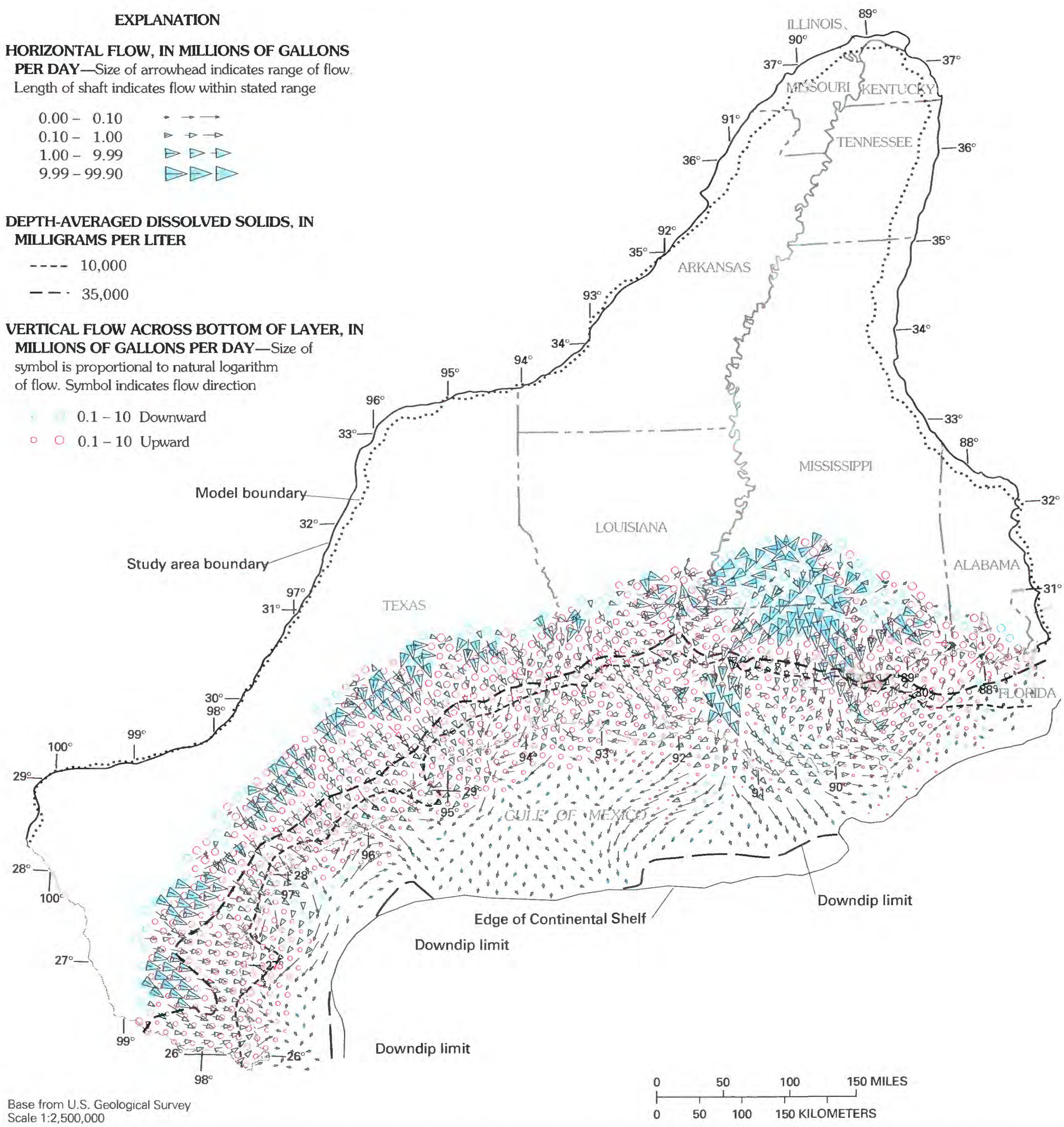


Plate 3C. Simulated predevelopment horizontal and vertical ground-water flow direction and magnitude, and depth-averaged dissolved-solids concentration: Layer 9, permeable zone C.

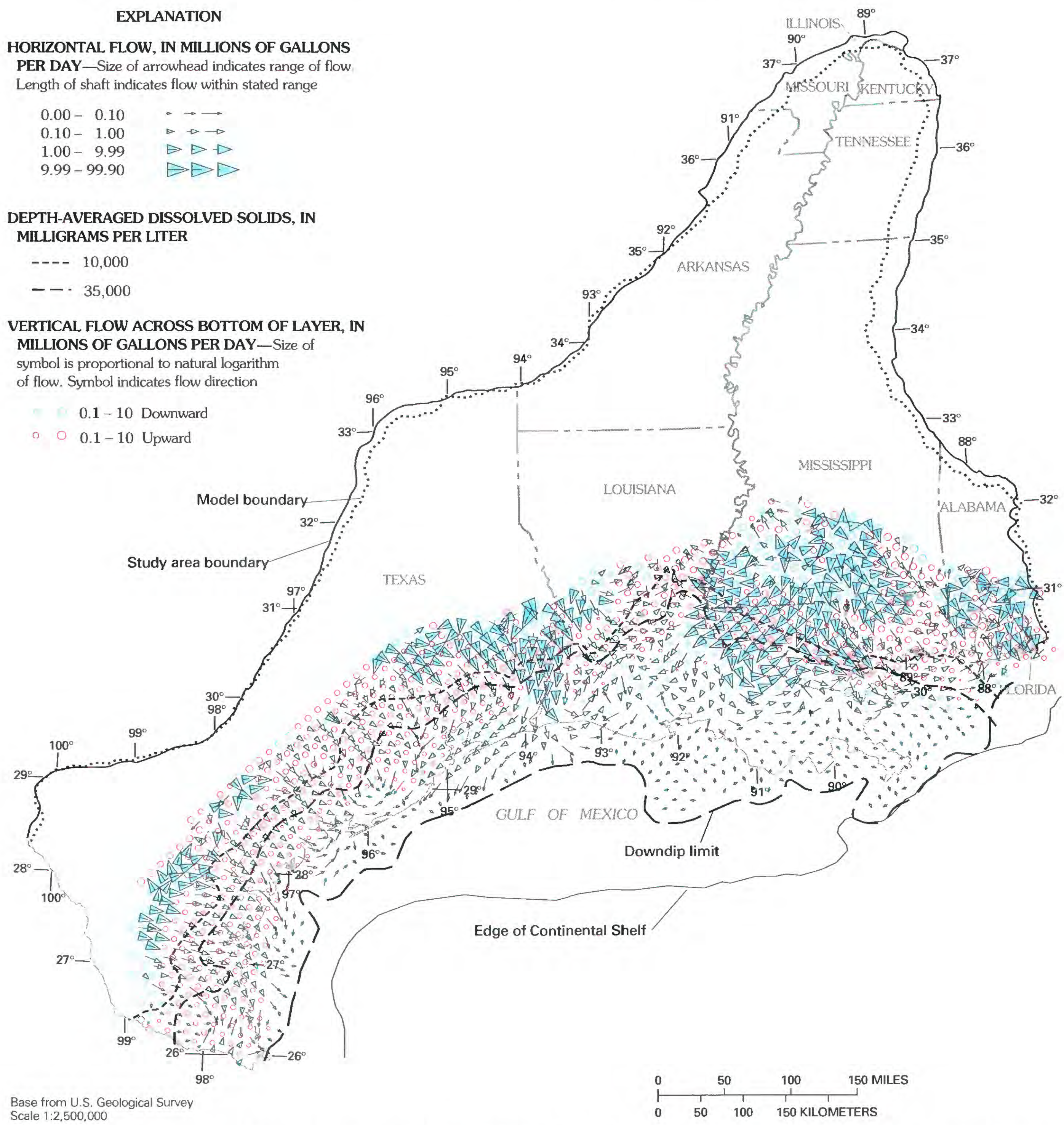


Plate 3D. Simulated predevelopment horizontal and vertical ground-water flow direction and magnitude, and depth-averaged dissolved-solids concentration: Layer 8, permeable zone D.

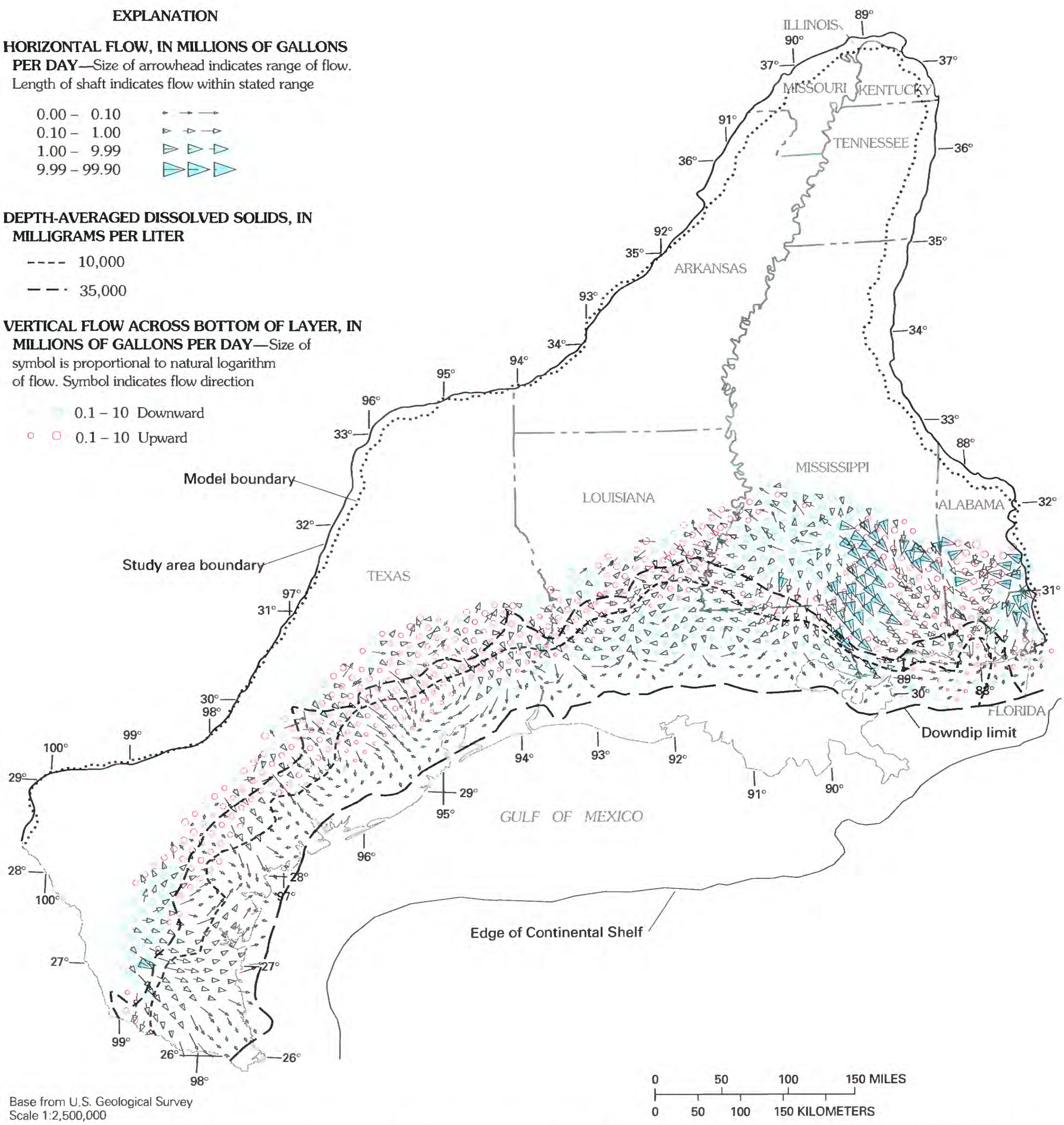
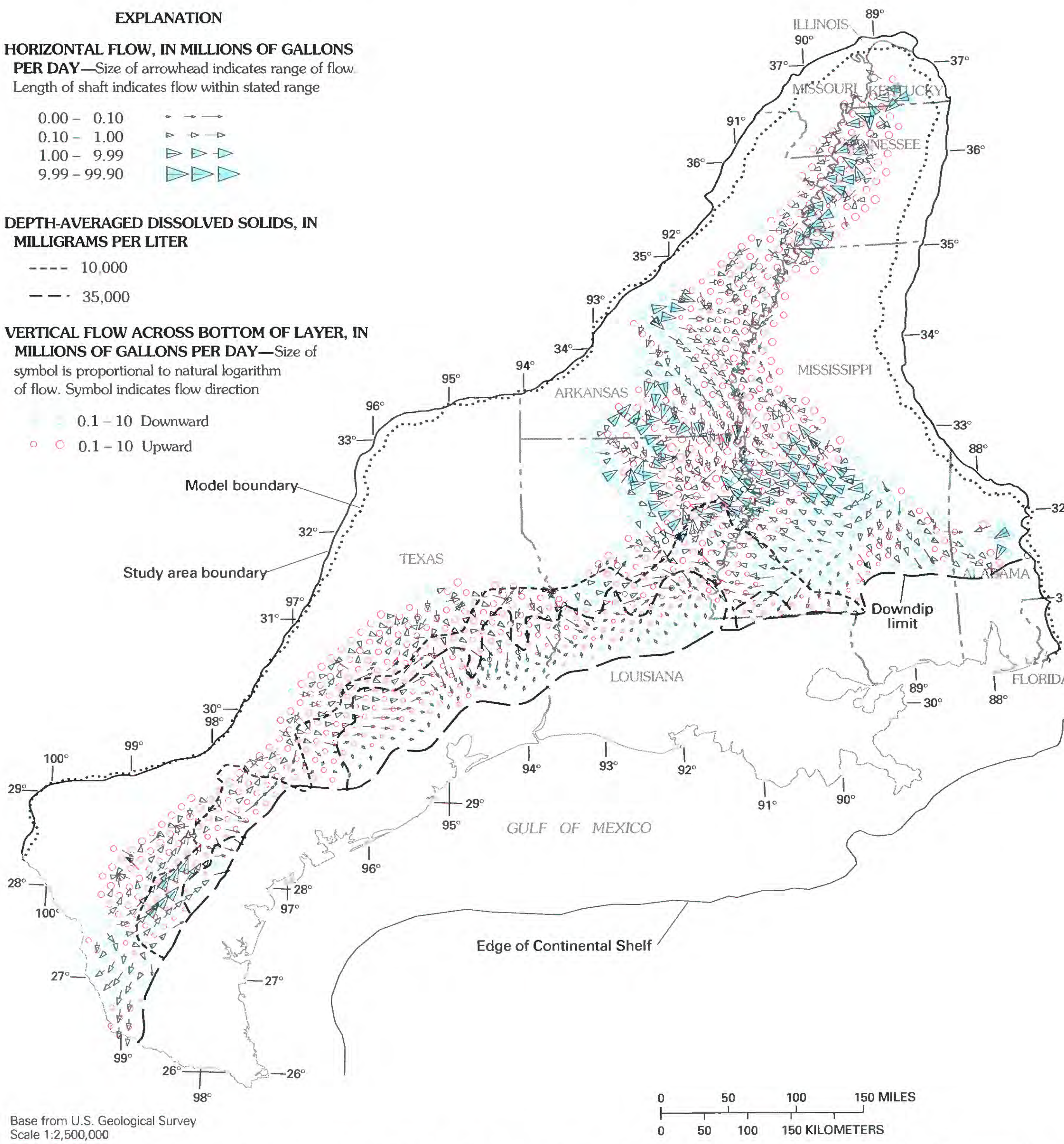


Plate 3E. Simulated predevelopment horizontal and vertical ground-water flow direction and magnitude, and depth-averaged dissolved-solids concentration: Layer 7, permeable zone E.



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

Plate 3F. Simulated predevelopment horizontal and vertical ground-water flow direction and magnitude, and depth-averaged dissolved-solids concentration: Layer 6, upper Claiborne aquifer.

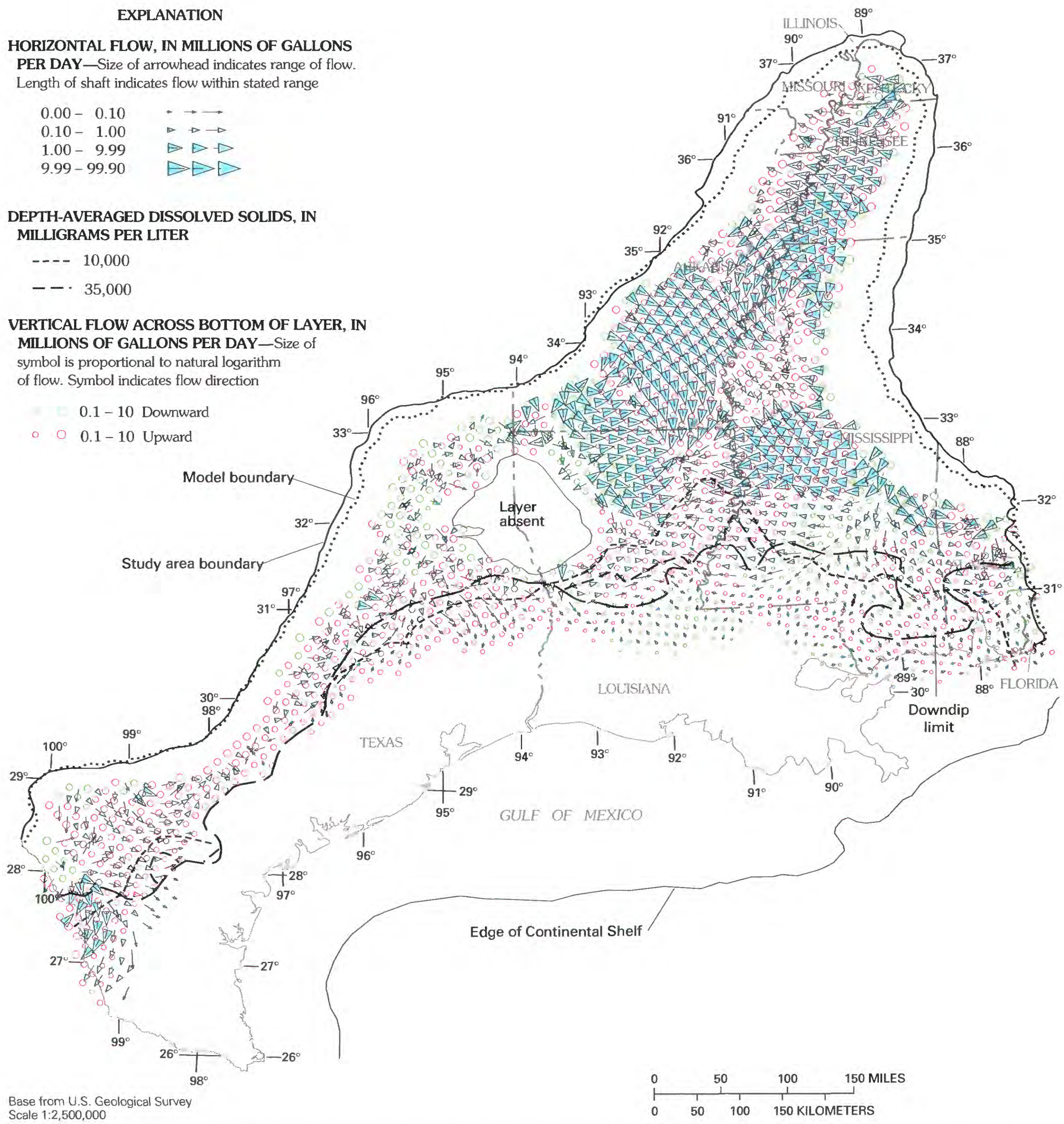


Plate 3G. Simulated predevelopment horizontal and vertical ground-water flow direction and magnitude, and depth-averaged dissolved-solids concentration: Layer 5, middle Claiborne aquifer.

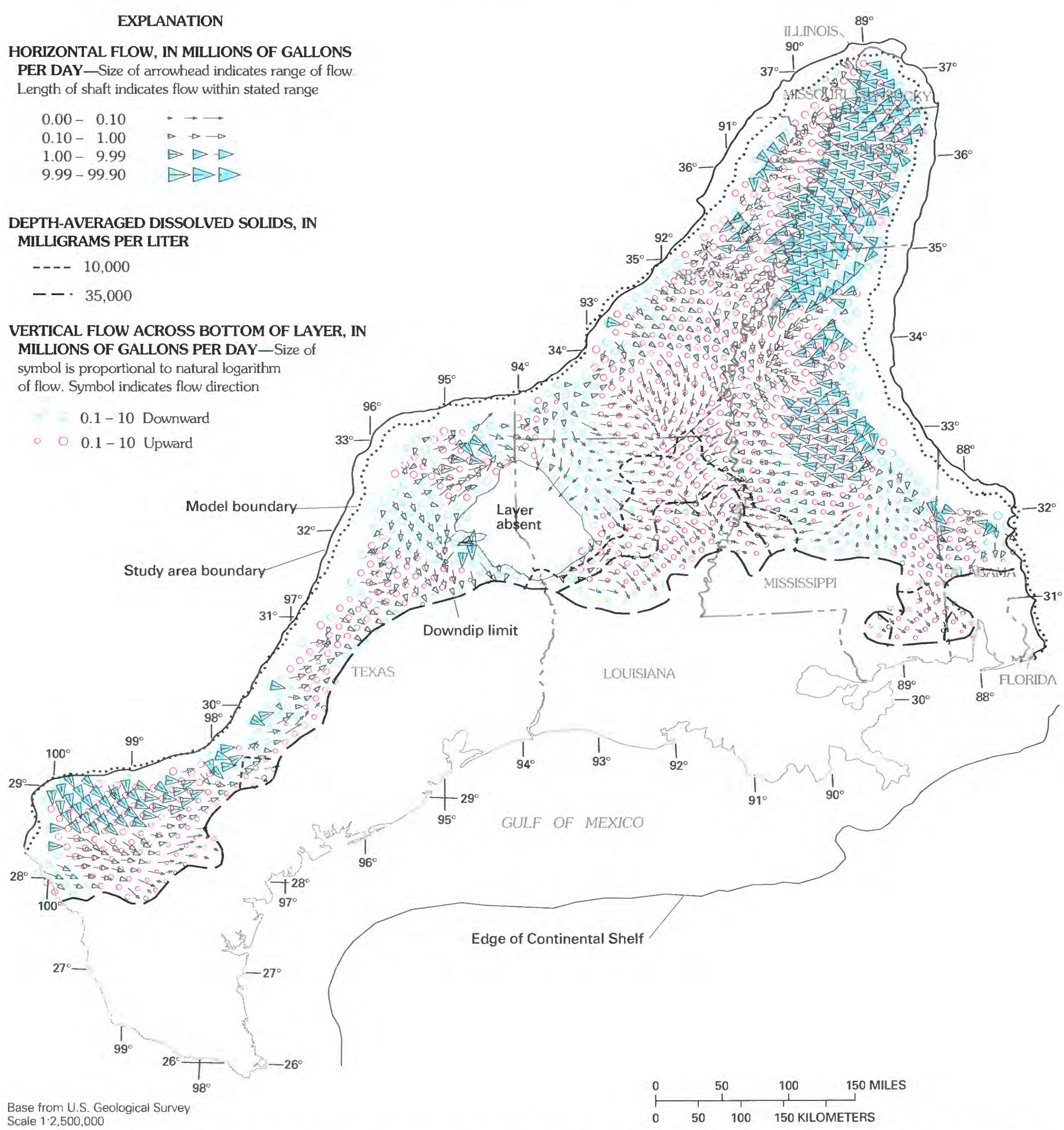


Plate 3H. Simulated predevelopment horizontal and vertical ground-water flow direction and magnitude, and depth-averaged dissolved-solids concentration: Layer 4, lower Claiborne-upper Wilcox aquifer.

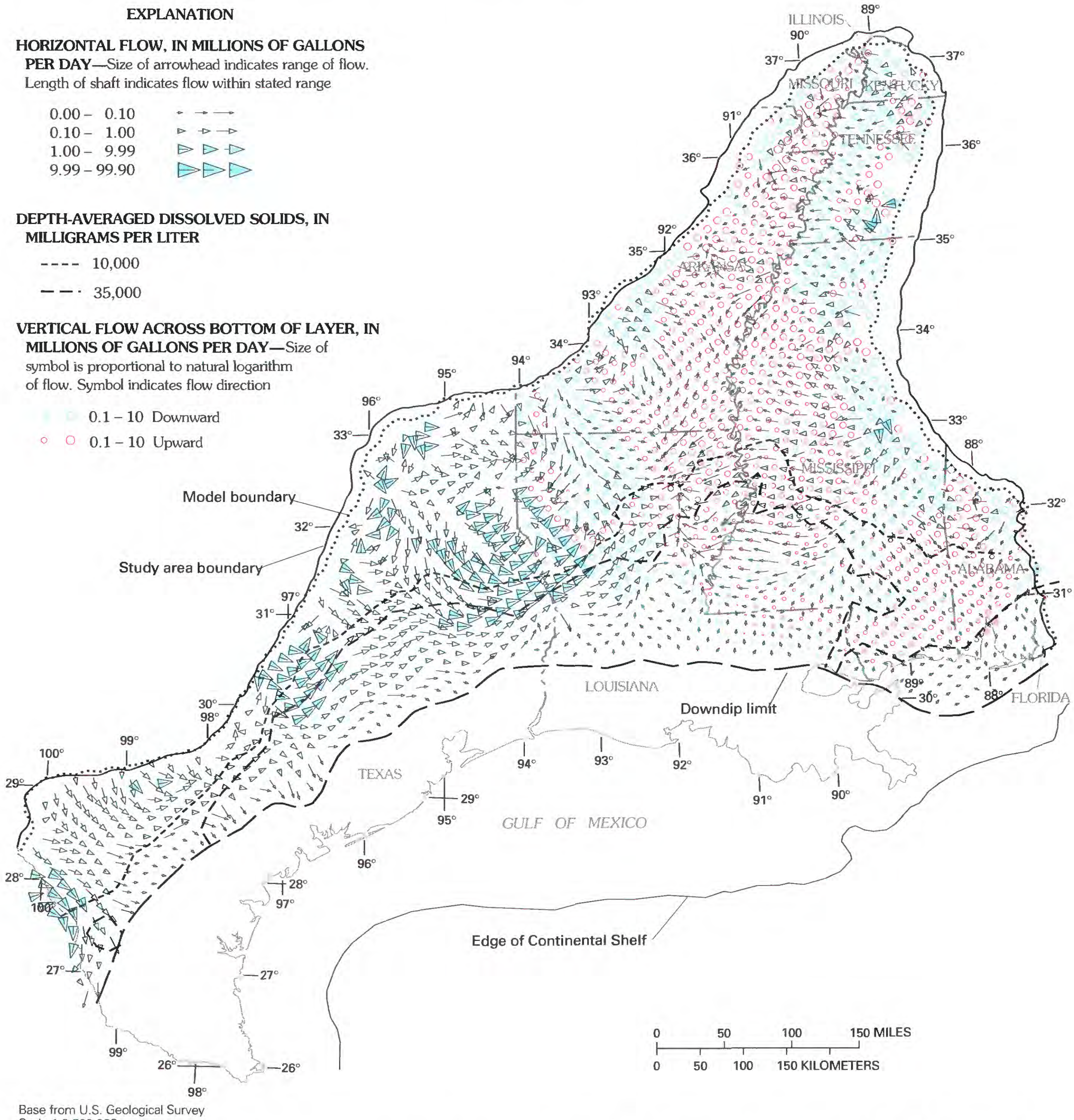


Plate 3I. Simulated predevelopment horizontal and vertical ground-water flow direction and magnitude, and depth-averaged dissolved-solids concentration: Layer 3, middle Wilcox aquifer.

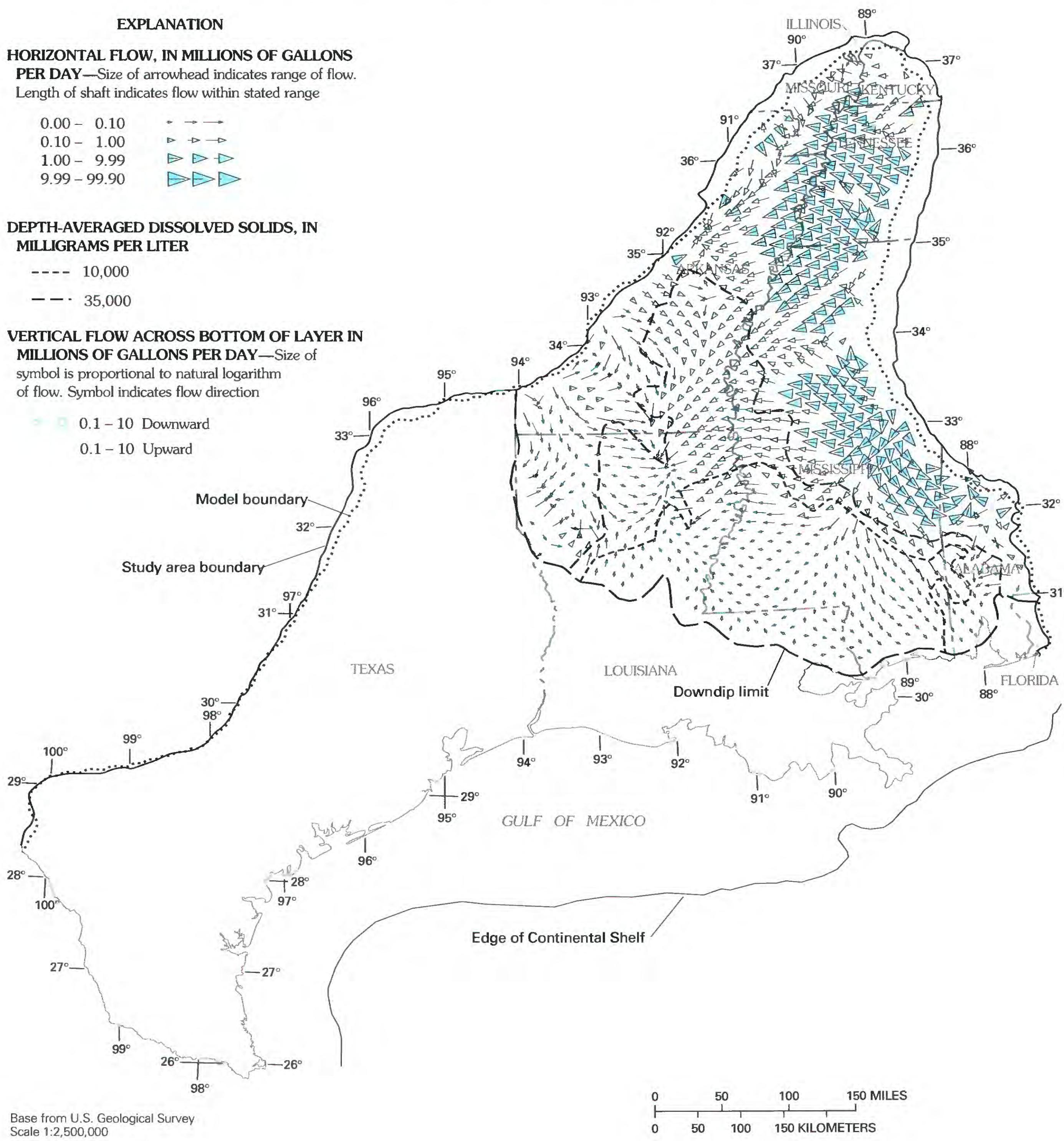


Plate 3J. Simulated predevelopment horizontal and vertical ground-water flow direction and magnitude, and depth-averaged dissolved-solids concentration: Layer 2, lower Wilcox aquifer.

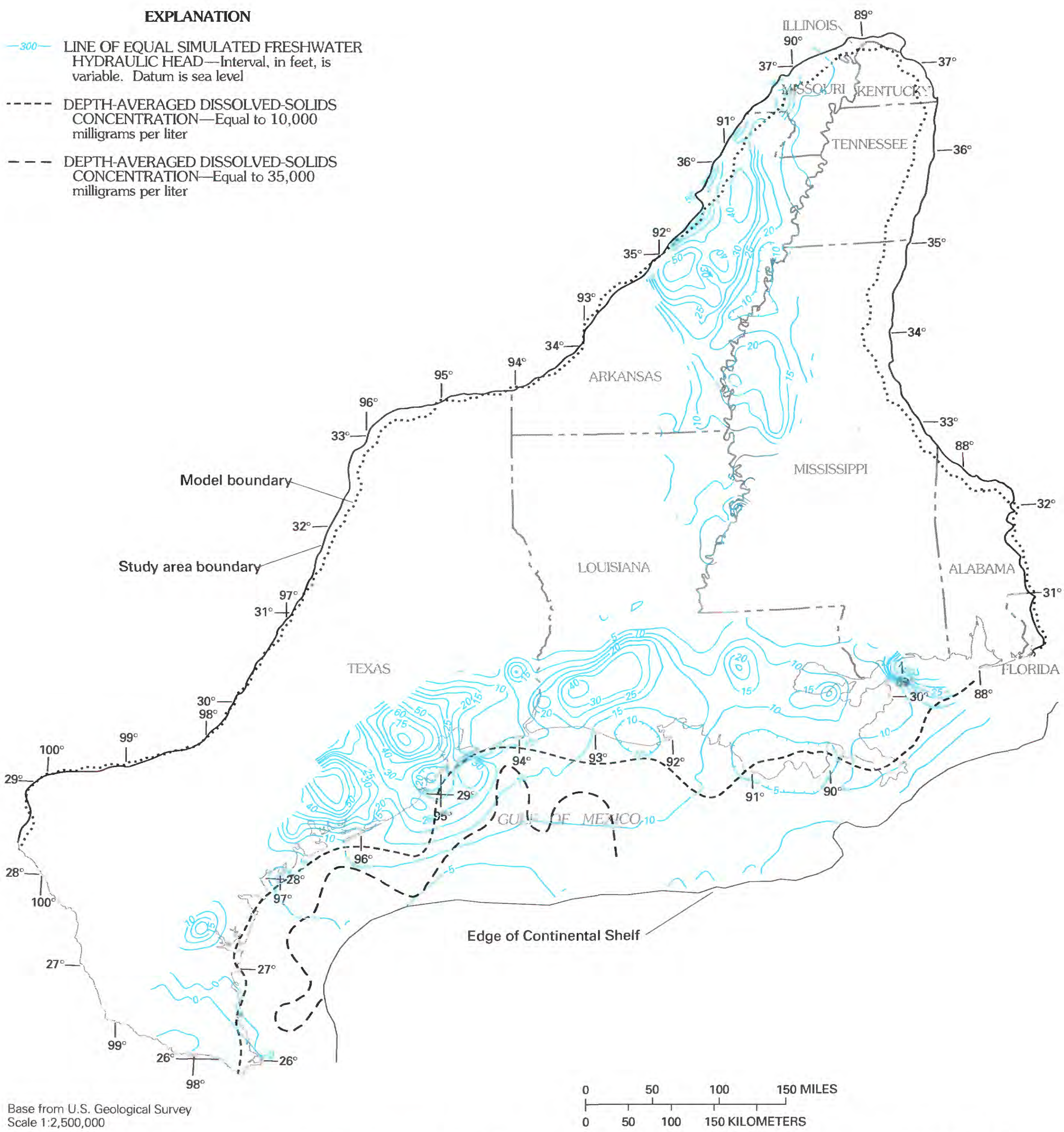


Plate 4A. Simulated drawdown from predevelopment to 1987 and depth-averaged dissolved-solids concentrations: Layer 11, permeable zone A, and Mississippi River Valley alluvial aquifer.

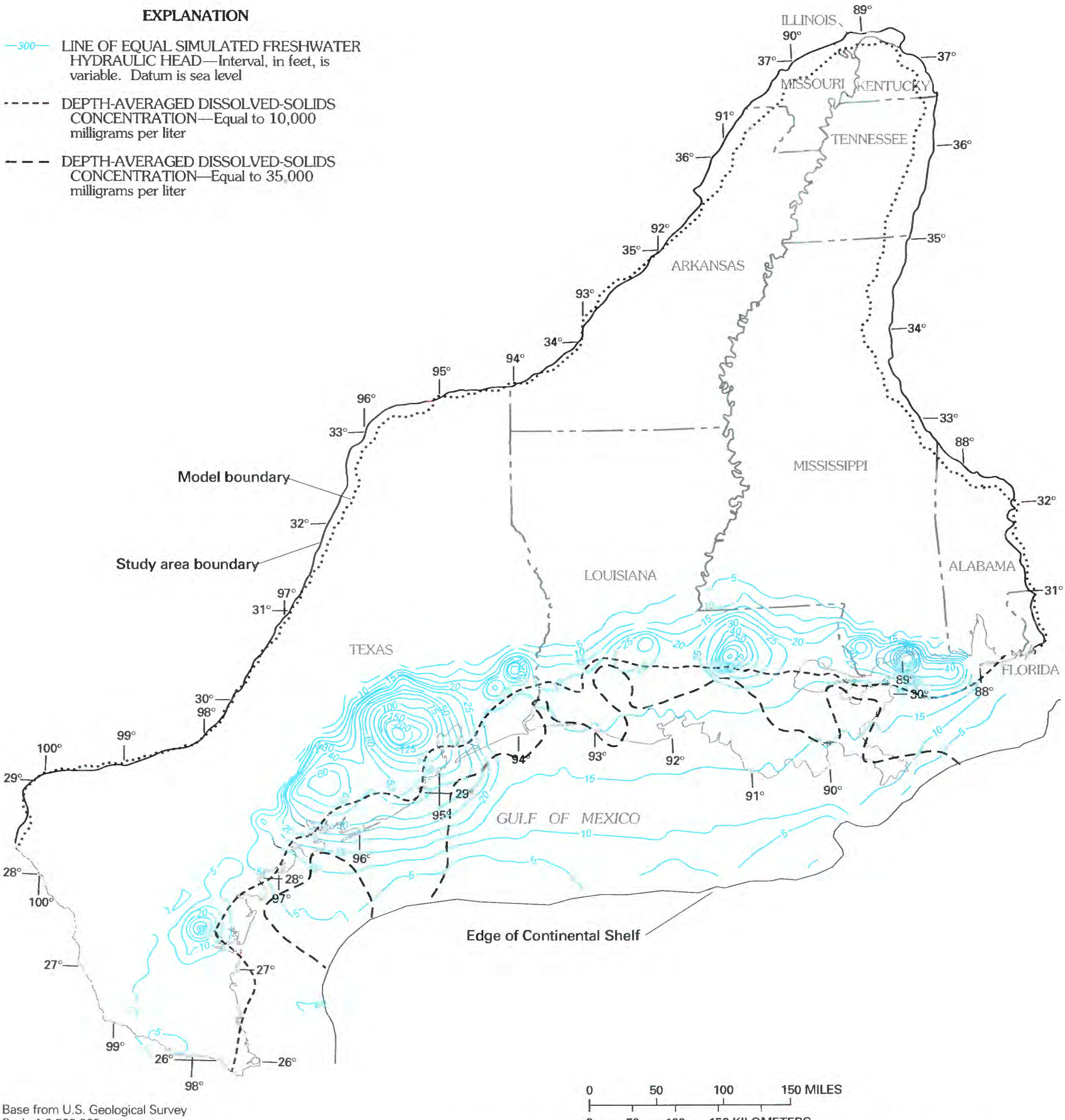


Plate 4B. Simulated drawdown from predevelopment to 1987 and depth-averaged dissolved-solids concentrations: Layer 10, permeable zone B.

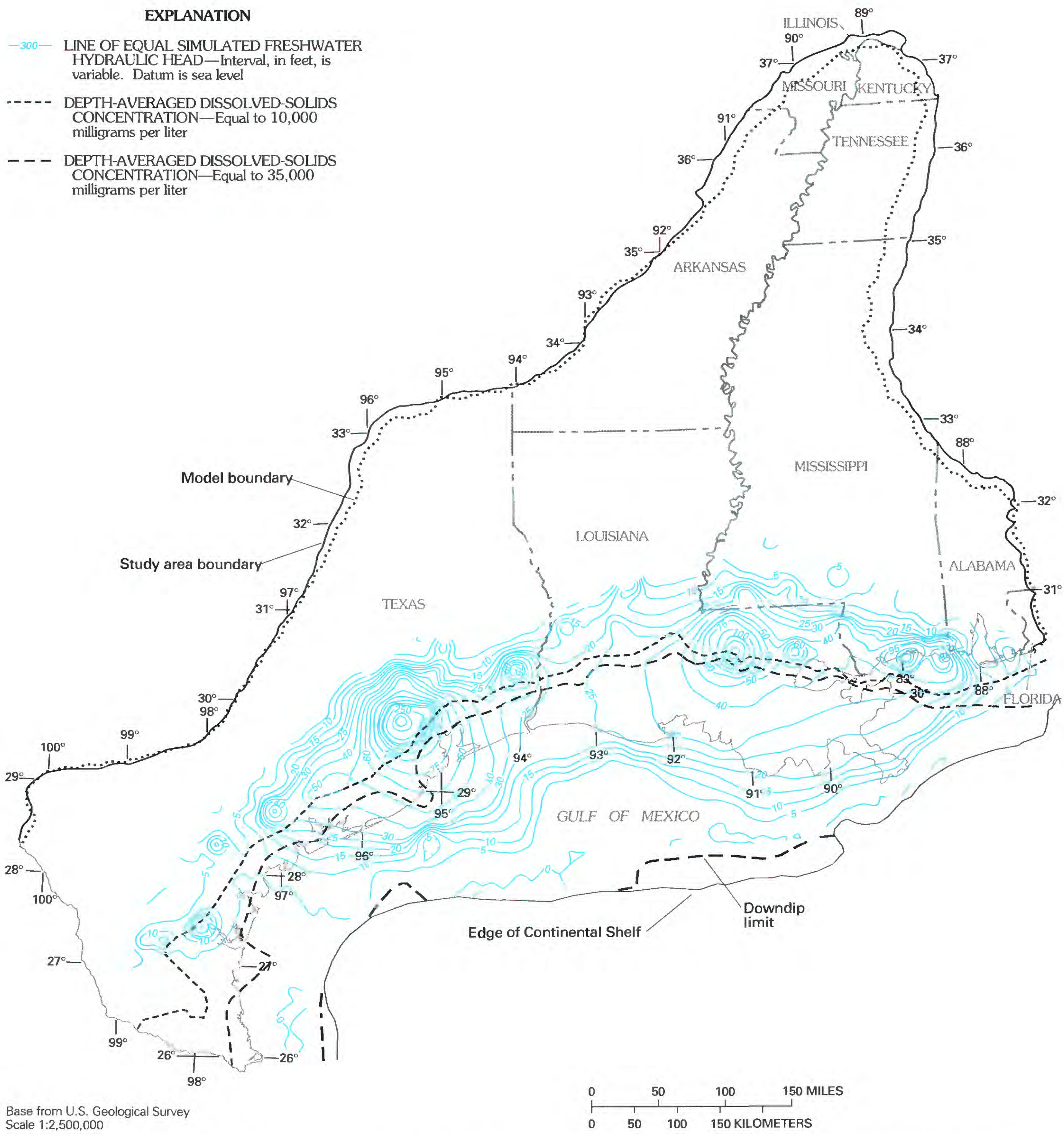


Plate 4C. Simulated drawdown from predevelopment to 1987 and depth-averaged dissolved-solids concentrations: Layer 9, permeable zone C.

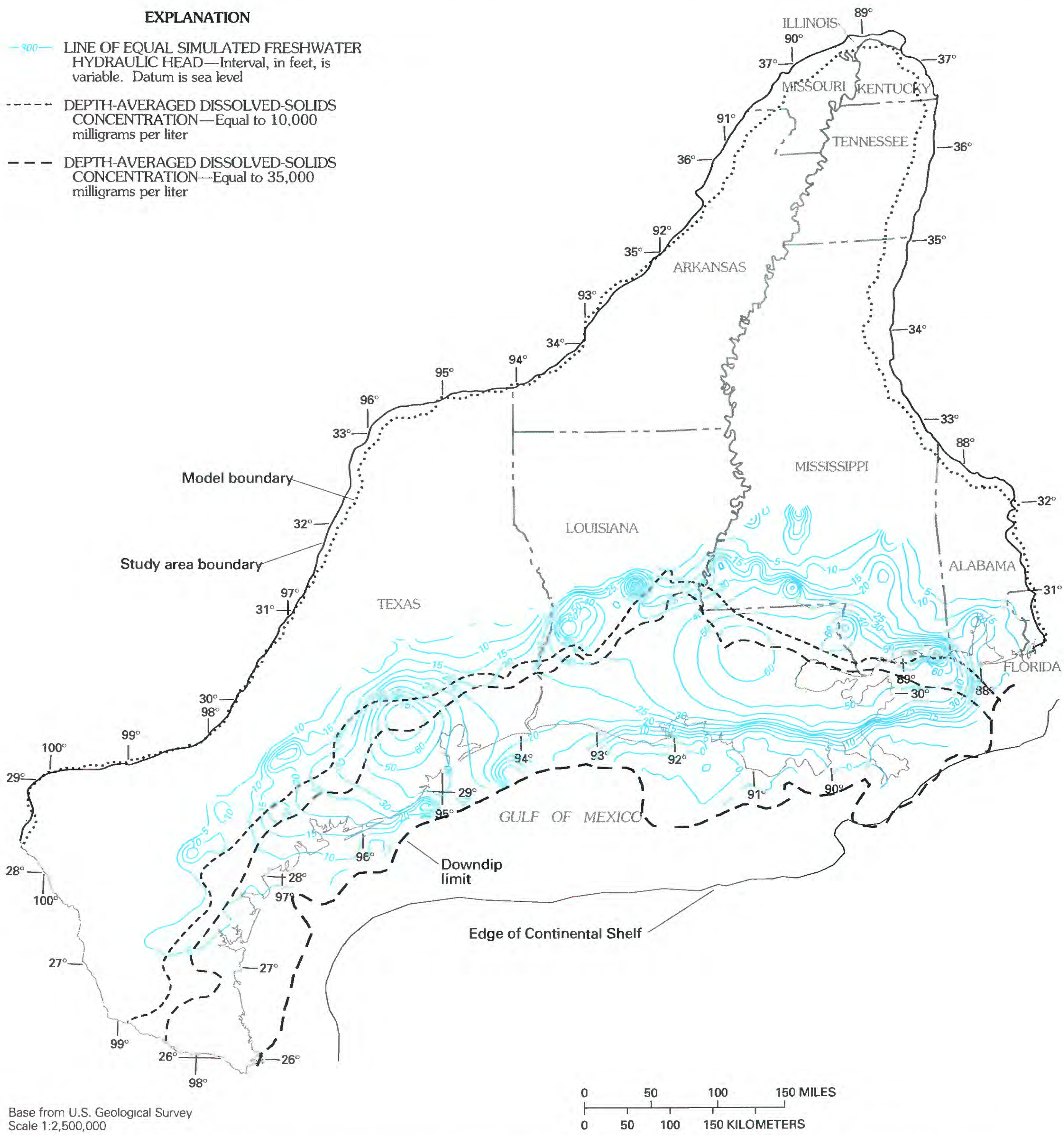


Plate 4D. Simulated drawdown from predevelopment to 1987 and depth-averaged dissolved-solids concentrations: Layer 8, permeable zone D.

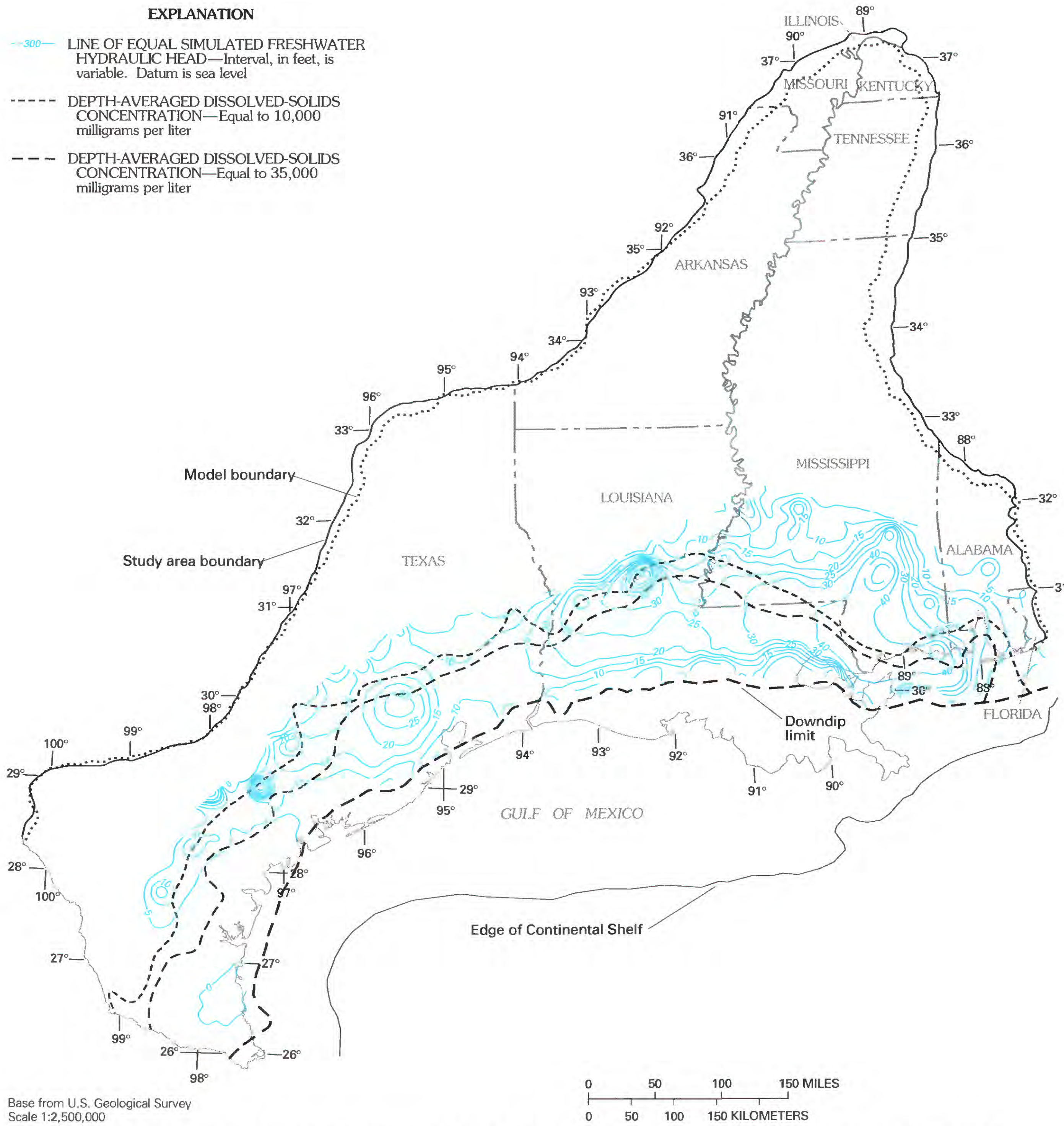


Plate 4E. Simulated drawdown from predevelopment to 1987 and depth-averaged dissolved-solids concentrations: Layer 7, permeable zone E.

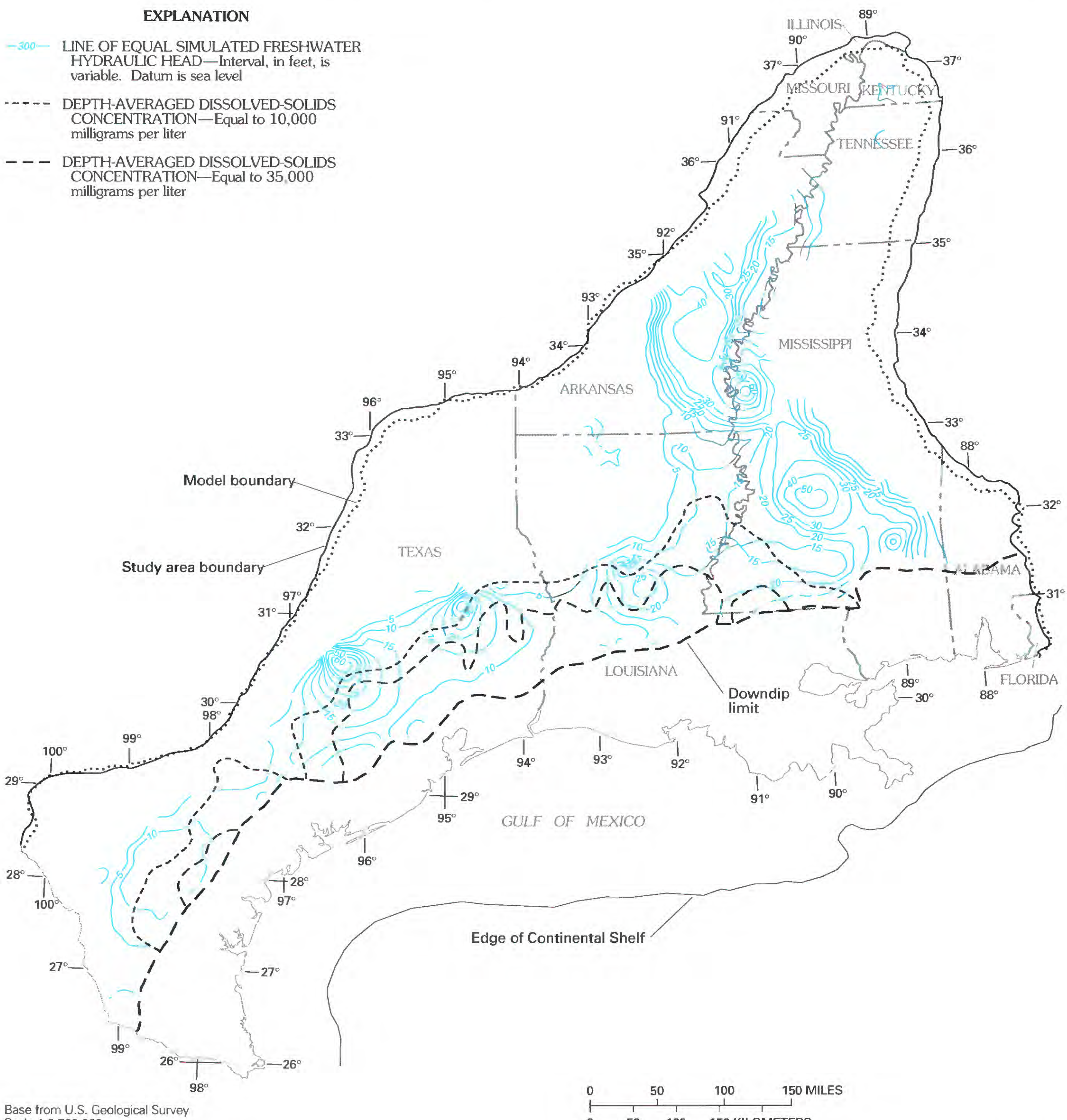


Plate 4F. Simulated drawdown from predevelopment to 1987 and depth-averaged dissolved-solids concentrations: Layer 6, upper Claiborne aquifer.

EXPLANATION

—300— LINE OF EQUAL SIMULATED FRESHWATER HYDRAULIC HEAD—Interval, in feet, is variable. Datum is sea level

----- DEPTH-AVERAGED DISSOLVED-SOLIDS CONCENTRATION—Equal to 10,000 milligrams per liter

— — — DEPTH-AVERAGED DISSOLVED-SOLIDS CONCENTRATION—Equal to 35,000 milligrams per liter

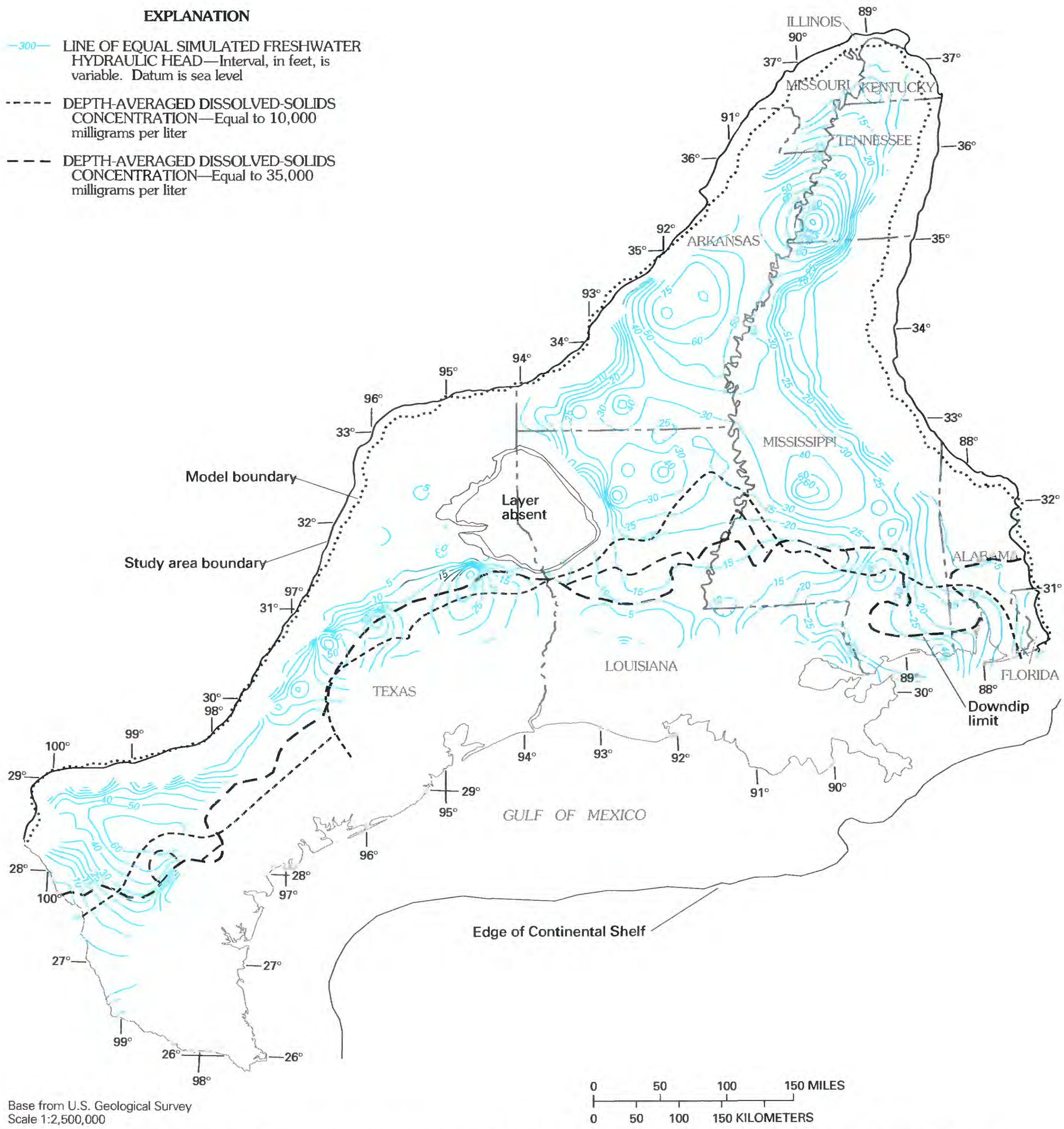


Plate 4G. Simulated drawdown from predevelopment to 1987 and depth-averaged dissolved-solids concentrations: Layer 5, middle Claiborne aquifer.

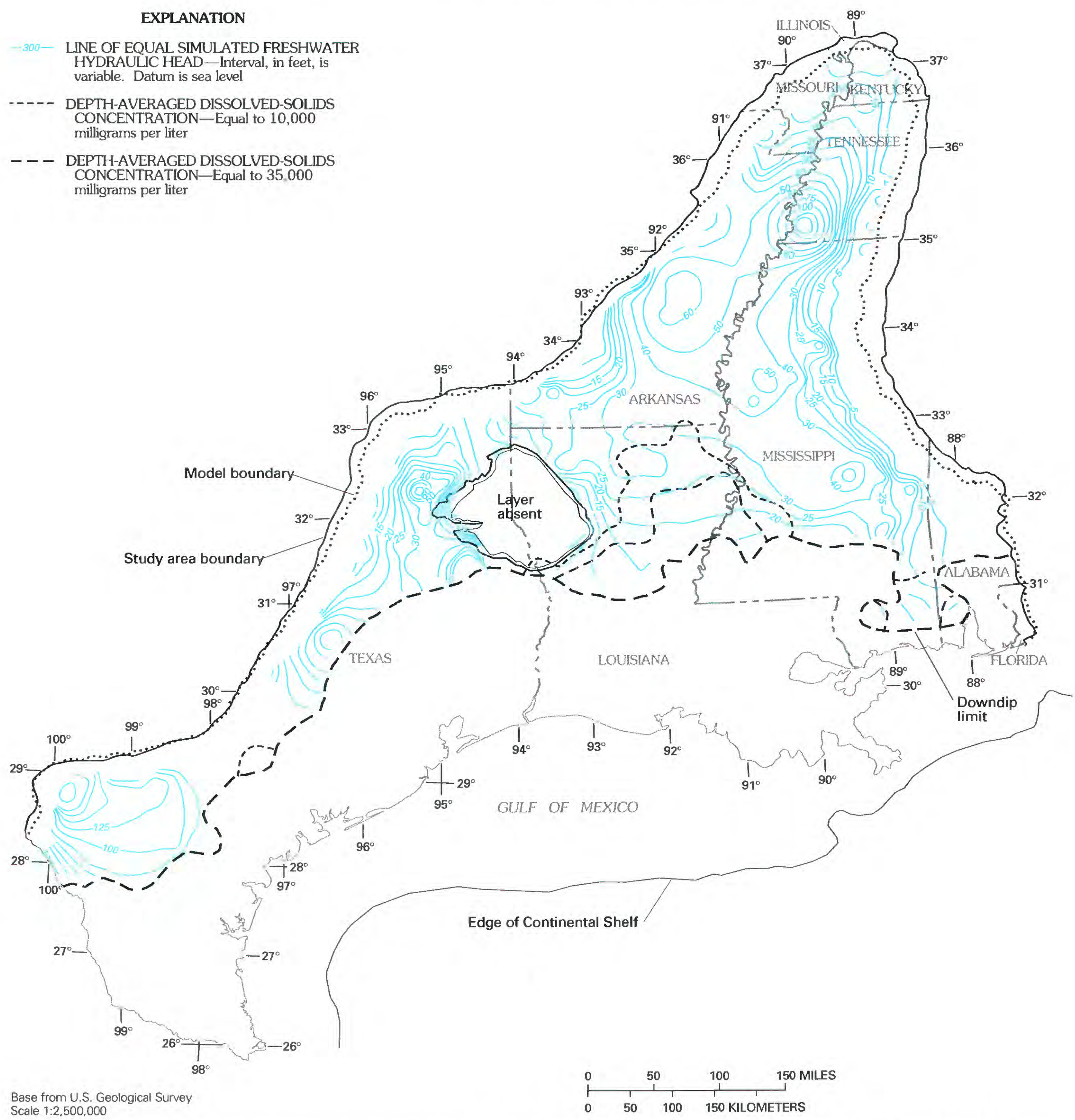


Plate 4H. Simulated drawdown from predevelopment to 1987 and depth-averaged dissolved-solids concentrations: Layer 4, lower Claiborne-upper Wilcox aquifer.

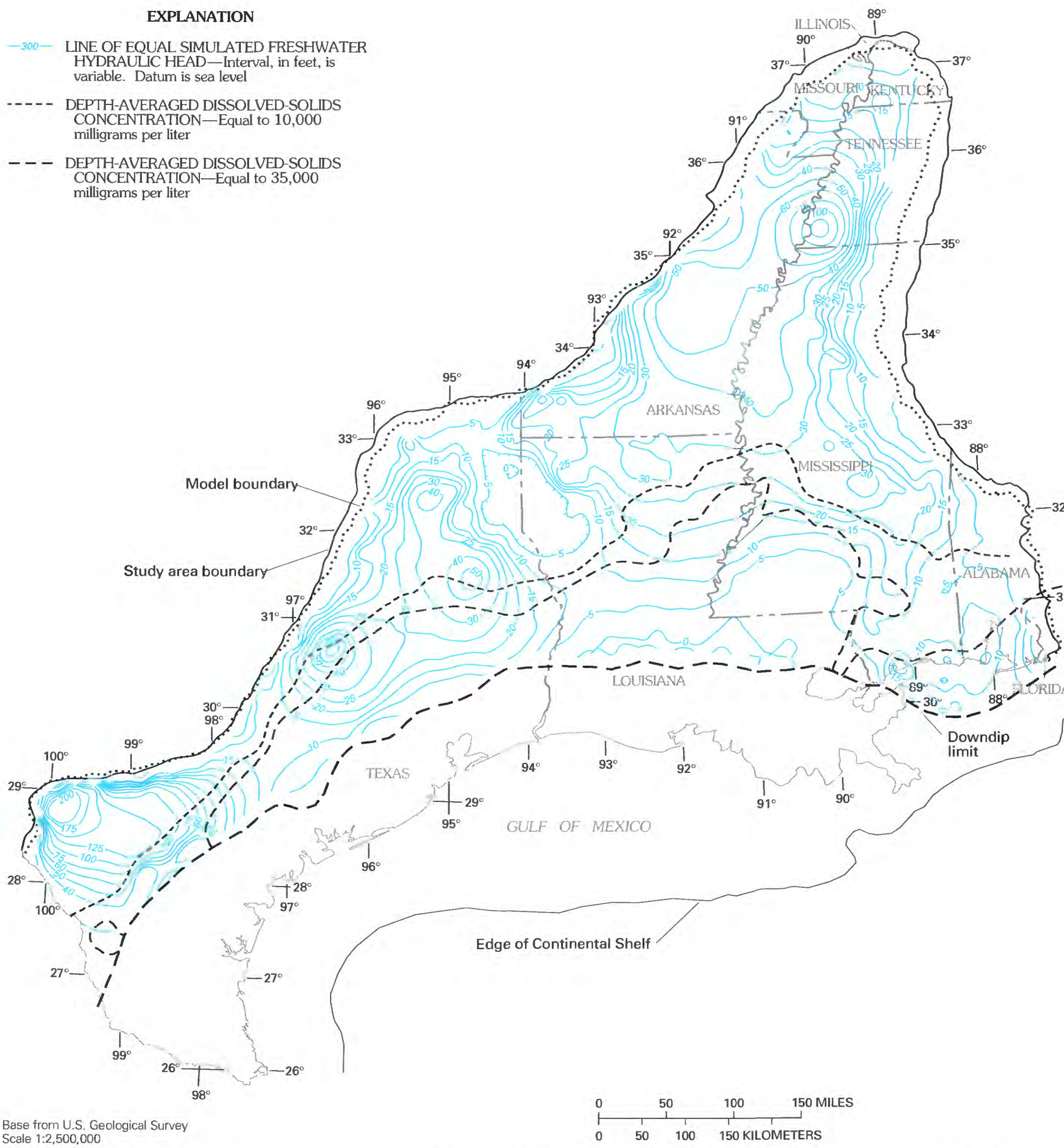


Plate 4I. Simulated drawdown from predevelopment to 1987 and depth-averaged dissolved-solids concentrations: Layer 3, middle Wilcox aquifer.

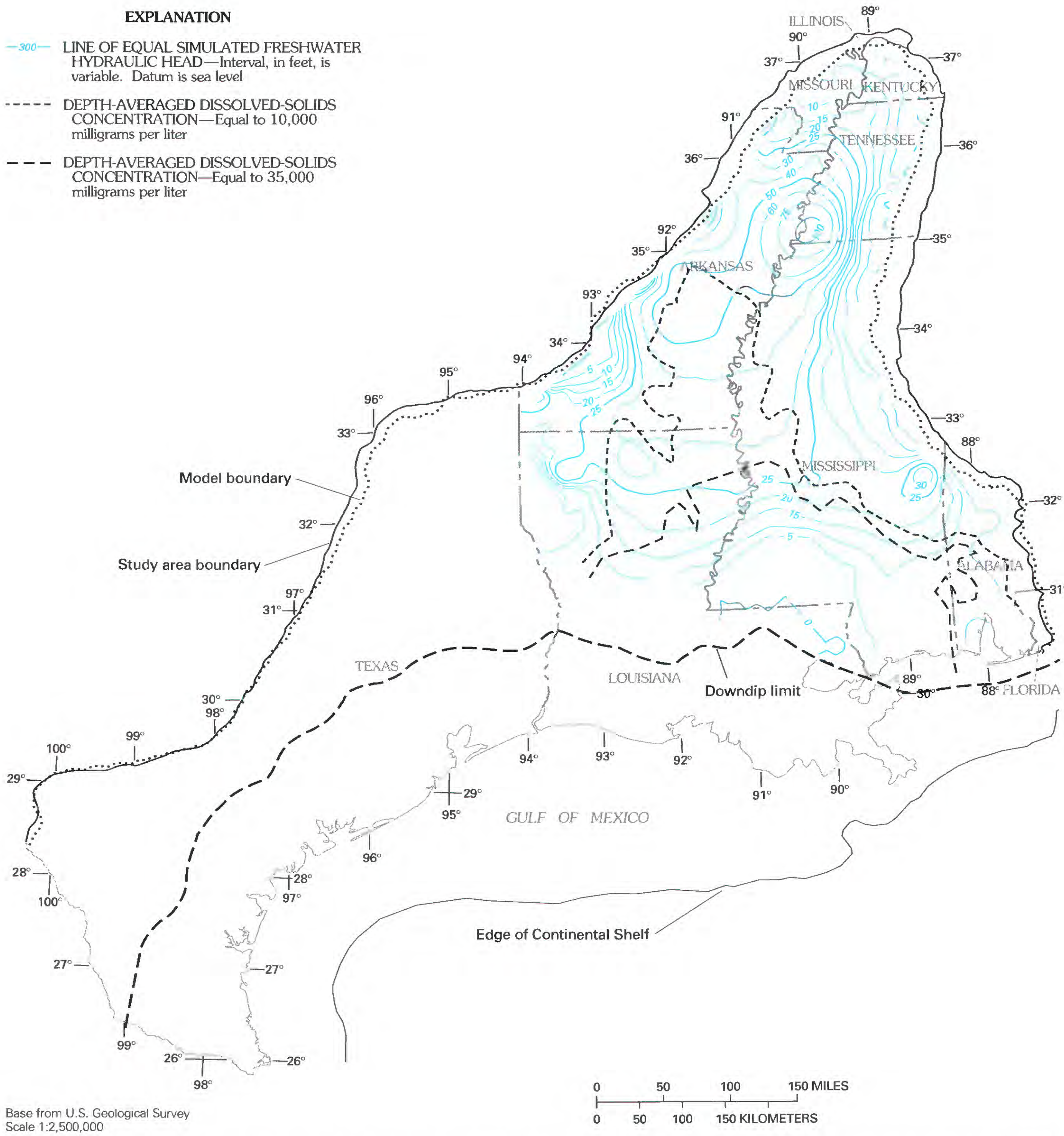


Plate 4J. Simulated drawdown from predevelopment to 1987 and depth-averaged dissolved-solids concentrations: Layer 2, middle Wilcox aquifer.

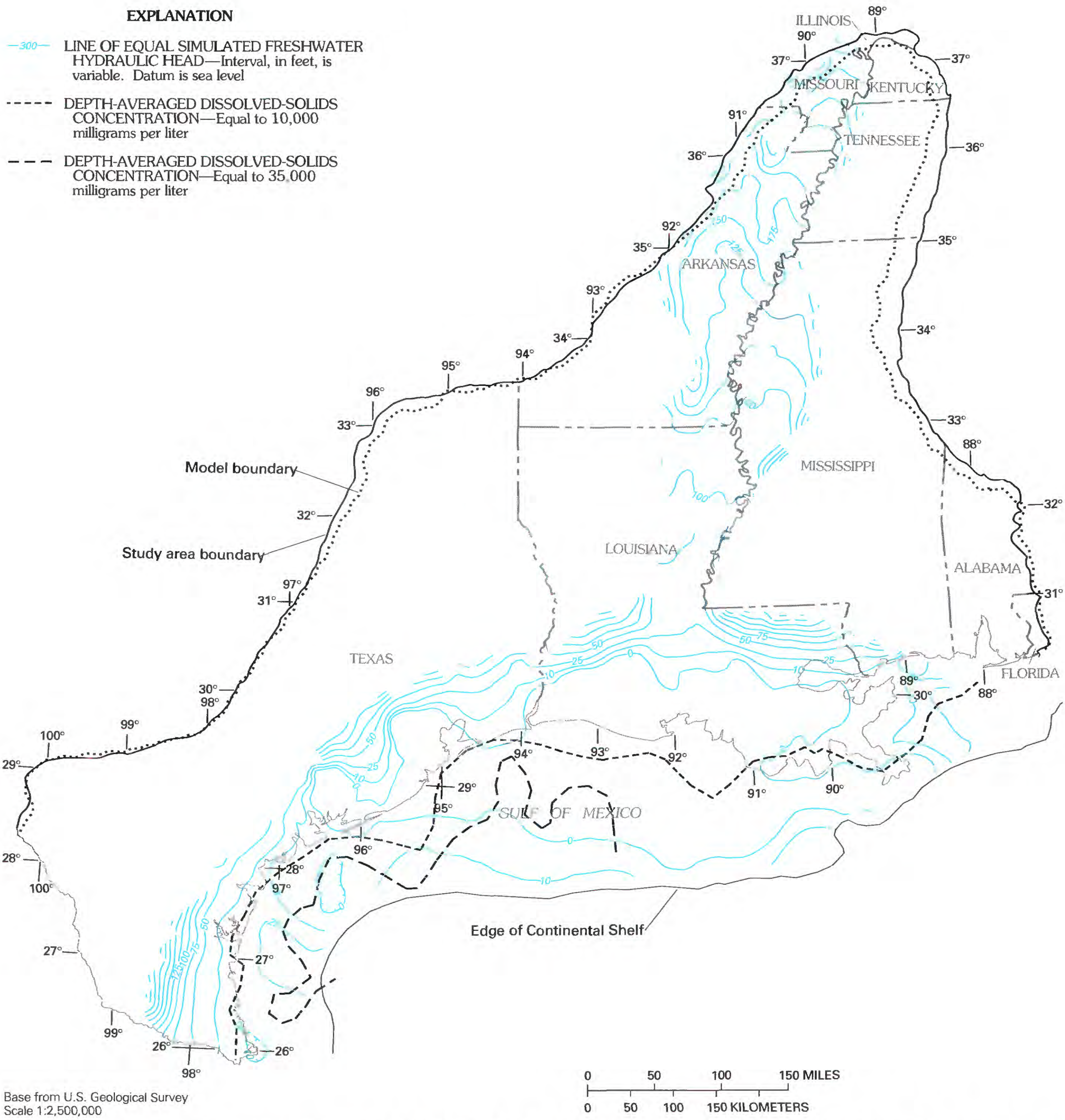


Plate 5A. Simulated freshwater hydraulic head, 1987, Layer 11: permeable zone A and Mississippi River Valley alluvial aquifer.

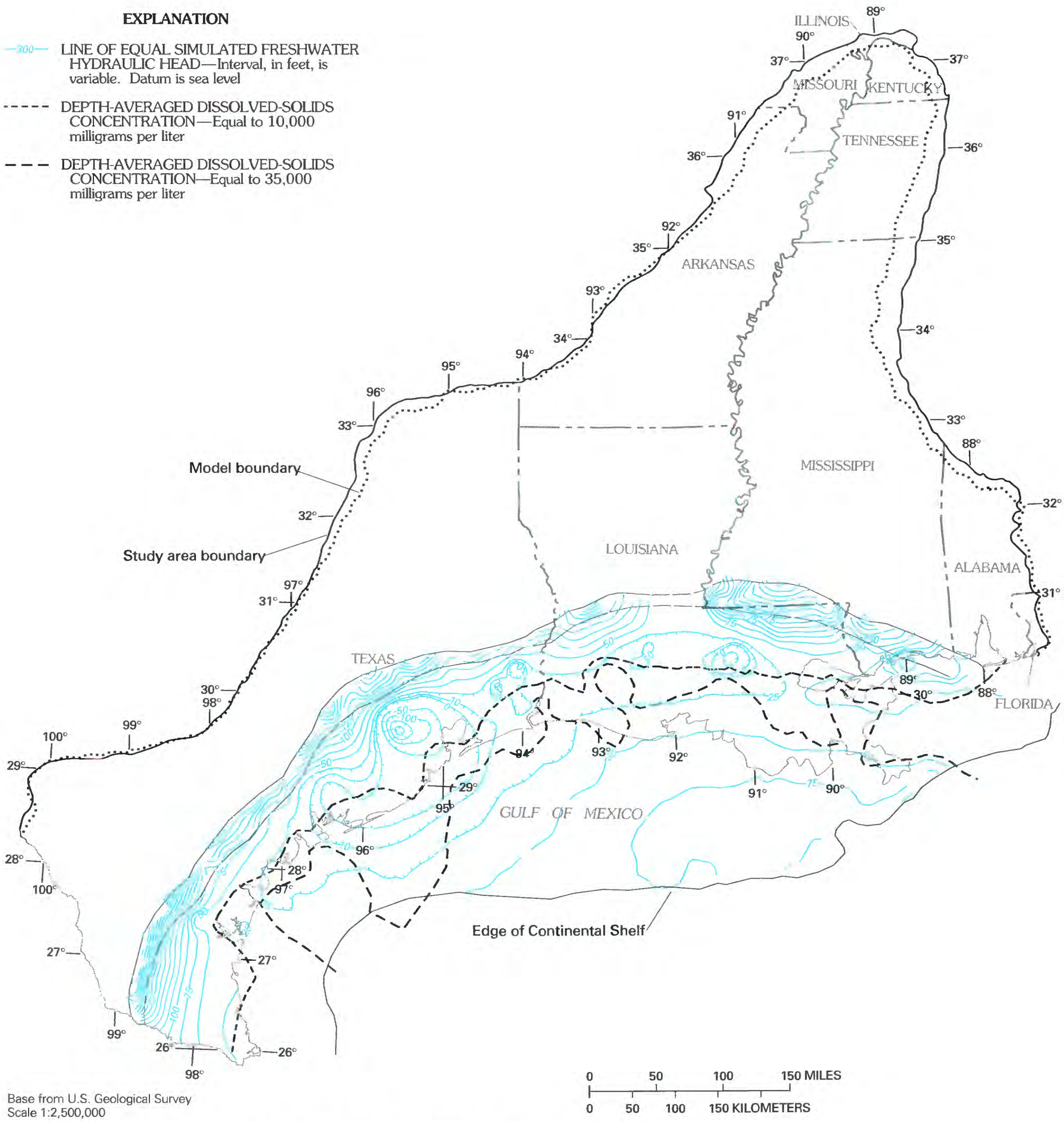


Plate 5B. Simulated freshwater hydraulic head, 1987: Layer 10, permeable zone B.

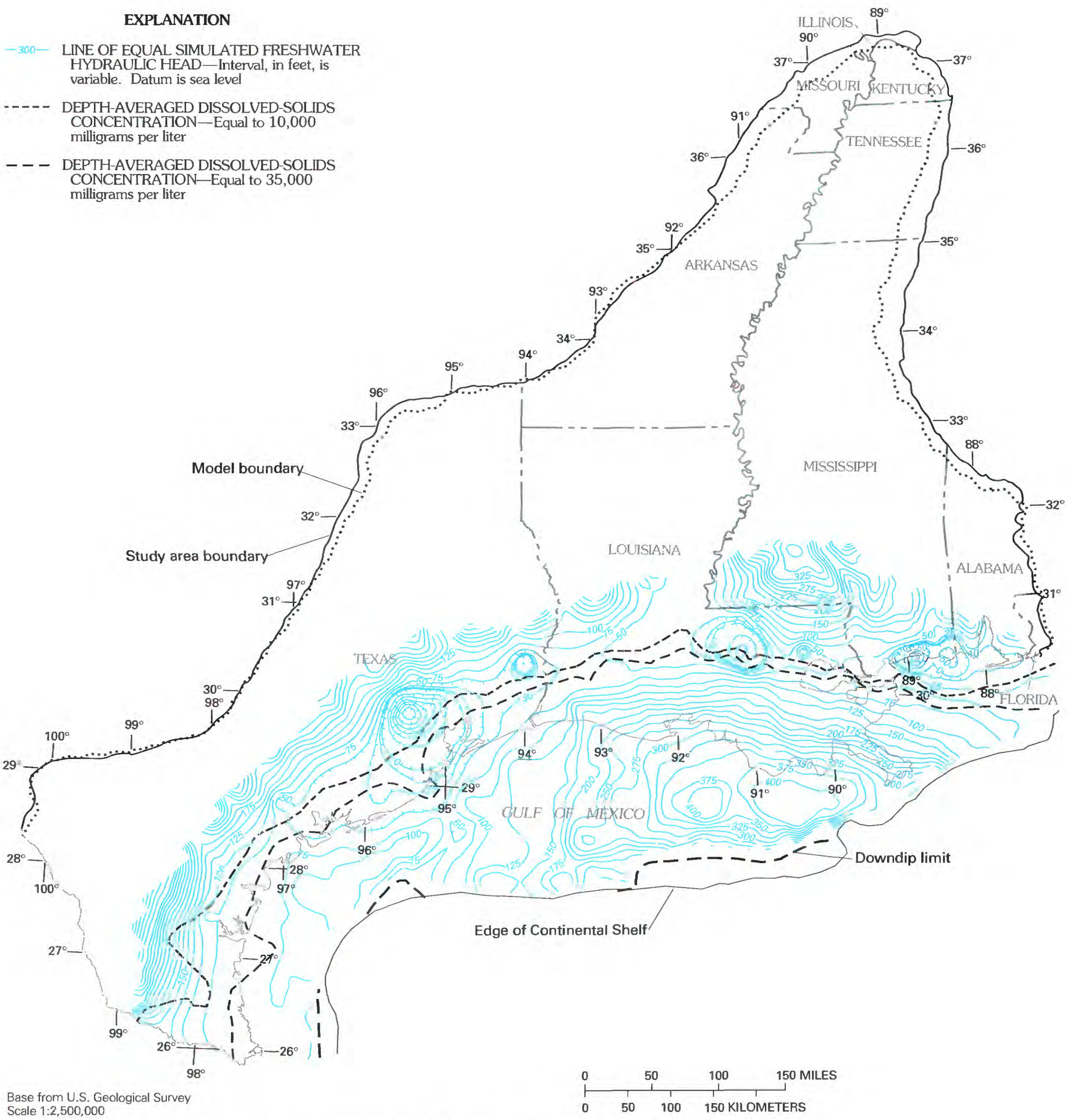


Plate 5C. Simulated freshwater hydraulic head: 1987, Layer 9, permeable zone C.

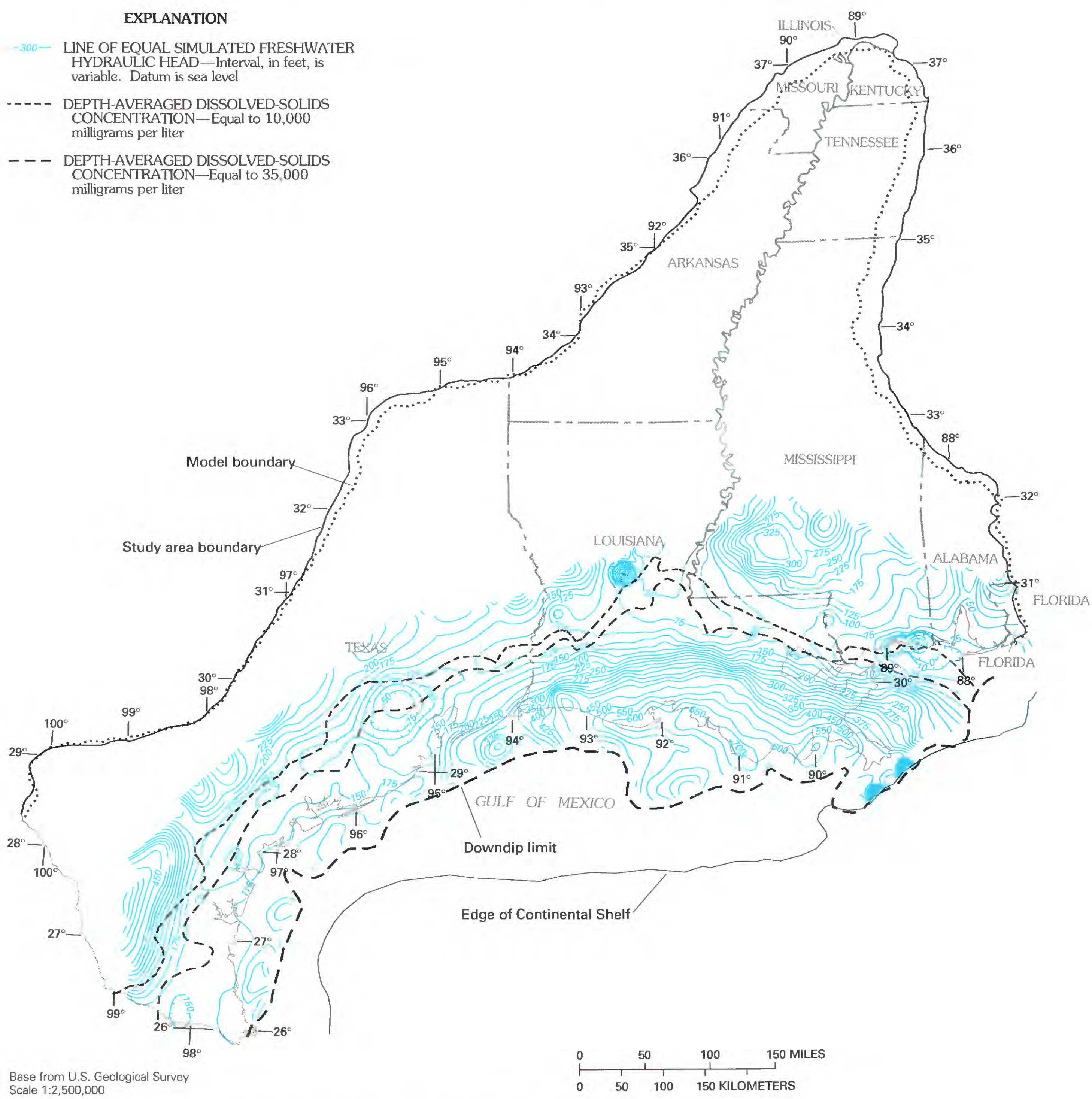
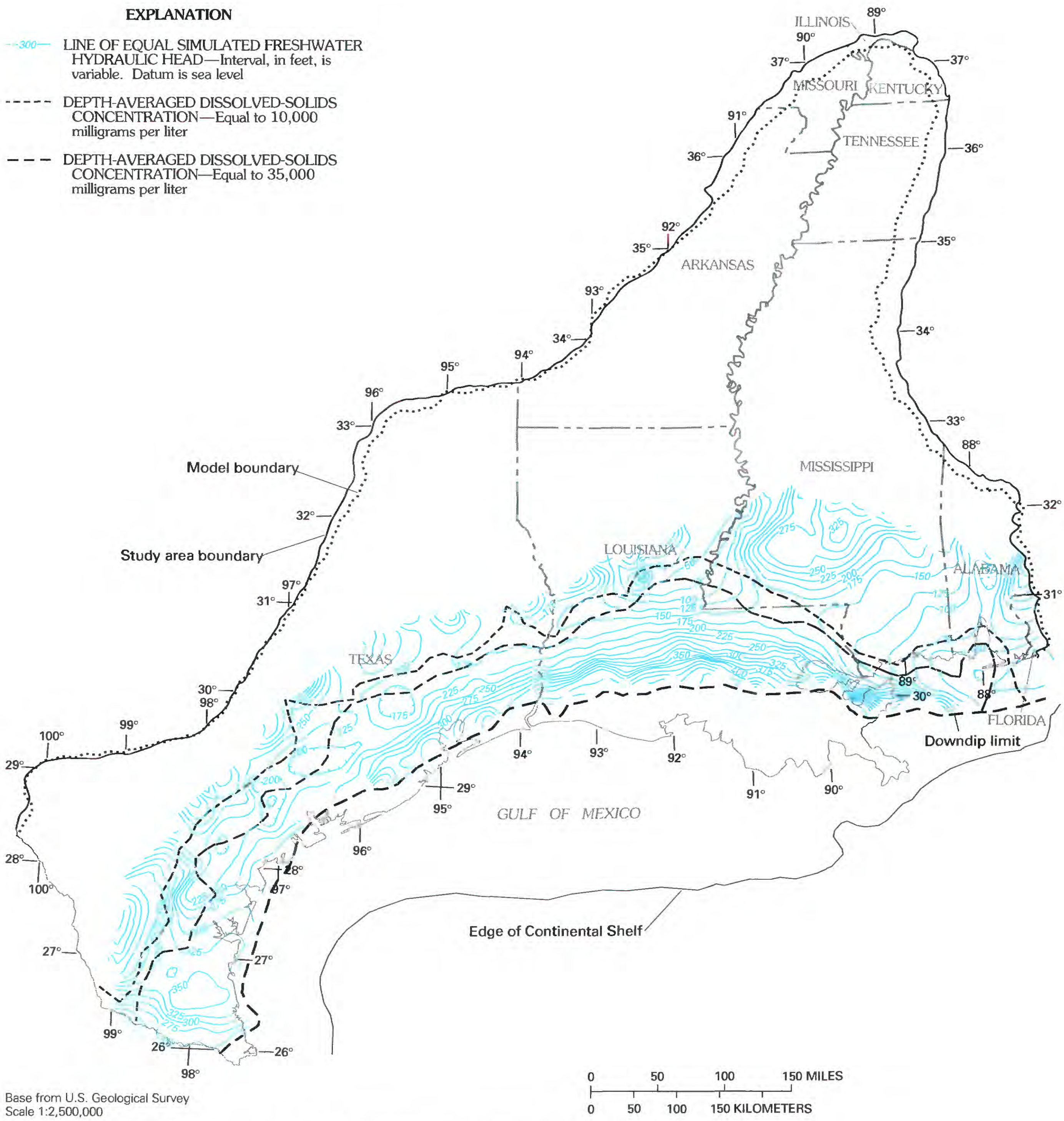


Plate 5D. Simulated freshwater hydraulic-head, 1987: Layer 8, permeable zone D.



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

Plate 5E. Simulated freshwater hydraulic head, 1987: Layer 7, permeable zone E.

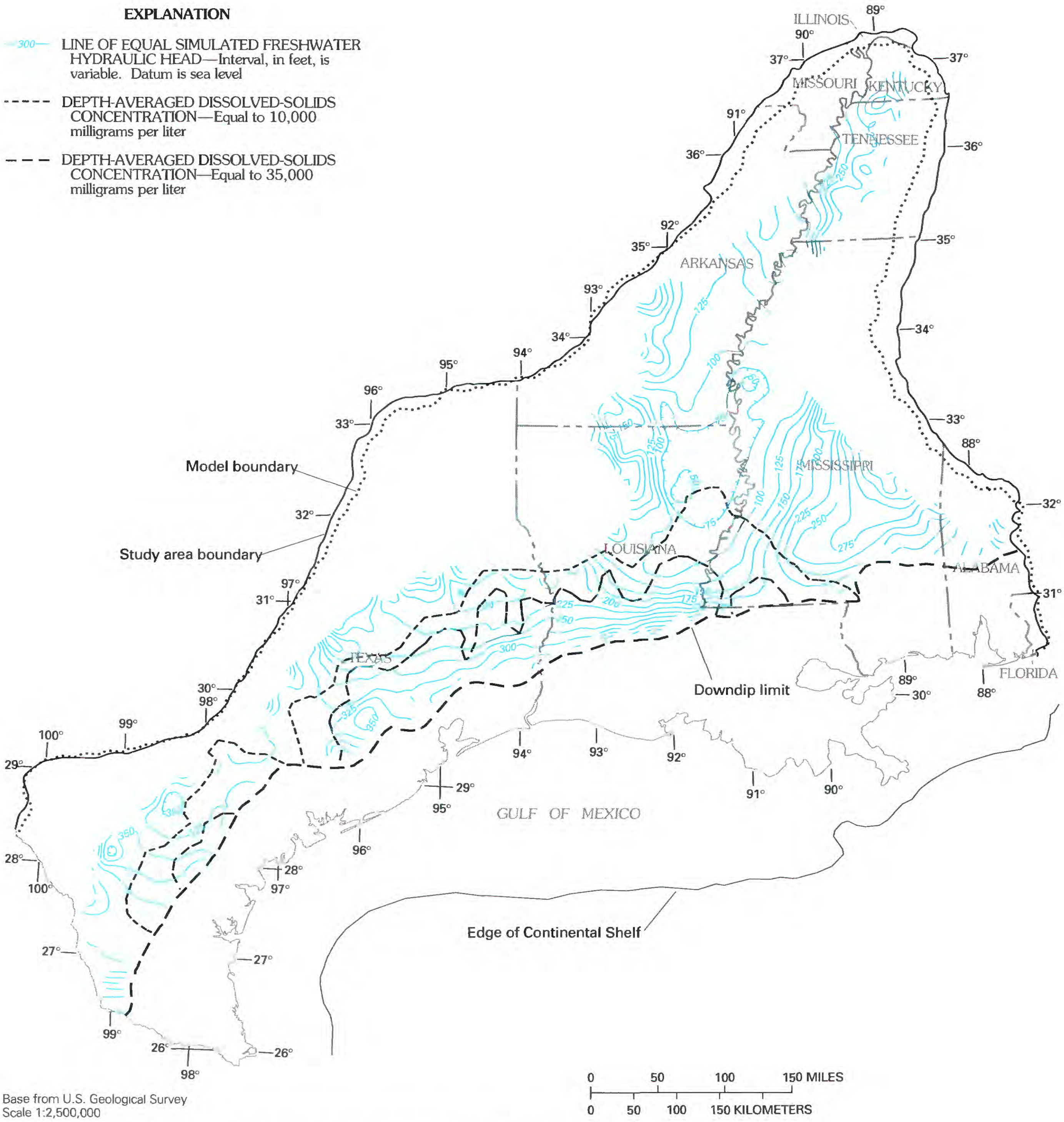


Plate 5F. Simulated freshwater hydraulic head, 1987: Layer 6, upper Claiborne aquifer.

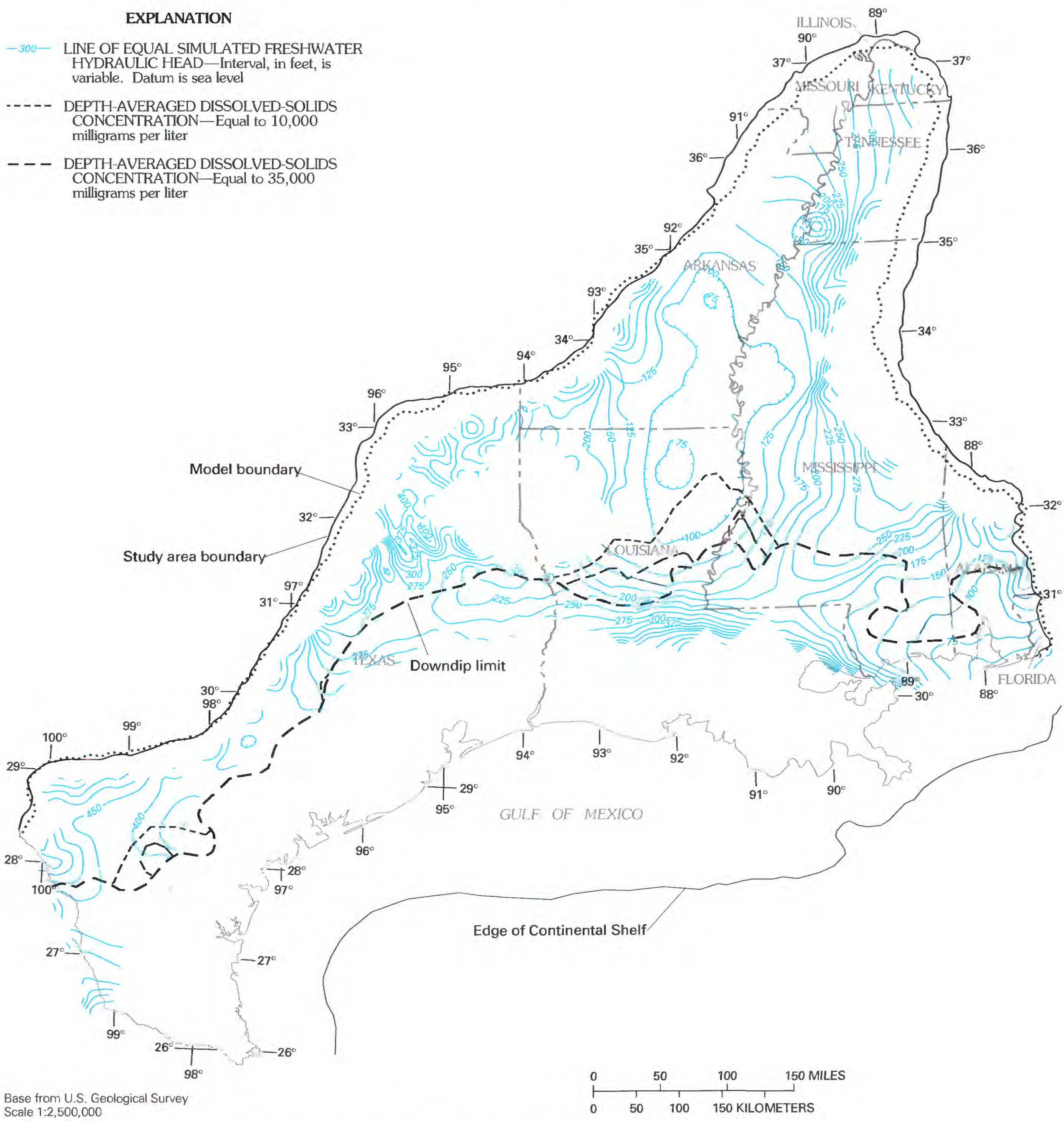


Plate 5G. Simulated freshwater hydraulic head, 1987: Layer 5, middle Claiborne aquifer.

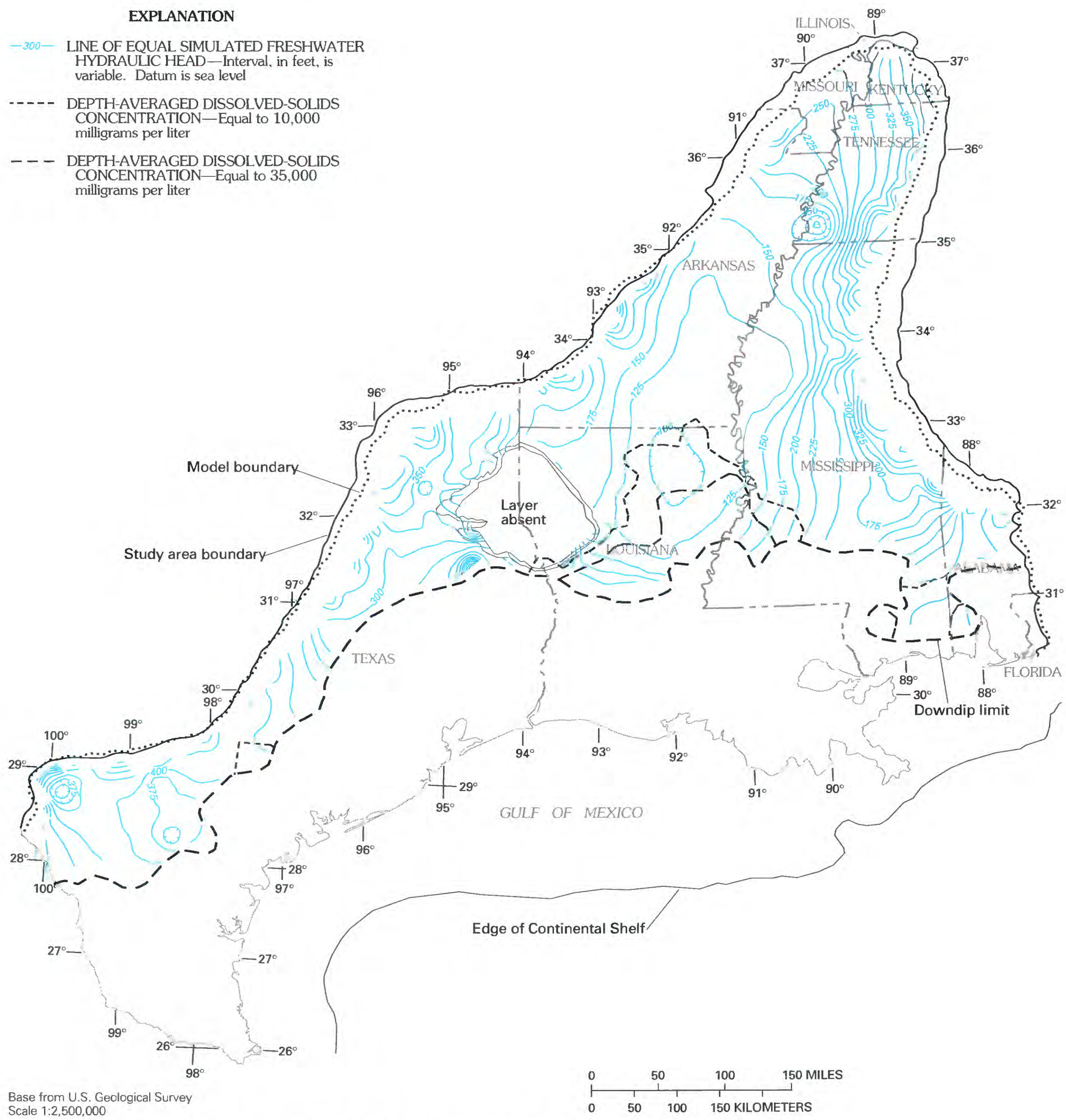


Plate 5H. Simulated freshwater hydraulic head, 1987: Layer 4, lower Claiborne-upper Wilcox aquifer.

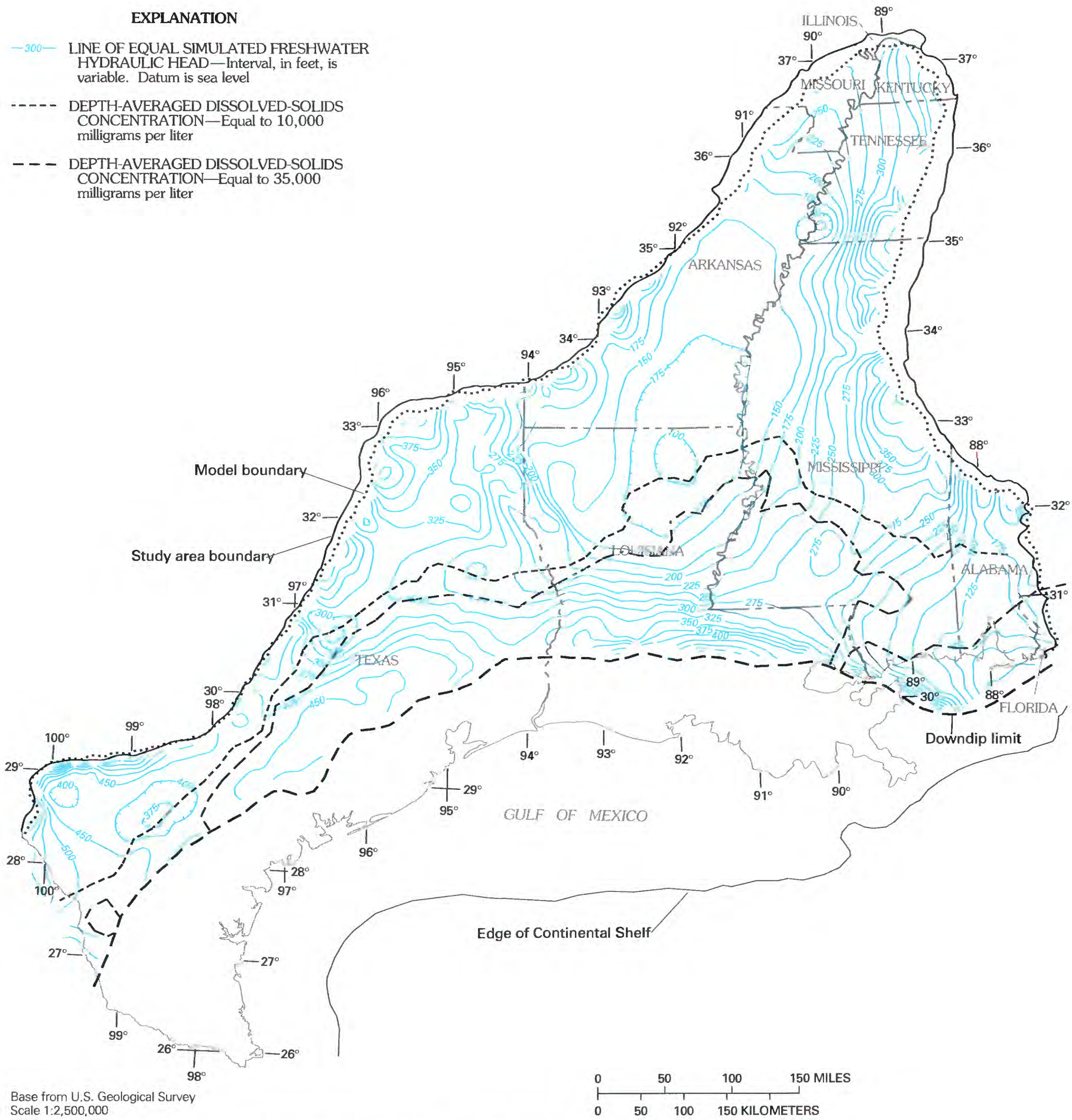


Plate 5I. Simulated freshwater hydraulic head, 1987: Layer 3, middle Wilcox aquifer.

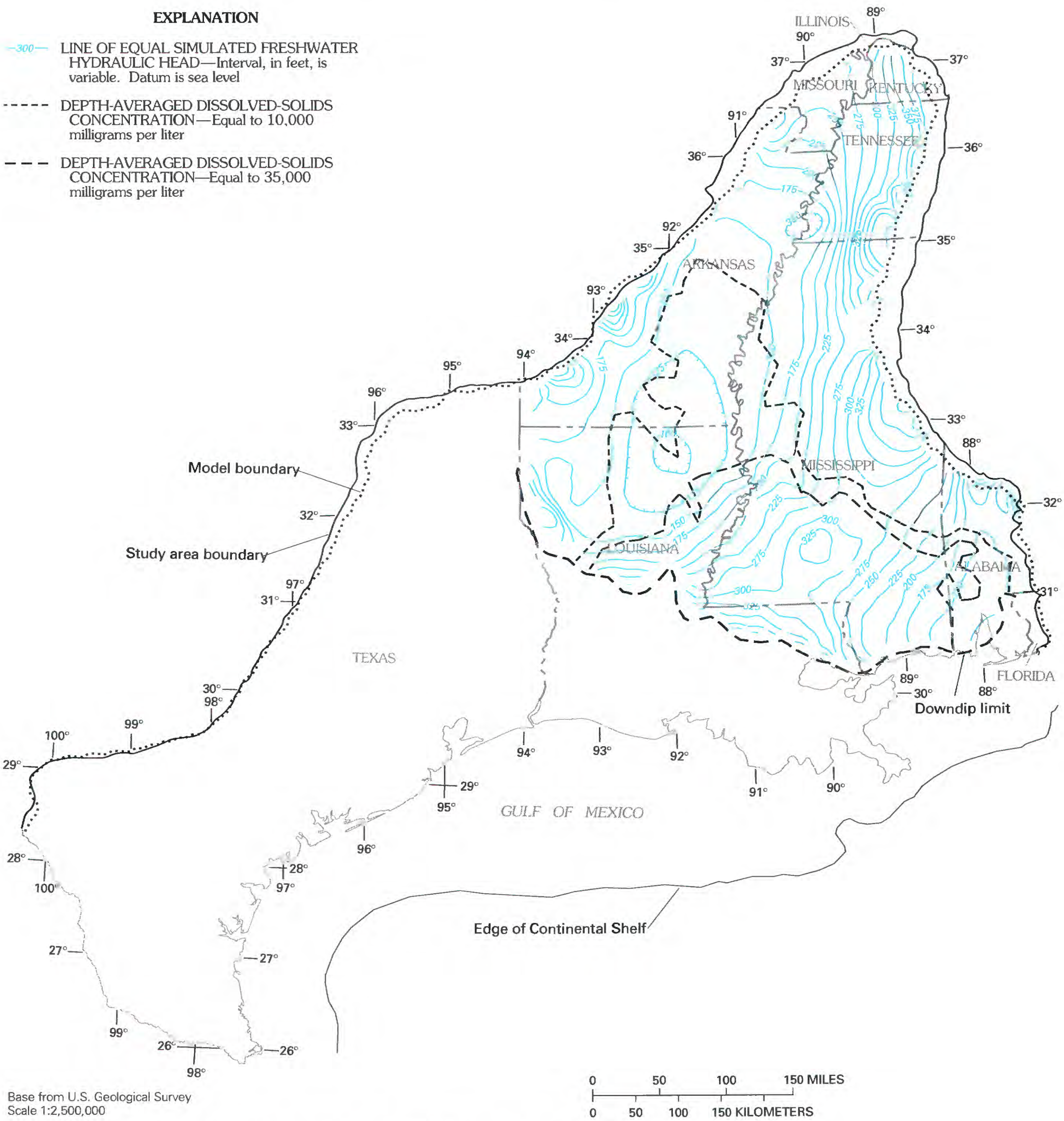


Plate 5J. Simulated freshwater hydraulic head, 1987: Layer 2, lower Wilcox aquifer.

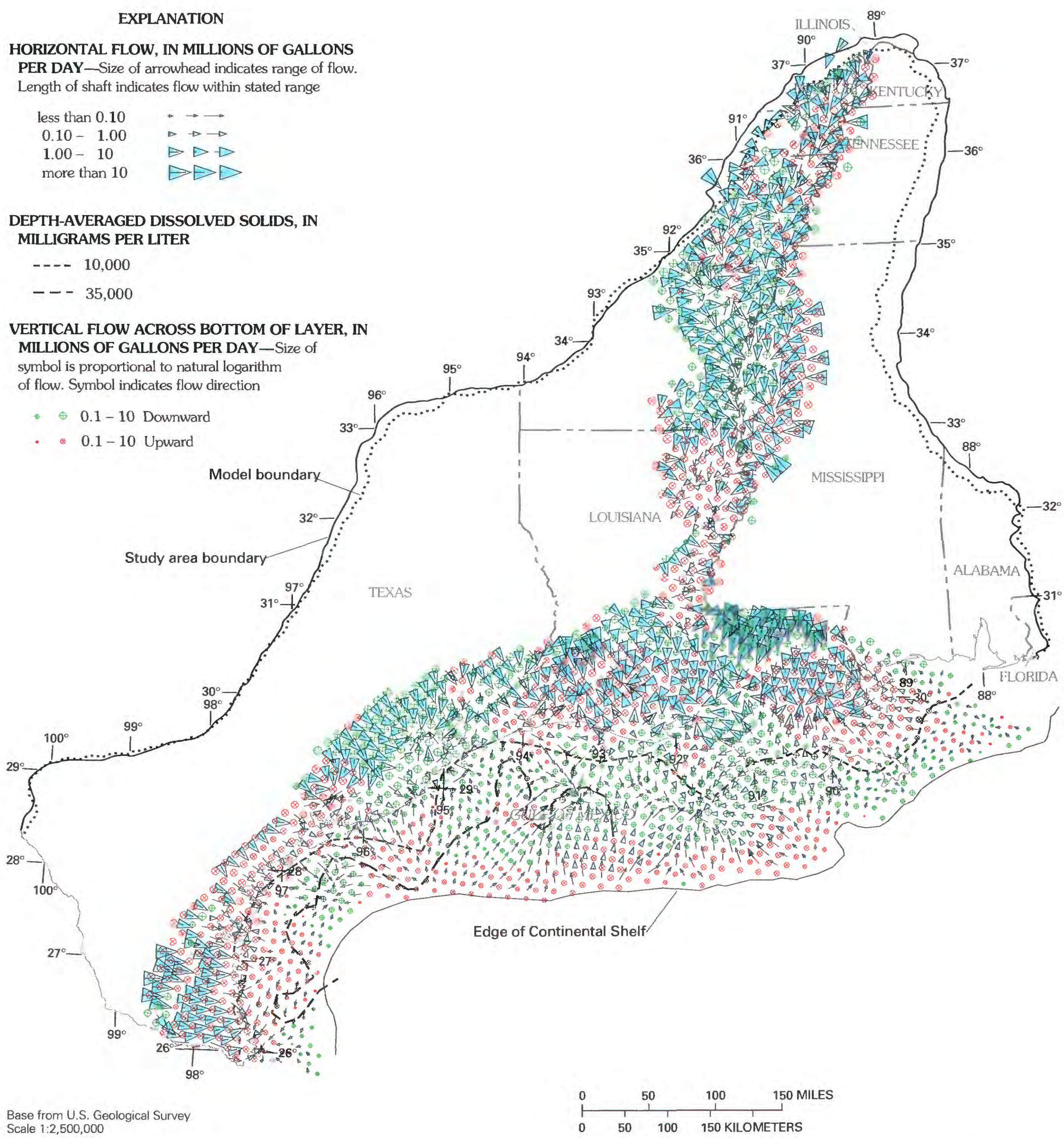


Plate 6A. Simulated 1985 ground-water flow directions and magnitudes and depth-averaged dissolved-solids concentrations: Layer 11, permeable zone A, and Mississippi River Valley alluvial aquifer.

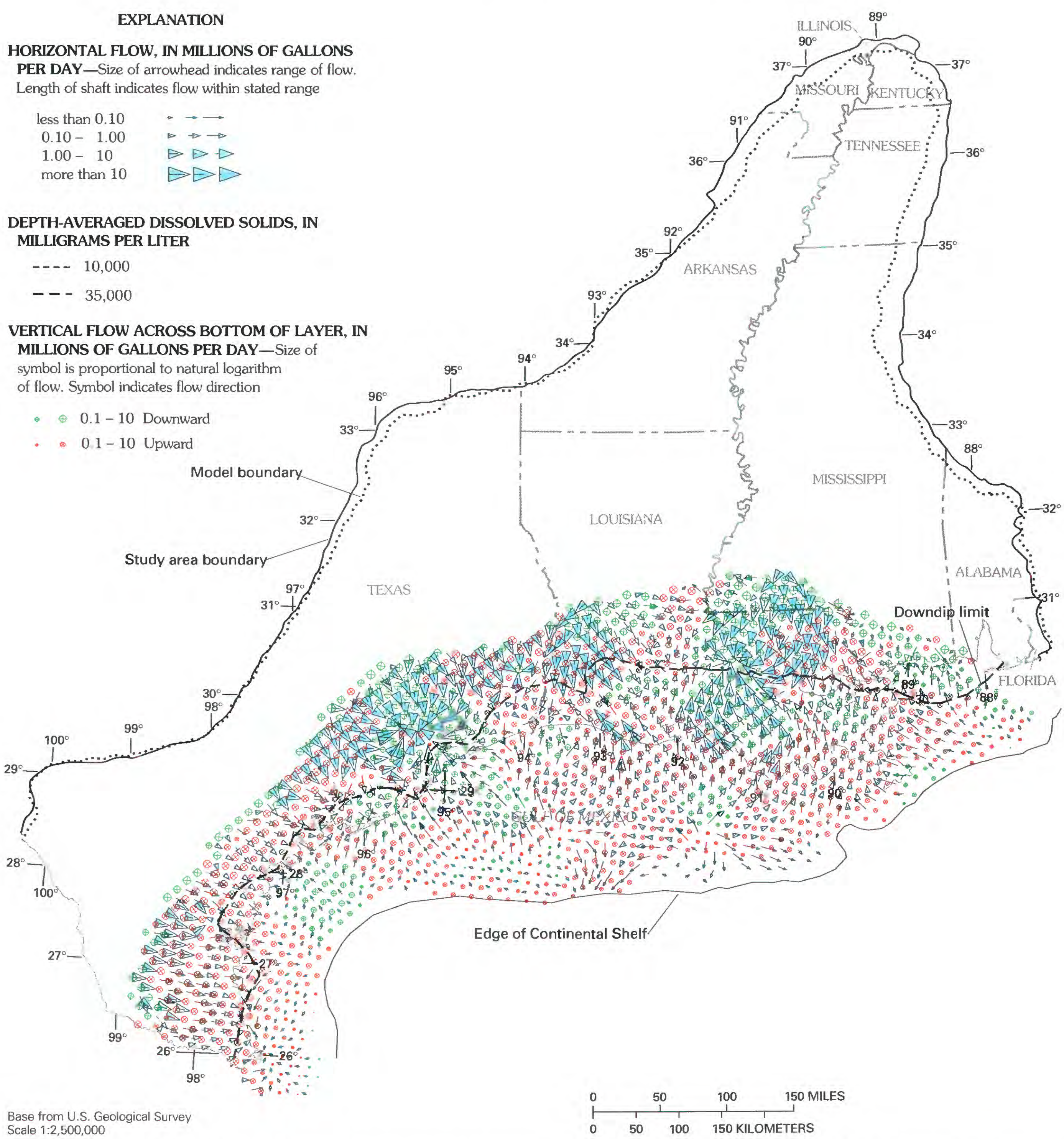


Plate 6B. Simulated 1987 ground-water flow directions and magnitudes and depth-averaged dissolved-solids concentrations: Layer 10, permeable zone B.

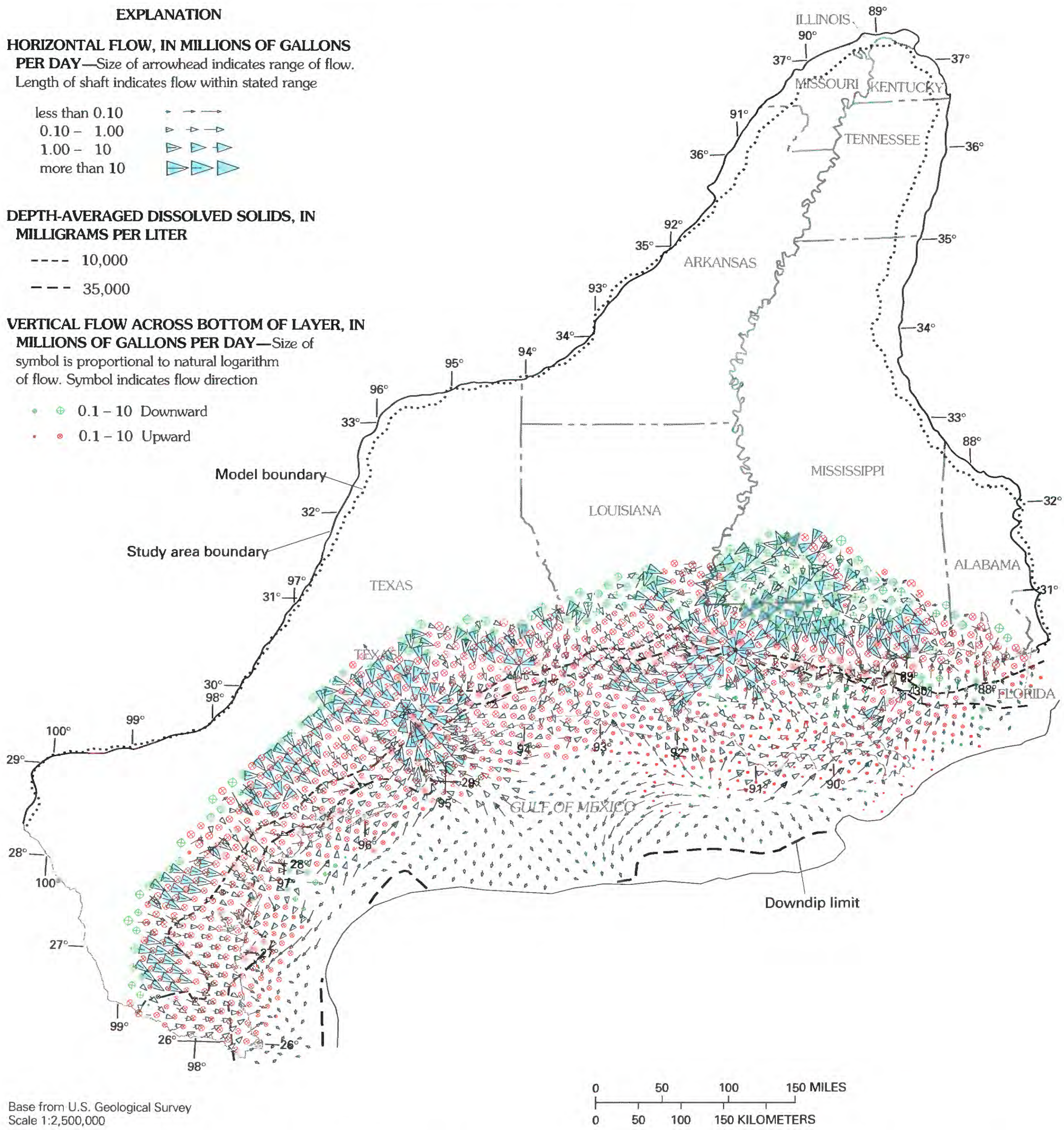


Plate 6C. Simulated 1987 ground-water flow directions and magnitudes and depth-averaged dissolved-solids concentrations: Layer 9, permeable zone C.

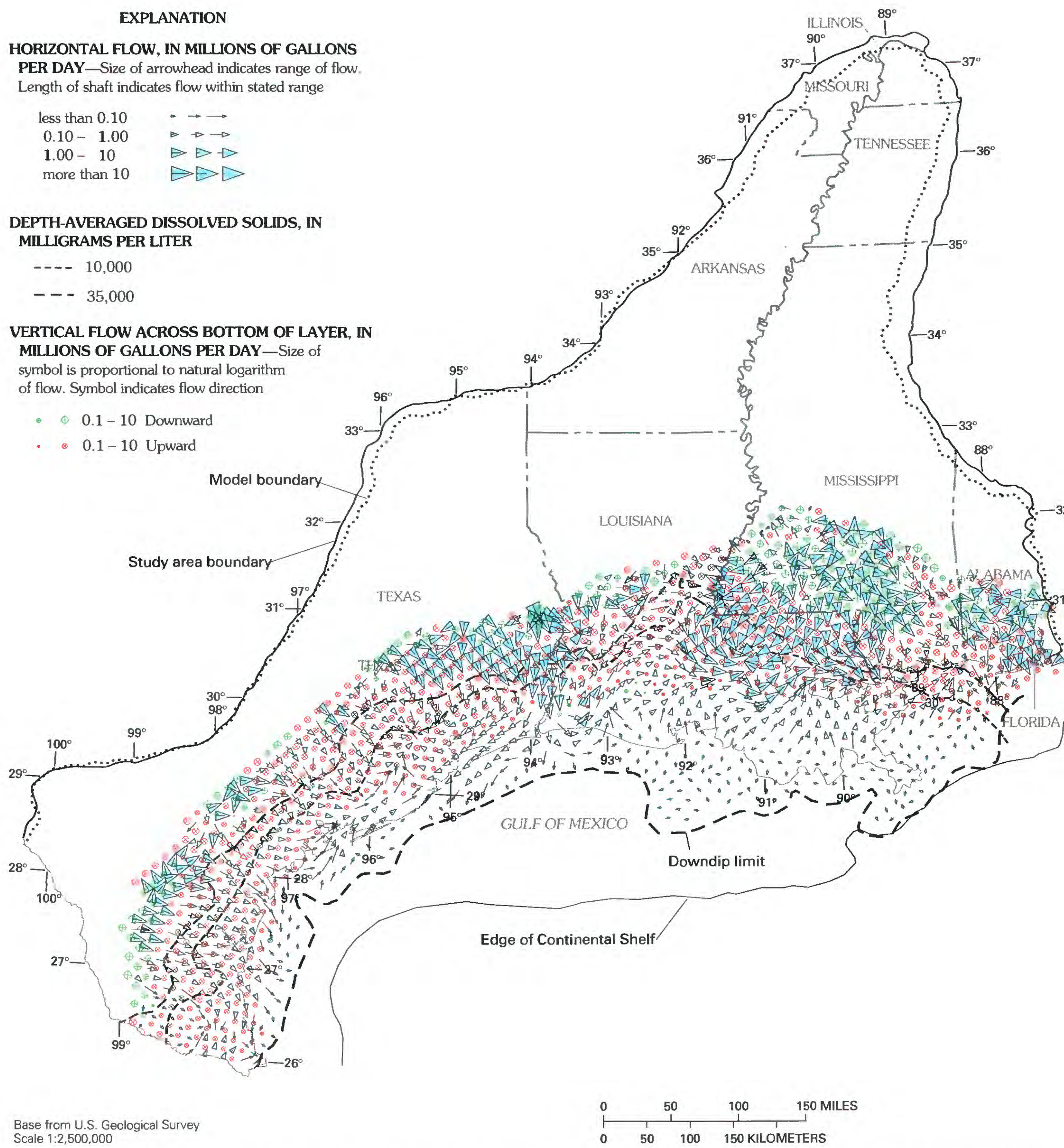


Plate 6D. Simulated 1987 ground-water flow directions and magnitudes and depth-averaged dissolved-solids concentrations: Layer 8, permeable zone D.

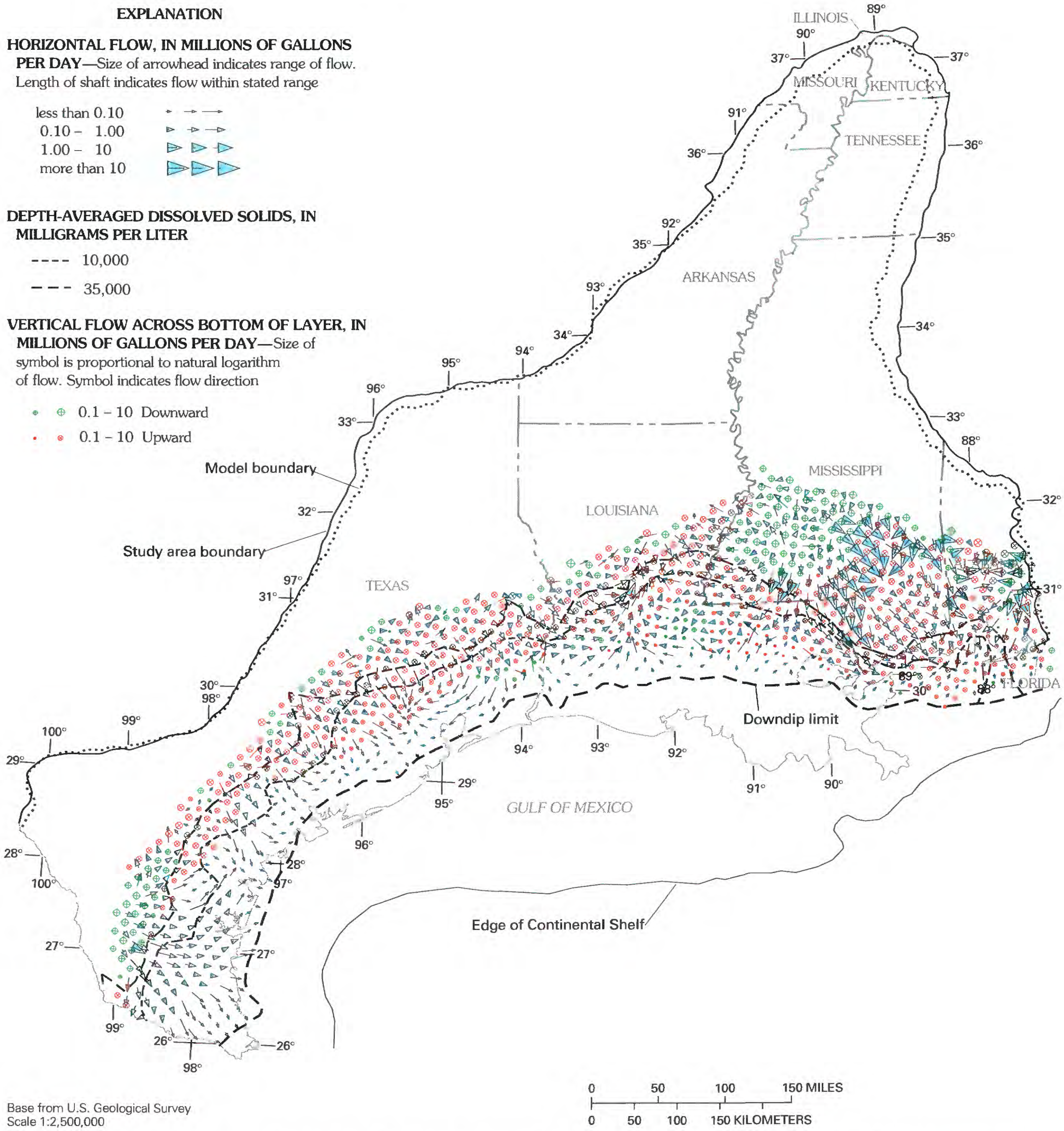


Plate 6E. Simulated 1987 ground-water flow directions and magnitudes and depth-averaged dissolved-solids concentrations: Layer 7, permeable zone E.

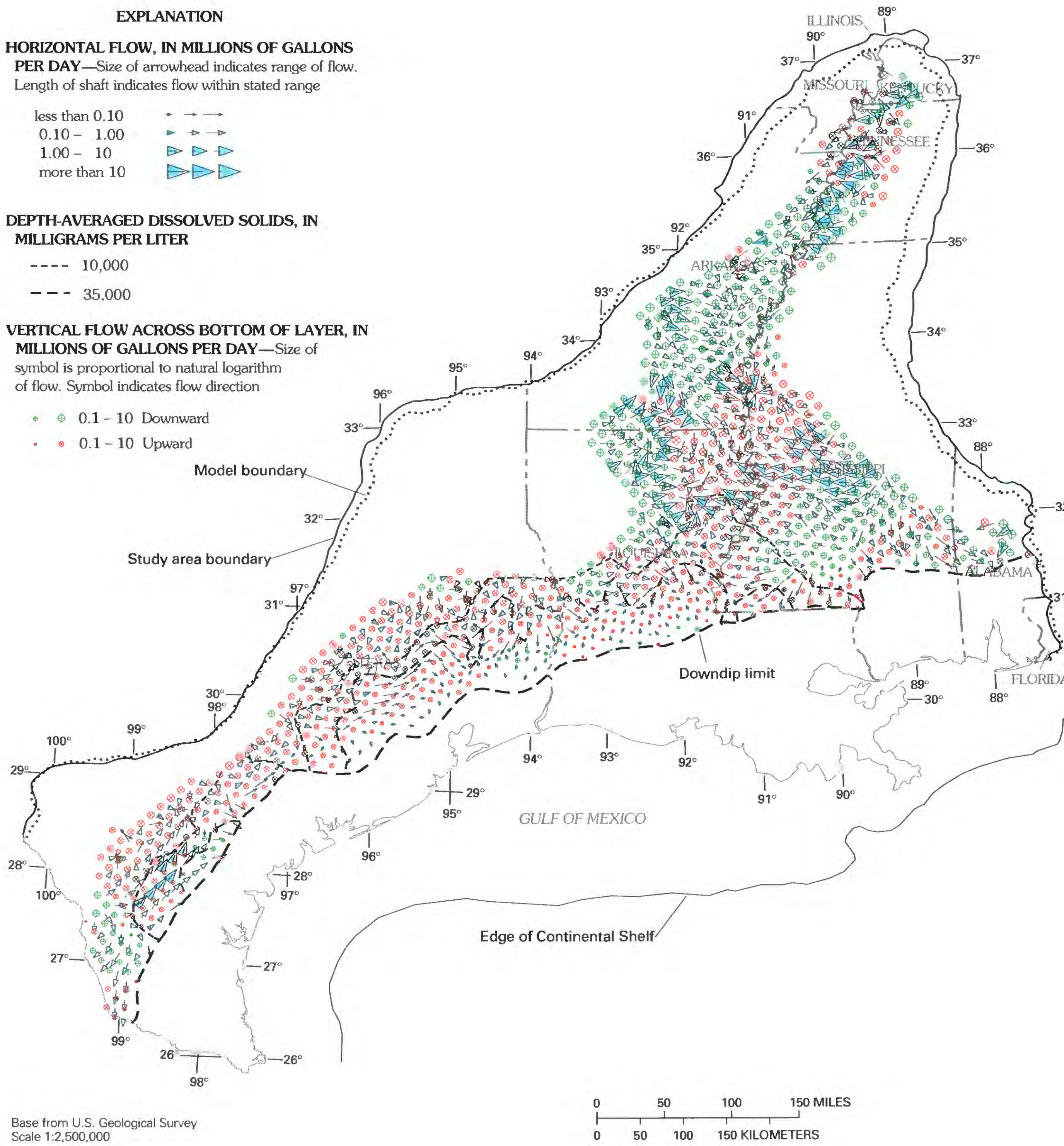


Plate 6F. Simulated 1987 ground-water flow directions and magnitudes and depth-averaged dissolved-solids concentrations: Layer 6, upper Claiborne aquifer.

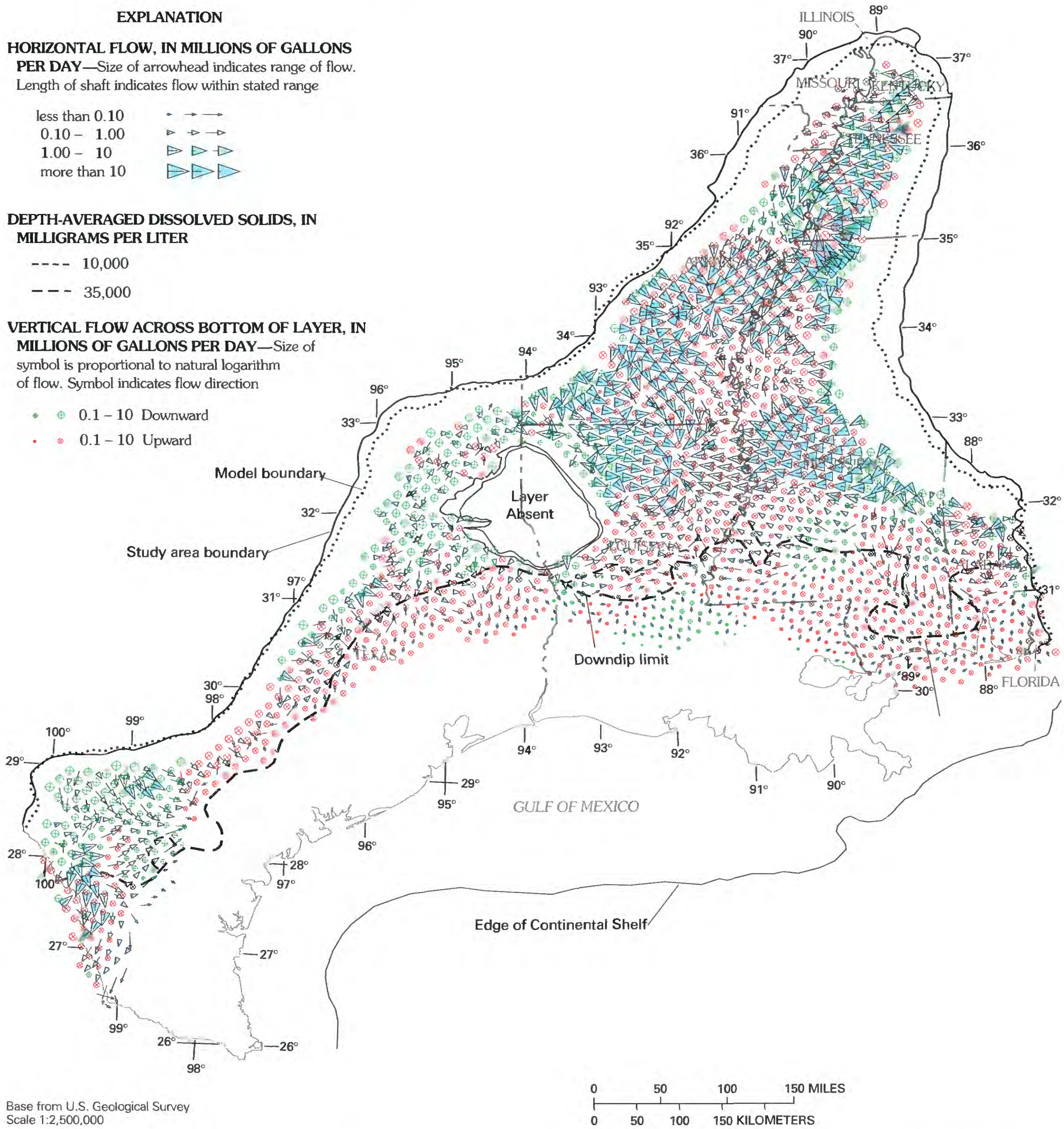


Plate 6G. Simulated 1987 ground-water flow directions and magnitudes and depth-averaged dissolved-solids concentrations: Layer 5, middle Claiborne aquifer.

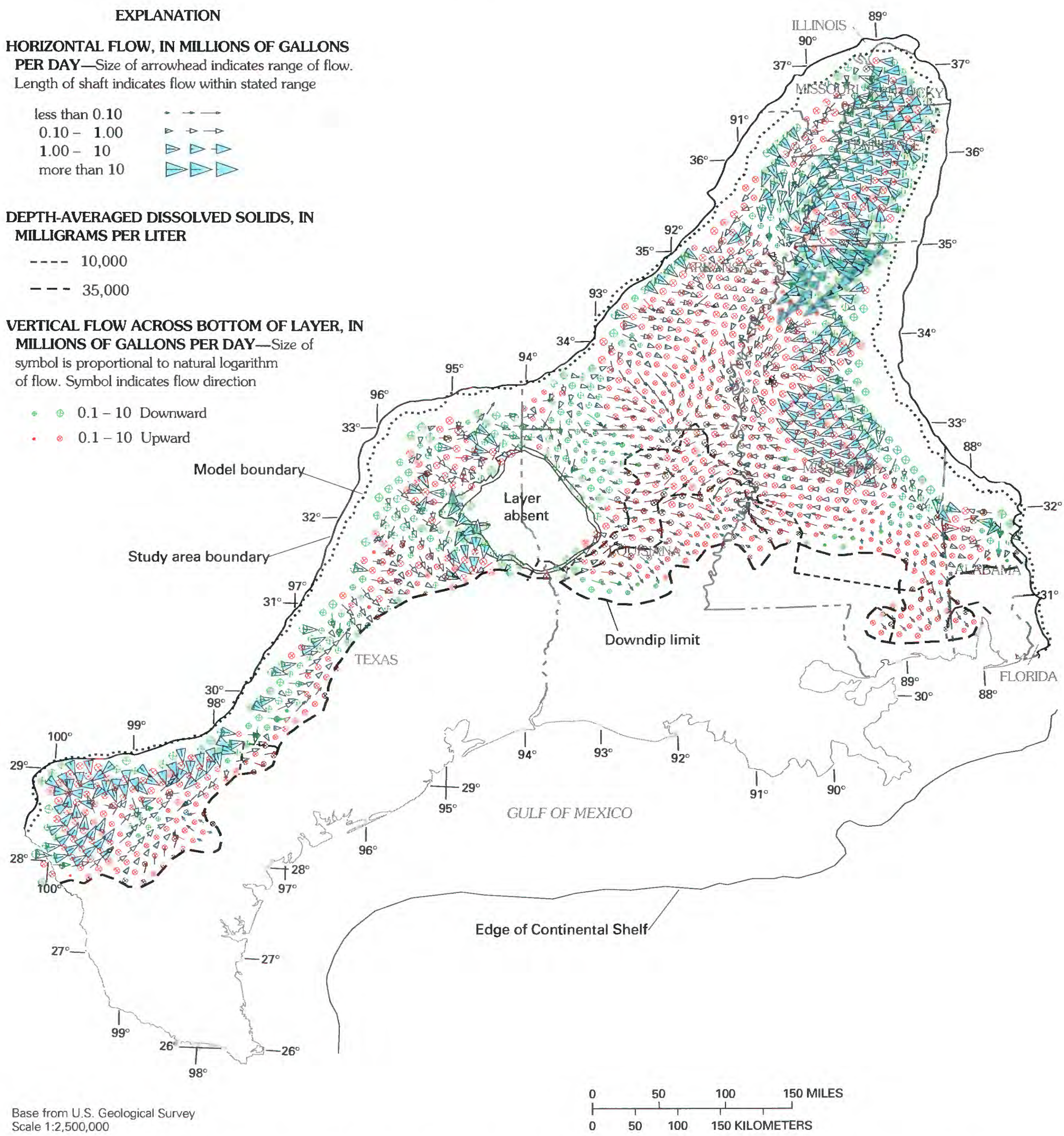


Plate 6H. Simulated 1987 ground-water flow directions and magnitudes and depth-averaged dissolved-solids concentrations: Layer 4, lower Claiborne-upper Wilcox aquifer.

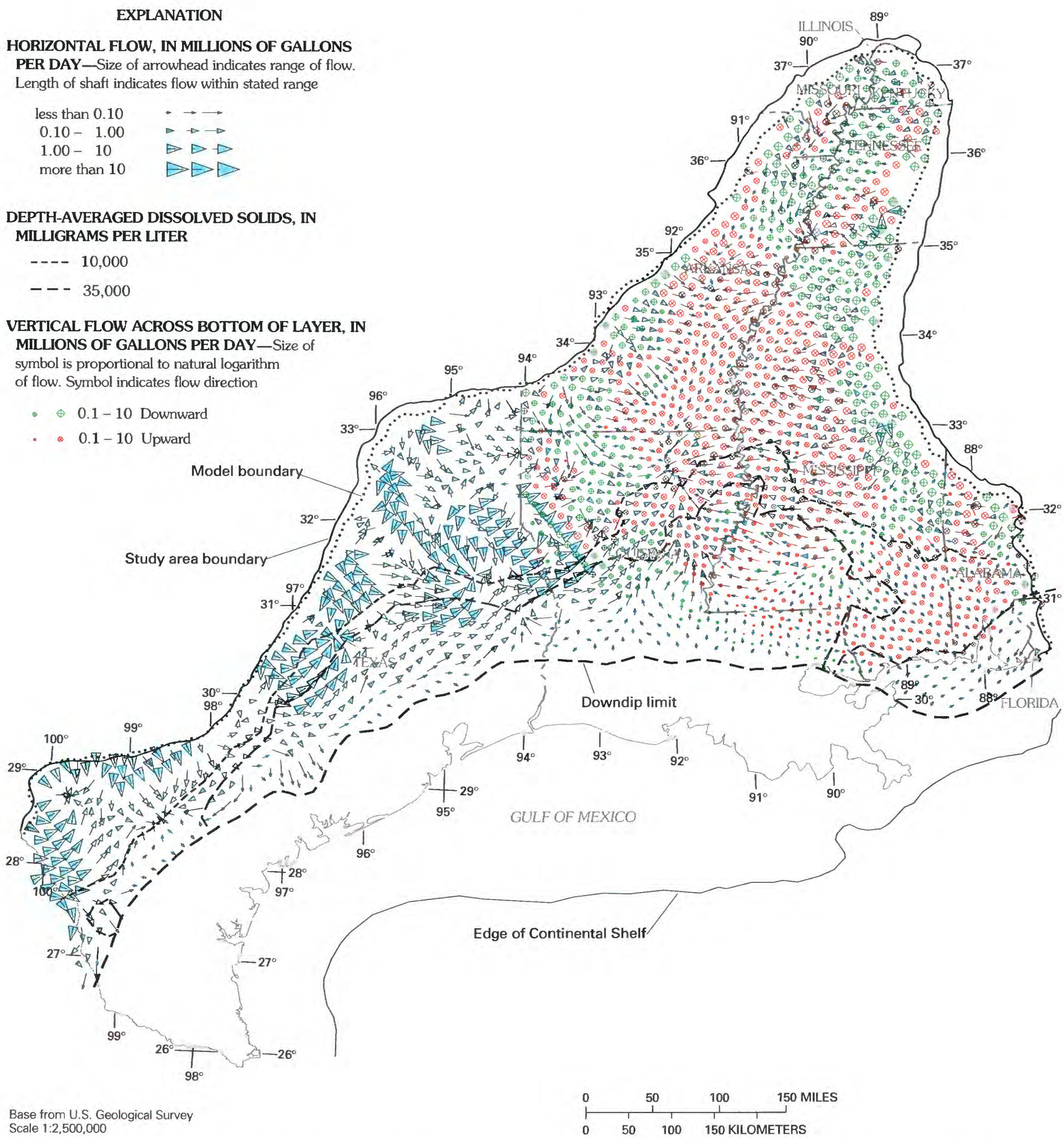


Plate 6I. Simulated 1987 ground-water flow directions and magnitudes and depth-averaged dissolved-solids concentrations: Layer 3, middle Wilcox aquifer.

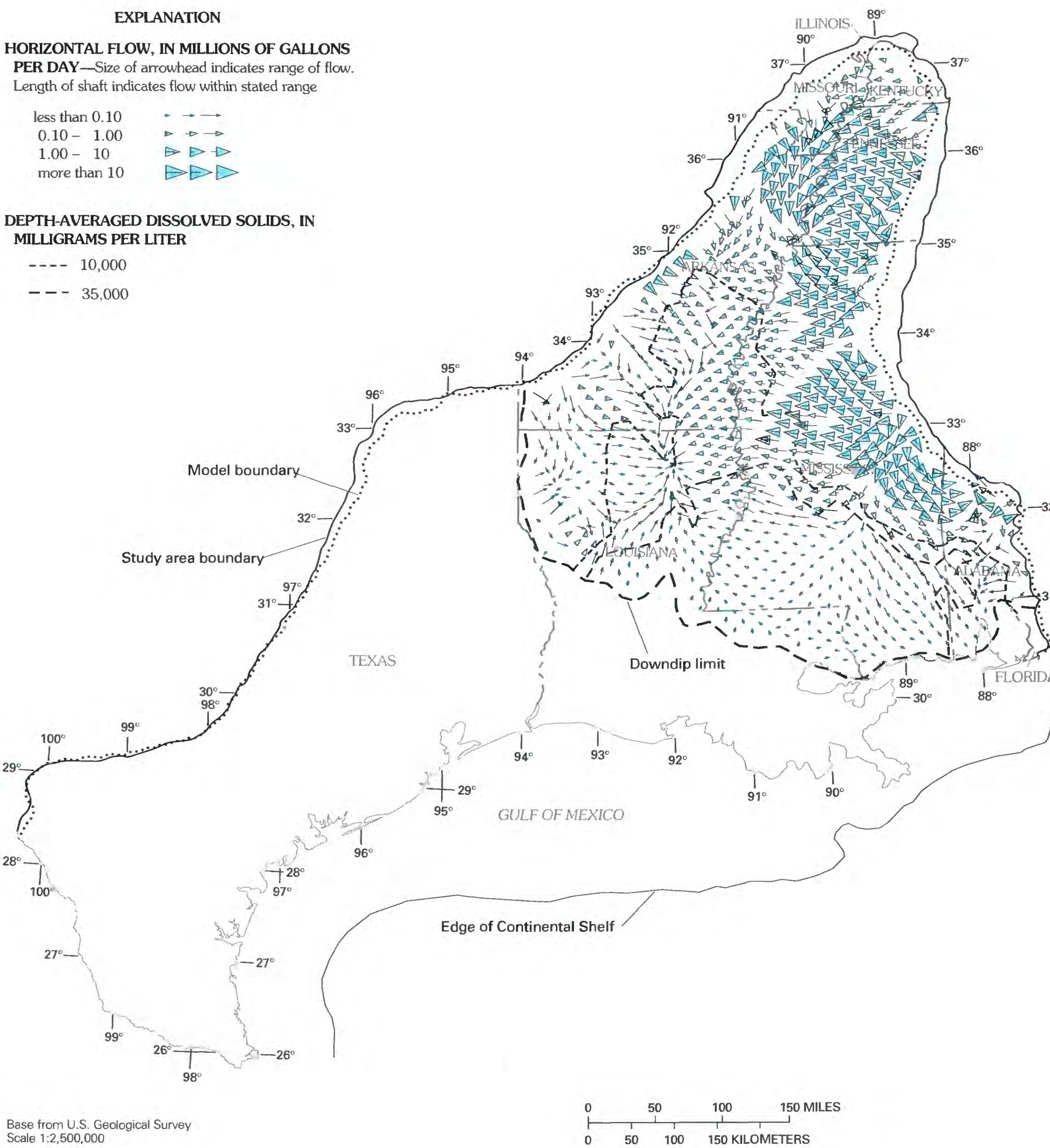


Plate 6J. Simulated 1987 ground-water flow directions and magnitudes and depth-averaged dissolved-solids concentrations: Layer 2, lower Wilcox aquifer.

Selected Series of U.S. Geological Survey Publications

Books and Other Publications

Professional Papers report scientific data and interpretations of lasting scientific interest that cover all facets of USGS investigations and research.

Bulletins contain significant data and interpretations that are of lasting scientific interest but are generally more limited in scope or geographic coverage than Professional Papers.

Water-Supply Papers are comprehensive reports that present significant interpretive results of hydrologic investigations of wide interest to professional geologists, hydrologists, and engineers. The series covers investigations in all phases of hydrology, including hydrogeology, availability of water, quality of water, and use of water.

Circulars are reports of programmatic or scientific information of an ephemeral nature; many present important scientific information of wide popular interest. Circulars are distributed at no cost to the public.

Fact Sheets communicate a wide variety of timely information on USGS programs, projects, and research. They commonly address issues of public interest. Fact Sheets generally are two or four pages long and are distributed at no cost to the public.

Reports in the **Digital Data Series (DDS)** distribute large amounts of data through digital media, including compact disc-read-only memory (CD-ROM). They are high-quality, interpretive publications designed as self-contained packages for viewing and interpreting data and typically contain data sets, software to view the data, and explanatory text.

Water-Resources Investigations Reports are papers of an interpretive nature made available to the public outside the formal USGS publications series. Copies are produced on request (unlike formal USGS publications) and are also available for public inspection at depositories indicated in USGS catalogs.

Open-File Reports can consist of basic data, preliminary reports, and a wide range of scientific documents on USGS investigations. Open-File Reports are designed for fast release and are available for public consultation at depositories.

Maps

Geologic Quadrangle Maps (GQ's) are multicolor geologic maps on topographic bases in 7.5- or 15-minute quadrangle formats (scales mainly 1:24,000 or 1:62,500) showing bedrock, surficial, or engineering geology. Maps generally include brief texts; some maps include structure and columnar sections only.

Geophysical Investigations Maps (GP's) are on topographic or planimetric bases at various scales. They show results of geophysical investigations using gravity, magnetic, seismic, or radioactivity surveys, which provide data on subsurface structures that are of economic or geologic significance.

Miscellaneous Investigations Series Maps or **Geologic Investigations Series (I's)** are on planimetric or topographic bases at various scales; they present a wide variety of format and subject matter. The series also incudes 7.5-minute quadrangle photogeologic maps on planimetric bases and planetary maps.

Information Periodicals

Metal Industry Indicators (MII's) is a free monthly newsletter that analyzes and forecasts the economic health of five metal industries with composite leading and coincident indexes: primary metals, steel, copper, primary and secondary aluminum, and aluminum mill products.

Mineral Industry Surveys (MIS's) are free periodic statistical and economic reports designed to provide timely statistical data on production, distribution, stocks, and consumption of significant mineral commodities. The surveys are issued monthly, quarterly, annually, or at other regular intervals, depending on the need for current data. The MIS's are published by commodity as well as by State. A series of international MIS's is also available.

Published on an annual basis, **Mineral Commodity Summaries** is the earliest Government publication to furnish estimates covering nonfuel mineral industry data. Data sheets contain information on the domestic industry structure, Government programs, tariffs, and 5-year salient statistics for more than 90 individual minerals and materials.

The Minerals Yearbook discusses the performance of the worldwide minerals and materials industry during a calendar year, and it provides background information to assist in interpreting that performance. The Minerals Yearbook consists of three volumes. Volume I, Metals and Minerals, contains chapters about virtually all metallic and industrial mineral commodities important to the U.S. economy. Volume II, Area Reports: Domestic, contains a chapter on the minerals industry of each of the 50 States and Puerto Rico and the Administered Islands. Volume III, Area Reports: International, is published as four separate reports. These reports collectively contain the latest available mineral data on more than 190 foreign countries and discuss the importance of minerals to the economies of these nations and the United States.

Permanent Catalogs

"Publications of the U.S. Geological Survey, 1879–1961" and **"Publications of the U.S. Geological Survey, 1962–1970"** are available in paperback book form and as a set of microfiche.

"Publications of the U.S. Geological Survey, 1971–1981" is available in paperback book form (two volumes, publications listing and index) and as a set of microfiche.

Annual supplements for 1982, 1983, 1984, 1985, 1986, and subsequent years are available in paperback book form.