STATE OF LOUISIANA
DEPARTMENT OF TRANSPORTATION AND DEVELOPMENT

WATER RESOURCES
TECHNICAL REPORT
No. 49

GEOHYDROLOGY AND SIMULATION OF GROUND-WATER FLOW IN THE "400-FOOT," "600-FOOT," AND ADJACENT AQUIFERS, BATON ROUGE AREA, LOUISIANA

Prepared by
DEPARTMENT OF THE INTERIOR
U.S. GEOLOGICAL SURVEY
In cooperation with
LOUISIANA DEPARTMENT OF TRANSPORTATION AND DEVELOPMENT

1989

STATE OF LOUISIANA

DEPARTMENT OF TRANSPORTATION AND DEVELOPMENT

and

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By

Eve L. Kuniansky

U.S. Geological Survey

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For the convenience of readers who prefer to use metric (International System) units rather than the inch-pound units used in this report, values may be converted by using the following factors:

Multiply inch-pound unit	By	To obtain metric unit
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
foot per day (ft/d)	0.3048	meter per day (m/d)
foot per mile (ft/mi)	0.1894	meter per kilometer (m/km)
foot per year (ft/yr)	0.3048	meter per year (m/yr)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
cubic foot per day (ft^3/d)	0.02832	cubic meter per day (m^3/d)
square foot per day (ft^2/d)	0.09290	square meter per day (m^2/d)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.59	square kilometer (km^2)
gallon (gal)	0.003785	cubic meter (m^3)
gallon per day per foot [(gal/d)/ft]	0.01242	square meter per day (m^2/d)
gallon per minute (gal/min)	0.06308	cubic meter per second (m^3/s)
million gallons per day (Mgal/d)	3,785	cubic meter per day (m^3/d)

Temperature in degrees Celsius (°C) can be converted to degrees Fahrenheit (°F) as follows: °F = 1.8 X °C + 32.

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

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GEOHYDROLOGY AND SIMULATION OF GROUND-WATER FLOW IN THE "400-FOOT," "600-FOOT," AND ADJACENT AQUIFERS, BATON ROUGE AREA, LOUISIANA

by Eve L. Kuniansky

ABSTRACT

The "400-foot" and "600-foot" aquifers at Baton Rouge, Louisiana, provide large quantities of water for both industrial and public supply. Pumpage from these aquifers by industrial users ranged from 22 to 36 million gallons per day from 1940 to 1960 and caused water-level declines of as much as 190 feet in the industrial district. Declines of artesian head resulted in subsidence of more than a foot in the industrial district between 1935 and 1965. Industrial use of water from these aquifers decreased to 14 million gallons per day in 1983 and compaction of the sediments has ceased.

Population growth southeast of the industrial area in recent years has resulted in new pumping centers in the "400-foot" and "600-foot" aquifers. Sources of water pumped from these aquifers are: recharge from the outcrop area, the Mississippi River through the Mississippi River alluvial aquifer, water stored in the confining beds and aquifer, and downward leakage (induced by pumpage) from the overlying deposits.

The effect of past and future ground-water development from the "400-foot" and "600-foot" aquifers was evaluated by using a three-dimensional ground-water flow model. The model was calibrated using water-levels from 1940 to 1948 and potentiometric maps for May and October 1984. Through calibration, it was determined that subsidence may have reduced the conductance of clay beds that underlie the industrial district. Simulations indicate that less than 1 inch per year of recharge in the outcrop area goes into the regional flow system. More than 95 percent of recharge in the outcrop area is discharged to streams, and there is no permanent change in storage in the aquifer in the outcrop area. Simulated flow to or from the Mississippi River through the Mississippi River alluvial aquifer is seasonal and dependent on river stage. Simulated flow into or out of the river to the aquifer was always less than 500 cubic feet per second for the reach of the river between St. Francisville and New Orleans, Louisiana. Simulated vertical flux to the "400-foot" aquifer varies from 1×10^{-5} to 1×10^{-3} cubic feet per day per unit area, depending on the hydraulic gradients. The "400-foot" and "600-foot" aquifers recover quickly from the effects of pumpage because of the proximity of the outcrop area and the Mississippi River. The Baton Rouge fault which is a barrier to ground-water flow in the deeper aquifers is a barrier to flow in the "400-foot" and "600-foot" aquifers west of the Amite

River. Simulation of future pumpage in southeastern Baton Rouge indicates declines of 40 to 60 feet from the simulated water levels of the undeveloped aquifers and about 20 feet from the simulated 1984 water levels.

INTRODUCTION

Pumpage from the "400-foot" and "600-foot" aquifers by industrial users ranged from 22 to 36 Mgal/d (million gallons per day) during 1940-60, resulting in water-level declines of as much as 190 ft (feet) in the industrial district. Because of ground-water conservation begun in the late 1970's and economic recession, industrial use of water from these aquifers decreased to 14 Mgal/d in 1983. Water levels have risen as much as 100 ft in the industrial district in response to this decrease in pumpage. While industrial use has decreased, public ground-water use has increased. Population growth southeast of the industrial area in recent years has resulted in new pumping centers in the "400-foot" and "600-foot" aquifers.

Sources of water pumped from the "400-foot" and "600-foot" aquifers are: recharge in their outcrop area, leakage of water from the Mississippi River through the Mississippi River alluvial aquifer, water stored in the confining beds and aquifer, and downward leakage (induced by pumpage) from the overlying near-surface deposits of Pleistocene and Holocene age. The induced recharge from downward leakage of water in the overlying near-surface deposits is of concern because there are several surface hazardous-waste sites in East Baton Rouge Parish. These sites are potential sources of contamination to the potable water in the "400-foot" and "600-foot" aquifers (Louisiana Department of Transportation and Development, 1984, p. 248-251).

To evaluate the effect of past and future ground-water development from the "400-foot" and "600-foot" aquifers, the U.S. Geological Survey, in cooperation with the Louisiana Department of Transportation and Development (DOTD), and the Capital Area Ground Water Conservation Commission, developed a three-dimensional ground-water flow model of the shallow aquifers in the five-parish study area shown in figure 1.

Purpose and Scope

This report summarizes the geohydrology of the deposits of the late Tertiary and Quaternary Periods at Baton Rouge, presents a geohydrologic framework of the "400-foot," "600-foot," and adjacent aquifers for digital modeling, and documents the results of a calibrated digital model.

Most of the aquifers at Baton Rouge are named for the depth from land surface to the bottom of each extensive sand and gravel unit in the industrial district (Meyer and Turcan, 1955, p. 12-13). The aquifers included in this study are the Mississippi River alluvial aquifer, shallow Pleistocene sands, "400-, 600-, 800-, and 1,000-foot" aquifers.

The digital model was used to quantify the sources of water flowing into the pumped areas of the "400-foot" and "600-foot" aquifers at Baton Rouge. The model was calibrated using hydrographs from observation wells from January

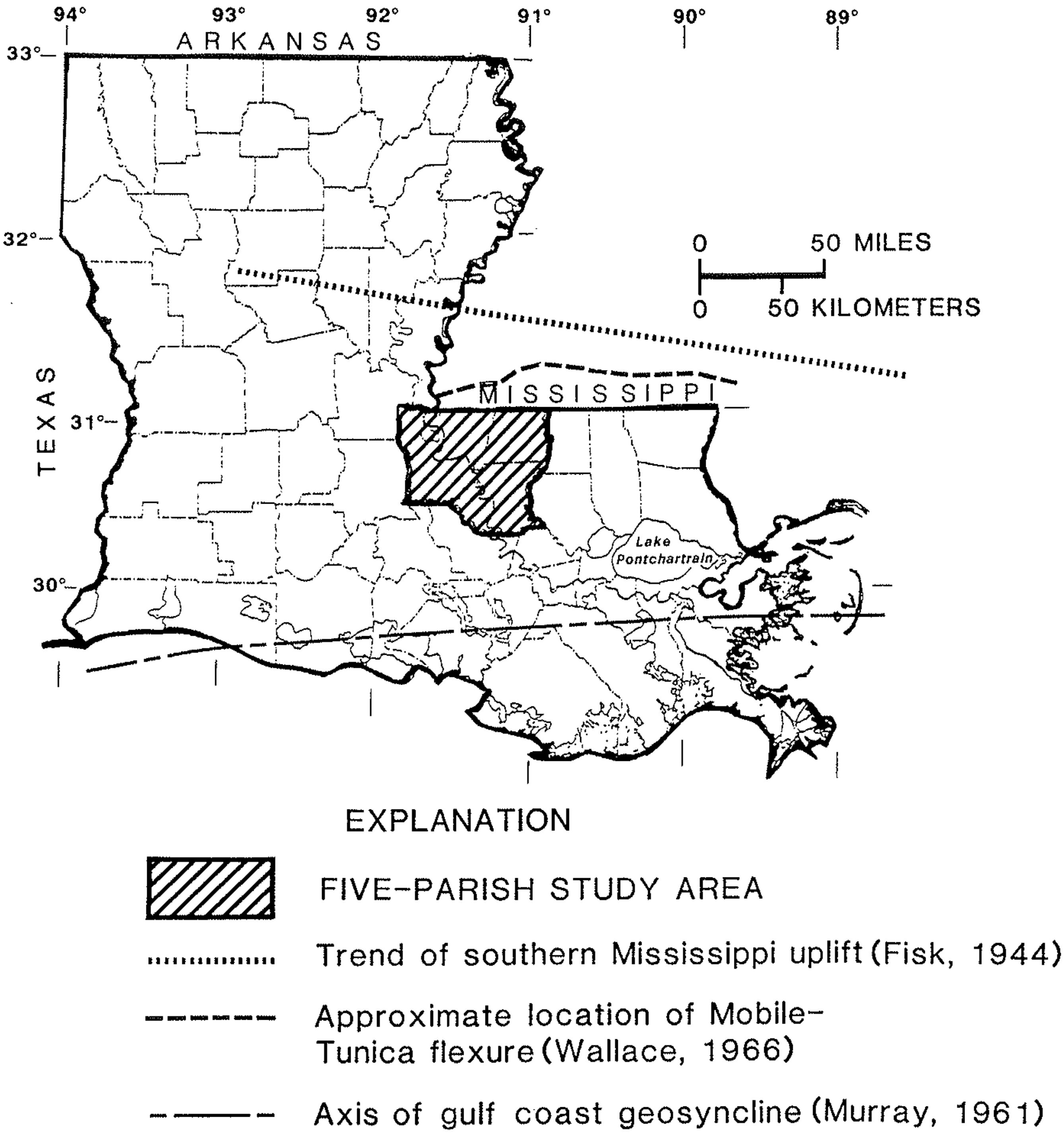


Figure 1.--Study area and location of structural features.

1940 to June 1948 and potentiometric maps of the "400-foot" and "600-foot" aquifers for May and October 1984. Sensitivity analyses performed on the calibrated model are presented graphically for each modeled aquifer to indicate how errors in model parameters affect computed water levels spatially, and a table is presented to show the parameters that have the greatest effect on water levels. Water-budget analyses from the calibrated model for past and future pumpage provide estimates of the quantity of water derived from recharge in the outcrop area, the Mississippi River, and downward leakage from near-surface deposits.

Acknowledgments

The assistance of John Overmeyer of Ethyl Corporation and Thomas Allain of Exxon Chemicals, U.S.A., in obtaining water levels for the May and October 1984 potentiometric maps is appreciated. Thanks are also due John McLaughlin, the Assistant State Climatologist of Louisiana, for providing climatological data.

PHYSIOGRAPHY AND GEOLOGIC SETTING

The study area is within the Gulf Coastal Plain. The unconsolidated clays, silts, sands, and gravels of Pleistocene and Pliocene age that occur within 1,000 ft of land surface at Baton Rouge are significant to this study. These shallow materials were deposited by fluvial, deltaic, and coastal processes that resulted in sand and gravel aquifers of irregular thickness and extent, interfingering with leaky confining beds of clay and silt.

The area encompasses two distinct physiographic features, the Mississippi River alluvial plain and the coastwise Pleistocene terraces. The Pleistocene terraces are slightly older alluvial deposits, higher in altitude, which were eroded by the Mississippi River. The eastern extent of the Mississippi River alluvial plain is shown in figure 2 by a solid line which parallels the remaining bluffs at the western edge of the Pleistocene terraces.

The terraces in East and West Feliciana Parishes have been eroded to rolling hills and decrease in altitude from approximately 300 ft in the north to about 100 ft above sea level in the south. The terraces in East Baton Rouge Parish are less hilly and decrease to 25 ft above sea level near the southern border of the parish.

The deposits which form the terraces dip southerly from the southern Mississippi uplift and thicken southerly toward the axis of the Gulf Coast geosyncline, forming a wedge of unconsolidated sediments. The dip increases to the south of the Mobile-Tunica flexure. The Mobile-Tunica flexure extends northwestward from Mobile Bay, Ala., to the Tunica Hills in the southwestern part of Mississippi near the Louisiana border (fig. 1). According to Fisk (1944, fig. 70), the dip of the base of the terrace deposits is 20 to 40 ft/mi (feet per mile) southward from the Mississippi-Louisiana border to Baton Rouge.

Figure 2.--Five-parish study area, Mississippi River alluvial plain, Pleistocene terrace deposits, and Baton Rouge fault.

In southern Mississippi, just north of West Feliciana Parish, Wallace (1966, pls. 1 and 2) located the Mobile-Tunica flexure (fig. 1) by mapping the contact of the sand and gravel deposits of Pleistocene age with the thick clay deposits of Pliocene age. The Mobile-Tunica flexure creates a topographic divide that controls both surface- and ground-water flow in the Pleistocene terrace deposits (Wallace, 1966, p. 19).

The Baton Rouge fault (fig. 2) is a growth fault, downthrown to the south, which has an impact on ground-water flow. The displacement increases from 30 ft at the surface to 225 ft at the top of the "400-foot" aquifer and to 350 ft at the top of the "2,000-foot" aquifer (Durham and Peeples, 1956, p. 65). The top of the "400-foot" aquifer south of the fault is adjacent to the bottom of the "600-foot" aquifer north of the fault at Baton Rouge. The displacement of the fault appears to decrease to the east. In Lake Pontchartrain, the displacement is only 200 ft at 2,000 ft below sea level (Cardwell and others, 1967, pl. 5).

The Mississippi River alluvial plain is essentially flat; it dips to the south at a lower rate than the Pleistocene terrace deposits. Land surface at the northern border of Pointe Coupee Parish is approximately 50 ft above sea level and is 25 ft above sea level in the southern part of West Baton Rouge Parish.

The Mississippi River alluvial aquifer is stratigraphically adjacent to the older Pleistocene and Pliocene deposits. At the northern border of the study area, the aquifer is in contact with deposits of Pliocene age. At the southern border of the study area, the Mississippi River alluvial aquifer is adjacent to the top of the "400-foot" aquifer and sediments above the "400-foot" aquifer of Pleistocene age. The generalized fence diagram of the sediments in the five-parish study area (fig. 3) shows the stratigraphic position of the Mississippi River alluvial aquifer to the deposits of Pliocene and Pleistocene age.

GEOHYDROLOGIC FRAMEWORK

Morgan (1963) developed a zonal concept for the freshwater-bearing aquifers of the terrace deposits in East and West Feliciana Parishes and correlated zones with the aquifers of East Baton Rouge Parish as named by Meyer and Turcan (1955). Figure 4, adapted from Morgan (1963), shows the general geohydrologic framework for the aquifers east of the Mississippi River in the study area and the generalized predevelopment flow. The "400-foot" and "600-foot" aquifers are in Morgan's Quaternary zone. The "1,000-, 1,200-, and 1,500-foot" aquifers comprise Morgan's zone 1. The "1,700-foot" and "2,000-foot" aquifers are included in his zone 2, and all the deeper freshwater-bearing sands of East Baton Rouge are in his zone 3. The "800-foot" aquifer actually is in the clayey zone which separates Morgan's Quaternary zone from his zone 1 as shown in figure 4.

For this study, the aquifers in East and West Baton Rouge Parishes and parts of the surrounding area were mapped as four layers: (1) the shallow Pleistocene sands and Mississippi River alluvial aquifer, (2) the "400-foot" aquifer, (3) the "600-foot" aquifer, and (4) the "800-foot" aquifer (Kuniansky and others, 1989).

The sands in the "400-, 600-, and 800-foot" aquifers tend to be interconnected in many places. The "800-foot" aquifer has some interconnection with the "1,000-foot" aquifer. Sands in the "400-foot" aquifer are the most continuous, and sands in the "800-foot" aquifer are the least continuous. Clayey surface deposits in the discharge area (fig. 4) may locally thicken and contain discontinuous sand and gravel lenses (the shallow Pleistocene sands) that reduce the confinement above the "400-foot" aquifer. The "800-foot" and "1,000-foot" aquifers are discontinuous and clays dominate the sediments deposited at this stratigraphic position; thus, these aquifers are part of a semiconfining zone between the "600-foot" aquifer and the zone 1 aquifers, where vertical leakage may occur through local interconnections between the aquifers.

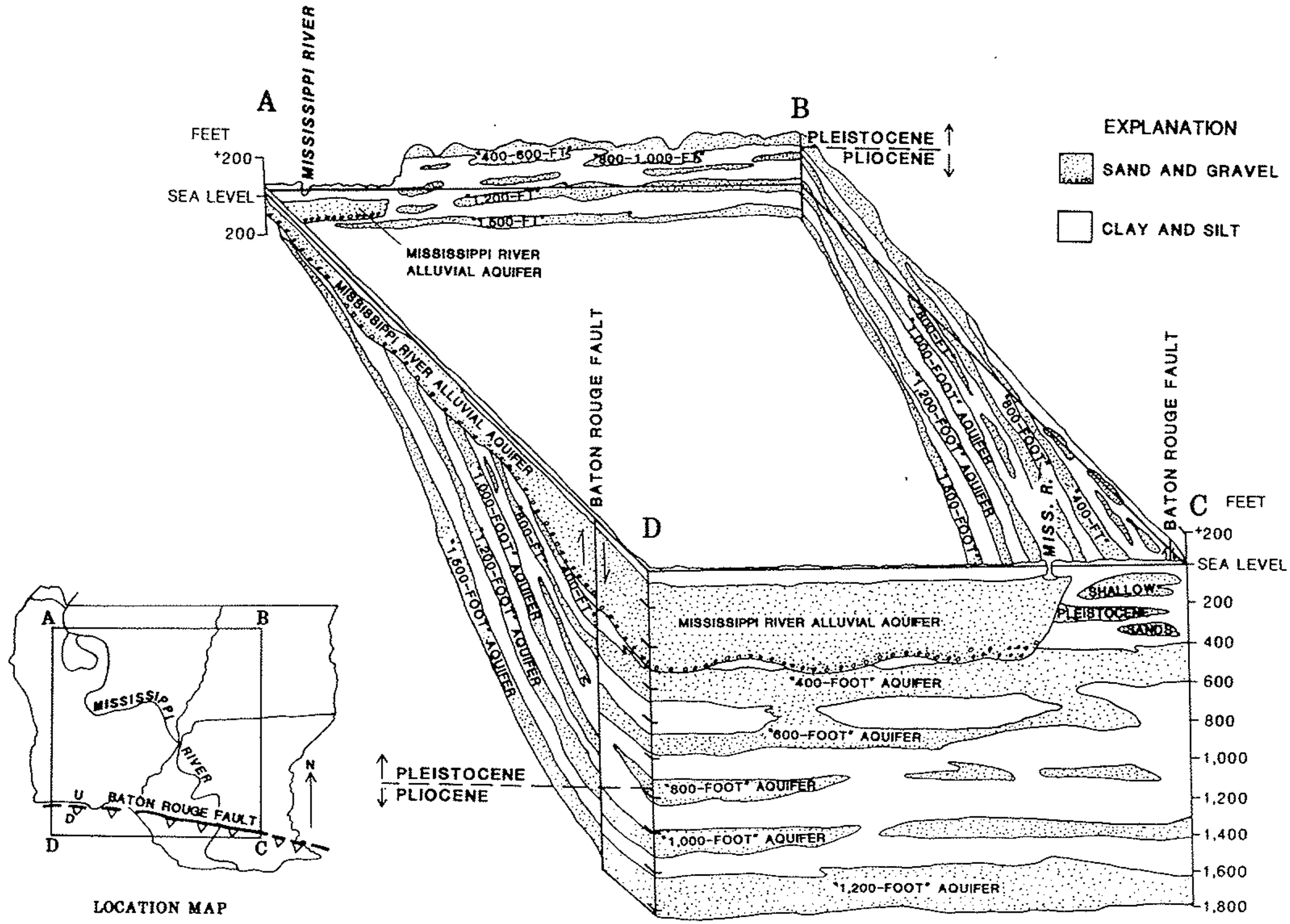


Figure 3.--Generalized fence diagram of the aquifers in the five-parish study area.

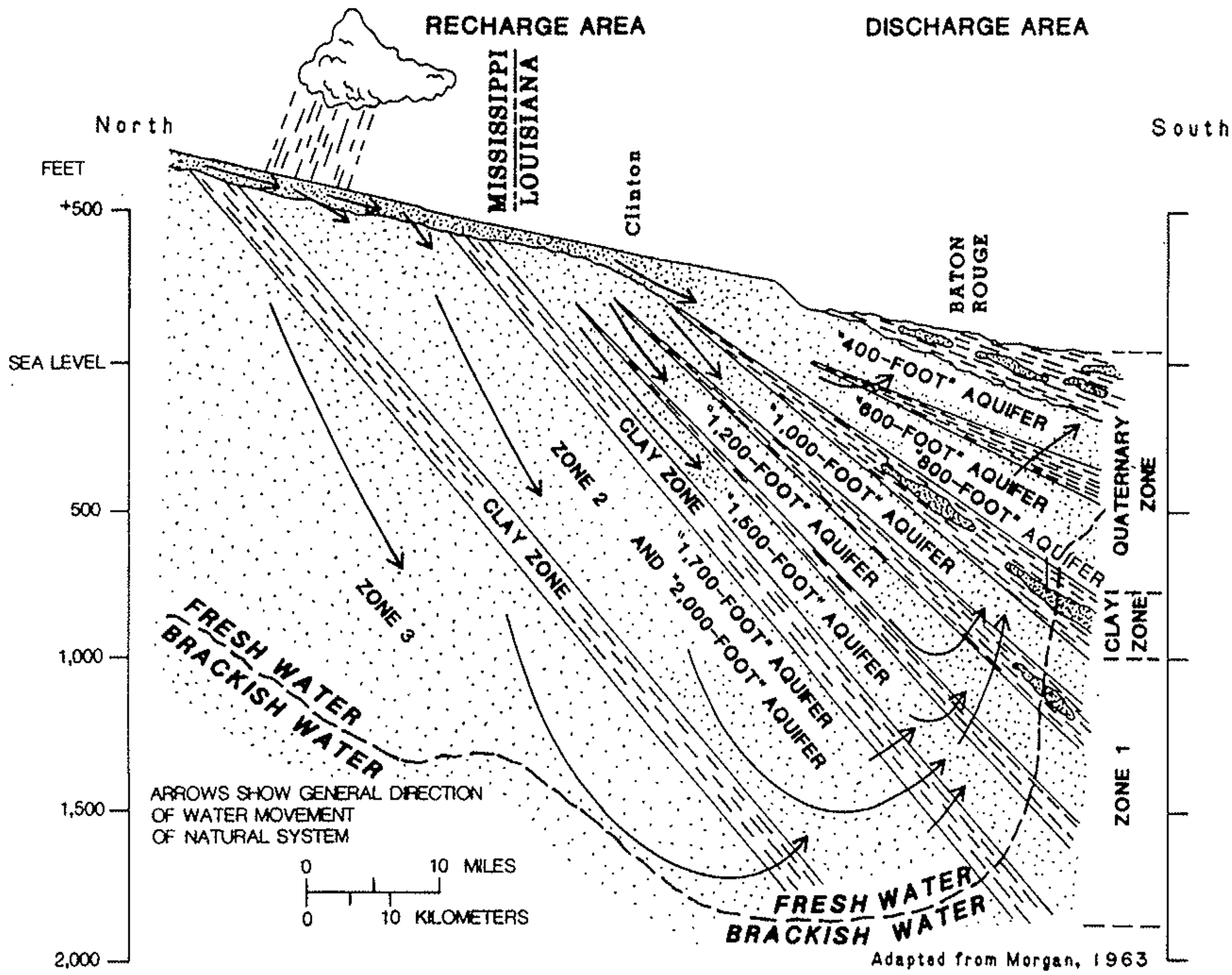


Figure 4.--Generalized section showing movement of ground water from the recharge area to the discharge area, southeastern Louisiana.

Direction of Ground-Water Movement

The outcrop area of the "400-foot" and "600-foot" aquifers in East and West Feliciana Parishes serves as a recharge area where rainfall can percolate into the aquifers (fig. 4). In this area, near-surface materials are predominantly sands and gravels.

The potentiometric contour map of the "400-foot" and "600-foot" aquifers developed by Morgan (1963, p. 27) indicates water-table conditions in the northern part of East and West Feliciana Parishes and confined aquifer conditions where the "400-foot" and "600-foot" aquifers are covered by surface deposits in the southern part of these parishes. Confined conditions are indicated by the decreasing percentage of base-flow contribution to average annual flow from the upstream gages to the downstream gages on both the Comite and Amite Rivers (table 1 and fig. 5). The general horizontal flow direction indicated on the map by Morgan (1963, p. 27) is in a north to south-southwest direction. Water levels in wells in the "400-foot" and "600-foot" aquifers tend to be at the same altitude in the northern part of East Baton Rouge Parish and in East and West Feliciana Parishes where the aquifers coalesce. Water levels in wells in the two aquifers differ by about 20 ft in the industrial district at Baton Rouge where they are separated by a clay layer and are pumped at different rates.

Table 1.--Continuous-record streamflow gaging stations, average annual flow, and estimated base flow

[mi^2, square miles; in., inches; ft^3/s, cubic foot per second; Q_{60}, discharge for 60 days]

Station name	Period of record	Drainage area (mi^2)	Average annual flow (in.)	Average annual flow (ft^3/s)	Estimated base-flow Q_{60} (ft^3/s)	Estimated base-flow runoff (in.)	Base-flow percentage of annual flow
Amite River near Darlington, La.	1951-84	580	21.6	922	356	8.33	39
Amite River near Denham Springs, La.	1939-84	1,280	21.4	2,019	692	7.33	34
Comite River near Olive Branch, La.	1943-84	145	21.8	233	67.6	6.33	29
Comite river near Comite, La.	1945-84	284	22.6	473	95.9	4.58	20
Tickfaw River near Liverpool, La.	1957-82	89.7	17.4	115	49.3	7.46	43

Water Budget and Recharge Estimate

Louisiana has a humid, subtropical climate. Average annual precipitation over the study area ranges from 54 to 60 in. (inches). Winter temperatures range from 4.4 to 18.3 °C. Summer temperatures range from 18.3 to 35.0 °C.

Even in this humid climate, there are periods of time when potential evapotranspiration, water loss that would occur if there were an unlimited amount of water available to the soil for use by vegetation and evaporation,

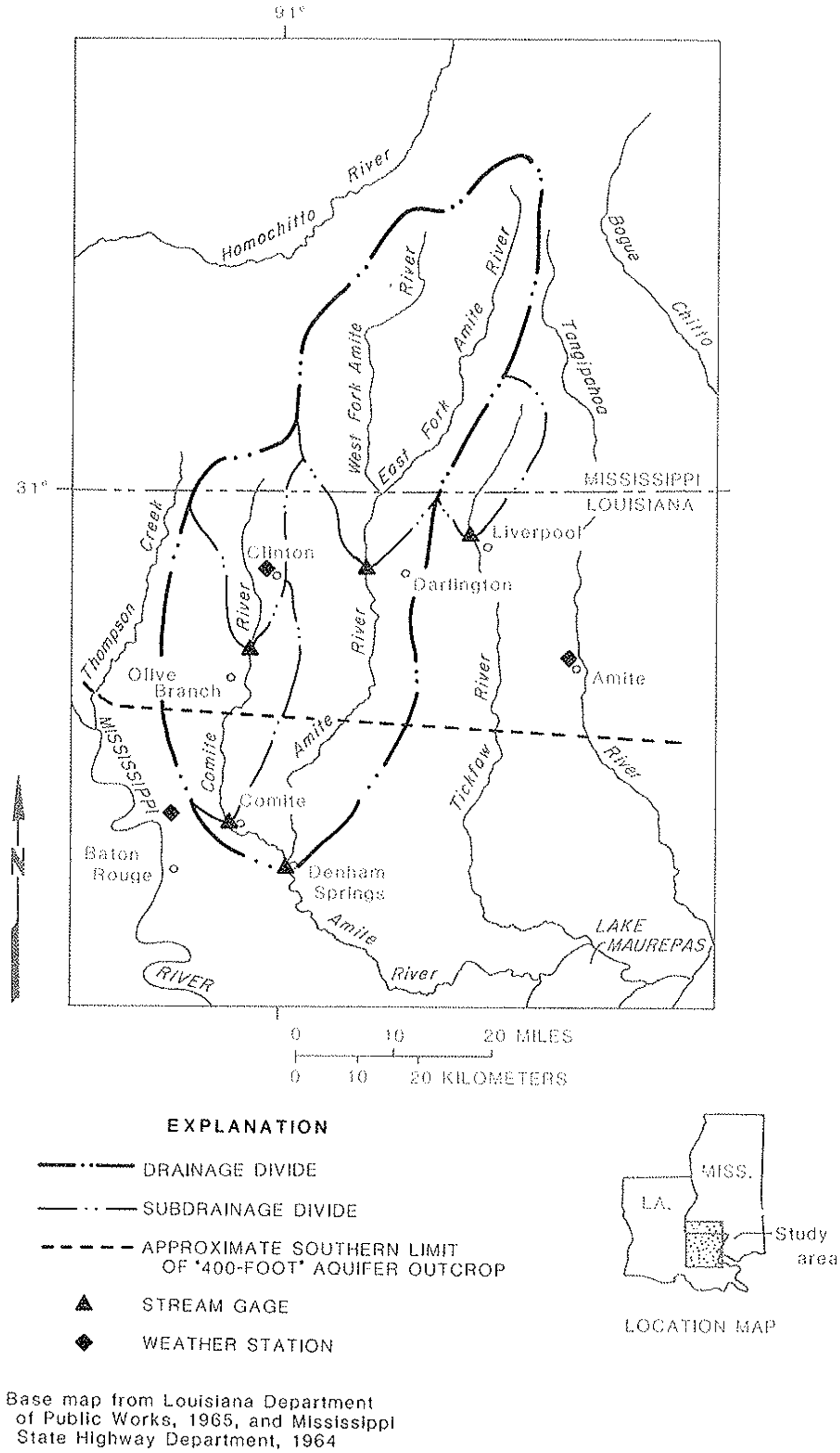


Figure 5.--Location of stream gages, weather stations, drainage divides, subdrainage divides, and southern limit of the "400-foot" aquifer outcrop.

exceeds precipitation. This results in a net deficit of water. The deficit reduces soil moisture which affects the rate of actual evapotranspiration. When rainfall exceeds potential evapotranspiration, there is a net surplus of water available for both surface runoff and ground-water recharge. Muller and Larimore (1975) prepared a seasonal water budget for Louisiana for 1941-70 using the Thornthwaite method. In Baton Rouge, it was determined that from December to April, a water-budget surplus was likely to occur; from June to October, a water-budget deficit was likely to occur. Table 2 lists monthly mean temperature, precipitation, water-budget deficit, and water-budget surplus for Baton Rouge. This information indicates that a water-budget surplus of about 19.3 in/yr (inches per year) may be available for surface-water runoff and ground-water recharge at Baton Rouge. Over the study area, the water-budget surplus ranges from 19 to 26 in/yr for the period of record, 1941-70 (Muller and Larimore, 1975, p. 19).

Many rivers, streams, and bayous drain the Pleistocene terrace deposits. Some of the streams that originate in the Pleistocene terrace deposits are the Comite, Amite, and Tickfaw Rivers (fig. 5). Both the Atchafalaya and Mississippi Rivers originate outside of the Pleistocene terrace deposits.

Table 2.--Mean temperature, precipitation, and water-budget data for Baton Rouge, Louisiana

Month	Mean temperature[1]		Precipitation[1]	Water-budget deficit[2]	Water-budget surplus[2]
	Fahrenheit	Celsius	(inches)		
January......	51.0	10.5	4.40	0.0	3.5
February.....	53.9	12.2	4.76	.1	3.9
March........	59.7	15.4	5.14	.0	3.6
April........	68.4	20.2	5.10	.1	2.4
May..........	74.8	23.8	4.39	.3	.8
June.........	80.3	26.8	3.77	1.1	.2
July.........	82.0	27.8	6.51	.7	.3
August.......	81.6	27.6	4.67	1.1	.1
September....	77.5	25.3	3.79	1.0	.2
October......	68.5	20.3	2.65	.7	.2
November.....	65.8	18.8	3.84	.0	1.2
December.....	52.9	11.6	5.03	.0	2.9
Annual.......	67.4	19.7	54.05	5.1	19.3

1 Data for Ryan Airport at Baton Rouge, La., for period of record 1950-81 from National Oceanic and Atmospheric Administration (1981).

2 Data for period of record 1941-70 from Muller and Larimore (1975).

Streamflow records were examined for five gaging stations on rivers originating in the terrace deposits (table 1 and fig. 5). The streamflow hydrographs for each station were examined for the application of hydrograph separation techniques to determine base flow. After inspecting the entire period of record for each stream, it was determined that the slopes of the recession limbs for each stream were inconsistent. This can be caused by the streams interacting with a nonhomogeneous aquifer or by several aquifers contributing to the base flow of the stream (Bingham, 1982). The 60-percent duration of flow from the flow-duration curve, in cubic feet per second (ft^3/s) (Q_{60}), was used as an estimate of long-term mean annual base flow rather than hydrograph separation (fig. 6). The flattening of the curves is characteristic of streams that have well-sustained base flow and Q_{60} occurs near the flexure point on these curves in the flattened part of the curve. Table 1 provides information on average flows and estimated base flows for the five stations analyzed. The base flow over the drainage area above each gage ranged from 4.58 to 8.33 in/yr with an average value of 6.81 in/yr.

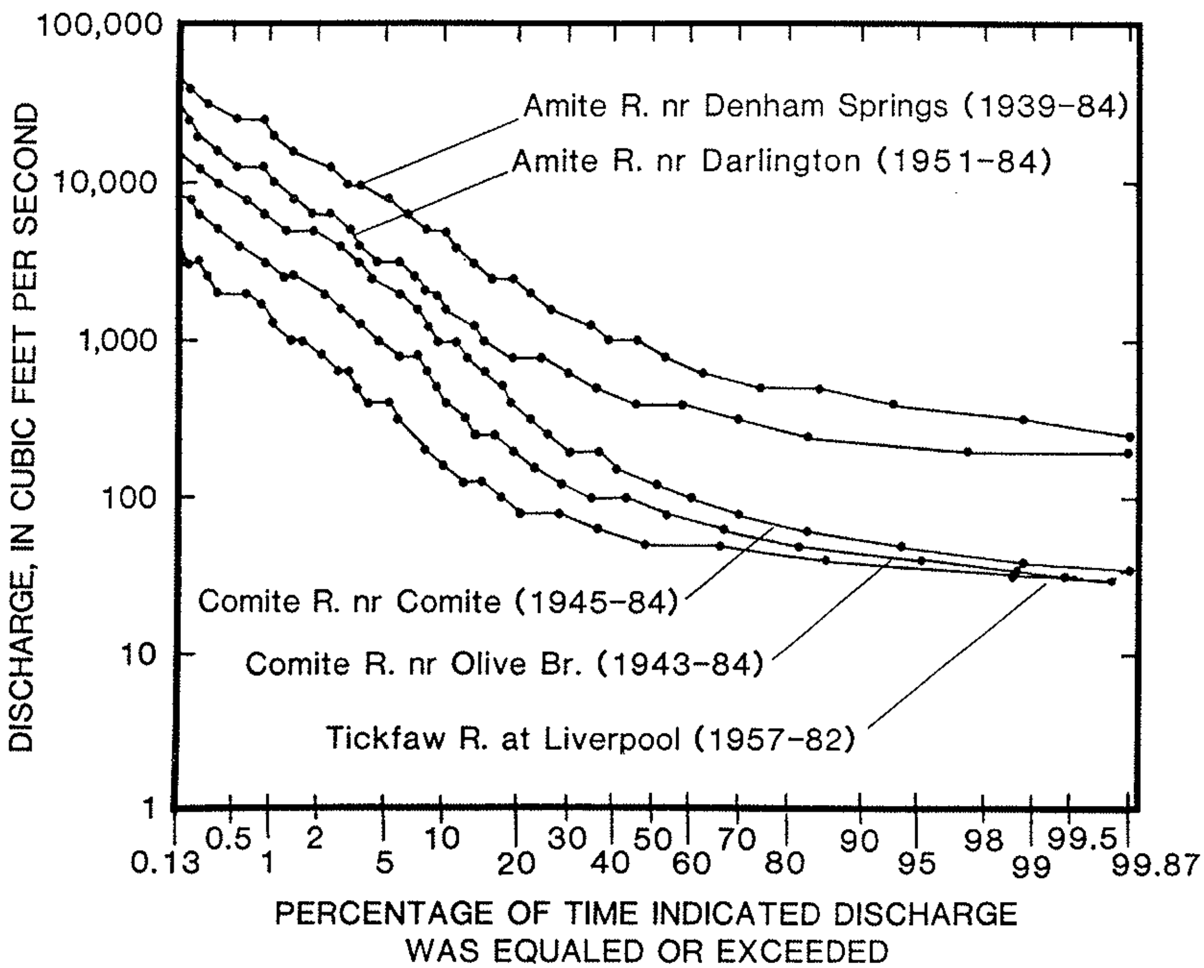


Figure 6.--Duration of daily flow at selected stations for three streams in the study area.

Annual regional aquifer recharge was estimated as the difference between mean annual water-budget surplus and mean annual streamflow for the same period of record. It is important to compare the same periods of record because of the large yearly variations in precipitation and temperature. Almost a 30-year period, 1951-79, was selected using the four gages on the Comite and Amite Rivers (fig. 5). The gage on the Tickfaw River was not used because of smaller drainage area and the shorter period of record.

The mean annual water-budget surplus from 1951 to 1979 at Baton Rouge, Clinton, and Amite, Louisiana, was 18.75, 24.84, and 26.80 in., respectively. These values were used to estimate the areal mean annual surplus for each of the drainage areas shown in figure 5. The estimated areal surplus for the drainage area at each gage was 25.0 in/yr for the Amite River near Darlington and the Comite River near Olive Branch, 23.5 in/yr for the Comite River near Comite, and 24.0 in/yr for the Amite River near Denham Springs. Because potential evapotranspiration is underestimated by about 10 percent (Muller and Larimore, 1975, p. 2), the estimated water-budget surplus for each drainage area is assumed to be the high end of the range in water-budget surplus; the low end of the range is assumed to be the estimated value reduced by 10 percent.

The average annual streamflow derived from each basin from 1951 to 1979 was 20.66 in. at the Amite River near Darlington, 20.31 in. at the Comite River near Olive Branch, 20.74 in. at the Comite River near Comite, and 20.19 in. at the Amite River near Denham Springs. A range in streamflow for each area was computed by assuming the discharge error is plus or minus 5 percent.

The range in deep regional aquifer recharge or discharge was computed for each drainage area by subtracting the minimum streamflow from the maximum surplus for each area for the estimate of maximum recharge and subtracting the maximum streamflow from the minimum surplus for the estimate of minimum recharge. A positive value indicates recharge and a negative value indicates discharge from the deep regional aquifer. The ranges for deep regional recharge for each gaged area are 5.3 to 0.8 in/yr for the Amite River near Darlington, 5.7 to 1.2 in/yr for the Comite River near Olive Branch, 3.8 to -0.6 in/yr for the Comite River near Comite, and 4.6 to 0.2 in/yr for the Amite River near Denham Springs. Because three of the stream gages represent drainage areas which drain into the gaged area of the Amite River near Denham Springs, 4.6 to 0.2 in/yr may be a more representative range for the deep regional aquifer recharge.

The local ground-water flow is represented by the base-flow runoff which is included in the mean annual streamflow. Thus, to get the total annual recharge to the aquifer, mean annual base flow must be added to the deeper annual aquifer recharge. The estimate of mean annual base flow was 6.8 in. for the drainage areas. Thus, total ground-water recharge could range from 12.5 to 6.2 in/yr in the outcrop of the Baton Rouge aquifer system.

Hydrologic Boundaries of the "400-foot" and "600-foot" Aquifers

The major hydrologic boundaries for the "400-foot" and "600-foot" aquifers are the northern limit of the aquifers in southern Mississippi, the Baton

Rouge fault in the south, and the Mississippi River through the Mississippi River alluvial aquifer, which is hydraulically connected with the sands of the "400-foot" and "600-foot" aquifers.

The Baton Rouge fault is a significant barrier to ground-water movement in the "400-foot" and "600-foot" aquifers at Baton Rouge. The "400-foot" aquifer south of the fault is connected to the "600-foot" aquifer north of the fault. The differences between water levels across the fault in these aquifers indicate that the fault restricts flow from the south toward the cone of depression beneath the industrial district of Baton Rouge. Whiteman (1979, p. 12) determined that there is a northward component of flow in the "400-foot" aquifer south of the fault which indicates that some leakage does occur through the fault. Effects of the fault on ground-water flow between the "600-foot" and "800-foot" aquifers have not been studied.

East of East Baton Rouge Parish, the Baton Rouge fault may not be a barrier to ground-water flow in the "400-foot" and "600-foot" aquifers. A north-south geohydrologic section east of the Amite River indicates that the "400-foot" and "600-foot" aquifers are merged and thick sands and gravels of these aquifers are adjacent to each other across the fault (D.J. Tomaszewski, U.S. Geological Survey, oral commun., 1985).

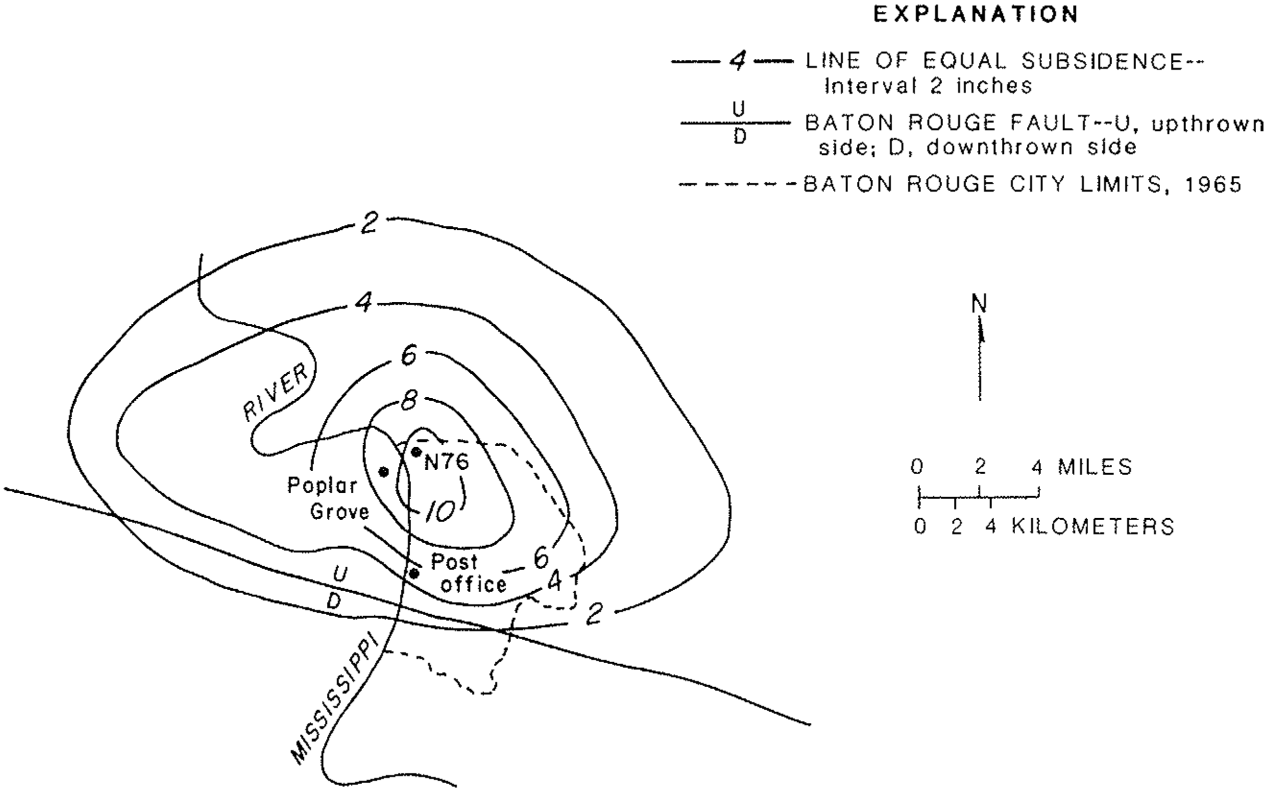
South of the Baton Rouge fault at Baton Rouge, the "400-foot" aquifer is hydraulically connected with the Mississippi River alluvial aquifer. Water levels in the "400-foot" aquifer fluctuate with the stage of the Mississippi River, as do the water levels in the shallow Pleistocene sands near the river.

North of the Baton Rouge fault in the industrial area, the "400-foot" and "600-foot" aquifers are not connected to the Mississippi River at the present river channel. The "400-foot" aquifer is hydraulically connected to the river through the Mississippi River alluvial aquifer west of the present river channel.

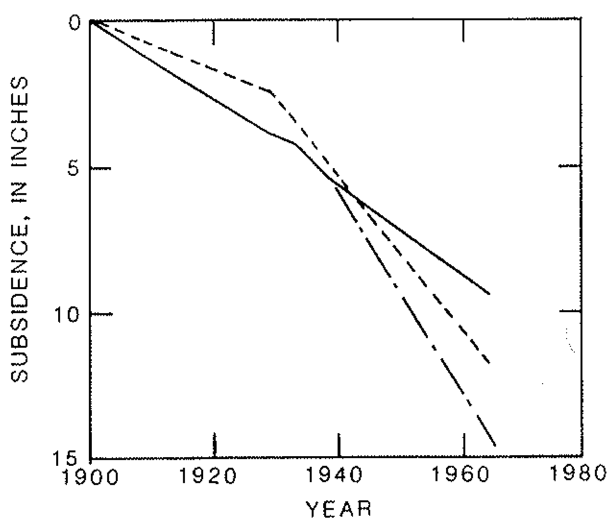
Land Subsidence Related to Ground-Water Withdrawals

Land subsidence at Baton Rouge, caused by the draining of water from the finer sediments (clay and silts) when pumpage of the aquifers lowered the artesian head significantly, was first documented by Davis and Rollo (1969) (fig. 7). A bowl-shaped 250 mi^2 (square miles) area of subsidence greater than 2 in. (fig. 7a) centered at the drawdown cone at Baton Rouge occurred between 1934-65. There was some subsidence prior to 1934 as can be seen from the plot of subsidence versus time at three benchmarks (fig. 7b). A maximum subsidence of 1.3 ft occurred at benchmark N76 (fig. 7b) between 1900-64. Wintz and others (1970, p. 12) attributed 0.9 ft of subsidence at benchmark N76 to ground-water withdrawals for the period 1938-64. Smith and Kazmann (1978, p. 10) reported additional subsidence of 0.42 ft at benchmark N76 for the period 1964-76.

The multiaquifer nature of the water-bearing sediments at Baton Rouge make it difficult to determine which clay confining beds compacted and when the compaction took place. Water-level altitudes between 1900-60 are shown in



a. Lines of equal subsidence at Baton Rouge, 1934-65.



b. Subsidence as a function of time, Baton Rouge, 1900-64.

Figure 7.--Subsidence at Baton Rouge, 1900-65 (from Davis and Rollo, 1969, p. 175-176).

figure 8 for the "400-, 600-, 800-, 1,500-, and 2,000-foot" aquifers. These water levels are composites from several wells in the industrial district at Baton Rouge. Figure 9 shows the history of ground-water development at Baton Rouge as compiled by Morgan and Rollo, 1890-1966, and the Capital Area Ground Water Conservation Commission, 1975-84. The decline in water levels (fig. 8) corresponds to the measured pumpage from the different aquifers (fig. 9).

Extensometers were placed near benchmark N76 and monitored from 1975-79 (Whiteman, 1980, p. 11, and pl. 2) to examine the compaction and rebound of three intervals of sediments. Between land surface and 833 ft below land surface no permanent compaction was observed from 1975 to 1979. Between 833 and 1,700 ft below land surface about 0.02 ft of permanent compaction occurred. The deepest interval 1,700-2,997 ft below land surface about 0.03 ft of permanent compaction occurred. All wells show about 0.1 ft of elastic compaction and rebound.

Examination of the records of the extensometers for 1981-82 indicates no permanent compaction or rebound for sediments from land surface to 833 ft below land surface, which indicates that these sediments were fully compacted from the applied pumping stress prior to 1975. The deeper sediments are still compacting but at a very slow rate. The past studies of subsidence indicate that the majority of subsidence caused by ground-water withdrawals occurred prior to 1964 in the industrial district.

The compaction of the confining beds could significantly reduce the porosity, void ratio, and hydraulic conductivity of the confining beds, resulting in less vertical movement of water between aquifers at the areas of maximum compaction. After compaction of confining beds, the same drawdowns can be attained with less pumpage from the "400-foot" and "600-foot" aquifers, resulting from the reduction in vertical leakage from overlying sediments. Maximum drawdowns occurred in the "400-foot" and "600-foot" aquifers between 1953-56 although maximum yearly pumpage and peak monthly pumpage never exceeded previous pumping rates (fig. 10).

Hydraulic Properties of the Aquifers and Confining Beds

The hydraulic properties of nonindurated, water-deposited materials are related to their grain-size distribution and percent silt and clay-size particles. Table 3 provides ranges of hydraulic conductivity for types of materials found in the study area. The variability in hydraulic conductivity of similar sized material is due to "differences in packing of the particles or differences in the amount of clay binding between particles" (Davis, 1969, p. 78). Bedding of the deposits can create anisotropy in hydraulic conductivity. Horizontal hydraulic conductivity is generally 2 to 10 times greater than vertical hydraulic conductivities in alluvial deposits (Davis, 1969, p. 32).

Transmissivity and storage coefficient are two hydraulic properties which can determined through aquifer tests, although transmissivity can be estimated by multiplying aquifer thickness times hydraulic conductivity of the aquifer material. Table 4 summarizes 23 aquifer tests which were conducted in the industrial district, approximately 2 mi north of downtown Baton Rouge.

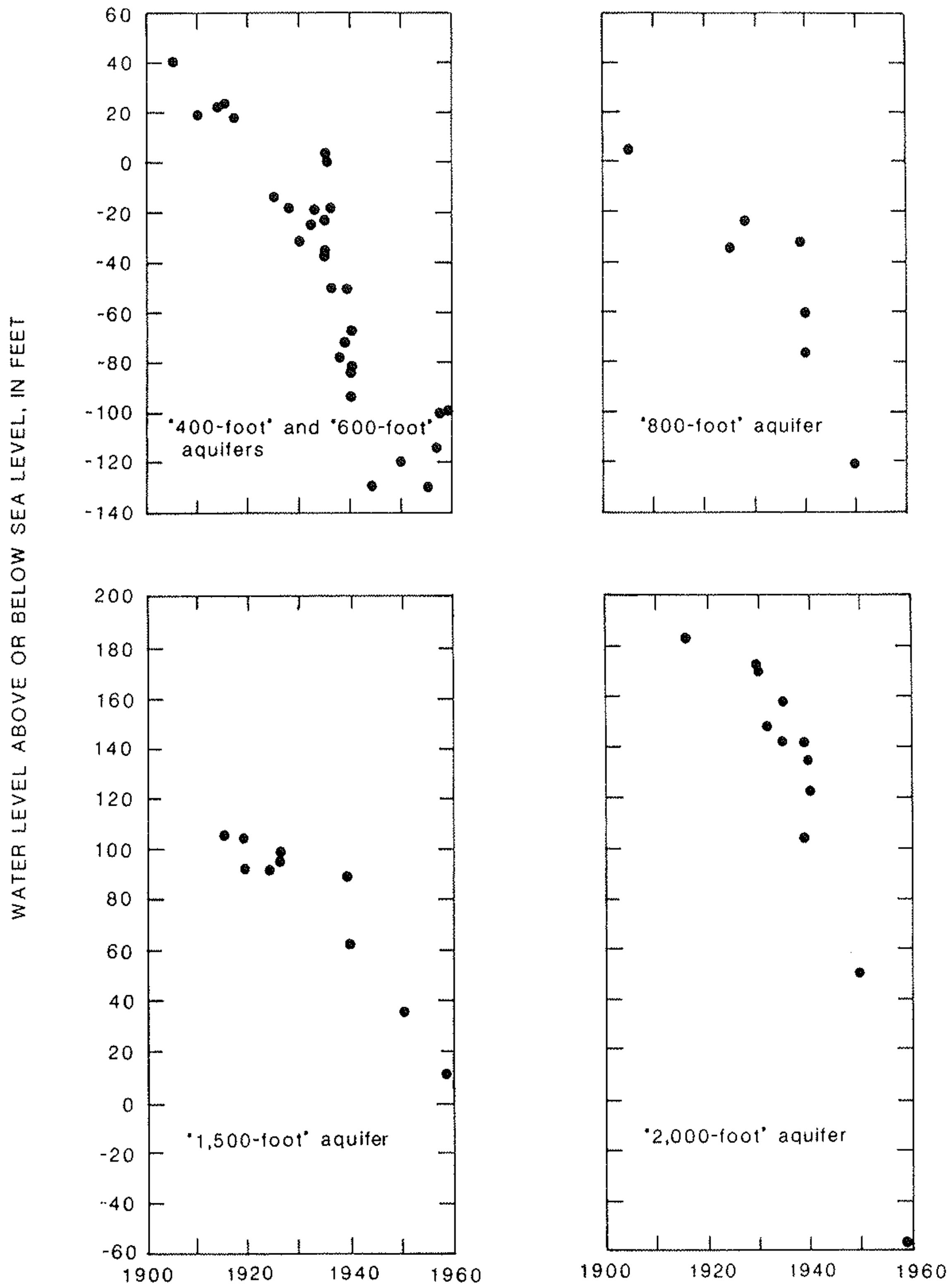


Figure 8.--Water-level declines from 1905 to 1960 at the industrial district, Baton Rouge.

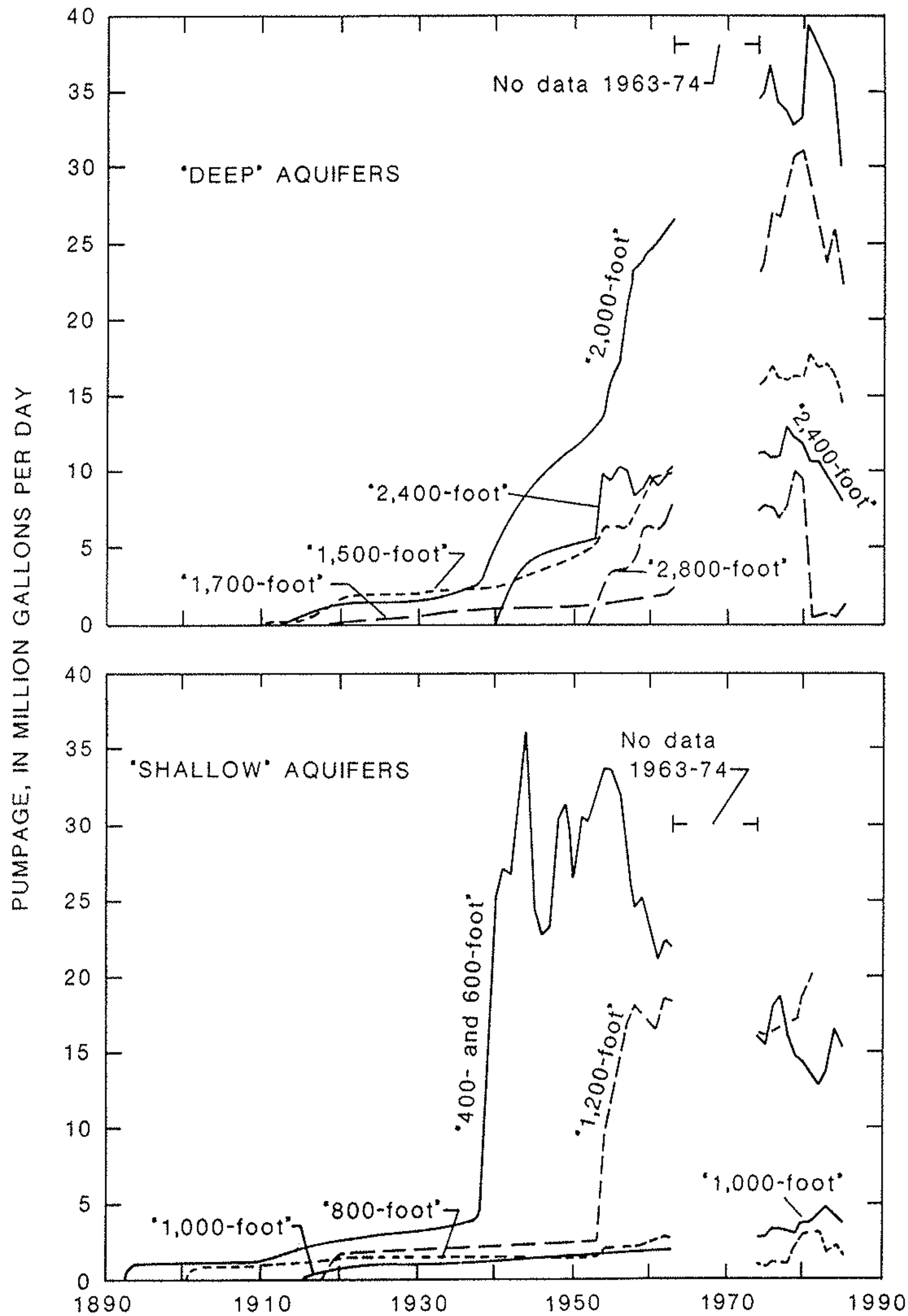


Figure 9.--Pumpage from aquifers in the Baton Rouge area, 1890-1985.

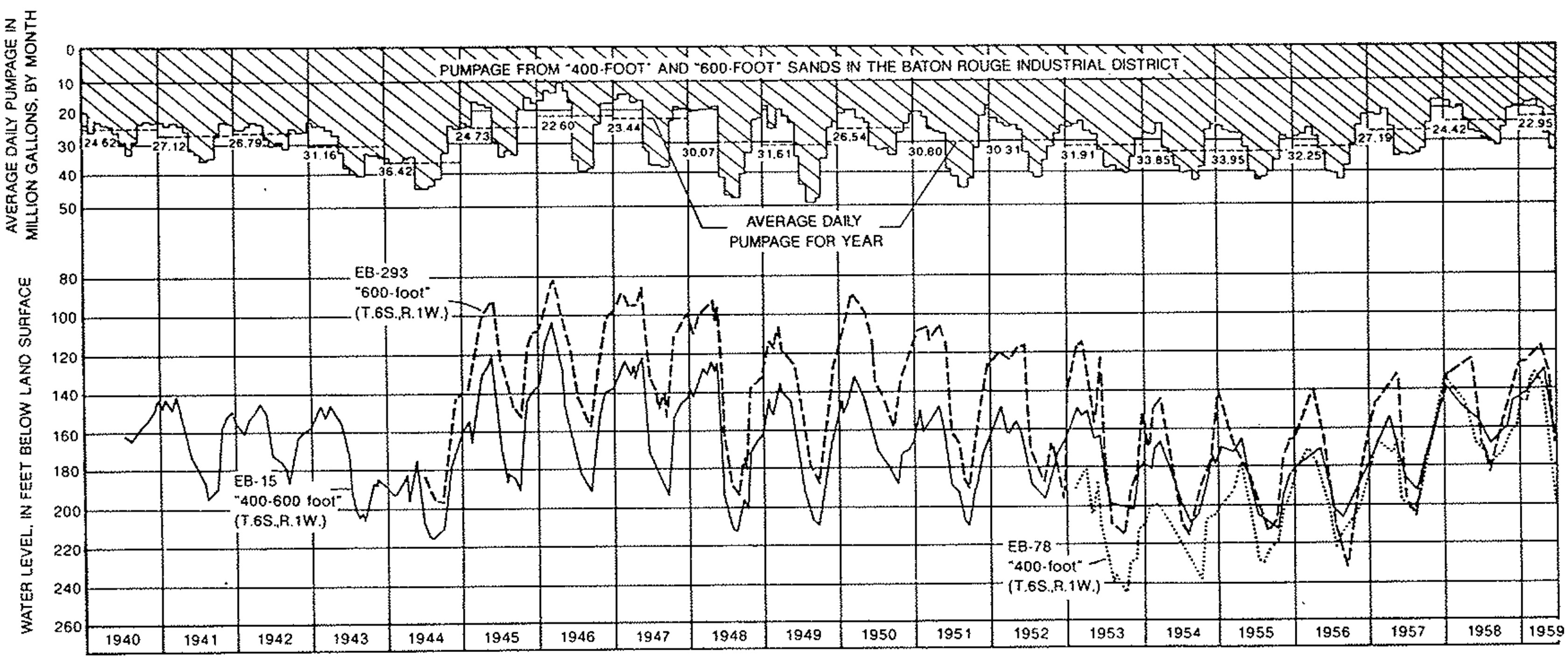


Figure 10.--Relation of pumpage to water levels in wells screened in the "400-foot" and "600-foot" aquifers in the Baton Rouge industrial district. (From Morgan, 1961, fig. 11.)

Table 3.--Range of hydraulic conductivity for gravel, sand, silt, and clay

[All values in feet per day]

Clay soils (surface)[1]	0.3	-	0.7
Deep clay beds[1]	3×10^{-8}	-	3×10^{-2}
Fine sand[1]	3	-	20
Medium sand[1]	20	-	70
Coarse sand[1]	70	-	300
Gravel[1]	300	-	3,000
Sand and gravel mixes[1]	20	-	300
Silt[2]	.0006	-	.9

[1] Bouwer, 1978, p. 38.
[2] Davis, 1969, p. 80-81.

Table 4.--Summary of hydraulic properties from aquifer-test data for Pleistocene aquifers at Baton Rouge, Louisiana

[Meyer and Turcan, 1955, p. 96-97]

	No. of aquifer tests	Transmissivity (feet squared per day)			Hydraulic conductivity (feet per day)			Storage coefficient		
		minimum	maximum	average	minimum	maximum	average	minimum	maximum	average
Mississippi River alluvial aquifer.	6	18,700	28,100	22,933	150	374	218	0.0009	0.02	0.007
"400-foot" aquifer..	11	4,280	10,300	6,747	32	71	48	.00026	.00097	[a].00042
"600-foot" aquifer..	5	12,700	16,400	14,660	74	107	90	.00034	.00057	[b].00047
"800-foot" aquifer..	1	------	------	3,210	---	---	36	-------	-------	-------

[a] Only 9 values were used to average storage coefficient.

[b] Only 4 values were used to average storage coefficient.

Analyses of two clay core samples collected in the industrial district in 1974 indicate that the confining beds at Baton Rouge are composed of mixed layer clay minerals (36-39 percent by weight), quartz (35-38 percent by weight), and montmorillonite (12-20 percent by weight). Vertical hydraulic conductivities of the cores were 1×10^{-5} and 1×10^{-6} ft/d, and specific storage of the cores were 1×10^{-5} ft^{-1} (Whiteman, 1980, p. 16-18).

The Mississippi River alluvial aquifer grades from coarse gravels at the bottom to fine silts at the top and ranges from 200 to 400 ft in thickness. It is confined or semiconfined at the top by a clayey natural levee and backswamp deposits. The clay separating the Mississippi River alluvial aquifer from the older Pleistocene aquifers is generally less than 50 ft in thickness, and in many places there is no clay separation.

The "400-, 600-, and 800-foot" aquifers south of the outcrop area are composed of fine- to medium-grained sands or medium- to coarse-grained sands (Morgan, 1961, p. 18-29). The outcrop area of the "400-foot" and "600-foot" aquifers is composed of medium- to coarse-grained sands with some gravels and silts (Morgan, 1963, p. 14). Clay layers that separate the aquifers range from 0 to 300 ft in thickness.

The order of magnitude for horizontal hydraulic conductivity of aquifers range from 10 to 100 ft/d and for vertical hydraulic conductivity of confining beds range from 1 X 10^{-2} to 1 X 10^{-7} ft/d. Values of storage coefficient range from 0.01 to 0.0001 from the confined aquifers in the study area.

Transient Leakage

Transient leakage of water to or from the confining beds is not instantaneous. The time, for completion of virtually all transient changes in water stored in the confining beds when a drop or rise in aquifer head occurs, can be estimated from the equation:

$$t = \frac{0.5\ S_s B^2}{K_v} \qquad (1)$$

where t is time, S_s is specific storativity, B is confining bed thickness, and K_v is the vertical hydraulic conductance of the confining bed (Bear, 1979, figs. 5-12a). After this time has elapsed, steady flow as computed in the current model occurs and no changes in storage occur in the confining bed. When t is small the effects of transient leakage are negligible and steady leakage occurs quickly.

The volume of water which can come from storage in the confining bed can be significant if water-level declines are large, specific storativity of the confining bed is large, the confining bed is thick, and the decline occurs for a period of time greater than t. Between 1937 and 1940 significant water-level declines occurred beneath the industrial district at Baton Rouge (fig. 8). The cone of depression that developed persisted through 1960.

The volume of water which would come out of storage from the confining beds in the cone of depression was estimated by multiplying the total volume of clay within a contour of the potentiometric surface of October 1966 (Rollo, 1969) by the mean water-level decline by the specific storage for clay and summing these together. Using the specific storage value of 1 X 10^{-5} ft^{-1} (Whiteman, 1980), the computed volume of water from storage was 13,346 Mgal (million gallons) or 10 percent of the total pumpage from the "400-, 600-, and 800-foot" aquifers. If the specific storage for clay were 1 X 10^{-4} ft^{-1}, then 100 percent of the volume pumped from the "400-, 600-, and 800-foot" aquifers could come from storage. Using equation 1, the time it would take for this transient leakage to take place for a 100 ft thick clay lens with a specific storage of 1 X 10^{-5} ft^{-1} would be 50 days for K_v = 1 X 10^{-3} ft/d or 14 years for K_v = 1 X 10^{-5} ft/d.

SIMULATION OF THE "400-FOOT," "600-FOOT," AND ADJACENT AQUIFERS

Five layers were used to simulate flow in the shallow aquifers at Baton Rouge. The U.S. Geological Survey finite-difference model (McDonald and Harbaugh, 1984) was used in a quasi three-dimensional manner in which aquifers are simulated as active layers with lateral flow separated by confining beds. Hydraulic head is computed for the active layers but not for the confining beds. Steady vertical leakage (no storage effects) is computed through confining beds by adding a head dependent source-sink term to the active model layer. The top four layers: (1) the Mississippi River alluvial aquifer and shallow Pleistocene sands, (2) the "400-foot" aquifer, (3) the "600-foot" aquifer, and (4) the "800-foot" aquifer are modeled as active layers. Layer 5, "1,000-foot" aquifer, is included as a specified-head layer in order to estimate upward or downward leakage between the "800-foot" and "1,000-foot" aquifers. Thus, layer 5 is a boundary condition for the "800-foot" aquifer and not an active model layer. All active layers were modeled as confined aquifers. This requires arrays of transmissivity (T) and storage coefficient (S) (for transient simulations) for each active layer. An array for vertical leakage coefficient (K_z/b, where K_z is vertical hydraulic conductivity and b is confining bed thickness) is required beneath all active layers to simulate vertical leakage.

Model Development

A three-dimensional model was necessary because of the complex geology and the difference in water-levels between the "400-foot" and "600-foot" aquifers at Baton Rouge. The U.S. Geological Survey modular finite-difference model was selected because the model (1) is well documented and tested, (2) allows for the deformation of the grid in cross-section (McDonald and Harbaugh, 1984, p. 55), (3) allows for specifying water levels beneath the bottom active model layer, and (4) is three dimensional.

Finite-Difference Model Description

The modular finite-difference code was developed with many different subprograms, called packages, for handling a variety of problems. The model packages used for this study are the basic package, the block-centered flow package, the river package, the recharge package, the well package, the general-head boundary package, and the strongly implicit procedure package. Both the basic and block-centered flow packages are required for any modeling problem. The strongly implicit procedure package is one of two mathematical algorithms currently (1986) available with the model for solving the large set of finite-difference equations. It is an iterative solution algorithm which tends to be numerically stable for most problems. All of the other packages mentioned above are part of the optional packages provided with the model (McDonald and Harbaugh, 1984).

The finite-difference formulation is block-centered in which a grid is developed by breaking the modeled plan area into a series of rectangles. Vertical deformation of the thickness of the aquifer can be accounted for by using the average thickness of the aquifer for each rectangle. The head computed by the block-centered formulation is the head of the centroid of each rectangular block or cell. The equation for flow at each cell is derived from the continuity equation in which "the sum of all flows into and out of the cell must be equal to the rate of change in storage within the cell" (McDonald and Harbaugh, 1984, p. 12). The well and recharge packages add or subtract flows at specified model cells. The river and general-head boundary packages add or subtract flows to a cell dependent on the specified river stage or head, riverbed conductance or general-head boundary conductance; and the computed model cell head. Because leakage between layers, river leakage, and the general-head boundary leakage (used in this study as leakage between two layers) are head dependent source-sink terms, an iterative solution process is required.

A limitation of the finite-difference model is the way in which confining beds can be modeled. To model a confining bed accurately, many thin active model layers would be required for each confining bed in order to properly account for changes in storage of the confining bed. The amount of computation for a multiaquifer system is not feasible for this study. Thus, changes in storage of water in the confining beds are not simulated. Only steady flow through the confining beds based on Darcy's law is computed at each time step.

Finite-Difference Grid, Model Boundaries, and Layering Scheme

The modeled area extends beyond the study area on the west, south, and east to include the Atchafalaya, Mississippi, and Tangipahoa Rivers, respectively, and on the north to include the area south of the Mobile-Tunica flexure (fig. 11).

The finite-difference grid developed for this study is variably spaced with the smallest cells at Baton Rouge, Louisiana (fig. 11). The smallest cells are 5,000 by 5,000 ft or 0.9 mi^2. The largest cell, row 27 column 1, is 49,000 by 45,000 ft or 79 mi^2. The smaller cells are placed over the area where the heaviest pumping from the "400-foot" and "600-foot" aquifers occurs.

The model grid is oriented about 3 degrees east of true north. This orientation places the segment of the Mississippi River at Baton Rouge parallel with column 9 rows 12-20 and the Baton Rouge fault parallel with row 18 columns 1-14 (fig. 11).

For the top layer, the western boundary is the Atchafalaya River and the eastern boundary is the Tangipahoa River. Both are treated as constant-head boundaries with mean annual river-stage values computed for each cell. The northern boundary is the northern limit of the shallowest Pleistocene terrace

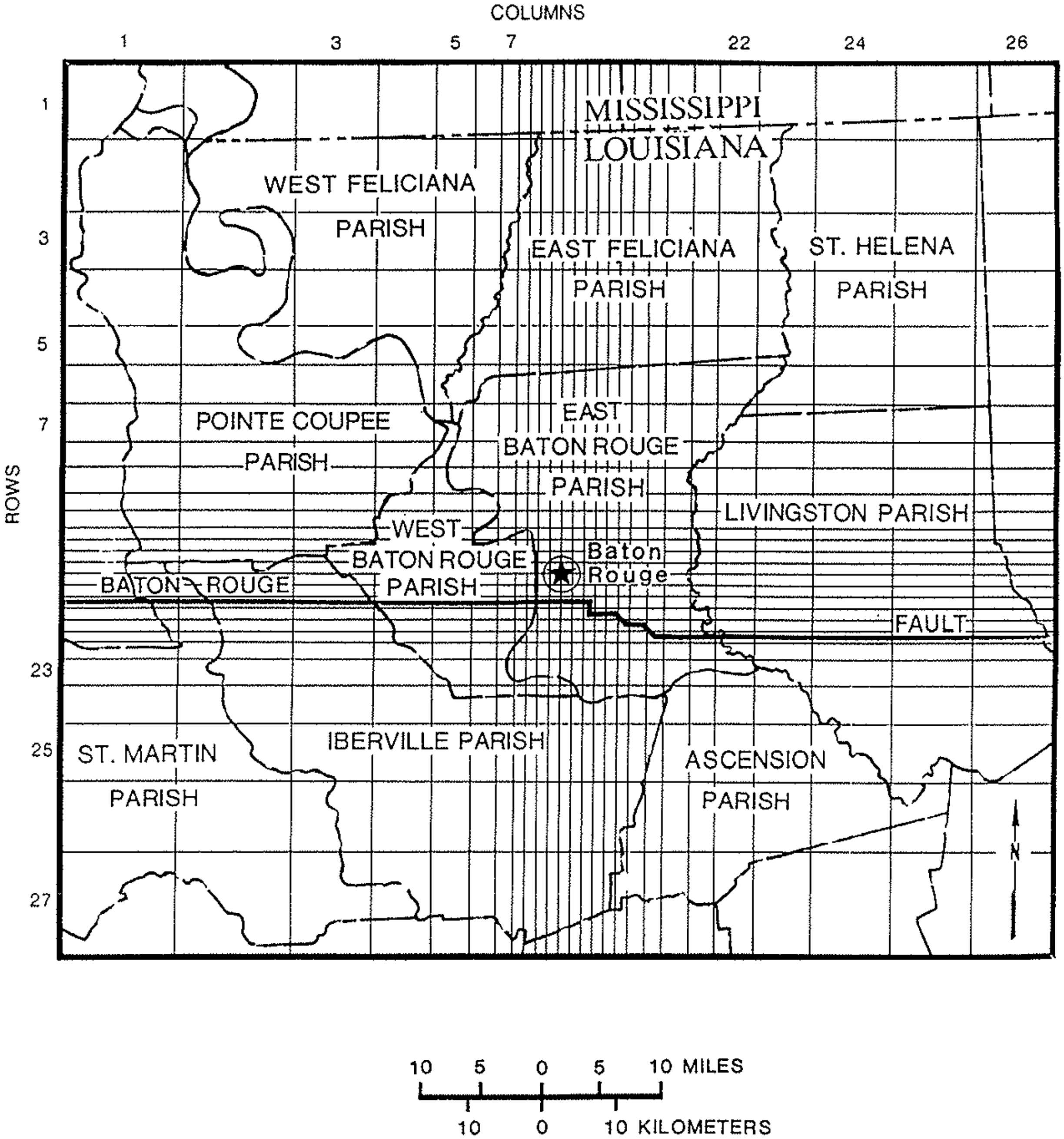


Figure 11.--Model grid and location of the Baton Rouge fault.

and the Mississippi River along the border of Pointe Coupee and West Feliciana Parishes. The Mississippi River also bounds the Pleistocene terrace deposits on the south but was input using the river package (McDonald and Harbaugh, 1984), in order to change river stage during transient simulations. Constant-head cells also were used in cells of the shallow Pleistocene aquifer because water-level data indicate that water levels remain close to land surface and river stage. These constant heads were estimated using surface topography and mean annual river stage. The Mississippi River was the only river simulated within the top layer. Only 248 model cells represent the Mississippi River alluvial aquifer and are actively modeled in layer 1.

The Mississippi River alluvial aquifer (layer 1) is hydraulically connected to lower aquifers over much of the western area. Also, in the southern area south of the fault, the Mississippi River alluvial aquifer and shallow Pleistocene aquifer were horizontally connected to the "400-foot" aquifer. (See figs. 12 and 13.)

The "400-foot" aquifer (layer 2) is interconnected with the "600-foot" aquifer over two-thirds of the area north of the Baton Rouge fault outside of Baton Rouge (fig. 13). South of the fault, these aquifers are not directly connected, but clay thickness between layers 2 and 3 is often less than 20 ft. A no-flow boundary was used at the northern border, and rivers were simulated in the outcrop area of layer 2, rows 1-5 (figs. 11 and 12). Water-level measurements made in 1984 indicated that the potentiometric surface in the outcrop area has remained the same since 1960 (Morgan, 1963), which justifies modeling layer 2 as a confined aquifer in the outcrop area (rows 1-5) with values of transmissivity and specific yield. The Atchafalaya and Tangipahoa Rivers simulated as constant-head boundaries in the top layer also serve as the west and east boundary of the deeper layers by creating a ground-water divide which can be modeled as a no-flow boundary. Recharge was distributed over the outcrop area, rows 1-5.

The "600-foot" aquifer (layer 3) has the same boundaries as the "400-foot" aquifer. The "600-foot" aquifer is more irregular than the "400-foot" aquifer, as can be seen by the number of cells in which the aquifer is thin or absent (fig. 12).

The "800-foot" aquifer (layer 4) also has the same boundaries as the "400-foot" aquifer. In the northern area (rows 1 and 2), the rivers are incised into Pliocene sediments. The "800-foot" aquifer is absent in many places (fig. 12). This aquifer also merges with "1,000-foot" aquifer in the northern part of the study area (fig. 13).

The 1,000-foot" aquifer (layer 5, not shown) was assigned constant heads during steady-state simulations. During transient simulations, potentiometric maps were developed for each stress period and the average water level for each cell was specified for computing vertical leakage to or from layer 4 with the general-head boundary package. For more information refer to McDonald and Harbaugh (1984, p. 343-346).

The geohydrologic complexity of the four modeled aquifers is simulated by adjusting the transmissivity and vertical leakage coefficient arrays for each active layer. Aquifer interconnections were incorporated in the model by adjusting the vertical leakage coefficients such that an equalization of water levels occurs between the interconnected active layers. An upper limit on the value of the vertical leakage coefficient at interconnections is the vertical hydraulic conductivity of the aquifer material divided by unit thickness (K_z/1 ft). Where the aquifer is thin or absent low transmissivity values were used. These values were of the order of magnitude of the hydraulic conductivity of mixed silt and clay multiplied by unit thickness (K_h X 1 ft). Figures 12 and 13 illustrate best how the model layering scheme incorporates the geohydrologic complexity of the Pleistocene and Pliocene aquifers. Figure 14 shows the relation among model layers.

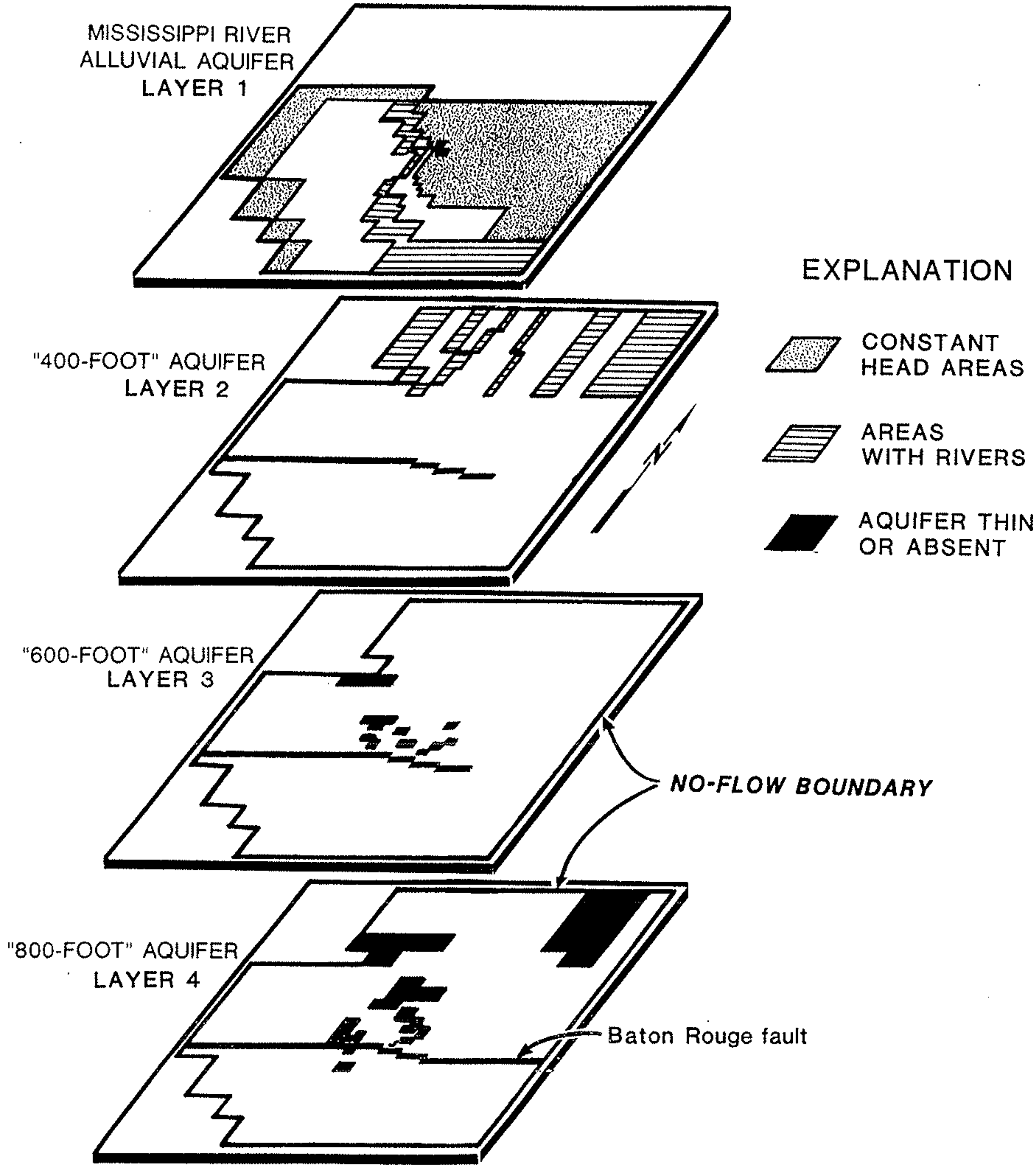


Figure 12.--Top four model layers showing constant head areas, areas with rivers, missing aquifer areas, and location of the Baton Rouge fault.

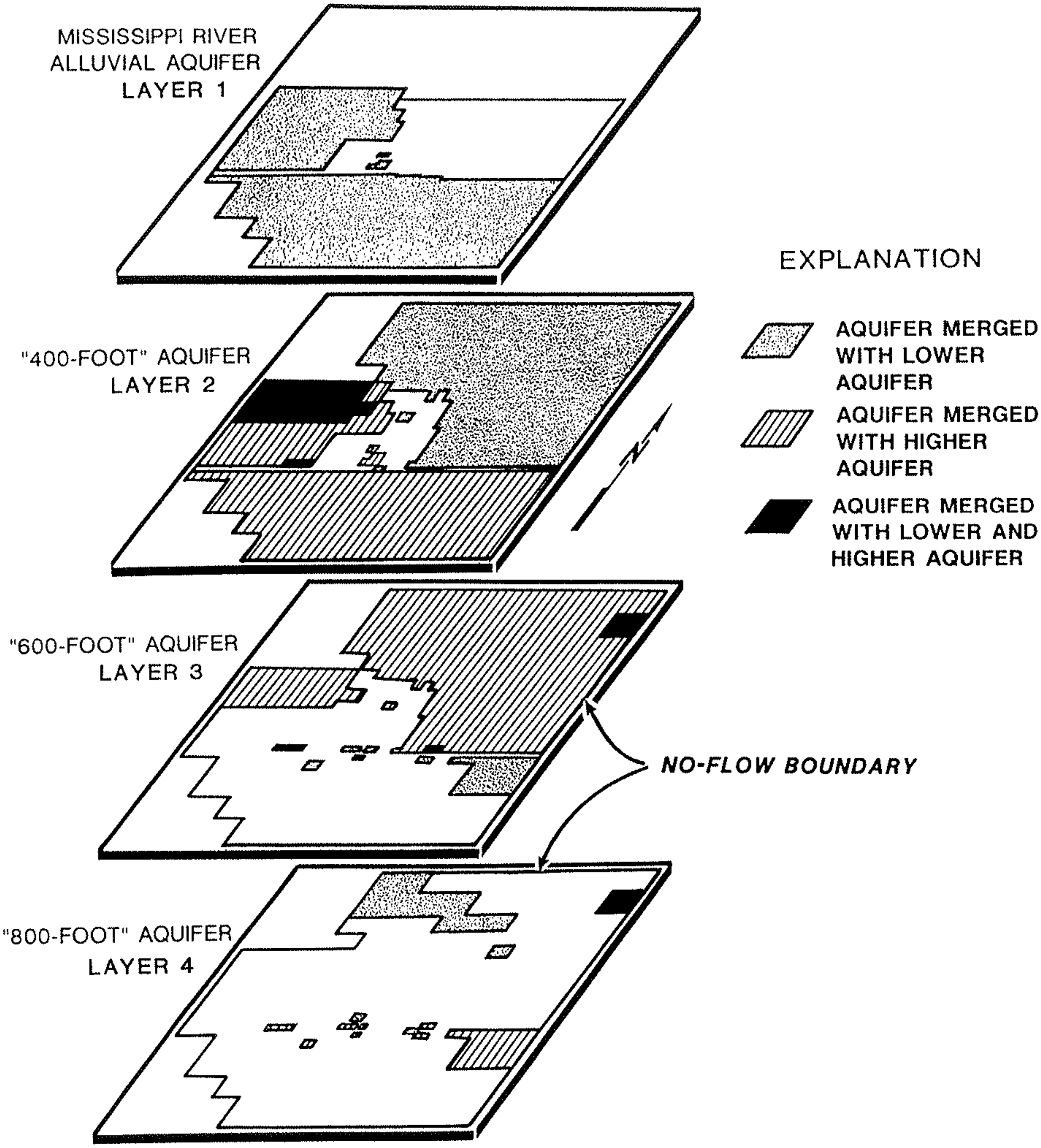


Figure 13.--Top four model layers showing locations of merged aquifer areas.

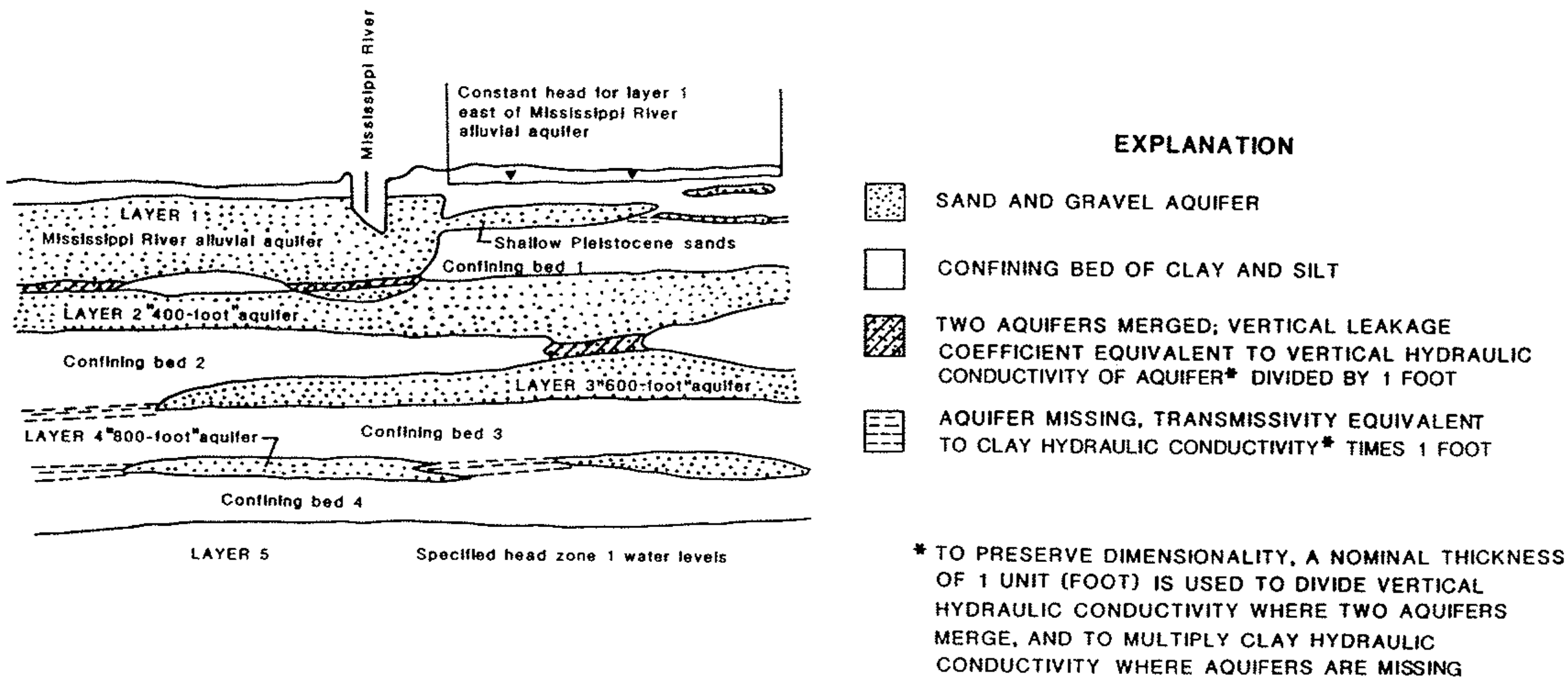


Figure 14.--Generalized cross section showing model layers.

Incorporating the Baton Rouge Fault into the Model

Whiteman (1979) published information on leakage across the Baton Rouge fault. Water-level data were collected from wells placed in the "600-foot" aquifer north of the fault and the "400-foot" aquifer south of the fault (modeled as layer 3). These data were used to analytically estimate the transmissivity across the fault.

During the fall of 1971, the Mississippi River stage at Baton Rouge remained fairly constant (6.8 to 8.8 ft above sea level) from August 22 to December 2. Whiteman (1979, p. 12) analyzed water levels in the "400-foot" aquifer south of the fault and determined northward flow from south of the fault to be 78,000 ft^3/d using the formula $Q = TIL$ where: $T = 8,000\ ft^2/d$ (transmissivity); $I = 3$ ft/mi (hydraulic gradient); and $L = 3.25$ mi (length of closed contour). An estimate of the transmissivity between wells EB-871 (screened in the "400-foot" aquifer south of the fault) and EB-869 (screened in the "600-foot" aquifer north of the fault) was made by assuming that Q and L remain constant and using the gradient between the two wells (750 ft/mi) during that period. The estimated transmissivity between wells EB-871 and EB-869 was 32 ft^2/d.

In the model, the fault is simulated as being 5,000 ft wide (the width of the finite-difference cell) so the estimated transmissivity obtained analytically (Ta) must be adjusted to this width. Because the water levels on either side of the fault and the flow across the fault should remain the same in the model, then by rearranging Darcy's law the modeled fault transmissivity (Tm) would be equal to the estimated fault transmissivity (Ta) multiplied by 5,000 ft and divided by 400 ft (the distance between wells EB-871 and EB-869). This results in a Tm of 400 ft^2/d which is between one and two orders of magnitude less than the transmissivity of the "600-foot" aquifer (table 4).

The location of the modeled fault is shown in figure 11. The line is dashed to the east in Livingston Parish where the fault does not form a barrier to flow in the "400-foot" and "600-foot" aquifers. The fault is believed to be a barrier in deeper aquifers in Livingston Parish and was included in the "800-foot" and "1,000-foot" aquifers (layers 4 and 5). There were no data to compute transmissivity of the fault analytically for the other aquifers in the model. Because displacement of the fault increases with depth, it was assumed that the reduction in transmissivity increases with depth.

Model Calibration

Purpose and Procedure

The purpose of model calibration is to refine the conceptual model of the "400-foot" and "600-foot" aquifers and to learn more about the geohydrologic system. Calibration is accomplished by the systematic adjustment of values for model parameters (transmissivity, vertical leakage coefficient, and storage coefficient), so that there is a good fit between model response and observed data. The modeled aquifer layers respond to stresses (pumpage, recharge, and river stage) in the form of computed water levels at each model cell. Calibration requires reliable estimates of the stresses applied to each layer and observed water-level data for comparison with simulated water levels.

Steady-state calibration was unacceptable for this model study because Mississippi River stage varies as much as 30 ft seasonally. This creates non-equilibrium conditions near the river which flows adjacent to the industrial district.

Transient calibration was performed for the period 1940-48 using hydrographs from wells Li-11, EB-293, -53, -74, -78, -45, -4, -9, -20, -10, -22, -15, -125, and -128, and for the periods May and October 1984 using observed potentiometric surfaces for May and October 1984. Starting water levels were estimated for each model layer for 1935 and 1983. Monthly stress periods were used from 1940-48 and 1983-84. The decision to use monthly stress periods rather than daily or quarterly stress periods was based on the ability to obtain monthly tabulations of pumpage from water-use records. Mean monthly river stage along each reach of the Mississippi River was computed for each stress period, and recharge was input monthly according to the seasonal water-budget surplus shown in table 2.

Time was discretized on a monthly basis from 1940-48 and 1983-84. It was found that having several time steps per monthly stress period only increased computations and produced negligible changes in model response.

An iterative approach between the two calibration periods was used. First a preliminary calibration was made on the 1940-48 hydrographs. Then all parameters were adjusted for calibration to the May and October 1984 data. It was learned through this procedure that two sets of vertical leakage coefficients in the area of subsidence were necessary to calibrate the two separate time periods. The iterative approach between the two time periods led to an investigation of subsidence, caused by ground-water withdrawals at Baton Rouge and the estimation of the volume of water removed from storage in the confining beds caused by the significant lowering of water levels during the beginning of major ground-water development in the 1940's and continuing through 1960.

Man-induced subsidence could have a permanent effect on the hydraulic conductance of the clays due to the compaction of the clays. Therefore, the only parameter difference between the two calibrated periods is in the values for the vertical conductance of the confining beds within the area of maximum subsidence. Calibration resulted in obtaining one set of transmissivity, storage-coefficient, and river leakage-coefficient data and two sets of vertical-leakage coefficient data for all four modeled layers.

Preliminary Simulations

Several steady-state and transient simulations were made to determine general model sensitivity, to test for the effects of error in initial conditions, and to test how long it takes for water-level declines from pumpage to approach steady-state water levels. To test general model sensitivity, some model parameters were varied and different recharge rates applied to rows 1-5 in layer 2 (outcrop area of the "400-foot" aquifer). It was determined that transmissivity and vertical leakage coefficients had the largest effect on water levels. The river leakage coefficient (river length times width times bed conductance divided by bed thickness) was adjusted by orders of magnitude until the water levels in the aquifer were within 0.5 ft of the river stage. This resulted in a river leakage coefficient equal to the river reach length times 10 for rivers simulated in layer 2 (fig. 12) and the river reach length times 100 for the Mississippi River simulated in layer 1 (fig. 12). Recharge rates applied to rows 1-5 in layer 2 ("400-foot" aquifer) affected these rows the most, having little or no effect on water levels at Baton Rouge. A rate of 16 in/yr caused mounding of water between rivers. Rates between 4 to 8 in/yr produced reasonable water levels in the outcrop area during steady-state simulations and over 90 percent of recharge discharged to the rivers when 5 in/yr was input to the outcrop area.

To test for error in initial conditions, several simulations were made by extending the estimated pumping period, prior to 1940, using the pumping rate of January 1940 (20 Mgal/d) and comparing the hydrographs from the simulations with and without pumpage prior to 1940. It was found that an additional 1 year of pumping prior to 1940 lowered water levels in January 1940 by less than 5 ft at pumped cells. An additional 10 years of pumping prior to 1940 lowered water levels for January through June 1940 only with the maximum lowering of less than 20 ft in January 1940. A simulation with initial starting water levels increased by 10 ft, except at constant head cells, showed no change in the hydrographs for 1940-48. Thus, errors in initial starting water levels have a negligible impact on simulated water-levels, and error in pumpage estimates prior to 1940 would affect simulated water levels for less than 6 months.

By comparison of a steady-state simulation with a transient simulation, both pumped at the same rate as January 1940, it was found that steady-state water levels at pumped cells were reached within 6 ft in 5 years and within 1 ft after 20 years in the "400-foot" and "600-foot" aquifer. Thus, using initial aquifer parameter values, water-level declines from pumpage approach steady-state water-level declines in about 20 years.

Calibration of the Period 1940-48

The period 1940-48 was selected for calibration because of the existence of detailed water-level records from wells screened in the "400-foot" and "600-foot" aquifers in the industrial district at Baton Rouge and from one well in the "600-foot" aquifer in Livingston Parish. Monthly pumpage data had been tabulated for each water user beginning in January 1940. Monthly mean Mississippi River stage data at six gaging stations along the part of the river in the modeled area were available.

Pumpage data prior to 1940 were not available. From examination of the water-level data (fig. 8) and the history of the petrochemical industry, it was concluded that major industrial use began about 1935. The simulation was started in 1935 with a 5 year stress period containing estimated pumpage of

7.8 Mgal/d from the "400-foot" aquifer, 10.2 Mgal/d from the "600-foot" aquifer, and 1.2 Mgal/d from the "800-foot" aquifer. The location of model cells with pumping stress from 1935-48 is shown in figure 15. Monthly pumpage from the "400-foot" and "600-foot" aquifers was reported by each industry for several wells summed together. Total monthly pumpage from the "400-foot" and "600-foot" aquifers is shown in figure 10. Many of the chemical plants estimated water use by the amount of product produced. It is uncertain how much or if any of the tabulated water-use data was from metered wells. Several attempts were made in splitting up the reported pumpage for the "400-foot" and "600-foot" aquifers. Pumpage from the "800-foot" aquifer was estimated from limited water-use data and from the schedules of drilled wells reported by Meyer and Turcan (1955).

Water levels in the "1,000-foot" aquifer were estimated for each year from 1940-48. These hydraulic heads were assigned to layer 5 and leakage between layers 4 and 5 were simulated using the general-head boundary package in each cell of layer 4, the "800-foot" aquifer, by having the general-head conductance term be the vertical leakage coefficient of confining bed 4 multiplied by each cell area. (See fig. 14.)

It was determined through preliminary steady-state simulations that most recharge in the outcrop area of the "400-foot" aquifer discharged to the streams as base flow. Because there were no observation well in the outcrop area, seasonal fluctuations in the outcrop area were not simulated. Recharge was input at a constant rate throughout the transient simulation. In rows 1-3 a rate of 8.76 in/yr, in row 4 a rate of 4.38 in/yr, and in row 5 a rate of 2.19 in/yr was input for all stress periods. This is equivalent to 6.85 in/yr over the 1,592 mi^2 outcrop area.

The major water-level declines in the "400-, 600-, and 800-foot" aquifers occurred between 1937-40 (fig. 8). Most water from storage in the confining beds would have been released during 1935-49 because of the significant water-level declines (fig. 8). The distribution in time of water released from the clay confining beds and lenses is unknown because of variations in the hydraulic properties of the clays and in their thickness. It was concluded that the simplest way to account for this water entering the aquifer system was by a linear reduction of pumpage. A more complex analysis would require information about the confining beds that is not available. The amount of reduction was determined after calibration of the model for the May 1984 data in which the best potentiometric data were available for much of the aquifer system. A 20 percent reduction (equivalent to 12,260 Mgal of water, January 1935 to June 1948; this is 92 percent of the volume of water estimated as coming from storage; see section on transient leakage) gave a reasonable fit to hydrographs, and a 40 percent reduction was too large.

Figure 16 shows the location of the observation wells used in the transient model calibration of January 1940 to June 1948. Figures 17-24 show the simulated water-level hydrographs and the measured water-level data for each model cell with an observation well. These hydrographs were from a simulation where reported pumpage had been reduced by 20 percent for each stress period. The simulated hydrographs fit observed data well. Plate 1 shows the simulated potentiometric surface of the "400-foot" and "600-foot" aquifers in August 1944 when the aquifers were drawn down the most during the calibration period (fig. 10).

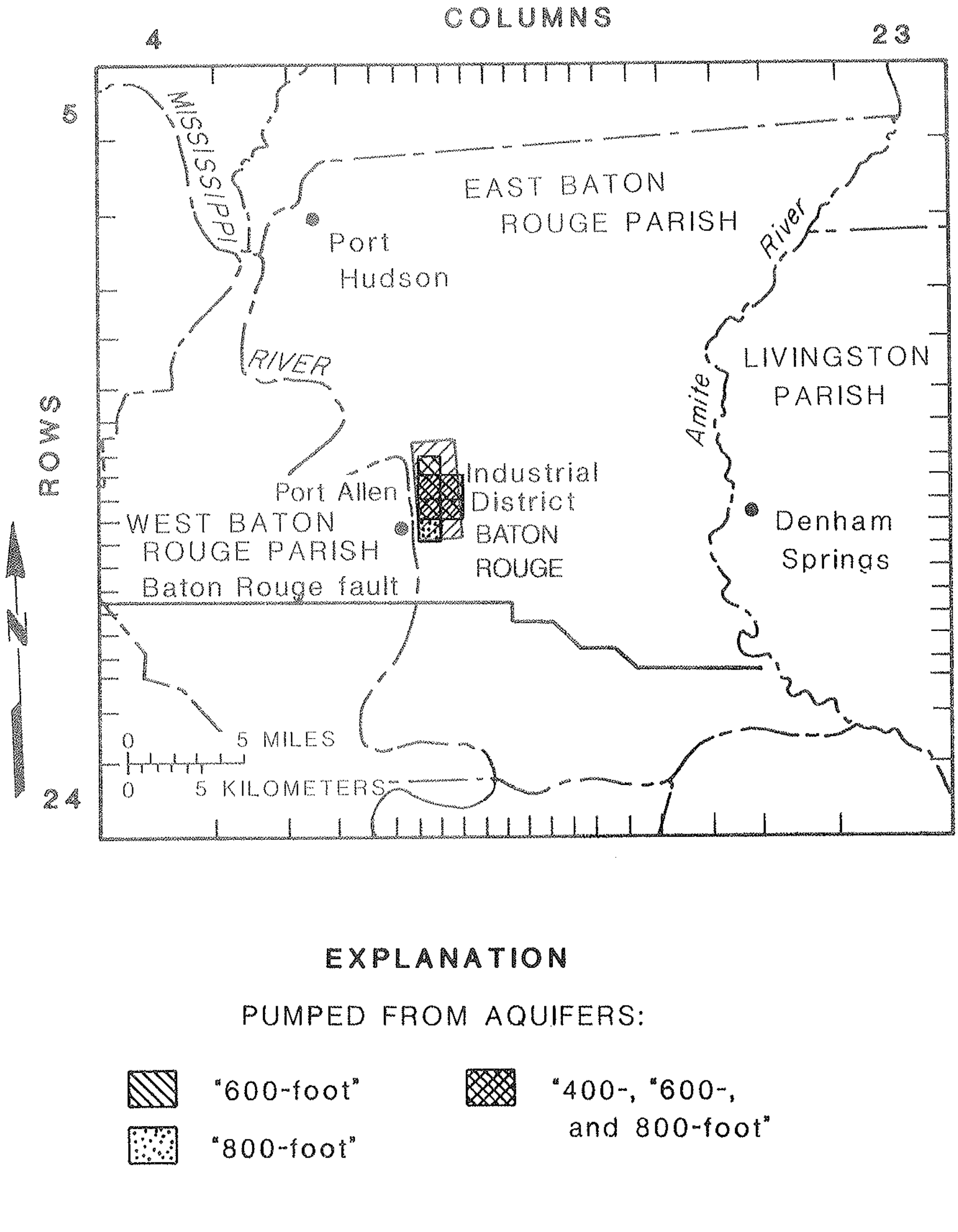


Figure 15.--Location of finite-difference cells with pumping stress, 1935-48.

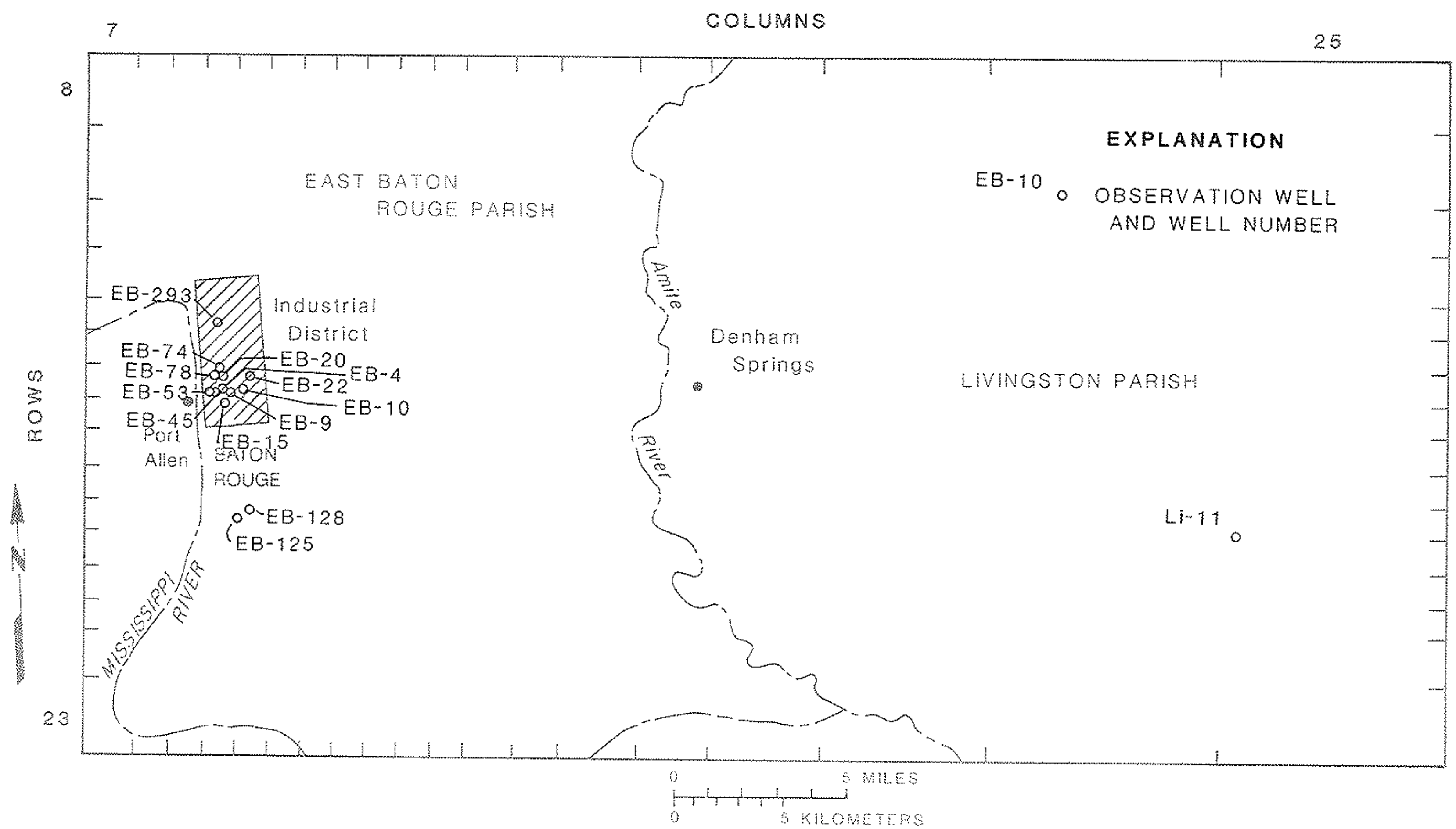


Figure 16.--Location of observation wells used in the transient model calibration, January 1940 through June 1948.

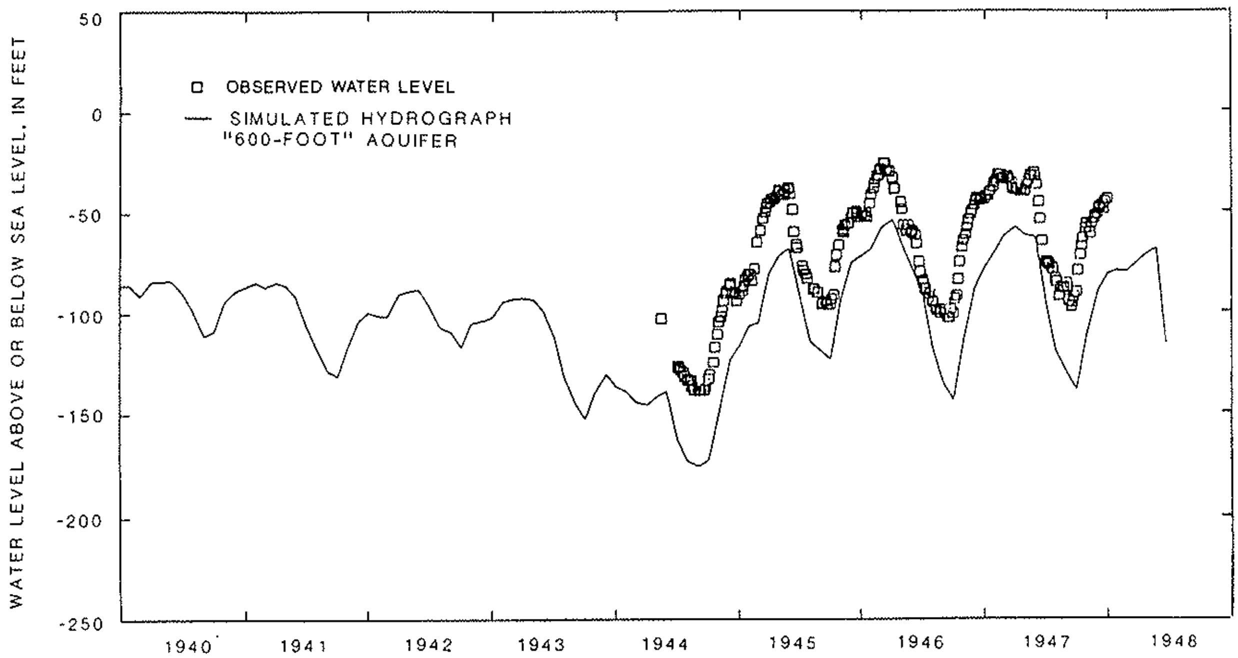


Figure 17.--Simulated hydrograph for the "600-foot" aquifer, row 12 column 10, and observed water level in well EB-293, 1940-48.

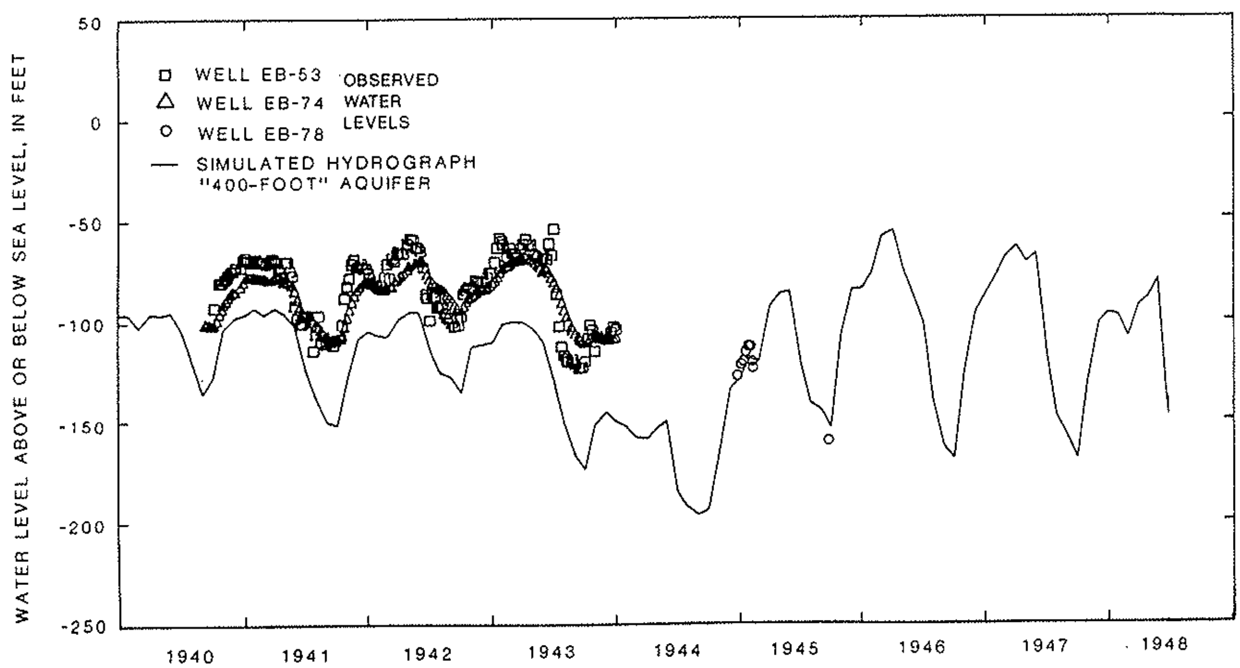


Figure 18.--Simulated hydrograph for the "400-foot" aquifer, row 14 column 10, and observed water level in wells EB-53, EB-74, and EB-78, 1940-48.

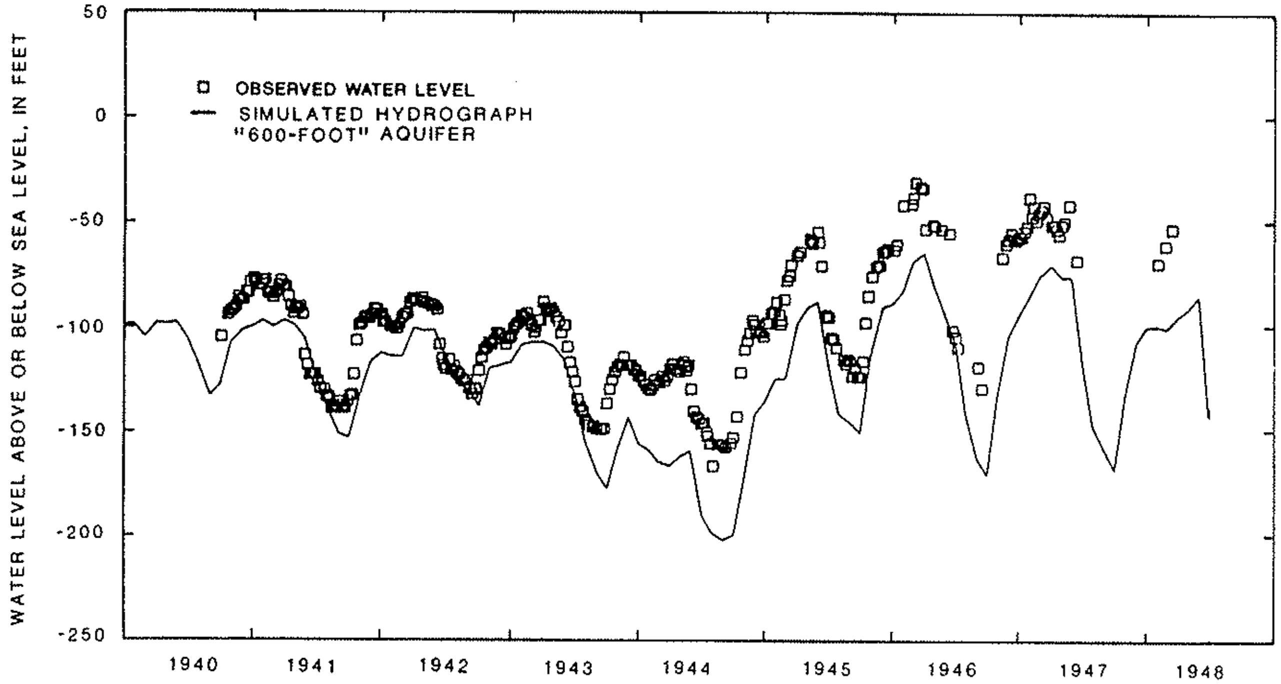


Figure 19.--Simulated hydrograph for the "600-foot" aquifer, row 14 column 10, and observed water level in well EB-45, 1940-48.

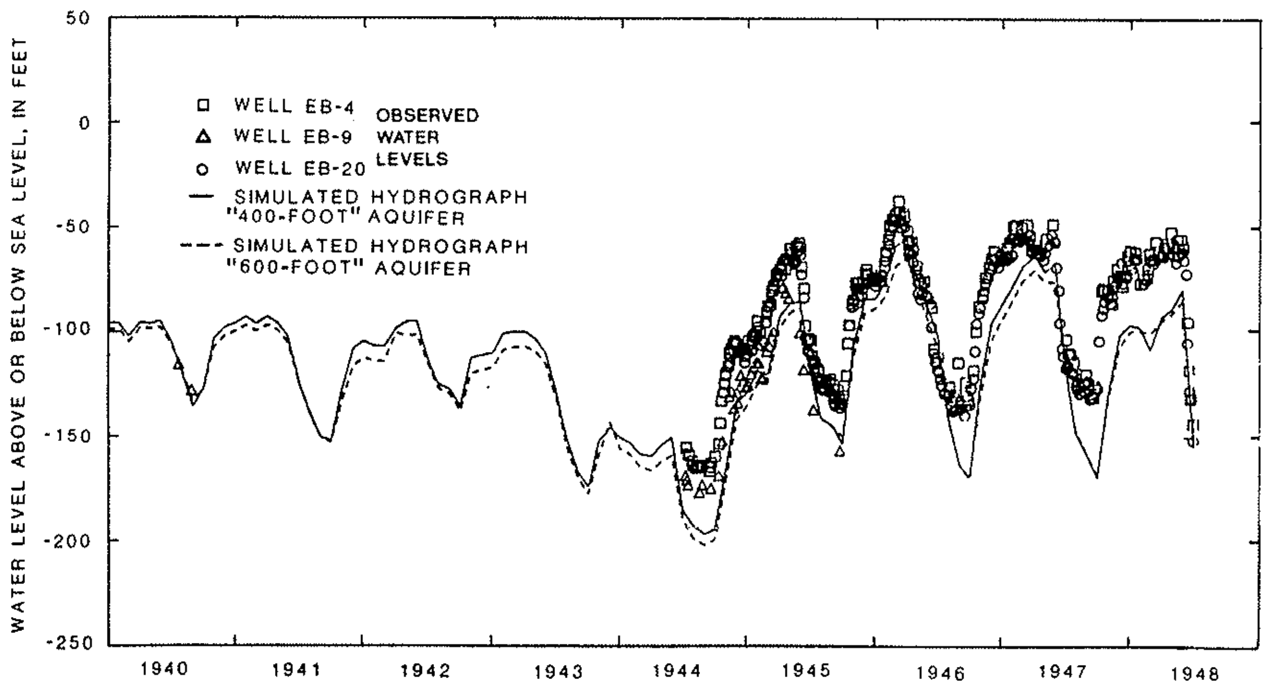


Figure 20.--Simulated hydrographs for the "400-foot" and "600-foot" aquifers, row 14 column 10, and observed water level in wells EB-4, EB-9, and EB-20, 1940-48.

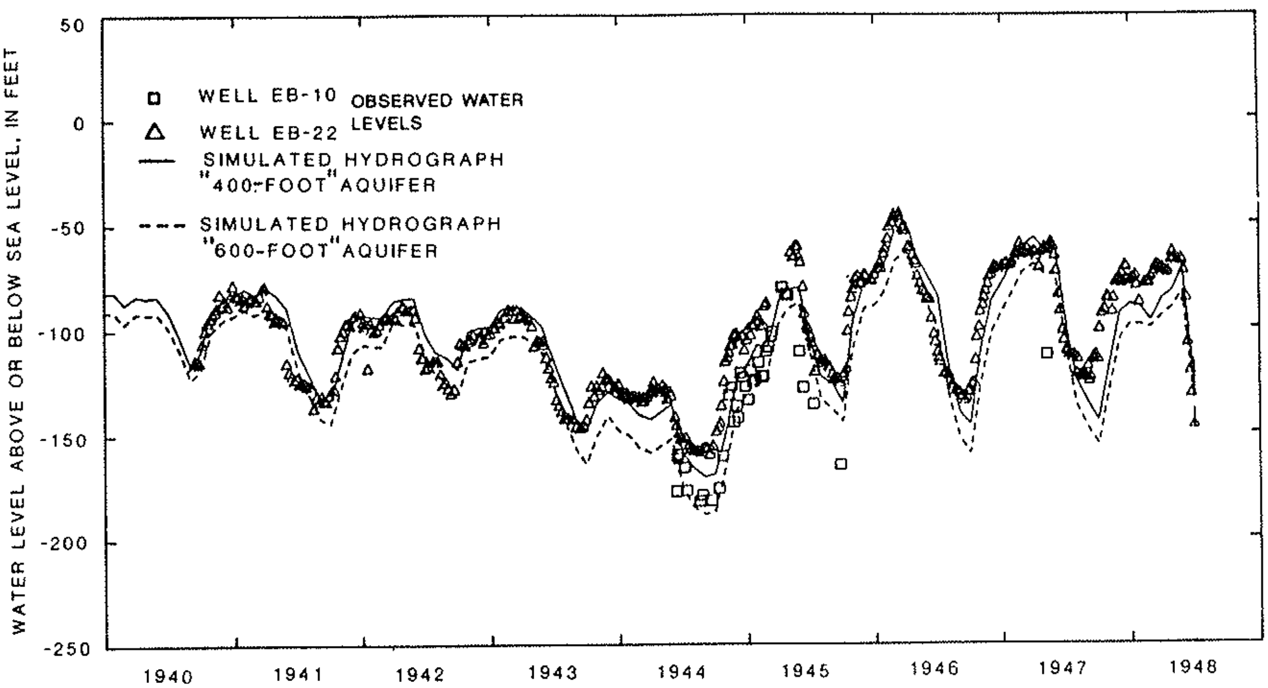


Figure 21.--Simulated hydrographs for the "400-foot" and "600-foot" aquifers, row 14 column 11, and observed water level in wells EB-10 and EB-22, 1940-48.

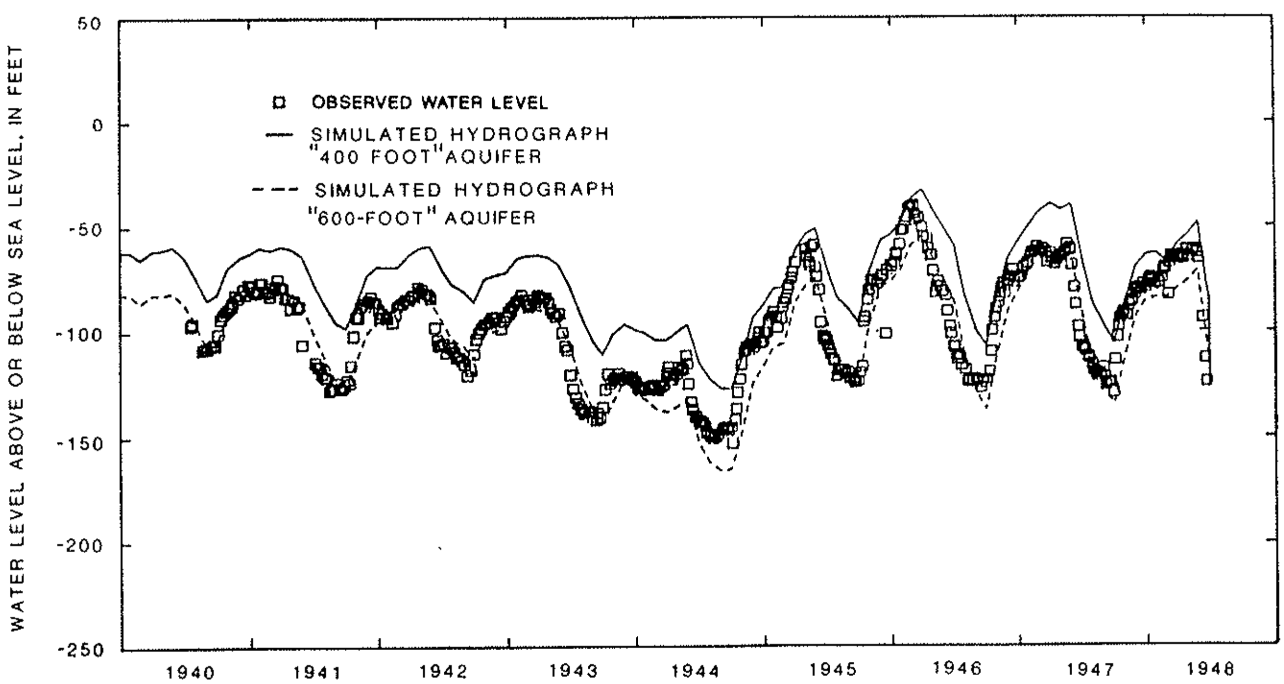


Figure 22.--Simulated hydrographs for the "400-foot" and "600-foot" aquifers, row 15 column 10, and observed water level in well EB-15, 1940-48.

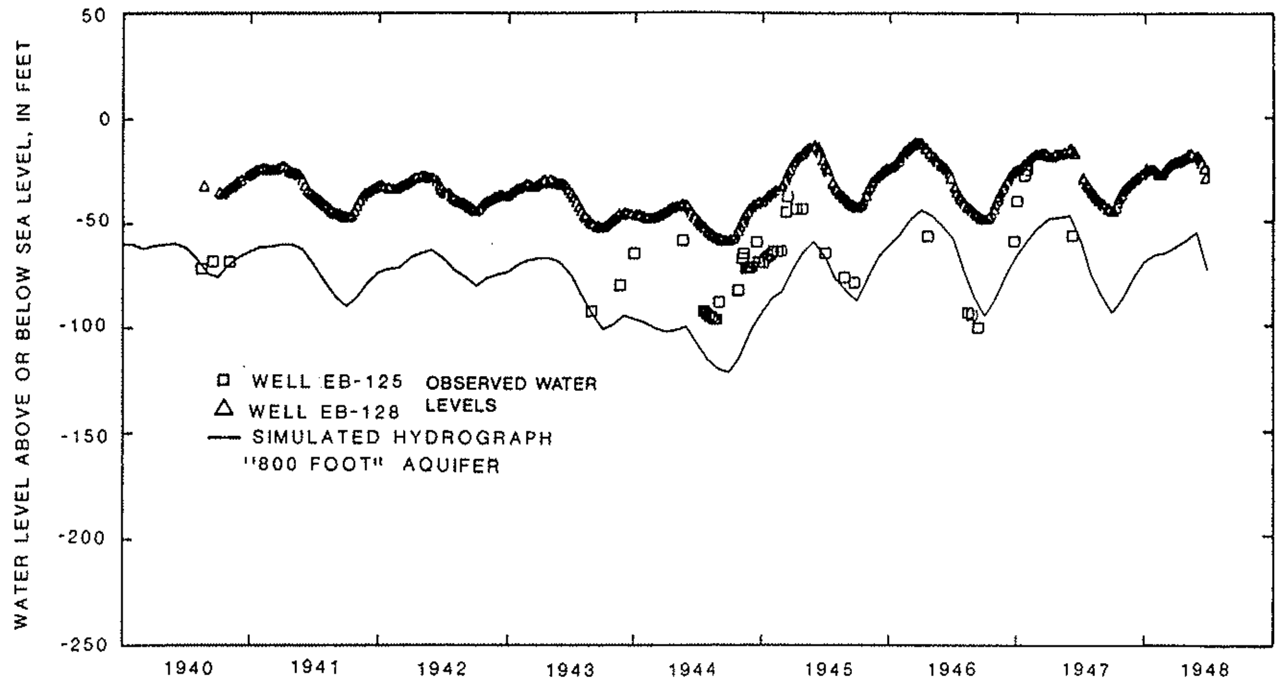


Figure 23.--Simulated hydrograph for the "800-foot" aquifer, row 17 column 11, and observed water level in wells EB-125 and EB-128, 1940-48.

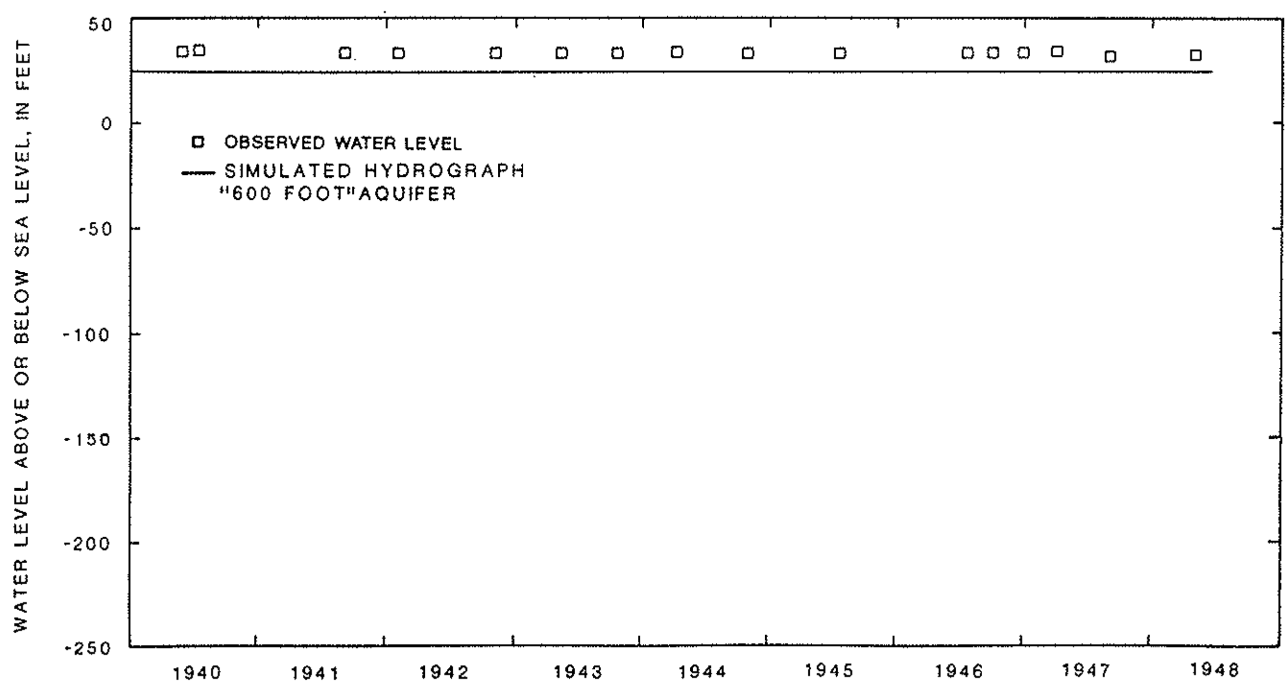


Figure 24.--Simulated hydrograph for the "600-foot" aquifer, row 18 column 25, and observed water level in well Li-11, 1940-48.

Calibration of the Period May and October 1984

Water-level data were collected in May and October 1984 to make a potentiometric map of the "400-foot" and "600-foot" aquifers. These two periods were selected because the Mississippi River stage peaks in May, and in October the stage is low.

The transient simulation was started in January 1983 by estimating water-levels for each layer. Preliminary simulations indicated that any major error in initial conditions would not effect model-simulated water levels after about 6 months. For this reason, the simulation was started 16 months prior to the month of calibration and consisted of 24 monthly stress periods, 1983-84. Throughout the simulation, water levels in the "1,000-foot" aquifer, layer 5, were assigned constant values, using the water levels published for 1980 (Martin and Whiteman, 1985).

Monthly pumpage data were obtained from the Capital Area Ground Water Conservation Commission. Figure 25 shows the location of model cells with pumpage. Mean river stage for each month was computed for each model reach of the Mississippi River.

Recharge rates were varied seasonally to account for seasonal variations in evapotranspiration. Recharge rates were decreased in rows 4 and 5 to account for the thickening surficial silts and clays in the transitional area from outcrop to fully confined conditions. During December through March, 14.45 in/yr was assigned to rows 1-3, 7.22 in/yr to row 4, and 3.61 in/yr to row 5 of layer 2, the "400-foot" aquifer. During April and November 10.07 in/yr was assigned to rows 1-3, 5.0 in/yr to row 4, and 2.52 in/yr to row 5 at layer 2. During May, 7.01 in/yr was assigned in rows 1-3, 3.51 in/yr in row 4, and 1.75 in/yr in row 5 of layer 2. During June through October, only 4.38 in/yr was assigned to rows 1-3, 2.19 in/yr to row 4, and 1.09 in/yr to row 5 of layer 2. This is an average annual rate of 6.97 in/yr over the 1,592 mi^2 area.

Plate 2 shows simulated and observed potentiometric maps for the "400-foot" and "600-foot" aquifers for May 1984. Plate 3 shows model simulated and observed potentiometric maps for the "400-foot" and "600-foot" aquifers of October 1984. Water levels were simulated with a maximum error of about 20 ft.

Calibrated Values of Hydraulic Properties

Three hydraulic properties are associated with each model layer: transmissivity, vertical leakage coefficient, and storage coefficient.

Maps of transmissivity for model layers 1-4 are shown in figures 26-29. Transmissivity is defined as sand thickness multiplied by hydraulic conductivity. Hydraulic conductivity is adjusted during calibration within the typical range associated with the lithology of the sediments (tables 3 and 4). For a mixed coarse sand and gravel, such as the Mississippi River alluvial aquifer which is the western part of layer 1, a hydraulic conductivity of

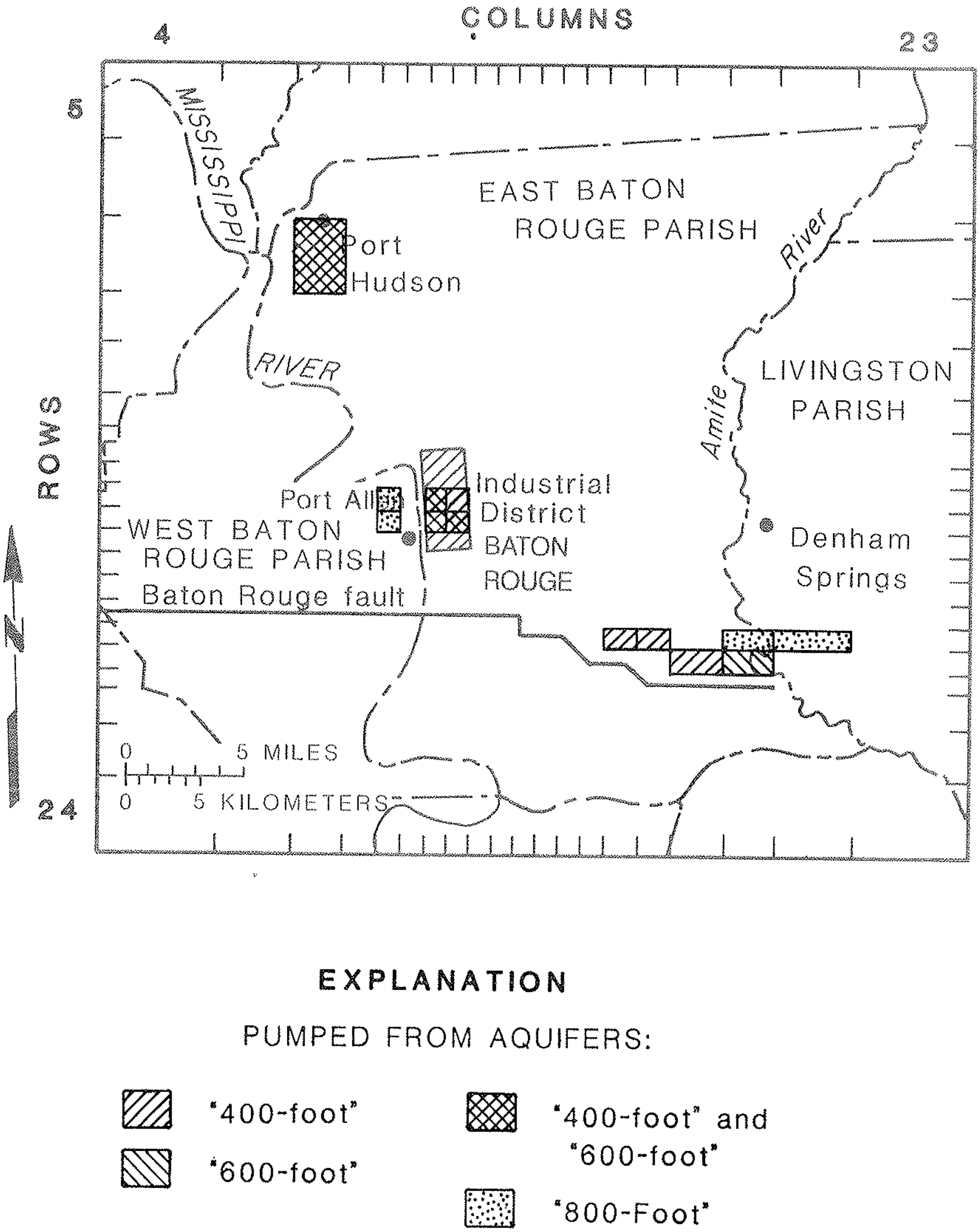
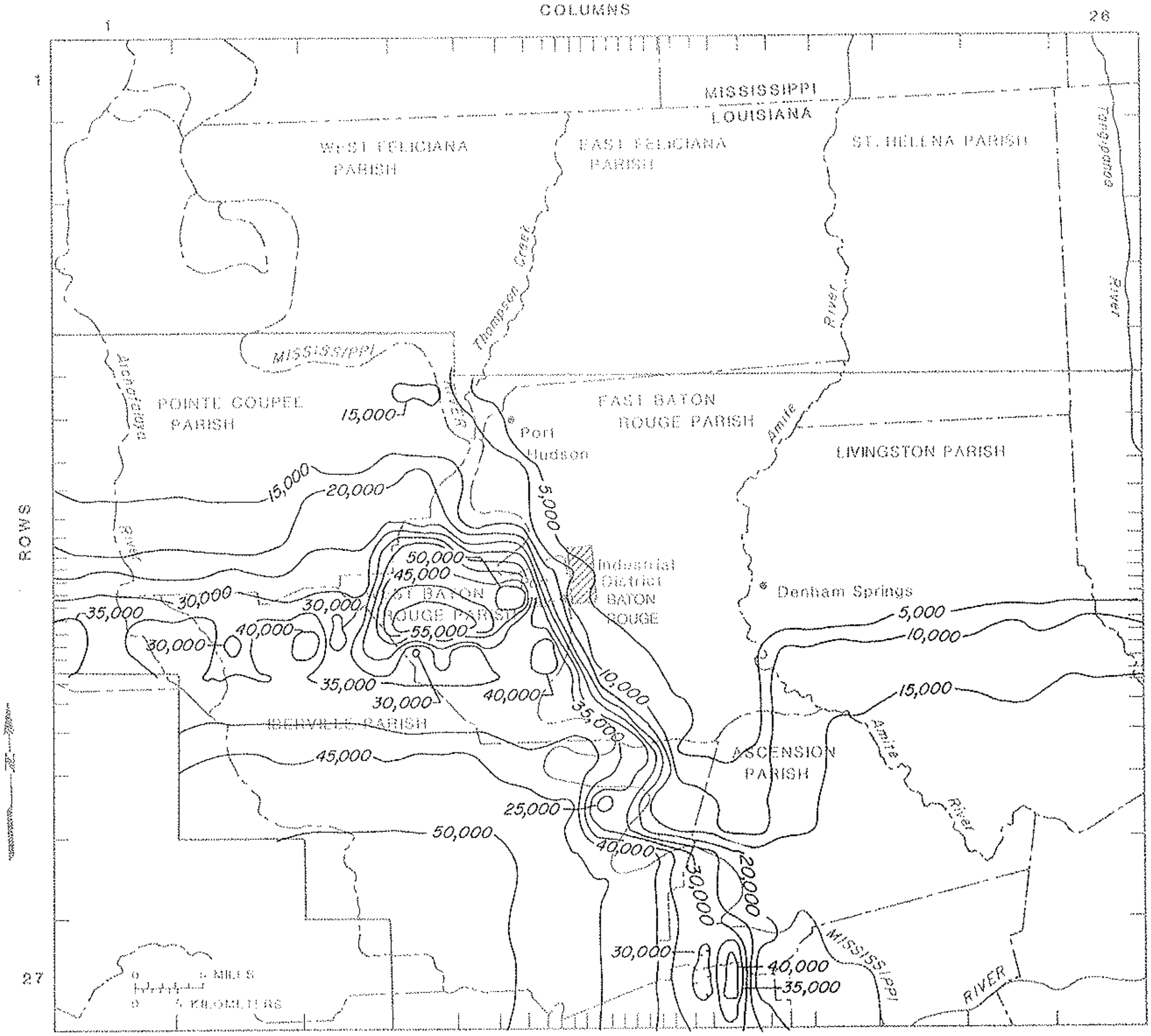


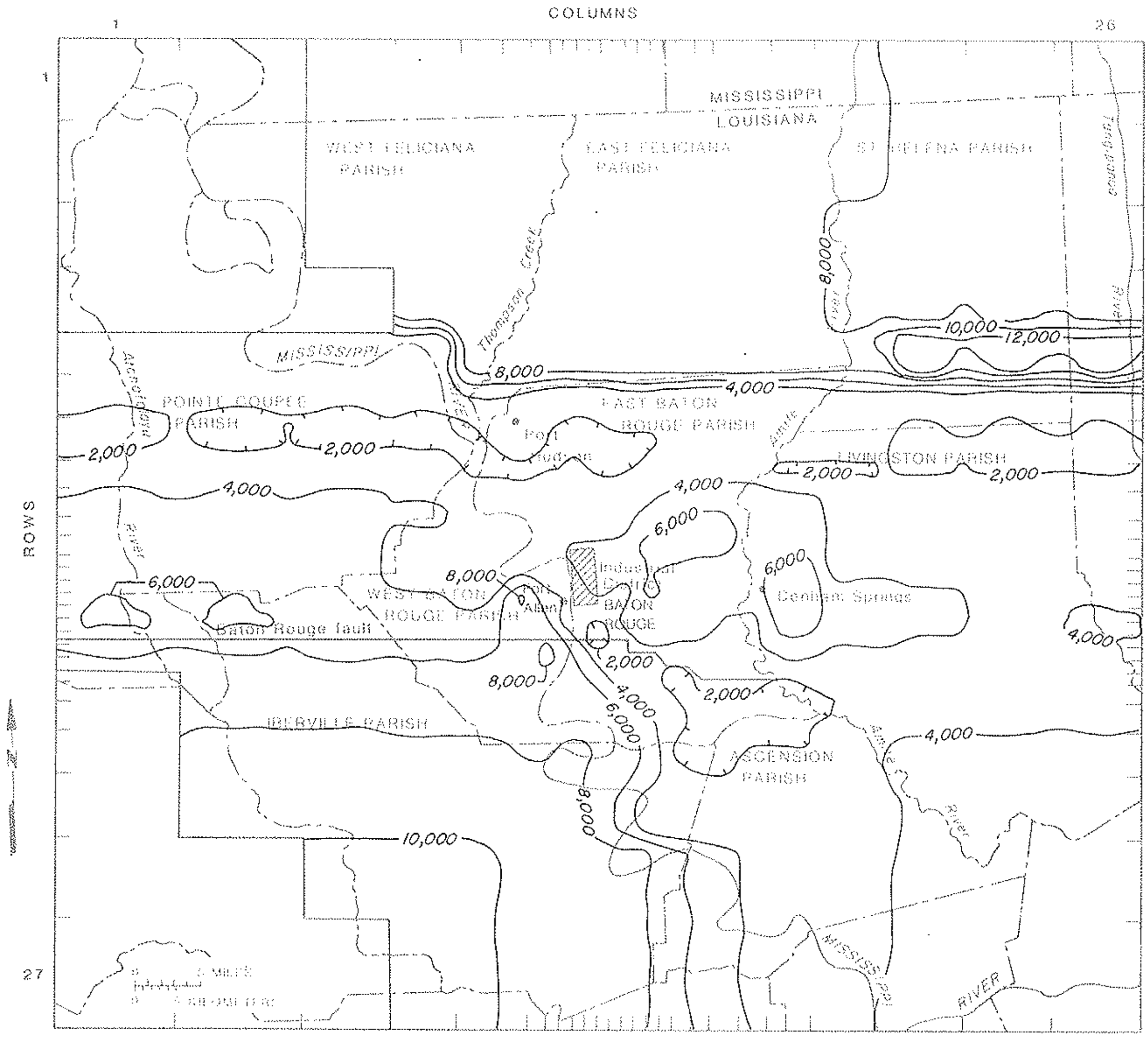
Figure 25.--Location of finite-difference cells with pumping stress, 1983-84.



EXPLANATION

—— 20,000 —— LINE OF EQUAL TRANSMISSIVITY--
Hachures indicate depression.
Interval 5,000 feet squared per day

Figure 26.--Transmissivity of the Mississippi River alluvial aquifer and shallow Pleistocene sands, model layer 1.



EXPLANATION

—— 2,000 —— LINE OF EQUAL TRANSMISSIVITY--
Hachures indicate depression.
Interval 2,000 feet squared per day

Figure 27.--Transmissivity of the "400-foot" aquifer, model layer 2.

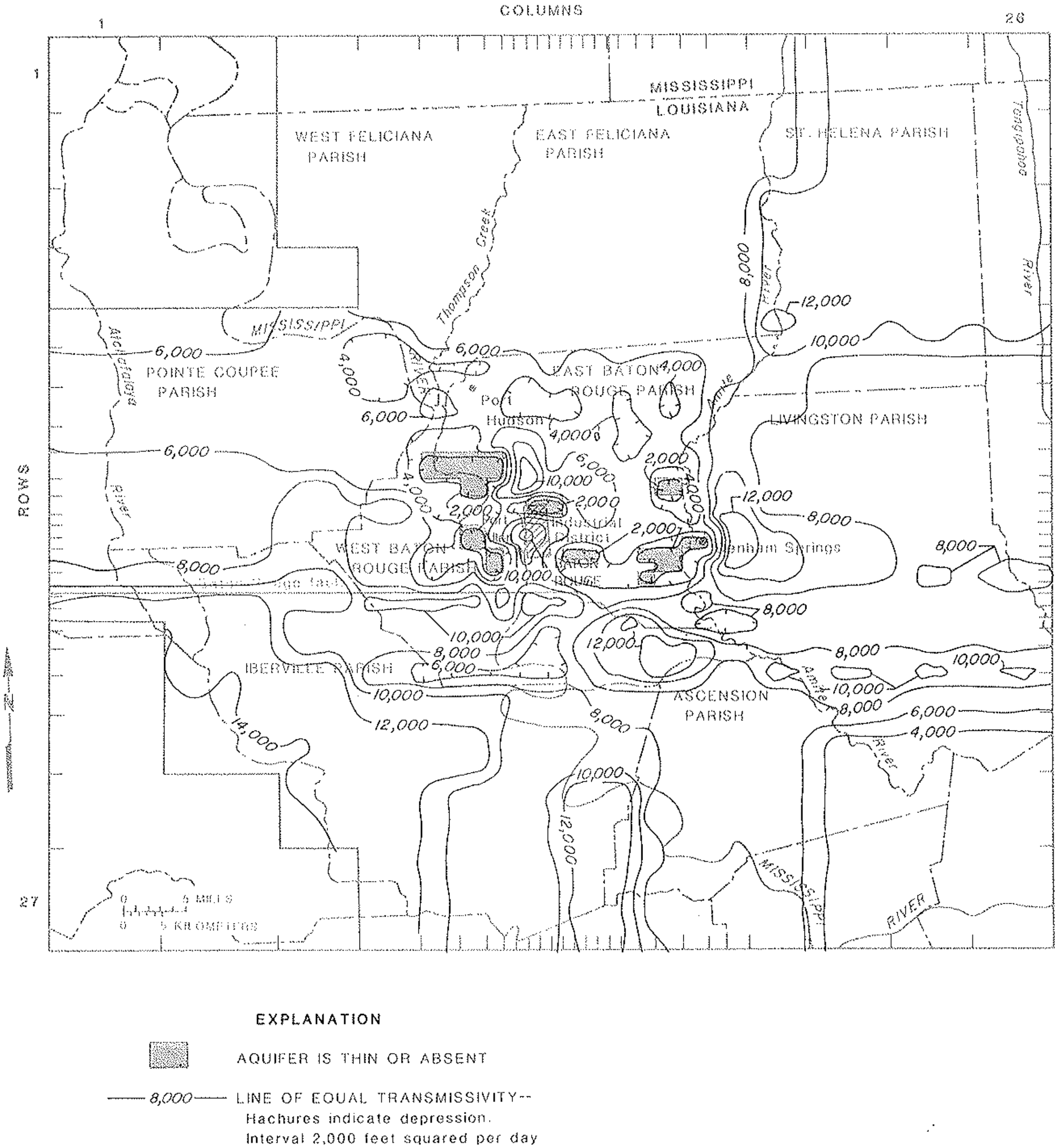
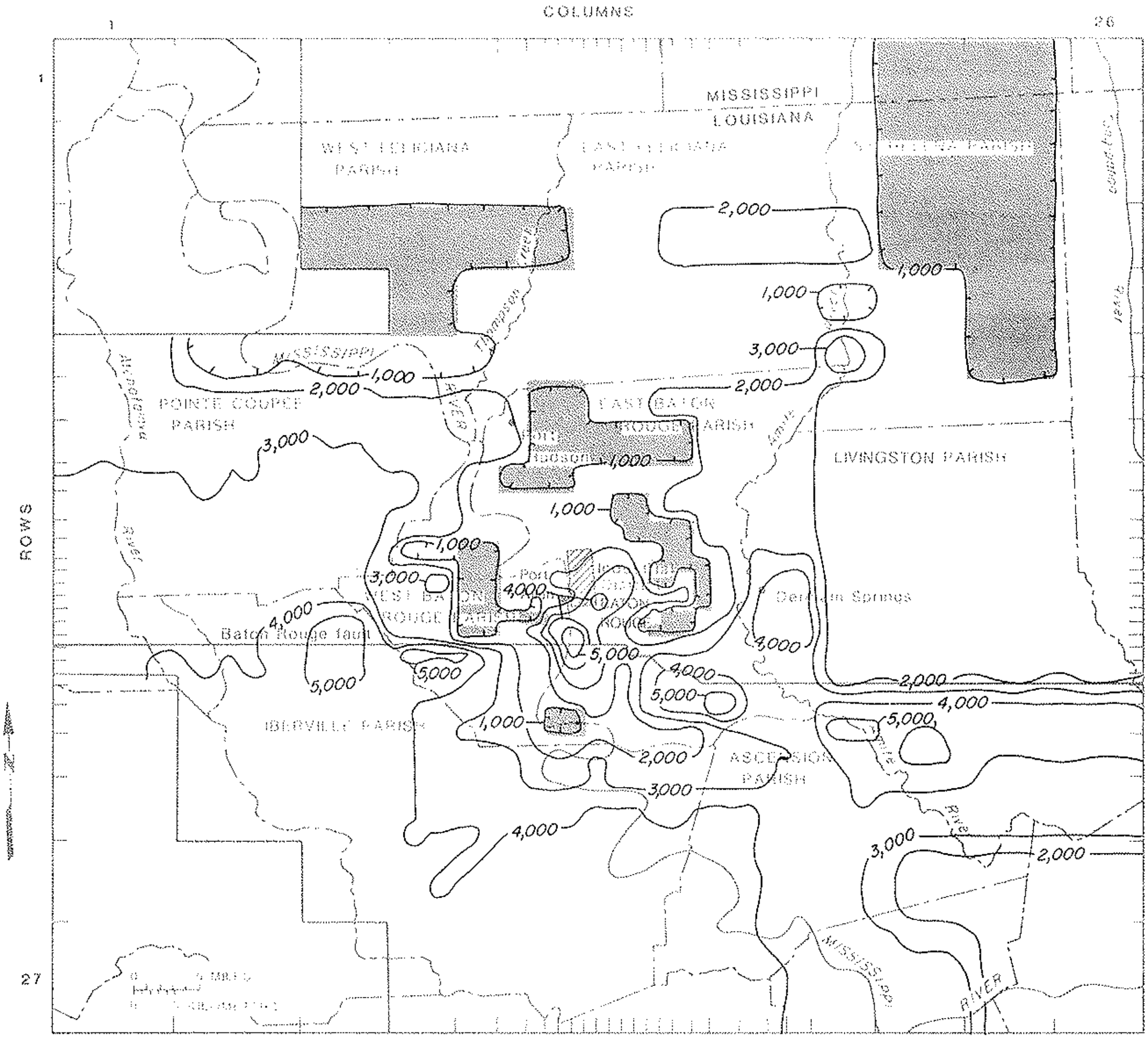


Figure 28.--Transmissivity of the "600-foot" aquifer, model layer 3.



EXPLANATION

AQUIFER IS THIN OR ABSENT

—— 2,000 —— LINE OF EQUAL TRANSMISSIVITY--
Hachures indicate depression.
Interval 1,000 feet squared per day

Figure 29.--Transmissivity of the "800-foot" aquifer, model layer 4.

90 ft/d is used in calibration. For the mixed sands and gravels in the outcrop area of the "400-foot" and "600-foot" aquifers about 80-90 ft/d was used as the hydraulic conductivity. The calibrated hydraulic conductivity for the fine to medium sands was 35-40 ft/d and for medium to coarse sands was 60-70 ft/d. The transmissivity of the Baton Rouge fault was decreased with depth to account for the increased displacement with depth. For layers 2, 3, and 4 the values used were 35, 6, and 0.4 ft^2/d, respectively.

Maps of vertical leakage coefficients for model confining layers 1-4 (fig. 14) are shown in figures 30-33. Vertical leakage coefficient is defined as vertical hydraulic conductivity of the confining bed divided by the thickness of the confining bed. Vertical hydraulic conductivity is adjusted during calibration within the typical range associated with silts and clays (table 3). Vertical hydraulic conductivity was decreased from 0.001 to 0.00075 ft/d for confining beds 1 and 2, to 0.00075 ft/d for confining bed 3, and to 0.00001 for confining bed 4 (fig. 14). In the area of maximum subsidence, (fig. 7a), the vertical conductivity was reduced according to subsidence for the calibration period 1983-84, in order to obtain the water-level declines shown on plates 1 and 2. The absolute maximum reduction in conductivity was 2.0 to 2.9 orders of magnitude over the four model cells in rows 13 and 14 columns 10 and 11. The dashed contours in figures 30-33 in the area of subsidence show the vertical leakage coefficients used in simulations after subsidence.

Maps of storage coefficients are not shown. Layer 1 was simulated with a coefficient of 0.005. Layer 2 was simulated with a coefficient of 0.0005 for rows 6-26 and 0.1 for the outcrop area in rows 1-5. Layers 3 and 4 were simulated using a storage coefficient of 0.0005. These calibrated values provide a reasonable fit to measured water-level data given the known aquifer stresses and are within reasonable ranges of storage coefficients (table 4).

Sensitivity of the Calibrated Model

Purpose and Procedure

Sensitivity analysis provides an indication of how the model parameters affect the model response. Generally, sensitivity analysis is accomplished by changing a parameter such as transmissivity and computing the change in model response from the calibrated water levels. Often this is displayed graphically by plotting the different percent change of parameter values used versus the corresponding mean change in water level from the calibrated sets of parameters. This form of analysis provides minimal information in that no information on the spatial distribution of the change in model response for a given change in parameter is presented. Because many of the parameters of this complexly interbedded aquifer system are highly correlated, normalized sensitivity analysis is presented for 16 sensitivity simulations. Normalized sensitivity, defined as the change in water level from the calibrated water level divided by the percent change in input parameter, is shown spatially through a series of contour maps of equal water-level change divided by percent change in parameter (definition modified from Sykes and others, 1985, p. 262). The same percent perturbation from the calibrated-parameter value is

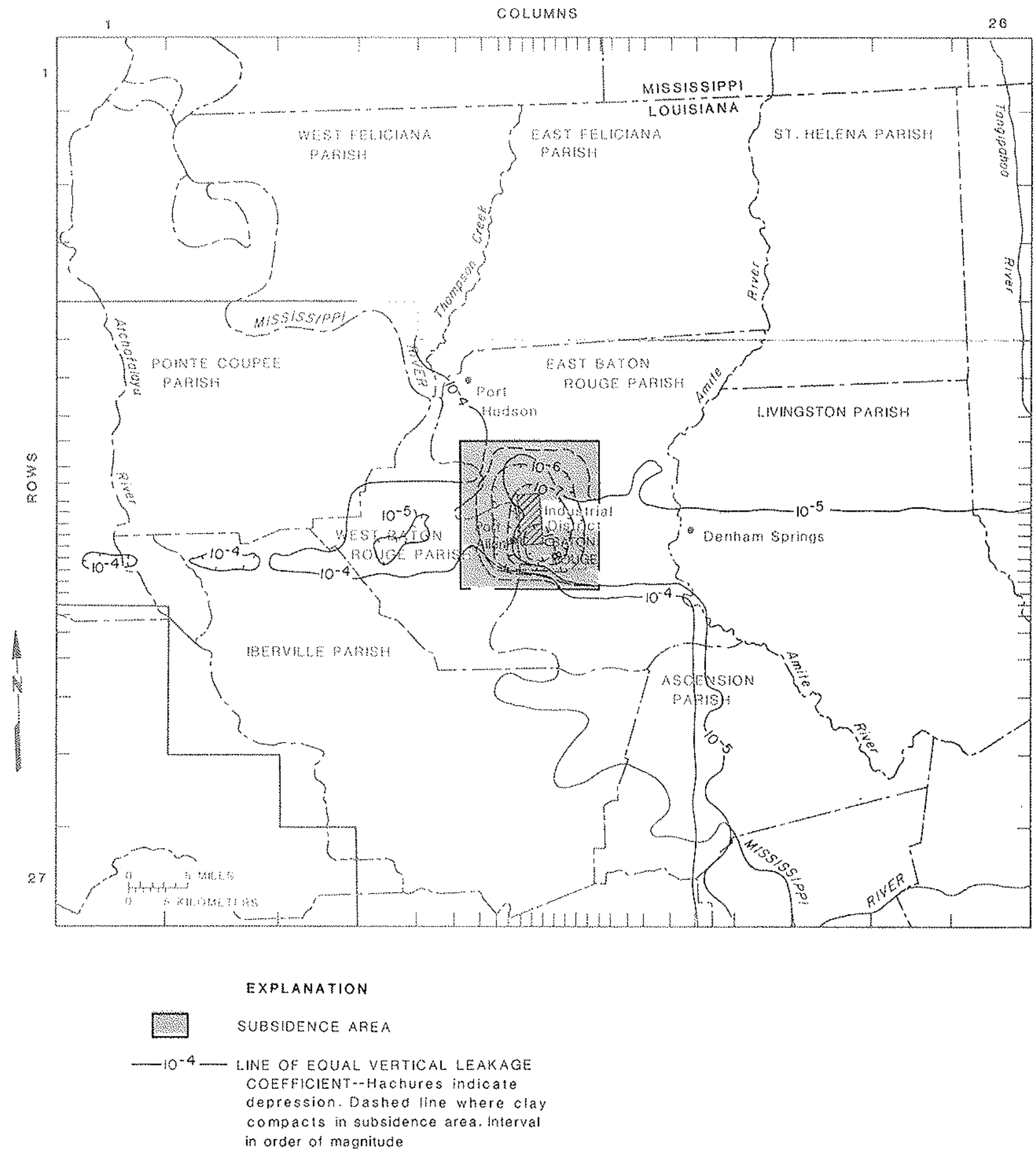


Figure 30.--Vertical leakage coefficient for confining bed 1.

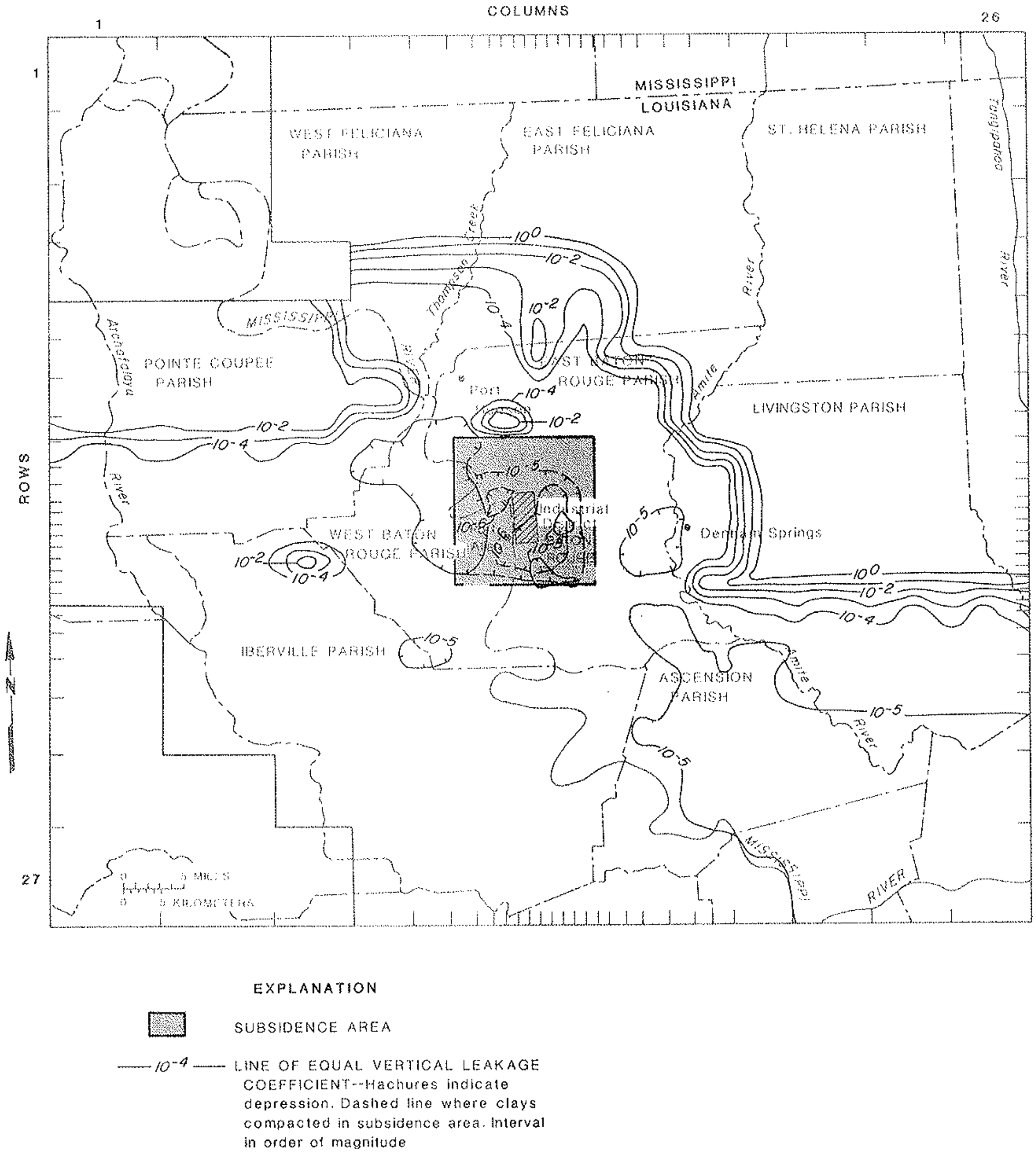
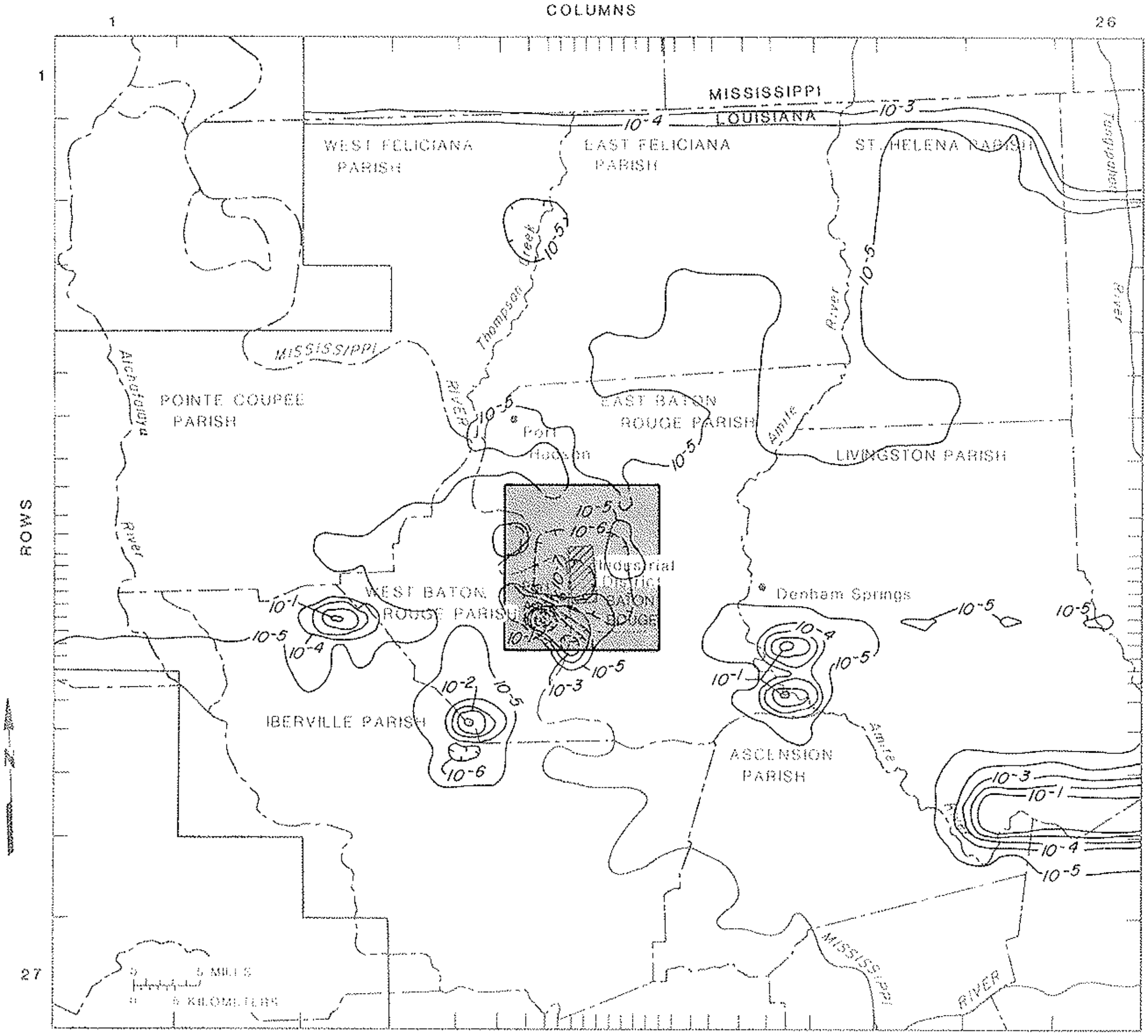


Figure 31.--Vertical leakage coefficient for confining bed 2.



EXPLANATION

SUBSIDENCE AREA

10^{-4} LINE OF EQUAL VERTICAL LEAKAGE COEFFICIENT--Hachures indicate depression. Dashed line where clays compacted in subsidence area. Interval in order of magnitude

Figure 32.--Vertical leakage coefficient for confining bed 3.

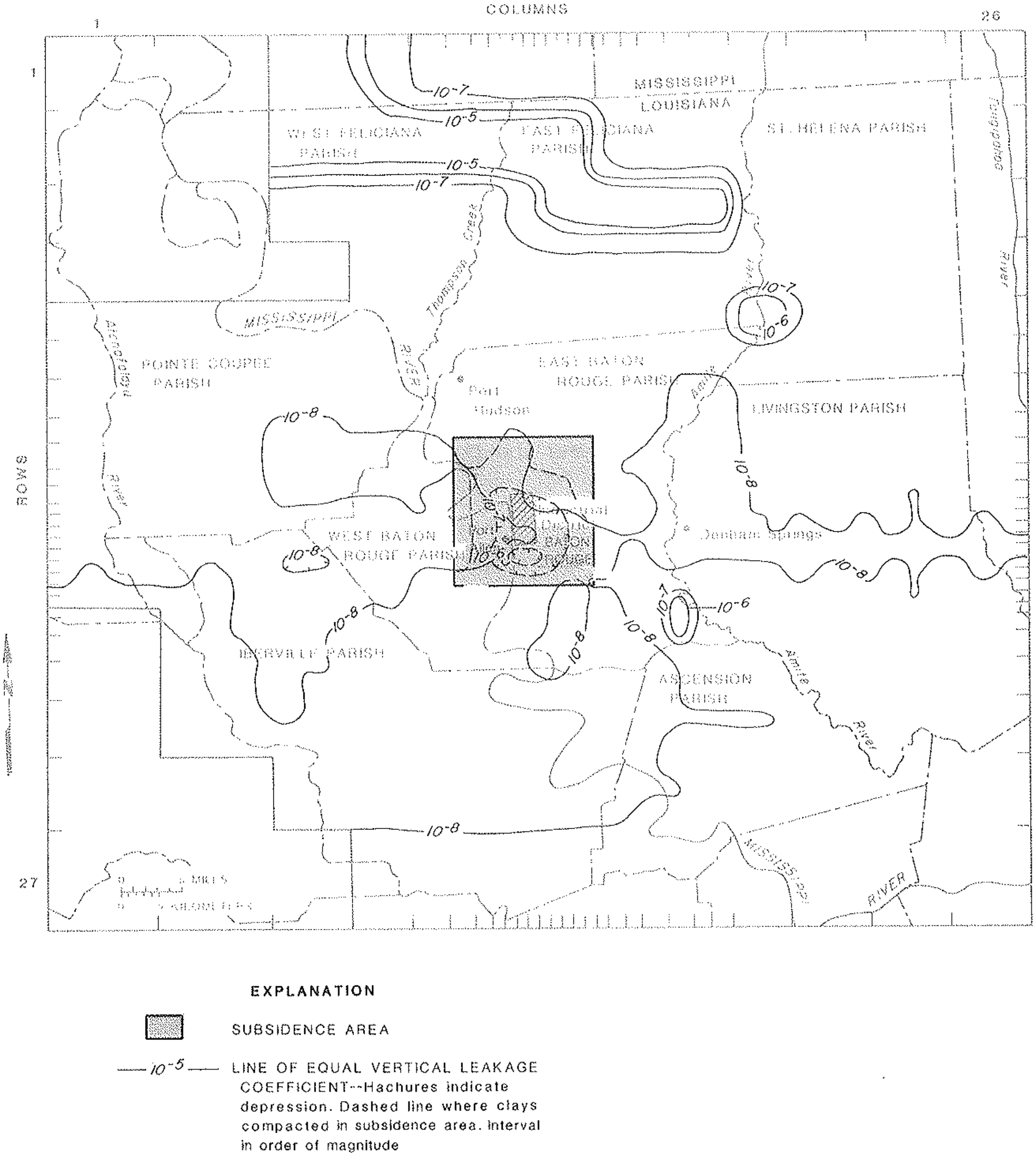


Figure 33.--Vertical leakage coefficient for confining bed 4.

used for all sensitivity simulations in order to determine the relative impact of error in any one parameter on the calibrated model response. Because the sensitivity analysis is normalized, the magnitude of the percent change in parameter is not critical. The analysis was performed on the simulation of May 1984. Both the May and October 1984 stress periods were examined with each of the four active layers contoured separately; of the resulting 128 contour maps, only selected ones are shown or discussed.

Transmissivity, storage coefficient, and vertical hydraulic conductivity for each active layer were increased by 1 percent, resulting in 12 sensitivity simulations. The sign of the contour, positive or negative, indicates whether water levels increased or decreased, respectively, given an increase in the parameter. Four additional sensitivity simulations were made. The transmissivity of the fault in layers 2 and 3, the "400-foot" and "600-foot" aquifers, was increased by 1 percent. The river leakage coefficient for the Mississippi River (layer 1) and for the rivers in the outcrop area of the "400-foot" aquifer (layer 2) were increased by 1 percent. Recharge inserted into the outcrop of the "400-foot" aquifer and the pumping rate from all aquifers were increased by 1 percent.

Results of Sensitivity Analysis

Table 5 provides a summary of the sensitivity analysis. The mean and mean absolute value of water-level changes from the calibrated water levels is tabulated for each active model layer for the May 1984 stress period. From this table it can be quickly ascertained that changes in transmissivity and pumping rate have more of an impact on water levels than changes in vertical hydraulic conductance and changes in storage coefficient. Changes in transmissivity of the fault in layers 2, 3, and 4 had less of an impact than changes in storage coefficient for each entire model layer. Changes in the river leakage coefficient and changes in the recharge rate had about the same order of magnitude of impact on each entire model layer as the changes in storage coefficient.

Layer 1, the Mississippi River alluvial aquifer, in this model is not sensitive to most parameter changes. This is mainly because of the water levels in layer 1 are constrained by simulated river stage of the Mississippi and Atchafalaya Rivers. The tabular summary (table 5) of the sensitivity analysis can be misleading, because of the spatial distribution of the changes in water level.

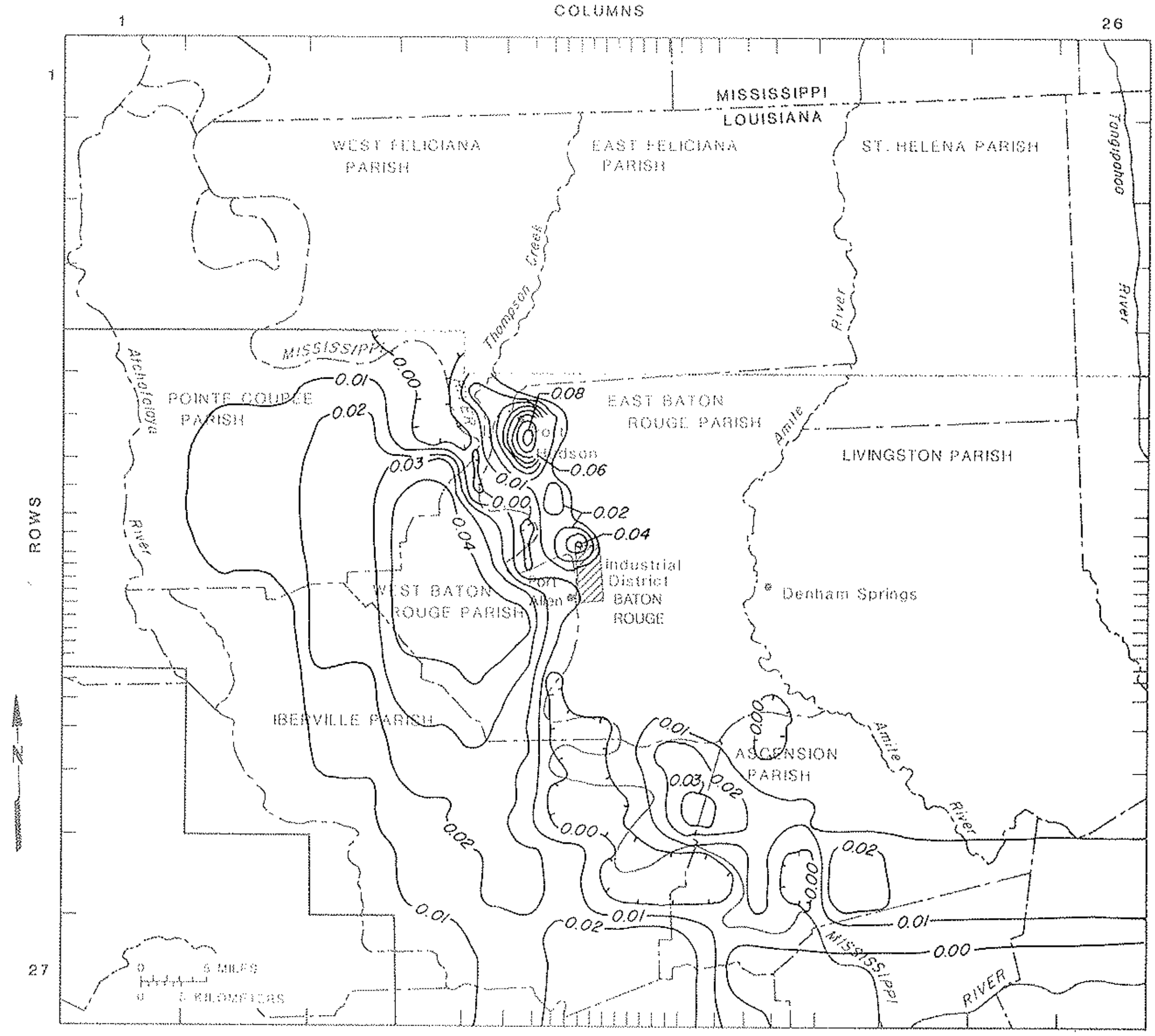
Transmissivity

A 1 percent increase in transmissivity of a layer produces the greatest change in model response for that layer, but also affects water levels in the surrounding layers. Figure 34 shows the change in water level for the Mississippi River alluvial aquifer (shallow Pleistocene sands also in layer 1 but treated as constant-head cells) for May 1984. An increase in transmissivity allows water to flow with less resistance through the aquifer, this produced a rise in water levels over much of the Mississippi River alluvial aquifer, when

Table 5.--Mean and mean absolute value of the change in water level from the calibrated model simulation for May 1984 when an input parameter or stress is increased 1 percent

[In feet]

Input parameter or stress	Layer 1 248 active model cells Mississippi River alluvial aquifer		Layer 2 684 active model cells "400-foot" aquifer		Layer 3 684 active model cells "600-foot" aquifer		Layer 4 684 active model cells "800-foot" aquifer	
	Mean	Mean absolute value	Mean	Mean absolute value	Mean	Mean absolute value	Mean	Mean absolute value
Transmissivity of layer 1........	0.164E-1	0.171E-1	0.616E-2	0.623E-2	0.430E-2	0.430E-2	0.334E-2	0.334E-2
Transmissivity of layer 2........	.177E-2	.199E-2	.283E-1	.355E-1	.256E-1	.319E-1	.228E-1	.278E-1
Transmissivity of layer 3........	.138E-2	.143E-2	.259E-1	.317E-1	.391E-1	.487E-1	.356E-1	.419E-1
Transmissivity of layer 4........	.299E-3	.314E-3	.396E-2	.440E-2	.600E-2	.728E-2	.138E-1	.213E-1
Vertical leakage of confining bed 1.	.616E-4	.635E-3	.351E-2	.217E-1	.241E-2	.206E-1	.227E-2	.199E-1
Vertical leakage of confining bed 2.	-.212E-3	.701E-3	-.146E-2	.354E-2	.179E-1	.188E-1	.151E-1	.156E-1
Vertical leakage of confining bed 3.	.105E-5	.667E-4	-.894E-3	.108E-3	-.166E-2	.310E-2	.172E-1	.247E-1
Vertical leakage of confining bed 4.	-.404E-4	.432E-4	-.978E-3	.104E-2	-.157E-2	.164E-2	-.819E-2	.837E-2
Storage coefficient of layer 1...	-.167E-1	.167E-1	-.583E-2	.583E-2	-.404E-2	.404E-2	-.298E-2	.298E-2
Storage coefficient of layer 2...	-.126E-2	.126E-2	-.603E-2	.625E-2	-.415E-2	.437E-2	-.263E-2	.305E-2
Storage coefficient of layer 3...	-.803E-3	.803E-3	-.413E-2	.424E-2	-.113E-1	.114E-1	-.849E-2	.873E-2
Storage coefficient of layer 4...	-.586E-3	.586E-3	-.598E-3	.246E-2	-.361E-2	.576E-2	-.812E-2	.139E-1
Fault transmissivity, layers 2, 3, and 4.	.688E-5	.188E-4	.767E-3	.149E-2	.444E-3	.166E-2	.372E-3	.171E-2
River leakage coefficient, layers 1 and 2.	.248E-2	.249E-2	-.402E-2	.570E-2	-.437E-2	.553E-2	-.403E-2	.484E-2
Recharge input layer 2 rows 1 to 5.	.273E-4	.273E-4	-.118E-1	.118E-1	-.120E-1	.120E-1	-.105E-1	.105E-1
Pumping rate, layers 2, 3, and 4.	-.305E-2	.305E-2	-.602E-1	.602E-1	-.793E-1	.793E-1	-.879E-1	.875E-1



EXPLANATION

—0.02— LINE OF EQUAL WATER-LEVEL CHANGE--Hachures indicate depression. Interval 0.01 foot

Figure 34.--Equal change in water levels simulated for the Mississippi River alluvial aquifer, model layer 1, when the transmissivity of layer 1 is increased 1 percent, May 1984.

river stage was rising. The greatest increase in water levels occurs above the pumping center near Port Hudson (row 7 column 6). There is a slight drop in water levels in layer 1 at some of the cells which contain reaches of the Mississippi River. The "400-foot" aquifer (not shown) had increases in water levels similar to layer 1. At Port Hudson the increase was only 0.04 ft. The "600-foot" aquifer (not shown) had water-level increases of only 0.01-0.02 ft. The "800-foot" aquifer (not shown) had water-level increases of 0.005-0.015 ft west of the Mississippi River and north of the Baton Rouge fault. During October 1984 the Mississippi River stage was falling and this produced a lowering of Mississippi River alluvial aquifer water levels in layer 1.

The increase in transmissivity of layer 2, the "400-foot" aquifer, resulted in the changes in water levels shown in figure 35. In the outcrop area (rows 1-5) the increased transmissivity allowed more water to flow back to the rivers resulting in a lowering of water levels of about 0.04 ft. The greatest change occurred at the pumping centers where water levels increased. Layer 3, the "600-foot" aquifer (not shown), had an identical response as the "400-foot" aquifers in the outcrop area where the two aquifers merge; but at the pumping centers, the increases in water levels for the "600-foot" aquifer were less. Layer 4, the "800-foot" (not shown), responds similarly to the "400-foot" aquifer but with less change than the "600-foot" aquifer. Layer 1, the Mississippi River alluvial aquifer (not shown), was insensitive to the increase in transmissivity of the "400-foot" aquifer. For the October 1984 stress period, water-level changes were similar to those for May.

The increase in transmissivity of layer 3, the "600-foot" aquifer, resulted in the water-level changes shown in figure 36. Decreases in water levels occurred in the outcrop area (rows 1-5) but these decreases were less than 0.04 ft. The maximum increase in water level of the "600-foot" aquifer was 0.4 ft, occurring at the Baton Rouge industrial district. The "400-foot" aquifer (not shown) had identical water-level changes where the aquifers merge (fig. 13) but only had a water-level increase of 0.2 ft in the industrial district. The "800-foot" aquifer (not shown) had a decrease in water levels in row 1 in response to the decrease of layer 3, but had larger increases in water levels in the industrial district, with a maximum increase of 0.25 ft. The Mississippi River alluvial aquifer was insensitive to the increase in transmissivity of the "600-foot" aquifer.

The change in water levels from the calibration simulation for May 1984 caused by the 1 percent increase in transmissivity of the "800-foot" aquifer, layer 4, is shown in figure 37. Water levels increased mostly in pumped cells. Water levels in the "600-foot" aquifer (not shown) increased by 0.06 ft in the industrial district. The "400-foot" aquifer (not shown) was less sensitive with a maximum increase of only 0.02 ft. The Mississippi River alluvial aquifer (not shown) was insensitive to the increased transmissivity of the "800-foot" aquifer.

Vertical hydraulic conductivity

The 1 percent increase in the vertical hydraulic conductivity of a confining bed often caused equal but opposite water-level changes in the two model layers above and below the confining bed. For example, the Mississippi

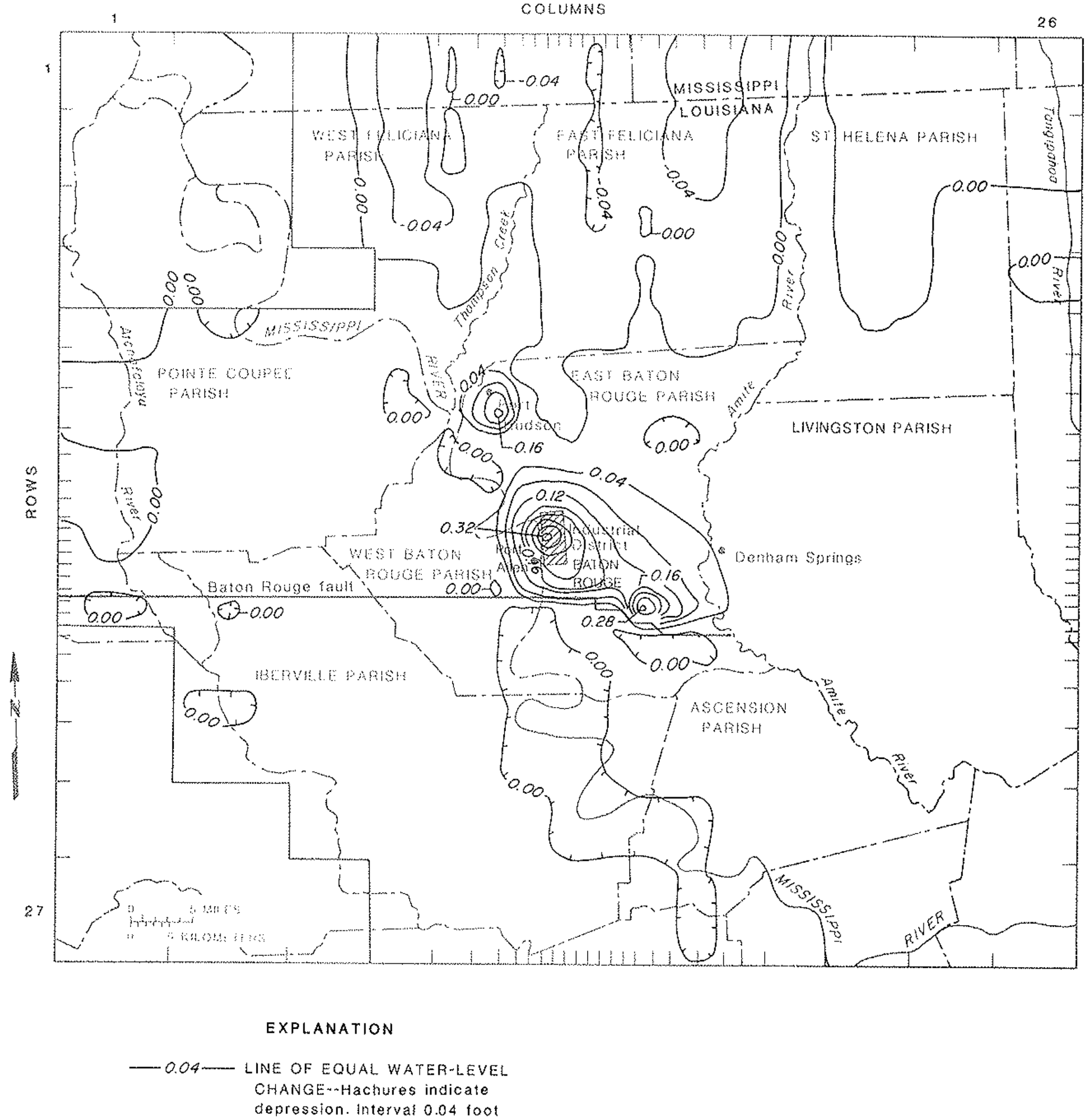
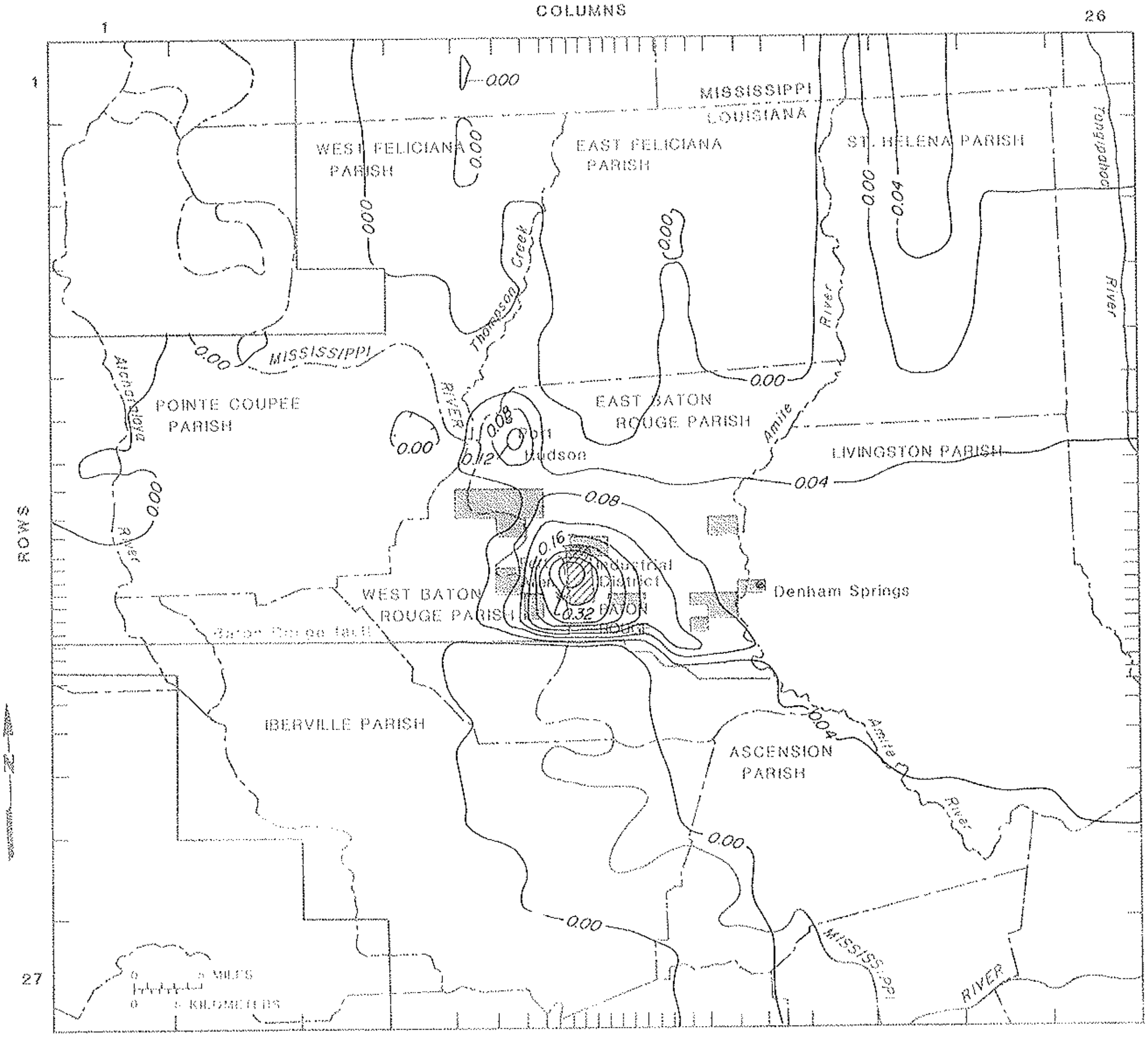


Figure 35.--Equal change in water levels simulated for the "400-foot" aquifer, model layer 2, when the transmissivity of layer 2 is increased 1 percent, May 1984.

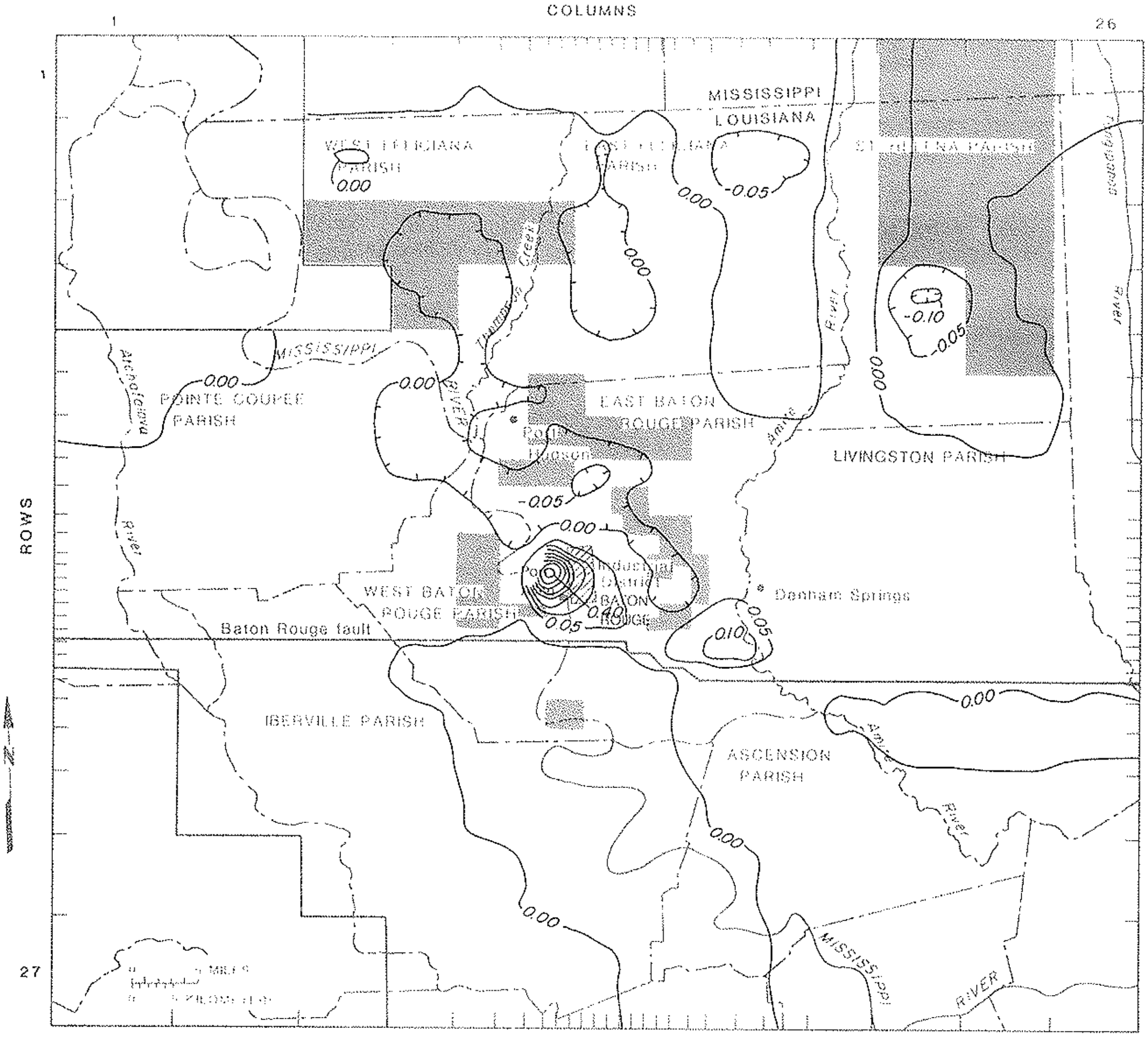


EXPLANATION

AQUIFER IS THIN OR ABSENT

0.04 LINE OF EQUAL WATER-LEVEL CHANGE--Hachures indicate depression. Interval 0.04 foot

Figure 36.--Equal change in water levels simulated for the "600-foot" aquifer, model layer 3, when the transmissivity of layer 3 is increased 1 percent, May 1984.



EXPLANATION

AQUIFER IS THIN OR ABSENT

0.05 — LINE OF EQUAL WATER-LEVEL CHANGE--Hachures indicate depression. Interval 0.05 foot

Figure 37.--Equal change in water levels simulated for the "800-foot" aquifer, model layer 4, when the transmissivity of layer 4 is increased 1 percent, May 1984.

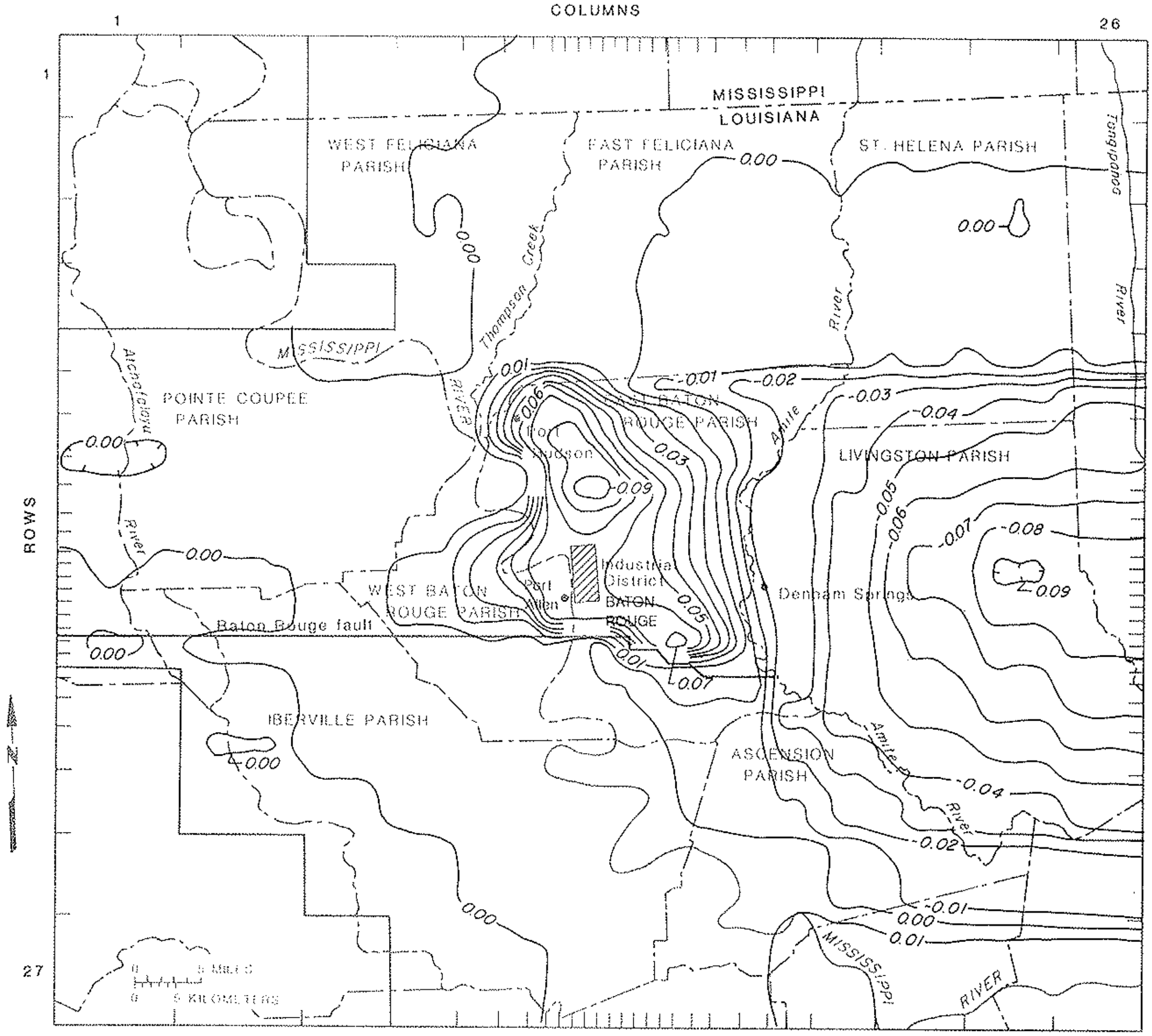
River alluvial aquifer, layer 1, was insensitive to the increased conductivity of the confining bed between it and the "400-foot" aquifer, layer 2, because of the influence of the Mississippi and Atchafalaya Rivers. The increase in confining-bed conductivity had a significant effect on water levels in all three layers beneath the top layer. Only the water-level changes in the "400-foot" aquifer, layer 2, are shown in figure 38. The "600-foot" aquifer, layer 3, had identical water-level changes as the "400-foot aquifer where they coalesce. In the eastern part of the modeled area where the aquifer system is not stressed significantly by pumpage, more water discharged through the confining bed which separates the "400-foot" aquifer from the shallow Pleistocene sands. This results in a reduction in water levels in the "400-, 600-, and 800-foot" aquifers in the southeast (fig. 38). Where the aquifer system is stressed by pumpage, more water leaks into the deeper aquifers through the top confining beds, causing water levels to rise in the deeper aquifers.

Figures 39 and 40 show the water-level changes from the calibrated model for the "400-foot" and "600-foot" aquifers, respectively, when the conductivity of the confining bed between the two aquifers was increased by 1 percent. Water levels in the "400-foot" aquifer decreased at the Baton Rouge industrial district by about 0.03 ft and increased slightly at the Port Hudson pumping center. The "600-foot" aquifer had significant water-level increases over most of the Mississippi River alluvial valley, especially at the Baton Rouge industrial district where the aquifer is pumped heavily and at the interconnection between the Mississippi River alluvial aquifer with the deeper aquifers (fig. 13). Decreases in water levels occurred at the two pumping centers where the "600-foot" aquifer was pumped less than the "400-foot" aquifer in May 1984. The "800-foot" aquifer (not shown) had water-level increases where the "800-foot" aquifer is merged with the "600-foot" aquifer (fig. 13). The Mississippi River alluvial aquifer was insensitive to the change in confining-bed conductivity between layers 2 and 3, the "400-foot" and "600-foot" aquifers.

The water-level change in the "800-foot" aquifer, layer 4, when the conductivity of the confining bed between the "600-foot" and "800-foot" aquifers was increased by 1 percent is shown in figure 41. Water levels in the "600-foot" aquifer (not shown) decreased slightly in much of the area outside the industrial district. Water levels rose in the "600-foot" aquifer at the industrial district where it is pumped less than the "800-foot" aquifer.

Most of the area where the "600-foot" aquifer merges with the "400-foot" aquifer, there is no difference in water levels in layers 2 and 3, the "400-foot" and "600-foot" aquifers. The "800-foot" aquifer had a decrease in water levels in the southeastern part of the model area (fig. 41) where more water was allowed to discharge upward through the confining bed. All layers were less sensitive to this particular change in parameter value.

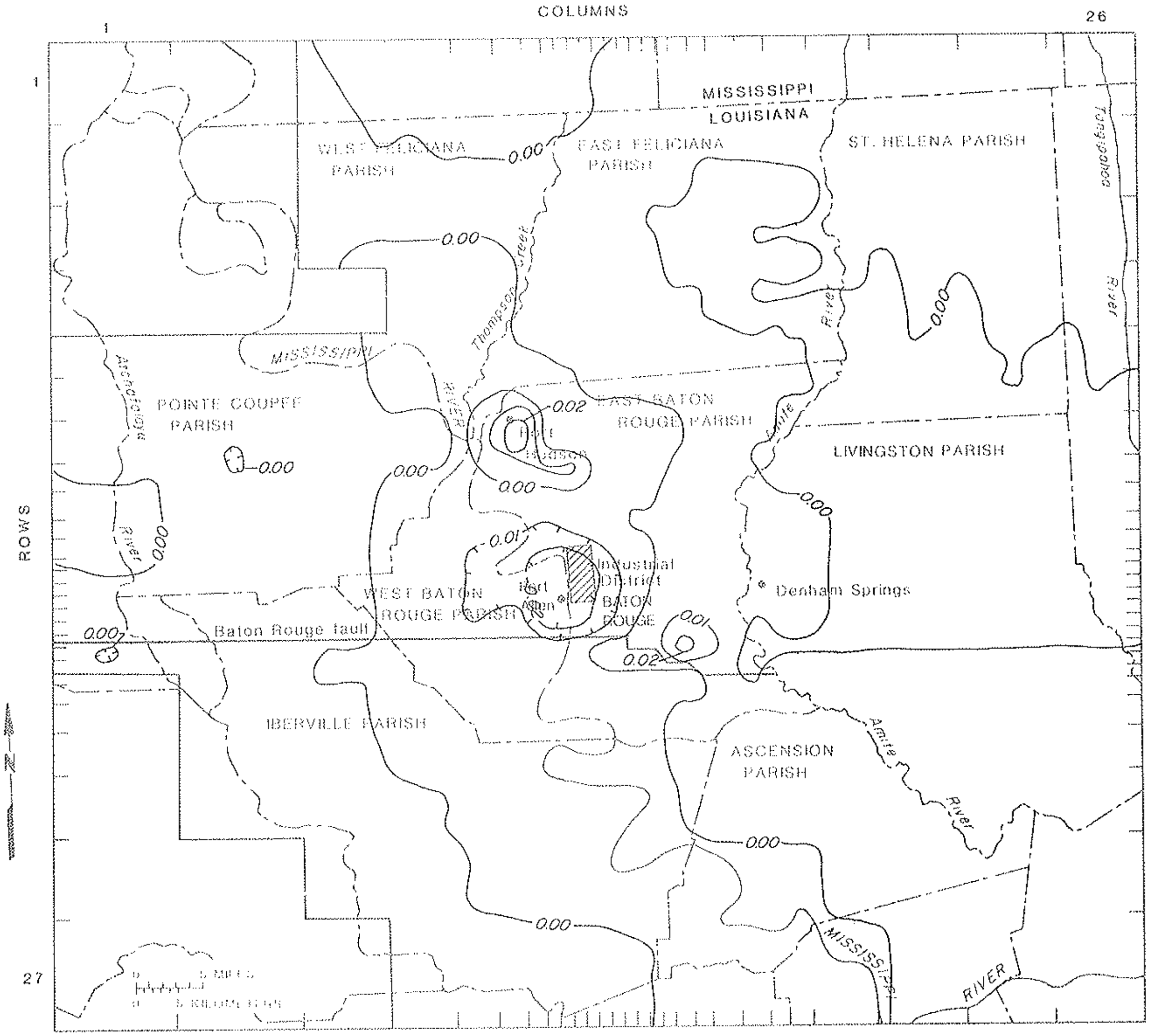
The water levels in all four active model layers were lowered by the 1 percent increase in the vertical hydraulic conductivity of confining bed 4 (fig. 14) between layer 4, the "800-foot" aquifer, and the specified heads in layer 5, the "1,000-foot" aquifer, but the lowering was small. (See table 5.) This is perhaps due to the fact that the "800-foot" and "1,000-foot" aquifers are not heavily stressed by pumpage in the Baton Rouge area (fig. 9). Also, the confinement between the "800-foot" and "1,000-foot" aquifers is greater than between the shallower aquifers.



EXPLANATION

—— 0.02 —— LINE OF EQUAL WATER-LEVEL CHANGE--Hachures indicate depression. Interval 0.01 foot

Figure 38.--Equal change in water levels simulated for the "400-foot" aquifer, model layer 2, when the vertical hydraulic conductance of confining bed 1 is increased 1 percent, May 1984.



EXPLANATION

—— 0.02 —— LINE OF EQUAL WATER-LEVEL CHANGE--Hachures indicate depression. Interval 0.01 foot

Figure 39.--Equal change in water levels simulated for the "400-foot" aquifer, model layer 2, when the vertical hydraulic conductance of confining bed 2 is increased 1 percent, May 1984.

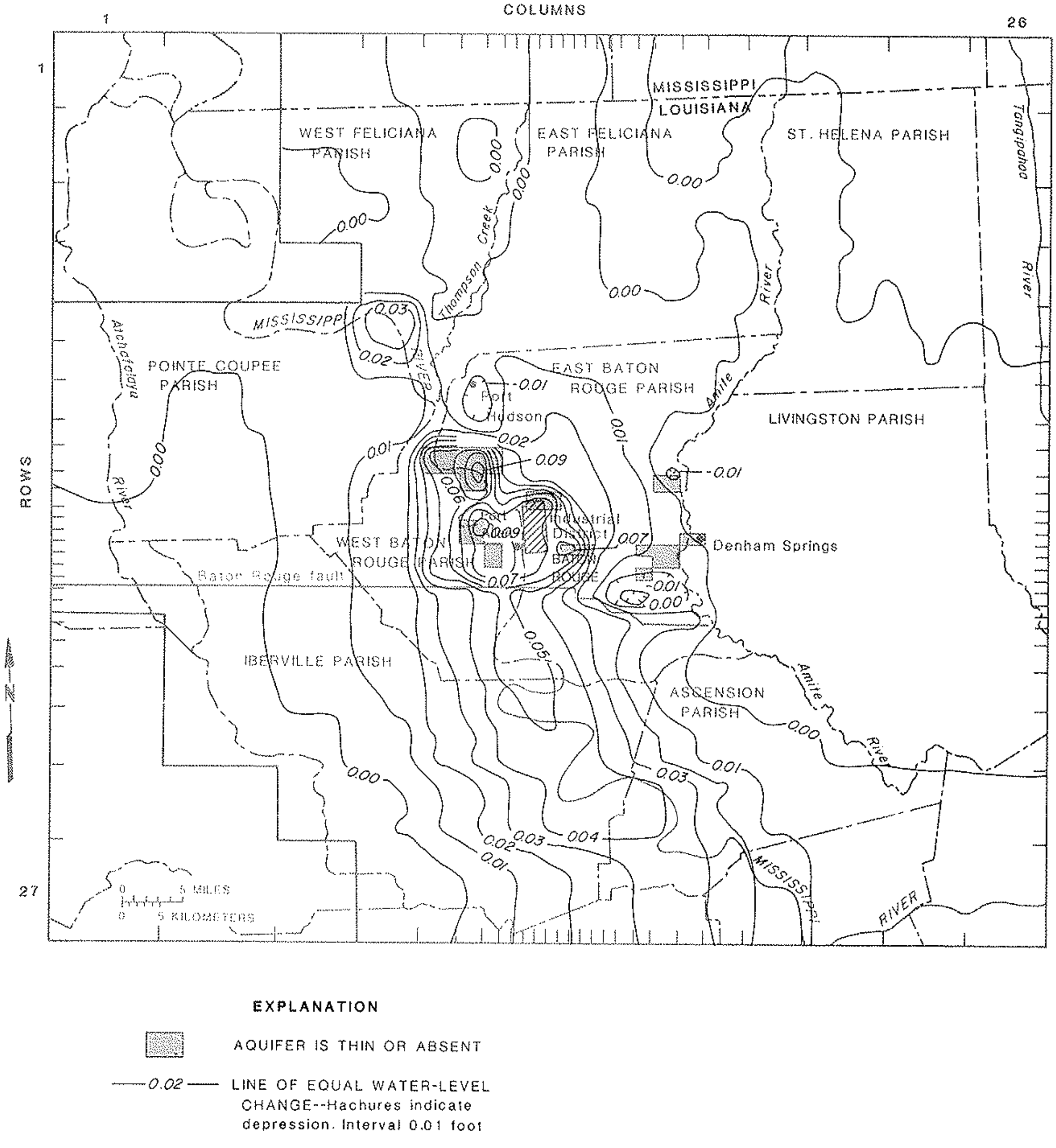
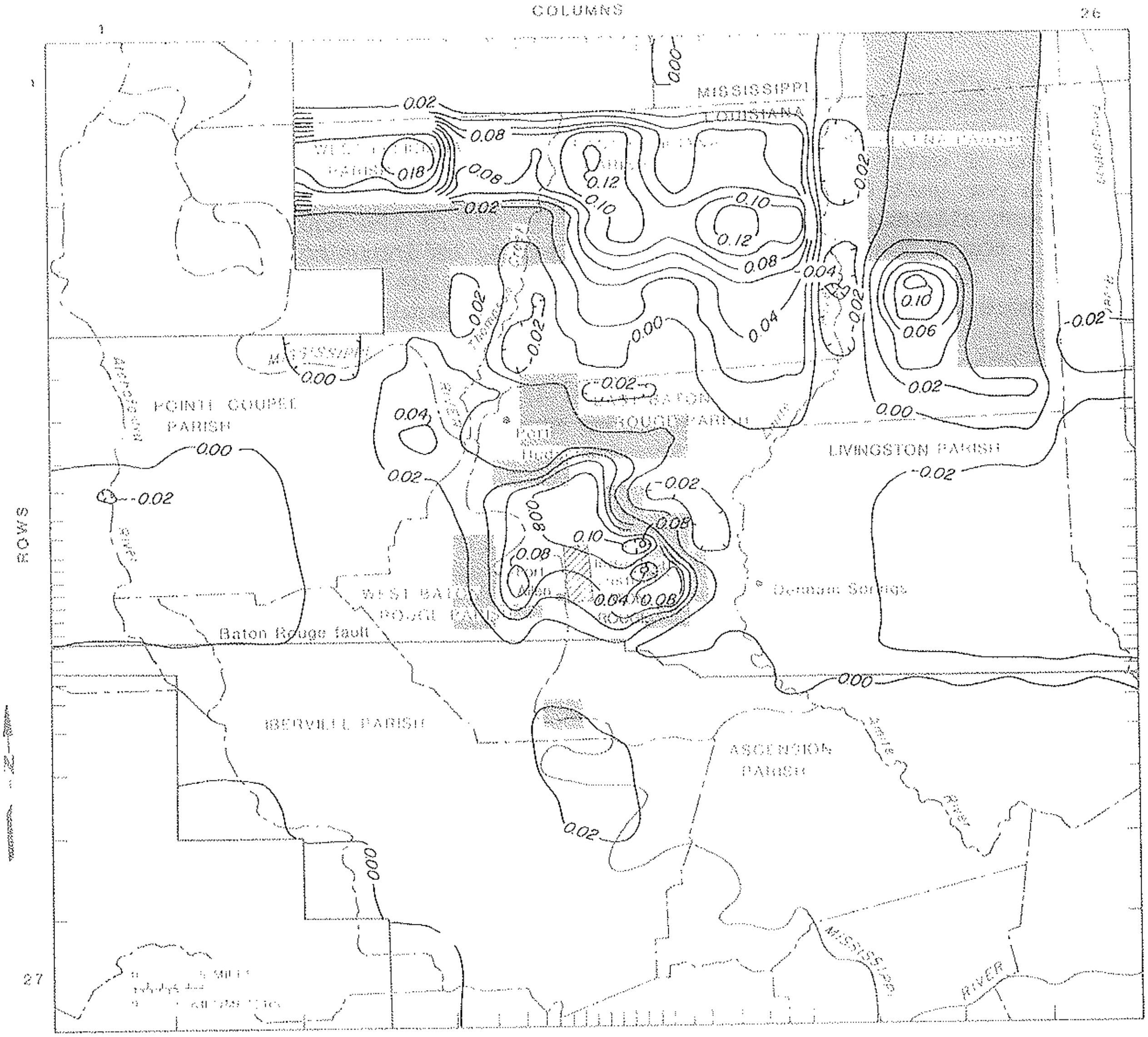


Figure 40.--Equal change in water levels simulated for the "600-foot" aquifer, model layer 3, when the vertical hydraulic conductance of confining bed 2 is increased 1 percent, May 1984.



EXPLANATION

AQUIFER IS THIN OR ABSENT

0.04 LINE OF EQUAL WATER-LEVEL CHANGE--Hachures indicate depression. Interval 0.02 foot

Figure 41.--Equal change in water levels simulated for the "800-foot" aquifer, model layer 4, when the vertical hydraulic conductance of confining bed 3 is increased 1 percent, May 1984.

Storage coefficient

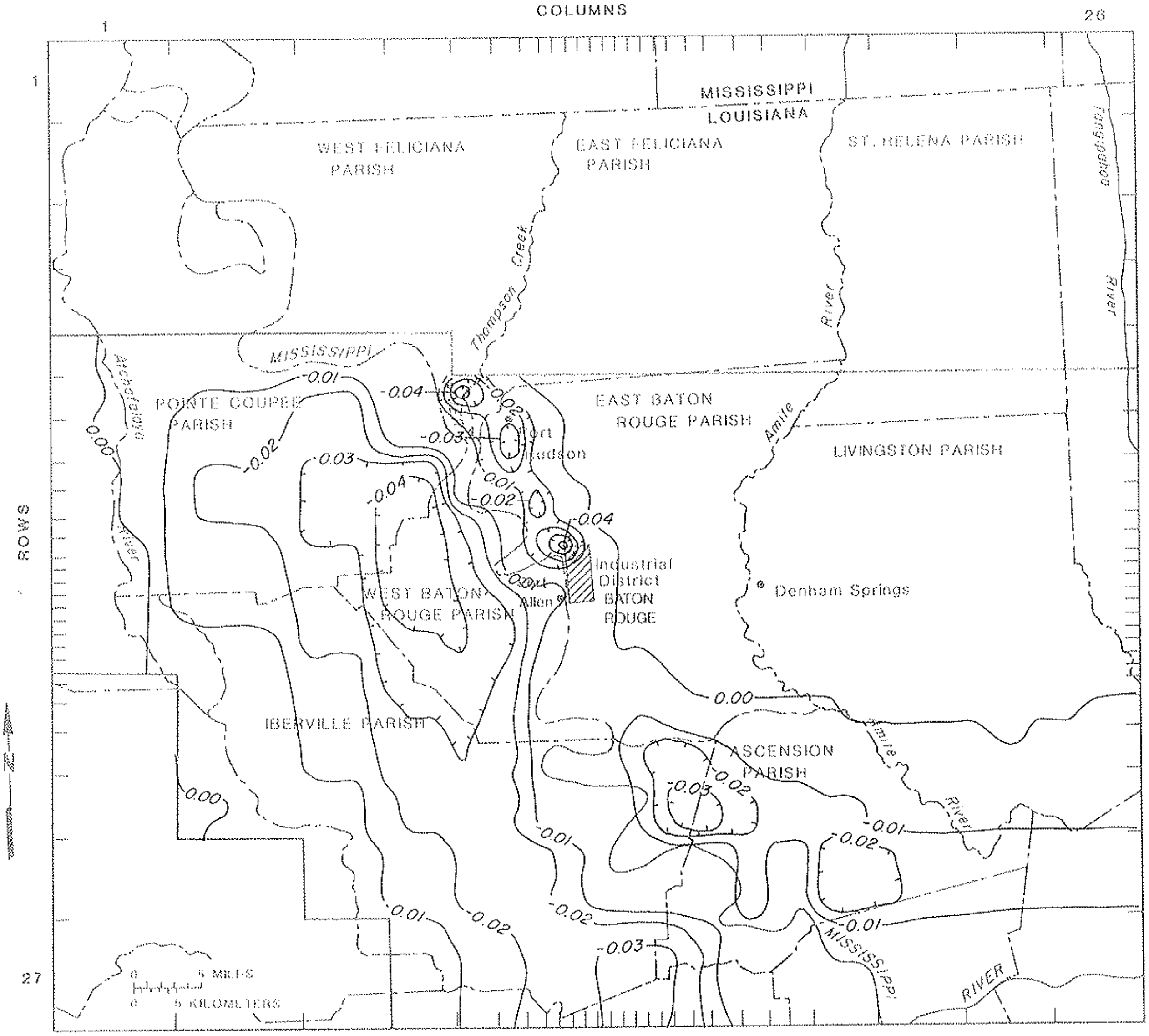
The 1 percent increase in storage coefficients for each layer results in very different water-level changes for the May and October 1984 stress periods. The storage coefficient is a dimensionless parameter representing the volume of water which can go into or out of storage from the aquifer per unit area per unit change in head (water level). An increase in the storage coefficient allows more water to go into or out of the aquifer given a change in head. For a change in stress (for example, an increase in pumpage) water levels will not drop as rapidly with an increase in the storage coefficient because more water can come from storage in the aquifer.

The change in water level in layer 1, the Mississippi River alluvial aquifer, for May 1984 is shown in figure 42 for the increased storage coefficient. At this time, Mississippi River stage is rising. Because of the increase in storage capacity of the aquifer, water levels in the alluvial aquifer are lower for this simulation than for the calibration simulation; more water is required to fill the increase in storage capacity of the aquifer. In October 1984, river stage is declining. Water levels are higher than in the calibrated simulation for layer 1 in October (fig. 43) because more water can come from storage to go back to the river. Because of the large areas where the Mississippi River alluvial aquifer and the "400-foot" aquifer are merged, water-level changes are similar to layer 1 for layer 2, the "400-foot" aquifer in May and October 1984. Layers 3 and 4, the "600-foot" and "800-foot" aquifers, were affected but to a lesser degree than layers 1 and 2 by the increase in storage coefficient of layer 1.

The change in water levels for the 1 percent increase in storage coefficient for the "400-foot" aquifer, layer 2, for May 1984 is shown in figure 44. The decreases in water levels at the center of figure 44 are related to the change in pumpage from April to May (fig. 25). The "600-foot" aquifer (not shown) also responded to the changes in water levels in the "400-foot" aquifer as did the "800-foot" aquifer, but with smaller changes in water level. The Mississippi River alluvial aquifer was insensitive to the change in storage coefficient. For the October stress period, water levels are higher in the "400-foot" aquifer than in the calibration simulation. In October when river stage is declining, a larger part of the water discharging from the "400-foot" aquifer to the Mississippi River alluvial aquifer comes from the increase in storage capacity of layer 2, thus, the higher water levels.

The increased storage coefficient of the "600-foot" aquifer, layer 3, resulted in the largest water-level change in cells with pumpage for both the May and October 1984 stress periods. Figure 45 shows the change in water levels for May. The "800-foot" aquifer responded more to these changes than the "400-foot" aquifer, and the Mississippi River alluvial aquifer was insensitive to the change.

Figure 46 shows the change in water levels in the "800-foot" aquifer, layer 4, when the storage coefficient for the aquifer was increased 1 percent for the May 1984 stress period. For the October 1984 stress period, similar water-level changes occurred in the "800-foot" aquifer.



EXPLANATION

—— 0.02 —— LINE OF EQUAL WATER-LEVEL CHANGE--Hachures indicate depression. Interval 0.01 foot

Figure 42.--Equal change in water levels simulated for the Mississippi River alluvial aquifer, model layer 1, when the storage coefficient of layer 1 is increased 1 percent, May 1984.

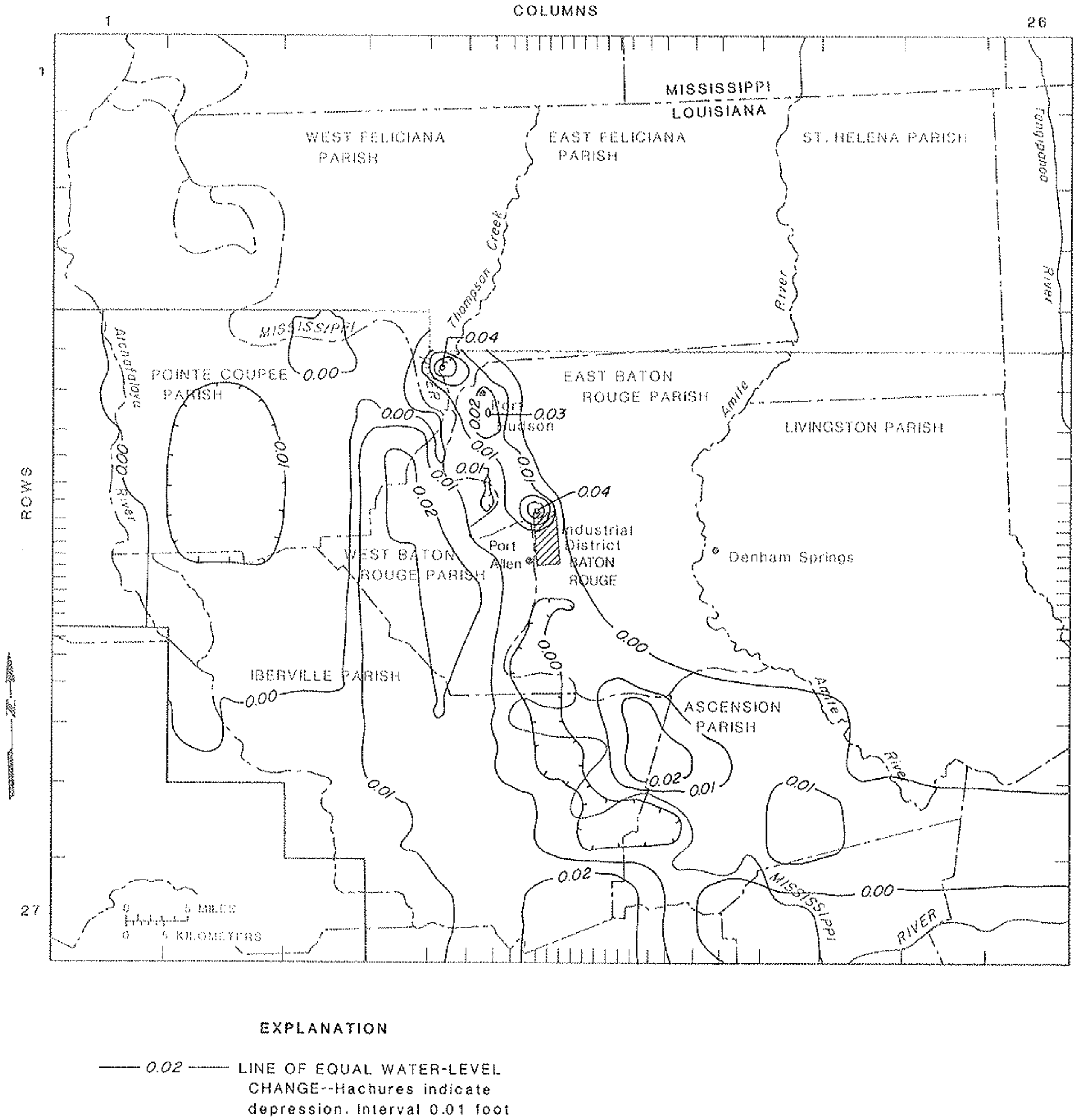


Figure 43.--Equal change in water levels simulated for the Mississippi River alluvial aquifer, model layer 1, when the storage coefficient of layer 1 is increased 1 percent, October 1984.

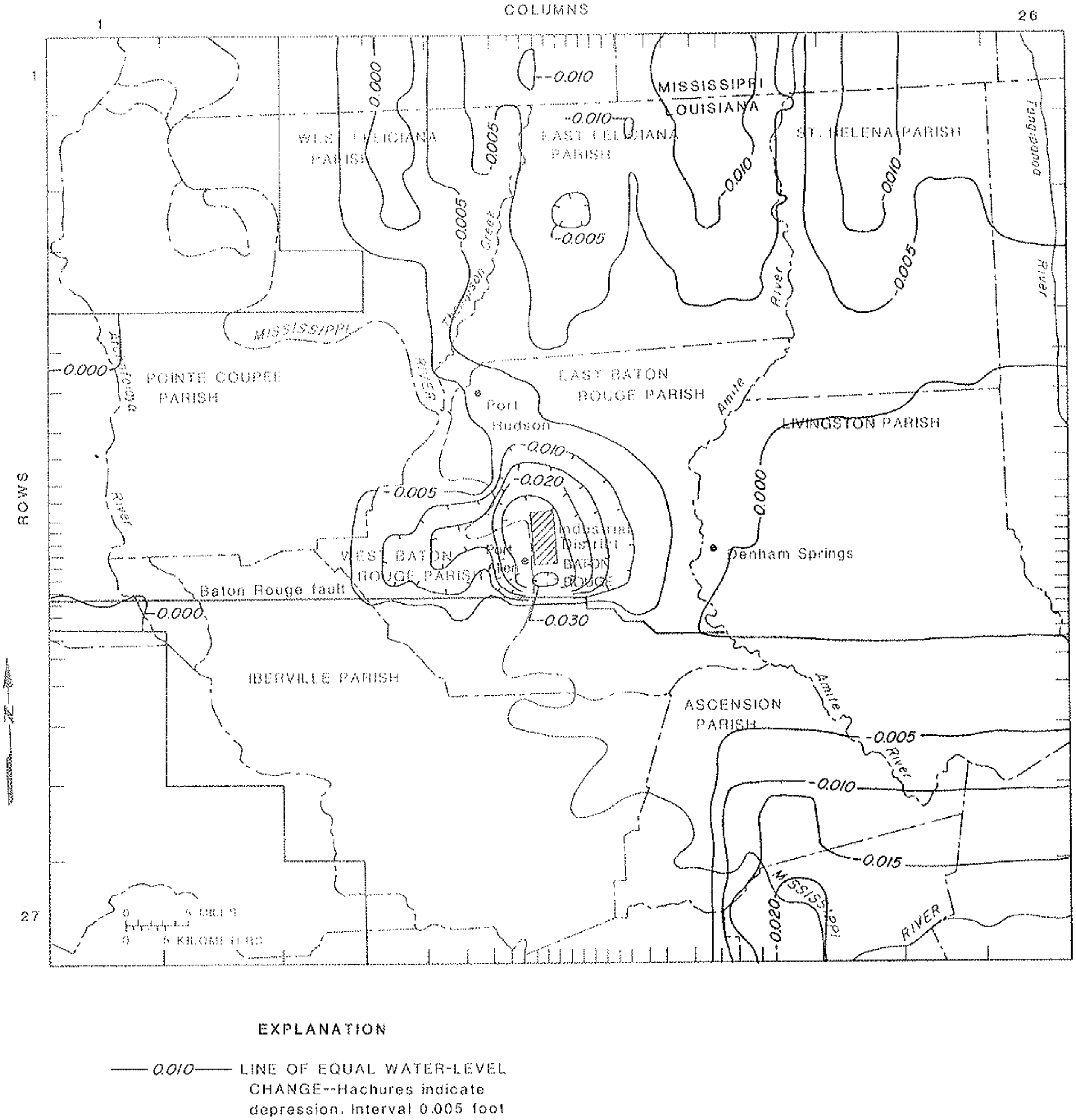


Figure 44.--Equal change in water levels simulated for the "400-foot" aquifer, model layer 2, when the storage coefficient of layer 2 is increased 1 percent, May 1984.

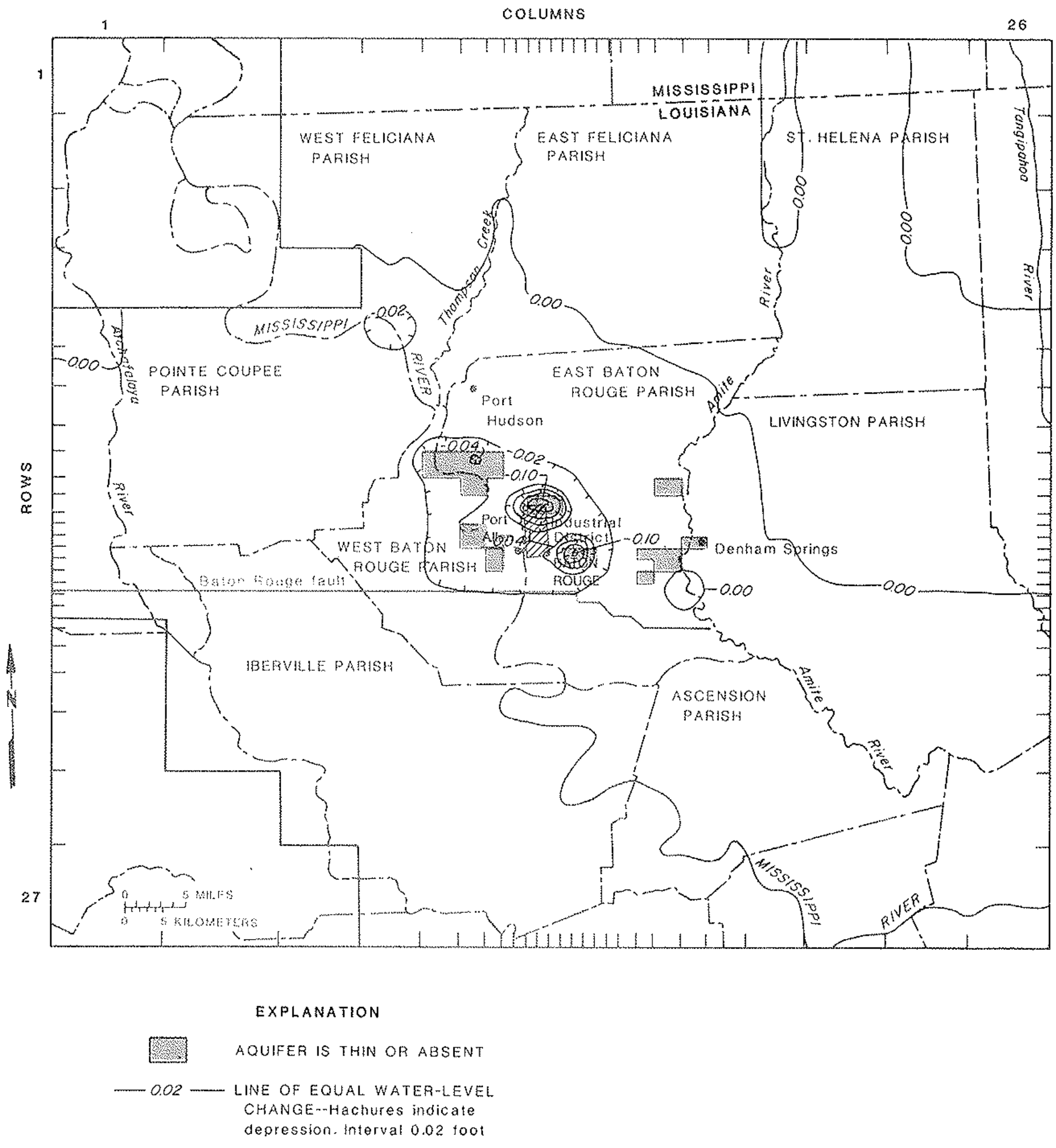
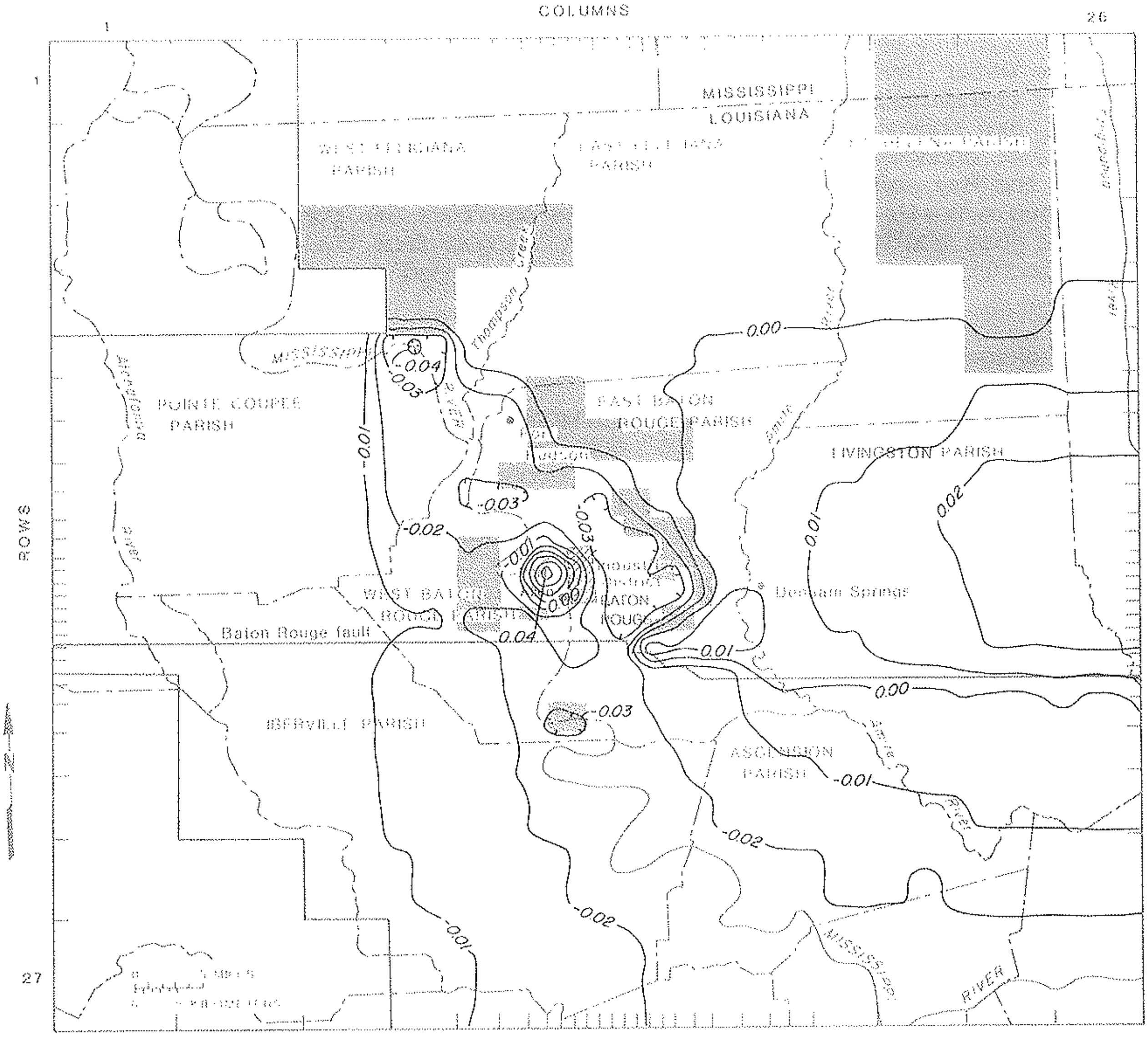


Figure 45.--Equal change in water levels simulated for the "600-foot" aquifer, model layer 3, when the storage coefficient of layer 3 is increased 1 percent, May 1984.



EXPLANATION

AQUIFER IS THIN OR ABSENT

—— *0.01* —— LINE OF EQUAL WATER-LEVEL CHANGE--Hachures indicate depression. Interval 0.01 foot

Figure 46.--Equal change in water levels simulated for the "800-foot" aquifer, model layer 4, when the storage coefficient of layer 4 is increased 1 percent, May 1984.

Transmissivity of the Baton Rouge fault

The water-level changes, which occurred for the May 1984 stress period when the transmissivity of the fault in layers 2, 3, and 4 was increased by 1 percent, are shown for the "400-, 600-, and 800-foot" aquifers in figure 47. Water-level changes in the Mississippi River alluvial aquifer were negligible. In both the May and October stress periods, there is a slight increase in water levels north of the fault where the aquifers are pumped and a slight decrease just south of the fault. These changes from the calibrated model are small, occurring only where the aquifers are pumped near the fault.

River leakage coefficient

Figure 48 shows the water-level changes for May 1984 that occurred in both layers 2 and 3, the "400-foot" and "600-foot" aquifers, when the river leakage coefficient was increased by 1 percent. This resulted in reductions in water levels in the outcrop area (rows 1-5) of these aquifers with the maximum reduction at the rivers (fig. 12). This is because of the increased ability for water to flow into the river from the aquifers. The "800-foot" aquifer (not shown) also showed a reduction in water levels but to a lesser degree because it merges with the shallower aquifer along the top row only. Water-level changes in the Mississippi River alluvial aquifer were negligible.

Recharge

Figure 49 shows the water-level changes that occurred in layers 2 and 3 when recharge to layer 2 was increased by 1 percent for the May 1984 stress period. This results in water level increases with maximum increases between rivers (fig. 12). A similar but lesser water-level rise occurred in the "800-foot" aquifer (not shown). The Mississippi River alluvial aquifer was insensitive to this change.

Pumpage

The effect on water levels from a 1 percent increase in pumpage is shown for all model layers in figure 50. It is apparent that errors in pumpage would cause the greatest localized error in water levels of all of the model input data. It is also important to note that water levels in layer 1, although it is not pumped, are affected by the increased pumpage in layer 2.

Limitations of the Model

The major limitations of this model of the Baton Rouge area are the inability to mathematically approximate all of the physical processes that have an impact on the geohydrology. Transient leakage of water into or out of storage from the confining beds cannot efficiently be simulated with this finite-difference model code. Subsidence and the resulting changes in the properties of the confining beds caused by compaction of the clays are not

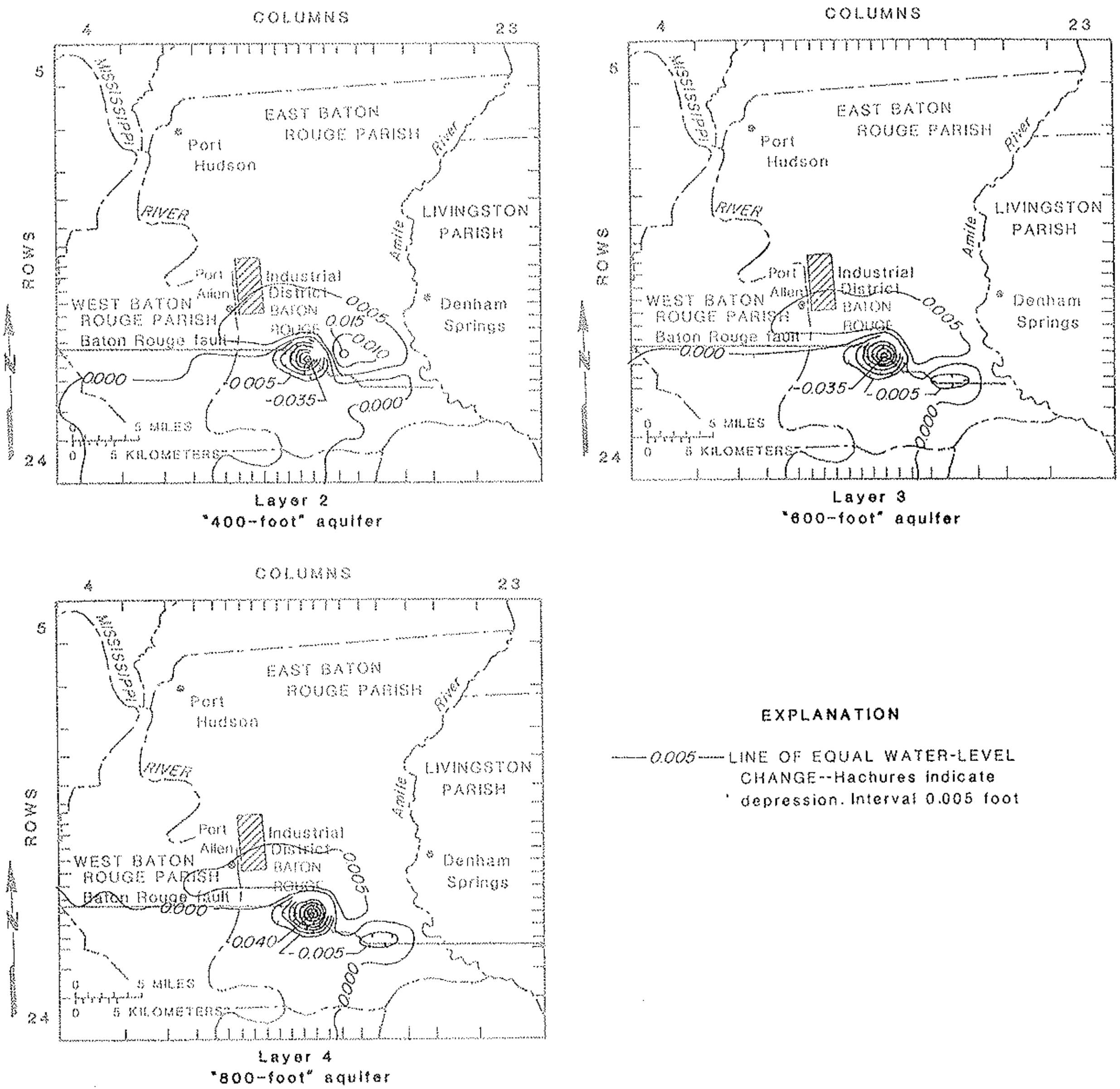


Figure 47.--Equal change in water levels simulated for the "400-, 600-, and 800-foot" aquifers (layers 2, 3, and 4, respectively), when the transmissivity of the fault is increased 1 percent, May 1984.

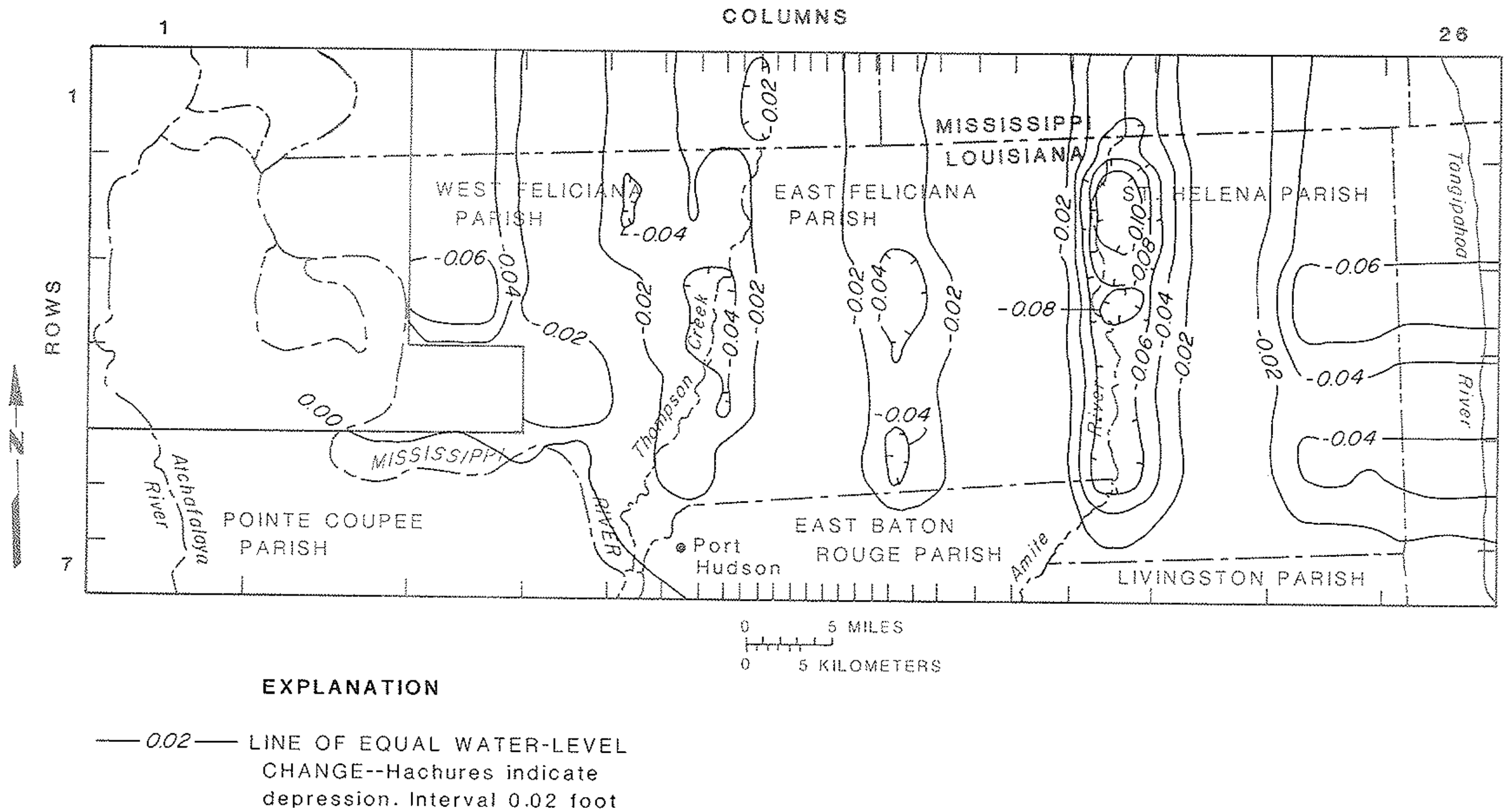
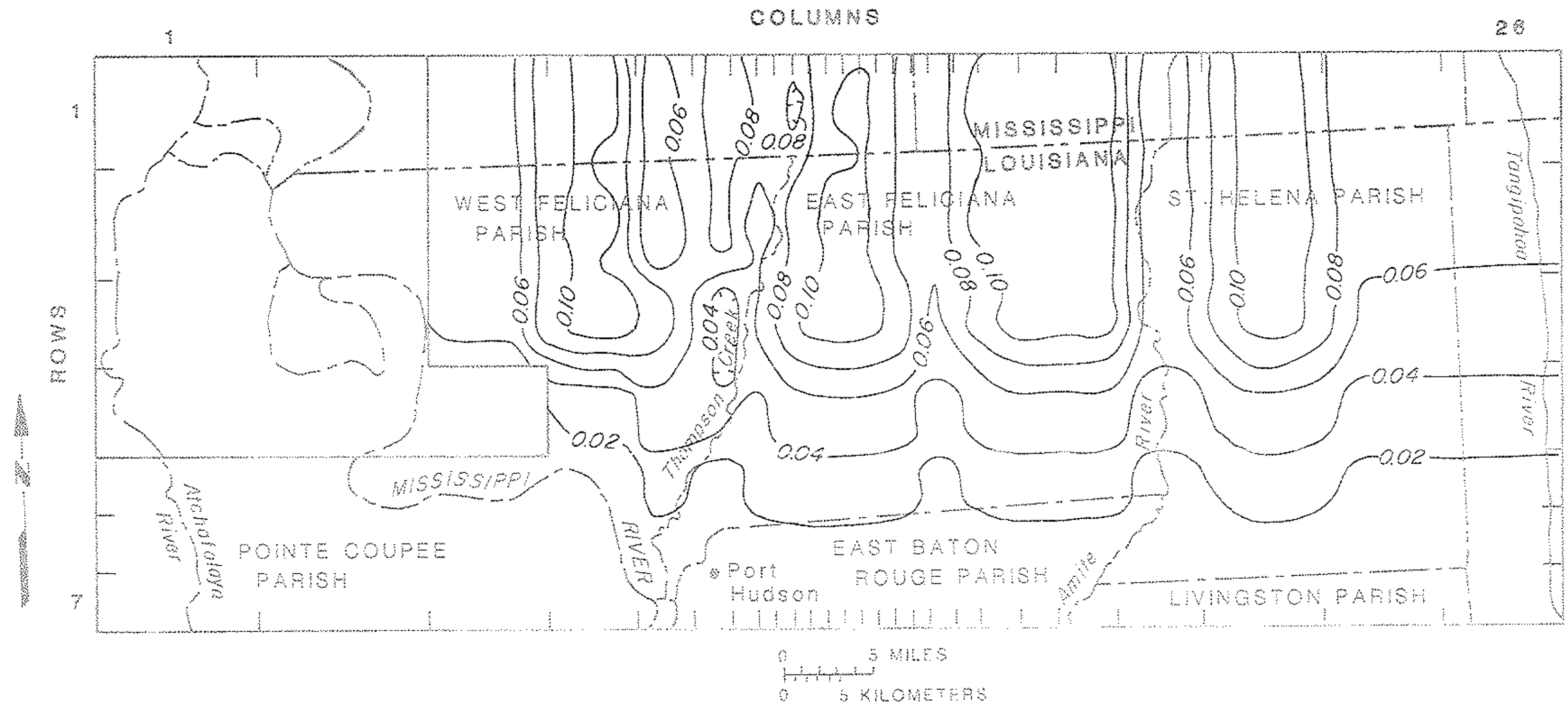


Figure 48.--Equal change in water levels simulated for the "400-foot" and "600-foot" aquifers (layers 2 and 3), when river leakage coefficients are increased 1 percent, May 1984.



EXPLANATION

—— 0.04 —— LINE OF EQUAL WATER-LEVEL CHANGE--Hachures indicate depression. Interval 0.02 foot

Figure 49.--Equal change in water levels simulated for the "400-foot" and "600-foot" aquifers (layers 2 and 3), when recharge is increased 1 percent, May 1984.

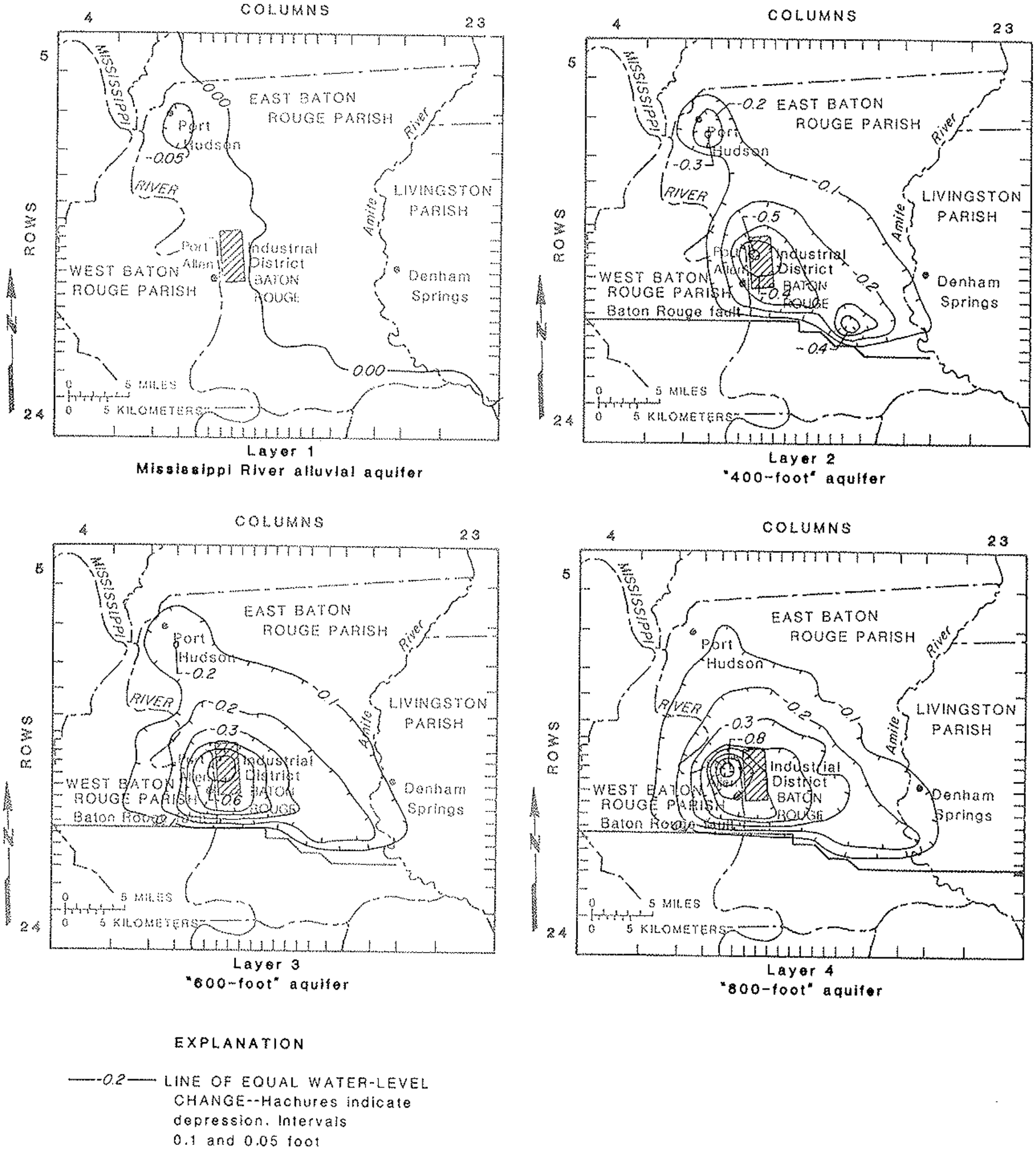


Figure 50.--Equal change in water levels simulated for the Mississippi River alluvial aquifer and the "400-, 600-, and 800-foot" aquifers (layers 1, 2, 3, and 4, respectively), when pumpage is increased 1 percent, May 1984.

incorporated into the finite-difference model code. It is not within the scope of this project to develop mathematical algorithms that account for these processes; thus, these problems are accounted for by analytical estimation of the volume of water that would have come out of the confining beds during the initial major ground-water decline in 1940 and an educated guess at the maximum change in the vertical hydraulic conductivity of the confining beds at the area of maximum subsidence.

Transient leakage in the confining beds do not present as critical a problem in simulating the long-term declines in water levels brought about by increased pumpage as do changes in the properties of the confining beds caused by compaction of the clays during subsidence. Subsidence resulting from declines in artesian head presents far greater problems. The resulting permanent changes in the properties of the confining bed are difficult to estimate. From the sensitivity simulations of change in vertical hydraulic conductivity of the confining beds, it can be noted that changes do have a significant impact on simulated water levels. (Refer to figs. 38-41 and table 5.) Therefore, predicted water-level declines from pumpage in previously unstressed areas may be erroneous.

ANALYSIS OF SIMULATIONS OF THE "400-FOOT" AND "600-FOOT" AQUIFERS

Four steady-state simulations were made with the model. All four simulations had mean Mississippi River stage input to the river cells in model layer 1. A predevelopment simulation was made with the model using the confining-bed properties of noncompacted clays shown by the solid lines in figures 30-33. Two additional simulations were made with the noncompacted confining beds. The locations of pumpage for these two hypothetical simulations are shown in figure 51. One of these hypothetical simulations has 50 Mgal/d of pumpage the other 100 Mgal/d. The fourth steady-state simulation was made using the confining-bed properties with compacted clays at the area of subsidence shown by the dashed lines in figures 30-33. The locations of pumpage for this fourth hypothetical simulation are shown in figure 52. The last simulation had the pumping rate from wells in the model area for 1985 increased 20 percent plus new public-supply wells pumping at maximum capacity. Potentiometric maps of the "400-foot" and "600-foot" aquifers are shown for these steady-state simulations.

The two transient calibration simulations and the steady-state simulations of the undeveloped aquifers provide information on changes in the ground-water flow system. The stress period of August 1944 is analyzed to show quantities of water flowing from the outcrop area, net flux to or from the Mississippi River, and vertical flux through confining bed 1. The 12 stress periods of 1984 are analyzed for net flow in the outcrop area and net flux to or from the Mississippi River. Net flux through confining bed 1 is shown for May and October 1984.

Potentiometric-Surface Maps from the Steady-State Simulations

The predevelopment potentiometric-surface maps indicate that there is relatively little difference in water levels between the "400-foot" and "600-foot" aquifers (pl. 4). In the outcrop area water levels were identical.

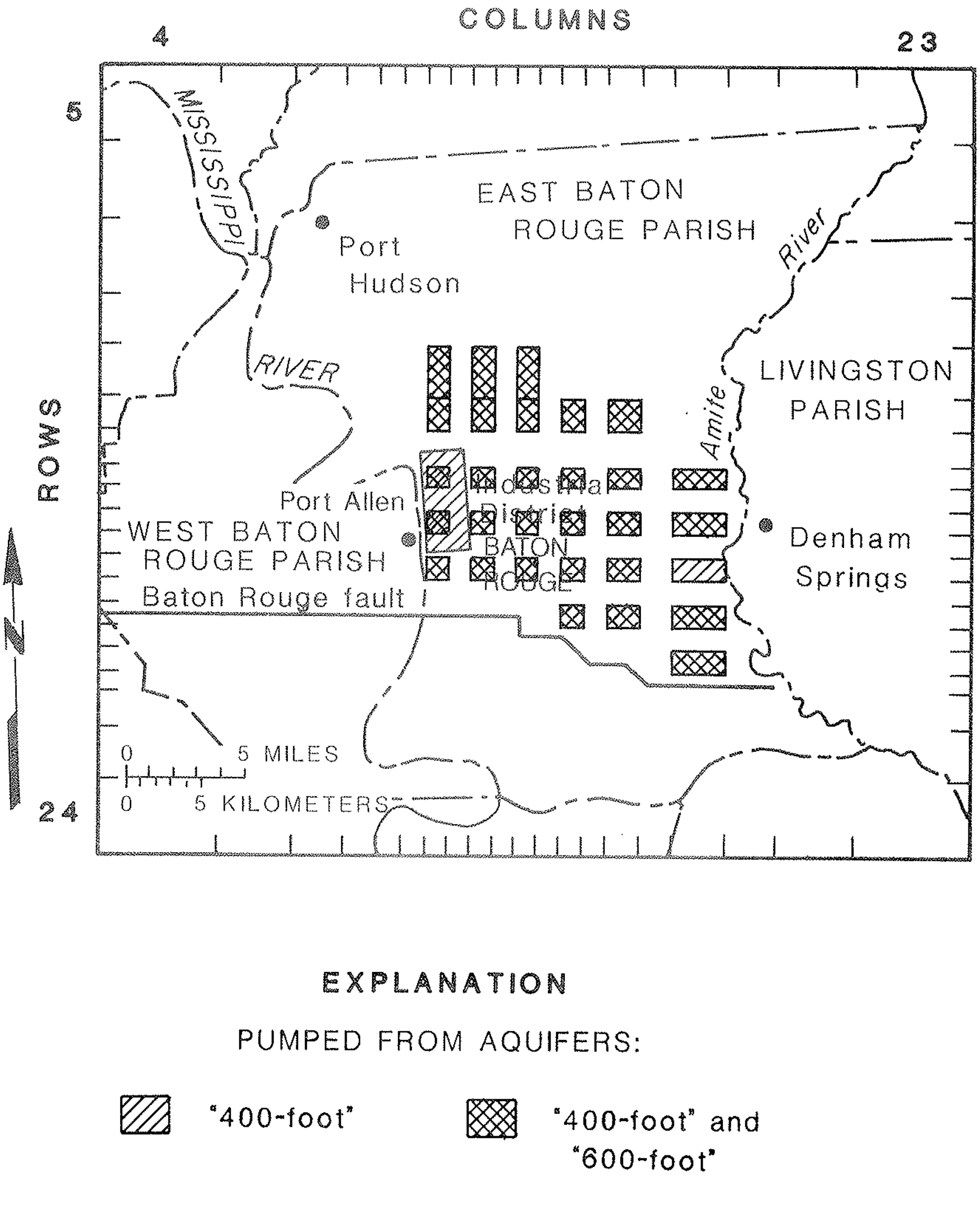


Figure 51.--Location of finite-difference cells with pumping stress for two alternative simulations, 50 and 100 million gallons per day.

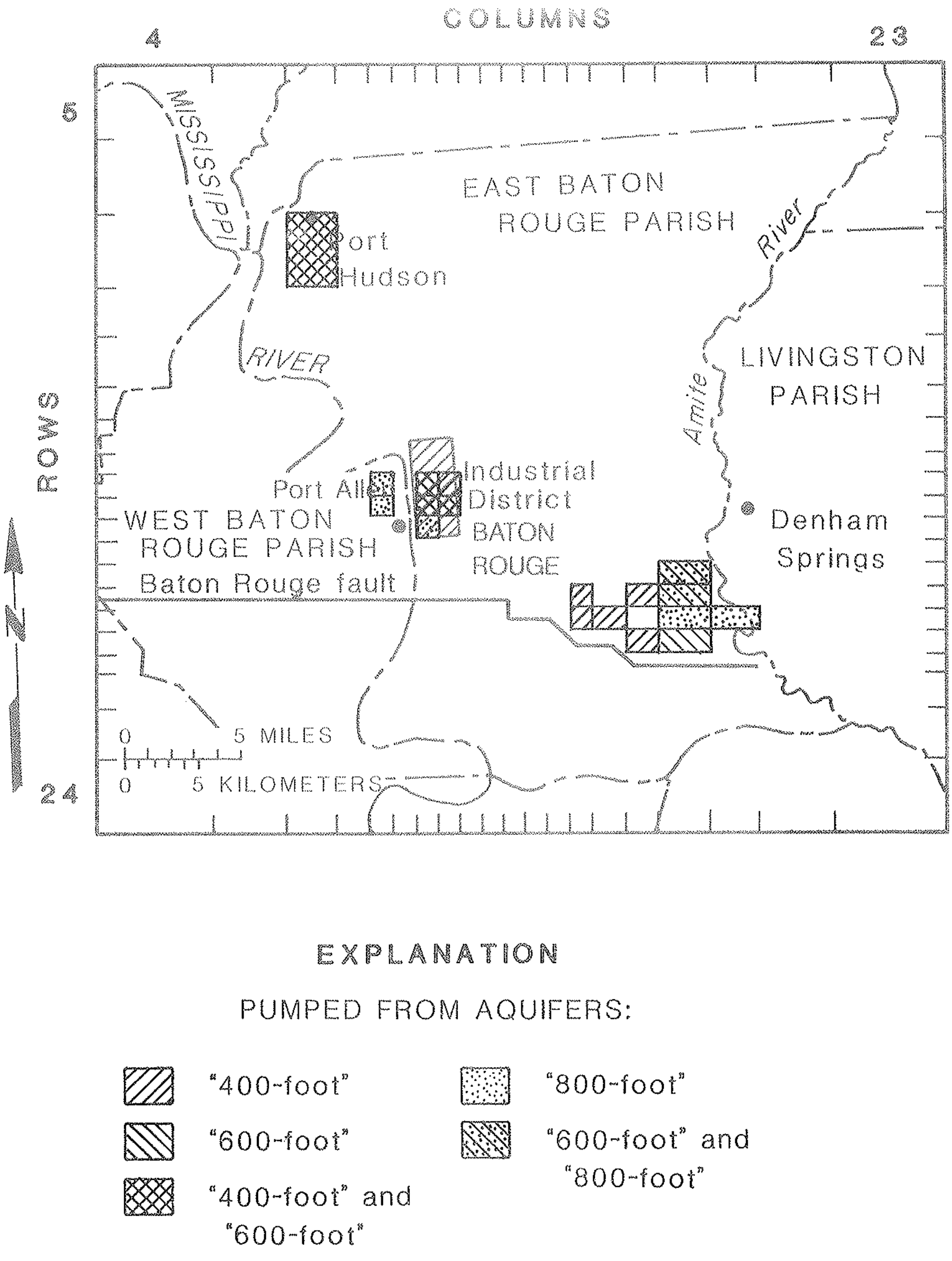


Figure 52.--Location of finite-difference cells with pumping stress for the alternative simulation of all wells pumped in 1985 and new wells pumped at full yield.

There is a 10- to 12-ft difference in head between the "400-foot" and "600-foot" aquifers near the western edge of the Pleistocene terraces near the Mississippi River, with higher water levels in the "600-foot" aquifer.

The steady-state predevelopment simulation indicates an upward flow of water from the aquifers throughout most of the modeled area before pumping began. The "400-foot" aquifer near the Mississippi River has water levels close to the average annual river stage.

The simulations with pumpage distributed over the entire Baton Rouge metropolitan area are hypothetical (fig. 51). These simulations are made to demonstrate how distributing pumping sources over a large area can reduce maximum drawdowns. During August 1944, maximum simulated drawdowns of about 200 ft below sea level resulted from pumping 36 Mgal/d (actual pumping rate 45 Mgal/d) from wells in the "400-foot" and "600-foot" aquifers located in the Baton Rouge industrial district. (Refer to pl. 1, which shows the simulated potentiometric surface, August 1944, and fig. 15, which shows the locations of pumpage.) With pumpage distributed over the metropolitan area (fig. 51), a pumping rate of 50 Mgal/d resulted in maximum simulated drawdown of about 100 ft below sea level and resulted in the potentiometric surfaces shown on plate 5. Only by doubling pumpage to 100 Mgal/d are drawdowns greater than 200 ft below sea level obtained in the "400-foot" and "600-foot" aquifers. The potentiometric-surface maps for this second simulation with pumpage locations shown in figure 51 are shown on plate 6.

The last steady-state pumping simulation was made to estimate drawdowns resulting from the new pumping centers in southeastern Baton Rouge and at Port Hudson (fig. 52). For this simulation pumpage for 1985 was increased 20 percent and all new public-supply wells were simulated at maximum pumping rates. Pumpage was simulated from the "400-foot" and "600-foot" aquifers for the industrial district, Port Hudson, and Southeast Baton Rouge at 13.25, 5.65, and 7.86 Mgal/d, respectively. The "800-foot" aquifer was simulated with a total pumpage of 5.30 Mgal/d. This results in maximum drawdown of about 60 ft below sea level for the "400-foot" aquifer and 80 ft below sea level for the "600-foot" aquifer with a cone of depression of larger areal extent than in May and October 1984 (pls. 2 and 3). The potentiometric-surface maps from this simulation are shown on plate 7. The simulated decline in southeast Baton Rouge is 40 to 60 ft from levels of the predevelopment simulation and 20 ft from levels of the 1984 simulation.

Changes in Ground-Water Flow

Changes in the ground-water flow system do occur over time because of the development of the aquifers at Baton Rouge for industrial and public water supply. Other changes in ground-water flow occur seasonally, resulting from changes in recharge and changes in Mississippi River stage. Long-term changes in flow were examined by comparison of the predevelopment steady-state simulation with the transient calibration simulations. Seasonal changes in flow were examined by comparison of month to month changes for the transient calibration simulation of 1984.

Recharge in the Outcrop Area of the "400-Foot" and "600-Foot" Aquifers

The outcrop area of the "400-foot" and "600-foot aquifers was simulated as the 1,592 mi^2 area north of the inactive model cells of layer 1 in rows 1-5, where rivers and recharge are assigned to model layer 2 (fig. 13). Recharge was assigned at a constant rate of 6.86 in/yr for all simulations except the transient simulations of 1983-84. For this simulation recharge was assigned seasonally and the average annual rate is 6.97 in/yr.

The main change in the ground-water flow system is that, in the undeveloped ground-water flow system, water flows into the "800-foot" aquifer from deeper, zone 1 aquifers at a rate of 0.01 in/yr; but by 1985 when the deeper, zone 1 aquifers have been developed, water is flowing out of the "800-foot" aquifers to the deeper aquifers at a rate of about 0.35 in/yr in the outcrop area. This change occurs sometime after the June 1948 simulated stress period and prior to the January 1983 stress period.

Another change in the ground-water flow system in the outcrop area is in the rate at which water flows out of the area horizontally from layers 2, 3, and 4, the "400-foot," "600-foot," and "800-foot" aquifers, respectively. This change is summarized in table 6. For all simulations the total horizontal flow leaving the outcrop area for each active model layer was about 0.4 in/yr. This rate stayed constant no matter how much recharge was applied in the outcrop area. The "400-foot" aquifer had the most variation because of changes in Mississippi River stage, which created less of a gradient for discharge to the river during high stage, as in May 1984.

Table 6.--Horizontal flow leaving the simulated outcrop area for the "400-, 600-, and 800-foot" aquifers

[In inches per year]

	Layer 2 "400-foot" aquifer	Layer 3 "600-foot" aquifer	Layer 4 "800-foot" aquifer	Total horizontal flow
Steady-state simulation of undeveloped system.	0.171	0.204	0.035	0.410
Transient simulation August 1944.	.177	.209	.034	.420
Transient simulation May 1984.	.150	.206	.032	.388
Transient simulation October 1984.	.180	.209	.033	.422

Seasonal changes in flow in the outcrop area are complicated, but most of the changes occur in the "400-foot" aquifer, layer 2. Figure 53 is a bar graph of the simulated monthly water budget showing the seasonal change in flow into and out of the outcrop area of the "400-foot" aquifer. During the winter months when evapotranspiration is low some of the recharge to the aquifer goes into storage (indicated by negative storage for November through April, fig. 53). During the summer months when recharge is reduced by greater evapotranspiration, water moves out of storage into the aquifer (indicated by positive storage May through October, fig. 53). The net result is that base flow, ground-water discharge to the major rivers, is sustained at a fairly constant average rate, dropping off in the summer, reaching a low at the end of summer, and building back up to a peak at the end of winter.

Net Flux into or out of the Mississippi River

The Mississippi River is in direct connection to the Mississippi River alluvial aquifer and some of the shallow Pleistocene sands, model layer 1. The alluvial aquifer and the shallow Pleistocene sands are not heavily stressed by pumpage in the Baton Rouge area because of relatively poor water quality. During high stage, usually March through May, no model cells contribute water to the river. During low stage, usually July through October, almost all model cells contribute water to the river. The net flow into or out the Mississippi River alluvial aquifer from the Mississippi River along with a graph of mean monthly river stage at Baton Rouge is shown in figure 54. This graph is for the reach of the river in the model from St. Francisville to New Orleans, La., simulated in 1984.

The interaction of the Mississippi River with the alluvial aquifer is similar for all the transient simulations. The amount of flux through the riverbed is stage dependent, as the riverbed conductance is a constant and the river is connected to an aquifer that is not stressed significantly by pumpage. Peak flow into or out of the river was less than 500 ft^3/s for all simulations. The seasonal changes in flow into or out of layer 1 were similar for all years.

Vertical Movement of Water from Aquifers above the "400-Foot" Aquifer

Prior to ground-water development, flow through confining bed 1 (fig. 14) was upward. Water was discharging from the "400-foot" aquifer to the swamps and streams at land surface in the southeastern part of the modeled area. Also, over much of the Mississippi River alluvial aquifer (part of layer 1 not modeled as constant head, fig. 12), upward flow from the "400-foot" aquifer to the alluvial aquifer was about 12 Mgal/d. There is downward flow from the shallow Pleistocene sands to the "400-foot" aquifer in the area near the western edge of the Pleistocene terraces (fig. 2). In this area the aquifers above the "400-foot" aquifer tend to have water levels close to land surface and the "400-foot" aquifer, which has connection to the Mississippi River alluvial aquifer west of the flood-plain boundary (fig. 2), has water levels approaching mean annual river stage. Figure 55 shows contours of equal vertical flux between layers 1 and 2 for the steady-state simulation of the undeveloped aquifer system.

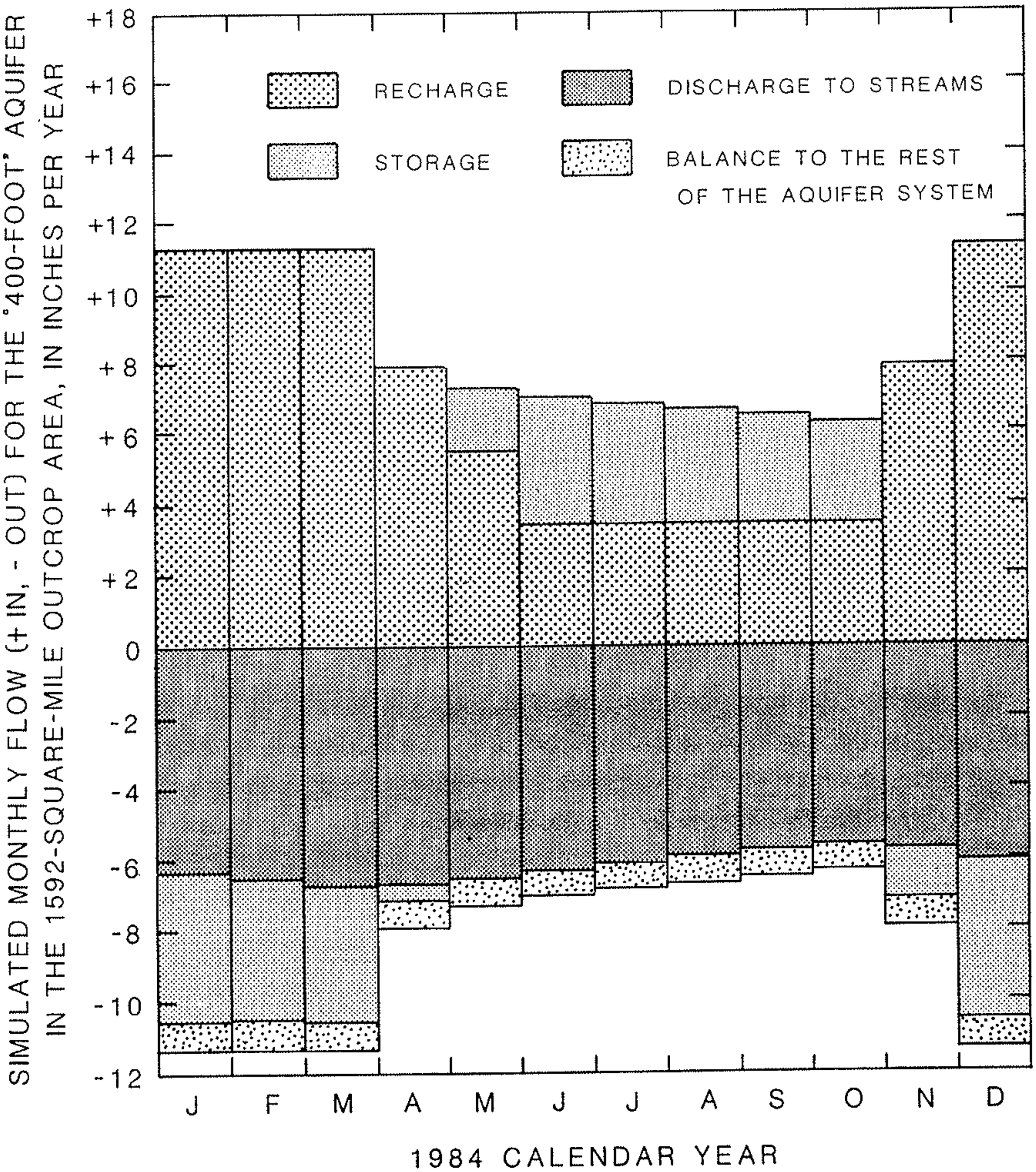


Figure 53.--Simulated monthly water budget for the outcrop area of the "400-foot" aquifer, 1984.

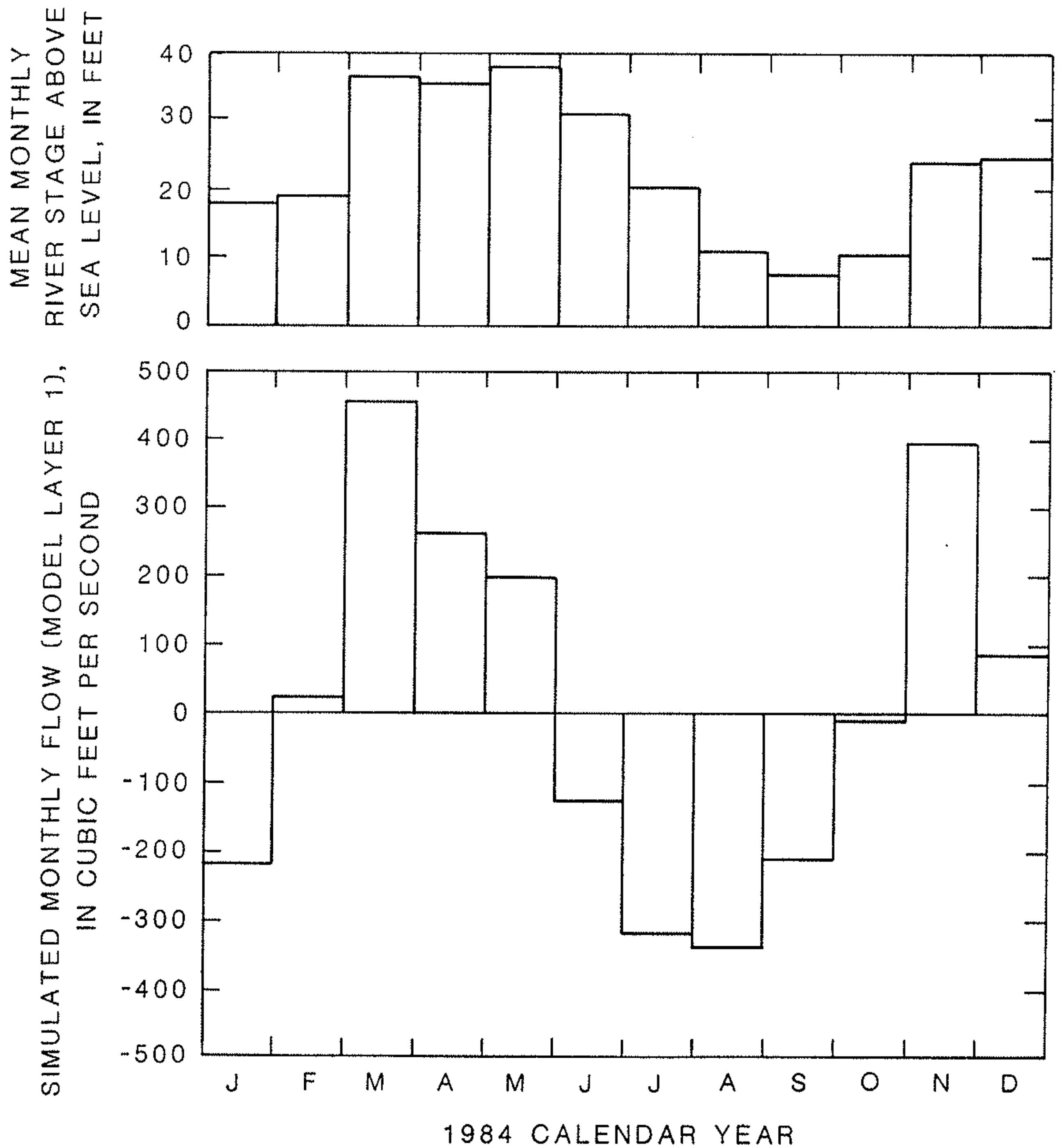
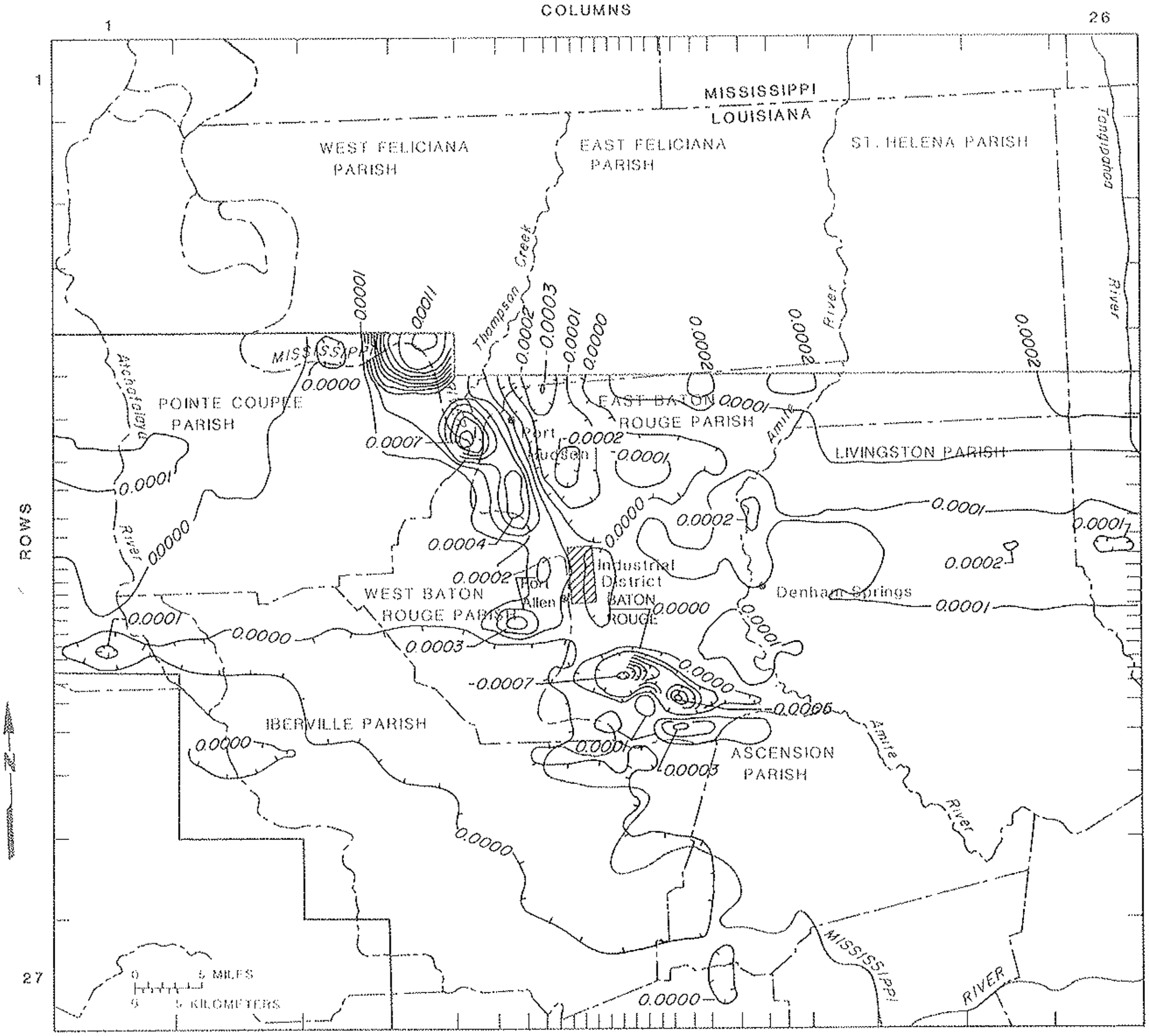


Figure 54.--Mean monthly Mississippi River stage at Baton Rouge, and simulated monthly net flow into or out of the Mississippi River alluvial aquifer from the Mississippi River reach between St. Francisville and New Orleans, Louisiana, 1984.



EXPLANATION

—0.0002— LINE OF EQUAL VERTICAL FLUX--
Hachures indicate depression.
Interval 0.0001 cubic foot per day per unit area

\- INDICATES FLOW IS DOWN

Figure 55.--Equal vertical flux between the shallow aquifers and the "400-foot" aquifer from the simulated undeveloped ground-water system, Baton Rouge area, Louisiana.

After development there are larger areas of downward flow between layers 1 and 2. Contours showing vertical movement of water between layers 1 and 2 for the simulation of August 1944 are shown in figure 56. Vertical flux ranges from 1 X 10^{-5} to 1 X 10^{-3} ft^3/d per unit area between the shallow Pleistocene sands and the "400-foot" aquifer in most of East Baton Rouge Parish.

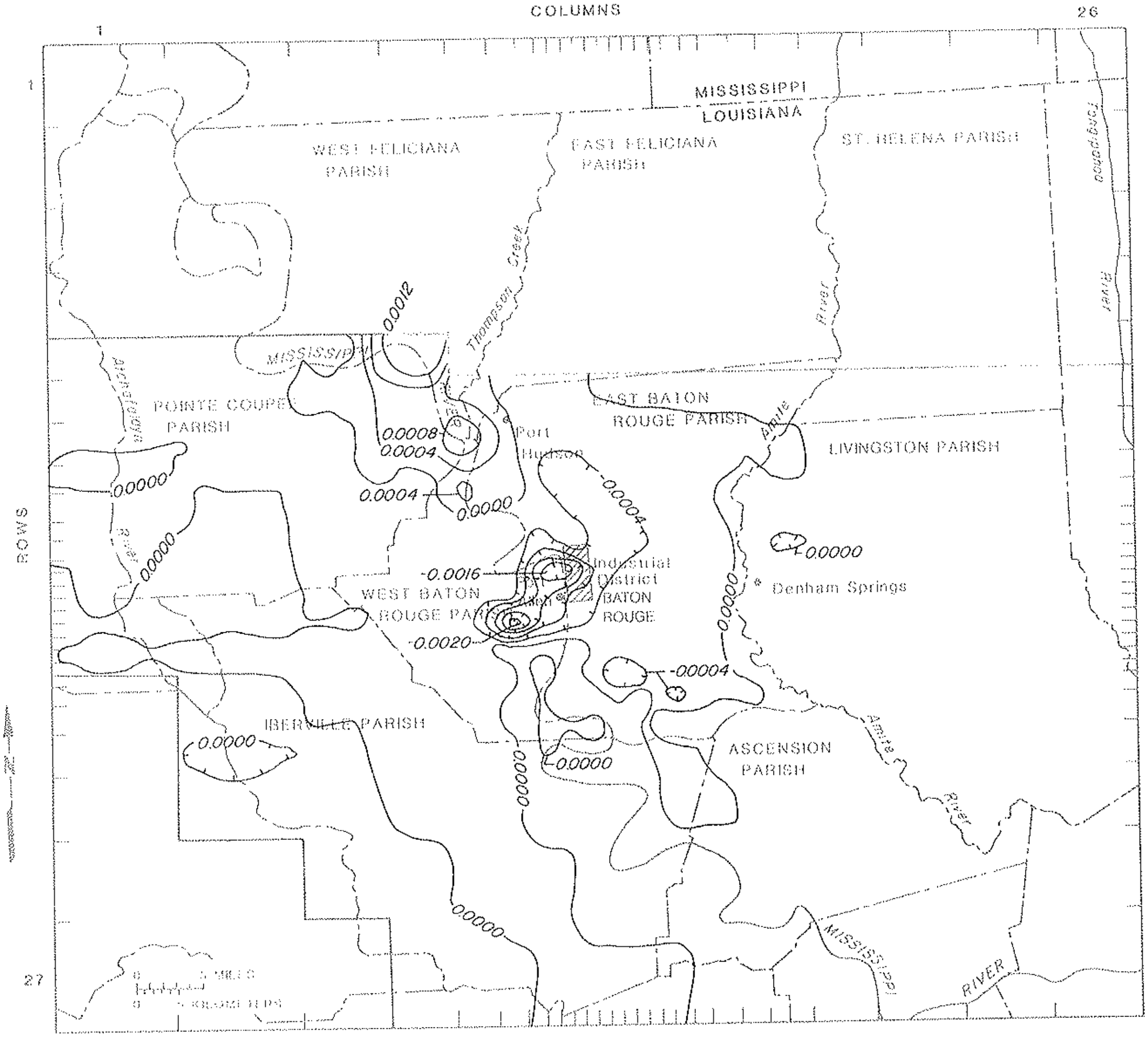
After compaction of the confining beds caused by the decline in artesian head, there is still downward flow at the pumping center in the industrial district, but it is not as great due to the reduction in vertical hydraulic conductivity of the confining bed. Figures 57 and 58 show the contours of equal vertical flux between the overlying shallow aquifers and the "400-foot" aquifer for May and October 1984, respectively. In May, when river stage is high, about 25 Mgal/d flows downward from the Mississippi River alluvial aquifer to the "400-foot" aquifer. In October, when river stage is low, about 12 Mgal/d leaves the "400-foot" aquifer and flows upward to the Mississippi River alluvial aquifer. The net flow for 1984 was about 1 Mgal/d of recharge to the "400-foot" aquifer by downward flow from the Mississippi River alluvial aquifer, a reversal from predevelopment conditions.

Water Budget for the Pumped Aquifers

The active model layers that are pumped are layers 2, 3, and 4, the "400-foot," "600-foot," and "800-foot" aquifers, respectively. Thus far, ground-water flow from many subareas of the model have been quantified. Water flow into and out of all three pumped aquifers is presented in table 7 for the steady-state predevelopment simulation, the simulated stress periods August 1944, May and October 1984, and the steady-state simulation of 1985 pumpage increased by 20 percent plus new public-supply wells pumping at maximum capacity. All values in table 7 are computed in million gallons per day; negative values represent discharge from the layers; positive values represent inflow to the layers.

A major change that occurs due to ground-water development after 1948 is that the net flow from the deeper aquifers (zone 1) is downward out of the "800-foot" aquifer, model layer 4. In the model area the natural gradient of deeper aquifers discharging upward shown in figure 4 has been reversed due to ground-water withdrawals.

The net flow from the "400-foot" aquifer to the Mississippi River alluvial aquifer and the shallow Pleistocene sands is seasonal and difficult to interpret. The steady-state simulation of the undeveloped system indicates net flow is upward. The May and October 1984 simulations indicate there is net flow into the "400-foot" aquifer from the shallower aquifers in May 1984, when river stage is high and net discharge to the shallower aquifers in October 1984 when river stage is low.

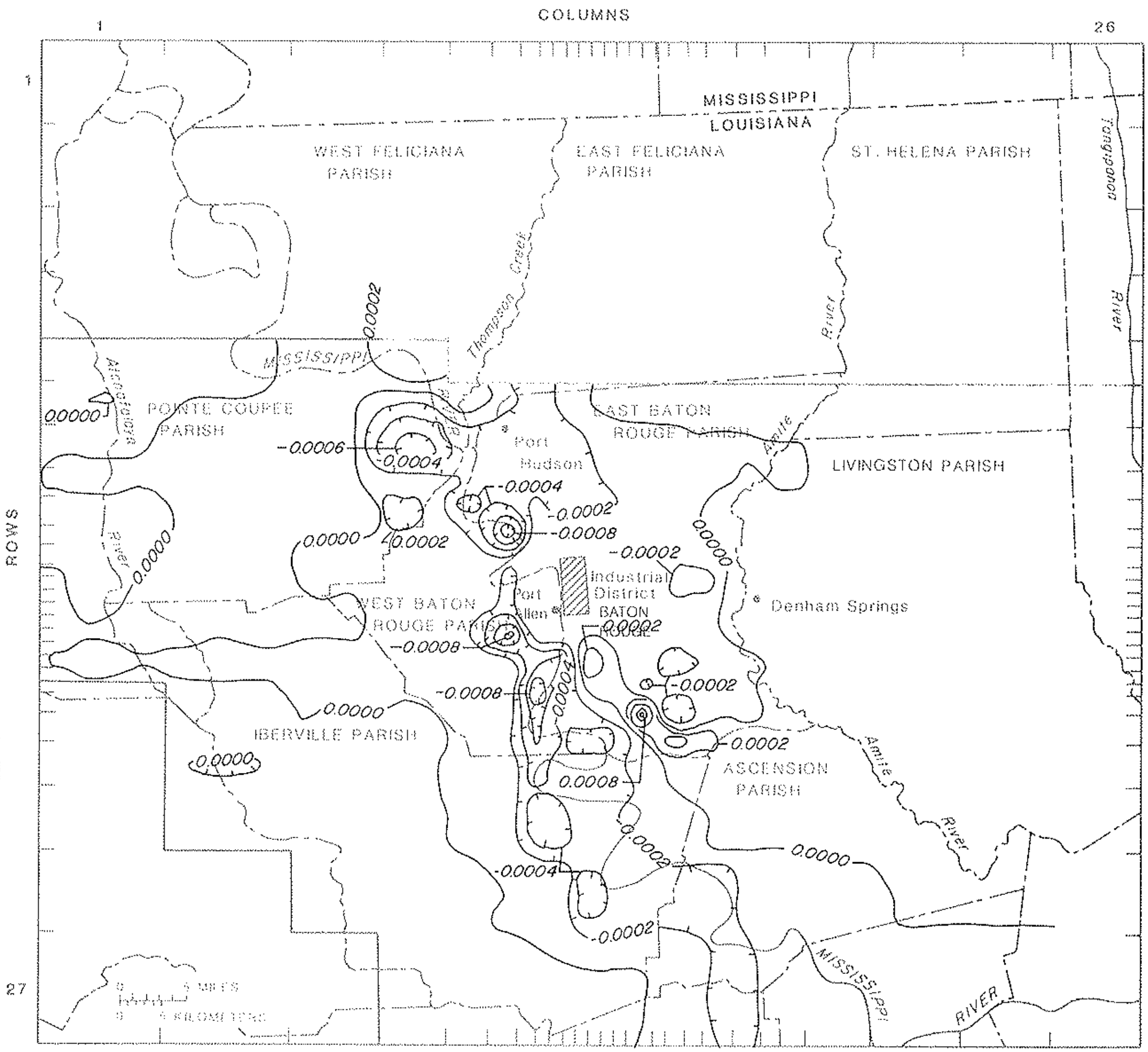


EXPLANATION

—— 0.0004 —— LINE OF EQUAL VERTICAL FLUX--
Hachures indicate depression.
Interval 0.0004 cubic foot per day per unit area

- INDICATES FLOW IS DOWN

Figure 56.--Equal vertical flux between the shallow aquifers and the "400-foot" aquifer for the simulated stress period for August 1944, Baton Rouge area, Louisiana.

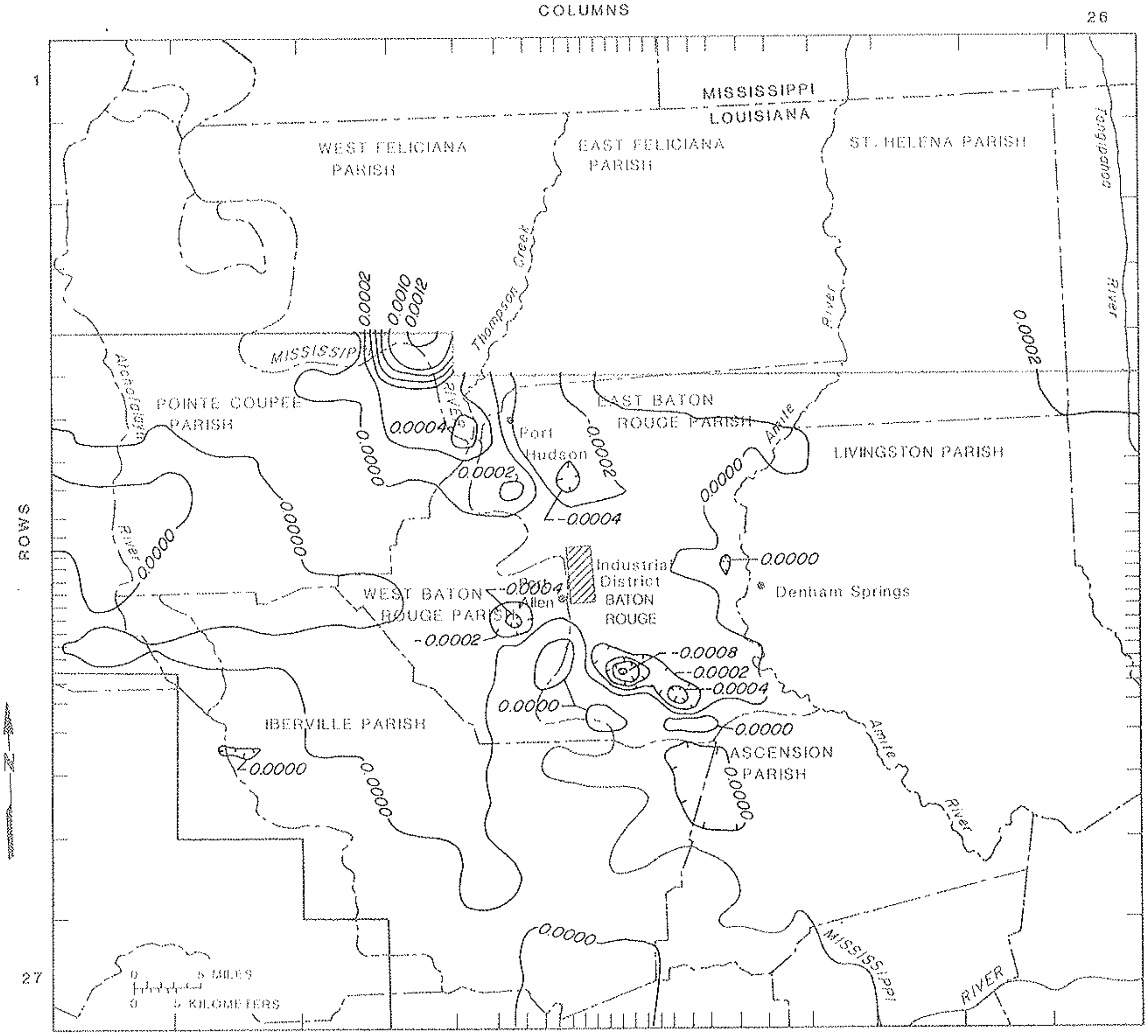


EXPLANATION

—0.0002— LINE OF EQUAL VERTICAL FLUX--
Hachures indicate depression.
Interval 0.0002 cubic foot per day per unit area

– INDICATES FLOW IS DOWN

Figure 57.--Equal vertical flux between the shallow aquifers and the "400-foot" aquifer for the simulated stress period for May 1984, Baton Rouge area, Louisiana.



EXPLANATION

—0.0002— LINE OF EQUAL VERTICAL FLUX--
Hachures indicate depression.
Interval 0.0002 cubic foot per day per unit area

\- INDICATES FLOW IS DOWN

Figure 58.--Equal vertical flux between the shallow aquifers and the "400-foot" aquifer for the simulated stress period for October 1984, Baton Rouge area, Louisiana.

Table 7.--Water budget for the pumped "400-, 600-, and 800-foot" aquifers (layers 2, 3, and 4)

[In million gallons per day; plus sign, equals inflow; minus sign, equals outflow]

	Recharge layer 2	Storage layers 2, 3, and 4	Rivers, layer 2	Wells, layers 2, 3, and 4	Net flow between layers 1 and 2	Net flow between layers 4 and 5
Steady-state simulation of undeveloped system.	+519.728	--------	-489.504	-------	-31.962	+1.713
Transient simulation August 1944.	+519.728	+23.659	-487.998	-36.303	-18.399	+.568
Transient simulation May 1984.	+415.784	+114.714	-494.856	-17.920	+9.621	-27.334
Transient simulation October 1984.	+259.864	+230.531	-423.692	-15.958	-24.316	-26.788
Steady-state hypothetical simulation increased 1985 pumpage.	+519.728	--------	-460.930	-32.054	-.032	-26.712

SUMMARY AND CONCLUSIONS

The "400-foot" and "600-foot" aquifers are a valuable source of water for both industrial and public use. Population growth southeast of the industrial area in recent years has resulted in new pumping centers in the "400-foot" and "600-foot" aquifers. Sources of water pumped from these aquifers are recharge from the outcrop area, the Mississippi River through the Mississippi River alluvial aquifer, water stored in the confining beds and aquifer, and downward leakage (induced by pumpage) from the overlying deposits.

The effects of past and future ground-water development from the "400-foot" and "600-foot" aquifers was evaluated by using a three-dimensional ground-water flow model. The model was calibrated by using hydrographs of water levels from 1940 to 1948 and potentiometric-surface maps for May and October 1984. Through calibration, it was determined that subsidence may have reduced the conductance of clay beds that underlie the industrial district. Simulations indicate that less than 1 in/yr of recharge in the outcrop area goes into the regional flow system. More than 95 percent of recharge in the outcrop area returns to streams, and there is no permanent change in storage in the aquifer in the outcrop area.

Because the Mississippi River is in direct hydraulic connection with the shallower unstressed aquifers above the "400-foot" aquifer, flow into or out of the river is seasonal. Peak flow into or out of the river was always less than 500 ft^3/s for all simulations for the simulated reach of the river between St. Francisville and New Orleans, Louisiana. During high stage, usually March through May, flow is from the river into the Mississippi River alluvial aquifer and the shallow Pleistocene sands for the whole model reach. During low stage, usually July through October, flow is out of the aquifers into the river over the entire model reach.

In Livingston Parish, ground water is still discharging upward to the surface swamps and streams not affected by ground-water development at Baton Rouge. In the southeastern part of East Baton Rouge Parish, ground-water development has reversed this natural movement, and there is downward flow from the shallow sands to the "400-foot" aquifer. Simulated vertical flux to the "400-foot" aquifer varies 1×10^{-5} to 1×10^{-3} ft^3/d per unit area, depending on the hydraulic gradients. The "400-foot" and "600-foot" aquifers recover quickly from the effects of pumpage because of the proximity of the outcrop area and the Mississippi River.

Over the Mississippi River alluvial plain in Pointe Coupee and West Baton Rouge Parishes, there is seasonal discharge and a recharge cycle from the "400-foot" aquifer to the Mississippi River alluvial aquifer that is related to river stage. The steady-state simulation, using average river stage and pumpage for 1985 increased to 32 Mgal/d, indicated that there is net recharge to the "400-foot" aquifer and deeper aquifers from the Mississippi River alluvial aquifer. Prior to development of the aquifers there was a net discharge to the Mississippi River alluvial aquifer from the deeper aquifers.

The Baton Rouge fault, which is a barrier to ground-water flow in the "800-foot" and deeper aquifers throughout southern Louisiana and Mississippi, is a barrier to flow in the "400-foot" and "600-foot" aquifers west of the Amite River. The Baton Rouge fault does not affect flow in the Mississippi River alluvial aquifer and does not appear to affect flow in the "400-foot" and "600-foot" aquifers in Livingston Parish.

About 1 ft of subsidence due to ground-water withdrawals occurred at Baton Rouge during 1935-65. The subsidence was the result of a decline in artesian head of about 200 ft. The resulting compaction of the sequence of sediments modeled was completed prior to 1975 in the industrial district of Baton Rouge. Compaction of the clays and silts results in reduced porosity and hydraulic conductivity of the confining beds. Compaction of the confining beds allows less vertical movement of water from the shallow sands above the "400-foot" aquifer. Past water-level and pumpage data and the calibration of the model for 1940-48 and 1983-84 indicate that more drawdown occurs with less pumpage in the industrial district at present than occurred in 1944-48 as a result of compaction. Compaction of sediments between land surface and 833 ft below land surface has ceased in the industrial district.

It was not within the scope of this study to predict subsidence caused by future ground-water withdrawals. The simulation of the area where newer pumpage occurs in southeastern Baton Rouge indicates declines of 40 to 60 ft from the simulated water levels of the undeveloped aquifer and of about 20 ft from the 1984 simulated water levels. The effects of both transient leakage and possible subsidence are not accounted for in the area where new wells are located.

The reduction of ground-water withdrawals from 36 Mgal/d in 1940 to 14 Mgal/d in 1983 in the industrial district and the spreading out of the major pumping centers to other parts of the metropolitan Baton Rouge area resulted in an overall rise of water levels. Declines of as much as 190 ft occurred in the 1940's and 1950's. The present (1986) maximum water-level decline is less than 100 ft.

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