Estimates of Hydraulic Conductivity from Aquifer–Test Analyses and Specific–Capacity Data, Gulf Coast Regional Aquifer Systems, South–Central United States

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CONVERSION FACTORS AND ABBREVIATIONS

"Inch-pound" units of measure used in this report may be converted to metric (International System) units by using the following factors:

Multiply	*By*	*To obtain*
foot (ft)	0.3048	meter
inch (in)	25.40	millimeter
mile (mi)	1.609	kilometer
square foot (ft^2)	0.09290	square meter
square mile (mi^2)	2.590	square kilometer

ALTITUDE DATUM

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.

ESTIMATES OF HYDRAULIC CONDUCTIVITY FROM AQUIFER-TEST ANALYSES AND SPECIFIC-CAPACITY DATA, GULF COAST REGIONAL AQUIFER SYSTEMS, SOUTH-CENTRAL UNITED STATES

By David E. Prudic

ABSTRACT

Hydraulic conductivities were estimated from more than 1,500 aquifer-test analyses and more than 5,000 specific-capacity data from wells drilled into Tertiary and younger sediments of the Gulf Coast region in the south-central United States. The values are assumed to represent the coarser-grained sediments in the aquifer systems. The purpose of estimating hydraulic conductivities for this area is to compare these estimates to hydraulic conductivities determined from the simulation of regional ground-water flow as part of the Gulf Coast Regional Aquifer-System Analysis project. In the simulation model, hydraulic conductivities are separated into two groups: coarse-grained sediments (sands) and fine-grained sediments (silts and clays).

Values for hydraulic conductivity range from less than 1 foot per day to more than 1,000 feet per day. The values are log normally distributed; thus, the geometric mean was used to represent a typical hydraulic conductivity. The geometric mean hydraulic conductivity for the entire study area was 55 feet per day from aquifer-test analyses and 71 feet per day from specific-capacity data.

A two-way analysis of variance was performed on the combined estimates of hydraulic conductivity that were grouped into 10 model layers and 9 areas within the overall study area. Results of this analysis indicate that area, layer, and the interaction of area and layer were all significant in explaining the variation of hydraulic conductivity at a probability level of 0.001. Thus, comparisons of means were done for each area and layer combination. Overall, the highest geometric means generally were in model layer 11 which corresponds to the upper Pleistocene and younger deposits along the coast of the Gulf of Mexico and the alluvium of the Mississippi River. Within each model layer, the geometric mean increased from areas along the western part of the study area to the eastern part, which indicates that the deposits near the Mississippi River might be more permeable than elsewhere.

Two separate analysis of covariance were performed on the estimates of hydraulic conductivity to determine if variations within each area and layer combination could be explained by depth of the well or by the thickness of sand beds throughout the perforated interval of the well. Results of these analyses indicate that depth to the middle of the perforated or screened interval was significant at the probability level of 0.02 and that sand-bed thickness was not significant at the probability of 0.10. In the analysis with depth, hydraulic conductivity decreased as a function of depth in a majority of area and layer combinations.

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INTRODUCTION

The method commonly used to represent a heterogeneous aquifer system in the simulation of three-dimensional ground-water flow is to conceptualize the system as an equivalent homogeneous anisotropic system. An aquifer system is represented in model simulations as discrete blocks in which an effective hydraulic conductivity is assigned to each of the three principle directions (two horizontal, one vertical). These effective values correspond to some combination of the hydraulic conductivities of the different lithologies present in each block.

The effective hydraulic conductivity assigned to model blocks depends on the geometry of the different lithologies and on the scale used in the simulation of ground-water flow (Freeze, 1975). In the case where a heterogeneous aquifer system is composed of alternating beds of different lithologies (for example sand and clay beds) whose extent exceeds that of a model block (fig. 1a), the effective hydraulic conductivity parallel to the layering of sediments (usually the horizontal directions) is commonly calculated as being equal to the arithmetic mean of the hydraulic conductivity of the individual beds weighted by the thickness of each (Bear, 1972, p. 154). The effective hydraulic conductivity perpendicular to the layering (usually vertical) is calculated as being equal to the harmonic mean of the hydraulic conductivity of the individual beds.

In aquifer systems where individual beds of both coarse- and fine-grained deposits are considerably smaller in extent than model blocks, and the beds are randomly distributed within the blocks (fig. 1b), the effective hydraulic conductivity in any direction is equal to the geometric mean of the hydraulic conductivity of the individual beds. However, in aquifer systems where the individual beds vary from smaller than to larger than the model blocks (fig. 1c), the effective hydraulic conductivity in the horizontal direction is between the geometric and arithmetic means, and the effective hydraulic conductivity in the vertical direction is between the harmonic and geometric means (Fogg, 1989, p. 46). The latter case is typical of aquifer systems in the Gulf Coastal Plain, and may be common in aquifer systems elsewhere.

Several investigators (summarized by Neuman, 1982, p. 83) have observed that the frequency distribution of hydraulic conductivity is generally log normal for a variety of aquifer materials. This suggests the arithmetic mean may not be appropriate for determining the hydraulic conductivity of a particular bed; rather the geometric mean (mean of the log transformed values) of the hydraulic conductivities may be more appropriate. The geometric mean was determined by Warren and Price (1961) to be the most appropriate value for the effective hydraulic conductivity when a sand was composed of heterogeneous sizes with no spatial correlation. Gutjahr and others (1978) derived equations for two-dimensional flow in an isotropic system where the distribution of hydraulic conductivity was assumed log normal. They concluded that the geometric mean was the appropriate value when representing the system with one representative hydraulic conductivity. This was the same conclusion as discussed by Matheron (1967). For three-dimensional flow, however, Gutjahr and others (1978, p. 956) determined that the effective hydraulic conductivity was slightly more than the geometric mean.

Thus, for a layered system, it may be best to approximate the hydraulic conductivity of each layer using the geometric mean prior to converting it into an equivalent homogeneous anisotropic system.

The main purpose of this report is to present the spatial distribution and statistical summaries of hydraulic conductivities estimated from numerous aquifer tests and specific capacities within the aquifer systems of the Gulf Coastal Plain (fig. 2). These aquifer systems

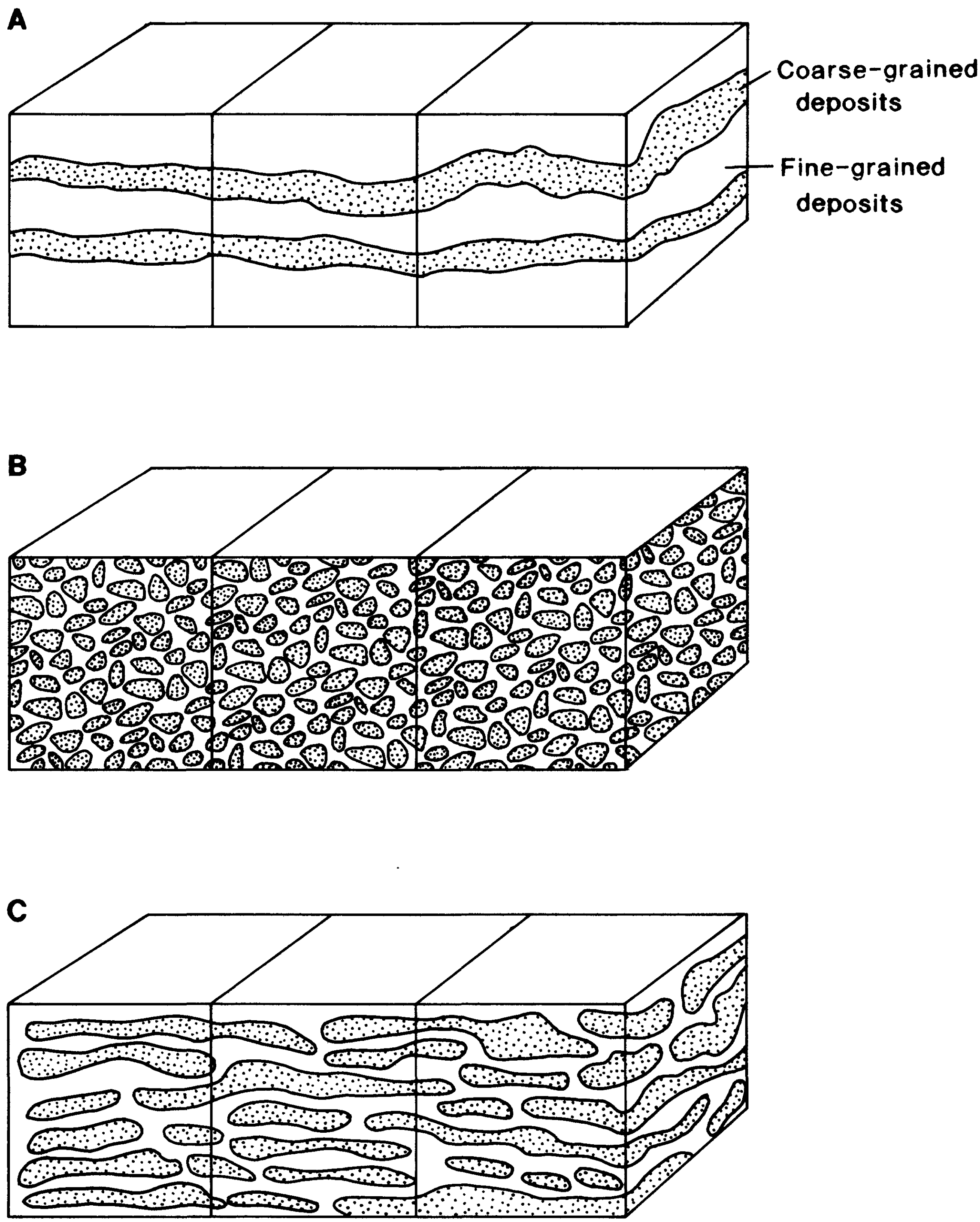


Figure 1.--Diagram showing model cells with varying distributions of coarse- and fined-grained deposits.

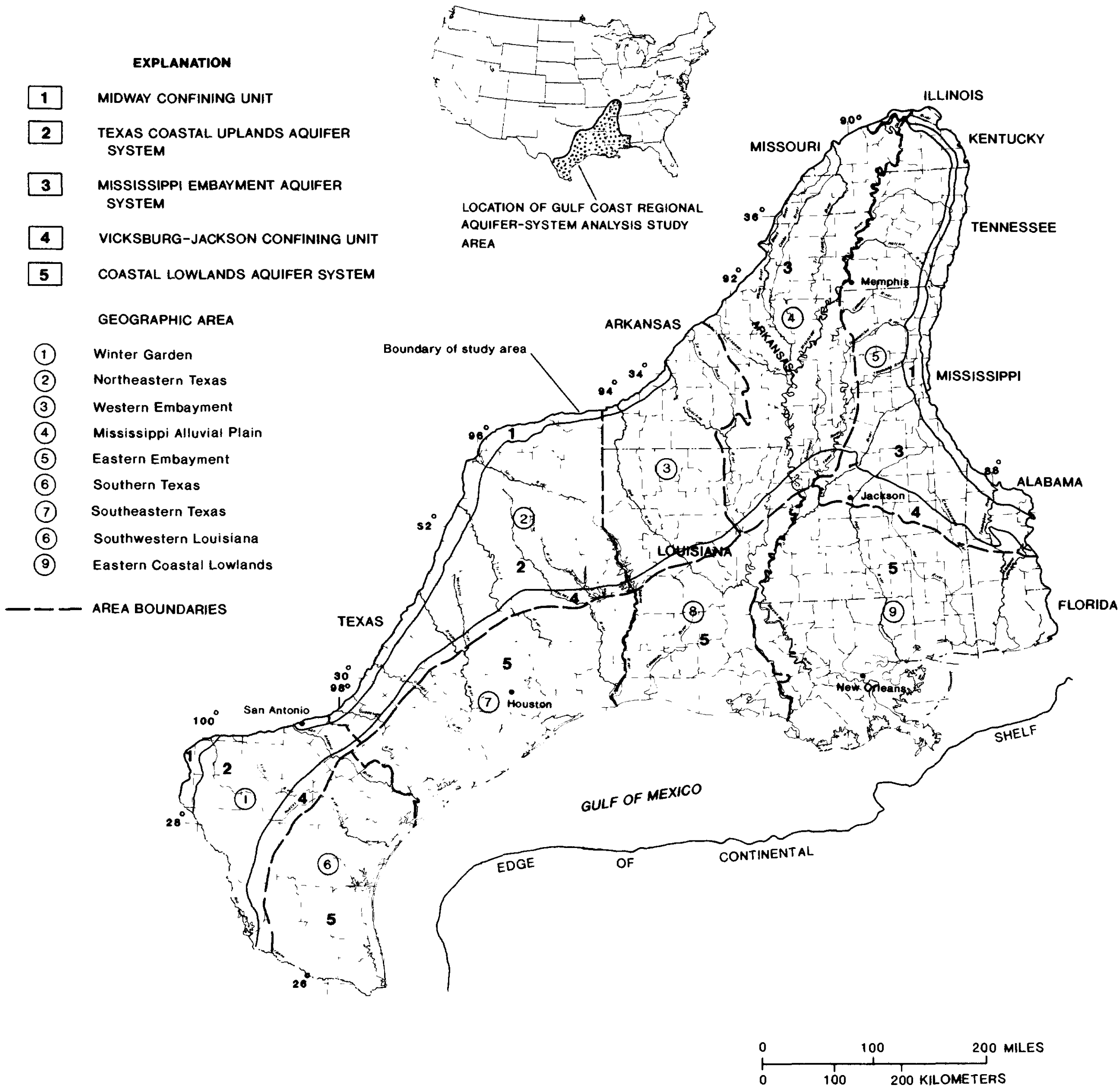


Figure 2.--Location of study area, aquifer systems, and geographics areas. (Modified from Mesko and others, 1990)

are being studied as part of the Regional Aquifer-System Analysis (RASA) program of the U.S. Geological Survey (Grubb, 1984). Included in the study is the regional simulation of ground-water flow.

The simulation model uses a finite-difference numerical scheme to solve the differential equation of ground-water flow in three dimensions (Kuiper, 1985). The aquifer systems are divided into model blocks with horizontal dimensions of 10 miles by 10 miles. Ten model layers are used to simulate vertical flow (Williamson, 1987). The effective horizontal and vertical hydraulic conductivity for each finite-difference block is calculated in the simulation on the basis of the sand percentage and a hydraulic conductivity for sand and fine-grained beds using a technique described by Desbarats (1987). The estimates of hydraulic conductivity from aquifer tests and specific capacities are assumed to represent the hydraulic conductivities of the sand beds because most wells are screened opposite the more productive zones.

This report describes the methods used to estimate hydraulic conductivities from aquifer tests and specific capacity data, and discusses the results of statistical analyses of the data. The distributions of estimated hydraulic conductivities presented herein will be used in evaluating the results of the model simulations.

DESCRIPTION OF THE STUDY AREA

The study area includes the Gulf Coastal Plain in parts of Alabama, Arkansas, Florida, Illinois, Kentucky, Mississippi, Missouri, Tennessee, Texas, and all of Louisiana (fig. 2). It covers an area of about 230,000 square miles on land and an additional 60,000 square miles of sediments beneath the Gulf of Mexico.

Three regional aquifer systems have been delineated in the study area (Grubb, 1984). These are the coastal lowlands, Texas coastal uplands, and Mississippi embayment aquifer systems (fig. 2). The aquifer systems are further divided into 9 areas and 17 geohydrologic units on the basis of geology, hydrology, and topography (Mesko and others, 1990; Grubb, 1987).

The aquifer systems are comprised of Tertiary and younger sediments that are predominately alternating beds of sand and clay with some interbedded gravel, silt, lignite, and limestone. The sediments generally dip towards the Gulf of Mexico geosyncline becoming thicker and finer-grained downdip. The downdip extent of the aquifer systems are terminated where the sediments grade into clays or where above normal pressures (geopressures) are present (Hosman and Weiss, in press).

The Mississippi embayment and Texas coastal uplands aquifer systems underlie the northern part of the coastal lowlands aquifer system and are separated from it by the Vicksburg-Jackson confining unit. This unit is primarily a marine clay, with marl and limestone and is present throughout much of the study area. The boundary between the Texas coastal uplands and Mississippi embayment aquifer systems is along the Sabine arch and uplift (Grubb, 1984; Hosman and Weiss, in press). Quaternary alluvium of the Mississippi River is included in both the Mississippi embayment and coastal lowlands aquifer systems. These deposits overlie an extensive area of Tertiary sediments and provide a lateral hydraulic connection between the Mississippi embayment aquifer system and the coastal lowlands aquifer system in east-central Louisiana.

The three aquifer systems were divided into geohydrologic units to quantitatively describe the ground-water hydrology of the study area (Hosman and Weiss, in press and Weiss, in press). The geohydrologic units for the aquifer systems are summarized in table 1. The

Table 1.--*Summary of aquifer systems, geohydrologic units, and model layers for the Gulf Coast Regional Aquifer-System Analysis study area (from Williamson, 1987, table 1)*

Aquifer system	Regional model layer number	Regional geohydrologic units
Coastal lowlands	11	Permeable zone A (Holocene-upper Pleistocene deposits)
	10	Permeable zone B (lower Pleistocene-upper Pliocene deposits)
	9	Permeable zone C (lower Pliocene-upper Miocene deposits)
	17	Zone D confining unit
	8	Permeable zone D (middle Miocene deposits)
	16	Zone E confining unit
	7	Permeable zone E (lower Miocene-upper Oligocene deposits)
	15	Vicksburg-Jackson confining unit [1]
Texas coastal uplands and Mississippi embayment [3]	11	Mississippi River Valley alluvial aquifer [2]
	6	upper Claiborne aquifer
	14	middle Claiborne confining unit
	5	middle Claiborne aquifer
	13	lower Claiborne confining unit
	4	lower Claiborne-upper Wilcox aquifer
	3	middle Wilcox aquifer
	2	lower Wilcox aquifer [2]
		Midway confining unit [1]

[1] The Midway confining unit was referred to as the Coastal Uplands confining system and the Vicksburg-Jackson confining unit was referred to as the coastal lowlands confining system by Grubb (1984, p. 11).

[2] Not present in Texas coastal uplands aquifer system.

[3] The confining units and aquifers in descending order are in the Oligocene Vicksburg Group, the Eocene Jackson and Claiborne Groups, the Eocene and Paleocene Wilcox Group, and the Paleocene Midway Group.

general outcrop area of the units is shown in figure 3. The numbering scheme presented in table 1 and figure 3 correspond to layer numbers used in the simulation model. Model layers 2 through 11 identify aquifers and are numbered sequentially from oldest to youngest deposits (generally from north to south in outcrop). Layer 11 also includes the Mississippi River Valley alluvial aquifer. Model layers 12 through 17 identify principle confining units and also are numbered from oldest to youngest deposits.

The relation of model layers is shown in a schematic section through the study area (fig. 4). Regional trends in hydraulic conductivity for the study area were evaluated on the basis of these model layers.

The geohydrologic units are overlain by recent alluvial sediments in many areas but particularly along the Mississippi River. The geohydrologic units for the Texas coastal uplands and Mississippi embayment aquifer systems were delineated by Hosman and Weiss (in press) on the basis of predominant lithology (sand or clay). Generally, stratigraphic units consist of a predominant lithology. Thus, stratigraphic and geohydrologic units generally coincide. However, the boundaries between geohydrologic units do not always correspond to the boundaries between stratigraphic units in areas where the lithology near the top or base of a stratigraphic unit is different from the predominant lithology used to define the geohydrologic unit.

The coastal lowlands aquifer system was more difficult to divide into geohydrologic units because distinct lithologic units can not be correlated from one area to the next. Because of this difficulty, the aquifer system was divided into five permeable zones on the basis of permeability contrasts as inferred from geophysical logs, intervals of pumped zones, and variations of vertical hydraulic gradients (Weiss, in press, and Weiss and Williamson, 1985). Two confining units were also delineated but the units do not extend over the entire area.

METHODS OF ESTIMATING HYDRAULIC CONDUCTIVITY

Two types of data were available for estimating hydraulic conductivity in the study area: (1) Data from aquifer tests compiled especially for this study, and (2) specific-capacity tests that had been entered into the U.S. Geological Survey's WATSTORE database prior to the beginning of this study. How each of these data sources were analyzed to provide regional estimates of hydraulic conductivity is described in this section.

Aquifer Tests

Data from aquifer-test analyses kept in files of each of the U.S. Geological Survey offices within the study area were compiled and entered into a computer file. Data entered into the computer file for each test included: (1) the location of the test (state, county or parish, and 25-square-mile block coordinates of the pumped well); (2) altitude of land surface; (3) the number of wells used to measure water-level changes; (4) total time of test; (5) pumping rate; (6) drawdown in pumped well; (7) depth to bottom of pumped well; (8) screened interval of pumped well; (9) thickness of sand in test interval; (10) estimate of transmissivity and storage coefficient; (11) specific capacity of pumped well; (12) type of test (drawdown or recovery); and (13) a subjective rating of test (good, fair, and poor). An example of the format used to enter the data into the computer file is presented by Martin and Early (1987, p. 3).

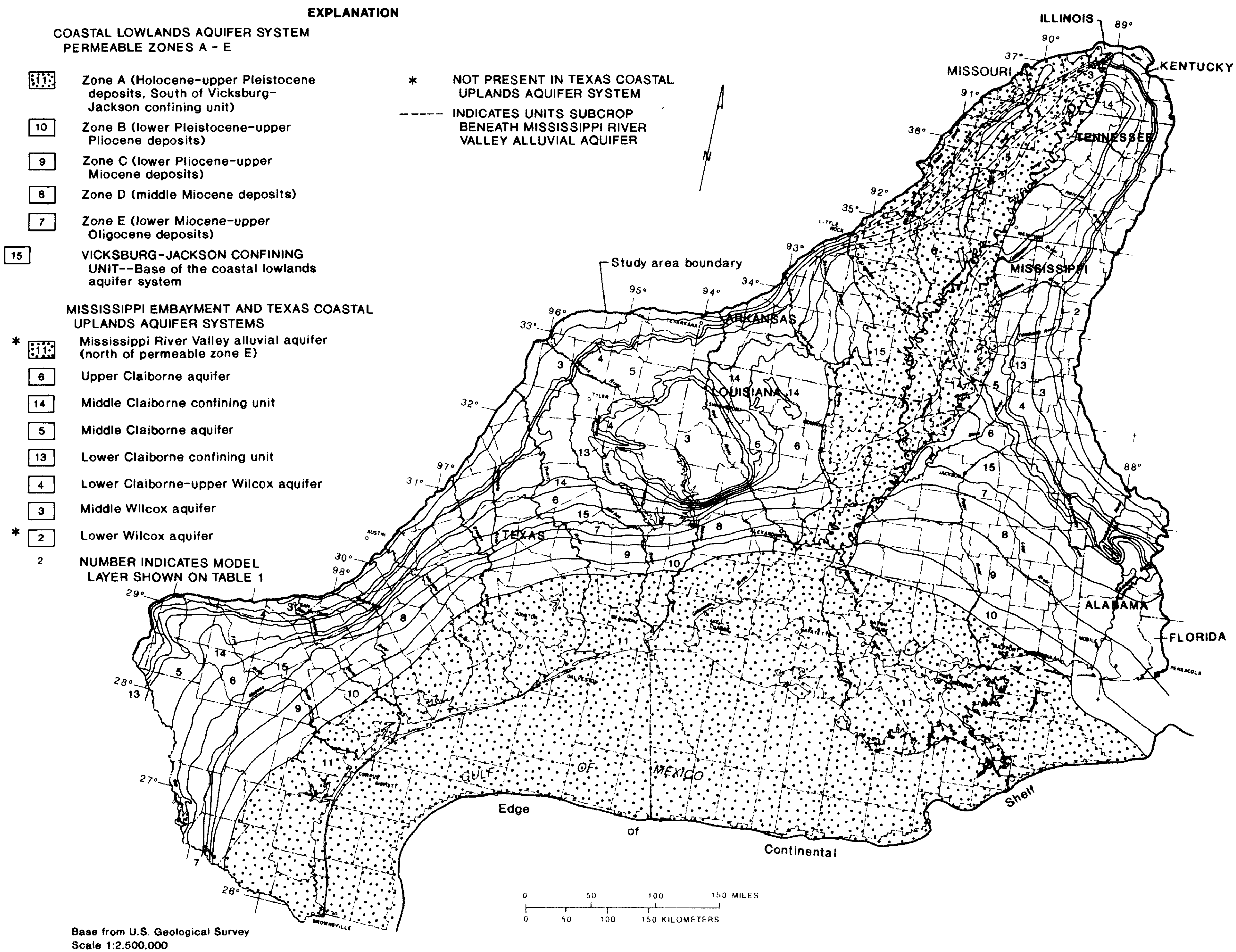


Figure 3.--Generalized outcrop of regional geohydrologic units. (From Grubb, 1987)

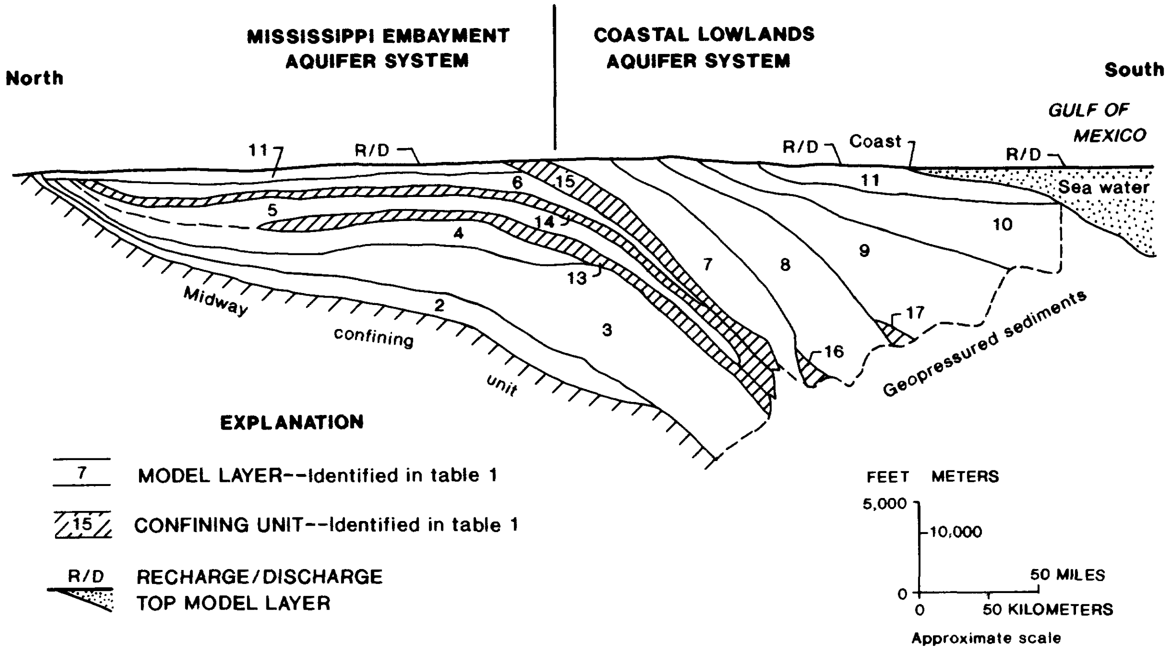


Figure 4.--Idealized diagram from northern edge of study area to edge of Continental Shelf showing vertical relation of model layers. (From Williamson, 1987)

Estimates of transmissivity were determined from drawdown or recovery parts of the tests. Most of the analyses from aquifer tests have been previously published. Results of aquifer tests from Tertiary sediments in the Mississippi embayment were published by Hosman and others (1968); results of aquifer tests in Texas by Myers (1969); in Mississippi by Newcome (1971); and in Louisiana by Martin and Early (1987).

Data from more than 2,500 aquifer tests were entered into the computer file. The tests were conducted by a variety of private companies and government agencies including well drillers, private consultants, state and local government agencies, and the U.S. Geological Survey and other federal agencies. The number of wells used in the tests ranged from a single pumped well to tests with six observation wells. Some of the tests were simple, using crude equipment and methods of measurement and analyses; others were more sophisticated.

In some wells, duplicate tests were conducted and consequently the computer file includes multiple estimates of transmissivity. An average transmissivity value was determined for wells with multiple estimates. In addition, many tests were not used in the statistical analyses of hydraulic conductivity because not all the necessary data were available to estimate hydraulic conductivity. Estimates of hydraulic conductivity were made from 1,557 wells. The distribution of these estimates are shown in figure 5. Most of the tests are concentrated in Louisiana, Mississippi, and Texas. The distribution of aquifer tests assigned to model layers generally follows the outcrop or subcrop (areas beneath the Mississippi River alluvium) and slightly downdip of geohydrologic units shown in figure 3.

Hydraulic conductivity was estimated from the aquifer-test analyses by dividing the estimate of transmissivity with:

1. The screened interval for tests in which only the pumped well was used in the analyses of transmissivity;

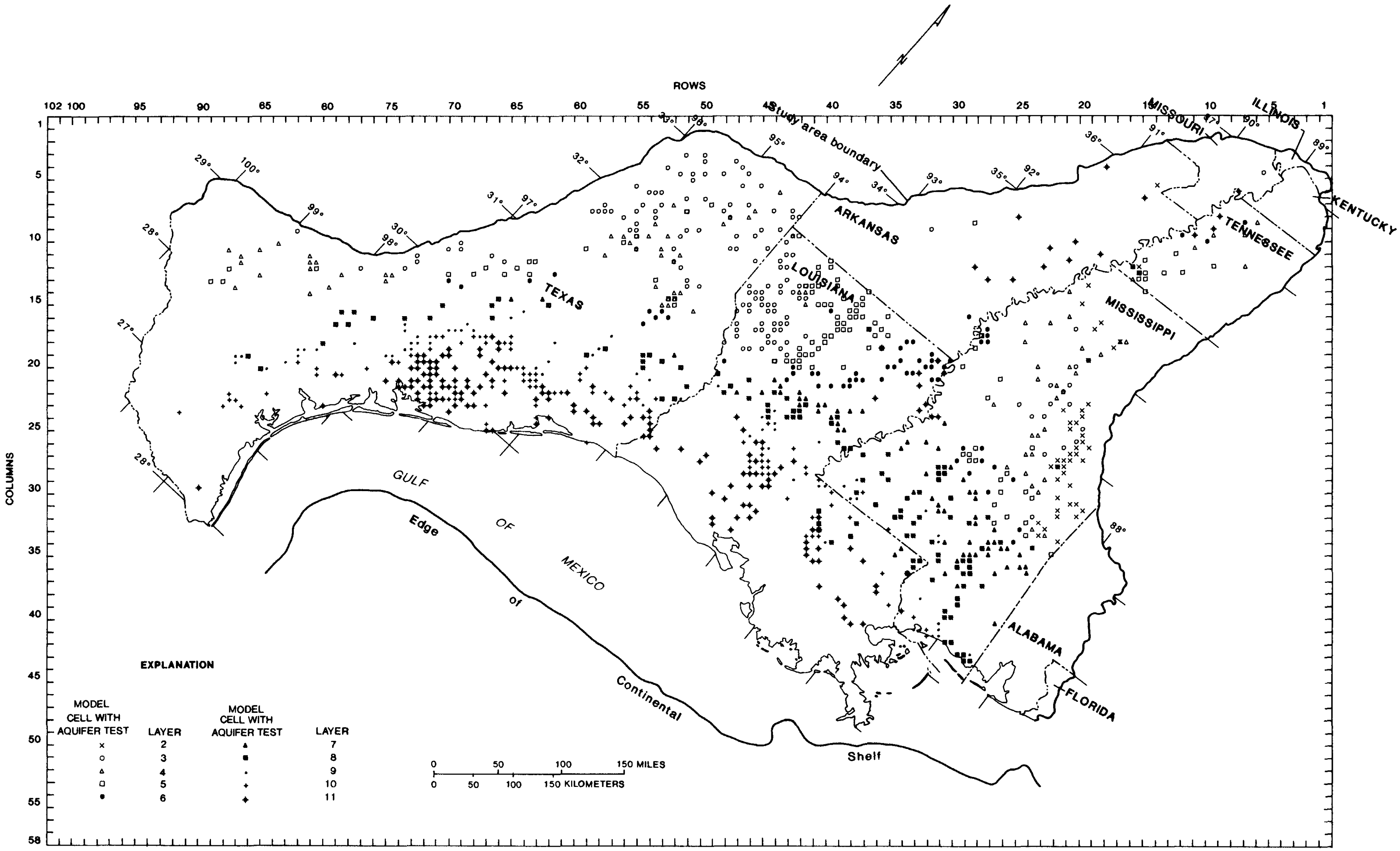


Figure 5.--Location of wells with aquifer tests used to estimate hydraulic conductivity.

2. The thickness of the sand beds for tests in which estimates of transmissivity were determined from observation well(s); and

3. Twice the screened interval for tests in which estimates of transmissivity were determined from observation well(s) but the sand-bed thickness was unknown.

The factor of 2 was determined as the average ratio between the sand-bed thickness and screened interval from tests in which both were known.

Specific-Capacity Data

Specific-capacity data from the aquifer-test file and data retrieved from the U.S. Geological Survey's WATSTORE (Baker, 1977) database also were used to estimate hydraulic conductivities of sediments in the study area. Transmissivities from specific-capacity data were first estimated from data in the aquifer-test file for wells that also included an estimate of transmissivity from an aquifer test. The two estimates of transmissivity were then compared by regression to develop a relation between transmissivity estimates from specific-capacity data to those from aquifer tests. The relation was used to adjust all transmissivities estimated from specific-capacity data.

A total of 1,068 tests out of 2,518 tests in the aquifer-test file included both an estimate of transmissivity and a specific capacity value. Transmissivity was calculated from specific capacities using the method described by Theis and others (1963, p. 331-341). The equation used to estimate transmissivity is modified from equation 1 (p. 332) in their report to convert to units used in this study and is an abbreviated form of Theis's equation. The equation is:

$$T = 15.32\ (Q/s)\ \left(-0.577 - \log_e \left[\frac{r^2 S}{4Tt}\right]\right), \qquad (1)$$

where

Q/s = specific capacity of a pumped well in gallons per minute per foot of drawdown;
r = effective radius of pumped well in feet;
S = storage coefficient in cubic feet of water per cubic feet of aquifer;
t = time in days, and
T = transmissivity in feet squared per day.

An iterative process was used to solve the equation because transmissivity is on both sides of the equation. An initial estimate of transmissivity of 1,000 ft^2/day was assumed for T on the right side of equation 1 and a new transmissivity estimated from the equation (T on left side of equation). The new value of T was then substituted into the right side of the equation and the process repeated until the difference between transmissivity values on the right and left sides of the equation was less than the specified value of 10 ft^2/d.

Several assumptions were used to calculate transmissivity from specific capacity data. Storage coefficients used in the equation were estimated by assuming an average specific yield of 0.15 whenever the depth of the well was less than 150 ft or whenever the depth to the top of the perforations was less than 100 ft. Otherwise, a specific storage of 4×10^{-6} per foot was assumed, which was multiplied by the length of the perforated section of the well. Wells where the perforated length was unknown were not included in the analyses. The effective radius of the well was assumed to be equal to the radius of the well. This assumption may result in too small of an estimate of the effective radius when the well is highly developed and in unconsolidated materials. For wells where the radius is unknown, the average well radius for all wells of .33 ft was used. Uncertainties in the storage coefficients and the effective radius result in generally small differences in the estimate of transmissivity because both are within the log term in equation 1.

Transmissivities calculated from specific-capacity data were regressed with corresponding estimates determined from aquifer-test analyses. Results of a simple least squares regression of the $\log_{10}$ transformed values is shown in figure 6. The correlation coefficient for the regression is 0.82 and the coefficient of determination (r-squared value) is 0.68. The correlation coefficient of non-transformed values is 0.43 and the coefficient of determination is 0.19.

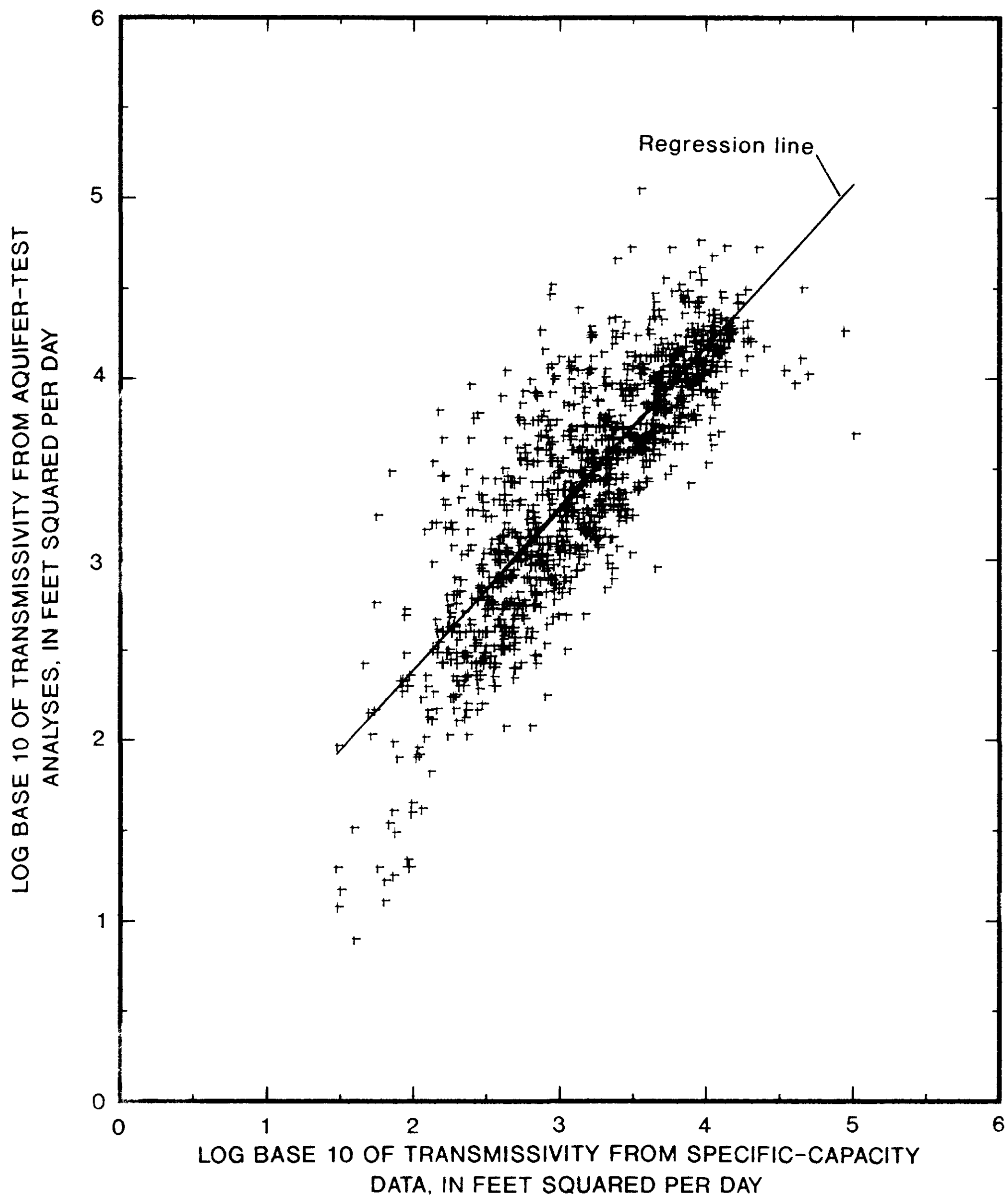


Figure 6.--Results of a least-square linear regression between transmissivities estimated from aquifer tests and specific-capacity data.

Comparison of transmissivities estimated from specific-capacity data with those determined from aquifer-test analyses indicates that estimates determined from specific-capacity data generally are less than those from aquifer-test analyses, although there is considerable scatter (fig. 6). The regression equation used to adjust transmissivities estimated from specific-capacity data is as follows:

$$TRANS = 3.89\ (TSPECAP)^{0.896}\ , \tag{2}$$

where

TRANS = adjusted transmissivity, in feet squared per day, and
TSPECAP = estimated transmissivity from specific-capacity data, in feet squared per day.

The adjusted transmissivity value in equation 2 is increased by a lesser factor as the transmissivity from specific-capacity data values increase. For example: the adjusted transmissivity increases 2.4 times to 240 ft^2/d when the estimate of transmissivity from specific-capacity data is 100 ft^2/d. The adjusted transmissivity increases only 1.29 times to 51,700 ft^2/d when the estimate of transmissivity from specific-capacity data is 40,000 ft^2/d,.

Specific-capacity data from a total of 5,920 wells in the study area were used to calculate transmissivity using equations 1 and 2. Most of the data (5,429) were retrieved from the U.S. Geological Survey's WATSTORE database. The remaining data (491) were in the aquifer-test computer file. Calculations of transmissivity from specific-capacity data in the aquifer-test file were limited to tests of a single pumped well which did not have an estimate of transmissivity from aquifer-test analyses.

Hydraulic conductivity was estimated for each test by dividing the calculated transmissivity value by the length of the perforated interval. Hydraulic conductivity was not estimated for tests that did not have a value for the length of the perforated interval. Thus, the number of tests in which a hydraulic conductivity was estimated was decreased to 5,131. The distribution of tests that resulted in an estimate of hydraulic conductivity is shown in figure 7. Figure 7a shows the distribution of tests where depth intervals of wells correspond to model layers 2 through 10. Figure 7b shows the distribution of tests where depth intervals of wells correspond to model layer 11. The distribution of tests for each model layer generally coincides with the outcrop areas of geohydrologic units shown in figure 3 and in areas somewhat downdip. The greatest number of tests are located in Louisiana and Mississippi.

DISTRIBUTION OF HYDRAULIC CONDUCTIVITIES

Attempts to contour estimates of transmissivity and hydraulic conductivity for each geohydrologic unit proved impossible because of the wide variation in values over short distances and because of uneven distribution of data. A statistically based approach was then used to describe the regional distribution of hydraulic conductivity in the study area and to determine regional values of effective hydraulic conductivity for sand beds.

Estimates of hydraulic conductivities from aquifer tests and specific-capacity data range from less than 1 ft/d to more than 1,000 ft/d. The frequency distribution of hydraulic conductivity is shown in figure 8. A majority of hydraulic conductivities are less than 100 ft/d. The median values are 58 ft/d and 74 ft/d for estimates determined from aquifer test analyses and specific-capacity data, respectively, whereas the mean values are 138 ft/d and 163 ft/d, respectively. Neuman (1982, p. 83) summarized results of several investigators and concluded that the distributions of hydraulic conductivities, and transmissivities in aquifers generally is log normal. Logarithmically transforming the estimates of hydraulic conductivities from both

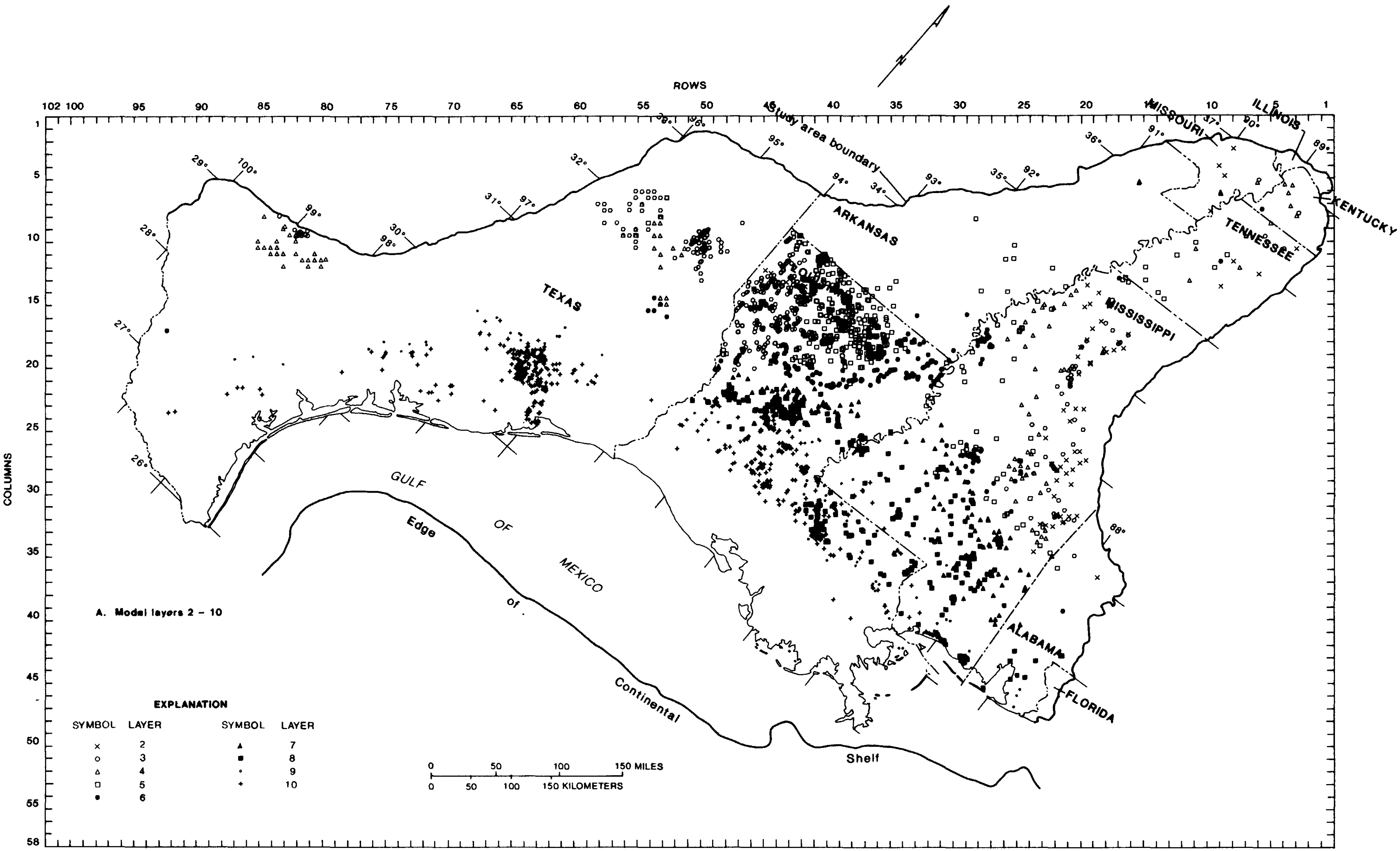


Figure 7a.--Location of wells with specfic-capacity data used to estimate hydraulic conductivity; well depth intervals correspond to model layers 2 - 10.

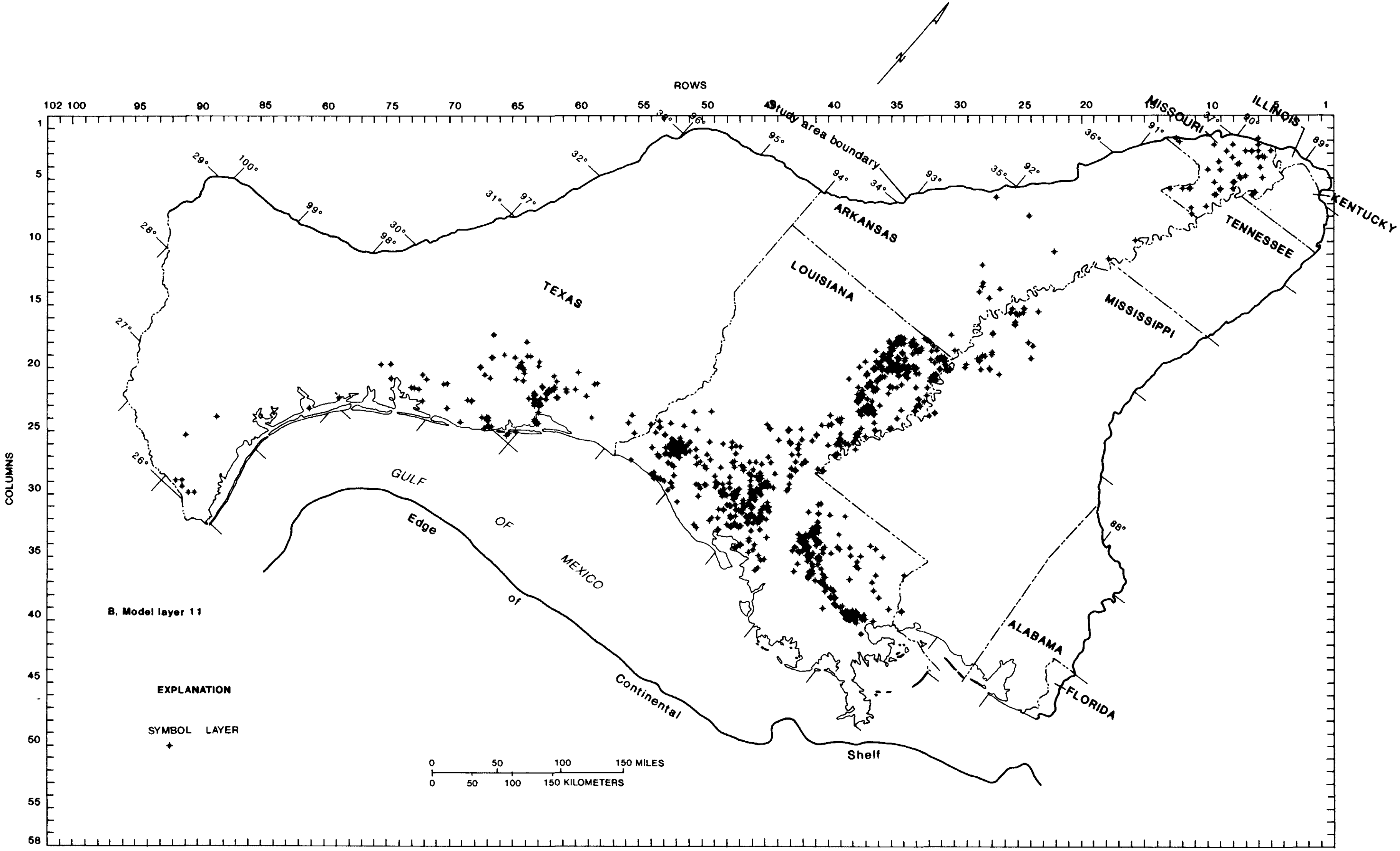
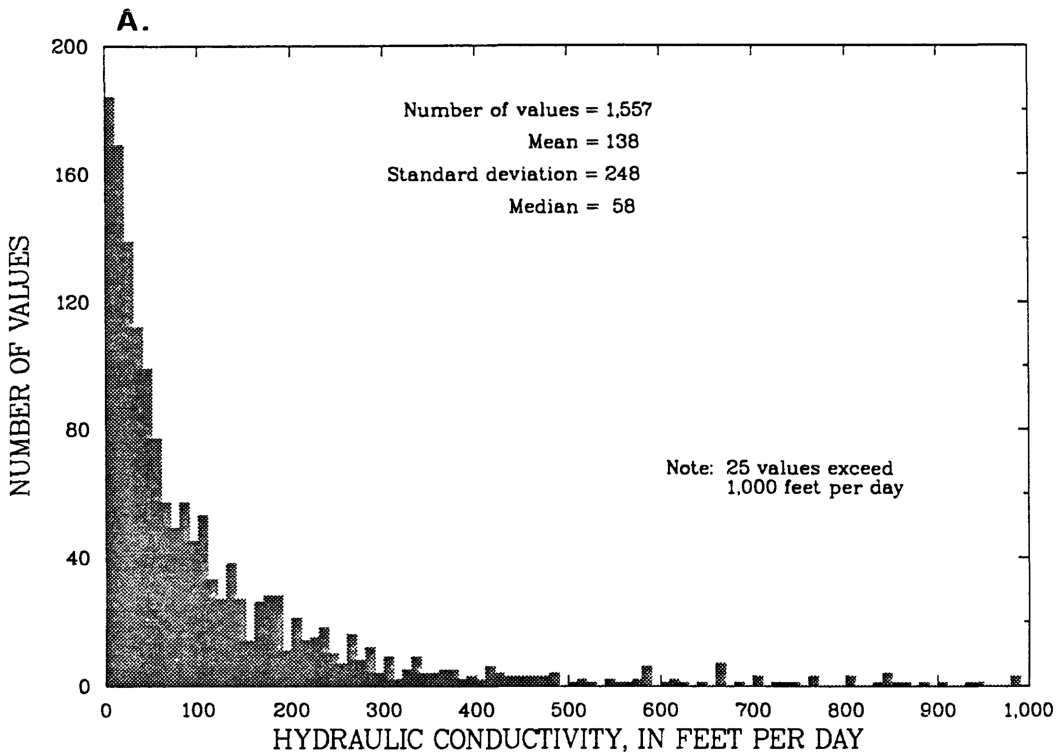


Figure 7b.--Location of wells with specfic-capacity data used to estimate hydraulic conductivity; well depth intervals correspond to model layer 11.



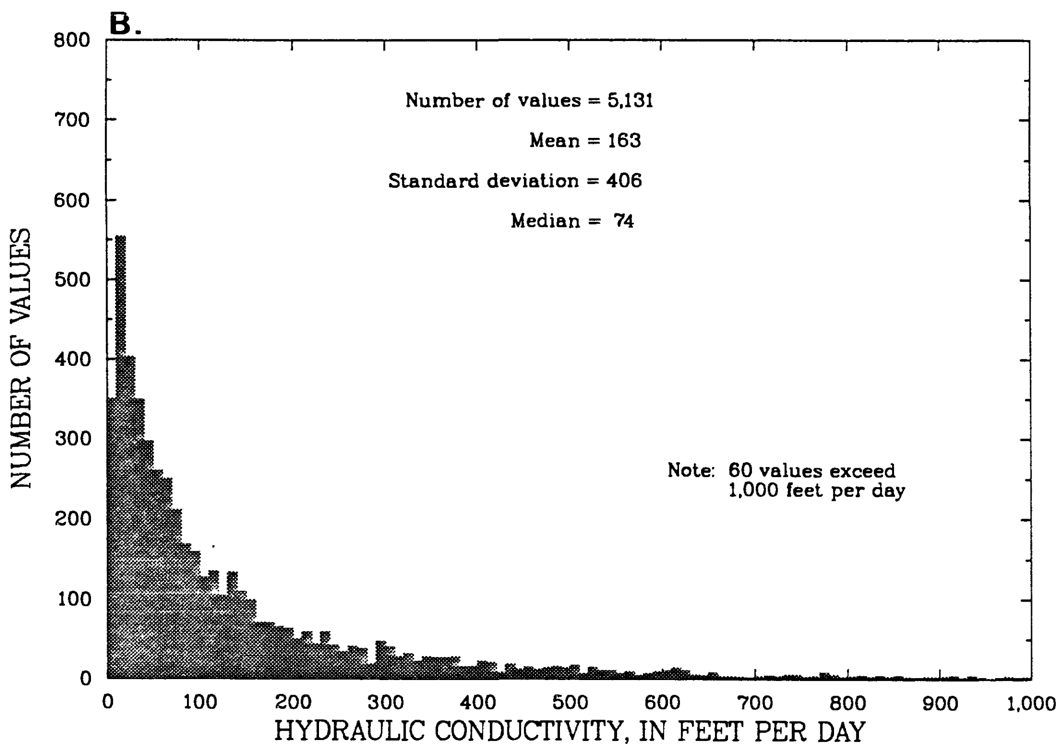


Figure 8.--Frequency distributions of hydraulic conductivities estimated from (a) aquifer-test analyses, and (b) specific-capacity data.

aquifer-test analyses and specific-capacity data resulted in normal distributions (fig. 9). Because the distributions are log normal, geometric means were used to represent a typical hydraulic conductivity from individual values.

The mean of the log-transformed values estimated from 1,557 aquifer-test analyses is 1.74 (units are in $\log_{10}$ ft/d) with a standard deviation of 0.63. The mean of the log-transformed values estimated from 5,131 specific capacities is 1.85 with a standard deviation of 0.58. The geometric mean from aquifer tests is 55 ft/d and from specific capacity is 71 ft/d; less than half the arithmetic means. Excluding estimates of hydraulic conductivities that exceed 1,000 ft/d resulted in a geometric mean from aquifer tests of 52 ft/d and from specific capacities of 68 ft/d. Values that exceed 1,000 ft/d were considered unreasonable for sediments in the study area and were not included in the final analyses. These values are less than 2 percent of the total number of estimates.

Estimates of hydraulic conductivity were first grouped by model layer for the purpose of comparing to parameter estimation results from simulations of regional ground-water flow in which sand hydraulic conductivities were allowed to vary by model layers. The hydraulic conductivities estimated from aquifer-tests analyses and specific capacity data are assumed to be the same as the sand hydraulic conductivities as wells are usually perforated next to the more permeable deposits. Statistics of hydraulic conductivity distribution by model layers are presented in table 2 and summarized in figure 10.

The number of hydraulic conductivity estimates from aquifer-test analyses ranged from 58 in model layer 2 to 231 in model layer 11 (table 2). The number of estimates from specific-capacity data ranged from 78 in layer 2 to 1,514 in layer 11, which was by far the largest group of estimates. Model layer 11 includes aquifers in the Mississippi River alluvial deposits and in the more recent deposits in the coastal lowlands (table 1, fig. 2). The range of hydraulic conductivity estimates from both aquifer-tests analyses and specific-capacity data were similar for each model layer (fig. 10).

The absolute difference in geometric means between the two methods ranged from 0 in model layer 9 to 77 in layer 11. Although estimates of transmissivity from specific-capacity data were adjusted to account for differences in estimates between aquifer-test analyses and specific-capacity data, the adjustment is only based on a comparison of transmissivity from wells that include both estimates. Estimates of hydraulic conductivity from specific-capacity data do not include wells which also have estimates of hydraulic conductivity from aquifer-test analyses. Therefore, a paired T test was performed comparing difference in mean $\log_{10}$ hydraulic conductivity between aquifer-test analyses and specific-capacity data for each layer to determine if the means from the two methods could be from the same population. Results from the paired T test indicate that the $\log_{10}$ means from the two methods could be from the same population, thus, individual estimates from both aquifer-tests analyses and specific-capacity data were combined and statistics generated for all estimates in a model layer (table 2). The statistics of the combined values generally approximate the statistics from specific-capacity data because of the greater number of estimates.

The geometric mean of the combined values is least for model layer 3 (20 ft/d); greatest for layer 11 (156 ft/d), and nearly the same for layers 4 through 10 (43 ft/d to 66 ft/d; fig. 10). The 25 and 75 percentiles and median values are also similar for layers 4 through 10 suggesting, on a regional basis, that these layers may have similar aquifer properties.

Estimates of hydraulic conductivity were also grouped by areas delineated within the study area (fig. 2) and statistics calculated to determine if there were differences in hydraulic conductivity between areas which could be useful in the regional simulation of ground-water flow. Statistics of hydraulic conductivity distribution by areas are presented in table 3 and summarized in figure 11.

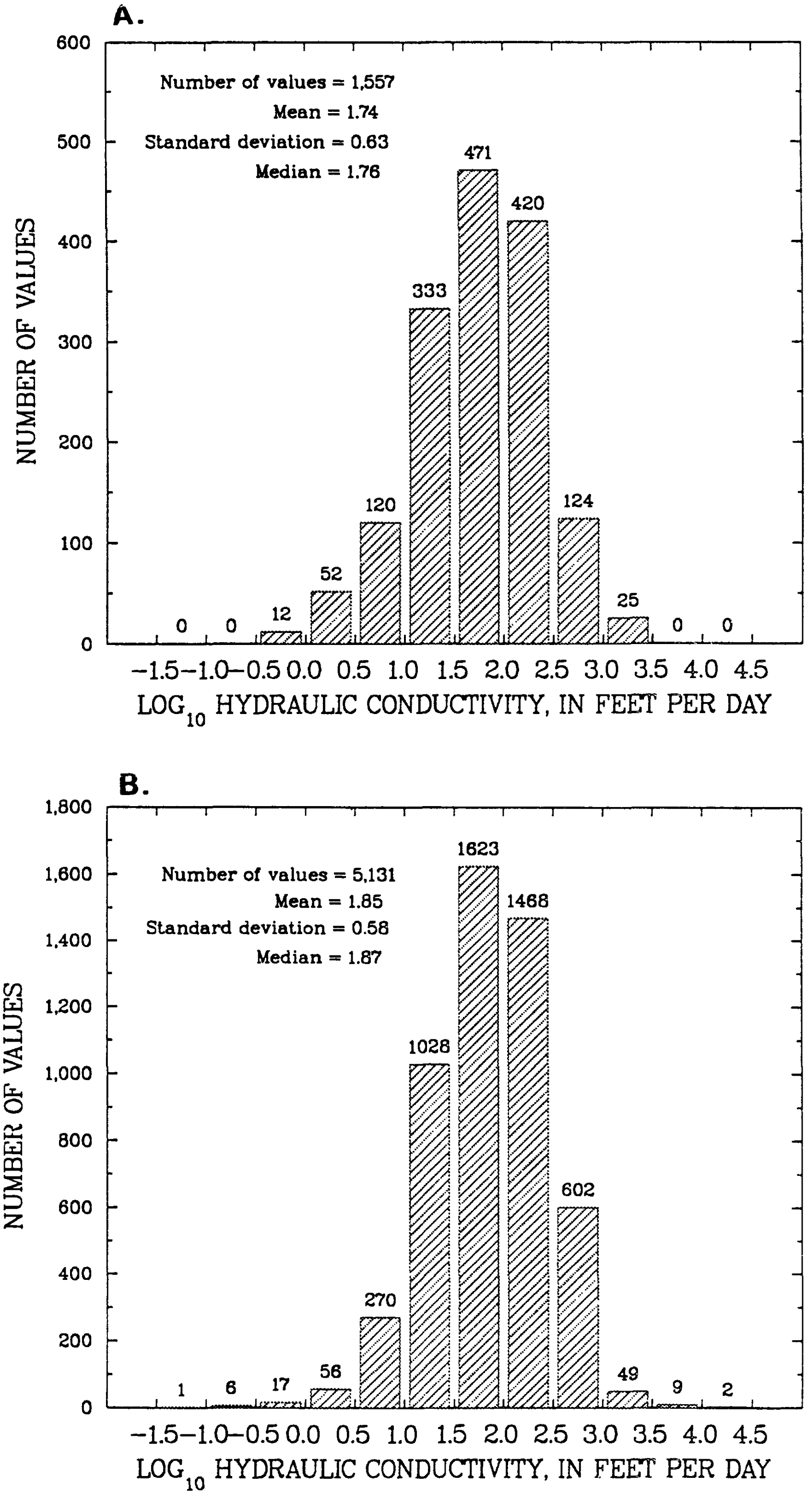


Figure 9.--Frequency distributions of logarithmically (base 10) transformed hydraulic conductivities from (a) aquifer-test analyses, and (b) specific-capacity data.

Table 2.--*Estimates of hydraulic conductivity by model layers in the Gulf Coast Regional Aquifer-System Analysis study area. Values are statistical summaries from aquifer-test analyses and specific-capacity data*

[Values are in feet per day; (AQ)--results from aquifer-test analyses; (SC)--results from specific-capacity data; (COMB)--combined results from both analyses][1]

Source	Number of values	Arith-metic mean	Standard deviation	Har-monic mean	Geo-metric mean	Percentile 1	25	Median	75	99
					Layer 2					
(AQ)	58	158	181	32	95	1.0	60	91	170	720
(SC)	78	129	149	4.2	65	.06	34	77	190	710
(COMB)	136	141	164	6.6	76	.43	44	84	180	720
					Layer 3					
(AQ)	213	43	94	5.2	14	.52	5.6	13	40	710
(SC)	569	48	75	7.4	22	.47	9.9	24	54	430
(COMB)	782	47	81	6.6	20	.50	8.5	20	51	440
					Layer 4					
(AQ)	104	67	70	16	39	.69	20	43	83	390
(SC)	151	82	112	16	46	.69	26	47	88	800
(COMB)	255	76	97	16	43	.84	23	45	84	580
					Layer 5					
(AQ)	167	148	170	20	72	1.3	31	93	190	960
(SC)	609	88	94	18	53	1.6	30	63	110	510
(COMB)	776	101	117	18	57	1.5	30	66	120	590
					Layer 6					
(AQ)	99	104	110	15	55	.58	27	66	130	560
(SC)	232	81	92	27	51	2.1	29	52	97	590
(COMB)	331	88	99	22	52	1.5	28	55	110	550
					Layer 7					
(AQ)	115	137	163	35	77	1.7	37	79	160	850
(SC)	317	85	90	19	53	.85	26	58	110	520
(COMB)	432	99	116	22	59	1.6	30	65	120	620
					Layer 8					
(AQ)	146	144	173	26	78	1.3	36	94	170	960
(SC)	408	97	108	34	62	3.3	36	69	120	660
(COMB)	554	109	130	31	66	1.6	36	74	130	750
					Layer 9					
(AQ)	211	102	134	20	50	1.7	24	50	130	840
(SC)	511	94	120	25	50	2.4	18	54	130	630
(COMB)	722	97	124	23	50	2.3	19	53	130	720

Table 2.--*Estimates of hydraulic conductivity by model layers in the Gulf Coast Regional Aquifer-System Analysis study area. Values are statistical summaries from aquifer-test analyses and specific-capacity data*--Continued

Source	Number of values	Arith-metic mean	Standard deviation	Har-monic mean	Geo-metric mean	Percentile 1	Percentile 25	Percentile Median	Percentile 75	Percentile 99
					Layer 10					
(AQ)	188	100	151	26	49	2.0	23	42	120	840
(SC)	682	94	113	29	52	4.9	21	47	120	510
(COMB)	870	95	122	28	51	4.7	21	46	120	640
					Layer 11					
(AQ)	231	164	181	45	92	4.4	39	99	230	880
(SC)	1514	256	208	86	169	9.2	94	202	360	910
(COMB)	1745	243	207	77	156	8.7	82	186	350	910
					ALL VALUES					
(AQ)	1532	115	153	16	52	1.2	22	57	140	800
(SC)	5071	136	163	21	68	2.3	29	73	170	790
(COMB)	6603	131	161	20	64	1.7	27	70	170	790

1 Analyses do not include estimates of hydraulic conductivity that exceed 1,000 feet per day. Values exceeding 1,000 feet per day are considered unreasonable for sediments in the study area. A total of 25 hydraulic conductivities estimated from aquifer-test analyses exceed 1,000 feet per day whereas 60 values estimated from specific-capacity data exceed 1,000.

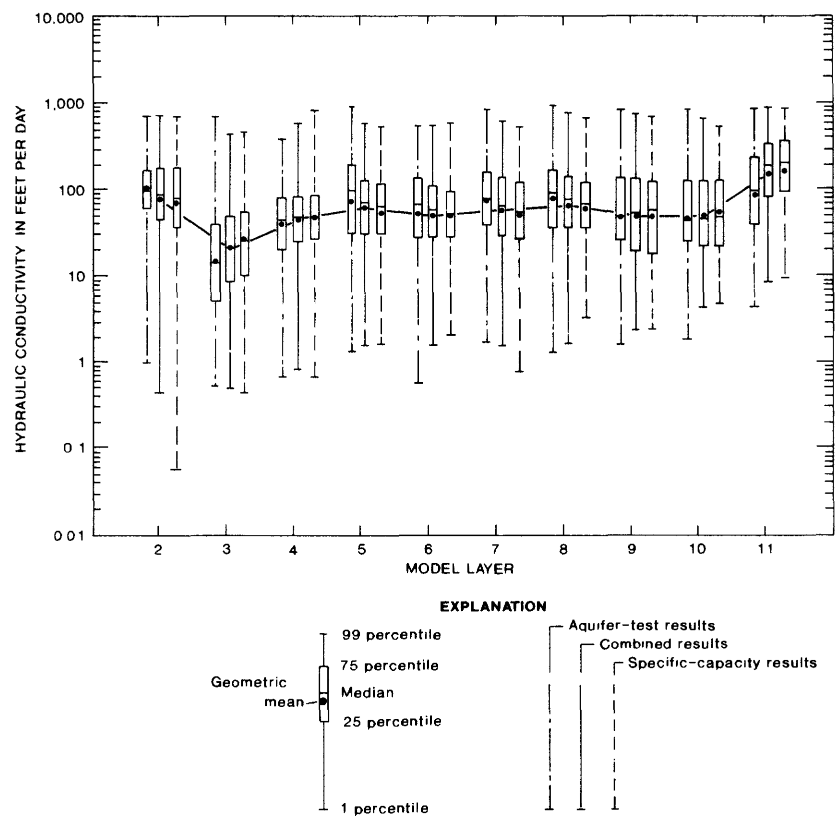


Figure 10 -- Variation among model layers of hydraulic conductivities determined from aquifer-test analyses and specific-capacity data

Table 3.--*Estimates of hydraulic conductivity by areas in the Gulf Coast Regional Aquifer-System Analysis study area. Values are statistical summaries from aquifer-test analyses and specific-capacity data.*

[Values are in feet per day; (AQ)--results from aquifer-test analyses; (SC)--results from specific-capacity data; (COMB)--combined results from both analyses][1]

Areas	Number of values	Arithmetic mean	Standard deviation	Harmonic mean	Geometric mean	Percentile 1	Percentile 25	Percentile Median	Percentile 75	Percentile 99
1										
(AQ)	23	46	59	5.6	18	0.61	7.8	17	84	220
(SC)	43	47	49	7.2	25	.39	13	33	50	180
(COMB)	66	47	52	6.5	22	.39	9.8	28	54	220
2										
(AQ)	185	19	25	5.0	10	.57	5.3	10	23	170
(SC)	177	27	27	10	18	.79	9.6	17	37	140
(COMB)	362	23	26	6.6	14	.58	6.9	14	29	140
3										
(AQ)	185	113	170	11	45	.63	20	50	130	950
(SC)	869	66	73	15	37	1.1	20	44	82	370
(COMB)	1,054	74	99	14	39	.97	20	46	88	530
4										
(AQ)	106	184	148	91	133	16	74	134	260	870
(SC)	641	257	218	68	157	5.9	71	203	370	910
(COMB)	747	246	211	71	153	7.0	72	194	360	910
5										
(AQ)	148	148	146	39	94	1.9	58	100	190	710
(SC)	210	119	145	7.1	57	.32	23	68	160	780
(COMB)	358	131	146	11	70	.97	36	83	180	740
6										
(AQ)	43	24	63	6.0	10	1.5	3.7	11	18	410
(SC)	29	102	189	8.3	24	1.6	6.1	18	88	670
(COMB)	72	56	134	6.8	14	1.5	4.8	12	30	670
7										
(AQ)	340	53	74	19	33	1.8.	19	35	57	440
(SC)	465	39	52	19	26	4.3	14	21	39	230
(COMB)	805	45	63	19	28	2.7	15	26	50	390
8										
(AQ)	166	205	218	51	120	4.9	56	135	240	930
(SC)	1,330	178	181	49	104	4.9	49	114	240	850
(COMB)	1,496	181	185	49	106	5.2	49	116	240	860
9										
(AQ)	335	166	160	62	111	8.5	62	120	200	830
(SC)	1,302	135	142	21	81	1.6	43	94	160	750
(COMB)	1,637	141	147	25	86	2.2	48	98	170	770

[1] Hydraulic conductivities exceeding 1,000 feet per day were not included in the analyses. Five values from specific-capacity data and one value from aquifer-test analyses were assigned to area 10 which is the area just offshore from the coastline. Either the locations of these wells are slightly in error or the wells were drilled on small islands.

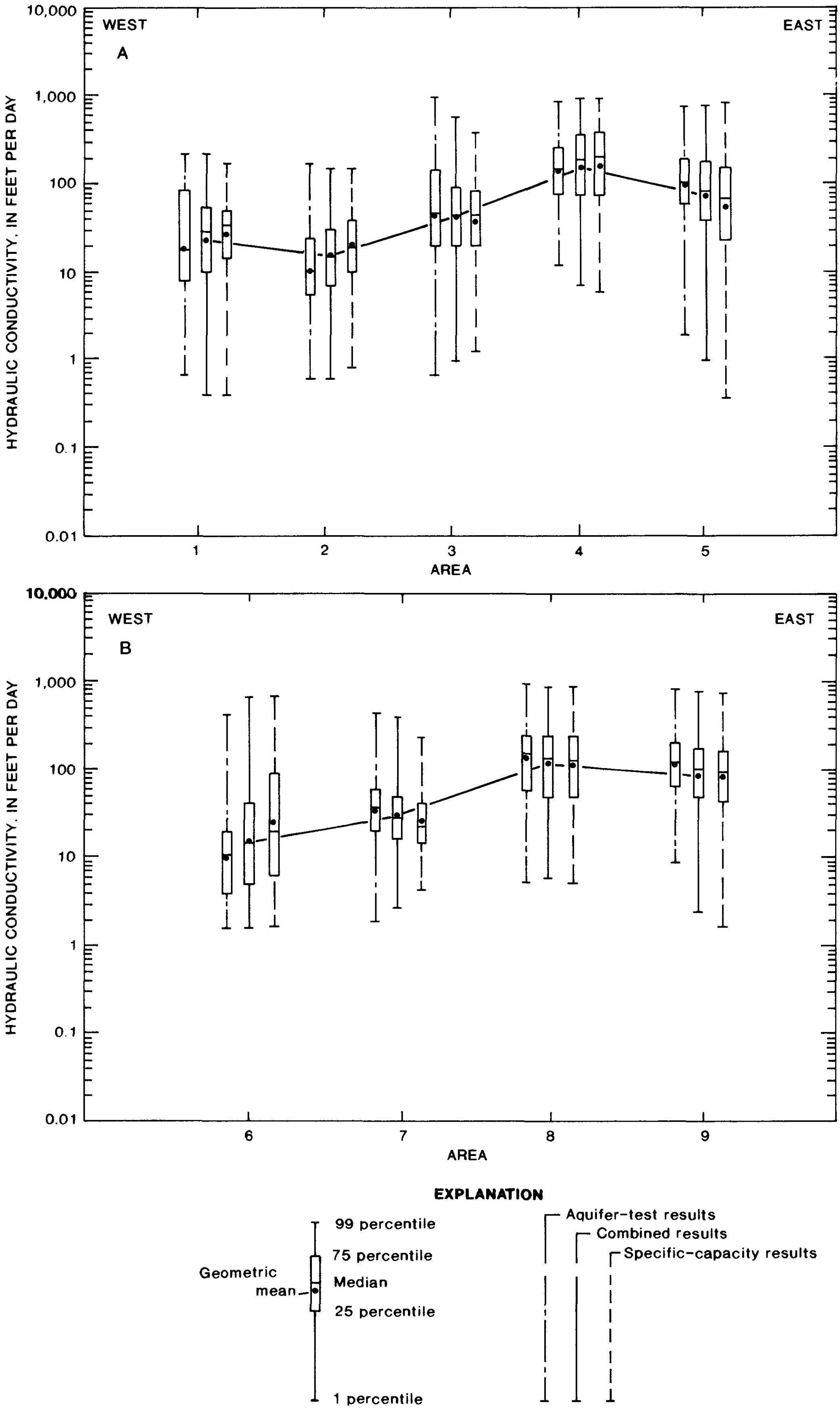


Figure 11.--Variation among (A) areas 1-5 and (B) areas 6-9 of hydraulic conductivities from aquifer-test analyses and specific-capacity data.

The number of hydraulic-conductivity estimates from aquifer-test analyses ranged from 23 in area 1 to 340 in area 7. The number of estimates from specific-capacity data ranged from 29 in area 6 to 1,330 in area 8. About 50 percent of all the estimates from specific capacity data are in areas 8 and 9, which includes southern Louisiana and Mississippi.

The range of hydraulic conductivity from both aquifer-test analyses and specific-capacity data were similar for all areas (fig. 11). In some areas (for example, areas 3 and 7), the range in hydraulic conductivity from aquifer-test analyses is greater than the range in hydraulic conductivity from specific capacity and in other areas (for example, areas 5 and 9), the range in hydraulic conductivity from specific-capacity data is greater.

The absolute difference in geometric mean hydraulic conductivity between aquifer-test analyses and specific-capacity data ranged from 7 in areas 1 and 7 to 37 in area 5. A paired T test comparing the difference in mean $\log_{10}$ of hydraulic conductivity between aquifer-test analyses and specific-capacity data for each area indicates the $\log_{10}$ means from the two methods could be from the same population. Thus, individual estimates of hydraulic conductivity from both methods were combined and statistics calculated for all the estimates in an area (table 3).

In general, variations in the geometric mean and median between areas from the combined estimates were as large as variations in the geometric mean and median between layers (compare figs. 10 and 11). The geometric mean of the combined values generally increased from areas in the western part of the study area to areas in the eastern part (fig. 12). Similar trends were observed between geometric mean and median values calculated from both aquifer-test analyses and specific-capacity data (table 3 and fig. 11).

Because the variation in geometric mean hydraulic conductivity for the combined estimates from aquifer-test analyses and specific-capacity data was as much by areas as by model layers, a two-way analyses of variance procedure (Steele and Torrie, 1980, p. 146-171) was performed on the $\log_{10}$ transformed hydraulic conductivities to determine if the main factors of area and layer and the interactions of these factors (combinations of layers and areas) were significant in explaining the variation in hydraulic conductivity. Results from this test (table 4) indicate that the variation of means for the combination of areas and layers are significant at a probability level of 0.001. Therefore, statistical comparisons of means must be made on the combinations of layers and areas and not on the basis of layers or areas alone.

To compare the significance between means of layers within areas, each layer and area combination was categorized for all $\log_{10}$ transformed hydraulic conductivities and a one-way analyses of variance procedure performed on values within each category. Means for each combination of layer and area were compared using the Duncan multiple-range test (Steele and Torrie, 1980, p. 187-188) at the probability level of 0.05. Results from the Duncan multiple-range test are listed in table 5. This comparison indicates that the mean $\log_{10}$ hydraulic conductivity in model layer 11 of area 4 (alluvial deposits of the Mississippi River) is significantly higher than all other means except for layer 11 in areas 8 and 9, layer 2 in area 4, and layer 5 in area 5. The lowest means ($\log_{10}$ hydraulic conductivity less than 1.0) were from model layer 3 in areas 1 and 9; and layer 9 in area 6.

The distribution of geometric mean hydraulic conductivity for all combinations of layers and areas is illustrated in figure 13. Means of layers within areas are excluded when the number of hydraulic conductivity estimates are less than 5. General statistics for each combination of layers and areas are given in table 6. The general trend of increasing geometric means within layers from areas in the western part of the study area (fig. 13) to the east is similar to the overall trend when all estimates of hydraulic conductivity are grouped by areas (fig. 12). The trend is most notable in model layer 9 where the geometric mean increases from 6 ft/d in area 6 to 85 ft/d in area 9.

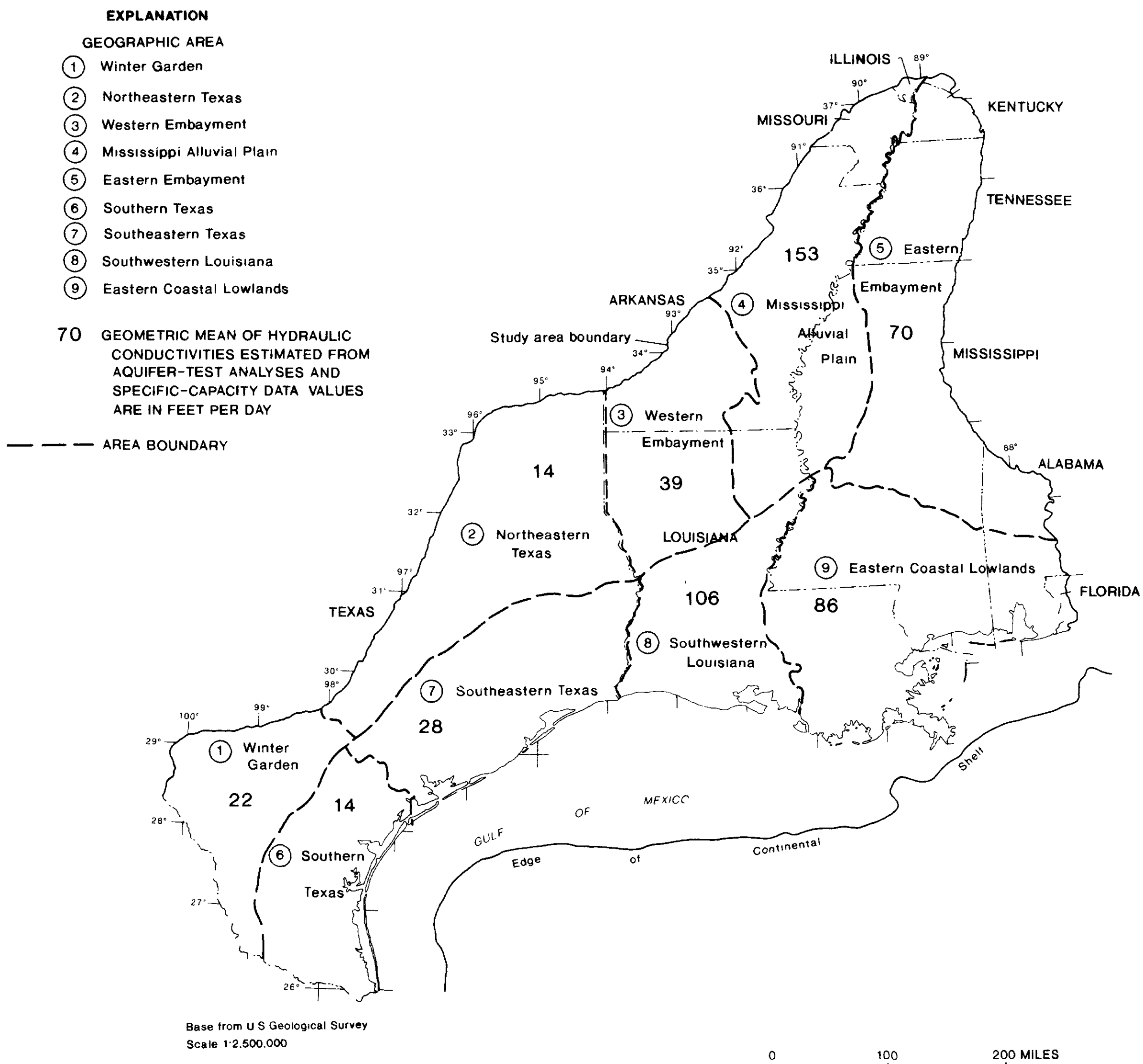


Figure 12.--General trend of geometric means of individual hydraulic conductivities estimated from aquifer tests and specific-capacity data by geographic area.

Table 4.--*Analyses of variance for log_{10} hydraulic conductivity for combined estimates from aquifer-test analyses and specific-capacity data as influenced by layer and area*

Source of of variation	Degrees of freedom	Sum of squares	Mean square	F value
Model (overall)	52	869.286	16.717	82.67***
Error	6,544	1,323.288	0.202	
Corrected total	6,596	2,192.573	---	
By model components				
Area	8	565.488	70.686	349.56***
Layer	9	257.714	28.635	141.61***
Area by layer	35	46.084	1.314	6.51***

*** F value significant at the probability level of 0.001

Model layer 11 represents the alluvial deposits of the Mississippi River as well as the more recent deposits along the Gulf Coast (table 1 and fig. 3). The Mississippi River alluvial deposits are represented in area 4, whereas the coastal sediments of layer 11 are represented in areas 6 and 7. Areas 8 and 9 include both the alluvial deposits and the coastal sediments. The hydraulic conductivity of the coastal sediments in model layer 11 may be considerably less than those of the Mississippi River alluvial deposits, as indicated by the differences in the geometric means of layer 11 in areas 6 and 7 (69 ft/d and 49 ft/d, respectively) to the mean in area 4 (316 ft/d). The results in table 5 indicate these differences are significant. The somewhat smaller geometric means in areas 8 and 9 (189 ft/d and 132 ft/d, respectively) compared to area 4 may be the result of combining the coastal sediments with the Mississippi River alluvial deposits. However, the smaller values in areas 8 and 9 may also be caused by the sediments becoming finer grained as the Mississippi River approaches its delta.

Relation of Hydraulic Conductivity to Depth

The hydraulic conductivity of an unconsolidated sediment should decrease with depth as a result of sediment compaction caused by increasing overburden pressures. The relation between hydraulic conductivity and effective stress (effective overburden pressure or grain to grain load) is discussed by Helm (1976, p. 378-379.)

An attempt was made to relate estimates from aquifer-test analyses and specific-capacity data to the depth of the middle of the perforated or screened interval of the wells. Thirty-seven estimates of hydraulic conductivity could not be related to depth because there were no data on the depth of the well or the depth to the top of the perforated interval. The distribution of depth below land surface to the middle of the perforated interval is shown in figure 14 for the 6,500 wells that also had an estimate of hydraulic conductivity. Depth to the middle of the perforated interval was less than 800 ft for about 80 percent of the wells. The mean was 580 ft and the standard deviation was 510 ft.

Table 5.--*Number of observations, means of log_{10} hydraulic conductivity, and standard deviation for combined estimates of hydraulic conductivity from aquifer-test analyses and specific-capacity data by layers and areas*

[The letters following the mean are used to indicate means that are not significantly different as determined from a Duncan multiple-range test (Steele and Torrie, 1980, p. 187-188). Means with the same letter are not significantly different at a probability level of 0.05. Dashes indicate layer and area combinations that do not exist.]

Layer	Area: West				East	West			East
	1	2	3	4	5	6	7	8	9
11	---	---	3	374	---	16	231	601	515
			1.67b-j	2.50a		1.84b-g	1.69b-j	2.28ab	2.12a-c
			0.0	0.29		0.80	0.43	0.42	0.40
10	---	---	---	---	---	27	330	293	219
						1.84k-p	1.41e-n	1.95b-e	1.91b-e
						.39	.31	.47	.45
9	---	---	---	---	---	23	213	152	334
						.76o-q	1.28g-p	1.91b-e	1.93b-e
						.41	.30	.47	.44
8	---	1	---	---	---	6	25	196	326
		0.18r				1.22h-p	1.47d-n	1.77b-i	1.89b-f
						.23	.53	.47	.43
7	---	---	---	---	---	---	4	242	186
							1.17i-p	1.79b-h	1.74b-i
							.75	.45	.51
6	1	22	73	174	32	---	2	11	16
	1.00m-p	1.04k-p	1.61c-l	1.89b-f	1.78b-h		.35qr	1.38e-n	1.52c-m
	---	.45	.42	.39	0.49		.21	.29	.49
5	4	32	492	147	71	---	---	1	29
	0.34qr	1.14j-p	1.76b-i	1.87b-g	2.03a-d			.72o-r	1.29f-o
	.54	.35	.46	.45	.51			---	.94
4	44	83	18	31	78	---	---	---	1
	1.62c-k	1.46d-n	1.49d-n	1.68b-j	1.85b-g				0.34qr
	.42	.40	.60	.47	.54				---
3	17	224	458	3	70	---	---	---	10
	.92n-p	1.02l-p	1.41e-n	1.20h-p	1.61c-l				.69p-r
	.60	.47	.59	.77	.60				.84
2	---	---	10	18	107	---	---	---	1
			1.40e-n	2.11a-c	1.90b-e				.97m-p
			.45	.37	.59				---

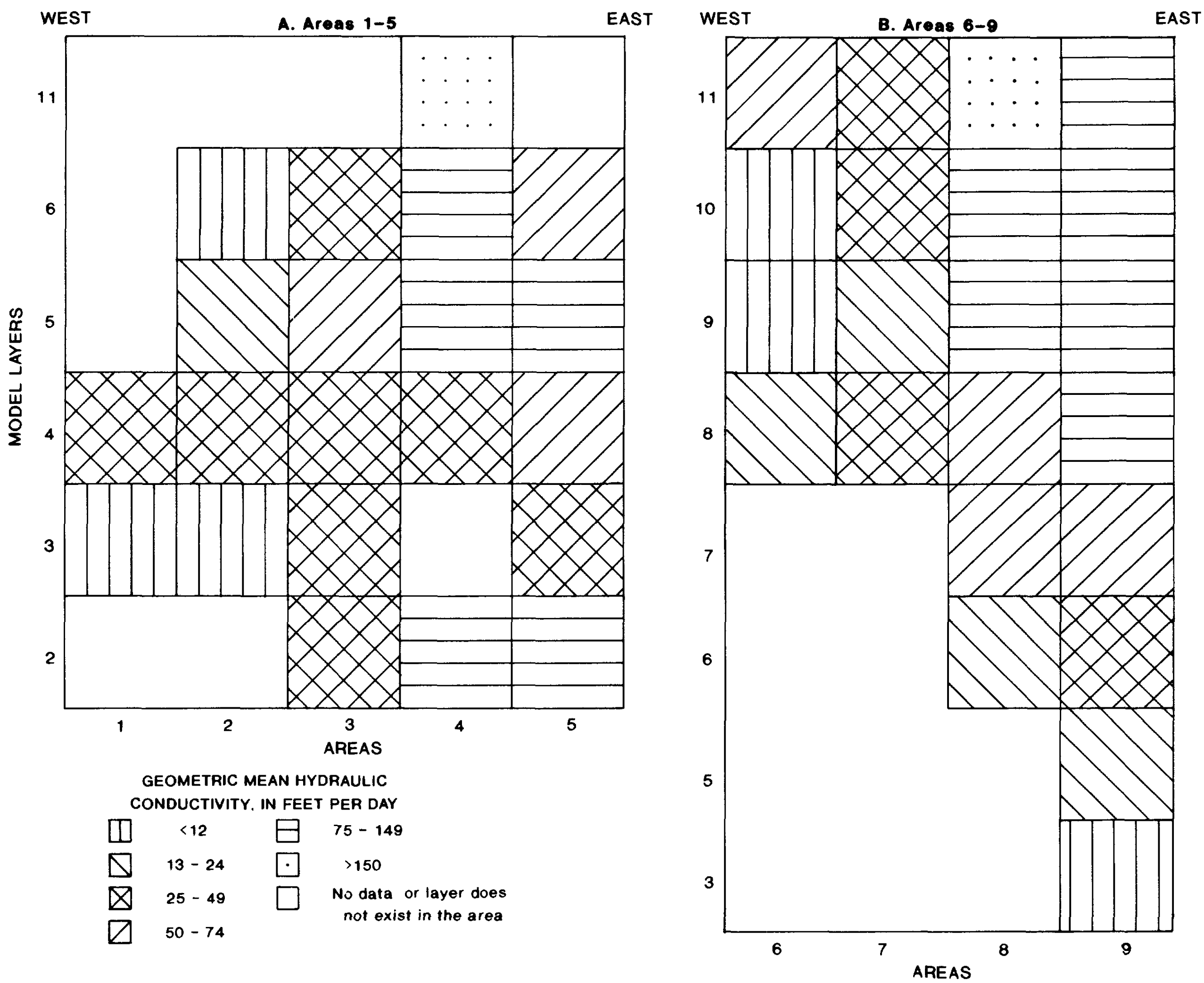


Figure 13.--Distribution of the geometric mean of individual hydraulic conductivities from aquifer tests and specific-capacity data by model layers within (a) areas 1-5 and (b) areas 6-9.

Table 6.--*Estimates of hydraulic conductivity by model layer and area for the Gulf Coast Regional Aquifer-System Analysis study area. Statistical summaries are the result of combining the aquifer-test analyses with analyses from specific-capacity data.*

[All values are in feet per day.[1] Summaries are not included for layers in an area with less than five estimates of hydraulic conductivity.]

Model layers	Number of values	Arithmetic mean	Standard deviation	Harmonic mean	Geometric mean	Percentile 1	Percentile 25	Percentile Median	Percentile 75	Percentile 99
Area 1										
3	17	17	23	3.2	8.3	0.4	4.2	6.4	26	92
4	44	63	56	25	42	2.9	21	41	93	220
Area 2										
3	224	18	24	5.5	10	0.5	5.5	11	20	160
4	83	40	31	14	29	0.7	21	34	54	230
5	32	19	17	10	14	3.0	6.4	16	21	75
6	22	16	15	5.3	11	0.6	8.0	13	20	58
Area 3										
2	10	42	54	17	25	5.6	11	19	56	190
3	458	56	85	8.7	26	0.6	11	31	64	470
4	18	47	29	5.3	31	0.4	30	42	66	110
5	492	94	112	30	57	3.0	31	64	110	590
6	73	64	75	24	41	2.0	27	40	65	450
Area 4										
2	18	187	193	95	128	38	70	118	220	710
4	31	76	73	24	48	2.8	26	44	96	320
5	147	119	123	41	75	5.3	41	78	160	680
6	174	116	113	51	78	8.3	44	82	140	620
11	374	376	203	227	316	20	230	340	480	970
Area 5										
2	107	145	162	5.4	79	0.1	54	85	180	720
3	70	92	137	9.8	40	0.2	17	40	99	790
4	78	129	146	25	71	1.0	35	79	190	830
5	71	170	142	42	106	3.2	69	131	230	630
6	32	90	73	19	60	0.9	34	81	120	360

Table 6.--*Estimates of hydraulic conductivity by model layer and area for the Gulf Coast Regional Aquifer-System Analysis study area. Statistical summaries are the result of combining the aquifer-test analyses with analyses from specific-capacity data*--Continued

Model layers	Number of values	Arith-metic mean	Standard deviation	Har-monic mean	Geo-metric mean	Percentile 1	Percentile 25	Percentile Median	Percentile 75	Percentile 99
					Area 6					
8	6	19	9.1	15	17	7.3	11	17	29	30
9	23	8.6	7.2	3.9	5.8	1.5	2.7	7.4	14	27
10	27	17	18	8.1	12	1.9	6.4	11.4	18	71
11	16	201	235	15	69	2.7	25	88	380	670
					Area 7					
8	25	50	43	13	30	1.6	14	40	68	150
9	213	24	18	15	19	2.5	13	18	31	100
10	330	35	39	20	26	6.2	17	24	39	230
11	231	79	96	30	49	4.3	26	55	96	610
					Area 8					
6	11	28	13	19	24	6.9	12	34	37	46
7	242	103	121	36	62	4.2	31	65	130	620
8	196	97	116	26	59	1.6	34	69	110	790
9	152	141	168	41	82	3.5	46	78	150	900
10	293	147	154	48	88	6.8	42	94	210	840
11	601	270	207	103	189	9.8	120	221	360	920
					Area 9					
3	10	13	14	0.9	4.9	0.2	0.9	8.3	22	42
5	29	65	70	1.9	20	.2	6.0	38	130	250
6	16	49	38	14	33	1.6	28	36	77	150
7	186	95	110	15	56	.5	28	66	110	680
8	326	123	140	45	78	6.7	44	84	150	760
9	334	129	121	45	85	5.0	47	98	170	750
10	219	126	122	39	82	2.9	42	100	160	710
11	515	194	175	84	132	12	72	136	250	870

[1] Hydraulic conductivites exceeding 1,000 feet per day were not included in the analyses. Five values from specific-capacity data and one value from aquifer-test analyses were assigned to area 10 which is the area offshore from the coastline. Either the locations of these wells are slightly in error or the wells were drilled on small islands.

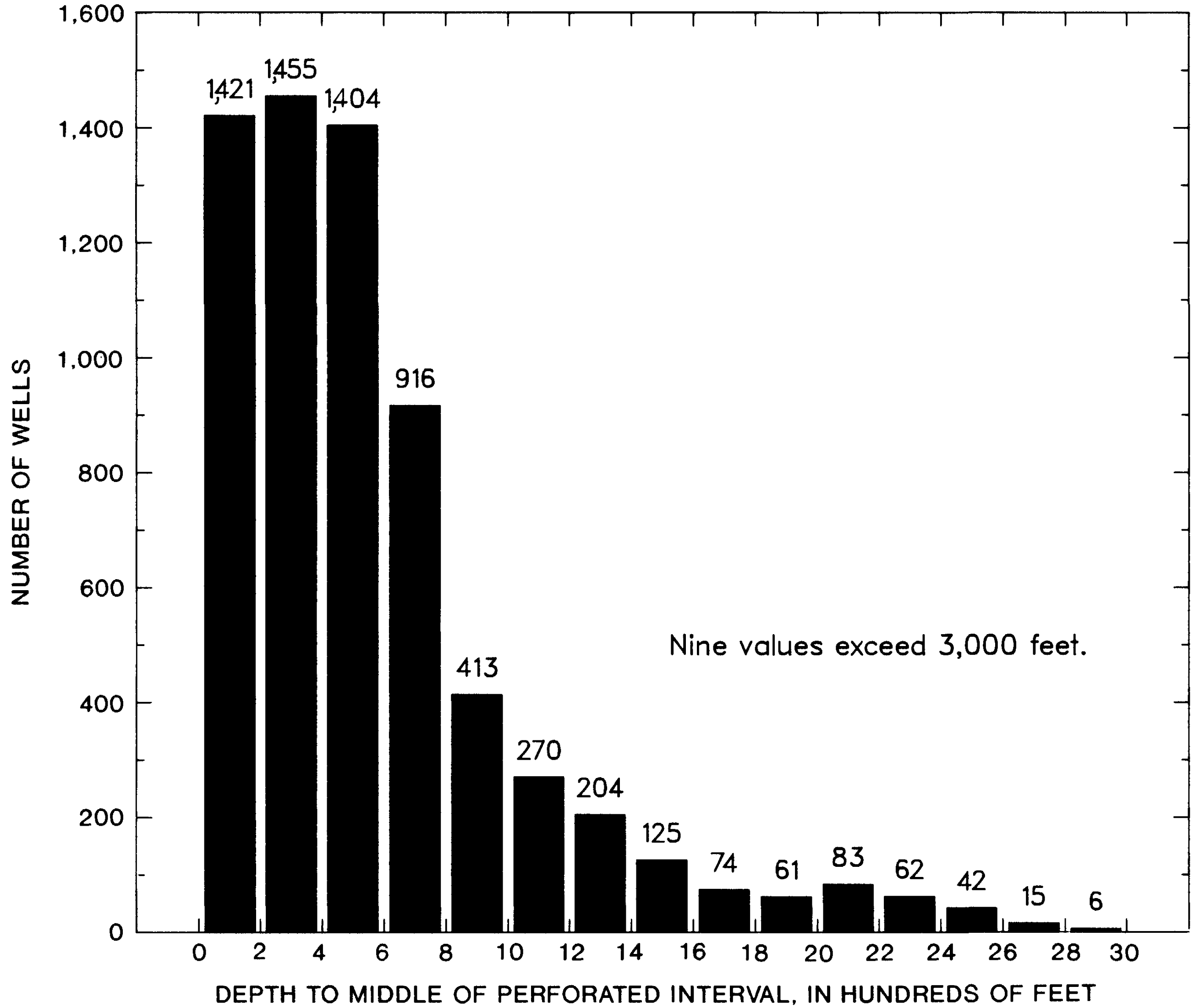


Figure 14.--Distribution of depths to middle of perforated or screened interval of wells from which hydraulic conductivities were determined from aquifer-test analyses and specific-capacity data.

An analyses of covariance was performed on the estimates of hydraulic conductivity in which area and model layers were the main factors and depth to the middle of perforated interval was the covariate. Two separate analyses were done on the $\log_{10}$ transformed hydraulic conductivities; one used non-transformed depths, the other used $\log_{10}$ transformed depths. Results from the analyses of covariance indicate that area, layer, depth, and the interactions (area by layer, depth by area, depth by layer, depth by layer and area) were all significant at the probability level of 0.02. The regression model using $\log_{10}$ transformed depths was not better than the model that used non-transformed depths. The coefficient of determination (r-squared values) for the model using non-transformed depths is 0.423, whereas the coefficient of determination for the model using $\log_{10}$ transformed depths is 0.426. Therefore, non-transformed depths were used for the analyses of covariance.

Equations relating hydraulic conductivity to depth are listed in table 7 for each layer and area combination. Equations are not listed for combinations in which there were fewer than 10 estimates. Included in the table is the number of values with depth estimates and the range in depths. Because of the poor fit of the regression model in the analyses of covariance, the values listed in table 7 only show general trends. Actual hydraulic conductivities may vary considerably from those calculated by the regression equations. The results, however, indicate that for a majority of layer and area combinations in which there were more than 10 hydraulic conductivity estimates, hydraulic conductivity decreased as depth increased (constant in equations listed in table 7 are divided by an exponential value).

In 11 of 41 equations, hydraulic conductivity increases with depth (constants in equations listed in table 7 are multiplied by an exponential value), and is constant in one equation (layer 10, in area 9). For 6 of the 11 equations, in which hydraulic conductivity increases with depth the increase is a factor of 2 or less over the range of data.

Although the analyses of covariance of $\log_{10}$ transformed hydraulic conductivity indicates that depth may be a significant factor in the variation of hydraulic conductivity, the estimates can still vary considerably at a given depth in a model layer and area. The large variations may be caused by other factors not accounted for in this analysis.

Relation of Hydraulic Conductivity to Sand Thickness

Payne (1968, 1970, and 1975) described the geohydrology of major sands (Cockfield and Yegua Formations, Sparta Sand, Carrizo and Meridian Sands) within the Claiborne Group. These units generally correspond to the upper Claiborne (model layer 6), middle Claiborne (model layer 5) and lower Claiborne-upper Wilcox (model layer 4) geohydrologic units in table 1. In his discussion of the permeability and transmissivity of each major sand unit, Payne compared the thickness of the sand section in which the well was screened to the hydraulic conductivity estimated from aquifer tests and specific-capacity data. He noted (1968, p. 5) that for sands deposited in stream channels, the hydraulic conductivity varied directly with the sand thickness. He further stated that such a relation was reasonable because the thicker sands were deposited where flow was persistent and where flow velocities were sufficient to produce a generally better sorted and coarser sand than those deposited along the margins of a channel or in the floodplain. Similarly, Fogg (1986) discussed the complexities of hydraulic conductivity variation within Wilcox Group (includes layers 2 through 4). Fogg concluded that the channel-fill sands within the Wilcox Group were more permeable and continuous than sands deposited in the adjacent floodplain and interchannel basins.

Table 7.--*Relation of hydraulic conductivity estimated from aquifer-test analyses and specific-capacity data to depth, in feet below land surface, to middle of perforated or screened interval by area and model layer*

[Abbreviations: K, hydraulic conductivity in feet per day; D, depth below land surface, in feet, to middle of perforated or screened interval. Equations are not included for area and layer combinations that had fewer than 10 estimates of hydraulic conductivity and depth. Results are based on analyses of covariance where area and model layers were the main factors and depth was the covariate.]

Layer	Equation [1/]	Number of estimates	Range in depth to middle of perforated interval (feet)
		Area 1	
3	$K = 8.7/10^{0.00022D}$	17	34 - 3,536
4	$K = 110/10^{0.00030D}$	44	105 - 3,890
		Area 2	
3	$K = 9.1/10^{0.00010D}$	224	67 - 2,200
4	$K = 15(10^{0.00044D})$	83	91 - 1,370
5	$K = 20/10^{0.00030D}$	31	49 - 1,560
6	$K = 19/10^{0.00046D}$	22	66 - 1,810
		Area 3	
2	$K = 84/10^{0.00125D}$	10	178 - 700
3	$K = 35/10^{0.00057D}$	456	35 - 2,180
4	$K = 18(10^{0.00074D})$	18	66 - 685
5	$K = 58/10^{0.00002D}$	490	45 - 928
6	$K = 40(10^{0.00001D})$	73	28 - 900
		Area 4	
2	$K = 318/10^{0.00037D}$	18	96 - 1,710
4	$K = 47/10^{0.00004D}$	31	85 - 1,920
5	$K = 53/10^{0.00022D}$	146	253 - 1,780
6	$K = 66(10^{0.00022D})$	174	85 - 615
11	$K = 398/10^{0.00192D}$	363	20 - 162
		Area 5	
2	$K = 81/10^{0.00021D}$	107	110 - 2,330
3	$K = 47/10^{0.00013D}$	69	88 - 1,520
4	$K = 102/10^{0.00034D}$	78	47 - 1,440
5	$K = 109/10^{0.00004D}$	71	30 - 1,290
6	$K = 47(10^{0.00021D})$	32	73 - 766

Table 7.--*Relation of hydraulic conductivity estimated from aquifer-test analyses and specific-capacity data to depth, in feet below land surface, to middle of perforated or screened interval by area and model layer*--Continued

Layer	Equation [1]	Number of estimates	Range in depth to middle of perforated interval (feet)
		Area 6	
9	$K = 12/10^{0.00037D}$	23	280 - 684
10	$K = 24/10^{0.00054D}$	27	228 - 803
11	$K = 200/10^{0.00212D}$	16	44 - 475
		Area 7	
8	$K = 23(10^{0.00024D})$	22	184 - 1,600
9	$K = 13(10^{0.00015D})$	212	60 - 2,140
10	$K = 31/10^{0.00011D}$	325	66 - 1,345
11	$K = 74/10^{0.00051D}$	229	59 - 680
		Area 8	
6	$K = 53/10^{0.00070D}$	11	293 - 850
7	$K = 65/10^{0.00004D}$	242	26 - 2,760
8	$K = 51(10^{0.00009D})$	195	38 - 2,400
9	$K = 135/10^{0.00042D}$	152	62 - 1,700
10	$K = 184/10^{0.00081D}$	293	82 - 1,470
11	$K = 266/10^{0.00044D}$	596	28 - 969
		Area 9	
3	$K = 1.18 \times 10^5/10^{0.00148D}$ [2]	10	2,510 - 3,530
5	$K = 459/10^{0.00100D}$ [2]	29	740 - 2,520
6	$K = 54/10^{0.00029D}$	16	145 - 1,160
7	$K = 46(^{0.00010D})$	186	98 - 2,680
8	$K = 74(10^{0.00002D})$	325	77 - 2,940
9	$K = 92/10^{0.00003D}$	333	333 - 2,700
10	$K = 81$ [3]	219	50 - 2,050
11	$K = 209/10^{0.00053D}$	515	90 - 865

[1] Because of the poor fit of the overall regression to the analyses of covariance (r-squared value of 0.43), equations only provide a general relation between hydraulic conductivity and depth. In the equation, hydraulic conductivity (K) decreases with depth (D) when the constant is divided by D; K increases with depth when the constant is multiplied by D. Equations are only useable in the range of depths used in the analysis.

[2] Equations are unreasonable for depths less than actual data values. For layer 3, in area 9, K decreases from 20 feet per day at depth of 2,510 feet to 0.6 feet per day at depth of 3,530 feet. For layer 5 in area 9, K decreases from 84 feet per day at a depth of 740 feet to 1.4 feet per day at a depth of 2,520 feet.

[3] Hydraulic conductivity does not correlate with depth.

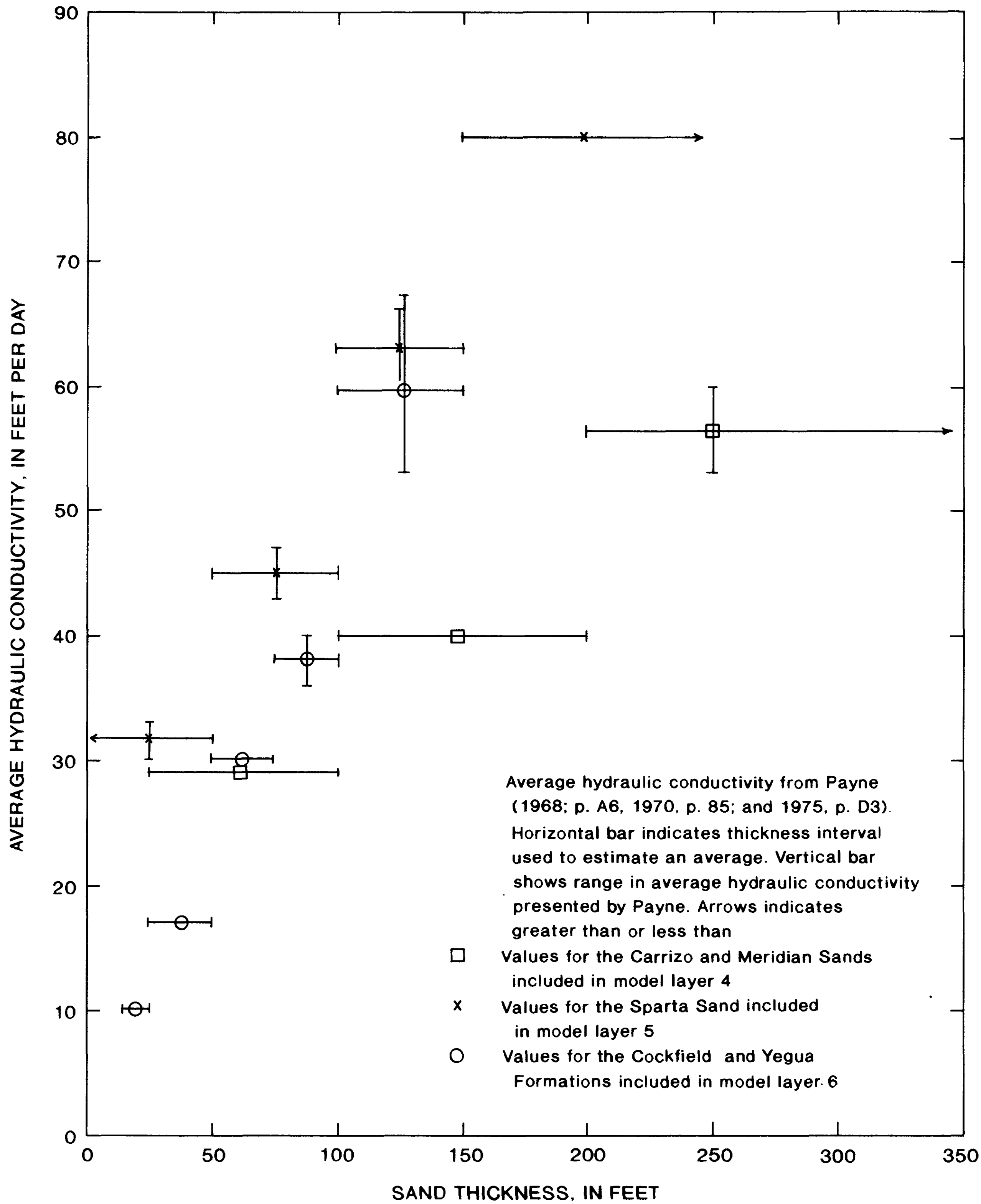


Figure 15.--Relation between average hydraulic conductivity and sand thickness for selected units of the Claiborne Group.

Payne (1968, 1970, and 1975) selected categories of sand thickness for selected units in the Claiborne Group and then averaged the estimated hydraulic conductivities within each category. Results of his analyses are shown in figure 15. In general, average hydraulic conductivity increased as the sand thickness increased in each of the selected units.

Because of the relations described by Payne, an analysis of covariance was performed on the estimates of hydraulic conductivity over the entire study area. The main factors used in the analysis of covariance were area and model layer; the covariate was the thickness of sand beds. The analysis used the $\log_{10}$ transformed hydraulic conductivities for consistency with the other analyses presented in this report. Only estimates of hydraulic conductivity from aquifer-test analyses were used because most estimates of hydraulic conductivity from specific-capacity data did not include an associated estimate of the sand-bed thickness. Sand-bed thickness was estimated for wells by summing the thickness of all sand beds throughout the perforated interval.

A total of 1,195 estimates of hydraulic conductivity had an associated estimate of sand-bed thickness. Estimates of sand-bed thickness ranged from 10 to 1,560 ft with 54 percent of the values being in the range from 26 to 100 ft. The mean thickness was 166 ft with a standard deviation of 239 ft.

Results from the analysis of covariance indicate neither the sand-bed thickness nor the $\log_{10}$ transformation were significant at the probability level of 0.10. Thus, on a regional basis, a relation between hydraulic conductivity and sand-bed thickness could not be determined from the available information. This does not mean that a relation between hydraulic conductivity and sand-bed thickness does not exist, particularly for specific formations such as the Sparta Sand. However, for this large regional study, the results were inconclusive.

SUMMARY

Hydraulic conductivities were estimated from more than 1,500 aquifer tests and more than 5,000 specific capacities from wells drilled into the Tertiary and younger sediments of the Gulf Coast region in the south-central United States. The purpose for the estimates was to determine distributions of hydraulic conductivity of the coarser sediments (sands) for use in the simulation of regional ground-water flow as part of the Gulf Coast Regional Aquifer-System Analysis.

Estimates of hydraulic conductivities range from less than 1 ft/d to more than 1,000 ft/d. The hydraulic conductivities from aquifer test and specific-capacity data have a log normal distribution. The mean $\log_{10}$ of hydraulic conductivity (units are in feet per day) estimated from 1,557 aquifer tests within the study area is 1.74, with a standard deviation of 0.63. The mean $\log_{10}$ value of hydraulic conductivity estimated from specific-capacity data from 5,131 wells is 1.85, with a standard deviation of 0.58.

Because the distributions of hydraulic conductivity are log normal, the geometric mean was used as a representative hydraulic conductivity. The geometric mean for hydraulic conductivities estimated from aquifer tests is 55 ft/d and from specific capacity is 71 ft/d. Excluding estimates of hydraulic conductivities that exceed 1,000 ft/d resulted in a geometric mean of 52 ft/d from aquifer-test analyses and 68 ft/d from specific-capacity data. Values that exceed 1,000 ft/d were excluded because they are considered unreasonable for sediments in the study area. Hydraulic conductivities exceeding 1,000 ft/d are less than 2 percent of the total number of estimates.

Estimates of hydraulic conductivity from aquifer-test analyses and specific-capacity data were combined and the values grouped into 10 model layers that generally correspond to geohydrologic units, and into 9 geographic areas delineated within the overall study area. A two-way analysis of variance was performed on the estimates to determine if the main factors of area and layer were significant in explaining variation in the estimates. Results of the analysis indicate that the main factors of area, model layer, and the interaction of area and layer were significant at a probability level of 0.001. Because the interaction of area and layer was significant in the analysis, comparison of means was done on every area and layer combination using the Duncan multiple-range test.

Results of the comparison of means indicate that the largest geometric means were in model layers 2, 5, and 11 in areas along the eastern part of the study area that included Mississippi River alluvial deposits. The smallest geometric means were generally in layers 3, and 5 through 10 in areas along the western part of the study area. Within each model layer, the geometric mean hydraulic conductivity increased from areas along the western part of the study area to areas in the eastern part near the Mississippi River, which indicates that the deposits near the Mississippi River might be more permeable than elsewhere in the study area.

An analysis of covariance was performed on the estimates of hydraulic conductivity to determine if variations in hydraulic conductivity within each area and layer combination could be explained by well depth. Results of the analysis indicate that depth to the middle of the perforated or screened interval was significant at the probability level of 0.02, and that for a majority of area and layer combinations, hydraulic conductivity decreased as a function of depth. A second analysis of covariance was performed to determine if variation in hydraulic conductivity could be explained by the thickness of sand beds encountered over the perforated interval. Results of this analysis indicate that sand-bed thickness was not significant at a probability level of 0.10. These results were limited to estimates of hydraulic conductivity from aquifer-test analyses because sand-bed thickness was not estimated for wells with only specific-capacity data. Furthermore, the results do not mean a relation can not exist between hydraulic conductivity and sand-bed thickness for individual geohydrologic units in the study area.

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